

Ministry of Energy and Water Resources
Geological Survey of Israel

The timing and the environmental and palaeoclimatic significance of the late Quaternary dune encroachments into the northwestern Negev Desert, Israel



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This work is dedicated to my late father,
(Dr.) Michael Roskin (1940 – 2005)

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ABSTRACT

Palaeoclimatic research based on ice cores, marine and lacustrine sediment cores, and speleothems has mainly focused on past precipitation and temperatures. Dunes serve as palaeoclimate archives and are also unique recorders of past windiness. The vegetated linear dune (VLD) fields which appear in mid-latitude (Central Asia) and mainly in low-latitude areas of the southern hemisphere (Australia, southern Africa, and South America) are partially vegetated, generally extend in a linear fashion and are separated by interdune corridors (valleys). Though currently stable, these VLDs are believed to have formed when still partially vegetated.

Stabilized VLDs extend over 1,300 km² to form the northwestern (NW) Negev dunefield, which comprises the eastern end of the northern Sinai Peninsula – NW Negev Erg. In the past, in different climates and environments, dunes have encroached into the NW Negev desert from northern Sinai.

The thesis identifies the spatial and temporal characteristics of dune encroachments from northern Sinai into the NW Negev desert, their consequent stabilization, and their triggering palaeoclimates and environments. Several sedimentological and methodological issues are also investigated: the relation between dune morphology and sand redness intensity to optically stimulated luminescence (OSL) ages, the significance of dune-sand redness, and the utilization of ground penetrating radar (GPR) for studying linear dune stratigraphy.

As part of a reconnaissance study for targeting sampling sites, dunefield mapping and digital terrain analysis show that VLDs exhibit varying cross-section geometries. Accordingly, the NW Negev dunefield was classified into a dozen geomorphic units. These were merged into three parallel encroachment corridors: north, central, and south that follow the west-east VLD orientation. Primary sampling sites were defined at the western and eastern ends of each corridor to provide a comprehensive set of data for the dunefield's dune and sand properties and to evaluate encroachment ages and rates. Altogether, forty stratigraphic sections from twenty sites were analyzed and sampled. These included exposed dune and interdune sections that revealed important stratigraphy, and drilled dune and interdune sections, ten of them with stratigraphy of interchanging aeolian, fluvial and standing water deposits. GPR profiles approaching 1 km of dune cross-sections and interdune sediments, collected mainly using a 100-MHz antenna, penetrated 5-10 meters and did not identify the dune substrate. Accordingly GPR was not found to be a dependable tool for sampling-oriented identification of stratigraphic units. Over 300 samples were collected of which nearly

all were spectroscopically measured for sand redness intensity: 118 were analyzed for particle-size distribution (PSD), 97 were dated by OSL, and 36 were analyzed for mineralogy.

Negev VLD substrates are mainly calcic sandy (20-72%) to silty loam palaeosols. The VLDs themselves lack palaeosols and only two dunes showed calcium carbonate concretions. Interdunes are found to comprise 1-10 m thick aeolian sand. Dune and interdune sand showed unimodal particle-size distributions, with 75-95% sand, 5-20% silt, whereas clay rarely surpassed 4%. The grain-size modes of the western and eastern sections of the dunefield corridors are generally similar. The dunes are quartz-rich with small amounts of plagioclase. The central encroachment corridor sands are purer than those of the south and north corridor dunes and contain less fines, calcite, and plagioclase.

OSL ages for the Negev sands and sediments date the sediments burial age, i.e., the last time it was exposed to light. The ages were found to be reliable by standard tests for the luminescence behavior of aeolian quartz. Dose recovery tests, recycling ratios and preheat plateaus showed that the sediments are well-suited for the SAR protocol.

The OSL age distribution displayed 3 age clusters; ~24-10 ka, ~2-0.8 ka and 150-10 years that fit into the chronostratigraphic units of the VLD axis and represent the main encroachment and mobilization episodes. OSL dating of exposed sections provided insights into and a general estimation of VLD elongation and accretion dynamics and rates, while spatial and vertical age density dated the dunefield encroachment episodes. Interdune sands have experienced limited accretion and VLDs have undergone minute lateral migration since their deposition. VLDs generally accrete sand along their axis during major mobilization episodes that partially erode the underlying sand to form horizontal unconformable contacts.

Each dune mobilization episode may include several smaller mobilization events. For each episode, the resultant stratigraphic VLD (axis) sand unit represents both mobilization and stabilization ages; the lower sands slightly post-date initial sand encroachment and burial, and the upper sand unit ages are close to sand stabilization time. Though the VLDs did not elongate between mobilization episodes, local reworking, accretion, and erosion of slopes was found—mainly in the crests (dune activation). This process, which was mainly regulated by strong wintertime storm winds, droughts, and vegetative and biogenic crust cover, resets the luminescence signal of the upper dune sands. Northeastern facing slip faces and dunelets superimposed on dune crests imply a dominant recent WSW-SW sand-transporting wind direction, supported by wind measurements, exemplify this current process.

In global terms, the NW Negev dunefield is relatively young. Although sand has been intermittently transported into the Negev for over 100 ka, *dunes* reached the southwestern part of the Negev dunefield only at ~24 ka. The major dune encroachment episode occurred at 18-11.6 ka. Most of the OSL ages cluster at 16-13.7 ka and 12.4-11.6 ka, which are synchronous to the Heinrich 1 and Younger Dryas cold events, respectively. At 16-13.7 ka the main encroachment event deposited the main bulk of the Negev's sand. Aeolian sand deposits lacking dune morphology accumulated in the central encroachment corridor and were subsequently overridden by late Holocene dunes. At 12.4-11.6 ka, dunes reached the easternmost extent of the dunefield. Generalized dune sand transport rates during encroachment episodes differ between corridors; ~25 m/yr around 15 ka in the northern corridor and 10 m/yr in the southern corridor between 23 to 12 ka and ~5-10 m/yr at 12.4-11.6 ka in the eastern part of the central corridor. These estimates incorporate the range of the OSL age errors. However, based on OSL-dated stratigraphy, the rates probably incorporate several shorter rapid incursion events.

Dune encroachment in the southern dunefield dammed wadis, forming short-term standing-water bodies that deposited light-colored loam units. Between these units and dune bases abundant Epipalaeolithic camp sites have been found suggesting a unique connection between these water-bodies and prehistoric man. Spectrally mapped, the exposed surfaces of standing-water deposits show similarities between the southwestern Negev dunefield and the northeastern Sinai in the vicinity of Wadi Al-Arish. It is therefore suggested that Wadi Al-Arish was also blocked during the major dune encroachment, causing extensive upstream and interdune flooding and deposition of fines.

The late Holocene (2-0.8 ka) dune mobilization episode included transverse and VLD incursion in the western part of the central corridor and surface remobilization and possible accretion of 1-2 m of sand in other parts of the dunefield and interdunes, possibly reworking Late Pleistocene sands. As these episodes coincide with the late Roman, Byzantine, and Early Arab periods, dune erodibility may have been assisted by anthropogenic decimation of dune vegetation and destruction of the biogenic crusts by cutting and trampling, respectively. The thick dune sections in the west and intermittent OSL-age clustering also imply periods of high sand-transporting windiness along with sand availability.

Intermittent sand activity and stabilization in the last 150 years ago reactivated dune crests and slopes but dune elongation did not ensue. The OSL ages are consonant with anthropogenic land-use changes. Dune activity due to Bedouin presence and grazing

gradually came to a stop in the early 1950's, and initial dune re-stabilization followed the ban on Bedouin grazing after the re-establishment of the Egypt-Israel border in 1982.

Sand redness, spectroscopically defined by the redness index (RI) ($RI=R^2/(B \cdot G^3)$), reflecting the amount of iron-oxide quartz grain coatings, does not vary greatly across the Negev dune sections and encroachment corridors. No correlation was found between RI intensity (i.e., redness) and the depositional OSL age of the sand. For most stratigraphic sections, RI values and sand sedimentological properties were relatively uniform. Therefore it is probable that either sand rubification developed rapidly following deposition in the Late Pleistocene Negev climate, which was claimed to have been wetter than today (300-350 mm annual rainfall), or that sand grain coating development and consequent rubification have probably been minimal since deposition.

Based on analysis of northern Sinai sand samples, remote sensing, and previous studies, it is suggested that the attributes of sand grain RI were inherited from upwind sources. Sand grain coatings formed at an early diagenetic stage and the sand has had the same redness since its Late Pleistocene aeolian departure from the middle and upper Nile Delta. This, along with radiocarbon dates of Late Pleistocene Nile Delta sands units and prehistoric sites in northern Sinai sands, provide the currently available evidence suggesting that the dune sand originated from the Nile delta and its availability in Sinai was initiated by the exposure of Nile Delta sands during glacial lower Mediterranean sea-levels.

It has been suggested by Enzel, Y. et al. (The climatic and physiographic controls of the eastern Mediterranean over the late Pleistocene climates in the southern Levant and its neighboring deserts. *Global and Planetary Change*, 60(3-4) (2008), 165-192) that Late Pleistocene wintertime Eastern Mediterranean (EM) cyclonic storms, bringing rainfall with fine sand and dust-transporting winds to the northern Negev, were more intense and/or frequent than today. Assuming ample sand supply, this climate model fails to explain the episodic and rapid encroachments of the Negev dunes and their synchronicity to the Heinrich 1 and Younger Dryas cold events. As in the last glacial maximum, these cold events are characterized by increased dust mass as found in Northern Hemisphere ice cores. This dustiness is explained by high entrainment rates in low-latitude dust sources caused by increased windiness due to steeper meridional gradients. The increased windiness probably intensified during EM storms that lasted no more than several days. Since wind velocities for dust entrainment exceed sand saltation velocities, we suggest that the Negev dunes may also have responded to this increased low-latitude windiness. Based on the orientations of the

VLDs, we can infer strong westerly palaeowinds that differ from the current WSW-SW sand-transporting dominating winds, suggesting that Late Pleistocene winds also differed slightly in direction. The Negev VLDs encroached in a generally wetter and mainly windier climate than today: the wetter conditions provided better conditions for vegetation growth, although during windy periods when dunes elongated, dune vegetation cover was partially suppressed. Dunes did not invade the Negev before 24 ka, probably due to lack of sediments reaching into the Negev from Sinai, even though previous glacial periods might have been sufficiently windy for dune transport.

Often-vegetated, low-latitude luminescence-dated dunes, mainly from the southern hemisphere, were found to be active since the LGM and had stabilized by the onset of the Holocene. As with the Negev dunes' synchronicity with the Heinrich 1 and Younger Dryas cold-events, the global dune data was found consistent with the sharp drop in southern and northern hemisphere dust deposition in ice cores. It is therefore suggested that the gradual and intermittent decrease in global windiness between the LGM and onset of the Holocene also determined global dune activity and subsequent quiescence.

Based on the suggested link between global low-latitude dunefield activity and glacial and cold-event induced windiness there is no expectation that these dunefields will be highly re-mobilized in the near future due to global warming. Droughts may decimate vegetation but the lack of strong winds will probably not cause substantial encroachment of Late Pleistocene magnitudes. Current winter storms in the northern Negev cause local reworking of the upper VLD section as identified by the OSL-dated GPR profiles. Therefore, intense EM storms may cause limited dune elongation of a thin sand section. Local and short-term climate changes in the form of increased windiness especially after drought may induce increased dune activity and the partial dune-damming of wadis.

The recurring discontinuous Late Quaternary aeolian sedimentation pattern found in OSL-dated VLDs provides new and important chronological and sedimentological insight into significant dune mobilization and stabilization processes while demonstrating the sensitivity of dunes located along the (northern) fringe of the sub-tropical desert belt to climate change (wind) and sediment supply. The suggested link between global reductions in cold-event windiness and low-latitude dune stabilization episodes emphasizes the dominant effect of windiness on major dune mobilizations in low-latitude dunes even if they are partially vegetated.

ABBREVIATIONS, DICTIONARY and AEOLIAN GLOSSARY

Abbreviations

VLD: Vegetated linear dune

DF: Prefix for Negev dunefield samples

EM: Eastern Mediterranean

GIS: Geographic Information Systems

a, yr: An age measured from the present

ka: An age measured from the present (in thousand years)

kyr: Thousand years, a time interval

LGM: Last Glacial Maximum

OSL: Optically Stimulated Luminescence

SAR: Single Aliquot Regenerative dose protocol for OSL dating

IRSL: Infrared Stimulated Luminescence

GPR: Ground penetrating radar

TL: Thermoluminescence

XRD: X-Ray Diffraction

RGB: Red-Green-Blue color scheme

Dictionary

Nahal: Ephemeral stream in Hebrew

Rama, Ramat: Plateau in Hebrew

Wadi: Ephemeral stream in Arabic and Hebrew

Gebel: Mountain or mountain ridge in Arabic

Aeolian glossary

Aeolian-sedimentological terms are mainly derived from other geological fields, mainly fluvial and sedimentological ones. As an aeolian glossary is absent, the following terms in thematic order, have been concisely described, in partial conjunction with some studies.

Linear dune: Also known as a longitudinal dune. A sand body of considerable length that elongates or have elongated in the past, relative straightness, parallelism, regular spacing and low ratio of dune to inter-dune areas (Lancaster, 1982; after Pye and Tsoar, 2009).

Seif (sword in Arabic) dune: An unvegetated and usually sinuous-shaped linear dune with a sharp crest.

Vegetated linear dune (VLD): Also known as sand or parallel ridges. A linear dune, usually hosting a 10 % to 20 % vegetation cover.

Transverse dune: Linked, sinuous-crested barchan dunes that develop in conditions of increased sediment-supply and unidirectional effective winds (after Pye and Tsoar, 2009).

Barchan dune: A crescent-shaped dune that develops under unidirectional winds and low supply of sand (after Pye and Tsoar, 2009).

Dune activation: Sand and dune activity.

Dune mobilization: Sand and dune activity and transport, sufficient to advance or elongate the dune in accordance to its type.

Dune stabilization: A dune or dune section not undergoing activation and/or mobilization because of vegetation and/or crust cover.

Dune incursion: Mobilization and elongation of dunes over new ground.

Dune encroachment: Massive mobilization and elongation of dunes over new ground.

Dune accretion: Net accumulation of sand upon a dune, usually one that has an upwards sedimentation patterns such as VLDs.

Dune construction/buildup: Vertical or/and lateral increase of a (linear) dune cross-section.

Downdune/updune: Location upon a dune axis relative to its elongation orientation.

Downwind/upwind: Spatial location (in an aeolian environment) relative to the dominant (sand-transporting) wind direction.

Sand sheet: A flat-surface of aeolian sand or horizontally-bedded sedimentological body of aeolian sand.

Dunefield (dune field): Sand bodies, similar to sand seas, but smaller.

Erg (sand sea): Large basins in deserts that are mantled by sand dunes.

Windiness: Increased strength and prevalence of strong winds.

Gustiness: Strength and prevalence of strong winds (after McGee et al., 2010).

Dunal: Attributed to dunes.

Erosivity: The potential for sediment transport (Chase and Brewer, 2009).

Erodibility: The availability of sediment for deflation (Chase and Brewer, 2009).

Last Glacial Maximum: A period in the Earth's climate history when ice sheets were at their maximum extension, between 26,500 and 19,000 years ago (after Wikipedia).

TABLE OF CONTENTS

Acknowledgments.....	i
Abstract.....	iii
Abbreviations, dictionary and aeolian glossary.....	xiii
Table of contents.....	x
List of figures.....	xiv
List of tables	xv
1. Introduction.....	1
1.1. The significance and state of the art of dunefield research.....	1
1.2. The study area.....	7
1.3. Research hypotheses and goals.....	11
1.4. Thesis outline.....	11
2. Methodology and methods.....	13
2.1. Reconnaissance work.....	13
2.2. Sampling site strategy, selection and stratigraphy.....	13
2.3. Sampling.....	15
2.4. Ground-penetrating radar (GPR) profiles.....	17
2.4.1. Background.....	17
2.4.2. GPR survey goals and site locations.....	22
2.4.3. GPR profiling and processing.....	22
2.5. Particle size distribution and moisture content analysis.....	23
2.6. Relative mineral abundances.....	24
2.7. Optical stimulated luminescence (OSL) dating.....	24
2.7.1. Introduction.....	24
2.7.1.1. Background.....	24
2.7.1.2. Equivalent dose measurement.....	26
2.7.1.3. Dose rate measurements.....	26
2.7.2. OSL age measurements and age determination.....	27
2.7.2.1. Sample preparation.....	27
2.7.2.2. Equivalent dose determination.....	28
2.7.2.3. Dose rate determination.....	28
2.7.2.4. Age calculations.....	31
2.8. Spectroscopic analysis of sand grain redness.....	31
2.8.1. Background.....	31
2.8.1.1. The temporal significance of sand redness.....	31
2.8.1.2. Spectroscopy of sand redness.....	32
2.8.2. Spectroscopic measurement.....	33
2.9. Landsat image processing.....	34
3. Age, origin and climatic controls on vegetated linear dunes in the northwestern Negev Desert (Israel)	35
3.0. Abstract.....	36
3.1. Introduction.....	36

3.2.	The research area.....	41
3.3.	Methods.....	43
3.3.1	Field methods.....	43
3.3.2	OSL dating.....	46
3.2.1.1.	Sample preparation.....	46
3.2.1.2.	Dose rate determination.....	47
3.3.3	Particle-size distribution and mineralogy.....	47
3.4.	Results.....	47
3.4.1.	Dune morphology and field relations.....	47
3.4.2.	VLD stratigraphy and internal structure.....	49
3.4.3.	Particle-size distribution and mineralogy.....	58
3.4.4.	OSL ages.....	58
3.4.4.1.	Analytic OSL precision	58
3.4.4.2.	Comparison to previous dates and ages.....	61
3.4.4.3.	OSL dated landforms types.....	62
3.4.4.4.	OSL age clustering.....	63
3.5	Discussion.....	63
3.5.1	Aeolian sand incursion episodes.....	63
3.5.1.1	Earliest evidence for aeolian sand deposition.....	63
3.5.1.2	Initial dune incursion.....	68
3.5.1.3	The main dune incursion.....	69
3.5.1.4	Dune damming in the southern incursion corridor.....	73
3.5.1.5	Summary of the late Pleistocene events.....	77
3.5.1.6	Late Holocene dune activity.....	77
3.5.2	The temporal and spatial aspects of sediment supply for dune into the Negev.....	79
3.5.2.1	The inferred source and dynamics of the northern Sinai dunefield.....	79
3.5.2.2	The chronology of sand transport in northern Sinai.....	81
3.5.2.3	Last glacial luminescence-dated global linear dune activity.....	83
3.6	Conclusions.....	84
3.7	Acknowledgments.....	85
4.	Do dune sands redden with age? The case of the northwestern Negev dunefield, Israel.....	86
4.0.	Abstract.....	87
4.1.	Introduction.....	87
4.1.1.	Sand color.....	87
4.1.2.	Spectroscopy of sand redness.....	89
4.2.	Study goals.....	90
4.3.	Study area.....	92
4.4.	Field and laboratory methods.....	95
4.4.1.	Sampling methods.....	95
4.4.2.	Spectroscopic measurements and indices.....	102
4.4.3.	Spectroscopic indices.....	102
4.4.4.	Landsat imagery.....	103

4.4.5. OSL dating laboratory procedures.....	104
4.4.6. Sedimentology.....	104
4.5. Results.....	105
4.5.1. Redness index properties.....	105
4.5.2. Sedimentology and RI.....	109
4.5.3. Sinai sand data.....	109
4.6. Discussion.....	110
4.6.1. Controls of in-situ sand rubification.....	110
4.6.2. Spatial and vertical distribution of sand redness.....	112
4.6.3. Sinai sand redness.....	113
4.6.4. Nile Delta sand-grain coatings.....	115
4.7. Conclusions.....	117
4.8. Acknowledgments.....	117
5. Palaeoclimate interpretations of Late Pleistocene vegetated linear dune mobilization episodes; evidence from the northwestern Negev Desert, Israel.....	118
5.0. Abstract.....	118
5.1. Introduction.....	120
5.1.1. Dunes as palaeoclimate records.....	120
5.1.2. Episodes of northwestern Negev dunefield activity.....	122
5.1.3. Northern Negev Late Pleistocene palaeoclimate interpretation.....	124
5.1.4. LGM – Holocene transition climate changes.....	128
5.1.5. Study goals.....	130
5.2. Northwestern Negev dune encroachment episodes.....	130
5.3. Negev vegetated linear dune dynamics, structure and chronology.....	134
5.3.1. Vegetated linear dune formation.....	134
5.3.2. Negev VLD mobilization-stabilization episodes.....	135
5.3.3. Rapid accretion and elongation.....	138
5.3.4. The change in Negev sand-transporting wind orientations since the Late Pleistocene.....	139
5.4. Regional palaeoclimate records.....	140
5.4.1. Speleothems and Lake Lisan records.....	140
5.4.2. Aeolian sand records.....	143
5.4.3. Northern Negev loess records.....	144
5.4.4. Northern Negev prehistoric sites.....	145
5.4.5. Summary: Late Pleistocene VLD mobilization-stabilization environment...	146
5.5. The global palaeoclimate connection.....	147
5.5.1. Coincidence of GISP H1 and YD dust fluxes with Negev dune mobilization-stabilization episodes.....	147
5.5.2. Post-LGM - Holocene global luminescence-dated dune mobilization and stabilization.....	148
5.5.3. Post-LGM – Holocene palaeoclimatic control of NW Negev windiness...	150
5.6. Conclusions.....	151

6. Summary.....	153
6.1. Synopsis.....	153
6.1.1. Overview.....	153
6.1.2. Methods for dune studies.....	154
6.1.2.1. Exposed VLD stratigraphy	154
6.1.2.2. OSL age performance	154
6.1.2.3. Sand age- redness ratio.....	154
6.1.2.4. GPR applicability for VLDs	155
6.1.3. Evolution of the NW Negev dunefield.....	155
6.1.4. VLD dynamics.....	156
6.1.5. The relationship of the Negev dune ages to loess ages.....	159
6.1.6. Negev dune-driving palaeoclimates.....	162
6.1.7. Evolution of the Sinai-Negev erg.....	163
6.1.8. Global dune-driving windiness.....	164
6.2. Overview research contribution.....	164
6.2.1. General.....	164
6.2.2. Detail.....	165
6.2.2.1. Research methods.....	165
6.2.2.2. Late Quaternary landscape evolution and palaeoclimate implications.....	166
6.2.2.3. Vegetated linear dune structure, dynamics and sedimentology.....	166
6.3. Future research.....	167
7. References.....	168
 Appendices	 188
Appendix A. Topographic and ground-penetrating radar sections.....	189
A.1. Dune topographic cross-sections.....	189
A.2. Northwestern Negev dunefield sampling site information table.....	190
A.3. Ground-penetrating radar site data table.....	191
A.4. Geomorphological map of the BM site.....	193
A.5. BM VLD west GPR profile and interpretation.....	194
A.6. BM VLD east GPR profile and interpretation.....	195
A.7. BM playa east GPR profile and interpretation.....	196
A.8. Tzidkiyahu site orthophoto.....	197
A.9. Tzidkiyahu west-east GPR profiles and interpretation.....	198
A.10. Tzidkiyahu north-south GPR profiles and interpretation.....	199
A.11. Retamim orthophoto, GPR profile and interpretation.....	200
Appendix B. Sedimentology data.....	201
B.1. Particle-size analysis result table.....	201
B.2. Dune moisture profiles.....	205
B.3. XRD mineralogy results of the NW Negev dunefield sands.....	205
 Appendix C. Remote sensing procedures.....	 206
C.1. Pre-processing of Landsat images – radiometric and atmospheric corrections.....	206

C.2. Resampling ASD Redness index to Landsat TM5 spectral resolution.....	209
C.3. Classified Landsat TM5 mineral enhancement	209

Hebrew abstract תקציר	א
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LIST OF FIGURES

1.1 World sand sea map.....	3
1.2 Examples of damage by sand mobilization.....	3
1.3 Dust storm in the Negev dunefield.	3
1.4 Landsat image of the Sinai-Negev erg	4
1.5 Landsat image of northeastern Sinai and the NW Negev dunefield.....	4
2.1 Geomorphic unit classification map.	16
2.2 Dune slope distribution of geomorphic units.....	16
2.3 Negev dunefield geomorphic unit map and dune encroachment corridors and sampling site map.....	18
2.4 Field drilling methods.....	20
2.5 OSL and GPR field work methods and spectroscopic measurement setup.....	21
2.6 Energy level diagram illustrating the luminescence process.....	29
2.7 OSL SAR protocol.....	29
2.8 Dose recovery ratio.....	30
3.1 Sinai-Negev erg and northwestern Negev dunefield maps.....	39
3.2 Dune cross-sections.....	44
3.3 Sediment texture, particle size distribution and mineralogy.	48
3.4 Luminescence signals, dose response curves, preheat plateaus, relative probability...	60
3.5 Stratigraphic logs and OSL ages.....	65
3.6 Time-slice maps.	66
3.7 Geomorphological map of the Halamish region.....	70
3.8 Relative probability plot of the OSL ages.....	71
3.9 A compilation of stratigraphic logs along a transect of the western dunefield.	75
3.10 Spatial and temporal evolution of the NW Negev dunefield.....	76
4.1 Regional map of the Sinai-Negev erg and northwestern Negev dunefield.....	91
4.2 Reflectance spectra of sand samples from different parts of the Negev dunefield.....	104
4.3 The Haluzit 1 exposed section.	105
4.4 Plot of redness index vs. depth of sections per dune encroachment corridor	106
4.5 Redness index maps of the northern Sinai and NW Negev dunefield.....	107
4.6 Scatter-plot of the OSL ages vs. redness index.....	108
5.1 Regional and synoptic map of the Sinai-Negev erg and Levant.....	121
5.2 Late Pleistocene development of the NW Negev dunefield.....	125
5.3 Histogram of Late Pleistocene Negev and global dune ages and sediment records...	129
5.4 Analysis of the two NW Negev main encroachment episodes.....	132
5.5 VLD formation and stratigraphy.....	132
5.6 Superimposed dunelets on Sinai and Negev linear dunes.....	141
6.1 The December 10-12 th , 2010 Eastern Mediterranean cyclone storm.....	160

6.2 Eastern Mediterranean Cyprus cyclone track density.....	161
6.3 Time slice palaeogeographic maps of the northern Nile Delta	161

LIST OF TABLES

3.1 Previous ages in the NW Negev dunefield.	40
3.2 Morphological characteristics of the three dune incursion corridors.	45
3.3 Optically stimulated luminescence (OSL) ages.	52-57
3.4 Comparison between OSL ages and previous ages.	61
4.1 Main sedimentological, RI, and OSL results	96
4.2 Redness index band data and their relationship to Fe sand-grain coatings.....	103

1. INTRODUCTION

1.1. The significance and state of the art of dunefield research

Dunes are wind driven (aeolian) sand accumulations and can be considered the most common and dynamic terrestrial landform. Dunes that cover over 10,000 km² are known as sand seas or ergs whereas smaller dune bodies are referred to as dune fields or dunefields. Dune bodies can largely be divided into two main types; coastal and inland. Coastal dunes are usually spatially limited and relatively non-complex. Inland dunes cover between 25% and 40% of the desert regions found in North Africa, Asia, and Australia, and approximately one-third of the arid regions of the Earth (Lancaster, 2007) though the extent, classification of global dunefields is not complete (Fig. 1.1).

Inland dune mobilization and encroachment causes significant environmental change, often interpreted as desertification processes. Sand and dune incursions physically erode soil leading to sand and dust hazards and decimating crops and infrastructure. Local sand movement, let alone regional incursions, are also known to inflict severe damage to infrastructure (Hagedorn et al., 1977; Tsoar and Zohar, 1985; Abdel-Galil et al., 2000) (Fig. 1.2). Dune incursions often block and divert streams, resulting in substantial morphological changes such as ponding and consequent sediment deposition (McCauley et al., 1982; Magaritz and Enzel, 1990; Kusky and El-Baz, 2000; Krapf et al., 2003). Devegetation causing sand exposure, and soil deflation due to strong winds (Fig. 1.3), may also precede sand encroachment (Tsoar, 1995). Downwind loess deposits accumulate in response to dust release, suggested to result from sand-grain abrasion during dune encroachment (Crouvi et al., 2008, 2009, 2010; Enzel et al., 2010).

Dunes upon the surface of Earth are geologically young landforms and dunefield chronology infrequently exceeds 80,000-100,000 BP (Stokes et al., 1997; Preusser et al., 2002; Lancaster and Tchakerian, 2003; Fitzsimmons et al., 2007). Disregarding methodological biases and dating limitations, dune ages are usually younger than the last glacial cycle (Fitzsimmons et al., 2007; Telfer and Thomas, 2007). Some dunefields were re-activated even in historical times, such as in the central and southwestern USA (Muhs and Holliday, 1995; Arbogast, 1996; Arbogast and Muhs, 2000), in some cases subsequent to changes in wind orientation (Arbogast and Muhs, 2000).

In order to interpret dunefield evolution we need to understand the sedimentological properties of dune encroachment from the level of the sand-grain through the single-dune to the dunefield level, and to have accurate chronological control of activity from the single-

dune level to the dunefield level. This is crucial for predicting future dune encroachment and increased dustiness.

There are three conditions for dune development: sufficient available and transportable sediment; conditions for aeolian sediment accumulation such as moderate topography, and winds above the threshold level for sand transport (5-6 m/s)(Pye and Tsoar, 2009). When wind power diminishes in an active dune environment, dunes usually become stabilized, and even in arid environments, vegetation will appear (Lancaster, 1995; Lancaster and Tchakerian, 1996; Pye and Tsoar, 2009).

As young landforms, dunes should be easy subjects for landform evolution interpretation. However, the methodologies for dating dunes and dunefields are rapidly developing. Dune morphology (O'Connor and Thomas, 1999; Fitzsimmons, 2007; Fitzsimmons et al., 2007) and spatial trends (Kocurek and Ewing, 2005) were recently suggested as important for analyzing dune and dune field evolution respectively though this has not been proven with luminescence ages. Dune stratigraphy and internal structure (Tsoar, 1983a; Nanson et al., 1992; Lancaster and Tchakerian, 1996), supported by dating (Nanson et al., 1992; Thomas and Shaw, 2002), are important for studying dune development though exposed dune sections are rare due to their friable nature. Most works have relied on luminescence ages from cored sections and have not sampled following stratigraphic classification of an exposed dune section and the dynamic character of dunes often alters geomorphic features and resets luminescence signals. Dune sand often lacks palaeosols and organic material that are important features for stratigraphy and (radiocarbon) dating. The main sedimentological factors relevant to studying dunes are sand geochemistry and mineralogy (Muhs, 2004) and sand grain color, (Gardner and Pye, 1981; Wopfner and Twidale, 1988; Goudie et al., 1993; White et al., 1997) which has not been analyzed in regard to luminescence age. .

There are many unanswered questions regarding the dynamics and control of dunes, and specifically linear dune development. Many researchers accept linear dunes to be extending dunes (O'Connor and Thomas, 1999; Telfer, 2011), including linear dunes in the NW Negev (Tsoar and Moller, 1986; Ben David, 2003; Tsoar et al., 2004; Tsoar et al., 2008), though other works suggest that linear dunes are produced by parallel helical vortices in unidirectional wind regimes (Bagnold, 1953; Folk, 1971) or wind-rifting of sand from a proximal source (Hollands et al., 2006). Two main types of linear dunes are recognized, unvegetated and vegetated (VLD) (Tsoar and Moller, 1986; Tsoar, 1989; Pye and Tsoar, 2009), though this distinction has not been fully accepted. Large proportions of the low to mid-latitude dune bodies in Australia (Nanson et al., 1992; Fitzsimmons et al., 2007; Cohen

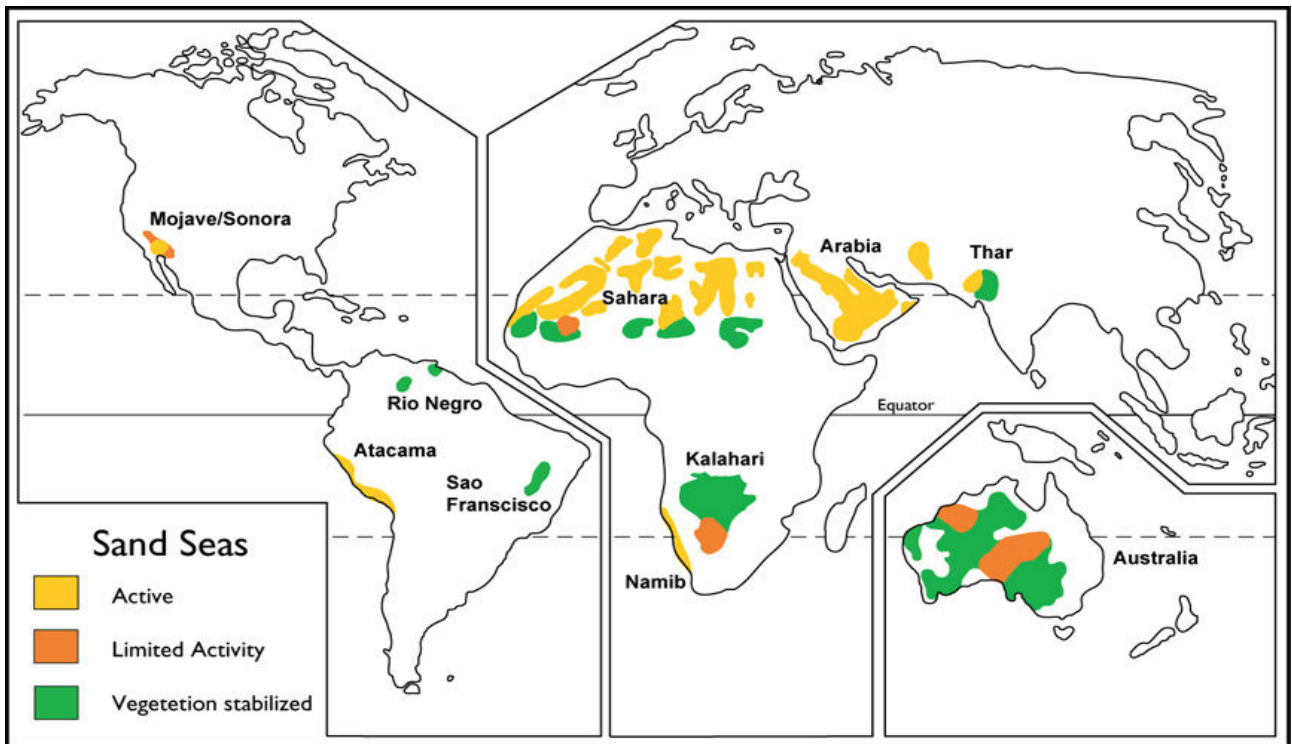


Figure 1.1 Global map of sand seas (ergs) as online for the Digital Dune Atlas (<http://inquadunesatlas.dri.edu/background.htm> after Thomas, 1997). There is at this point, no official published global dune map. This map demonstrates the recent progress but it still lacks the Sinai-Negev Erg (i.e. this work), several Chinese ergs (i.e. Sun and Muhs, 2007) and Argentinian ergs (i.e. Tripaldi et al., 2007) .



Figure 1.2 Examples of damage from sand and dune mobilization in Africa.

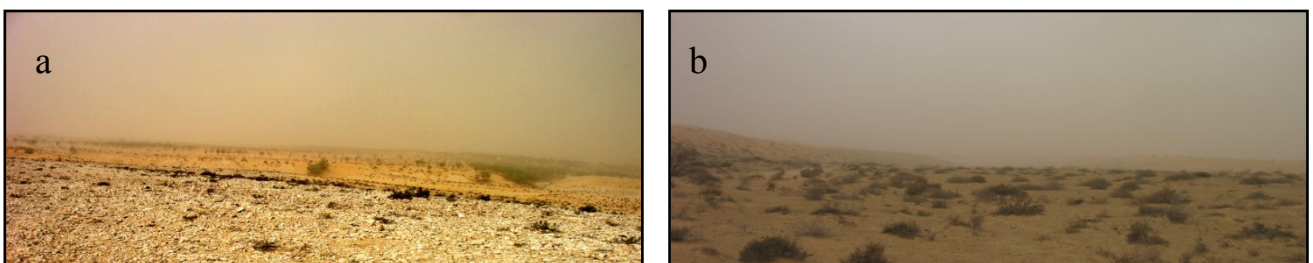


Figure 1.3 Photographs of dust transport during winter storms above the eastern edge (a) and interdune corridor (b) of the northwestern Negev dunefield.

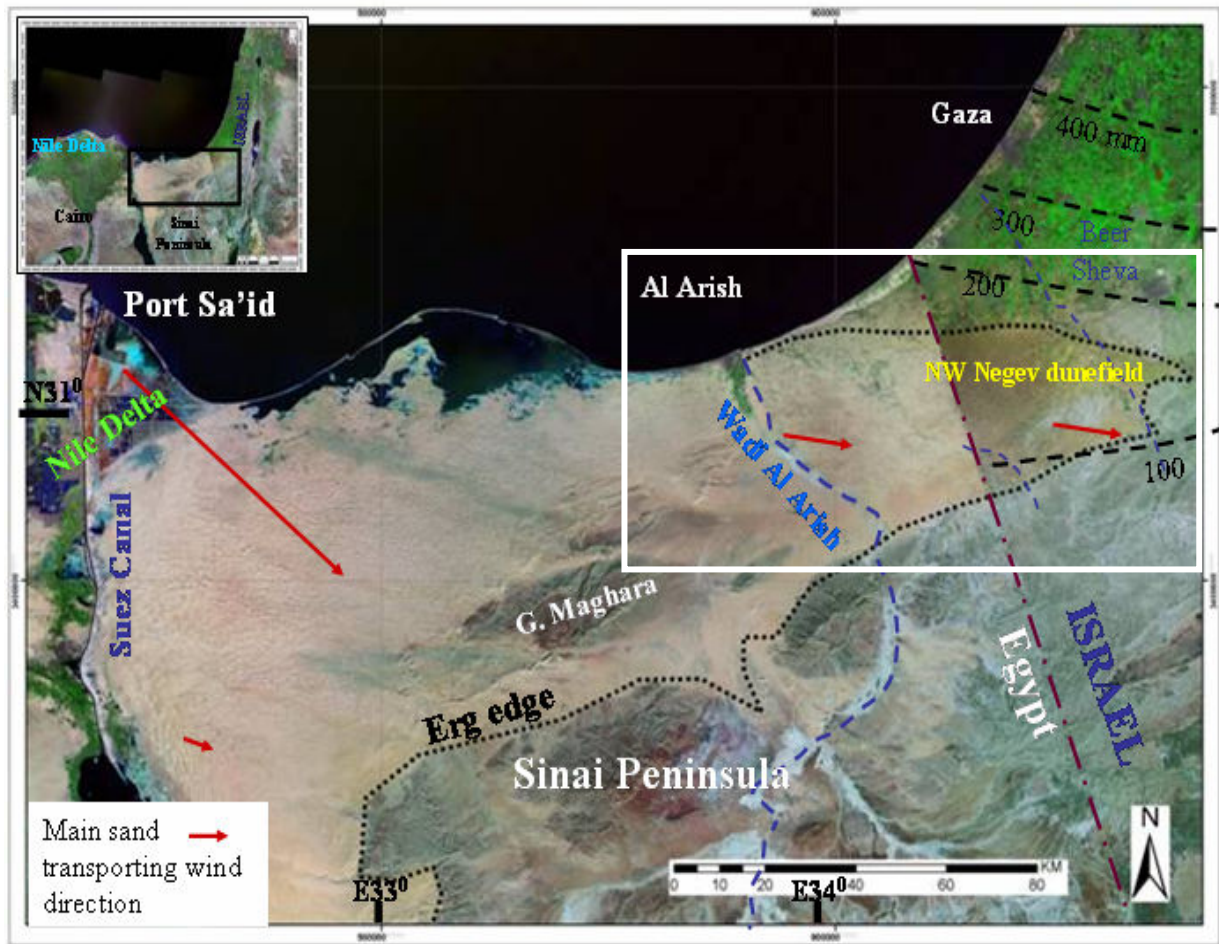


Figure 1.4 False Landsat composite image of the Sinai-Negev erg and the northwestern (NW) Negev dunefield at its eastern edge. The dashed black lines in Israel are isohyets of annual precipitation in mm. The white box depicts figure 1.5.

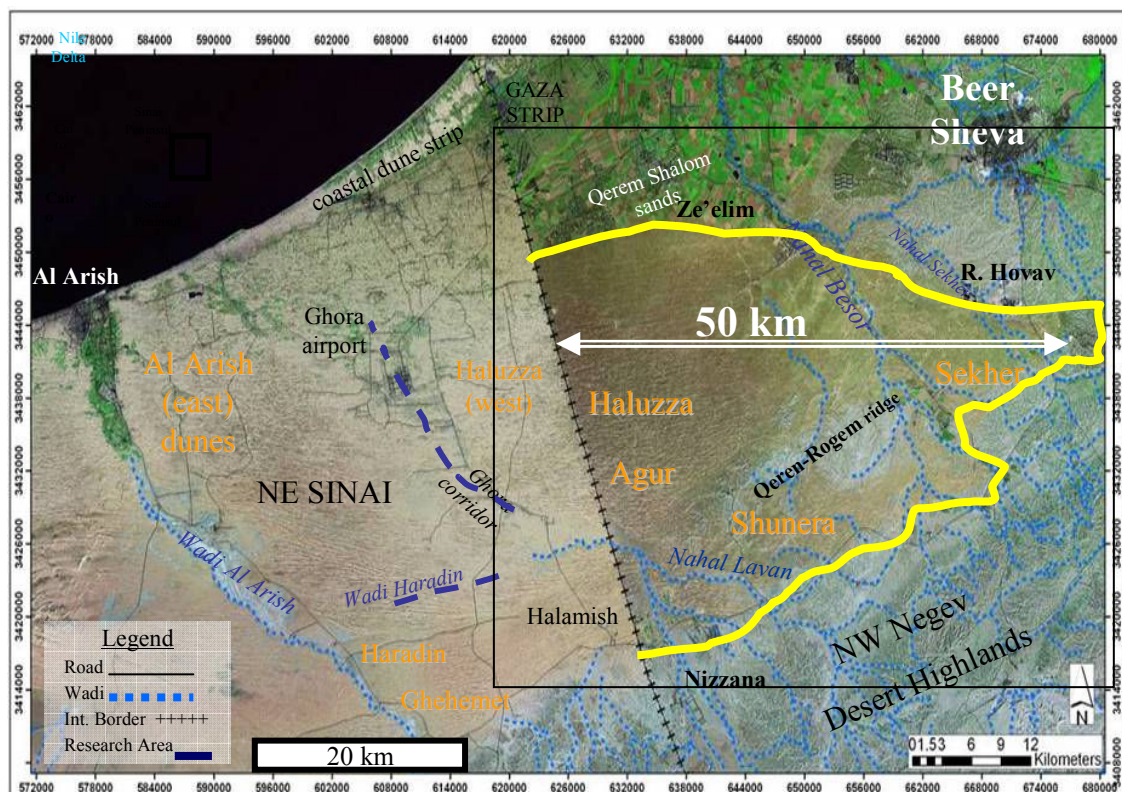


Figure 1.5 False Landsat composite image of the northeastern Sinai-Negev erg and the northwestern (NW) Negev dunefield. Main dune regions are marked in orange. Black box outlines figure 2.1.

et al., 2010), southern Africa (Telfer and Thomas, 2007), and South America (Tripaldi and Forman, 2007; Tripaldi et al., 2011) and Central Asia (Maman et al., 2011) comprise VLDs that are currently stable in regions with low wind power. Accordingly, in past climates and environments, the dunes, either with or without vegetative cover, were active until stabilizing at their current position. Accordingly, a better understanding of the processes of VLD formation and elongation, whether the VLD is partly or fully vegetated, is needed. Vegetation cover is assumed to be the main reason behind VLD formation, for which several theories have been proposed. In contrast to the sinuous elongation that occurs with linear, unvegetated seif dunes, VLDs are thought to lengthen along straight lines and approximately in the direction of the prevailing wind (Tsoar and Moller, 1986; Tsoar, 1989; Tsoar et al., 2008). The "tuning fork" pattern (Tsoar et al., 2008) or Y-junctions (Kocurek and Ewing, 2005) is common in VLDs but missing from seifs (Tsoar and Moller, 1986; Tsoar et al., 2008). This coalescence, though not clearly understood, has been attributed to deflection by cross-winds of the extreme of the dune ridge during the elongation process in order to preserve dune spacing (i.e. Tsoar et al., 2008 and references within). It seems, therefore, that VLDs have always been vegetated to some degree, though probably more sparsely during colder periods when wind power was greater (Hesse and Simpson, 2006; Hollands et al., 2006; Cohen et al., 2010).

The application of luminescence dating to aeolian sands in desert regions has revolutionized our understanding of the dynamics of these systems on centennial to millennial time scales. The availability of increasingly precise numerical ages for periods of aeolian deposition has provided information on rates of dune migration and accumulation of sand and supported new models of dune development (Lancaster, 2008). As a result of advances in optically stimulated luminescence (OSL) single aliquot regenerative-dose (SAR) protocols (Murray and Wintle, 2000) improvements have occurred in the quality and quantity of OSL-based age estimates and in the chronological control of episodes of dunes and dunefield activity (e.g., Fitzsimmons et al., 2007; Telfer and Thomas, 2007). Despite the greater accuracy of OSL in dating dunes, however, the reliability of OSL ages in representing episodes of dune activity and their palaeoclimatic significance has been questioned on several grounds: OSL dating cannot pinpoint the onset of dune activation (Nanson et al., 1992; Telfer and Thomas, 2007; Fitzsimmons and Telfer, 2008), and sampling does not always include a full dune section due to technical limitations (Bateman et al., 2003) and so is often not systematic and lacks sufficient spatial resolution (Telfer et al., 2010). Moreover, due to their dynamic and erosive nature dunes present discontinuous records (Telfer and Thomas, 2007) and we cannot always

distinguish between episodic and continuous sedimentation (Bateman et al., 2003; Telfer and Thomas, 2007; Chase, 2009).

The relationship between global climate change in terms of windiness, dustiness, and dune activity has profound palaeoclimate and future climate implications. However, the study of global palaeoclimate change has relied mainly on temperature and rainfall fluctuations as recorded in ice, marine and lacustrine cores, the traditional proxies of choice due to their high resolution, sensitivity, continuity, and ability to archive extensive areas of Earth. On land, speleothems have become the leading terrestrial palaeoclimate proxies of palaeotemperatures, water vapor sources, and rainfall amounts (Bar-Matthews et al., 1999) as they comprise direct, environmental records of in-situ rainfall (Enzel et al., 2008).

Terrestrial aeolian loess deposits are relatively continuous palaeoclimatic proxies of glaciogenic and desert dust transport and deposition (Chen et al., 2003; Muhs et al., 2008). However, their palaeoclimatic interpretations are complicated due to their complex sedimentologies, varying grain-size distributions, mineralogy from mixed-transport distances and sources, and post-depositional processes (Kohfeld and Harrison, 2001) such as pedogenesis (Jacobs and Mason, 2007). Although high resolution dating of global and local windiness has been less studied, recent works suggest that changes in dust deposition are driven by global changes in wind gustiness (McGee et al., 2010). This is based on dust proxies in ice, marine, and lacustrine cores providing continuous records of changes in global dustiness. It has not been tested for dunes, however.

Fluctuation from mobilization to stabilization in dunes and dunefields indicates their sensitivity to palaeoclimate, climate, and human-inflicted environmental change. Thus, dunes and dunefields are often valuable archives of past climatic changes and conditions (Sarnthein, 1978; Lancaster, 1995; Muhs and Holliday, 1995; Lancaster, 2008), weather systems, precipitation, pluvial cycles, and sea-level fluctuations (Sarnthein, 1978; Goring-Morris and Goldberg, 1990; Lancaster, 1990; Lancaster, 1995; Lancaster, 1999; White et al., 2001; Glennie and Singhvi, 2002; Preusser et al., 2002). In general, mobile inland dunes have been used as indicators of arid conditions (e.g., Sarnthein, 1978; Munyikwa, 2005; Hesse and Simpson, 2006; Lomax et al., 2011). Dune mobilization thresholds linked to reduced precipitation and evaporation has been shown to exist in mid-latitude dunes (Muhs and Holliday, 1995). Earlier models suggested that dunes were controlled by both precipitation and windiness (Lancaster, 1988). However, more recent models show dune activity to be primarily the result of wind power, so that exposed dunes can be mobilized even in humid climates with annual precipitation well above 1000 mm (Tsoar, 2005; Chase, 2009; Tsoar et

al., 2009; Yizhaq et al., 2007, 2009). Zhou et al. (2009) suggest that both sand and loess deposits record episodes of activity during the Heinrich 5 and Younger Dryas cold-events. However, to the best of my knowledge, palaeoclimatic studies of dune bodies have failed to fully demonstrate the connection between global wind power and dune activity.

The links between linear dune formation dynamics, internal dune structure, chronological interpretation of luminescence ages, and the spatial chronology of an entire dunefield as a basis for palaeoclimatic reconstruction, have not yet been fully addressed. As a result, this study will investigate the palaeoclimate of the northwestern (NW) Negev dunes (Fig. 1.4) by building a temporal and spatial chronological framework of the NW Negev dunefield evolution and providing a chronostratigraphic interpretation of the dunes and their sedimentologies.

1.2. The study area

The general study area consists of the northern Sinai Peninsula- NW Negev erg (Sinai-Negev erg). The northern Sinai region has been and still is a transitional area between the Nile Delta of Egypt and the Levant (Israel) in many respects: geologically, culturally, geopolitically and geomorphologically. At present the erg is geopolitically divided by the Egypt-Israel border (Fig. 1.4). Unfortunately, sand sampling was not possible owing to restricted access to the Egyptian section of the erg and research on this section had to be based on remote sensing, archive samples, and data from previous research. Fieldwork was thus restricted entirely to the NW Negev dunefield. Located as it is at the downwind end of the erg, the dunefield is considered a suitable location for the study of dunefield evolution.

The source of the northern Sinai dunes is believed to be the Nile Delta (Goring-Morris and Goldberg, 1990; Tsoar, 1990; Hunt, 1991; Amit et al., 2011) though this has not been proven. Northern Sinai is comprised mainly of active and sparsely vegetated linear and seif dunes (Tsoar, 1995; Misak and Draz, 1997; Abdel-Galil et al., 2000; Rabie et al., 2000) that can be remotely imaged directly from space. The dunes extend in a general west-east orientation towards the NW Negev. Luminescence dating of the Sinai dunes has not been carried out.

The NW Negev dunefield (N30°/E34°) covers approximately 1,300 km² and has a generally isosceles triangle shape with its base parallel to the Egypt-Israel border and its eastern point at Ramat Beqa, an incised plateau composed of Lower Eocene carbonates (Avedat Group) (Zilberman, 1982), gently rising 10-50 m above the dunefield. The dunefield is divided by the Qeren-Rogem anticlinal ridge (Qeren Ridge) that trends WSW-ENE and protrudes 50-150 m

above the dunes. The ridge is the most northerly exposure of the Northern Negev Syrian-Arc anticlinal system (Zilberman, 1982, 1991). It is composed of the Avedat Group carbonates and is dissected, mainly on its northwest flanks, by steep, short, small drainage systems that are in turn dammed by dunes (Tsoar, 1983; Enzel et al., 2010).

The portion of the dunefield south of the Qeren Ridge fills an east-west synclinal depression and is locally bounded by wadis, Eocene chalk buttes and ridges. The dunes intercept and fill several wadis from the south (Blumberg et al., 2004). Dissected surfaces underlain by loamy sediments are evidence for palaeolakes created by dunes damming the wadis (Magaritz and Enzel, 1990; Harrison and Yair, 1998; Ben-David, 2003; Blumberg et al., 2004) and larger drainages with Late Pleistocene flood plains (Zilberman, 1992).

The main aeolian sand body lies north of the Qeren Ridge. It covers a gently seaward-sloping landscape that was established by the receding Pliocene shoreline and later covered by a sequence of Pleistocene calcareous loam palaeosols (Bruins and Yaalon, 1979; Zilberman et al., 2007). These sandy units overlay Senonian Mt Scopus Group soft marls and chalks (Emog, 1986).

Previous dune research has concentrated on the eastern, southern and mainly southwestern section and edges of the dunefield (Goring-Morris and Goldberg, 1990; Magaritz and Enzel, 1990; Zilberman, 1991; Harrison and Yair, 1998; Ben-David, 2003; Blumberg et al., 2004; Ben-David and Yair, 2008), depicted in figure 3.1c and table 3.2 of chapter 3. Detailed geomorphic, pedologic and ecologic analysis at a single dune level is thoroughly presented in Breckle et al. (2008) for dunes at the Nizzana research station. According to limited datasets of different ages relating to the southern dunefield, dunes encroached into the NW Negev in the Late Pleistocene (Zilberman, 1991; Rendell et al., 1993; Harrison and Yair, 1998; Ben-David, 2003), mainly during the Younger Dryas (Enzel et al., 2010). The dunes are mainly associated with Epipalaeolithic sites (uncalibrated) radiocarbon dated to ~18-10 ka (Goring-Morris and Goldberg, 1990). Over the past few centuries, the Sinai-Negev dunes have experienced several cycles of vegetation covering and removal (Tsoar, 1995; Meir and Tsoar, 1996; Tsoar, 2008).

The Negev dunefield was mapped at a 1:125,000 and 1:250,000 scale and the dunes were analyzed for military purposes in the early 20th century by the British War Office (Newcombe maps) (Levin et al., 2009). The NW Negev dunes, initially scientifically described and measured by Rosnan (1953) and Striem (1954), consists of stable vegetated linear dunes (VLD), with vegetation cover (Tsoar and Moller, 1986; Tsoar et al., 2008) of 5-15% (Siegal, 2009), which adds minute organic material to the dune section (Blume et al., 1995). Similar to

the linear dunes of the Sinai, the dunes are elongated in a general west-east direction (270°) with southern dunes having azimuths of 259° to 249° (Striem, 1954). The dune flanks are currently stabilized mainly by biogenic crusts (Danin et al., 1989; Karnieli and Tsoar, 1995; Karnieli et al., 1996; Karnieli, 1997; Kidron et al., 2000)

The dunefield south of the Qeren contains blocked and diverted ephemeral streams (wadis) (Goring-Morris and Goldberg, 1990; Magaritz and Enzel, 1990; Ben-David, 2003). The main dune body is north of a diagonal bend to the west of Nahal Lavan that currently drains via Wadi Haradein into Wadi Al-Arish (Figs. 1.4 and 1.5). In the past, the wadi course drained northwest through the Ghora corridor (Dan, 1977; Ben-David, 2003), possibly even in a direct path towards the Mediterranean Sea. Feldspar grains found amongst the fluvial gravels in several drills in Nahal Lavan were IRSL-dated to ~55-137 ka by Ben-David (2003). Two well-developed (stage II-III) palaeosol development periods at 35-30 ka and 27-24 ka, and weakly developed (Stage I-II) palaeosols at 14-12 ka have been identified and dated by (uncalibrated) radiocarbon along the southern edge of the Negev dunefield (Zilberman, 1992). This suggests that the ancient palaeocourse already led in a westerly direction even then. Nevertheless, the upstream standing or/and flood water deposits show that the main dune body contributes to the diversion and blocking of the Nizzana water course (Harrison and Yair, 1998).

NW Negev sand grains were found to be moderate to well rounded with no trend along their W-E extension (Hunt, 1991). A bi/polymodal grain size distribution predominantly (125-250 μ /63-125 μ) and (2.75 and 3.75 ϕ) are suggested by Hunt (1991) to be a result of either varying wind intensities or provenance. In the eastern part of the dunefield, a dune profile has been found to be predominately sand with approx. 80% fine sand (250-63 μ) (Gev, 1997). Sand mineralogy is dominated by quartz (95%) with 1-5% calcite (predominately in form of terrestrial snail shell fragments) and less than 1% of (unweathered) K-feldspar and plagioclase (Hunt, 1991). Clay mineralogy in the sands includes smectite and mixed-layer illite/smectite, attributed to external input (Hunt, 1991) and not to in-situ weathering.

Previous spectroscopic and remote sensing studies of sand redness in the NW Negev dunefield yielded general spatial trends but did not include any dating (Hunt, 1991; Campbell, 1999; Wenkart, 2006; Tsoar et al., 2008). Using laboratory spectroscopy, Wenkart (2006) and Tsoar et al. (2008) divided the dunefield into three sand units based solely on contouring a grid of the spectroscopic redness index ($RI = R^2/(B \cdot G^3)$). Wenkart (2006) and Tsoar et al. (2008) suggested that the west-central section of the dunefield north of the Qeren Ridge represents the latest dune incursion due to its lower RI values, whereas the redder northern and eastern

dunefields contain mature sands. This analysis was based on the assumption that when sand source factors and climatic conditions are homogenous, diversity of hues and spectral properties of red sand point to 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; White et al., 2001).

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

It has been suggested that the Late Pleistocene climate along the Sinai-Negev Erg, while similar to the Eastern Mediterranean synoptic configuration, was also stormier, wetter, and windier than today, and it was responsible for the Sinai-Negev dune mobilizations and Negev loess deposition (Enzel et al., 2008). In addition, an archaeobotanical study of the Central Negev Highlands south of the Negev dunefield has suggested a wetter Late Pleistocene between 18-10 ka (Baruch and Goring-Morris, 1997). Vaks et al. (2006) also maintain that prior to 14-13 ka, during the Late Pleistocene, the northern Negev received 300-350 mm of rain.

Since there is currently no dune incursion, despite the various spatial and temporal anthropogenic changes, one can assume that dune incursions transported sand and dunes eastwards into the Negev under different climatic and environmental conditions. The northern Sinai region provides an unlimited source of sand and this was not a constraining variable for dune incursion into the Negev. This research proposes that linear dune extension developing into dunefield encroachment is mainly the result of effective wind regimes transporting sand from Sinai. From the geometric, morphologic, and spectral properties of the NW Negev dunes and sand supported by luminescence ages, and from the spatially varied standing-water deposits and prehistoric site ages, we can infer several periods of incursion and mobilization controls resulting in distinct sand and dune bodies.

1.3. Research hypotheses and goals

The research hypothesis states that several sand incursions into the NW Negev took place during the late Quaternary creating dunefields of different geomorphic units (Karnieli and Tsoar, 1995). In general, each dunefield accumulated during a relatively rapid pulse in response to environmental factors marked mainly by strong wind intensities. In some cases, several dunefield sectors may have accumulated simultaneously since their spatial distribution probably resulted either from different sedimental supply sources or different wind directions of similar wind intensities.

The main aims of the research were to 1) identify the periods of sand and dune incursion into the NW Negev Desert that formed the NW Negev dunefield; 2) demarcate the spatial extent of these incursions; and, 3) interpret and conceptually model the dune-driving palaeoclimates.

Several tasks were involved.

- Spatial and temporal identification of the sand incursions and stabilization periods that formed the current NW Negev dunefields.
- Examination of the link between dune and sand grain color and spectral characteristics, and the significance of the spatial distribution of sand color in regard to, dune stability, dunefield evolution, and palaeoclimate.
- Evaluation of the importance of the major palaeoclimatic factors of wind regimes and/or rainfall patterns in triggering dune mobilization.
- Discussion of the results and inferred palaeoclimate in the context of global dune mobilization and global Late Quaternary climate change. This will help to identify and conceptualize a model of the conditions for dune mobilization and stabilization.
- Discussion of the impact of future environmental and climate change, for example increased aridity, upon the NW Negev dunefield and its environs.

1.4. Thesis outline

This thesis is based upon three peer-reviewed articles. The present chapter described the relevance and importance of global research, and particularly the Sinai-Negev dune and dunefield research, for defining the research hypotheses and goals. Chapter 2 systematically reviews the research methodologies and methods, and expands and partially overlaps the methods sections presented in the peer-reviewed articles in Chapters 3, 4, and 5. The results and various discussions of the thesis are found in the following articles:

Chapter 3 — **Roskin, J., Porat, N., Tsoar, H., Blumberg, D.G., Zander, A.M., 2011a. Age, origin and climatic controls on vegetated linear dunes (VLDs) in the northwestern Negev desert (Israel). *Quaternary Science Reviews*, 30: 1649-1674.**

The article examines the research regarding the Sinai-Negev Erg, discusses the reliability and accuracy of the OSL age results, presents stratigraphic and sedimentological data of the Negev dunefield dunes, sands, and palaeosols, and discusses the contribution of OSL ages and stratigraphy of the Negev to temporally and spatially map the dune incursions. It also attempts to decipher the sand and dune transport path and chronology from the Nile Delta through northern Sinai into the NW Negev.

Chapter 4 — **Roskin, J., Blumberg, D.G., Porat, N., Tsoar, H., Rozenstein, O. Do dune sands redden with age? – The case of the northwestern Negev dunefield, Israel. *Aeolian Research* 5: 63-75.**

The article deals with the sedimentology of the Sinai-Negev Erg sands and their transport route. Based upon the relationship between spectroscopic redness intensity and OSL depositional ages of sand samples taken from exposed and fully-drilled VLDs and interdunes of the NW Negev dunefield it is suggested that sand-grain redness does not progress with depositional time. Following spectroscopic analysis of northern Sinai sand samples, remote sensing of the Sinai dunefield, and data from previous studies on the Nile Delta sands, it is suggested that sand grain redness attributes were inherited from upwind sources and are not related to the age of stabilization.

Chapter 5 — **Roskin, J., Tsoar, H., Porat, N., Blumberg, D.G., 2011b. Late Pleistocene regional and global palaeoclimate of dune mobilization and stabilization; evidence from the vegetated linear dunes of the northwestern Negev Desert, Israel. *Quaternary Science Reviews* 30: 3364-3380.**

This article builds on the dune elongation and accretion understandings presented in Chapter 3 (Roskin et al. 2011a) and refine the analysis of VLD stratigraphy, mobilization and stabilization dynamics and periods. Combined with regional sediment records, global proxies, and global dune luminescence ages, this enables interpretation of Late Pleistocene Negev and global dune-driving palaeoclimatic controls such as increased low-latitude cold-event and glacial windiness.

Chapter 6 summarizes the main findings and conclusions of this study, its scientific importance and significance and proposed future research. The GPR profiles and interpretations are presented in Appendix A. As they did not considerably contribute to the research goals, the profiles are not discussed in the main body of the thesis.

2. METHODOLOGY AND METHODS

2.1. Reconnaissance work

The large Sinai-Negev Erg, between the Nile Delta and the northern Negev, along with the expanse of the NW Negev dunefield study area where field work took place, required a sound knowledge of the landforms, sediments and sites, as well as the details and methods used by previous researchers, in order to choose optimal sampling sites to serve the study goals. First, the NW Negev dunefield was studied using aerial photographs and Landsat image classifications and with the aid of reconnaissance surveys along dirt-roads. This also assisted in locating exposed dune sections. Dune crests were mapped from orthophotos at a scale of 1:5,000 in a GIS and validated by field surveys and fused images (Fig. 2.1a). Dune morphometries and morphologies were mapped and classified using ArcMap 3D, and spatial analysis modules on digital elevation models obtained from online SRTM and airborne LiDAR (Appendix A.1; Fig. 2.1b).

Geomorphic units were classified qualitatively based on dune crest orientation and spatial density, dune and cross-section morphology, and single and spatial dune slope distribution (Figs. 2.2 and 2.3a+b). This approach followed the hypothesis that stable, mature dunes could have degraded (O'Connor and Thomas, 1999; Lancaster, 2007), and different dune morphologies and related geomorphic units may represent different histories of dune mobilization, buildup, stabilization, and degradation. Ultimately, the geomorphic units merged into three main west-east trending dune bodies, delimiting discrete incursion or encroachment corridors as partially consistent with Tsoar et al. (2008) (Fig. 2.3c; Table 3.2 in Chapter 3).

2.2. Sampling site strategy, selection and stratigraphy

The sampling strategy for stratigraphy, sedimentological analyses and OSL dating was designed to identify the earliest dune incursions and to analyze dune elongation/advancement rates. The dunefield was sampled along 5 lines: western and eastern north-south transects and a west-east transect along each incursion corridor (Fig. 2.3c). The NNW-SSE sampling line ("western transect") at the western end of the study area along the Israel-Egypt border, transected and is almost perpendicular to the VLD orientation. Sampling was performed along this line in nearly every dunal geomorphic unit. In the eastern transect, the easternmost extent of each incursion corridor was sampled in order to date dune advance and cessation.

Sampling was conducted at defined sites. Sites often included several sampled exposed and/or drilled sections of the dune, interdune and upper dune substrates (Appendix A.2). All of

the exposed sections identified in the dunefield were described. Additional sites were sampled by drilling in keeping with the sampling strategy. Dune stratigraphy in the exposed sections was described using standard sedimentological and pedological methods (Dan et al., 1964; Birkeland, 1999). Ground-penetrating radar (GPR) surveys were conducted primarily to identify sedimentary units and determine their depths prior to sampling by drill.

Sampling was aimed at retrieving the dune-base sand (the lowest ~1 m) and its underlying substrate in order to detect the earliest sand activity and create a stratigraphically uniform age database which has been lacking from previous studies (Telfer and Thomas, 2007). Dune axes were generally targeted for sampling in order to obtain the dune core sediment, presumably least affected by slight possible lateral dune sand movement (Bristow et al., 2007).

The BM VLD flanks were sampled in order to provide a picture of VLD elongation, buildup and lateral migration dynamics across a full dune cross-section and to provide correlation with GPR profiles. The BM VLD was also sampled at two sections along its axis in order to investigate the longitudinal morphological changes, elongation, and narrowing of the linear dunes over time.

Splits of sixty-two northern Sinai sediment samples (for location of analyzed samples, see Fig. 4.5 in Chapter 4), mainly sands, along with general sampling site description data, were provided to Dr. Dan Muhs who split the sand with me. The sand was courtesy of Dr. Amihai Sneh (GSI emeritus) who sampled it for the Geological Survey of Israel during the late 1970s.

2.3. Sampling

Drilling techniques for sampling sand from drills were developed on the job as dune-drilling techniques for OSL samples has only recently been published (Munyikwa et al., 2011). Drilling was performed with Dormer Engineering hand augers (Figs. 2.4a-f). of diameters of 85, 95 and 105 mm (Figs. 2.4a, e & f). A Drillmite 6Hp hydraulic engine was found to be ineffective as it does not facilitate sample extraction (Fig. 2.4b). Drill auger retrieval was usually manual and as it becomes laborious beyond ~7 m, a maximum drill depth of 11.4 meters was achieved (Fig. 2.4c). Several horizontal auger probes of steep dune flanks were also carried out though these were not found to be advantageous over vertical drilling (Fig 2.4d). Beneath the upper 2-3 meters of the dune, the unconsolidated sand was usually slightly moist (~1%), providing easy penetration and drill hole stability. Drill hole surfaces were cased with 0.5 m-long PVC pipes and drill holes were cased with 3" and 4" PVC or Dormer 3" 1.5 m

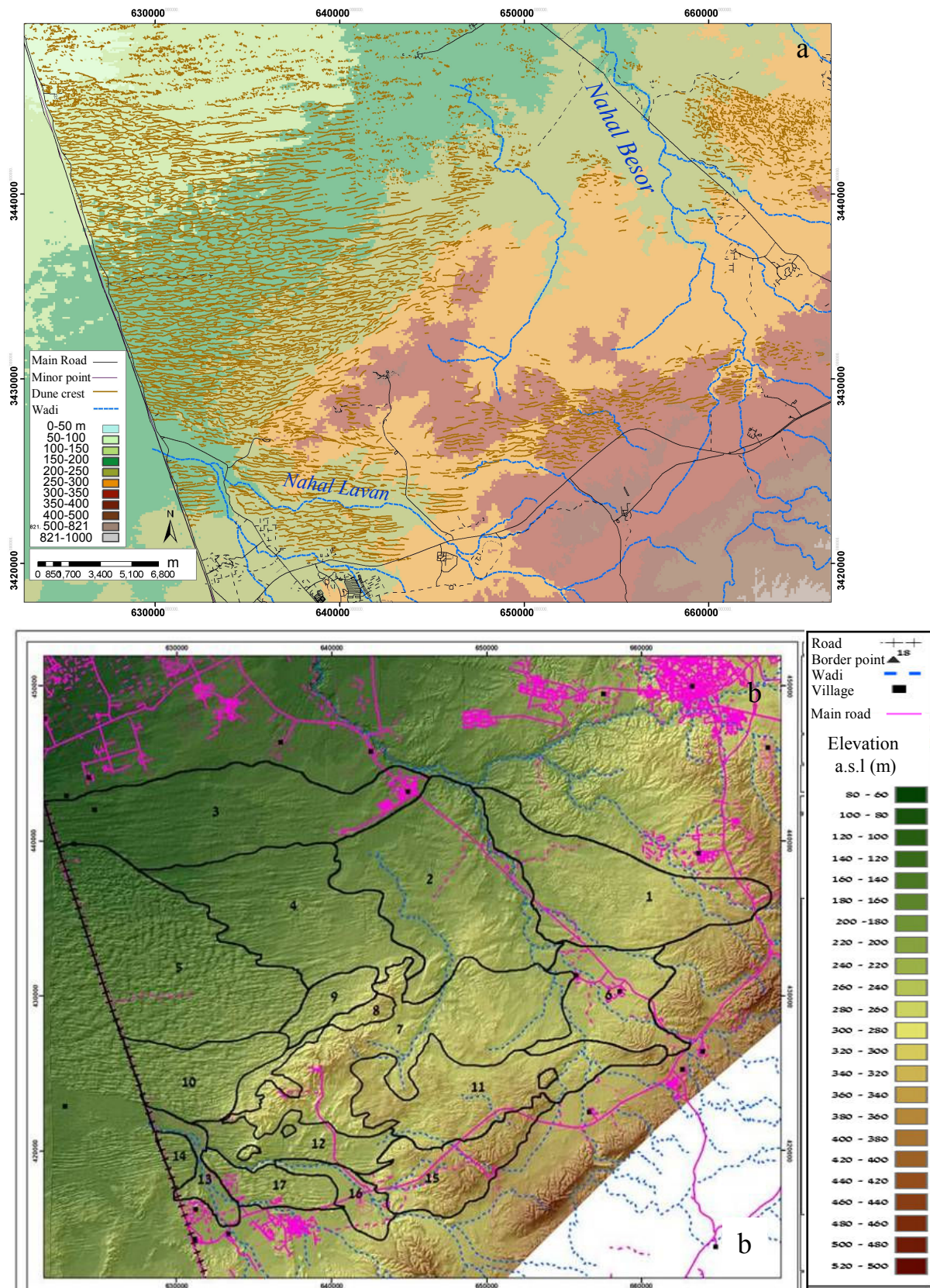


Figure 2.1 Topographic-geomorphic analysis of the NW Negev dunefield landforms

a. Dune crest line mapped upon shaded elevation map.

b. Geomorphic unit (g.u.) classification (numbers) upon shaded relief map. The geomorphic unit names signify their main features as follows: 1. Sekher dunes and sands. 2. Ze'elim basin. 3. Northern corridor (Haluzit – Ze'elim) dunes. 4. Haluzza east dunes. 5. Haluzza (west) dunes. 6. Revivim-Besor wadi confluence. 7. Qeren Ridge. 8. Dune-blocked wadis. 9. Crescentic dunes. 10. Agur dunes. 11. Shunera dunes. 12. North Nahal Lavan dunes. 13. Nahal Nizzana. 14. Halamish dunes. 15. Shivta Chalk Hills. 16. Nahal Lavan floodplain. 17. Lavan dunes.

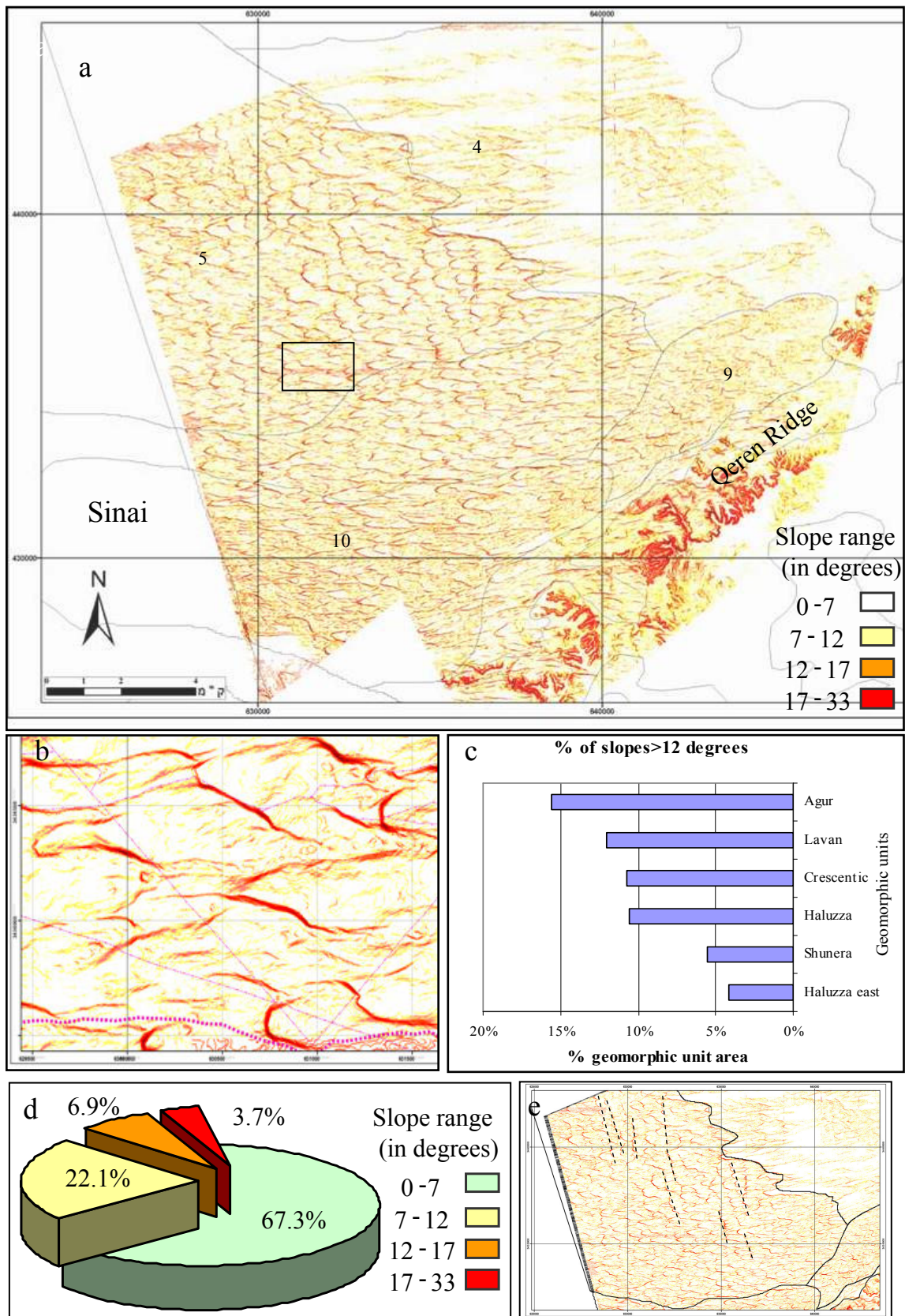


Figure 2.2 Slope analysis of the NW Negev dunes.

a. Slope distribution map and geomorphic classification (numbers of geomorphic units as in figure 2.1b) of the central NW Negev dunefield. For map location, see figure 2.3a.

b. Close-up of dune slopes (black triangle in 2.2a). Note the crescentic shaped transverse-like eastern-facing lee slip faces. c. Relatively steep slope frequency per dunal geomorphic unit. d. Slope distribution of the Haluzza geomorphic unit. e. North – south dashed lines of parallel eastern-facing crescentic slip faces in the Haluzza geomorphic unit.

sand-sampling metal pipes in order to stabilize the borehole and prevent sediment fall and contamination (Fig. 2.4a). Depth was measured by 10 cm-interval marks on the extension rods.

Sampling for OSL dating of drilled sediment was performed at 1.5 m intervals unless field examination of the samples indicated changes in sediment properties, in which case sample frequency was increased. An opaque cloth draped upon the extension rods and sampler protected the extracted OSL sample from sunlight (Figs. 2.4g+h). The lower section of the auger sediment was discarded due to suspected drill hole contamination (Figs. 2.4e & f). Sediment from the central section of the auger was immediately placed in opaque black bags. An additional sediment sample was collected from the same probe or the next probe for dose rate determination. OSL sampling from both exposed sections and drills was often executed at night in order to easily avoid exposure to light (Figs 2.5a+b).

Sampling for OSL dating of exposed sections usually began 1-2 m below the surface to avoid the bioturbated and active dune crests. Sampling points were chosen for each unit at least 10 cm from sedimentary contacts and usually in mid-unit. Sampling involved driving hard, opaque, 20 cm long by 3 cm diameter plastic pipes into the exposure by hand or hammer (Fig. 2.5c). Here, where possible, cosmic and γ dose rates were derived from *in situ* measurement using a calibrated portable gamma scintillator (Fig. 2.5d)

Altogether more than 300 samples for OSL dating and sedimentology were collected from 40 exposed sections and drills from 20 sites. A sampling density of one site per approximately 65 km² provides a relatively reliable numerical age dataset for environmental and palaeoclimatic interpretation (after Telfer and Thomas, 2007).

2.4. Ground penetrating radar (GPR) profiles

2.4.1. Background

Ground-penetrating radar (GPR) is an electromagnetic geophysical exploration technique for mapping shallow subsurface structures and locating underground objects. GPR uses short pulses of electromagnetic energy (radio waves) that are propagated into the ground by a transmitting antenna that is placed upon the ground surface. At the same time a receiving antenna detects the waves that are reflected up to the ground surface when the transmitted pulse encounters a subsurface interface across which exists an electromagnetic impedance contrast. The delay between the transmitted pulse and the arrival of a reflection is proportional to the depth of the subsurface feature (reflection surface) that generated the reflection.

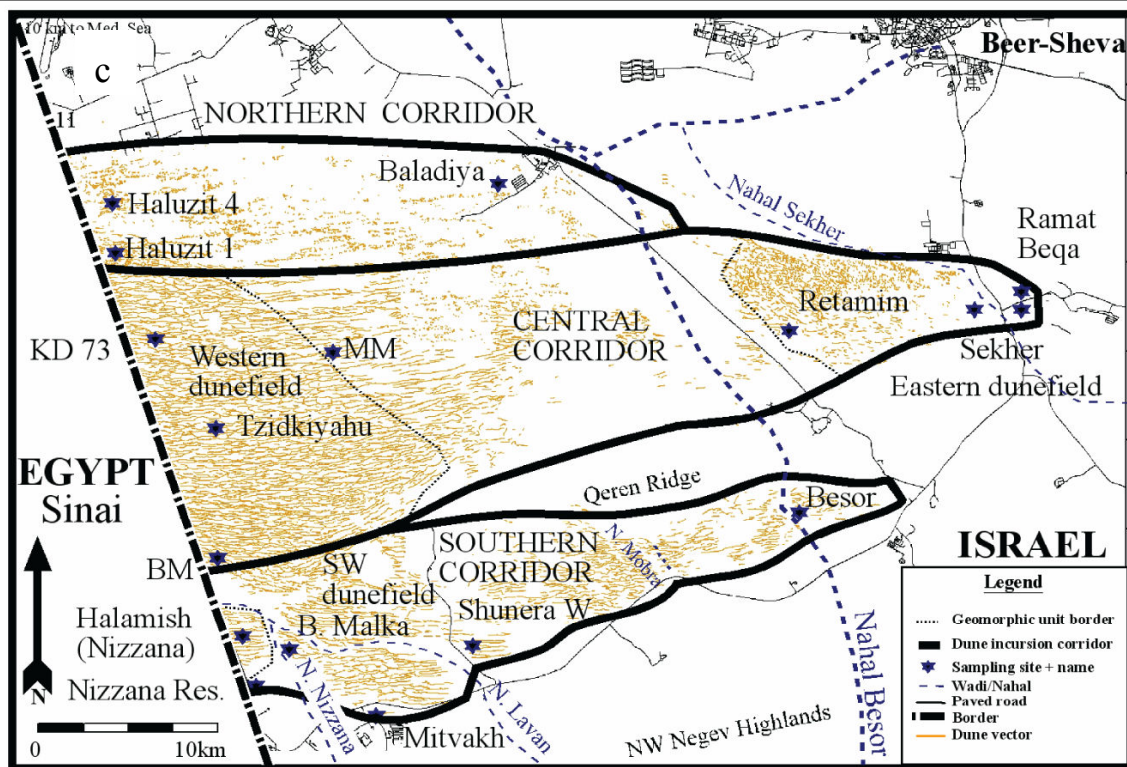
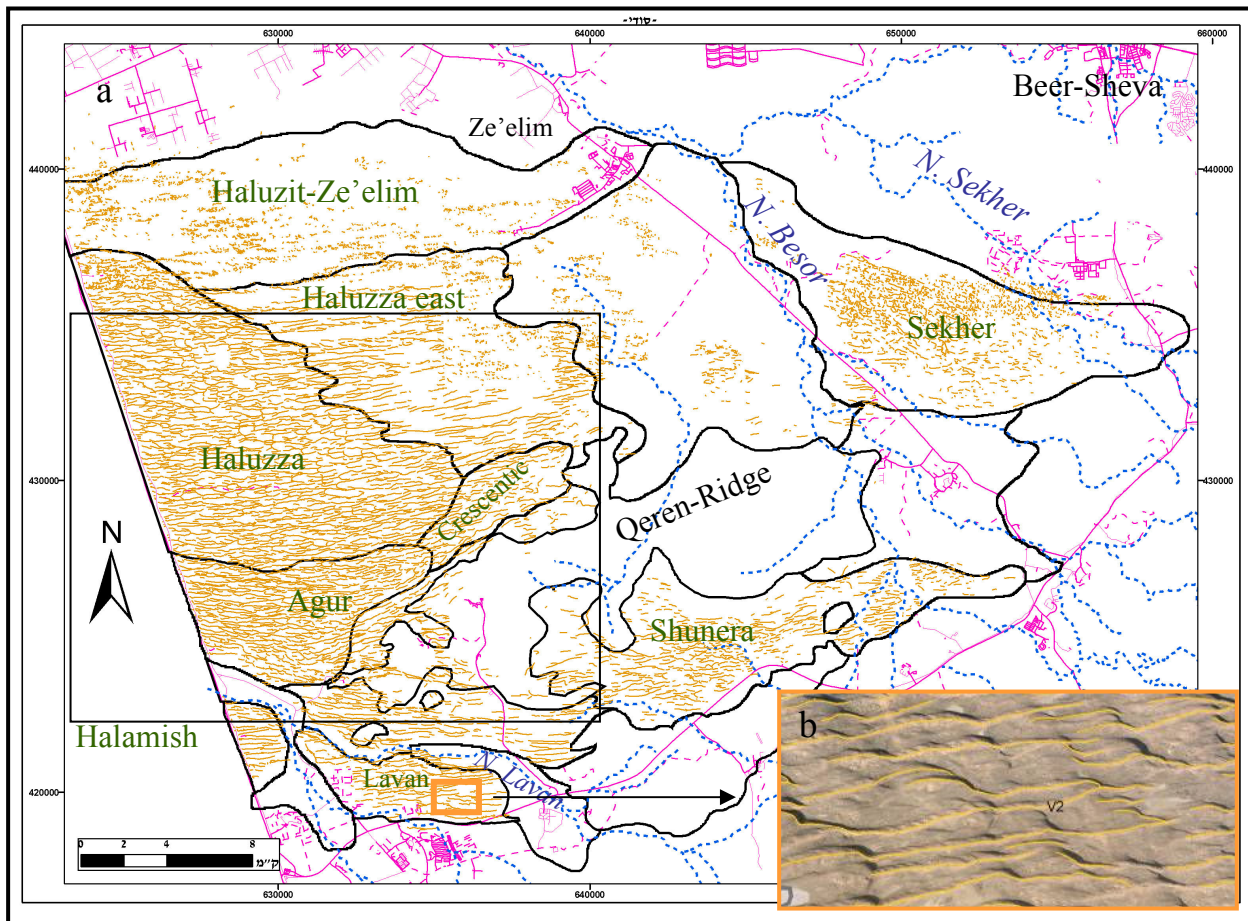


Figure 2.3 Geomorphic unit (green names) and dune orientation analysis of the NW Negev dunefield. a. Geomorphic units and dune crest line planar pattern and density.

The black box marks figure 2.2a.

b. Inset of orthophoto-mapped Lavan geomorphic unit VLD crest lines.

c. Dune encroachment (incursion) corridors according to VLD patterns and geomorphic units. Named sampling sites, usually including several sections or drills, are located near the dune encroachment corridor edges.

The past 25 years have seen a substantial growth in GPR applications for earth sciences, including aeolian deposits (Neal, 2004). Good results have been obtained for coastal dunes (Pedersen and Clemmensen, 2005; Barboza et al., 2009), though limited work has been published for inland linear dunes and specifically VLDs. Bristow et al. (2005, 2007) calculated historical unvegetated linear dune advance and lateral migration in Namibia using GPR facies analysis coupled with optical ages. They showed that in some cases GPR can further our understanding of dynamic sand movement in and along a single dune based on analysis of bounding surfaces and the field relations between the dune structure and its substrate. Clemmensen et al. (2007) used GPR to map the internal structures of a relatively low recent linear dune on an island off of the Denmark coast, where recent/historical ages revealed age inversion.

Until now, GPR surveys of the Negev dunes have been limited to mapping the moisture content of linear dunes (Basson, 1992; Gev, 1997), the relationship of the Nizzana VLDs to their substrate (Ben-David, 2003) and the comparison of GPR reflections to Single-Aperture Radar (SAR) reflections of sandy deposits in buried channels (Blumberg et al., 2004). Ben-David (2003) identified interfingering of dune "roots" with later, younger, silty, playa-like accumulations, but did not describe internal dune structures. The low electric conductivity (3-5mSec) of the Nizzana dune section (Blume et al., 1995, 2008) which is similar to the Negev dunes suggests low salinity, which should not inhibit GPR profiling. Bristow et al. (2007) reported limited penetration and poor resolution due to signal attenuation which prevented imaging of primary vegetated linear dune sedimentary structures in central Australia. This was attributed to a combination of physical factors: clay illuviation, salt, biological disturbance of the structure and sediment homogenization, rubification, and the age of the dunes. The Negev dunes, however, lack clay horizons, are younger, less red, have low salt contents, and were predicted to be more responsive to GPR imaging.

The limited accessibility of dune cores, mainly due to the unstable stature of exposed excavated dune deposits, restricts achieving detailed mapping and analysis of the internal structure of linear dunes (Bristow et al., 2007). GPR profiling was chosen to map the internal dune structure since unsaturated clean sand is considered a good material for GPR transmission due to its low relative dielectric permittivity (2.55-7.55) and conductivity values (0.01 mSm^{-1}), which limit attenuation (0.01-0.14) (after Neal, 1994) and enable good penetration. For an antenna central frequency of 100 MHz GPR resolution is calculated to be 0.25 m for damp sand and 0.375 for dry sand (Bristow, 2009). Interchanging cross-strata laminae, due to grain-size differences, have corresponding moisture differences that may cause changes in velocity,

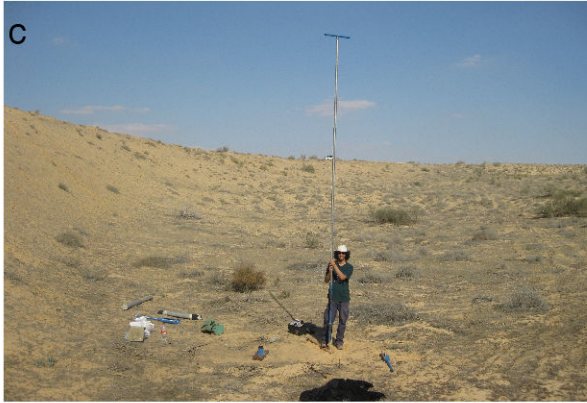
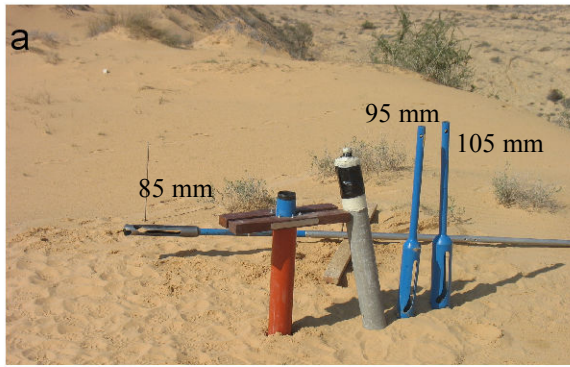


Figure 2.4 Field drilling methods. a. Sand drilling equipment and auger diameters. Note Dormer pipes inserted in orange PVC pipe. They grey PVC pipe sealed with a black bag eased extraction of the sediment from the auger. b. Dormer Drillmite engine. c. Auger extension poles following retrieval from Tzidkiyahu interdune section. d. Horizontal drilling at the southern base of the BM east VLD. e. Auger tip. f. The maximum sand fill in an auger. g. Drape cloth over drill rig and (h) during sampling, enabling extraction for OSL-dating without exposure to sunlight.



Figure 2.5 OSL and GPR field work methods and spectroscopic measurement setup.
a. OSL sample extraction by dune-drilling and from an exposed section (b) at night.
c. OSL sampling at the Sekher VI section with plastic tubes. d. Cosmic and gamma dose rate measurement with a calibrated portable Rotem P-11 gamma scintillator.
e. Topographic surveying with a Total Station, here positioned upon the BM transverse dune.
f. GPR profiling at the BM site.
g. Spectroscopic measurement set up including a (1) black box for the sample and (2) probe muzzle. h. Different redness intensities of sand samples being room-dried in plastic plates.

inducing reflections (after Bristow, 2009). Internal cross-strata bedding patterns and bounding surfaces were hypothesized to be possible reflection surfaces.

2.4.2. GPR survey goals and site locations

It was hypothesized, first, that the internal dune structure obtainable from GPR would contribute valuable data for analyzing VLD elongation, accretion, lateral migration and crest and surficial activity. Second, it would help to differentiate the more prominent upper young and active dune crest even deep in the dune, from the dune and dune substrate. The GPR survey goals included:

- Determining dune depth and contact with interdune sands, palaeosols, and playa loams, in other words, the palaeosol reflectors found by Hollands et al. (2006) at depths of 4 m in Australian linear dunes.
- Determining the VLD's internal structure, both along the dune axis and axis cross-section, in order to advance our knowledge of dune elongation and accretion dynamics in the NW Negev.
- Identifying the main dune units for OSL sampling.
- Identifying the internal structure of the inferred transverse dunes that infill the VLD interdune in order to validate their internal structure and advancement dynamics and identify units for OSL sampling.

2.4.3. GPR profiling and processing

- Sites were chosen based upon geomorphic unit mapping and identification of various dune sub-types (Appendix A.3). This was based on the reconnaissance work noted above.
- Data acquisition in the field included marking the GPR transects perpendicular to the VLDs axis using a measuring tape. Along this line, topographic cross sections were measured using a Total Station (Fig. 2.5e) and the GPR survey was conducted (Fig. 2.5f). Transects were marked upon high resolution orthophotos.
- GPR measuring used hand-held transmitter-receiver antennae with 1 m spacing at step-size increments of 10 cm.

- The GPR unit included a RAMAC system which transmitted mainly at 100 MHz for reaching dune depths and at 200 MHz to achieve higher resolution albeit with shallow differentiation.
- Figures of GPR settings, profiles and interpretations are presented in Appendix A.4-A.11.
- Data processing included several gain amplification levels (to increase depth of data), migration, bandpass filtering and decon. Data was processed without and with topographic corrections based on the cross-section profiles.
- The GPR display used both variable-intensity (color) and wiggle (b&w). Subjective image interpretation was mainly based on the migration and wiggle processed sections. Illustrations of the interpretations often only show the main reflection surfaces (see Appendix A.7).
- The GPR measurement and processing was led by Dr. Uri Basson of Geo-Sense Ltd.
- GPR terminology and interpretation techniques are according to Neal (2004).

2.5. Particle size distribution and moisture content analysis

Variations in particle size distribution (PSD) of sands exist at three scales; (a) on individual dunes; (b) between dunes types and dunes and interdunes; and (c) regionally across the dunefield (Pye and Tsoar, 2009). Beyond the fact that these scales are analyzed in this research, PSD analysis was undertaken to identify changes along the VLD axis and to identify post-depositional changes such as pedogenesis, i.e. addition of fines.

PSD analysis of 118 samples that included sand and loams was carried out by laser diffraction (using a Malvern Mastersizer MS-2000) at the Sedimentology Laboratory of the GSI (Appendix B.1). Compared to pipette results, laser diffraction measurements of PSD of sand fractions using different machines has been found to produce similar PSDs (Beuselinck et al., 2008; Cheetham et al., 2008) making the results comparable to other studies.

Samples were split into 5-g portions, sieved to < 2 mm, and stirred for dispersion in sodium hexametaphosphate solution for 10 min followed by ultrasonification for 30 s. Three replicate aliquots, later modified to two aliquots for sand samples 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 with optical parameters of RI=1.52 and A=0.1.

Sand and loam samples from representative sites and sections collected in sealed jars were measured for moisture content by oven-drying (Appendix B.2).

2.6. Relative mineral abundances

X-ray diffraction (XRD) was used to determine relative mineral abundances for Negev and Sinai sand. Relative abundances were established by measuring XRD peak heights using Materials Data Inc. JADE 5 software at the USGS in Denver.

The following XRD procedure was used:

- Sample splits of several grams were pulverized in a shatterbox for a standard time of 5 min.
- Slides for XRD were prepared as random mounts.
- Slides were X-rayed from $8^{\circ} 2\theta$ (most samples) or $20^{\circ} 2\theta$ (several samples) to $55^{\circ} 2\theta$ on a Phillips 3100 X-ray diffractometer, using copper radiation.
- The resulting diffractogram showed the relative intensity (count/s) versus the angle of radiation for calcite, quartz, and plagioclase. This was then interpreted to establish the relative abundance of the minerals.

The above method offers a low level of accuracy, sometimes with errors reaching tens of percent. The results are therefore taken to indicate "relative abundance" (Appendix B.3).

2.7. Optical stimulated luminescence (OSL) dating

2.7.1. Introduction

2.7.1.1. Background

Optically stimulated luminescence (OSL) dating is a method used extensively in the earth sciences and archaeology. It dates the burial age of sediment when it loses contact with sunlight. It is based on the accumulation of dose and emission of light, luminescence, mainly from sand-sized quartz grains. The method can be applied to a wide range of sediments and materials containing even small quantities of sand-sized quartz grains. The age range over which the method can be applied is from several hundred years to several hundred thousand years (Wintle, 2008). The fundamentals and status of luminescence dating are described in detail in major reviews (Aitken, 1998; Duller 2004, 2008), and specifically regarding drylands and aeolian sediments (Singhvi and Porat, 2008).

Luminescence age is determined by the ratio between the measure of the natural radiation dose (D_e) absorbed by the quartz grains and the rate at which the energy was delivered (dose rate). This is according to the simple formula:

$$\text{Luminescence age} = D_e / \text{dose rate} \quad \text{Equation 2.1}$$

Age = Time since the sediment was last exposed to sunlight

D_e = Equivalent dose measured in Gy ($\text{Gy} = 1 \text{ J kg}^{-1}$)

Dose rate = Radiation flux within the sediment (Gy a^{-1})

Natural (α , β , γ) radiation is largely emitted by radioactive potassium (K) isotopes and isotopes in the decay chains of uranium (U) and thorium (Th) present in the sediment. At a sub-atomic level, energy is stored by excited electrons trapped at sites within the crystal structure, where they cannot normally reside, but can be stored as a result of defects in the crystal structure. In general, the deeper the defect's location beneath the conduction band, the longer the electrons can be trapped (Fig. 2.6). When an electron is released, it loses the energy gained during burial and may emit part of that energy in the form of photons of light, termed luminescence. Measurement of the luminescence signal intensity can be used to calculate the amount of radiation to which the sample was exposed during burial. Measurement involves (optical) stimulation of the sand grain using light of blue or green wavelengths. The total absorbed dose of radiation is known as the equivalent dose (D_e).

In the natural environment, quartz OSL traps are emptied by exposure to wavelengths from ultraviolet to green of the electromagnetic spectrum present in sunlight (Aitken, 1998). This luminescence signal bleaching, also known as zeroing or resetting, is not always complete under all depositional circumstances. For example, sediments in fluvial environments (Olley et al., 1998, 2004; Rittenour, 2008), do not necessarily receive sufficient sunlight for complete bleaching. Aeolian sediments are often assumed to have been well bleached during transport. However, it cannot always be assumed that the signal was completely reset upon deposition (Aitken, 1998; Singhvi and Porat, 2008). Sand-grain mixing can introduce grains of different ages into the sections (Bateman et al., 2003, 2007). In this study it was assumed that the luminescence signals of saltating sand grains are fully reset.

2.7.1.2. Equivalent dose (D_e) measurement

The radiation doses given to sediments in the laboratory consist of β radiation only. The radiation is applied on small aluminum discs (aliquots) usually ranging between 2 to 10 mm that contain several hundred quartz grains.

The luminescence sensitivity of the aliquot – the amount of light it emits for each unit of radiation, i.e. light stimulation, heating, and radiation, to which it is exposed, changes depending on the laboratory procedures undertaken. This can create an inconsistent response and poses a problem for the accurate measurement of D_e . To overcome this problem, the Single Aliquot Regenerative dose (SAR) protocol was developed (Murray and Wintle, 2000, 2003). The SAR protocol is a regenerative-dose protocol, which measures both the natural and regenerated OSL signals (Fig 2.7).

The SAR protocol comprises a series of cycles. In the first cycle the OSL signal (denoted L) from the aliquot arises from the radiation dose to which the sample was exposed in nature, and hence is given the term L_N . In the second cycle the aliquot is exposed to an artificial (regenerated) source of radioactivity in the laboratory. The OSL signal is then measured. Subsequent cycles measure (L_x) as different regeneration doses are given to the aliquot. All of these measurements of luminescence are preceded by a preheat – heating the sample to a fixed temperature (usually between 160°C and 300°C) and holding it there for a short period of time (10s). This procedure removes unstable electrons from shallow traps so that the OSL signal comes only from electrons that would have been stored safely through the burial period. The brightness of these luminescence signals (L_x) are used to construct a dose response curve (Fig. 3.4b in chapter 3).

The luminescence sensitivity is measured by giving a small fixed radiation dose (test dose) in the second half of each SAR protocol and then by measuring the resulting OSL signal (T_x) (Fig. 2.6). The effect of any change in sensitivity is corrected by plotting a graph of the sensitivity corrected luminescence signal (L_x/T_x). D_e can then be calculated, based upon this luminescence signal, corrected for any changes in sensitivity that may have occurred (after Murray and Wintle, 2000, 2003).

OSL-SAR dating is the currently most established method for dating sands and dunes, especially where organic material for ^{14}C is lacking. Errors are usually in the range of 5-15%.

2.7.1.3. Dose rate measurements

Dose rate measurement tries to identify the amount of natural radiation the sediment underwent since deposition. As this cannot be accurately measured or assessed, dose rate

calculation includes uncertainties that affect the calculated age error. Dose rate is determined by concentrations or activities of K, Th, and U radioelements and cosmic energy. Radioelement concentrations can be measured using different methods; neutron activation analysis, different mass-spectrometry procedures like inductively-coupled plasma mass-spectrometry (ICPMS), atomic absorption or X-Ray fluorescence spectroscopy (XRF). The concentrations are converted into dose rates using conversion factors for radioactivity (Aitken, 1998).

Alpha (α) radiation dose rate can penetrate a few microns into sand grains. HF etching of the OSL samples makes the α dose rate negligible. Beta (β) derived dose rates have substantial influence on sand samples. Beta particles (electrons) penetrate several mm into sand grains. Gamma (γ) rays are high energy photons that penetrate approximately 0.3 m into sediment (Aitken, 1998). Both β and γ radiation experience scattering throughout sediment (after Aitken, 1998). The proximity of β -emitting grains to quartz grains can cause substantial dosage variation between grains. Beta particle attenuation by sediment-pore moisture content reduces the dose rate. Measured moisture content of selected samples along with taking a significant uncertainty into consideration are implemented in dose rate calculation. This uncertainty is crucial as the moisture content value incorporated represents constant moisture content in the sediment since burial. Calculation of dune ages with moisture content estimations from 2% to 12% was found to change age errors by 4% (Bubbenzer and Hilgers, 2003).

A proportion of the cosmic radiation energy does reach and penetrate the planet surface. The radiation dose reaching Earth decreases with depth, a particularly rapid process in the uppermost meter of the Earth (Duller, 2008) (a depth that was not usually sampled in the present study). The contribution of cosmic rays to sediment dose rate is assumed to be constant over time, and is a function of latitude, altitude and burial depth (Prescott and Hutton, 1994).

2.7.2. OSL measurement and age determination

2.7.2.1. Sample preparation

Thirteen preliminary samples from shallow depths (labeled ISR in Table 3.3; Chapter 3) were dated using OSL at the Marburg Luminescence Laboratory, Germany. An additional 84 samples were prepared and measured at the Luminescence Laboratory of the Geological Survey of Israel (GSI), Jerusalem. Sample preparation follows Porat (2007). Briefly, this involved dry-sieving to isolated grain-size fractions of 125-150 μm or 150-177 μm , followed by immersion in 8% HCl to remove carbonates. After washing and drying, heavy minerals and

most feldspars were separated from the quartz using a Frantz magnetic separator on high (1.5A) current on the magnet (Porat, 2006). Grains were then submersed in concentrated (40%) HF for 40 minutes to etch grain rinds affected by α particles and dissolve any remaining feldspars. This was followed by 16% HCl treatment to remove any precipitated fluorides. The quartz grains were mounted on 10 mm aluminum discs with 5 mm masks using silicon (oil) spray as an adhesive.

2.7.2.2. Equivalent dose (D_e) determinations

Equivalent dose (D_e) determinations used a modified single aliquot regenerative-dose (SAR) protocol (Murray and Wintle, 2000) that included a cleaning step of heating to 280°C for 100 s at the end of each measurement cycle. The protocol started with measuring the natural signal, followed by a zero dose point to test for thermal transfer; three beta dose points; a second zero; a repeated dose (recycling ratio), and a second repeated dose after infrared (IR) bleaching (IR depletion ratio) (Fig. 2.7). Measurements were made either on a DA-12 or a DA-20 TL/OSL Risø reader equipped with blue LED's. Irradiation was from a calibrated ^{90}Sr β source and the luminescence signal was detected through 7.5 mm U-340 filters.

Dose recovery tests included bleaching discs in the sun for 1 hour and then giving a dose of 15 Gy. After a pause of 3 hours, a SAR protocol was used to measure the given dose over a 280-280° C range of preheat temperatures. The ratio between measured and given doses is 0.9-0.95 at temperatures above 220° C (Fig. 2.8). These ratios, similar to values found by Murray and Wintle (2003) and Pruesser et al. (2007) for glacial deposits seem to represent the inherent nature of the Negev quartz grains and contribute to the age errors that often exceeded 10%. Thus, preheats of 220⁰-260⁰ C for 10 seconds were used before OSL measurement at 125⁰ C. Test dose cutheat was applied for 5 s at 20 °C lower than the preheat. The D_e of each aliquot was determined by fitting a linear + exponential curve to the data points.

2.7.2.3. Dose rate determination

Where possible, cosmic and γ dose rates were derived from *in situ* measurement using a calibrated portable Rotem P-11 gamma scintillator with a 2" sodium iodine crystal (Porat and Halicz, 1996). For drillholes, cosmic dose rates were estimated from burial depths, though changing burial depth over time was not considered. For most samples age calculation relied mainly on the cosmic dose rates based on the sample depth. Chemical analyses of radioactive elements (K, Th, and U) of the sediments were performed using inductively coupled plasma

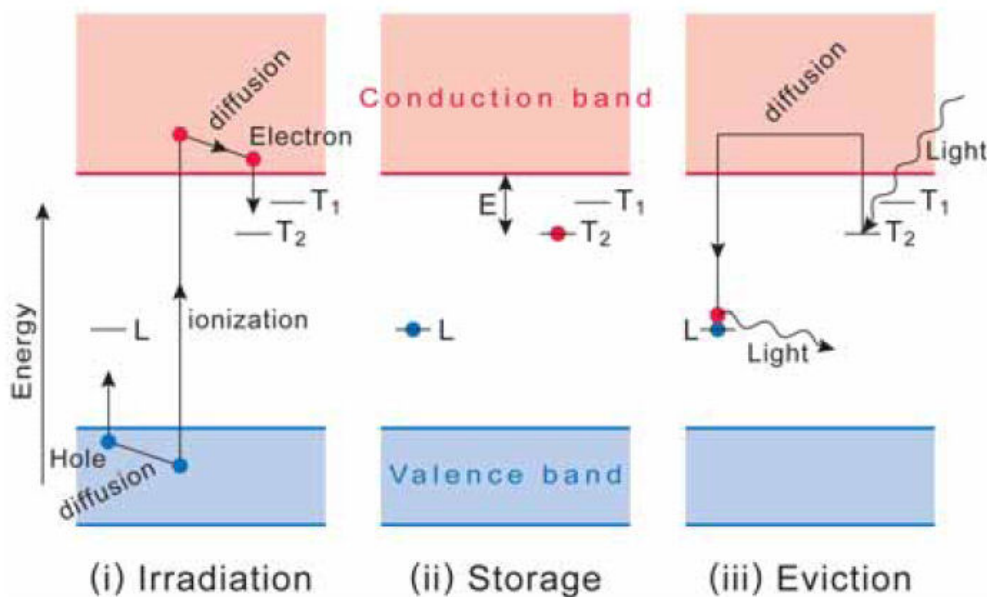


Figure 2.6 Energy level diagram illustrating the luminescence process (after Duller, 2008):
 (i) radiation interacts with the crystal (ionization), pushing electrons into the conduction band and leaving ‘holes’ in the valence band.
 (ii) electrons become trapped at defect sites (T1, T2, etc’) and remain for a period of time the trap below the conduction band (E) (eg T2) the more stable the electron and the longer it stays trapped.
 (iii) crystal is stimulated by heat or exposure to light, releasing electrons which recombine with holes at luminescence centers (L) and emit light photons = the luminescence signal.

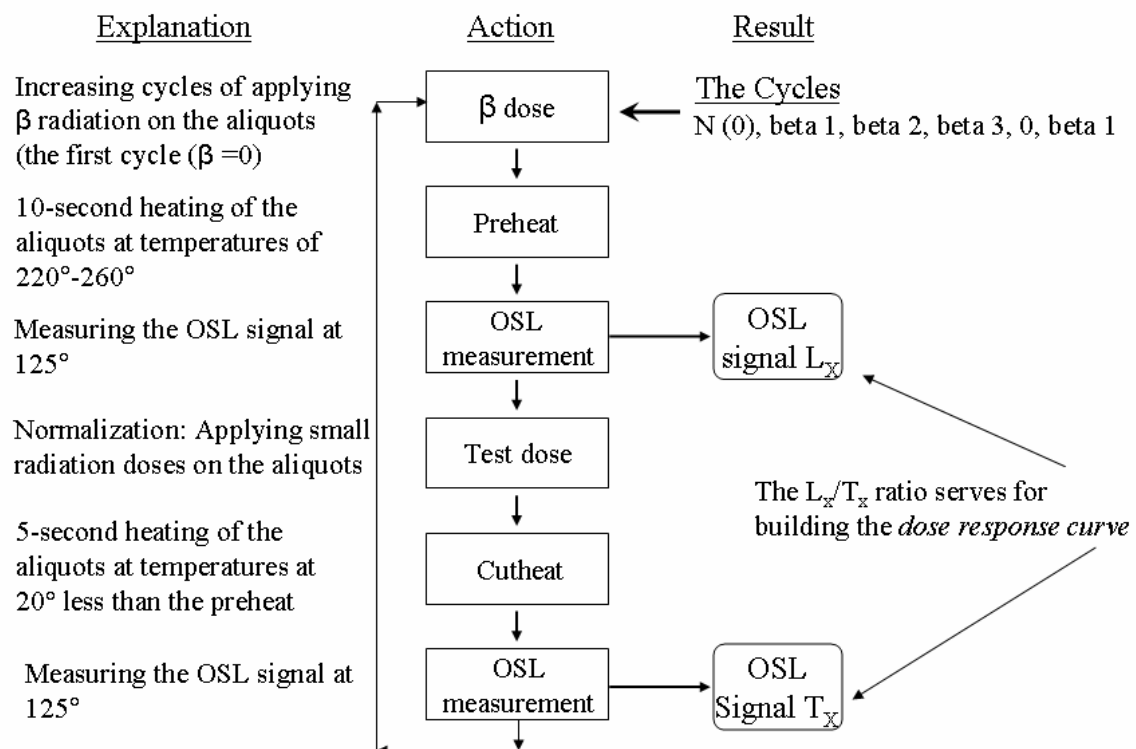


Figure 2.7 The Single Aliquot Regenerative dose (SAR) protocol used for this study.

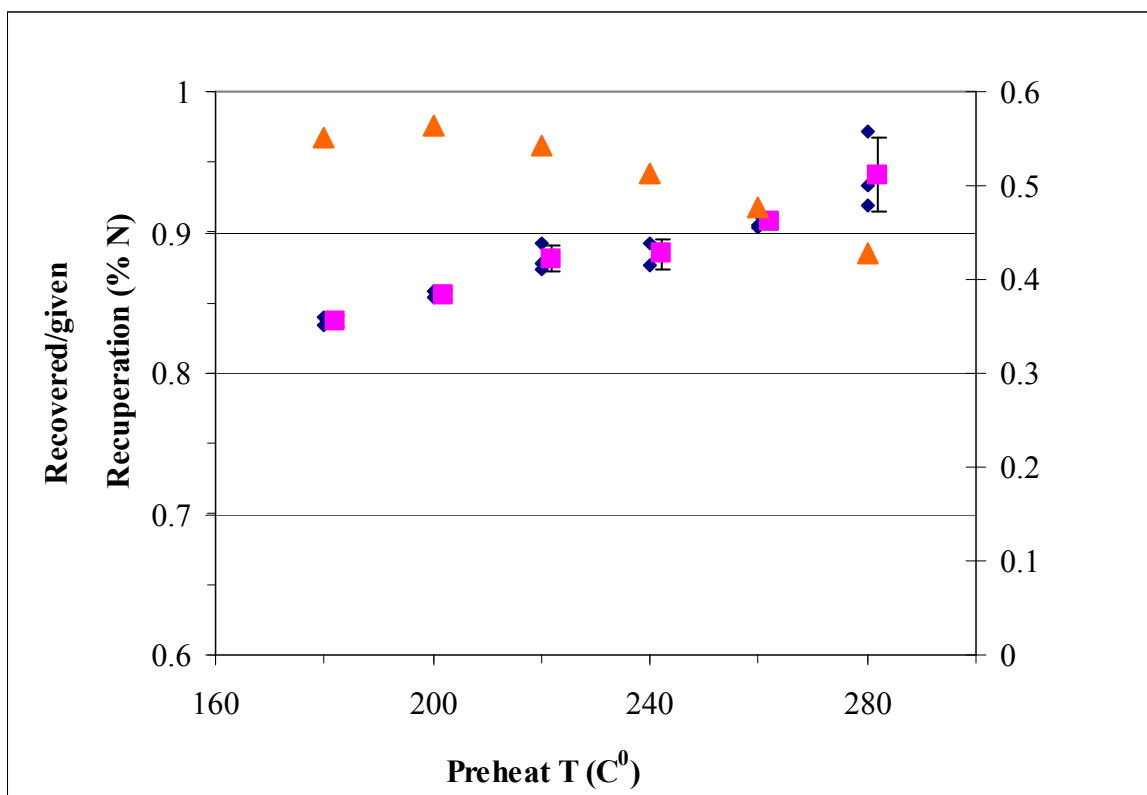


Figure 2.8 Dose recovery over a range of preheat temperatures for sample DF-625. Diamonds are individual measured aliquots and rectangles are the average for each temperature. Triangles show the average recuperation (right-hand y-axis) over the range of preheat temperatures.

mass-spectrometry or atomic emission spectrometry (ICP-MS/AES), and their concentrations converted into α , β , and γ dose rates using the factors given by Nambi and Aitken (1986). A moisture content of 2 ± 1 %, based on moisture measured for oven-dried samples, was used for age calculations of the sand samples. For samples with $>25\%$ fines, a moisture content of 6 ± 2 % was used (Appendix B2).

2.7.2.4. Age calculations

For most dune and interdune samples, age calculations relied on a D_e averaged from a minimum of 13 measurements (aliquots) per sample, with approximately half the samples relying on more than 17 measurements. However, very old (~ 100 ka) and some of the very young (0-150 years) samples were measured from only 8-13 aliquots as these ages were not the focus of this study. Several samples had a high scatter and further twenty-four to forty-eight 2 mm (~ 150 grains) aliquots were measured to assess the source of the scatter and obtain a more reliable age. Average D_e values and errors for each sample were calculated using the unweighted mean with Analyst (version 3.24) and LDBase software packages (developed by G.A.T. Duller). Several samples were also calculated using the central age model (CAM) in order to better identify age differences for dune accretion and elongation. The precision and accuracy of the OSL ages are analyzed in Chapter 3 (~~Section 4.4~~).

2.8. Spectroscopic analysis of sand grain redness

2.8.1. Background

2.8.1.1. The temporal significance of sand redness

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) images show the surface of reddened quartz sand to be covered in flakes and granular aggregates of hydrates of iron oxides with goethite (FeOOH) and hematite (Fe_2O_3), forming 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 known as rubification or brownification (Felix-Henningsen et al., 2008), defined as a change of soil color to yellow or red during intense weathering, thereby 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) usually 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 irrespective of 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, it is suggested that varying hues of red in sand indicate different depositional 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). 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. Thus, in some cases, sand redness quantification can potentially be a relative indicator of elapsed time.

It seems that time is an important factor for both laboratory experiments and the natural rubification process. However, there is no proof of a direct relationship between reddening and the age of sand using numerical dating. Grain residence time has been suggested as an important factor in reddening (Lancaster, 1989). Though inland sand rubification is a slow process in arid and semi-arid climates, distinguishable reddening can be attained in stable sand in less than 10 k years (Gardner and Pye, 1981). It has also been suggested that remotely sensed progressive rubification of presumably late Holocene Israeli coastal sands moving from the coast inland correlated to time (Ben-Dor et al., 2006), but this concept has not been proved by in-situ dating for either inland or coastal dunes.

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 homogeneous in color. Felix-Henningsen et al. (2008) reported that a section of a Nizzana VLD in the Negev had uniform brownification. These studies, which lacked spectral analysis, did not describe entire dune sections and neglected to include sufficient luminescence ages to investigate the relationship of sand redness to age.

2.8.1.2. Spectroscopy of sand redness

Laboratory spectroscopy provides a uniform measuring environment without the physical and spectral constraints of remote sensing and field spectroscopy — changing surface cover

(mixed pixels), variations in radiance relative to slope, atmospheric conditions, corrections, and varying observation angles. 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.

The redness index [(RI), $RI = R^2/(B \cdot G^3)$] was found to be a useful index for the quantitative spectral measurement of sand rubification in the laboratory (Ben-Dor et al., 2006; Levin et al., 2007). Tsoar et al. (2008) applied it to the Negev dunes. RI values, correlated to extractable iron oxide after Ben-Dor et al. (2006) and Tsoar et al. (2008) ($R^2 = 0.89, 0.67$ respectively) for Israeli coastal and Negev sand coatings, suggest that the index is suitable for quantifying sand grain coating redness. The RI was calculated using specific, albeit different, R, G, and B bands by Ben-Dor et al. (2006) and Tsoar et al., (2008). The dimensionless redness indices provide a ratio of relative redness. Based on the sand's 8-10 YR colors, it is suggested that goethite is the predominant iron-oxide sand grain coating in the NW Negev (Felix-Henningsen et al., 2008).

Continuum removal (CR) transformation of the NW Negev sands spectra showed a distinct absorption at 498 nm (Wenkart, 2006), which is close to that of goethite (485 nm) (Spectral Library, Grove et al., 1992), indicating the spectral potential to map this mineral. In the present research, the specific R, G, and B bands were selected after Ben-Dor et al. (2006), although both RI results are positively correlated ($R^2 = 0.94$) when the Ben-Dor et al. (2006) and Tsoar et al. (2008) bands are used.

2.8.2. Spectroscopic measurement

Laboratory spectroscopic preparation included carefully measuring 60 cc of split loose sand room-dried at 20° C for 24 h in plastic plates (Fig. 2.5g) in order to evaporate water 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 broken up by hand to reduce the pedons. Immediately prior to measurement, the sand samples were transferred to a 4 × 4-cm opaque, plastic black box and gently shifted to create a flat surface. Sand reflectance was measured with the 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 (1000W) 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 between the probe edge and the sand surface (Fig. 2.5h). Measurements from four directions were taken for each sample to avoid a Bidirectional Reflectance Distribution

Function (BRDF). All readings for each sample were averaged. 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's (EPIF) bias correction MatLab algorithm.

2.9. Landsat image processing

Landsat 5 TM images (Row 175, images 38, 39) from June 1987 (30 m/pixel) were used for mapping landforms, paludal deposits and sand redness. Since 1982, the relatively bare Negev dunes have been closed to Sinai Bedouin livestock grazing and wood gathering, leading to the rehabilitation of biogenic crusts and vegetation (Meir and Tsoar, 1996; Karnieli and Tsoar, 1995; Tsoar, 2008; Tsoar et al., 2008). By 1987, developing Negev dune vegetation and crust covers are presumed to have already created a bias in the Wenkart (2006) ferric index analysis based on Landsat imagery. To compare the Sinai results to those of the Negev and to minimize the effect of the biogenic crust on Negev surface reflectance, the 1987 images were nevertheless chosen because they were closest (earliest) to the land cover change that began in 1982. Another image, taken in August 2003, was examined for control and is mainly applicable to the relatively bare Sinai sands.

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) (Appendix C.1).

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's spectral resolution (Appendix C.2). An R^2 correlation of 90% was found between the ASD Fieldspec spectrometer RI and the resampled bands (Chapter 4; Table 4.2). Regional redness index maps of northeastern Sinai and NW Negev sands were processed using the RGB bands.

The spatial extent of the upper surficial standing-water/paludal deposits, believed to be formed by dune-damming of wadis (Magaritz and Enzel, 1990; Ben-David, 2003) in the northeast Sinai and the NW Negev, were mapped by supervised classification of a Landsat TM (2003) image mineral composite spectral enhancement (Appendix C.3).

Chapter 3: Age, origin and climatic controls on vegetated linear dunes in the northwestern Negev Desert (Israel)

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3.0 Abstract

The stabilized northwestern Negev vegetated linear dunes (VLD) of Israel extend over 1,300 km² and form the eastern end of the Northern Sinai – northwestern Negev Erg. This study aimed at identifying primary and subsequent dune incursions and episodes of dune elongation by investigating dune geomorphology, stratigraphy and optically stimulated luminescence (OSL) dating. Thirty-five dune and interdune exposed and drilled section were studied and sampled for sedimentological analyses and OSL dating, enabling spatial and temporal elucidation of the NW Negev dunefield evolution.

In a global perspective the NW Negev dunefield is relatively young. Though sporadic sand deposition has occurred during the past 100 ka, dunes began to accumulate over large proportions of the dunefield area only at ~23 ka. Three main chronostratigraphic units, corresponding to three (OSL) age clusters, were found throughout most of the dunefield, indicating three main dune mobilizations: late to post last glacial maximum (LGM) at 18-11.5 ka, late Holocene (2-0.8 ka), and modern (150-10 years). The post-LGM phase is the most extensive and it defined the current dunefield boundaries. It involved several episodes of dune incursions and damming of drainage systems. Dune advancement probably occurred in rapid pulses and the orientation of VLD long axes indicates long-term wind directions similar to the present. The late Holocene episode included partial incursion of new sand, reworking of Late Pleistocene dunes as well as limited redeposition. The modern sand movement only reactivated older dunes and did not lengthen VLDs.

This aeolian record fits well with other regional aeolian palaeoclimatic evidence. We suggest that sand supply and storage in Sinai was initiated by the Late Pleistocene exposure of the Nile Delta sands and by an increase in global gustiness during the LGM. Globally controlled LGM and continuing post-LGM gustiness transported the dune sands into the northwestern Negev.

Our results demonstrate the sensitivity of dunes located along the (northern) fringe of the sub-tropical desert belt to climate change (i.e. wind) and sediment supply.

3.1 Introduction

Dunes compose unique archives of past climates (Sarnthein, 1978; Lancaster, 2007; Telfer and Thomas, 2007; Lancaster, 2008; Telfer et al., 2010) and in many arid regions compose the main landform for palaeoclimatic research (Chase, 2009) that can infer on past winds upon the surface (Tsoar, 2005). Until recently, the chronological framework on dunes was constrained due to limited access to the dunes internal structure and lack of datable materials

(Singhvi and Porat, 2008). A recent proliferation of luminescence dating of quartz has highlighted the potential of dating of dunes. Recent works have addressed the need to achieve a certain density, quantity and depth of luminescence samples in order to reliably evaluate past periods of dune activity and stability (Bateman et al., 2003; Bateman et al., 2007; Telfer and Thomas, 2007; Stone and Thomas, 2008).

The northwestern (NW) Negev Desert dunefield constitutes the easternmost terminus of the 13,000 km² Northern Sinai Peninsula – NW Negev Erg (Fig. 3.1) (Tsoar et al., 2008), which includes the Northern Sinai dunefield and the NW Negev dunefield. The Erg is situated in the northern edge of global desert latitudes (N30°20' E32°15'– N31°10' E34°45') and has clearly defined borders (Fig. 3.1a). The northwest corner of the Northern Sinai dunefield is in fact part of the northeast Nile Delta stagnant Pelusiac branch (Sneh et al., 1986; Neev et al., 1987). Sand transported from the northeast Nile Delta into northwest Sinai is believed to be the Erg's sole sand source (Hunt, 1991) although this hypothesis requires testing.

The Northern Sinai dunes comprise mainly of active and elongating linear seif dunes and partially vegetated compound linear dunes (Abdel Galil et al., 2000; Rabie et al., 2000). Nevertheless, the Northern Sinai dunes of Egypt are currently not encroaching into the Israeli Negev section of the Erg. The Negev dunefield is composed of stable vegetated linear dunes (VLD) (Tsoar et al., 2008) that are covered by biogenic crusts (Danin et al., 1989; Kidron et al., 2009) although some dune crests are active. Thus it appears that the incursion of sands from Northern Sinai that created the dunefield occurred in an environment more conducive for sand mobilization and transportation than today (Tsoar et al., 2008). Situated at the downwind end of the Sinai-Negev Erg, the NW Negev dunefield constitutes an ideal setting for the study of dune encroachment chronologies.

Several geomorphological and archaeological studies have been published on the Northern Sinai and NW Negev dunefields. In Northern Sinai, the age of the dunes and sands were estimated at a few archaeological sites (Goldberg, 1977, 1986; Neev et al., 1987; Bruins, 1990; Gladfelter, 2000) (Table 3.1). In the northern sector of the Negev dunefield, artifacts have been found mainly on the stabilized dune and interdune surfaces, dating to the Byzantine period (100-400 years AD) and younger (Nahshoni and Aladjem, 2009). In the center of the Negev dunefield there is currently no archaeological finds. Archaeological dating has been conducted on sites in the south and eastern fringes of the dunefield (Table 3.1; Fig. 3.1c) mainly of Epilpaleolithic (22-11 ka) age (Goring-Morris and Bar-Yosef, 1987; Goring-Morris and Goldberg, 1990). Impressive remains of the Roman-Byzantine towns of Nizzana, Shivta, Saadon and Halussa delimit the dunefield in the south (Rubin, 1990). According to most of

these studies, the onset of dune encroachment into the NW Negev began in the Late Pleistocene (Magaritz and Enzel, 1990; Zilberman, 1991; Ben-David, 2003; Enzel et al., 2008; Tsoar et al., 2008) or, based on prehistoric sites, during the Epipaleolithic period (Goldberg, 1986; Goring-Morris and Bar-Yosef, 1987; Goring-Morris and Goldberg, 1990). It has also been suggested based on one site that the main dune incursion occurred during the Younger Dryas (Enzel et al., 2010). The Holocene, though interpreted as being generally more arid based upon archaeology (Goldberg, 1986), stream incision (Harrison and Yair, 1998) and speleothems (Vaks et al., 2006; Lisker et al., 2010) shows surprisingly limited and sporadic evidence of sand activity.

Aside from archaeological chronology, radiocarbon dating has been applied mostly to calcium carbonate deposits and nodules (Magaritz and Enzel, 1990; Zilberman, 1993) whose reliability is often questioned. Two well-developed (stage II-III) palaeosol development periods at 35-30 ka and 27-24 ka, and weakly developed (Stage I-II) palaeosols at 14-12 ka have been identified and dated by radiocarbon along the southern edge of the Negev dunefield (Table 3.1; Fig. 3.1c). The palaeosols are interpreted to postdate periods of sedimentation. (Zilberman, 1993), These periods have been suggested to be relatively humid in contrast with dune activity which was associated with a more arid climate (Goldberg, 1986; Goring-Morris and Goldberg, 1990; Zilberman, 1993). Thermoluminescence (TL) and infrared stimulated luminescence (IRSL) dating has been applied to scattered samples in the southern and southwestern part of the NW dunefield (Ben-David, 2003), mainly from interdune areas and stream terraces, but dunes and their underlying palaeosols were rarely targeted (Table 3.1; Fig. 3.1c). These ages suggest that dunes had limited lateral movement and that they have been in their current configuration throughout the Holocene.

A compilation of previous palaeoclimate interpretations of the northern Negev by Zilberman (1991) showed rapid humid-dry fluctuations with conflicting chronologies. This compilation distinguished between repeated cycles of Late Pleistocene climatic regimes with three phases: moist, characterized by dust (loess) deposition; semi-arid, allowing pedogenesis; and arid, characterized by stream incision and sand penetration. Ben-David (2003) suggested that as dunes have been in place since 25 ka, the wind regime has not changed since then. Vaks et al., (2006), based upon U-Th ages speleothem growth, shows that late Pleistocene rainfall in the northern Negev dropped below ~300 mm/a around 13-14 ka. Crouvi et al. (2008, 2009) shows loess deposition in the periphery of the Negev dunefield between 100-11 ka, a period covering several climate regimes. Accordingly, a methodological study on regional palaeoclimate and sedimentology of the northern Negev is necessary.

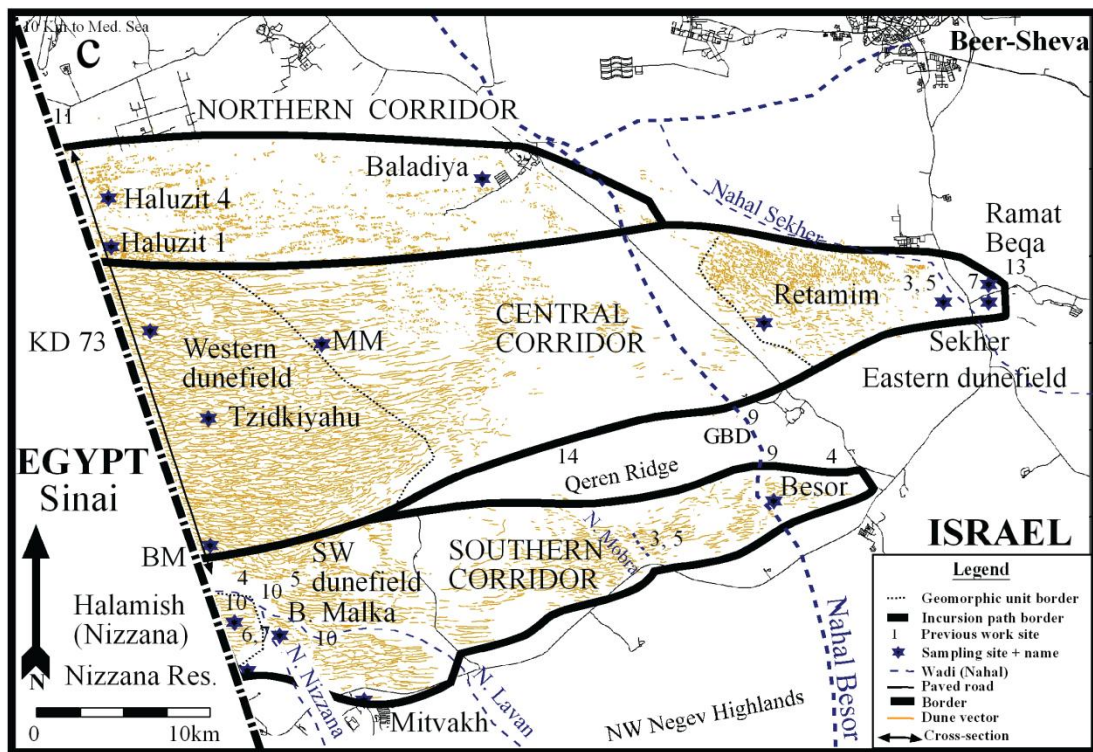
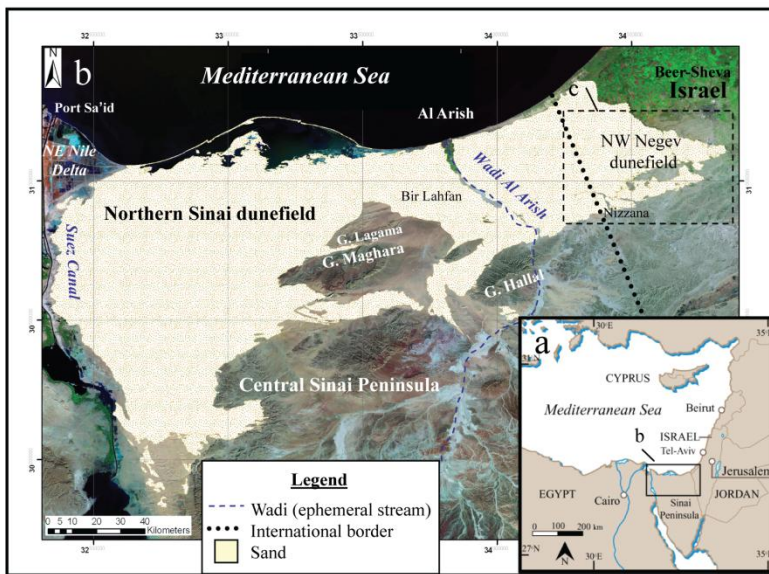


Figure 3.1

a. Location map of the eastern Mediterranean region. The Sinai-Negev Erg (Figure 1b) is outlined by a black box.

b. False Landsat (2000) composite image of the central and northern Sinai Peninsula and the western Negev Desert, Israel. The Sinai-Negev Erg, marked in light yellow, stretches south and parallel to the southeastern Mediterranean coastline from the northeastern Nile Delta across the Egypt-Israel border (dotted black line) into the northwest Negev. The northwest (NW) Negev dunefield which is the main research area is outlined in black. Note that in Sinai, the mountain ridges of Gebel (G.) Maghara and Lagama that block part of the dunes, and that Wadi Al-Arish is the only watercourse that crosses the entire Northern Sinai dunefield section.

c. Dune axis mapping results, sampling site names and incursion corridors (in capital letters). Dunefield regions [southwestern (SW), western and eastern] are also displayed and are referred to in the text. A geologic cross section of the central and northern incursion corridors east of the border appears in figure 7. Main sites of previous works are numbered in coordination with Table 1. Ben-David (2003), and Goring-Morris and Goldberg (1990) worked on several dozens of sites in the southwest and southern dunefield, respectively.

No.	Work	Year	Research location	Methods	Upper palaeosol (ka)	Initial sand encroachment (ka)	Dune buildup (ka)	Dune remobilization (ka)	Remarks
1	Goldberg (in Bar-Yosef and Phillips)	1977	Gebel Maghara, Northern Sinai	Radiocarbon dating of ostrich shells, charcoal and nodules.	>15.7-13.2	~40-33	Epipaleolithic 14.5-10		
2	Goldberg	1986	Southern Levant	"		25	20	Geometric Kebaran-Neolithic	Fluvial loess-silt deposition (Historic fill) 1.75-0.6 ka.
3	Magaritz and Enzel	1990	Nahal Mobara and Nahal Sekher	Radiocarbon dating of gypsum, ostrich shells, laminated carbonate minerals, charcoal and nodules.	12-14	15-20	17-17.5	2	Suggested sand incursion 30-34 ka.
4	Zilberman	1991, 1993	NW Negev	Radiocarbon dating of pedogenic carbonate nodules and charcoal.	12-14	30-25 (mid Late Paleolithic)	14-16 (Epipaleolithic)	2.2-3	Sand sheets
5	Goring-Morris and Goldberg,	1990	Southern NW Negev dunefield margins	Radiocarbon dating of carbonate in ostrich shells, charcoal and nodules.		22-16 (LGM)	Epipaleolithic (14.5-10)	Neolithic, Chalcolithic, Byzantine	
6	Rendell et al., Harrison and Yair	1993 1998	Halamish (Nizzana research site)	TL dating		9-43		6-10– since then stable	Age inversions in the dated section.
7	Tsoar and Goodfriend,	1994	Ramat Beqa quarry	Radiocarbon dating of hearths, amino acid epimerization of land snail-shells				6, 1.1-1.4 ka, 200yr.	
8	Gladfelter	2000	Wadi Gayifa, NE Sinai	Mainly archaeological artifacts	28±4.6 (Th/U)	20-14.5/10	15-9.5	in both, sand accumulated	22-15 ka dry; 15-11 ka - wet phases.
9	Greenbaum and Ben David	2001	Southeastern section of Negev dunefield	IRSL dating at GSI		98 (fine sand)	20		
10	Ben David	2003	Southwestern section of Negev dunefield	IRSL dating at GSI + TL dating		110 (sand)	30/25-12		One 67 ka sand exposure.
11	Zilberman et al.	2007	Qerem Shalom	OSL dating (SAR) at GSI	14.5±2.3 13.4±1.7	40-90			Composes base of upper sand soil unit.
12	Wieder et al.	2008	Ruhama	OSL dating (SAR) at GSI	13.6±1.2				
13	Crouvi et al.	2008, 2009	Three loess hilltop sections by dunefield periphery	OSL dating (SAR) at GSI	10.7±0.7 13.7±0.7	Pre/para-loess deposition	Not mentioned		
14	Enzel et al.	2010	Gulley in Qeren Ridge northern slope	OSL dating at GSI and calibrated radiocarbon dating on ash.			13	11-10	

Table 3.1: Previous ages in and adjacent to the study area. Radiocarbon dates of charcoal and ostrich shells were calibrated using Calib6.0. Radiocarbon dates from carbonate mineral and nodules are viewed only as a general estimation.

The seminal paper of Enzel et al. (2008) lays out a generalized palaeoclimatic scheme for the Eastern Mediterranean and southern Levant. It suggests a Mediterranean-controlled rainy and colder late Pleistocene north of the central Negev followed by a more arid Holocene throughout Israel,. The model though, lacks a detailed regional dating framework, especially for the northern Sinai and northwestern Negev dunes.

Advances in optically stimulated luminescence (OSL) dating (Murray and Wintle, 2000) have increased the feasibility, reliability and effectiveness of dating Quaternary deposits. This has generated a proliferation of research of inland quartzose dunefields. An OSL age provides the time of the end of exposure of quartz grains to direct sunlight, which occurs by burial by additional sediment. Therefore, OSL ages of dunes indicate the time of burial to a depth of several cm within an aeolian sand section. A large number of OSL ages facilitates the construction of a time-dependant framework of aeolian processes (Bateman et al., 2003; Chase and Thomas, 2007; Fitzsimmons et al., 2007; Miao et al., 2007; Telfer and Thomas, 2007).

This study is aimed at identifying primary and subsequent dune incursions and episodes of dune elongation by studying dune geomorphology/stratigraphy and OSL dating. The OSL chronostratigraphic framework of the study area has been compiled by designating sampling sites along VLD elongation corridors and by transecting the dunefield perpendicular to the transport pathways.

Here, we present sedimentological and stratigraphic attributes of the landforms in the Negev dunefield with emphasis on the vegetated linear dunes, combined with over 100 luminescence ages. These will be used to interpret the timing of genesis and growth episodes of the NW Negev dunefield. Our assumption at the onset of the study was that during the Late Quaternary there were several separate sand incursions into the NW Negev that created distinct geomorphic units in the dunefield. Each geomorphic unit may have accumulated in pulses mainly as a response to a certain climatic regime in which strong wind played a major role. Possibly, two or more geomorphic units accumulated simultaneously as their spatial distribution was dictated by different sediment supply sources or varying wind directions but similar pronounced wind intensities. We test these hypotheses through a detailed OSL chronology combined with field studies.

3.2 The research area

The NW Negev dunefield covers approximately 1,300 km² (Fig 1b) and is bordered on the west by the Northern Sinai dunefield of Egypt, in the south by floodplains and in the east by

an incised plateau composed of Lower Eocene carbonates (Avedat Group) (Zilberman, 1982), gently rising 10-50 m above the dunefield. However, this plateau does not appear to topographically block the migration of the dunes to the east. The dunefield is divided by the Qeren-Rogem anticlinal ridge (Qeren ridge) that trends WSW-ENE and protrudes 50-150 m above the dunes. The ridge is the most northerly exposure of the Northern Negev Syrian-Arc anticlinal system (Zilberman, 1982, 1991). It is composed of the Avedat Group carbonates and is dissected, mainly on its northwest flanks, by steep, short, small drainage systems that are in turn dammed by dunes (Tsoar, 1983; Enzel et al., 2010).

The portion of the dunefield south of the Qeren ridge fills an east-west synclinal depression and is locally bounded by wadis, Eocene chalk buttes and ridges. The dunes intercept and fill several wadis from the south such as Nahal (ephemeral stream in Hebrew, equivalent to wadi) Mobra (Blumberg et al., 2004). Dissected surfaces underlain by loamy sediments are evidence for palaeolakes created by dunes damming the wadis (Harrison and Yair, 1998; Ben-David, 2003; Blumberg et al., 2004) and larger drainages with Late Pleistocene flood plains (Zilberman, 1993).

The main aeolian sand body lies north of the Qeren ridge. It covers a gently seaward sloping landscape that was established by the receding Pliocene shore and later covered by a sequence of Pleistocene calcareous loam palaeosols (Bruins and Yaalon, 1979; Zilberman et al., 2007; Hatzor et al., 2009). Tsoar et al. (2008) classified the dunefield into three sectors, based on spectroscopic redness index of sand sampled from the surface. They suggested that the west-central part of the dunefield north of the Qeren ridge is the latest incursion due to its lower redness, while the northern and eastern fringes are the most mature.

The climate in the study area has been summarized by Littmann and Berkowicz (2008). Situated along a desert margin, between the semi-arid Mediterranean and the arid to hyper arid Negev, rainfall in the NW Negev mainly depends on the frequency and southerly extent of wintertime tracks of central and eastern Mediterranean cyclonic (Cyprus lows) fronts skimming the area. Some rain, mostly in the spring and autumn, is associated with the Active Red Sea Trough (ARST) systems (Kidron and Pick, 2000). The winds associated with the winter fronts mainly come from the southwest, west and northwest and have velocities of up to 20 m/s (Sharon et al., 2002). Summer winds are unidirectional with usually lower velocities (Allgaier, 2008; Tsoar et al., 2008). Nizzana, in the south of the study area, has drift potential (DP), directional variability wind index (RDP/DP) and resultant drift direction (RDD) values [terminology from Fryberger (1979)] that vary between 21 and 108 vector units, 0.48-0.73 and 241° - 289° respectively (Tsoar et al., 2008). The resultant drift potential (RDP) and RDD

indicate the main winds are from the west, consistent with the dune orientation. The low DP and RDP values are a measure of the low-energy wind environments which explains, at least in part, the current natural stable status of the dunes (Tsoar, 2005; Yizhaq et al., 2009). This observation emphasizes the extreme changes in environmental conditions that are required to initiate dune activity.

Average annual precipitation in the research area is approximately 150 mm in the north, decreasing to 80 mm in the south at Nizzana, though in the last decade rainfall has overall decreased by 40% (Siegal, 2009). At Nizzana, several storms have been found to bring 10-30 mm of precipitation in daily events (Kadmon and Leschner, 1995; Sharon et al., 2002; Almog and Yair, 2007). Potential evaporation is 2000-2200 mm/yr at Nizzana (Littmann and Berkowicz, 2008). Despite the low precipitation and high evapotranspiration, biogenic microphytic (soil) crusts preserve the dunes from reactivation by strong winds, unless they are trampled or covered (Almog and Yair, 2007; Kidron et al., 2008, 2009), even when perennials have wilted (Siegal, 2009).

3.3 Methods.

3.3.1 Field Methods: Site selection and sampling procedures

The dunefield was first studied using aerial photographs and Landsat images. Preliminary analysis using an ArcMap 3D module of DEM from SRTM, characterized dune morphometries. Dune crests were mapped from orthophotos at a scale of 1:5,000 in a GIS and validated by field surveys. Geomorphic units were qualitatively classified based upon dune crest orientation and spatial density, and dune morphology and cross-section morphology. This approach was motivated by the assumption that mature stable dunes may have degraded (O'Connor and Thomas, 1999; Lancaster, 2007) and thus different dune morphologies may represent different histories of buildup and stabilization. Ultimately, the geomorphic units were merged into three main west-east trending dune bodies that delimit discrete incursion corridors, partially consistent with Tsoar et al. (2008) (Fig. 3.1c; Table 3.2).

Sampling strategy for sedimentological analyses and OSL dating was designed to identify the earliest dune incursions and to analyze elongation/advancement rates (Table 3.3). The dunefield was sampled along 5 lines; western and eastern north-south transects and a west-east transect along each incursion corridor (Fig. 3.1c). The NNW –SSE sampling line (the "western transect"), located at the western end of the study area along the Israel-Egypt border, transects the VLD orientations sub-diagonally. Sampling was performed along this line in

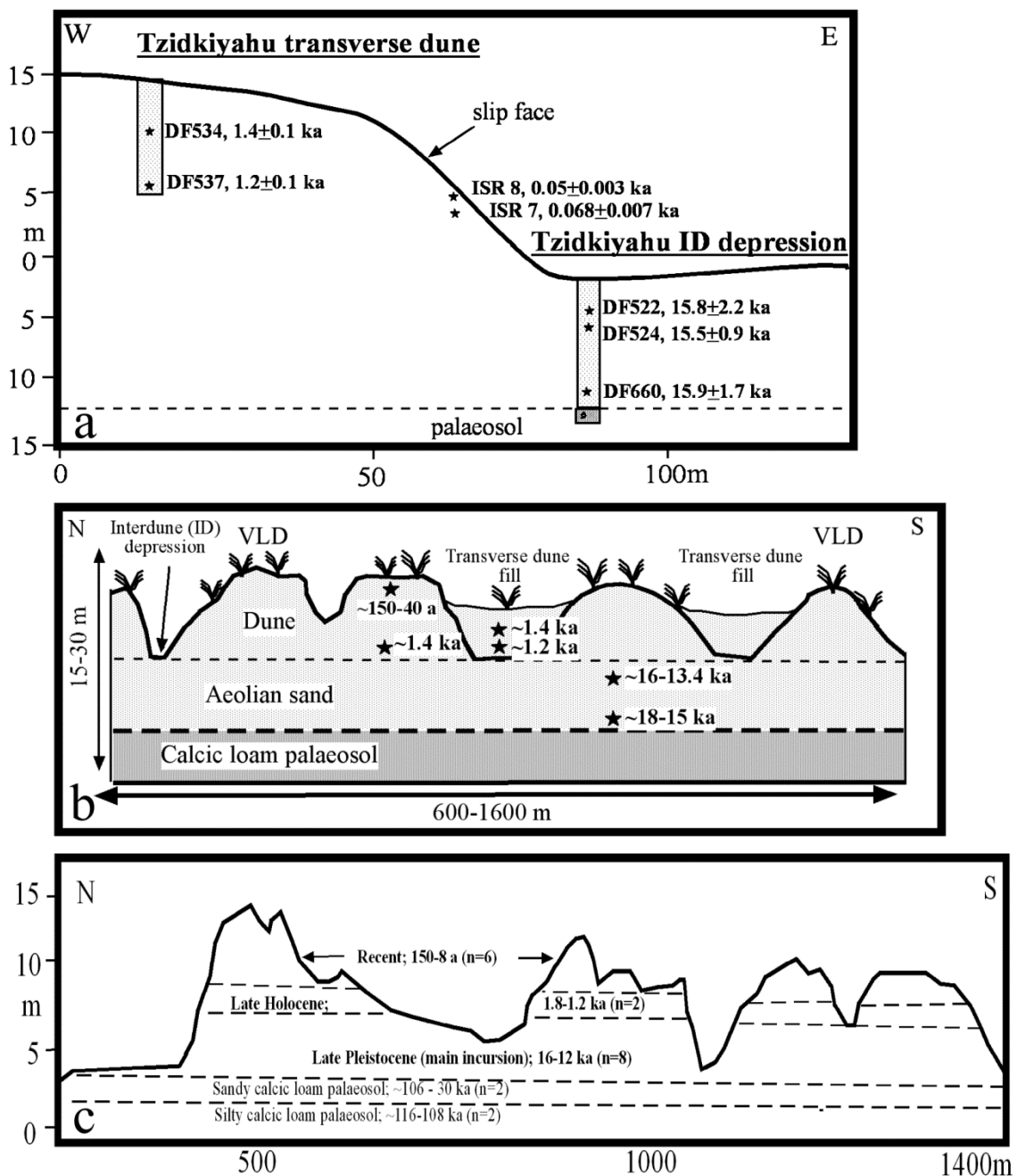


Figure 3.2

- Cross section of the late Holocene Tzidkiyahu transverse dune in the central incursion corridor. Dune advancement direction is from west to east. An older 5-10 m thick aeolian sand section underlies the transverse dune and overlies a calcic loam palaeosol.
- Schematic cross-section of aeolian sand and dunes in the western part of the central incursion corridor. Transverse dunes (see 2a) fill the interdunes between 8-12 m high late Holocene vegetated linear dunes. For a regional cross-section see figure 9.
- The chronostratigraphy of the Negev VLDs as found at the Haluzit 1 exposed dune section. Throughout the dunefield, unit thickness varies, though the general chronostratigraphy is similar.

every geomorphic unit, both from dune crests and interdune topographic lows. In the eastern transect, we sampled the easternmost extent of each incursion corridor, to date dune advance and cessation. Sampling was aimed at retrieving the dune-base sand (the lowest 1 m) and the underlying substrate, to detect the earliest sand activity and to create a stratigraphically uniform age database which has been lacking from previous studies (Telfer and Thomas, 2007). Dunes axes were generally targeted for sampling to obtain the dune core sediment presumed to be the least affected by possible slight lateral dune sand movement. Five dune flanks were sampled in order to understand VLD elongation and buildup dynamics. One VLD was sampled at two sections located along its axis in order to investigate the longitudinal plunge, elongation and narrowing of the dune over time. A majority of the sections were sampled from exposures. Dune stratigraphy was described using standard sedimentological and pedological methods (Dan et al., 1964; Birkeland, 1999). Sampling points were chosen for each unit at least 10 cm from contacts and usually in mid-unit. Sampling involved driving hard opaque 20 cm long by 3 cm diameter plastic pipes into the exposure by hand or hammer. Drilling was performed with Dormer Engineering hand augers, mainly manually or assisted with a Drillmite 6Hp hydraulic engine. Sampling for OSL dating was performed at 1.5 m intervals unless field examination of the samples revealed changes in sediment properties, in which case sample density was increased. OSL sampling usually began 1-2 m below the surface to avoid the bioturbated and active dune crests.

Incursion corridor	Annual rainfall (mm)	Incursion corridor length in the Negev (km)	Dune crest elevation (m)	Dune cross-section approximate widths (m)	Dune cross-section sand erodibility	Slip face orientation	Biogenic crusts thickness (mm)	Est. vegetation cover (%)	Interdune characteristics
Northern	140-160	25-30	W: 5; E: 12-18	W: 100-150; E: 400-500	Fully encrusted.	No slip faces.	7.6	12-17	1-3 m thick sands over palaeosol
Central	100-130	50-55	W: 20; C: 5; E: 5-10	W: 150; C: 50 + 200; E: moderate morphologies	Sporadic active crests.	Changes annually, mainly northern.	4.2	10-12	Tranverse dunes 200-500 m long between depressions with a 5-10 m thick sand sequence overlying a palaeosol.
Southern	80-90	30-35	W+C: 10-15; E: 5-10	W+C 50+200 E: 30-50 m, moderate morphology	Active 10-50 m. wide crests.	Mainly northern.	2.7	5-10	Fluvial and standing water deposits interchanging with sand.

Table 3.2: The morphological characteristics of the three dune incursion corridors. Rainfall data and vegetation cover is after Siegal (2009). Biogenic crust thickness is after Almog and Yair (2007). Dune crest elevations and widths were retrieved from measurements by a total station in regard to west (W), east (E) and central (C) parts of the incursion corridors. Sand erodibility estimations are based on field work and a supervised classification of a Landsat (2000) image. Slip face orientations are based on field work and aerial photograph interpretation.

3.3.2 OSL dating

3.3.2.1 Sample preparation

Thirteen preliminary samples (labeled ISR in Table 3.3) from shallow depths were dated by OSL at the Marburg Luminescence Laboratory, Germany. Additional 84 samples were prepared and measured at the Luminescence Laboratory of the Geological Survey of Israel (GSI), Jerusalem. Sample preparation follows Porat (2007). Briefly, dry sieving isolated grain fractions of 125-150 μm or 150-177 μm , followed by immersion in 8% HCl to remove carbonates. After washing and drying heavy minerals and most feldspars were separated from the quartz with a Frantz magnetic separator using a high (1.5A) current (Porat, 2006). Subsequently, quartz was etched with concentrated (40%) HF for 40 min, to etch grain rinds affected by α particles and dissolve any remaining feldspars, followed by 16% HCl treatment to remove any precipitated fluorides.

Approximately 1000 grains were mounted on 10 mm aluminum discs with 5 mm masks using silicon (oil) spray as an adhesive. Equivalent dose (D_e) determinations used a modified single aliquot regenerative-dose (SAR) protocol (Murray and Wintle, 2000) that included a cleaning step of heating to 280°C for 100 s at the end of each measurement cycle. The protocol started with measuring the natural signal, followed by a zero dose point to test for thermal transfer, three beta dose points, a second zero, a repeated dose (recycling ratio) and a second repeated dose after infrared (IR) bleaching (IR depletion ratio). Measurements were carried out either on a DA-12 or DA-20 TL/OSL Risø Readers equipped with blue LED's. Irradiation was from a calibrated ^{90}Sr β source and the luminescence signal was detected through 7 mm U-340 filters.

Dose recovery tests over a range of preheats showed that in the preheat range of 220-280°C the ratios between measured and given doses is 0.9-0.95, similar to values found by Murray and Wintle (2003) and by Pruesser et al. (2007) for glacial deposits. Thus preheats of 220⁰-260⁰ C for 10 seconds were used before OSL measurement at 125⁰ C. Test dose cutheat was applied for 5 seconds at twenty degrees lower than the preheat. The D_e of each sample was determined by fitting a linear+exponential curve to the data points.

For most dune and interdune samples, age calculations relied on a D_e averaged from a minimum of 13 measurements (aliquots) per sample, with approximately half of the samples relying on more than 17 measurements. However, very old (~100 ka) and some of the very young (0-150 years) samples were measured on only 8-13 aliquots as these ages were not the focus of this study. Several samples had a high scatter and further 24 to 48 2 mm (~150

grains) aliquots were measured to assess the source of the scatter and obtain a more reliable age. Average De values and errors for each sample were calculated using the unweighted mean.

3.3.2.2 Dose rate determination

Where possible, cosmic and γ dose rates were derived from *in situ* measurement using a calibrated portable Rotem P-11 gamma scintillator with a 2" sodium iodine crystal (Porat and Halicz, 1996). For drillholes, cosmic dose rates were estimated from burial depths, though changing burial depth over time was not considered. Age calculation relied mainly on the cosmic dose rates based on the sample depth. Chemical analyses of radioactive elements (K, Th and U) of the sediments were done by inductively coupled plasma mass-spectrometry or atomic emission spectrometry (ICP-MS/AES), and their concentrations were converted into α , β and γ dose rate using the factors by Nambi and Aitken (1986). A moisture content of 2 ± 1 %, based on moisture measured on oven-dried samples, was chosen for age calculations of the sand samples. For samples that have >25% fines, a moisture content of 6 ± 2 % was used.

3.3.3 Particle size distribution and mineralogy

Particle size distribution (PSD) was measured using a laser-diffraction Malvern Mastersizer MS-2000. Samples were split to 5 g, sieved to < 2 mm, and stirred for dispersion for 10 min in sodium hexametaphosphate solution followed by ultrasonification for 30 s. Three replicate aliquots for each sample were run, and after good reproducibility was achieved modified to two aliquots. Each aliquot was subjected to three consecutive 5-second runs at a pump speed of 1800 RPM. The raw laser diffraction values were transformed into PSD using the Mie scattering model. Optical parameters were RI=1.52 and A=0.1.

Semi-quantitative abundance of quartz, calcite and plagioclase were determined by X-ray diffractometry (XRD) peak heights (quartz= $20.8^{02\theta}$; plagioclase= $27.9^{02\theta}$; calcite= $29.4^{02\theta}$).

3.4 Results

3.4.1 Dune morphology and field relations

The NW Negev dunefield VLDs orientations show a general west-east orientation (Fig. 3.1). Dune lengths are usually limited to several kilometers before coalescing, often in Y junctions (Tsoar and Moller, 1986; Tsoar et al, 2008). The Negev VLDs differ from VLDs that extend for many kilometers, as found for example in Australia (Folk, 1971). This

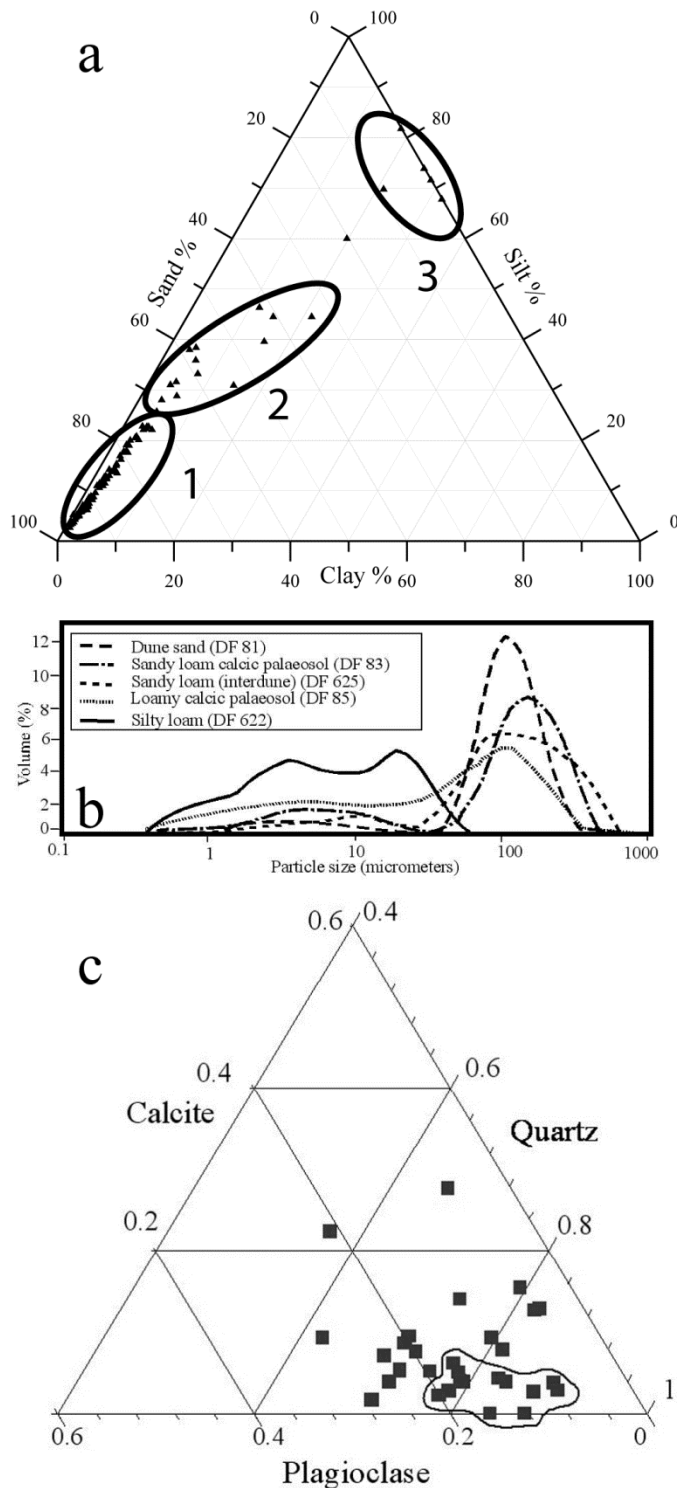


Figure 3.3

a. A ternary diagram showing the relative abundance of sand, silt and clay. Ellipse 1 contains dune sand, ellipse 2 contains loamy palaeosols, and ellipse 3 contains standing water silt loams.

b. Particle size distribution of sediments from various depositional environments in the NW Negev dunefield.

c. Ternary diagram showing relative abundance of minerals for selected sand samples, determined by XRD. Sand samples with relatively high calcite proportions are found along dunefield margins and mainly in the southern incursion corridor while in the central incursion corridor samples tend to have more than higher quartz values (black polygon).

constrains the ability to date accurately single dune elongation rates. Rather, dune elongation rates can be measured along short increments or along longer general west-east azimuths. Interdune (ID) spacing between VLDs ranges from 100-400 meters.

Interdune surfaces are composed of two different sediments. In the northern and southwestern sectors and north of the Qeren ridge, the interdune corridor is composed of flat loamy aeolian and playa sediments with a sparse sand cover. In the central incursion corridor, the interdunes are mostly filled with usually structureless, aeolian sand 0.5-10 m thick (Fig. 3.2a).

Each dune's morphology changes towards its easternmost end (Table 3.2). Dune width, height and width/height ratios also differ along and between the three major west-east incursion corridors. This differs significantly from linear dunes in the Namib Desert that have uniform height for great distances (Livingstone, 1989). The northern incursion corridor VLDs are broad and low, though at the northeast corner of the dunefield (Baladiya) the VLDs are broad and high and are covered by *Tamarisk aphylla* trees originally planted in the 1930's (Lipshitz and Biger, 2004). The central incursion corridor is characterized in the west by 5-10 m thick interdune aeolian sands that are overridden by VLDs and transverse dunes. Transverse dunes with 5-10-meter-high east-facing slip faces fill the interdune areas between the VLDs. Here, the cumulative aeolian sand thickness of the dunefield attains a maximum of 25-30 m. (Fig. 2a, b). Further east, the sands cross Nahal Besor and relatively deep dune and interdune sections are found at the Retamim and Baladiya sections, respectively (Fig. 3.1c). The furthest eastern lobe of the dunefield has dunes with less distinct morphology. Sands breach a short section of Nahal Sekher and fill several western facing wadis/depressions in the Ramat Beqa plateau (Fig. 3.1c).

In the southern incursion corridor (south of the Qeren ridge) interdune areas is mainly composed of fluvial sediments. Dune substrates include fluvial sediments, calcic palaeosols and Plio-Pleistocene terrace deposits (Zilberman, 1991; Ben-David, 2003).

3.4.2 VLD stratigraphy and internal structure

Understanding the stratigraphic setting and buildup of the internal dune structure is a prerequisite for interpretation the OSL ages of a single dune section and of the entire dunefield (Fitzsimmons et al., 2007). This is highly important as previous works such as in the Kalahari (Telfer and Thomas, 2007; Stone and Thomas, 2008) have not fit the OSL ages to an internal dune structure. Before discussing the dune structure there is a need to define a few key terms. Sand mobilization relates to any sand activity, may it be of a sand sheet or

dunes. Dune buildup refers to vertical accretion of sand, and dune elongation indicates extension of a linear dune along its axis. During dune buildup and elongation, the dune sand undergoes phases of erosion and accretion making the stratigraphic sequence discontinuous with unconformities (Bateman et al., 2003; Munyikwa, 2005) that are often difficult to recognize.

The wide extent of abundant exposed sections enabled us to identify stratigraphic contacts and thus specifically target our OSL sampling points. This significantly improved our understanding of the ages obtained from additional drilled sections. Exposed vertical outcrops of dune and interdune cross-sections, (Figs. 3.2c & 3.4) also enabled close examination of the dunes internal structure. At the Haluzit sections, bulldozed trenches of VLDs exceeding 1 km along and perpendicular to their axis, exposed rare and short-lasting full cross and longitudinal dune sections (Fig. 3.2c).

The three main stratigraphic units found in sections and cores drilled into a VLD along the dune axis are; 1. The substrate that underlies the dune. 2. The main bulk of the dune interior. 3: The upper 1-3 meters of the dune slopes and crest, named here the dune mantle. This division is based upon identification of horizontal sedimentary units of dune sand and substrate, employing criteria such as the completeness, size and relative abundance of land snail shells, bedding, hue and consistency of the sand, and carbonate contents. Laboratory analyses such as PSD and OSL dating further contributed to this division.

Dune substrates include in the center and north palaeosols that are easily identified in exposed sections (Haluzit 1 section) and in cores by a darker color, 2-20 mm concentric carbonate nodules and a finer sandy-silty loam texture due to a mixture of sand and loess (Fig. 3.3b). The presence of the carbonate nodules at the top of the palaeosol attests to truncation of the palaeosol's A and upper B soil horizons with thicknesses of several tens of cm, possibly by dune sand erosion. Lag deposits containing carbonate nodules and clay pellets at the base of some dunes attest to surface windiness and sand erosion.

In the southern incursion corridor dunes are underlain by floodplains (Ben-David, 2003). The only evidence of ancient watercourses beneath the central and northern incursion corridor dunes is found in the Baladiya drillings that penetrated gravels beneath the dune section (Machta, 2005). These may have been deposited by the lower Nahal Mobra prior to dune encroachment (Blumberg et al., 2004).

The VLD interior structure reveals several stacked sand units with up to two horizontal to sub-horizontal contacts which can be identified by slight changes in consistency, color and particle size. Bedding is rarely apparent in the Negev dunes (other than for the upper 2-3

meters), similar to observations in the Strzelecki Desert of Australia (Telfer and Thomas, 2007; Cohen et al., 2010) and the Kalahari linear dunes (Telfer and Thomas, 2007). Bioturbation may explain the lack of bedding in some dunes of the Negev. We observed krotovina and cicada burrows that have previously been described in the southwest dunefield (Halamish) (Filser and Prasse, 2008) and in other aeolian environments (O'Geen and Busacca, 2001). Bioturbation may also explain the scatter in some of the OSL data. At Haluzit, an exposure along the VLD long axis shows continuity of the stratigraphic units in dune interior and mantle, implying that the VLD axis core is relatively stable and is not constantly reworked by bi-directional winds that are characteristic of seifs and non vegetated linear dunes (Tsoar et al., 2004; Bristow et al., 2007). These observations suggest that the Negev VLD's are extending-elongating forms that at specific episodes deposit several horizontal to sub-horizontal stacked units that can be traced along the dunes axis. The horizontal unit contacts are suggested to be formed by wind erosion that usually initially includes erosion of the existing unit followed by a subsequent depositional phase. The finds show that the VLD internal structure has a net accumulative buildup that is not fully reworked, interrupted by hiatuses representing periods when only the upper dune section was partially active. Thick chronostratigraphic units exceeding 2-3 meters probably did not undergo bioturbation to an extent that penetrated and mixed the middle-lower parts of the unit.

The VLD mantle surface includes 5-15% vegetated dune crests that commonly lack a biogenic crust cover and are at least partially active (Tsoar et al., 2008). In contrast, even steep dune flanks host a biogenic crust. The dunes' internal structure reflects its external physiography. Cross-bedded sets (5° - 25°) and dips are identifiable and are separated by a clear contact from the main dune section. Roots are common along with approximately 1% organic material. In the upper 1-2 meters of some dune crests and slopes, where dune sand contains minimal moisture, thin remnants of covered biogenic crusts were found, indicating shallow burial by very recent reworked sands. Thus the upper 1-3 meters of the VLD, whether active or encrusted, contrasts sharply with the dune interior.

Dune mantles attest to recent surficial aeolian activity but the recent reworking and additions may not contribute to elongation and net accumulation (see Allgaier, 2008). These recent reworking of sand occurred mainly in a period when grazing was active and vegetation and crust cover was minimal (Meir and Tsoar, 1996; Tsoar, 2008). Similar young active dune mantles have been reported on dunes in the Kalahari Desert (Thomas et al., 1997; O'Connor and Thomas, 1999).

Palaeosol indicators such as carbonate horizons are a useful marker for dune stabilization (Fitzsimmons et al., 2007). The absence of any obvious palaeosol in the Negev hinders the possibility of extracting reliable dune accumulation OSL ages especially when sampling is from coring and at intervals of 1- 2 meters (Telfer and Thomas, 2007; Stone and Thomas, 2008). The discrete NW Negev VLD structure though, as found at several exposed sections, suggests that the base of the dune and the section beneath the dune mantle should give reliable ages of the main dune buildup and elongation episodes even when retrieved from drills.

Table 3.3 (next 5 pages): Optically stimulated luminescence (OSL) ages with field and laboratory data organized according to incursion corridors. VLD=vegetated linear dune; TD=transverse dune; D=Interdune; E=exposure; A=sampled by an auger. Dunes were sampled from their axis unless mentioned. γ +cosm – measured in the field; Calc. γ - calculated from radioelements; Cosm. - estimated from burial depth; sa – small (2 mm) aliquots. All ages are in thousands of years (ka) except for ages below 100 years (a) which are in years and italicized.

The ISR-labeled samples were analyzed and calculated by Anja Zander at the Marburg Luminescence Laboratory, Department of Geography, Philipps-University Marburg (see Supplementary Data Captions).

Site and Sample	Sampling method	Depth (m)	γ +cosm. (μ Gy/a)	Calc. γ (μ Gy/a)	Cosm. (μ Gy/a)	Grain size (μ m)	K (%)	U (ppm)	Th (ppm)	Ext. α (μ Gy/a)	Ext. β (μ Gy/a)	Total dose (μ Gy/a)	No. of discs	OD (%)	De (Gy)	Age (ka)	Morphology & Comments
Northern incursion path																	
<u>Haluzit 4</u>																	
DF-31	E	0.55	490			150-177	0.56	0.7	2.1	2	513	1005 $\pm$ 54	10/12	66	0.09 $\pm$ 0.05	85$\pm$45 a	VLD
DF-32	E	1.15	490			150-177	0.61	0.6	1.7	2	525	1017 $\pm$ 54	11/12	68	0.14 $\pm$ 0.1	0.14$\pm$0.09	
DF-34	E	1.9	576			150-177	0.71	0.8	2.8	3	643	1222 $\pm$ 62	10/13	23	1.8 $\pm$ 0.2	1.4$\pm$0.2	
DF-35	E	3.3	507			150-177	0.75	0.6	1.7	2	619	1128 $\pm$ 56	13/13	14	12.0 $\pm$ 1.8	10.6$\pm$1.6	
DF-301	A	3.25		317	141		0.73	0.5	1.9	2	605	1065 $\pm$ 28	13/13	10	9.4 $\pm$ 0.8	8.9$\pm$1.0	VLD continuation, 150 m downwind of exposure
DF-302	A	3.9		330	131	125-150	0.72	0.6	2.0	2	613	1077 $\pm$ 32	13/13	12	12.9 $\pm$ 1.7	12.0$\pm$1.6	
DF-304	A	5.1		319	115	125-150	0.73	0.6	1.7	2	613	1048 $\pm$ 27	17/17	12	13.4 $\pm$ 1.6	12.8$\pm$1.5	
DF-308	A	6.9		457	95	88-125	0.71	1.5	3.5	6	699	1256 $\pm$ 40	8/8	18	145 $\pm$ 30	116$\pm$23	
<u>Haluzit 4 Hothouse</u>																	
DF-41	E	1.0	490			150-177	0.7	0.6	2.0	2	592	1084 $\pm$ 54	13/13	7	10.4 $\pm$ 0.9	9.6$\pm$0.9	ID
DF-42	E	2.0	573			150-177	0.83	0.6	1.7	2	673	1247 $\pm$ 61	13/13	11	15.4 $\pm$ 1.9	12.3$\pm$1.7	
<u>Baladiya</u>																	
DF-75	E	2.4		333	156	150-177	0.75	0.6	1.9	2	624	1114 $\pm$ 27	7/24	67	3.4 $\pm$ 0.7	3.0$\pm$0.6	VLD
DF-76	E	3.2		321	142	150-177	0.55	0.8	2.2	3	521	987 $\pm$ 28	13/13	12	13.5 $\pm$ 1.6	13.7$\pm$1.7	
DF-714	A	5.7		271	107	125-150	0.72	0.36	1.3	2	566	946 $\pm$ 33	17/17	13	14.8 $\pm$ 2.2	15.6$\pm$0.7	
DF-715	A	8.0		266	85	125-150	0.66	0.4	1.4	2	532	885 $\pm$ 29	19/19	10	14.2 $\pm$ 1.6	15.9$\pm$0.7	
DF-719	A	9.8		345	72	125-150	0.76	0.6	2.1	3	642	1061 $\pm$ 33	17/19	28	15.6 $\pm$ 2.0	14.7$\pm$1.9	
<u>Haluzit1</u>																	
DF-53	E	1.8		350	168	125-150	0.66	0.7	2.5	3	597	1118 $\pm$ 27	9/13	53	0.09 $\pm$ 0.03	75$\pm$30 a	VLD
DF-60a	E	2.6	500			150-177	0.62	0.6	1.6	2	529	1031 $\pm$ 54	13/13	26	0.06 $\pm$ 0.02	60$\pm$20 a	Northern VLD flank
DF-802	E	2.9		365	147	125-150	0.76	0.7	2.3	3	660	1175 $\pm$ 30	12/12	16	2.1 $\pm$ 0.4	1.7$\pm$0.3	
DF-803	E	3.7		289	134	125-150	0.71	0.5	1.4	2	579	1004 $\pm$ 28	14/14	5	13.8 $\pm$ 0.9	13.7$\pm$0.9	
DF-804	E	4.5		324	122	125-150	0.76	0.5	1.9	2	625	1074 $\pm$ 31	14/14	6	14.7 $\pm$ 1.1	13.7$\pm$1.1	
DF-81	E	6.8		377	96	125-150	0.83	0.7	2.2	3	705	1180 $\pm$ 35	12/13	10	18.4 $\pm$ 2.6	15.5$\pm$2.2	

DF-83	E	7.5	320	89	125-150	0.65	0.7	2.5	3	537	949±33	13/13	29	101±17	106±19
DF-85	E	8.5	523	81	125-150	0.91	1.6	3.8	4	809	1417±33	11/11	17	153±30	108±22

Haluzit

ISR 6	E	0.24		207±10	150-200	0.72	1.09	3.02			1400±70	24/24		0.07 ± 0.01	51±4 a
ISR 5	E	0.57		202±10	150-200	0.77	0.64	2.35			1100±55	24/24		0.11 ± 0.01	96±7 a

VLD crest

Central incursion path

KD 73 depression

DF-681	A	2.0	276	164	125-150	0.61	0.6	1.4	2	524	966±27	11/13	15	12.8±1.3	13.3±1.4
DF-685 sa	A	6.0	216	104	125-150	0.46	0.5	1.1	2	402	724±31	15/24	69	13.0±1.9	17.9±2.8

ID

KD 73

DF-695	A	9.2	256	76	125-150	0.55	0.5	1.5	2	473	806±30	17/17	8	12.6±1.1	15.6±1.5
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VLD

MM

DF-11	E	1.25	665		150-177	0.79	0.6	2.1	2	655	1322±70	10/13	53	0.06±0.01	45±10 a
DF-13	E	2.6	886		150-177	0.71	0.6	1.7	2	592	1480±91	13/13	19	0.06±0.01	40±10 a
DF-16	A	5.7	610		150-177	0.79	0.7	2.0	2	665	1278±65	13/13	5	1.6±0.1	1.3±0.1
DF-17A sa	A	7.0	315	94	150-177	0.70	0.8	1.3	2	601	1012±27	24/25	27	9.4±2.0	9.3±2.0
DF-18 sa	E	1.1	777		125-177	0.83	1.0	3.5	4	751	1532±82	12/24	68	15.0±2.8	9.8±1.9

VLD

Retamim Plain

ISR 4	E	0.26		214±11	150-200	0.71	0.42	1.89			1200±60	24		1.33 ± 0.19	1.2±0.1
ISR 3	E	0.43		212±11	150-200	0.59	0.33	1.52			1000±50	24		1.18 ± 0.02*	1.2±0.1

ID

Retamim dune base

DF-541	A	1.7	250	170	125-150	0.63	0.4	1.2	2	507	928±27	17/17	10	1.4±0.1	1.5±0.2
DF-543	A	3.3	195	140	125-150	0.46	0.4	0.9	1	385	722±28	13/13	12	11.5±1.5	16.0±2.1
DF-545	A	4.6	181	121	125-150	0.42	0.4	0.8	1	355	659±26	13/13	8	12.7±1.1	19.3±1.8
DF-548	A	6.7	268	97	150-177	0.62	0.5	1.4	2	512	878±31	22/25	20	23.9±3.2	27.2±3.8
DF-700	A	7.6	429	88	125-150	0.83	1.0	2.6	4	753	1273±34	30/31	14	29.0±3.8	22.8±3.1

ID

Retamim VLD

Broad VLD

DF-568	A	7.8	208	87	150-177	0.50	0.3	1.2	1	402	697±29	13/13	12	7.5±1.0	10.7±1.5	
<u>Ramat Beqa quarry</u>																Infill of topographic trough
DF-579	E	4.3	252	125	125-150	0.64	0.4	1.2	2	514	893±24	13/13	9	5.0±0.5	5.6±0.6	
DF-578	E	4.85	291	118	150-177	0.66	0.5	1.7	2	546	956±33	13/13	12	4.6±0.6	4.8±0.7	
DF-575	E	8	458	85	125-150	0.83	1.1	3.0	4	765	1308±38	18/19	16	15.2±2.3	11.6±1.8	
ISR 2	E	1.36		200±10	150-200	0.70	0.41	1.67			1100±55	24		9.11 ± 0.28*	8.2±0.4	Quarry wall at mid-section
ISR 1	E	2.6		183±9	150-250	0.81	0.7	3.22			1400±70	24		12.0 ± 0.25*	8.8±0.5	"
<u>Nahal Sekher VI site, southern section</u>																Undulating vegetated sand cover
NS-1	E	0.5	247	210	125-150	0.61	0.4	1.3	2	490	949±34	15/17	25	3.0±0.5	3.2±0.5	Above artifacts
NS-2	E	0.75	258	193	125-150	0.59	0.5	1.4	2	492	944±34	16/17	15	11.3±1.3	11.9±1.4	Below Natufian artifacts
NS-3	E	1.5	235	174	125-150	0.57	0.4	1.2	2	467	878±31	17/17	8	12.0±1.1	13.7±1.3	
NS-4	E	2.65	316	151	125-150	0.65	0.34	2.9	2	534	1004±34	19/19	12	11.4±1.6	12.4±1.8	
<u>Nahal Sekher VI site, northern section</u>																Undulating vegetated sand cover
NS-5	E	1.6	290	172	125-150	0.64	0.6	1.6	2	542	1006±33	17/17	11	3.0±0.4	2.9±0.4	
NS-6 sa	E	1.8	310	168	125-150	0.66	0.7	1.7	2	571	1051±31	22/24	37	4.0±1.3	3.8±1.2	Lag + Natufian unit
NS-7	E	2	292	164	125-150	0.61	0.6	1.8	527	985	985±35	17/17	7	12.3±1.1	12.3±1.2	
<u>Nahal Sekher XXX site</u>																Edge of undulating vegetated sand cover at top of southern N. Sekher bank
NS-11	E	0.45	262	214	125-150	0.63	0.5	1.3	2	516	1004±29	15/17	17	11.8±1.0	11.5±1.3	
<u>Nahal Sekher</u>																Reworked loess (sandy loam) top adjoining edge of sand cover
NS-10	E	0.3	439	219	125-150	0.74	1.0	3.6	4	690	1352±36	19/19	14	11.2±1.8	9.0±1.5	
<u>Tzidkiyahu Transverse</u>																TD
DF-534	A	4.6	230	121	125-150	0.52	0.5	1.1	2	443	796±30	8/11	29	1.1±0.1	1.4±0.1	
DF-537	A	7.85	213	86	150-177	0.58	0.3	0.9	1	448	748±29	13/13	9	0.88±0.09	1.2±0.1	
ISR 8	E	0.28		208±10	150-200	0.74	0.28	1.66			1200±60	24		0.06 ± 0.00	50±3 a	
ISR 7	E	0.7		202±10	150-200	0.62	0.24	1.18			1000±50	24		0.07 ± 0.01	68±5 a	
<u>Tzidkiyahu</u>																VLD
DF-557	A	7.2	227	92	150-177	0.62	0.3	1.0	1	477	797±30	12/13	16	1.09±0.08	1.4±0.1	

<u>Tzidkiyahu depression</u>																ID
DF-522 sa	A	3.4		279	139	125-150	0.58	0.6	1.6	2	508	928±32	17/20	43	14.6±1.9	15.8±2.2
DF-524	A	4.5		254	122	125-150	0.63	0.4	1.3	2	510	888±30	13/13	3	13.8±0.7	15.5±0.9
DF-660	A	10.2		200	69	125-150	0.46	0.4	1.0	1	387	657±28	17/17	9	10.5±1.0	15.9±1.7
<u>BM west</u>																VLD
DF-509	A	5.5		217	110	150-177	0.58	0.3	1.0	1	451	779±29	9/9	14	0.36±0.1	0.5±0.1
DF-511	A	10		217	70	125-150	0.58	0.3	1.0	1	456	745±29	11/13	44	0.66±0.09	0.9±0.1
DF-153	A	5.0		247	116	125-150	0.58	0.4	1.4	2	478	843±26	10/12	100	0.01±0.005	8±5 a Northern VLD flank
DF-110	A	4.0		213	129	125-150	0.52	0.3	1.2	1	420	763±27	9/12	27	0.35±0.04	0.5±0.1 Southern VLD flank
DF-111	A	4.7		203	120	125-150	0.50	0.3	1.1	1	404	728±26	9/12	27	1.30±0.16	1.8±0.2 Southern VLD flank
<u>BM east</u>																VLD, 250 m. downwind and beyond plunge of BM west
DF-513	A	4.5		242	122	125-150	0.55	0.5	1.2	2	466	832±34	11/13	28	0.65±0.07	0.8±0.1
DF-514	A	6.3		290	101	125-150	0.61	0.6	1.7	2	531	924±35	13/13	7	0.78±0.06	0.8±0.1
DF-515	A	7.5		226	89	150-177	0.51	0.4	1.3	1	423	740±32	12/13	15	1.25±0.17	1.7±0.2
DF-122	A	3.0		253	145	125-150	0.56	0.5	1.4	2	477	878±28	11/12	26	0.13±0.03	0.15±0.03 Northern VLD flank
DF-134	A	2.5		213	154	125-150	0.54	0.3	1.1	1	431	799±25	12/12	11	0.93±0.11	1.2±0.1 Southern VLD flank
<u>BM depression</u>																ID
DF-506	A	6.5		224	99	125-150	0.58	0.4	0.9	1	466	790±31	16/19	22	13.9±1.6	17.7±2.1
DF-507 sa	A	6.8		466	96	125-150	0.75	1.3	3.1	5	748	1315±35	17/24	30	39.0±4.5	29.7±3.5
<u>BM west playa</u>																ID
DF-100	E	0.25	640			88-125	0.75	2.3	4.8	10	929	1579±69	12/12	8	2.86±0.24	1.8±0.2
DF-101	E	0.65	474			125-150	0.58	0.3	0.9	1	454	929±52	12/12	5	2.02±0.11	2.2±0.2
<u>BM east playa</u>																ID
DF-487	A	7.85		179	86	150-177	0.43	0.4	0.7	1	355	622±29	18/19	43	9.1±1.2	14.7±2.0
Southern incursion path																
<u>Halamish West</u>																VLD base-ID contact
ISR 13	E	1.05			199±10	150-200	0.82	0.68	2.19			1300±78	24		4.49 ± 0.16	3.4±0.2
ISR 12	E	1.58			197±10	150-200	0.85	1.39	3.58			1600±80	24		14.90 ± 0.40	9.5±0.4

ISR 11	E	1.90		187±9	150-200	0.74	0.66	2.31			1200±60	24		11.0 ± 0.34	11.0±0.5	
<u>Halamish</u>																VLD
DF-618	E	1.2	181	181	150-177	0.42	0.40	0.8	1	351	714±25	16/19	20	16.7±2.4	23.4±3.4	Northern flank base
<u>Halamish ID</u>																ID in trench
DF-621 sa	E	0.35	452	224	125-150	0.91	1.0	2.9	4	795	1474±35	18/18	22	17.5±1.4	11.9±1.0	
DF-625 sa	E	1.3	371	178	125-150	0.67	0.9	2.6	3	616	1169±33	18/19	22	16.7±2.1	14.3±1.8	
DF-626	E	2.3	250	158	125-150	0.62	0.5	1.2	2	497	906±32	19/21	17	17.3±2.6	19.1±2.9	
<u>Halamish VLD</u>																Crest
ISR 10	E	0.28		209±10	150-200	0.64	0.59	2.32			1200±60	24		0.03 ± 0.00	26±2 a	
ISR 9	E	0.58		205±10	150-200	0.61	0.71	2.04			1100±55	24		0.03 ± 0.00	27±7a	
<u>Halamish East</u>																VLD
DF-632	A	9.4	281	74	125-150	0.61	0.60	1.5	2	526	884±30	17/19	18	11.6±1.5	13.2±1.8	
DF-633	A	9.5	503	74	125-150	0.83	1.4	3.6	5	798	1379±38	13/13	13	13.4±1.9	9.7±1.4	Fluvial loam dune base
<u>Nizzana floodplain</u>																Sand mantle on base of hill
DF-516	E	2.3	527		125-150	0.61	0.80	2.0	3	550	1080±58	13/13	10	15.1±1.7	14.0±1.7	
DF-518	E	3.7	551		125-150	0.61	0.60	2.2	3	530	1083±60	15/17	17	22.1±2.7	20.4±2.8	
<u>Mitvakh</u>																VLD
DF-200	E	9.25	217	75	125-150	0.58	0.3	1.0	1	456	750±29	18/24	36	10.7±1.8	14.3±2.4	
<u>Beer Malka</u>																TD
DF-1	E	3	275		125-150	0.21	0.6	0.8	1	235	511±35	17/19	22	9.1±1.6	17.8±3.3	
DF-3	E	3.5	674	137	88-125	1.0	2.1	4.4	9	1066	1886±31	11/12	15	62.7±7.7	33.2±4.1	
DF-4	E	3	315		150-177	0.42	0.2	0.7	1	324	639±38	18/19	18	7.8±1.3	12.2±2.1	
<u>Besor terrace</u>																Stream-truncated dune-sand terrace
DF-639	A	3.5	281	137	125-150	0.65	0.50	1.6	2	537	957±34	15/16	5	11.8±0.7	12.3±0.9	

3.4.3. Particle size distribution and mineralogy

Unimodal distribution and fine to very fine sands characterize most dunes and interdune aeolian sands (Figs. 3.3a & 3.3b). PSDs along a dune's vertical section are often quite uniform. Transverse dunes contain slightly coarser grain size modes than VLDs. Some dunes are characterized by higher contents of fine-grained sediment at their base (Fig. 3.3b), suggesting incorporation of the underlying palaeosol substrate by abrasion and erosion. At fifteen localities the dune substrates are brown calcic palaeosols with bimodal textures ranging between sandy to silty loams (Fig. 3.3a).

The interdune fill of aeolian sands, mainly in the central incursion corridor, usually has similar PSD to the overriding VLDs and transverse dunes. In the southwest, the interdune deposits are aeolian-fluvial loamy sands to standing-water silty loams (Fig. 3.3b).

Dune sands of the central incursion corridor contain the least amount of fines. In contrast, dunes of the northern incursion corridor are characterized by finer grain size modes. There is no indication of sand grain size fining towards the eastern edge of the dunefield as suggested by Hunt (1991) and Enzel et al. (2010).

Negev dune sands are quartz rich with smaller amounts of plagioclase. The variability in mineralogy correlates with the incursion corridor classification. The central incursion corridor is more quartz rich than the northern and southern corridors (Fig. 3.3c). Several samples, mainly from dunes along the southern and northern fringes of the dunefield contain measurable calcite that was probably incorporated into the section from neighboring floodplain and loess deposits.

3.4.4 OSL ages

3.4.4.1 Analytical OSL precision

All the OSL ages, with their field and laboratory attributes, are presented in Table 3.3 and Fig. 3.5. Several lines of evidence increase our confidence in the reliability of the OSL ages. All the OSL ages were derived using the same SAR protocol on quartz grains of similar grain-size fractions, and can therefore be compared between the laboratories. All samples display a strong initial OSL signal and rapid decay, indicating the dominance of the fast OSL component. Samples showed good preheat plateaus in the range of 180-280 °C, with negligible recuperation (1-3%). IRSL depletion ratios were within 1.0 ± 0.1 , indicating negligible feldspar contamination, and 90% of the recycling ratios are within 1.0 ± 0.1 , suggesting that the SAR protocol corrects appropriately for any sensitivity changes.

The relatively homogeneous nature of the dune sands resulted in similar and low dose rates, reflecting the higher quartz content. The dose rates of the ISR samples, measured at the Marburg Luminescence Laboratory, show similar dose rates to the samples measured at the GSI and a comparable range of ages (Table 3.3). The modern samples with ages of 150-8 years indicate that the aeolian transport and deposition conditions have the potential to efficiently bleach any remnant doses in the quartz grains.

De distributions were usually normal, however samples may have had a few tailing aliquots of higher and/or lower De values. This is mainly attributed to contamination by bioturbation and minute contribution of underlying older sand. For example sample DF 685 contained 1-2 mm loam pellets that comprised ~1% of the bulk dune base and could be identified visually. These pellets are assumed to have originated from underlying palaeosols that are considerably older.

To assess the systematic uncertainty, 24 aliquots (2 mm) of six samples were bleached and dosed several times. All aliquots were then given the same laboratory β dose (10-16 Gy) and the OSL signal measured and normalized using the conventional SAR protocol. A scatter ranging from 2% to 11% was found on the L_x/T_x values, with the greater scatter for the smaller (2 mm) aliquots. These values represent systematic uncertainties and indicate that samples with less than 10% scatter on the natural De were probably well bleached at the time of deposition, and can be dated with errors of <15% on the ages.

Over-dispersion (OD) values represent the scatter beyond the systematic uncertainties (after Galbraith et al., 1999), and the values calculated for the samples measured at the GSI are given in Table 3.3. Fifteen samples from mid-dune sections have the lowest OD, less than 10%. The majority of the samples (54) have OD values below 20%. Another 17 samples, many being dune bases overlaying calcic palaeosols, have OD values between 20-30%. High OD (>30%) values are found for modern age samples and for samples from the bases of the dunes that had incorporated some underlying older material.

For dune base samples with ODs exceeding 13%, distinct outliers, usually 1-3 aliquots, were removed from the average De calculation (Fig. 3.4), thus lowering the OD of the remaining aliquots to below 13%.

To summarize, most of the ages show low OD values and narrow De distribution. Disregarding dune crests, the errors of dune sand ages usually do not exceed $\pm 15\%$ and the ages are considered reliable. Overall, the ninety-seven OSL ages are, within errors, in stratigraphic order; the rare cases of age reversals can mostly be attributed to reworking of sediment without sufficient solar resetting.

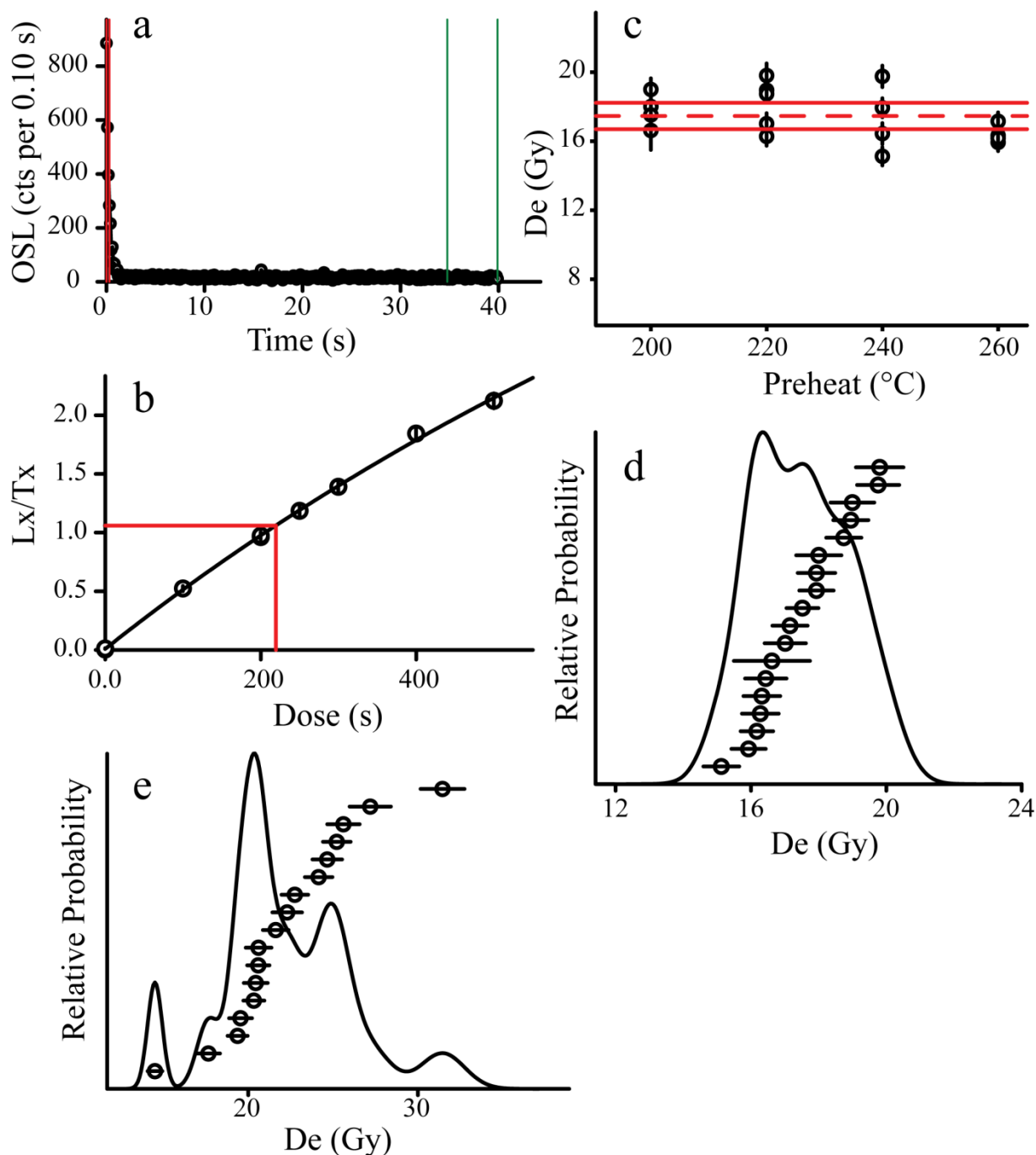


Figure 3.4

a. Natural OSL decay curve for a 2-mm aliquot of sample DF-200. Signal integration was for the first two channels, and background was subtracted from the last 10 channels.

b. Dose response curve for the same aliquot as in a. $D_e = \sim 10$ Gy.

c. Preheat plateau over the temperature range of 200-260°C for sample DF-621.

d. Relative probability plot for 18 aliquots of sample DF-621. Note normal distribution. $D_e = 17.5 \pm 1.4$ Gy.

e. Relative probability plot of sample DF-518 from a basal sand unit. The plot shows two outlying aliquots that caused over-dispersion of the sample to be 17%. Removal of these two aliquots reduced the over-dispersion to 11% and resulted in an age of 20.4 ± 2.4 ka.

Site	Source	Setting	Method & target-	Previous ages (ka)	This paper's OSL ages (ka)	Remarks
Halamish ID	Rendell et al., 1993; Harrison and Yair, 1998	Alternating silty loam and sandy loam units in interdune trench.	TL on KF	11.46±1.1 15.1±1.5	11.9±1.0 14.3±1.8	OSL sampling was in mid unit. OSL age of 19.1±2.9 ka at base of exposure.
Halamish Dune	Ben David, 2003	Cored base of VLD axis.	IRSL-SAAD (Ref) on KF	23.6±3.4	23.3±3.4	OSL age from same VLD 3 km down dune and taken from northern flank base.
Ramat Beqa quarry	Tsoar and Goodfriend, 1994	Hearths in sand quarry.	Charcoal, ¹⁴ C	4100 BC (6.1 B.P)	5.6±0.6 4.8±0.7	OSL sampled from mid section, estimated to be above ¹⁴ C site. Ceramics from Early Bronze I was also found (3,200 BC).
Sekher VI site	Barzilay, 2010	Two 3 m outcrops in shallow sands above northeastern bank of Nahal Sekher.	Joint layer of Early and Late Natufian microlith lunates. OSL ages sampled beneath layer.	Early Natufian 12.5-11.5 BP; Late Natufian 11.5-10.75 BP	12.3±1.2 and 11.9±1.4	This paper's samples were taken 10 cm below the Natufian artifacts. The surface was stabilized around 12 ka and underwent Early Natufian activity followed by later Late Natufian activity.
Sekher XXX site	Goring-Morris and Goldberg, 1990	Artificial 1 m exposure at brink of Sekher sands above southern bank of Nahal Sekher.	Small microlith lunates are presumed Harifian. OSL age was sampled beneath Harifian layer.	Harifian – 10.75-10.1 BP	11.5±1.3	Similar age to Sekher VI reaffirms surface sand stabilization around 12 ka.
Qeren Ridge	Enzel et al., 2010	Hearth over sand exposed in incised channel	Charcoal, ¹⁴ C	9.6-10.2 BP	11.4±1.3 (by Enzel)	OSL (performed at GSI) sampled at top of sand unit 50 cm beneath hearth.

Table 3.4 Comparison between the OSL ages obtained in this study for samples from the NW Negev and previous ages from the same area.

3.4.4.2 Comparison to previous dates and ages

In the Negev dunes where materials datable by radiocarbon or palaeosol indicators are absent, luminescence ages provide the only numerical chronology. In order to incorporate previously published ages into one chronological framework, these ages are compared to our OSL ages in places where similar units were dated (Table 3.4). The combination of ¹⁴C and OSL ages may also provide a better age control on features other than sand (Bubenzner et al., 2007). Chronological comparisons were made with interdune sediments and archaeological sites along the dunefield fringe, where charcoal and ostrich egg shells were dated by radiocarbon. Uncalibrated dates such as from Goldberg (1977) are now calibrated using Calib 6.0. Previous TL (Rendell et al., 1993; Harrison and Yair, 1998; Ben-David, 2003) and IRSL (Ben-David, 2003) ages sampled at identical or similar settings are also compared.

Table 3.4 presents side-by-side our OSL ages and previous luminescence and radiocarbon ages from similar units or locations. In most cases the ages agree very well, despite not being sampled at the same time or at the exact same location or stratigraphic section. This multiple

suite of concordant ages derived by different dating methods comprises positive evidence for the reliability and significance of the OSL ages and places the Northern Sinai and NW Negev Erg in one chronological framework, as well as harnessing ages from past luminescence protocols as elaborated by Chase (2009).

3.4.4.3 OSL dated landform types

OSL ages were obtained from a variety of landforms to improve our understanding of landscape evolution and to correlate the ages with previous studies that mainly targeted fluvial and interdune sections. Aside the VLDs, other landforms in the study area include: 1) mature palaeosol substrates beneath the dunes; 2) interdune (ID) fluvial-aeolian sediments; and 3) transverse dunes (Fig. 3.2). Section names are presented in Fig. 3.1c and their dated stratigraphic sections are presented in Fig. 3.5.

The oldest ages come from loamy palaeosols that underlie much of the dunefield in its western part (Fig. 3.5). In the north the OSL ages of this palaeosol range from ~116 to ~106 ka, broadly within the last interglacial period. Farther south a similar palaeosol was dated to ~30 ka (DF 507), suggesting aeolian accretion of fines and sand well after the last interglacial period. These data indicate that major dune-building in the Negev post-dates the last interglacial.

Interdune sediments in the southern and eastern dunefield that have lower sand fractions (Fig. 3.3a&b) gave ages between 33 ka and 2 ka (Fig. 3.5 sections 11, 20-25, 33c; Fig. 3.6f). The earliest age (DF 3; 33.2 ± 4.1 ka) comes from extensive floodplain deposits in the southern dunefield area, previously dated to 30-50 ka (Rendell et al., 1993; Harrison and Yair, 1998; Ben-David, 2003) (Fig. 3.7). The rest of the interdune sediments mainly display ages in the range of ~14 - 9 ka, similar to the dune section ages (Fig. 3.5). These interdune deposits are currently incised by streams. Ben-David (2003) inferred that these finer sediments were deposited in standing water due to extensive dune damming.

Two different types of transverse dune were dated. Transverse dunes, found mainly in the central incursion corridor, fill the interdune between VLDs (Fig. 3.2a). The lower sections of these transverse dunes at Tzidkiyahu show 25^0 - 33^0 eastern facing steep slip faces and date to 1.4-1.2 ka (Table 3.3). At Beer Malka, the western stoss base of an outstanding 1 km wide and 40 m high transverse dune on the eastern Nahal Nizzana floodplain gave ages of 17.8 ± 3.3 ka and 12.2 ± 2.1 ka (Figs. 3.5 & 3.7).

3.4.4.4 OSL age clustering

The main bulk of the ages is from within dune and interdune aeolian sections and range from 23 ka to recent. Three main age groups are easily discerned in the relative probability plot (Fig. 3.8): 18 ka to 11.5 ka (Late Pleistocene), 2 ka to 0.8 ka, (Late Holocene) from the upper dune sections and 150 to 8 years (modern) from dune mantles. Additional ages span from 23 ka to 18 ka, with a single age of 27.3 ± 3.8 ka. Several sand units, mainly in the east date to ~ 3 ka.

Twenty samples from the bases of dunes or interdunes throughout the dunefield overly loamy soils (Fig. 3.5) and date to 23-11 ka and were retrieved from the basal 0.5 m of the dunes. The dune-base age-span overlaps with the major Late Pleistocene incursion, substantiating this main and major encroachment event.

Deciphering dune activity solely upon age clustering analysis can be deceptive if not embodied in the dunes structure and stratigraphy and its dynamic implications. In the Kalahari, where exposed dune sections are unavailable, Stone and Thomas (2008) showed that dune age clustering is often a function of dune core sampling density. The consistent vertical internal sedimentary structure of the Negev VLDs (Fig. 3.2c) affirms the chronological clustering of the dunefield OSL ages, which overcomes the somewhat blind and inconsistent sampled method of dune drilling. Age-depth profiles (after Fig. 3.5) though, show that sediment ages cannot be directly correlated to dune depth, mainly due to the regional and local variation in dune height, implying differing spatial and possibly temporal rates of sediment supply as found in Australia (Cohen et al., 2010).

3.5 Discussion

3.5.1 Aeolian sand incursion episodes

3.5.1.1 Earliest evidence for aeolian sand deposition

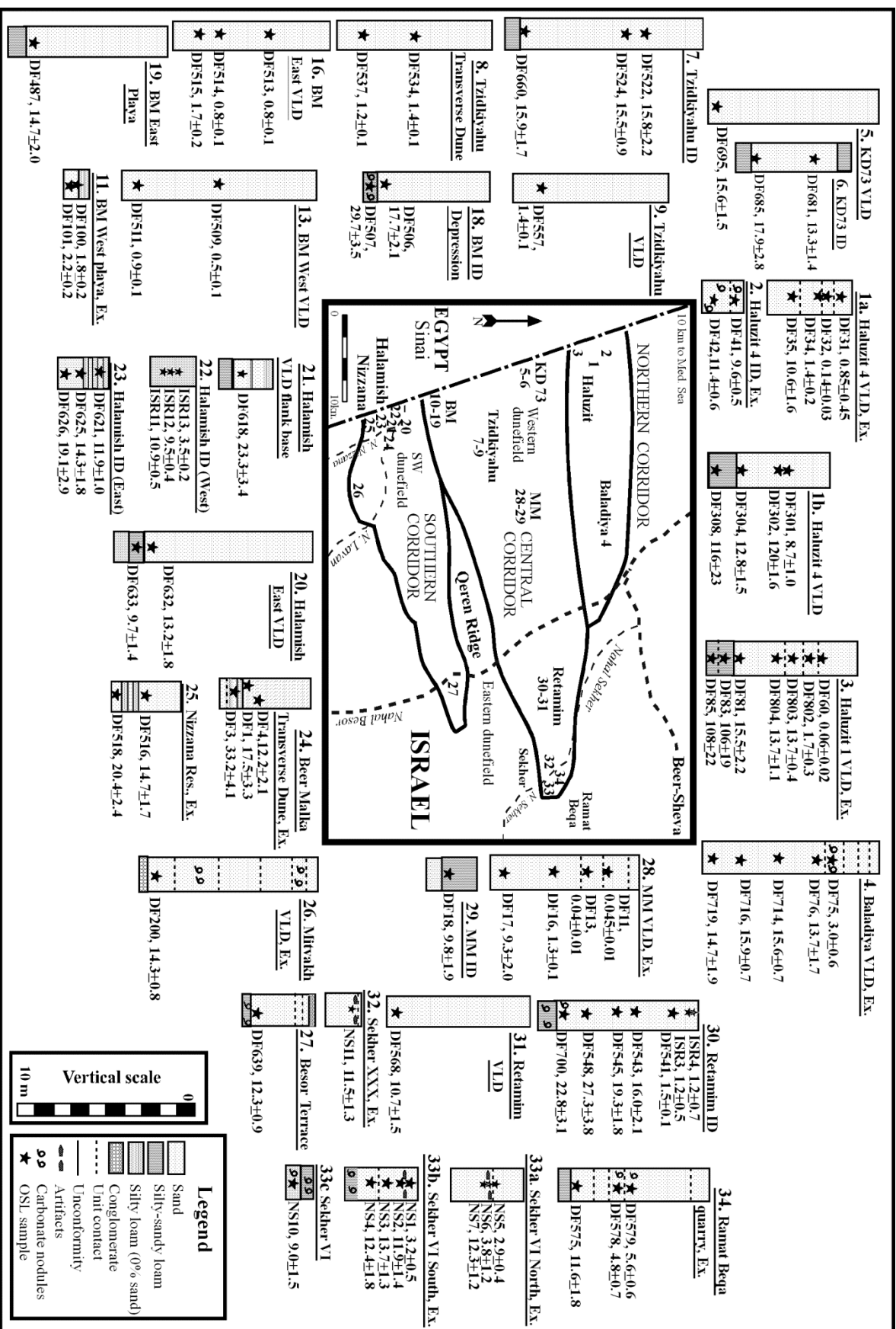
The palaeosols underlying the dunes contain aeolian sand (Fig. 3.3), but their loamy soil characteristics indicate different climate regime prior to dune encroachment and buildup. Four samples from the upper 1 m of the calcic palaeosol substrates date to ~ 116 -30 ka. They exhibit bimodal grain size distribution (Fig. 3.3b). Some of the palaeosols have low sand contents, similar to the upper loess unit ("L1") defined by Crouvi et al. (2008; 2009) and dated to 50-10 ka at three locations in the dunefield periphery. The sandy parent material of the palaeosols dated to 116-106 ka (Figs. 3.2c & 3.5) suggests some aeolian input prior to the last interglacial period. Sands may have encroached slowly on the region as thin sand sheets, later to be stabilized. Soil formation processes were, contemporaneous with an input of

aeolian fines (loess) that currently surround the study area to the east, south and north (Crouvi et al., 2009; Zilberman et al. 2007). Sand deposition would likely require a vegetation cover (Pye and Tsoar, 2009) and a semi-arid climate in contrast to the current arid climate, supporting palaeoclimatic interpretations of a moister Late Pleistocene (Bar-Matthews et al., 1999, 2003; Vaks et al., 2006). Vegetation impedes sand transport and is suggested to inhibit dune buildup, which results in sand sheet development (Kocurek and Nielson, 1986).

At Qerem Shalom, north of the study area, a 3-m-thick aeolian sandy loam deposit dated to 40-90 ka is attributed to a coastal source (Zilberman et al., 2007). Limited sand transport may also have been controlled by limited available sediment in the west. The Nile Delta section contains thick coarse quartz sand accumulations overlain by stiff clays dated no earlier than ~38 ka (Coutellier and Stanley, 1987; Stanley et al., 1996), suggesting limited availability of exposed interglacial sand. In any case, sand supply was also initially controlled by the erosivity and exposure of the Nile Delta sand source complexes despite the frequency and strength of sand transporting winds in Northern Sinai possibly being higher than today's (Enzel et al., 2008).

Additional evidence for incipient aeolian activity in the NW Negev without dune buildup is the presence of quartzose sand within fluvial sections draining the southern incursion corridor (Fig 1c; Table 3.1): An IRSL age of 98 ± 12 ka was obtained for a 1 m-thick fine sand unit at the Nahal Besor-Revivim confluence terrace (Greenbaum and Ben-David, 2001); a calcic gravelly sandy basal palaeosol in a Nahal Lavan fluvial section was dated to 67 ± 6 ka by TL; and several samples of fluvial sand matrix between cobbles in Nahal Nizzana were dated by IRSL to 115-76 ka (Ben-David, 2003) (Fig. 3.1c; Table 3.1). Zilberman (1993) also identifies sand within the Late Pleistocene flood plains, mainly in the southern dunefield, and concluded that sand has been in the system since ~100 ka.

The BM interdune calcic loamy palaeosol (DF-507; 29.7 ± 3.5 ka), overlain by aeolian sand (DF 506; 17.7 ± 2.1 ka) (Figs. 3.5 & 3.9) is the youngest palaeosol dated beneath the aeolian sands. This sample marks the shortest hiatus identified between a palaeosol and clean unconsolidated aeolian sand, indicating that while at ~116-30 ka sandy-silty loam pedogenesis occurred, substantial aeolian activity probably began only post ~30 ka. This is clear evidence that dune sands did not reach the western dunefield beforehand. Though being the only palaeosol in the dunefield dated to ~30 ka, its calcium carbonate content resembles the Stage II-III Early Upper Paleolithic palaeosols 5 km to the south at the Nahal Lavan-Nizzana fluvial confluence (Zilberman, 1993). It also resembles a calcic unit beneath the prehistoric site Azariq XIII along Nahal Lavan (Goldberg, 1986; Goring and Morris and



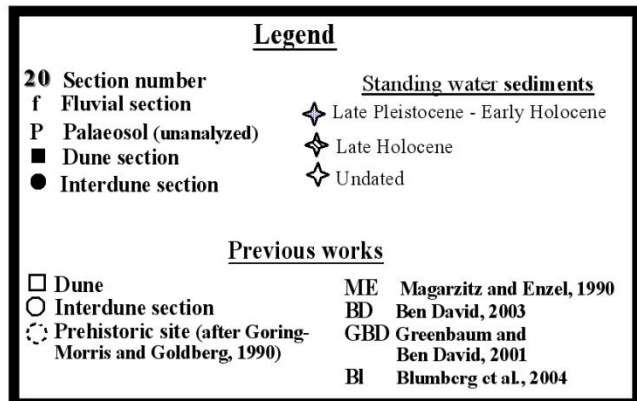
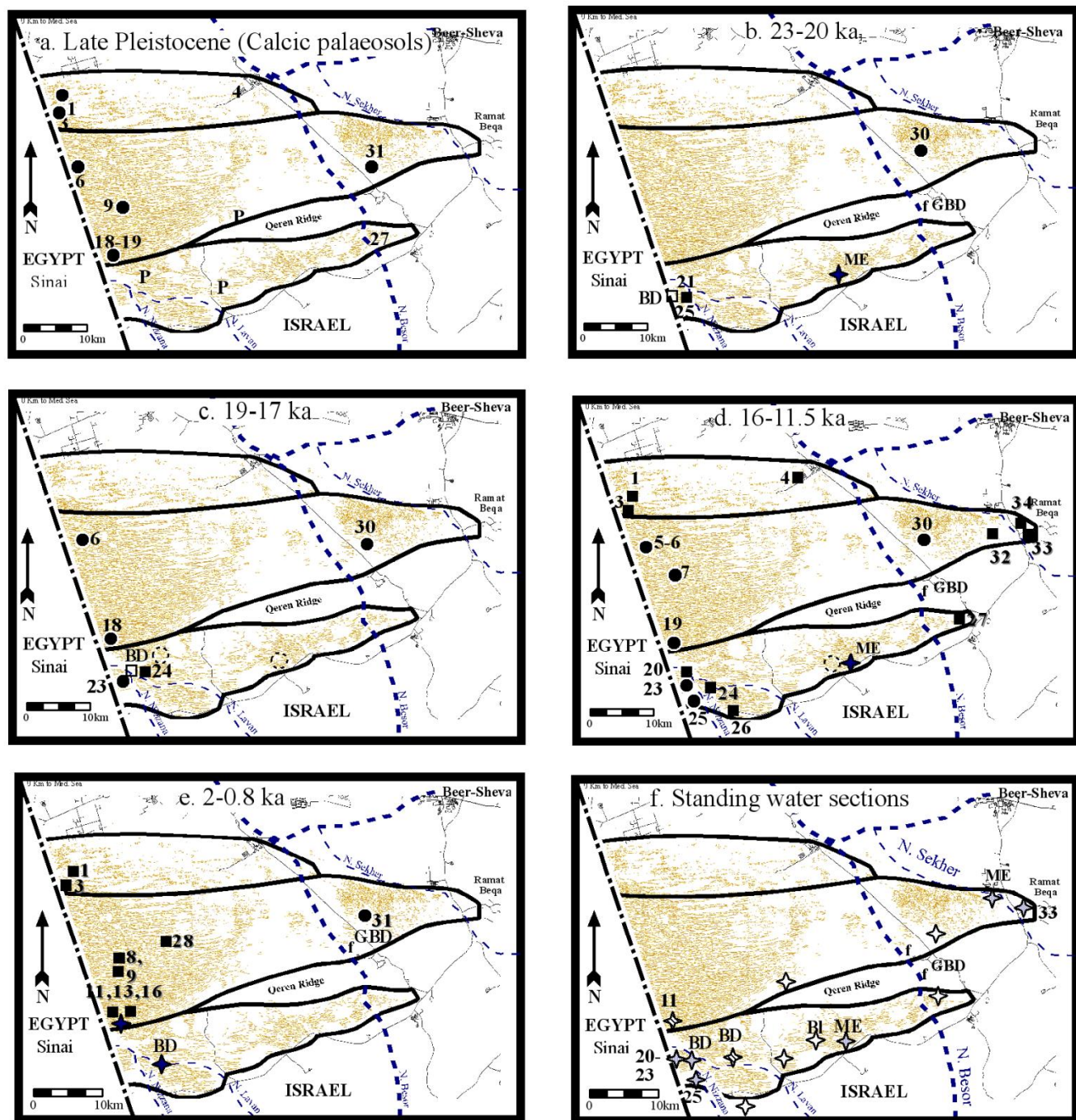


Figure 3.6

a-e: Time slice maps of stages in the evolution of the NW Negev dunefield, derived from the OSL ages. Dune section numbers are as on Fig. 5.

f.: Sections of standing water deposits. OSL ages of these deposits along with previous dating associates them with dune-damming, mainly during the main incursion episode.

were dated by radiocarbon to ~27-24 and -14-12 ka (Magaritz and Enzel, 1990; Zilberman, 1993). These ages suggest that roughly during episodes of dune encroachment, calcic horizons in local settings unaffected by sand may have developed, also implying that the dune-building climate was not arid.

In the east of the central incursion corridor, the base of the Retamim ID section preserves the oldest unconsolidated aeolian sand unit that overlies a calcic palaeosol, dated to 27.3 ± 3.8 ka with PSD similar to slightly younger dune sand (Figs. 3.5 & 3.6). This sand, which is only slightly younger than the youngest palaeosol, marks the onset of the aeolian phase in the northwest Negev that soon matured into the initial dune incursion. The Retamim ID base resembles the section at Halamish (Fig. 3.7), where Ben-David (2003) suggested initial sand accumulation at 25-27 ka. It also strengthens Zilberman's (1991) synthesis of the southern dunefield that suggested initial sand incursion followed by dune incursion, that, based upon Epipaleolithic artifacts, began evolving at 25-30 ka and overrode the fluvial stratigraphic sequences.

The Retamim ID section also preserves evidence of the thickness of the initial aeolian sand cover on the palaeosols. This sand contains small quantities of calcic nodules and stains that suggest that it formed as periodic sand sheets or deposits and not as dunes. The ~27-19 ka ages are not found at basal sections along the western transect of the central and northern incursion corridors (Fig. 3.9). This may mean that the western transect bases have been fully reworked and the OSL signal of the sand grains was fully reset. The Retamim basal ages could represent sands that were probably reworked in the west, possibly due to local physiography of local depressions or pockets (after Stone and Thomas, 2008), were locally preserved. The ~16 ka age of Retamim ID section attests that later incursions were recorded in the section and it is intriguing how much of the sand dated between ~27-19 ka was subsequently eroded.

The consistent lack of preservation of old basal units throughout the dunefield, considering the sampling resolution of dune bases, suggests that the thickness of the initial sand was thin. Furthermore, the thinning and final disappearance of the sand and dune cover east of Retamim does not support the notion that a thick sand unit was deposited there at ~27-23 ka or even earlier, and later transported and accumulated further east during later dune incursions. This comprises additional support that dunes did not cover the NW Negev before ~23 ka.

Archaeological evidence also supports the transition from a thin sand and loess - loam surface to dune encroachment. Archaeological remains at the Shunera XV –XVI sites by

Nahal Mobra overly fluvial loess and are embedded in a deflated sandy mixture (Fig. 3.1c; Table 3.1). The sites date to the Early Epipalaeolithic (~19 ka) and indicate that the surface was thinly covered with sand (Goring-Morris and Goldberg, 1990; Goring-Morris, 1998) that probably stabilized around 19 ka as found further northeast at the Retamim ID section. Southwest of the dunefield at Wadi Gayifa, west of the Egypt-Israel border, carbonate nodules from a palaeosol were dated by Th-U to 28 ± 4.6 ka, perhaps equivalent to the BM site palaeosol dated to 29.7 ± 3.5 ka (Fig. 3.5). These are overlain by thin sand (Gladfelter, 2000) (Table 3.1)

To summarize, episodic thin sand covered the region already by ~100 ka. The sand was incorporated and stabilized into calcic loamy palaeosol units until ~30 ka, though in areas preserved from erosion by advancing dune palaeosols may have continued to develop. The late Pleistocene palaeosols were eroded differentially by the dune incursion. There is evidence of aeolian sand accumulation as sheets in the central incursion corridor and southern incursion corridor that occurred between ~27-19 ka and was soon followed by dune encroachment.

3.5.1.2 Initial dune incursion

There are limited and spatially sporadic luminescence ages for the initial episode of dune buildup. The oldest (>20 ka) evidence for dune presence was found in the southwestern part of the southern incursion corridor at Halamish, limited to the west bank of Nahal Nizzana near the Egyptian Israeli-border (Figs. 3.5 & 3.7). A single dune-base IRSL age of 23.5 ± 1.4 ka is presented by Ben-David (2003) (Fig. 3.7). The exposed base of the northern dune flank, 3 km down the same dune to the east, OSL-dates to 23.3 ± 3.4 ka (DF 618; Fig. 3.9). A clayey-silt to sand interdune section nearby (Halamish West) was dated by IRSL to 24.5 ± 1.4 ka (Ben-David 2003, section KR1), probably pin-pointing the same sand burial event along with slack water sedimentation (Fig. 3.7).

Along this dune elongation corridor, on the eastern Nahal Nizzana floodplain, an eroded dune exposed in a wadi terrace and assumed to be part of a dune dam, was dated by TL to 18.4 ± 1.6 ka (Ben-David, 2003). On this same floodplain the next oldest dune age of 17.8 ± 3.3 ka is found at the stoss slope base of the Beer Malka transverse dune (Fig. 3.7). South of the Halamish dunes at the current dunefield fringe, sand covers a broad chalk hill on the western bank of Nahal Nizzana. The exposed colluvial hill base revealed aeolian sandy loam units interchanging with silty loam layers similar to those found in the Halamish interdunes (Fig. 3.5). The aeolian basal unit dates to 20.4 ± 2.4 ka (Fig. 3.5). As there is limited sand upstream,

this dated unit delimits the southern fringe of the initial incursion corridor which has not been changed since.

The limited spatial extent and quantity of corresponding archaeological finds support the age and limited spatial extent of the initial dune incursion. A single Late Upper Palaeolithic site (~20 ka) at Azariq XIII, 2 km east of the Nahal Lavan-Nahal Nizzana confluence, crop out in an eroded base of a VLD and overlies a weathered calcic palaeosol (Goldberg, 1986; Goring-Morris and Goldberg, 1990) (Fig. 3.1c). The stratigraphic relations point that a thin sand cover was present during artifact deposition, though its difficult to ascertain if the site was situated upon a sand sheet prior to dune buildup or upon a dune nose or flank toe that later slightly migrated eastwards. The rest of the archaeological sites in the southern dunefield post-date the initial dune incursion.

The examples cited above suggest that initial dune formation in the Negev began around 23-20 ka in the southwest corner of the current dunefield (Figs. 3.5 & 3.6b). The lack of dunes that are dated by OSL to this age in other parts of the dunefield could possibly be due to later erosion or remobilization resulting in resetting of sediment luminescence. The southwest dunefield is also where fluvial fine deposits buried sands and linear dune sand flanks, and protected the basal initial dune sand from later reworking.

3.5.1.3 The main dune incursion

The main dune incursion, which defined the current spatial extent of the dunefield and transported the main bulk of sand, took place around 18-11.5 ka (Figs. 3.5, 3. 6, 3.8 & 3.9). Evidence for this incursion is found throughout the dunefield, mainly at dune bases up to their mid-sections, in interdune sediments and sands (Figs. 3.5 & 3.6). Several dune bases, mainly in the west (Fig. 3.9), exhibit slightly earlier ages of ~19-17 ka (Fig. 3.6c), which may indicate an incipient stage of the main incursion that was concentrated at ~16-11.5 ka. Sporadic younger ages between ~11-9 ka are not found in dune bases but rather in the dunes mid-sections, and therefore do not represent the dune incursion but later reworking. Fluvial sand dated by TL and IRSI to 15.4-10 ka, with three older ages between 18-23 ka (Ben-David, 2003), support our finds.

The evidence for the main incursion along the western and eastern dunefield transects is presented below, followed by analysis for each incursion corridor. Along the western transect, dune base ages range from 14 to 18 ka (Figs. 3.2c, 3.5 & 3.9). In the southwest (Halamish), where preserved basal dune ages are >20 ka, a VLD axis base dates to 13.2 ± 1.8 ka (Halamish East) (Fig. 3.10). This younger age possibly represents elongation or reworking of the older

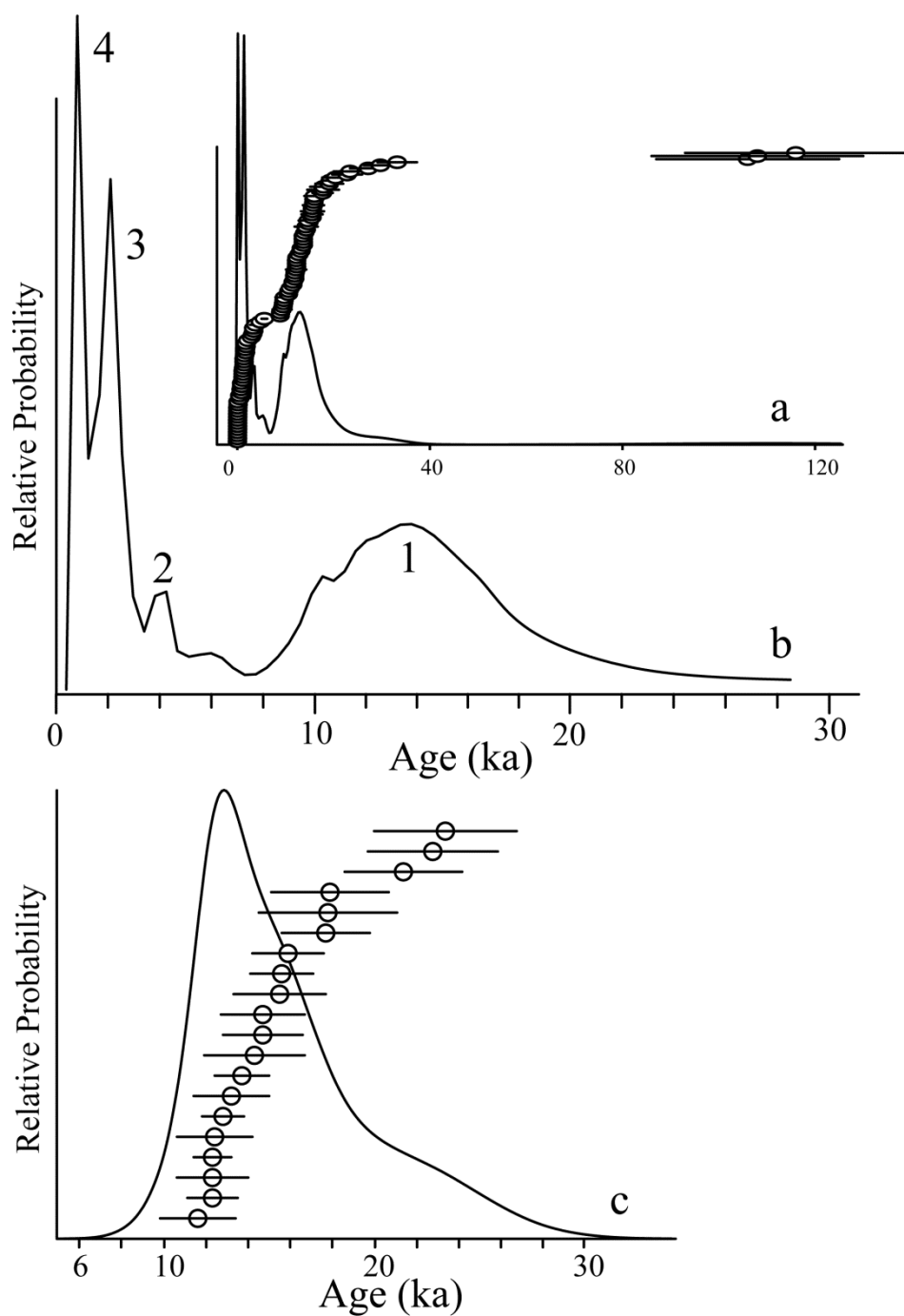


Figure 3.8

Relative probability plots of OSL ages:

a. The entire data set of 97 OSL ages from the NW Negev dune, sand and dune substrate.

b. Samples younger than 30 ka. The graph indicates age clustering at: (1) 18-11.5 ka, (2) 3 ka. (3) 2-0.8 ka. (4) 150-8 years.

c. Ages obtained from the base of dune and aeolian sand sections. Note how they fit into cluster (1) of graph b.

dunes which is also supported by the nearby Beer Malka transverse dune ages between 12.2 ± 2.1 and 17.8 ± 3.3 ka (Figs. 3.5 & 3.7).

The northeastern edge of the dunefield shows ages similar to those found in the western transect. The northern incursion corridor eastern edge at Baladiya displays exceptionally broad (200-400 m) and high (10-15 m) dunes (Fig. 3.5) with a 7 m thick lower section, dated to 15.9 ± 0.7 - 13.7 ± 1.7 ka (Fig. 3.5). The Late Pleistocene section is overlain by a 0.6 m sand unit lightly cemented by carbonate (DF 75; 3.0 ± 0.6 ka). The section signifies rapid elongation and buildup roughly over ~1,000-2,000 years, followed by extensive stability. In the easternmost extent of the central corridor, the basal sands at Ramat Beqa and Nahal Sekher display slightly younger ages than at Baladiya in the northeast (Fig. 3.5). In the southeast, the base of a stream terrace composed largely of dune sand is dated to 12.3 ± 0.9 ka. The formation of this terrace predates the early Holocene incision of Nahal Besor (Greenbaum and Ben-David, 2001; Ben-David, 2003). The limited thickness and spatial cover of sand east of this site make it unlikely that sand arrived to this section before ~12 ka (the basal age), as if so, later aeolian activity would have transported sand further east.

These results indicate that while dunes initially developed in the southwest, dunes accumulated in the northeast only at 14-15 ka. Slightly later at ~12.5-11.5 ka, dunes reached their easternmost extent in the central and south incursion corridors and stabilized (Fig. 3.6d).

Though the dunefield lacks continuous VLDs that elongate continuously as a single defined dune for many kilometers, OSL ages at the end points of the 30-50 km long incursion corridors indicate that west-east sand transport in this period was rapid. The basal and the mid-section ages of Haluzit 1 (Figs. 3.2c & 3.5, 3.9) in the west (15.5 ± 2.2 ka; 13.7 ± 0.9 ka, respectively) and the basal and the mid section ages of Baladiya (Fig. 3.5) in the east (14.7 ± 1.9 ka; 13.7 ± 1.7 ka, respectively) present essentially identical ages. This indicates rapid encroachment and settlement of sand across the northern incursion corridor within ~1,000 years. After ~14 ka, wind intensities probably subsided resulting in less sand input and dune growth. Nevertheless, based on the broad and 12 m high Baladiya section and the limited sand found further east, wind did not substantially erode the dune axes that accumulated during this period. The current presence of abundant vegetation throughout the northern part of the dunefield (Siegal, 2009), which may have also been present in the past, decreased sand input and may explain the low broad VLD morphologies of the northwestern section of the dunefield that seems to not have substantially changed since stabilization at ~14 ka.

Ages mainly around ~12 ka indicate the end of the main incursion and dune buildup with the NW Negev dunefield attaining its maximum and current spatial configuration (Figs. 3.5 &

3.6d). The Sekher VI OSL age of 13.7 ± 1.3 ka (Fig. 3.5) shows that at the end of the main incursion phase some sand reached the eastern edge of the present dunefield. This coincides with an OSL age of 13.1 ± 1.5 ka for basal dune sand on the Qeren ridge (Table 3.4) (Enzel et al., 2010). North of the research area at Qerem Shalom the upper sand sheet base (1.5 m depth) was dated to 14.5 ± 2.3 ka and 13.4 ± 1.7 ka (Zilberman et al., 2007). The three Sekher sites dated by OSL contain a Natufian assemblage ~ 12.5 - 10.75 ka (Goring-Morris et al., 1998; Barzilay et al., 2009; Barzilay, 2010)]. A surface with mixed prehistoric artifacts was dated to ~ 3 ka and interpreted to be a lag deposit that was exposed from ~ 12 ka to ~ 3 ka (Fig. 3.5; Tables 3.3 & 3.4).

Abundant sand accumulated in the western section of the present dunefield during this main phase of dune-building at 16-11.5 ka. In the central incursion corridor, at the Tzidkiyahu site, cumulative sand thickness reaches approximately 30 meters (Figs. 3.2a, b & 3.5), providing evidence for a rapid dune accretion episode of approximately up to 10 m/1000 yr, as inferred previously from the northern incursion corridor sections (Fig. 3.5). The basal 6 meters at Tzidkiyahu shows nearly identical ages of 15.9 ± 1.7 ka 15.5 ± 0.9 ka and 15.8 ± 2.2 ka (Figs. 3.2a, 3.5 & 3.9). The consistent PSDs of the section compliment the similar ages and indicate a uniform sedimentary aeolian environment. Ten km to the north, at KD 73, PSD and dune base ages (DF 695; 15.6 ± 1.5 ka) are similar to Tzidkiyahu (Fig. 3.5) suggesting that the whole basal section of the western part of the central incursion corridor accreted in a similar event.

The Negev rapid dune movement and high accumulation rates indicate rapid accretion and stabilization of a short but extreme episode. The rates are approximately an order of magnitude greater than those of the Egyptian Sand Sea (Libyan Desert, west of Cairo), with calculated net-accumulation rates of 30-100 cm/1000 years (Bubenzer et al., 2007). Average net sedimentation rates of 10 cm/ka for vegetated linear dunes in South Australia are also substantially lower (Lomax et al. 2011).

3.5.1.4 Dune damming in the southern incursion corridor

The southern incursion corridor contains additional evidence of dune migration that dammed wadi courses (Magaritz and Enzel, 1990; Ben-David, 2003). Dune-dammed paludal sediments are found throughout the southern dunefield in exposed sections commonly overlying basal dune flanks. Standing-water palaeolakes expanded into the interdunes (i.e. the Halamish sections) (Figs. 3.5 & 3.7) past the current Egypt-Israel border and upstream, reaching the Nizzana road (Nizzana reservoir section; Fig 1c). Ten interdune sections with

stratigraphy of interchanging aeolian, fluvial and standing water deposits (Fig. 3.5) were studied and dated from underlying and overlying sandy sediments, as the paludal fine-grained sediments themselves usually lack datable quartz sand (Figs. 3.3a & 3.3b).

Interdune basal paludal sediment overlies dunes dated to ~23 ka and 18 ka (Ben-David, 2003) (Figs. 3.5 & 3.7) and post-dates these ages. The upper-parts of five interdune sections were dated to 8-10 ka, younger than the main incursion stabilization age (Fig. 3.6f), suggesting accumulation in response to the main incursion period due to intensive dune damming. The earliest archaeological remains in the Halamish interdune surface are from the PPNB (Pre-Pottery Neolithic B period; 9.4-7.6 ka) (Goring-Morris, pers. comm.) and are slightly younger than the upper Halamish ID unit. Following the cessation of dune elongation, water-lain sediments continued to accumulate behind the dune dams. The unique magnitude of the Beer Malka transverse dune, with ages of 17.8±3.3 ka and 12.2±2.1 ka (Figs. 3.5 & 3.7), suggests an episode with strong unidirectional winds and substantial sand supply. The transverse dune seems to have advanced considerably since 12.2±2.1 ka, though this may be due to a surplus of available sand due to possible dune-damming of Nahal Nizzana.

Other than Nahal Besor, all the wadis crossing the southern incursion corridor were periodically blocked by dunes (Ben-David, 2003). Wadi Al-Arish, currently the only water course that crosses the Northern Sinai dunefield, may have also been periodically blocked. The wide floodplain silty and bright deposits (Sneh, 1983; Kusky and El-Baz, 2000; Ben-David, 2003), easily identified along the wadi by space imagery between Gebel Hallal and Al-Arish on the Mediterranean coast, are suggested to be an evidence of palaeolakes (Kusky and El-Baz, 2000). They differ from the underlying fluvial muds, sands and gravels that are exposed in the wadi's section as described by Sneh (1983) and appear similar to the dune dammed deposits in the Negev.

Thus, dune damming exemplifies the extent of environmental impact of a massive dune incursion. We suggest that the main dune incursion transported large sand volumes across Wadi Al-Arish and blocked it. Mid-sized (10^2 - 10^3 km²) NW Negev catchments breached and destroyed the dune dams in the Early Holocene (Fig. 3.6e) leaving residual standing water deposits (Harrison and Yair, 1998; Ben-David, 2003). In contrast, smaller drainage basins are still covered by dunes (Blumberg et al., 2004). This aeolian-fluvial history may explain the occurrence of Mid-Epipaleolithic [(~15-12.5 ka; Goring-Morris et al. (1998)] to Harifian [(~10.75-10.1 ka; Goring-Morris et al. (1998)] artifacts and camps in this part of the Negev (Goldberg, 1986; Goring-Morris and Goldberg, 1990; Barzilay et al., 2009). While it has been pointed out that those archaeological sites have a limited spatial extent in the region (Goring-

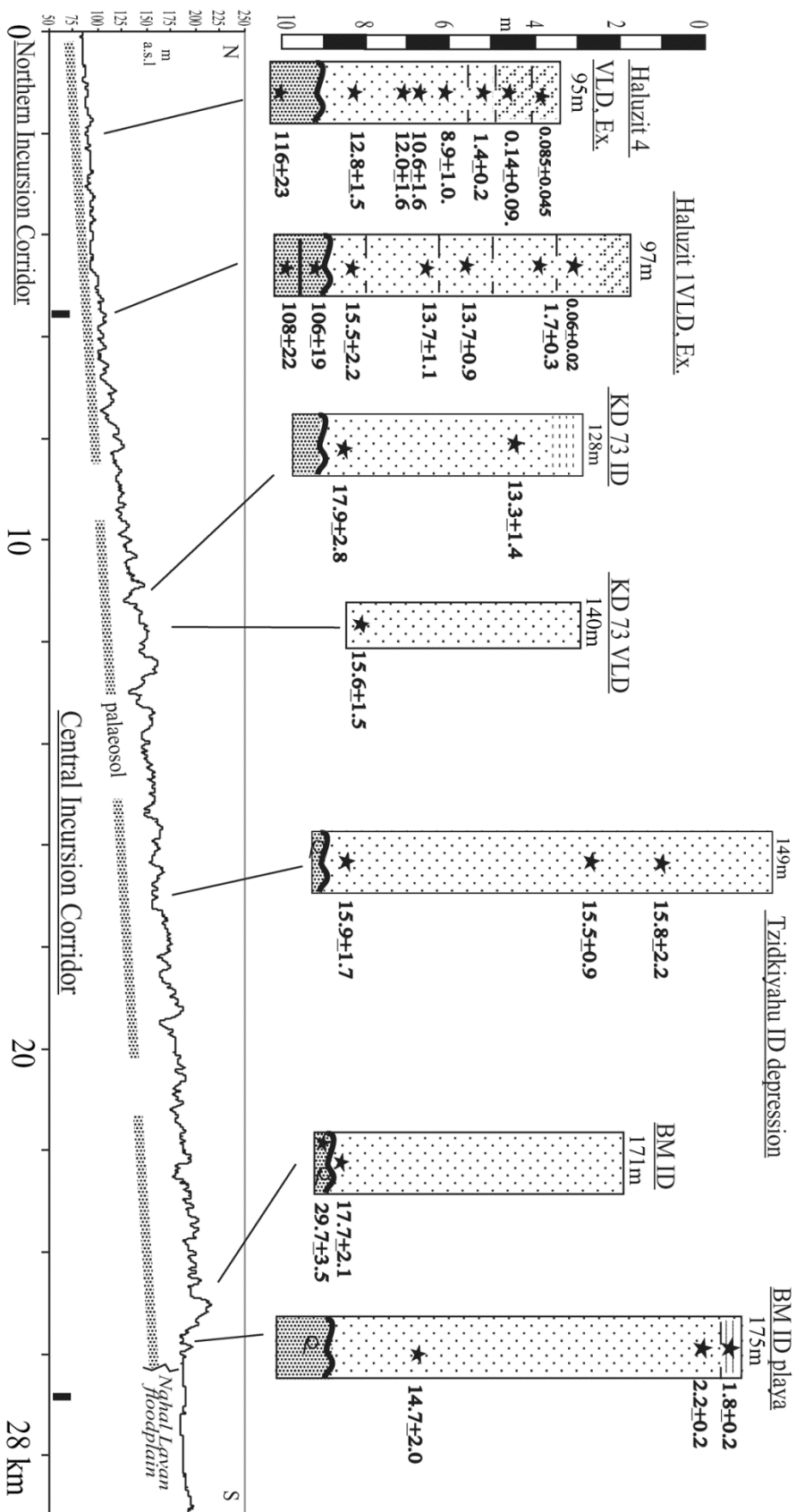


Figure 3.9 A compilation of all stratigraphic logs from the western transect of the northern and central incursion corridor shown on the same scale (see location on Fig. 1c). The topographic cross section (from 10m/pixel DTM) at the bottom of the figure exemplifies the broad and low dunes in the northern incursion corridor vs. the central corridor VLD and transverse dunes. The logs are aligned along the top of the palaeosol and not by elevation. Ages are in ka.

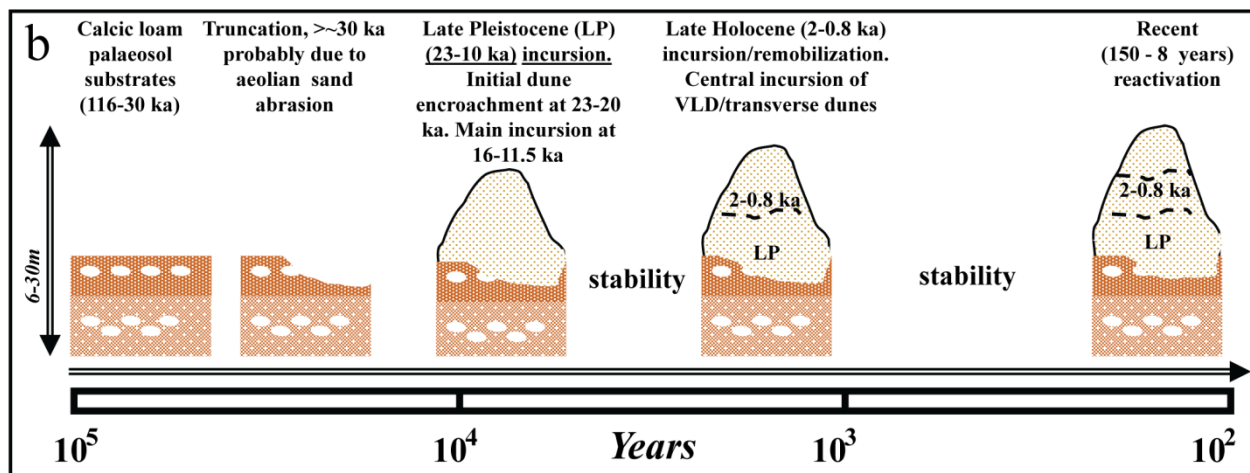
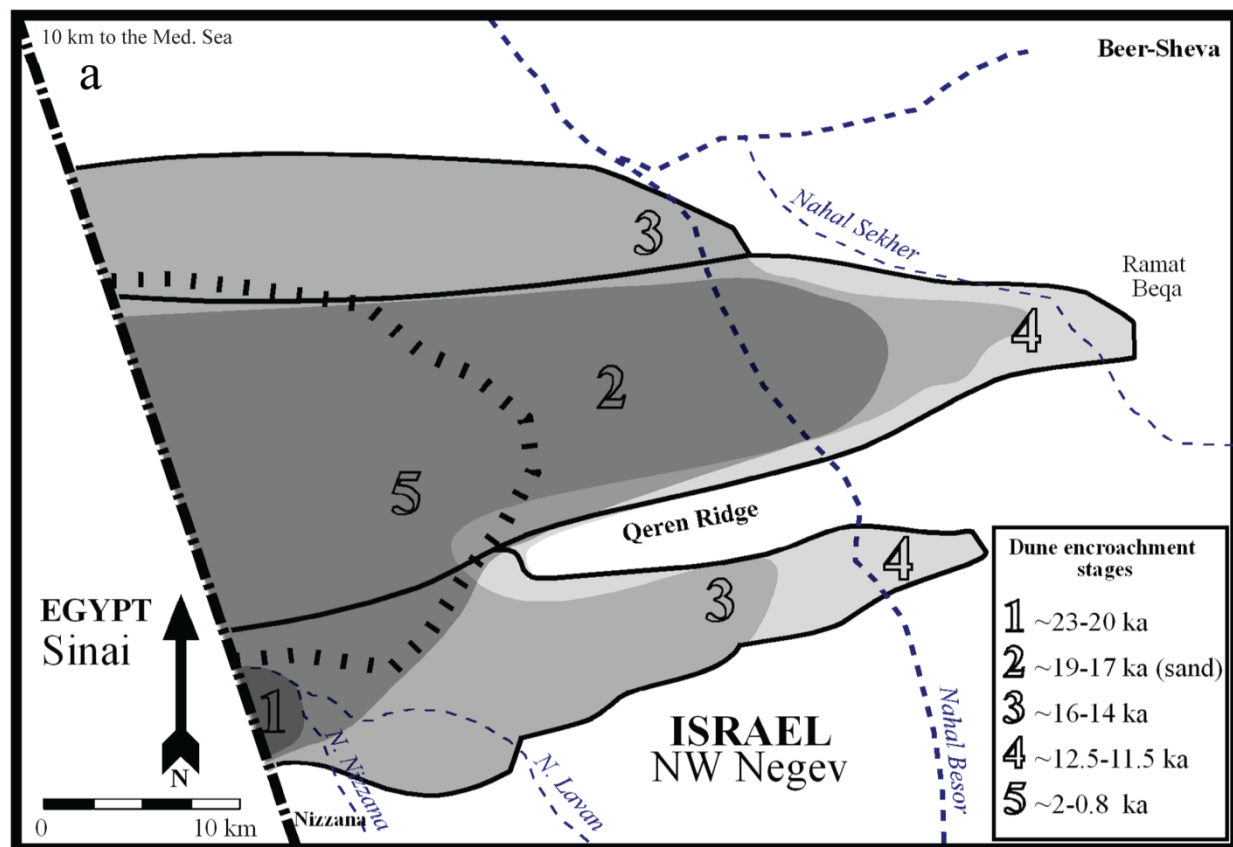


Figure 3.10 A summary of the evolution of the NW Negev dunefield.

a. A map showing the geographic extent of the different incursion phases. The OSL ages from this study were combined with previous ages (Table 1).

b. A cartoon showing the main sedimentological stages found throughout the study area.

Morris and Goldberg 1990), palaeolakes and ponds created by dune dams would have been favorable sites for at least short-term human settlement.

The color of interdune sand between paludal silts, based on a spectral redness index, is similar to the adjacent dune sand color (Roskin et al., 2010), indicating the short-term extent of these palaeolakes and ponds. If standing water remained for a long time and the water has contact with the sand underlying the silts, it is likely that bleaching of sand grain color would have occurred, though probably at a different rate than what was identified in active transverse dunes between lagoons in Brazil (Levin et al., 2007).

3.5.1.5 Summary of the Late Pleistocene aeolian episodes

Field and geochronological evidence indicates that aeolian sands have existed in the Negev Desert at least since 100 ka, the last interglacial (Fig. 3.10), however there is no evidence for dune remnants in the NW Negev earlier than ~23 ka. Our work shows that by ~23 ka there is initial dune buildup evidence only in the southwest, while by 19-17 ka, sand, but not necessarily dunes, advanced through the central incursion corridor (Fig. 3.10). From ~16 ka, several major and rapid incursion events occurred, marking the main incursion phase. Sand advanced in the north to its easternmost extent and thick, uniform-grained sand deposits accumulated in the western parts of the central and southern incursion corridors. The dunes dammed drainage systems, resulting in ponds or palaeolakes spreading upstream and laterally into the interdune areas, where fine-grained sediments were deposited. The main incursion is older than suggested by Enzel et al. (2010) and is not associated with the Younger Dryas cooler period of ~13-12 ka. In fact, the main dune incursion period rather ends with the Younger Dryas. By 13-11.5 ka though, in the south and center of the dunefield, the dunes and sand were remobilized and reached their easternmost extent. At some locations in the dunefield, small-scale sand movement slightly truncated the 15-11.5 ka surface and finally stabilized by ~10 ka. By 10-9 ka, many dune dams were breached due to stream incision.

3.5.1.6 Late Holocene dune activity

While dune ages from the middle and early Holocene are limited, a significant cluster of late Holocene ages (~2-0.8 ka) was found for the upper dune sections (Figs. 3.5 & 3.6e). This episode is preceded by a sporadic, ~3 ka event, identified only in the east, perhaps due to better dune preservability. This event may also be represented by dune damming between ~3.5-1.2 ka in Nahal Lavan (Ben-David 2003) in the south (Fig. 3.6f).

Though the late Holocene aeolian episode between 2 ka and 0.8 ka was extensive, it has not been previously identified in the Negev, although Hunt (1991) and Goldberg (1986) identified significant loessy silt accumulation at this period. The Tzidkiyahu VLDs (Fig. 3.2a) that date at 8 meters deep to ~1.4 ka show that the late Holocene episode, with substantial dune buildup, was as significant as the Late Pleistocene in the central incursion corridor. This incursion contributed new aeolian sand and not only remobilized Late Pleistocene sands (Fig. 3.5). At Tzidkiyahu, VLD and interdune transverse dune sands with identical ages (1.4 ka) cover the main incursion unit. The 8-12-m-thick late Holocene dunes do not contain any evidence of pedogenesis, indicating that this episode was probably rapid with strong and mainly uni-directional sand-transporting winds. Although by this time humans had long occupied the Northern Negev, the dune thickness argues against reactivation due to anthropogenic effect of decimation of the stabilizing biogenic crust and vegetation cover. The coeval formation of vegetated-linear and transverse dune types may be due to strong west-east winds that elongate VLD's (Tsoar et al., 2008). Allgaier (2008), studying dune dynamics by Halamish, found that during strong winds of cyclonic storms, sand is also transported through the interdunes. This process might explain the transverse interdune sand influx and their east-dipping slip faces. Additional evidence for late Holocene dune activity include pottery sherds at a Late Byzantine (~1.4-1.7 ka BP) gathering site discovered upon an exposed upper dune surface impregnated with calcium carbonate (the Mitvakh site) (Figs. 3.1c & 3.5). This indicates dune activity just prior to Byzantine presence. Finally, Tsoar and Goodfriend (1994) dated the upper Ramat Beqa aeolian sand section by radiocarbon to similar ages (1.57-1 ka).

There are abundant late Holocene palaeoclimatic stratigraphic, archaeological and historic data from the Negev. This has furnished the debate if climate change, i.e. increased aridity, induced the collapse of the Byzantine towns and the extensive agricultural infrastructure in the Northern Negev (Issar et al., 1989; Rubin, 1990; Avni et al., 2006). The ruins of the Byzantine city of Halussa are covered by 1-2 m of sand, and historical letters attest to sand incursion that decimated the grape vines (Meyerson, 1994). The ages from the Retamim ID section (Fig. 3.5) upper ages, 3 km to the east of the ruins, corresponds to this period. Byzantine sites along the Northern Sinai and southern Mediterranean coast of Israel have been covered by several meters of sand (Neev et al., 1987) which may imply stronger winds from Mediterranean winter storms.

This late Holocene episode of sand remobilization and partial incursion into the Negev illustrates unusual spatial characteristics and dune superposition. In the western part of the central incursion corridor, a thick 8-12 m sequence of late Holocene sand overlies the main

incursion sand unit, whereas in the east there is only limited evidence for late Holocene sand accretion. The dunefield did not extend eastward beyond its Late Pleistocene depositional limits during this episode (Figs. 3.5 & 3.10). Was this due to a local sand-supply surplus in the west or strong and possibly locally confined westerly winds that somehow mainly affected the western part of the dunefield? This question is beyond the scope of the present study but provides working hypotheses for future research.

The cluster of modern (150-8 yr) OSL ages of dune mantle samples probably documents remobilization without dune elongation. This short episode could be due to anthropogenic causes, mainly trampling of biogenic crusts by livestock (Tsoar and Moller, 1986; Meir and Tsoar, 1996; Tsoar, 2008) in a climate with wind power similar to today's.

3.5.2 The temporal and spatial aspects of sediment supply for dune encroachment into the Negev

3.5.2.1 The inferred source and dynamics of the Northern Sinai dunefield

Sediment supply and strong winds are both prerequisites for sand transport from the Nile Delta through the Sinai portion of the Sinai-Negev Erg towards the NW Negev, and it is crucial to evaluate the geomorphic and palaeoclimatic controls of the dune encroachment into the Negev. The mechanism of sediment delivery from the Nile Delta and Mediterranean coast to northwest Sinai has not been investigated in detail (Hunt, 1991), however Nile Delta sediment storage is affected by the contribution of the Nile tributaries (Williams et al., 2000), the Mediterranean Sea currents and sea level change that defines the position of the coastline.

Strong southeastern aeolian sand transport drift potentials ($DP=1139$; $RDP=529$) were calculated for the years 1987-1993 from meteorological data from the Port Said airfield, Egypt, at the northeast edge of the Nile Delta. These winds had annual monthly averages of 7.4-10 m/s between the years 1989-1999 (GASCO, 2007). If these winds occurred in the past, they could have been the driving mechanism behind transport of both deltaic and coastal sand inland into northwest Sinai. These winds may explain the occurrence of non-vegetated linear dunes in the western part of the Sinai dunefield south of Port Said, that are currently elongating to the SSE by several meters/yr (Tsoar et al., 2004). Wind data from central and eastern parts of Northern Sinai also indicate that wind power decreases to the east towards the Negev. The Sinai dunes, in contrast to the NW Negev dunes, are currently uncrusted and active (Tsoar, 1995; Abdel Galil et al., 2000; Rabie et al., 2000). Dunes near Gebel Maghara (Fig. 3.1b) in Northern Sinai advance only several m/yr (Goldberg, 1977) while reconnaissance dune hazard studies imply that northern Sinai dunes elongate between 5-15

m/yr (Abdel Galil et al., 2000; Rabie et al., 2000). Although there is evidence for measurable sand movement in Northern Sinai at present, there must have been times in the past when sand advanced at substantially higher rates.

Sand transport across the Northern Sinai is controlled by sand availability and transport strength. The northern Sinai dunefield is spatially continuous over substantial areas, with dune heights exceeding 30 meters (Gad, no date), suggesting essentially constant availability of sand supply. This implies that the dynamics of Sinai sand movement is controlled chiefly by wind strength.

The palaeoclimate that enabled the Negev dunefield development has traditionally been interpreted to be the result of past aridity (Goring-Morris and Goldberg, 1990; Magaritz and Enzel, 1990; Hunt, 1991; Harrison and Yair, 1998). Elsewhere, active inland desert dunes were also interpreted to be indicators of arid conditions (Sarnthein, 1978; Hesse and Simpson, 2006) and this paradigm is in accordance with the assumption that dune mobilization thresholds are defined in part by a decrease in effective precipitation (Lancaster, 1988). In many parts of North America dunes are active in hyperarid environments where wind strength is low, while vegetated dunes are stable in semiarid environments where wind strength is high (Muhs and Holliday, 1995; Muhs and Wolfe, 1999). Recent modeling shows however, that dune activity is controlled dominantly by wind power and dunes can be mobilized even in humid climates when stripped from vegetation (Tsoar, 2005; Yizhaq et al., 2007, 2009).

In the NW Negev, dune erosivity and activity is controlled to a great degree by the amount of biogenic crust and vegetation cover (Tsoar and Moller, 1986; Tsoar, 2008; Kidron et al., 2009; Siegal, 2009). The thick late Holocene VLDs and transverse dunes that overlie the Late Pleistocene 5-10 m thick dune sands suggest that in some cases later incursions cover previous depositions, rather than rework them. This can be explained by an influx of sand that covers the still-stabilized encrusted sand and dunes. Buried crusts identified in pits dug into recent dune slopes and crests in the central study area also imply that recent sand remobilization can cover and bury biogenic crusts quickly and thus neutralize their stabilizing effect.

In scenarios where there is substantial sand supply, dune vegetation can however, direct sand transport mainly along pre-existing dune crests and thus exerts a certain control on dune morphology (Tsoar and Moller, 1986; Tsoar et al., 2008). Thus, we suggest that in some scenarios, for sand to encroach from Sinai into and on to the current NW Negev VLD landscape, sand supply is sufficient and Negev dune erosivity is not a prerequisite. Strong

winds, therefore, may be the main driver for sand transport from the northeast Nile Delta to northwest Sinai, across northern Sinai and into the NW Negev. Thus, the past incursion episodes that initiated dune elongation and buildup were mainly characterized by increased windiness.

3.5.2.2 The chronology of sand transport in Northern Sinai

The chronology of sand transport in Northern Sinai is important in understanding the process that led to the encroachment of dunes into the NW Negev. Analysis of the existing chronological and sedimentological data from the northeast Nile Delta across Northern Sinai allows a construction of a general chronological framework of the controls on and episodes of aeolian activity in the Sinai-Negev Erg.

Sediment supply into Sinai is influenced by sea level change (Edgell, 2006), as found for other regions (Preusser et al., 2002; Lancaster, 2008). The significant glacial to interglacial Mediterranean sea-level oscillations most likely have affected northwest Sinai – northeast Nile Delta sand availability (Edgell, 2006). The last glacial period (35-18 ka) lowered global sea-level by approximately 120-130 m (Fairbanks, 1989; Bard et al., 1990). The Mediterranean Sea dropped similarly and retreated 40-50 km north of the Nile Delta's mouth (Stanley and Warne, 1993; Butzer, 1997). Enzel et al. (2008) suggest that this regression exposed Nile Delta sediments to erosion and aeolian transport and was closely followed by a 30 m entrenchment of the Nile River into its delta (Butzer, 1997). Dated Late Pleistocene sediments from the northeast Nile Delta are sparse as most of the cores are 20-40 m long and have penetrated the full Holocene record but only the top of the Pleistocene sediments (Stanley et al., 1996). Coutellier and Stanley (1987) described two major fluvial sand units dated by radiocarbon to ~42-24 ka and ~24-11.5 ka, the latter deposited concurrent with the Mediterranean Sea level rise and is overlain by finer Holocene sediments (Stanley et al., 1996). In accordance, between ~30-11.5 ka the Nile Delta was a sandy alluvial plain (Stanley and Warne, 1993; Butzer, 1997).

Carbonate deposits beneath northwest Sinai sand by the Nile Delta Pelusian branch were also dated to ~35-30 ka, providing a basal age for dune sand (Neev et al., 1987). The Sinai dunefield southern end is directly east of the apex of the Nile Delta (Fig. 3.1). This juxtaposition could imply that the Nile Delta sand is the source for the Sinai dunes. Furthermore, desert sand facies in the northwestern Sinai dunefield and in the western Delta show strong and distinct similarities in sedimentological properties with the Late Pleistocene Delta sands, such as grain coating intensity (Stanley and Chen, 1991). The redness intensity

of the Negev dune sand was found to be similar to the Sinai sands (Roskin et al., 2010), suggesting that they all draw from the same source.

The impressive Mesozoic carbonate ridges of Gebel Maghara and Gebel Lagama rise several hundred meters above the western-central part of the Sinai dunefield. The ridges are dissected by wadis of different sizes and are surrounded by linear dunes, climbing dunes and sand ramps that have attracted researchers (Bar-Yosef and Phillips, 1977; Goldberg, 1977; AMEA, 2006)(Fig. 3.1b). The geo-archaeological investigation of Late Quaternary sand stratigraphy at the basal slopes of Gebel Maghara and Gebel Lagama (Goldberg 1977) is the only detailed study of past sand activity and stabilization in northwestern Sinai, 150 km west of the Negev dunefield. The age estimates for the cultures in this work are derived from prehistoric sites with similar artifact assemblages and radiocarbon dates on charcoal and ostrich shells (Table 3.1). At Gebel Lagama, basal aeolian sand predates Lagaman sites; thus initial sand buildup is estimated to ~40-33 ka (calibrated) (Table 3.1; after Goldberg, 1977; Goldberg, 1986; Goring-Morris and Goldberg, 1990). This gives abundant time for dune buildup in northern Sinai prior to and during the LGM, and precedes the NW Negev dune incursion that starts after 30 ka. Based on this evidence, along with finds in northeastern Sinai (Goldberg, 1984, 1986; Gladfelter, 2000) and the corresponding NW Negev OSL age suite (Table 3.1), we suggest a basal age for the Sinai- Negev Erg sands of ~35 ka. Nevertheless, luminescence ages of basal Northern Sinai aeolian deposits and dunes are essential for validation.

Near Gebel Maghara, Mushabian and Geometric Kabaran (~15.6-13.2 ka) sites lay directly over a weak calcic palaeosol located in the middle of an aeolian sand section (Goldberg, 1977). Such young palaeosols were not found in the Negev dunefield, only around it (Zilberman, 1993; Zilberman et al., 2007; Crouvi et al., 2008; Wieder et al., 2008). The palaeosols may attest to a pause in sand mobilization, interpreted by Goldberg (1977) as a more humid climate. In addition, Mushabian artifacts are contemporaneous with a palaeolake. Thus a dune incursion that blocked a wadi occurred in northern Sinai at a time similar to dune damming in the NW Negev (Fig. 3.6f).

The Gebel Maghara-Lagama aeolian sand dates suggest similar dune activation and stabilization ages for Northern Sinai and the Negev, particularly in regard to the main dune incursion (16-11.5 ka) indicating rapid and episodic dune advancement across the Erg as found around ~15 ka in the Negev. The similar ages in the Sinai and Negev parts of the Erg, as well as the ages suggested for the northeast Nile Delta, which was the likely source of sand during lowered sea level (Coutellier and Stanley, 1987; Stanley and Chen, 1991; Stanley and

Warne, 1993; Stanley et al., 1996) leads us to hypothesize that the initiation of the Sinai-Negev dune incursion was possibly constrained by insufficient sand supply before ~35 ka.

Differential sand supply from Northern Sinai can also explain the differences between the three NW Negev incursion corridors (Fig. 3.1). The morphological characteristics of the dunes in the Negev incursion corridors, other than vegetation and biogenic crust cover, are similar westward in northeast Sinai, between Wadi Al-Arish and the Egypt-Israel border. The low and broad (Table 3.2) northern incursion corridor dunes are probably restricted by limited sand supply, as their upwind direction coincides with the Mediterranean Sea. The central incursion corridor though, which has the thickest sand section (Fig. 3.9), the relatively higher quartz content (Fig. 3.3) and the easternmost extent (Fig. 3.1c), is the easternmost part of the main Sinai sand transport corridor that crosses the center of the Northern Sinai dunefield north of Gebel Maghara. The similar basal dune incursion age throughout the western transect (Figs. 3.5 & 3.9) also suggest that at the same time and climate, sand influxes differed between the north and south parts of the dunefield, mainly due to sand supply controls upwind in Northern Sinai.

3.5.2.3 Last glacial luminescence-dated global linear dune activity

The short but main 16-11.5 ka episode of dune-building in the NW Negev (Fig. 3.8) is striking and invites comparison on the one hand with episodes of linear dune formation in similar low latitude arid regions, and on the other hand with global windiness. The concept of gustiness (McGee et al., 2010) attempts to explain the LGM global dust, and it had been suggested that many regions experienced stronger winds in the LGM (Petit et al., 1990; Mahowald et al., 1999). Previous compilations of luminescence-dated global dune building and activity episodes (Lancaster, 2007; Singhvi and Porat, 2008 and references therein) from varying dune types show possible distinct episodes, but with significant local variability.

Extensive luminescence dating of linear dunes has been conducted only in Australia (Fitzsimmons et al., 2007) and southern Africa (Telfer and Thomas, 2007; Stone and Thomas, 2008). These dune chronologies, along with the NW Negev OSL ages show a peak of dune activity from the end of the LGM until the beginning of the Holocene.

Global Holocene dune activity differs from the Late Pleistocene record and is characterized by a highly variable and spotty records of short-term mobilization episodes (Singhvi and Porat, 2008). This may be due to a higher resolution stratigraphy and sampling, anthropogenic influence and lack of differentiation between dune elongation and dune mobilization episodes. The distinct late Holocene episode of the NW Negev does not match

most of the global records. However, at the Godolla Hills, Hungary, an IRSL-dated sand section has been found to be strikingly similar to the NW Negev dunes with an upper 1.5-2.2 ka unit overlying a 14.3-15.5 ka unit (Novothny et al., 2010).

Beyond these observations, most global dunefield luminescence age suites lack a sufficient quantity of ages to configure a reliable palaeoclimatic scheme (Telfer et al., 2010). The global age suites may also partially be a result of sampling, where clustering of dune ages was biased to upper dune sections that were more commonly sampled (Bateman et al., 2003). In addition, conclusions were drawn from small age sets of 10-20 samples that represent vast areas from different sedimentary environments. Though there is no universally accepted minimum number or density of OSL ages recognized for dunefield studies, a data set of around 100 ages (depending on dunefield size and morphology), seems to be a minimal condition for a reliable reconstruction of dunefield evolution (Telfer et al., 2010).

Nevertheless, there seems to be a certain contemporaneous activity of dune in low-latitudes around the LGM and during the post-LGM period, with a dramatic cessation of activity with the commencement of the Holocene. This suggests that the Late Pleistocene - Holocene transition involved a long-term decrease in gustiness, a hypothesis that requires more testing.

3.6 Conclusions

In this study, we present a large and spatially dense OSL database for the NW Negev dunefield, supported by fully documented dune sections, sedimentological data and geomorphic attributes. The Sinai-Negev Erg is young in a global perspective. Encroachment episodes in the Sinai section of the Erg are suggested to be chronologically similar to the dune incursion episodes into the NW Negev. While evidence for sand is found since 100 ka in the NW Negev, sand supply that generated the most OSL-dated dune accretion and elongation is suggested to have begun only since ~35 ka due to ample Nilotic sand supply. Calcic soils in the NW Negev were exposed at the surface until ~30 ka, after which they were eroded and buried by the encroaching dunes, indicating stability prior to the LGM. A relatively short LGM and post-LGM (23-11.5 ka) gusty climate transported aeolian sand into the Negev in several incursion episodes. At ~23 ka, initial evidence of dune buildup is identified. Between 16-11.5 the NW Negev witnessed extensive dune incursion. A rapid accretion event around ~16-14 ka, followed by short-term stabilization, left impressive dune and sand sections in the Negev and suggests that the post-LGM period had a stronger aeolian imprint on the region. By ~11.5 ka, dunes covered the full extent of the present dunefield, and a significant drop in

the regional gustiness may have occurred shortly thereafter. These episodes generated dune damming of Negev fluvial systems and produced lakes and ponds that supported prehistoric short-term camps.

The distinct late Holocene (2-0.8 ka) dune mobilization and incursion episode developed under conditions very different than the Late Pleistocene dune incursion environment. It certainly illustrates the sensitivity of the aeolian dune system to external forcing mechanisms, and additional investigation is needed to define the controls for this episode.

The sedimentary archive of the Negev VLDs demonstrates that different VLD morphologies have similar chronostratigraphy and VLD buildup and elongation are probably reliable proxies for periods of regionally gusty climates, as long as sand supply is not a limiting factor. These finds reinforce the value of VLDs for evaluating past regional environments.

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Chapter 4: Do Dune Sands Redden with Age?

The Case of the Northwestern Negev Dunefield, Israel

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4.0 Abstract

The redness index (RI) ($RI = R^2/(B \cdot G^3)$) 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.

4.1 Introduction

4.1.1 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 (Fe_2O_3) 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 quantification 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 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 homogeneous 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.

4.1.2 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 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 (Madiera 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 quantify 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 dune 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.

4.2 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. 4.1). 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.

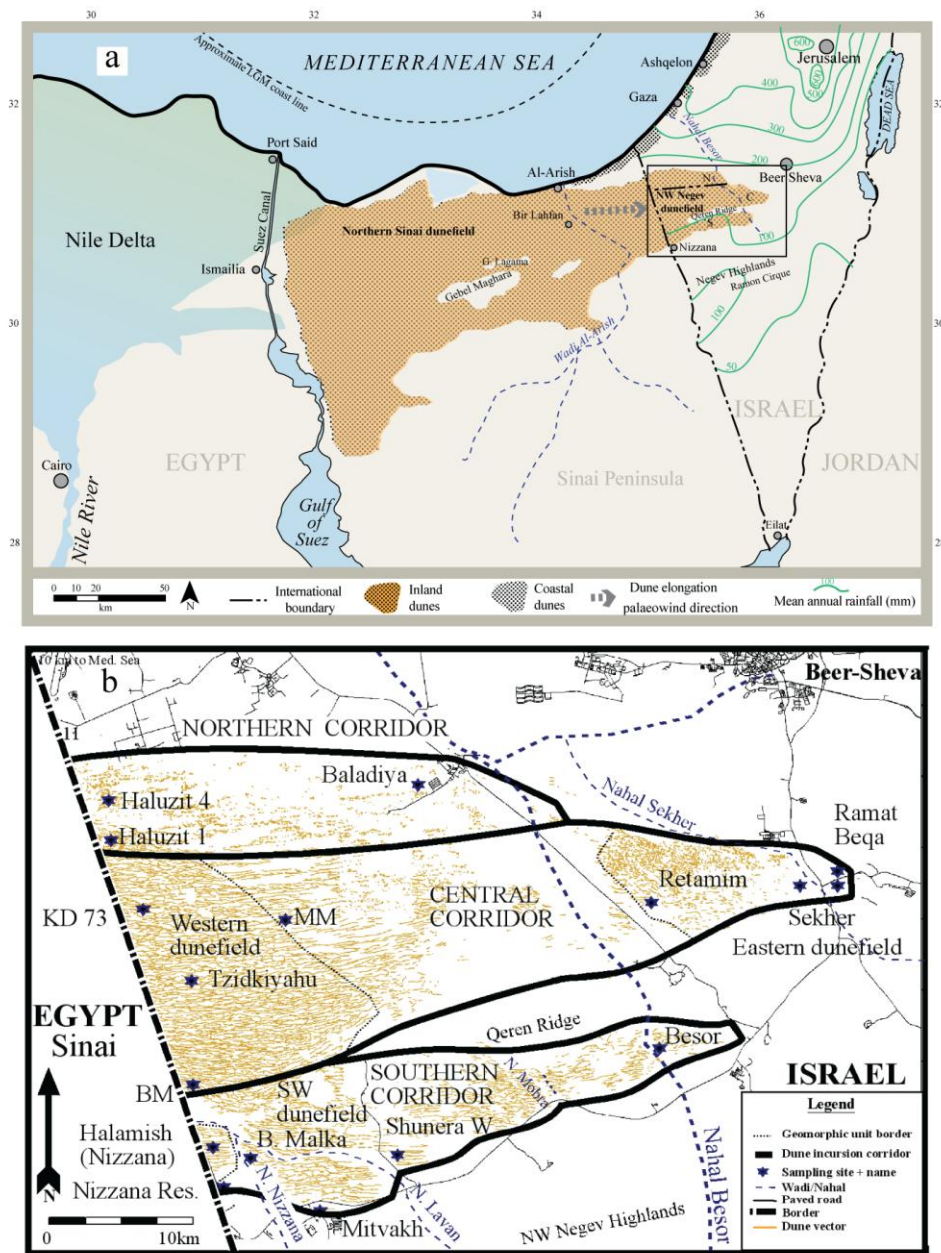


Figure 4.1. a. A regional map of the Sinai-Negev erg. The erg, stretches south and parallel to the southeastern Mediterranean coastline and extends eastward from the middle and upper northeastern Nile Delta, crosses the Egypt-Israel border (dotted black line) and extends into the northwestern (NW) Negev 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. Grey box depicts figure 1b. b. Dune axis mapping results, sampling site names and incursion corridors (in capital letters). Dunefield regions [southwestern (SW), western and eastern] are also displayed and are referred to in the text.

4.3 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. 4.1a). 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 Hillal 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, 1956) (Fig. 4.1).

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 longshore currents along and onto the northern Sinai and southern Israel coasts (Emery and Neev, 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, 1995; Misak and Draz, 1997; Abdel-Galil et al., 2000; Rabie et al., 2000) 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 dunefield (N30°/E34°) constitutes the Israeli section, covering approximately 1,300 km², of the Sinai-Negev Erg (Fig 1a, b). Its location at the downwind end of the erg, where sand has been deposited since the Late Pleistocene (Roskin et al., 2011), is considered a suitable location to study sand rubification. The dunefield comprises stable vegetated linear dunes (VLD), whose vegetation cover, ranging from 5 - 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 transects 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/yr as measured at Nizzana in the southwest corner of the NW Negev dunefield (Fig. 4.1) (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 Munsell 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 Munsell 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- μ m 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 these clay films and colorization of the sand grain. The extractable Fe range by DCB for the majority of the Negev sand samples

(0.06 - 0.14 % Fe) (Tsoar et al., 2008) was higher than that for Israeli coastal sands (0.02 - 0.05% Fe) (Ben-Dor et al., 2006).

The Negev dune redness ratio based on laboratory spectroscopy was found to have limited spatial consistency with the slight reddening trend observed from west to east (Campbell, 1999). However, these were confined to surface samples, and sampling sites were based neither on sand transport paths/corridors nor on statistical logic. Using laboratory spectroscopy, Wenkart (2006) and Tsoar et al. (2008) divided the dunefield into three sand units based solely on contouring a grid of the spectroscopic redness index ($RI = R^2/(B \cdot G^3)$) of sand sampled ~20 cm beneath the surface. Wenkart (2006) and Tsoar et al. (2008) suggested that the west-central part of the dunefield north of the Qeren Ridge represents the latest dune incursion due to its lower RI values, whereas the redder northern and eastern dunefields contain the most mature sands. However, these sand units only partially agree with sand-transporting wind directions and linear dune orientation, and therefore, they were subsequently modified into three dune encroachment or incursion corridors (northern, central, and southern) by Roskin et al. (2011a) based on GIS dune axis and slope mapping (Fig. 4.1). Dune corridor classification was further supported by mineralogical analysis and a spatially dense sampling campaign of 97 OSL ages from dune and interdune sections that generated several important findings. Sporadic aeolian sand deposition between ~116 - 30 ka stabilized and developed into calcic loamy palaeosols, pre-date the Negev dune encroachments. Aeolian sand cover, sufficient to form dunes, began to accumulate only at ~23 ka in the southwest dunefield corner. Following the last glacial maximum (LGM) (22 - 18 ka), the dunes invaded the NW Negev along the three main west-east encroachment corridors, and three main chronostratigraphic layers accumulated. The main dune encroachment occurred between 18 - 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 - 8 yrs), respectively (Roskin et al., 2011a). Beyond these episodes, 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.

4.4 Field and laboratory methods

4.4.1 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. 4.1). 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 conditions. Most of the sections are exposed, which improved the reliability of the chronostratigraphic analysis. Drilling was performed with Dormer Engineering hand augers, and drill holes were usually cased with PVC pipes to stabilize the borehole and prevent sediment fall and sample contamination. Sand was usually sampled approximately every 1.5 meters along drilled sections or according to stratigraphic units identified in exposed sections. More than 200 sand samples were gathered from 32 dune and interdune sand sections. Samples for OSL dating were chosen from 28 sites (Table 4.1).

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 analogues 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 were sampled (Fig. 4.1). Several samples were collected near representative sampling sites of Wenkart (2006) and Tsoar et al. (2008) and several samples from throughout the Negev dunefield, measured by Wenkart (2006) and Tsoar et al. (2008) were re-measured to ensure that our RI measurements were consistent with the previous work.

Table 4.1 List of the main sedimentological, RI, and OSL results for all sites and samples.

The main site names appear in Fig. 4.1b and the numbered sites appear in Fig. 4.5b.

Cc = calcium carbonate. A = Hand augered section. Ex = Exposure.

Section no' + physiography / sample (DF)	Main field description + location (for Sinai)	Section type and sampling depth (m)	Redness Index (RI)	% fines (silt+clay)	OSL age (ka)	Section RI average and Std. Deviation
NORTHERN CORRIDOR						
1. Haluzit4		Ex+A				
Broad VLD crest						
31	sand	0.6	37.7	13.2	0.085±0.04 5	36.3
32	sand	1.2	41.2		0.14±0.03	5.5
33	sand	1.5	45.2			
34	sand	1.9	37.6	14.8	1.4±0.2	
35	sand	3.3	28.1	12.6	10.6±1.6	
301	fine dry sand	2.9	30.8		8.7±1.0	
302	sand	3.8	30.3		12.0±1.6	
303	sand	4.5	36.6			
304	sand	4.9	34.6	13.9	12.8±1.5	
305	sand+clay peds	5.9	41.3			
2. Haluzit4		Ex				
Interdune (ID) sand						
40	disturbed sand	0.3	32.9			38.1
41	relatively coarse sand	0.6	36.6	19.1	9.6±0.5	5.5
42	slightly cemented dark sand	1.8	44.9	21.5	11.4±0.6	
3. Haluzit1		Ex				
VLD axis						
802	moist sand	2.9	32.5	20.4	1.7±0.3	33.3
803	sand	3.7	28.4	8.6	13.7±0.4	4.7
804	sand	4.5	27.3	10.2	13.7±1.1	
81	sand	6.8	35.3	19.3	15.5±2.2	
82	sand + dark cc nodules (lag deposits)	7.0	37.8	22.6		
83	clay-silty sand	7.5	38.4	27.4	106±19	
4. Baladiya		Ex+A				
Wide broad VLD						
71	fine sand	0.4	70.4			73.0
72	fine sand	0.7	69.2	24.1		9.8
73	fine sand	1.1	75.4			
74	fine sand	1.8	77.6	13.4		
75	cemented sand	2.4	50.8	24.2	3.0±0.6	
76	slightly coarser sand	3.2	66.2	17.9	13.7±1.7	
712	moist, 1-2% 1 mm cc stains	3.6	72.5			
713	sand, slightly moist, peds	4.7	71.7			
714	1% white cc stains, soft peds	5.7	82.9	8.7	15.6±0.7	
715	1% white cc stains, soft peds	6.7	87.3			
716	uniform brown sand, no cc	7.8	88.1		15.9±0.7	
717B	sand	8.0	66.0			

717A	single cc stains	8.5	71.7		
718	fine silty sand	9.3	60.0		
719	light uniform fine sand	9.8	76.4	20.3	14.7±1.9
720	condensed sand	10.3	81.9	26.1	

CENTRAL CORRIDOR

5. KD 73		A			
Interdune depression					
beneath transverse dune					
slip face					
681	dry fine sand	2.0	21.5	17.1	13.3±1.4
682	slightly moist coarse sand	3.0	25.9	13.2	26.6
683	sand	4.0	29.6	6.0	3.8
684	slightly moist coarse sand	4.5	25.2	9.4	
685	sand	6.0	31.0	14.0	17.9±2.8
6. KD 73		A			
VLD crest					
690	moist sand	1.5	21.0	11.4	22.4
691	moist like DF 690	3.0	22.7		1.5
692	increasingly moist	4.5	20.8	7.0	
694	slightly moist sand	8.0	23.1		
695	moist sand	9.2	24.5	6.1	15.6±1.5
7. Tzidkiyahu Interdune		A			
Depression beneath					
transverse dune slip face					
521		2.5	23.3	5.6	24.9
522		3.4	23.0		15.8±2.2
523	moist sand	4.0	22.9		1.5
524	moist sand	4.5	23.8	7.0	15.5±0.9
525	moist, slight brown streaks	5.5	25.6		
526	"	6.6	24.8		
527	"	7.3	26.7		
528	"	7.9	25.0		
529	"	8.7	25.1		
530	moist sand, brown pigmentation	9.7	26.4		
660	moist sand	10.3	26.9	4.0	15.9±1.7
Tzidkiyauh ID transverse		A			
dune base					
651	very slightly moist sand	1.6	23.4		23.5
653		4.0	24.0		0.4
654	moist sand	4.9	23.2	4.6	
8. Tzidkiyahu tranverse		A			
dune between VLDs					
531	sand	1.0	31.9		30.3
532	slightly moist sand	2.1	29.5		1.1
533	moist sand	3.1	30.6		
534	moist	4.6	28.5	5.3	1.4±0.1
535	more moist	5.8	30.6		
536	"brown sugar" texture	6.8	30.4		
537	more moist sand smears auger circumference	7.9	30.6	9.2	1.2±0.1
9. Tzidkiyahu VLD crest		A			

550		1.2	31.5	6.4		28.3
551	dry/initially cohesive sand	1.8	29.7			2.3
552	dry/initially cohesive sand	2.9	31.7			
553		3.8	27.5			
554	moist sand	4.9	26.7	4.8		
555	moist sand	6.2	25.9			
556		6.0	26.5			
557	moister sand	7.2	27.1	4.4	1.4±0.1	
10. BM west VLD crest		A				
8001		1.5	29.8			28.7
8002/508	moist sand	3.1	28.5	7.3		1.1
8003	"	4.0	29.8			
8004/509	"	5.5	27.7	6.3	0.5±0.1	
8005	"	6.6	27.3			
8006/510	"	7.7	30.0	12.7		
8007	slightly less moist sand	8.8	28.5			
8008/511	moist sand	9.8	27.6	13.1	0.9±0.1	
110	sand	4.0	28.7	9.0	0.5±0.1	
111	sand	4.7	28.4	13.5	1.8±0.2	
11. BM East VLD crest		A				
7001	slight moist sand	2.4	31.6			31.1
7002		3.1	32.6	6.6		1.3
7003	moist sand	3.8	32.7			
7004/513	moist sand	4.5	31.5	5.4	0.8±0.1	
7005a	slightly consolidated sand	5.3				
7005 (b)/514	moist sand	6.3	29.7	16.2	0.8±0.1	
7006	moist sand	6.8	29.7			
7007/515	moist sand	7.6	29.9	14.6	1.7±0.2	
BM East VLD flank		A				
130		1.0	29.3			29.9
135	sand	2.5	27.9		1.2±0.1	0.9
136	sand	~1.4	30.8			
137	sand	1.3	29.4			
138	sand	~1	29.9			
139	sand	0.8	29.0			
13. BM VLD interdune		A				
4000	slightly moist sand	1.5	27.5			35.6
4001	"	2.5	28.3			6.7
4003	quite moist sand	3.8	34.5			
4004	moist sand	5.0	43.2	9.7		
4005	"	5.6	42.1			
4008 / 506	less moist sand	6.5	37.8	8.7	17.7±2.1	
BM east Interdune playa, (South)		A				
485	moist sand	5.9	41.7	7.3		47.2
486	moist sand	6.9	48.3			5.0
487	sand	7.9	53.6	7.9	14.7±2.0	
488	moist sand	8.6	45.2			
BM east Interdune playa, (North)		A				
6000	silty sand	0.5	29.0			35.0

6001	sand with 5% silt	1.0	29.5			6.6
6002	dry sand	2.2	40.1			
6003	dry sand	2.8	41.2			

13. MM		Ex. + A				
VLD crest						
10	sand	0.5	31.4			35.8
11	sand	1.3	33.4	10.6	0.045±0.01	4.0
12	sand	2.0	36.4			
13	sand	2.6	34.9	9.1	0.04±0.01	
14	sand	3.6	44.7			
15	sand	4.7	36.4			
16	sand	5.7	36.3	9.3	1.3±0.1	
17a	sand	7.0	33.3	13.5	9.3±2.0	

14. Retamim		A				
Interdune						
540		0.8	30.7			36.9
541	dry sand	1.7	33.0	10.3	1.5±0.1	11.0
542		2.4	36.6			
543	moist sand	3.3	24.1	5.8	16.0±2.1	
544		3.8	21.3			
545	moist sand, 1-5cm cc nodules	4.6	39.0	3.5	19.3±1.8	
546		5.3	20.6			
547		5.7	42.9			
548	dark brown moist sand	6.7	49.0	8.3	27.3±3.8	
549	clean sand	7.4	45.7			
700	white stains sand friable peds	7.6	50.3	17.1	22.8±3.1	
701	peds with red rusty stains inside	7.7	53.0	9.0		
702	silt y blocky peds, no nodules	7.8	34.0	26.5		

15. Retamim Broad VLD		A				
crest						
560		0.4	40.0	13.1		37.6
561	moist sand	1.5	35.2	12.4		2.0
562		2.3	40.1			
563	moister sand	2.9	38.9	8.8		
564		4.0	35.4			
565		4.6	38.3	6.5		
566		6.1	37.9	5.1		
567		7.1	37.5			
568		7.8	35.0	4.4	10.7±1.5	

16. Ramat Beqa quarry		Ex.				
575	silty-sand	8.0	55.9	21.9	11.6±1.8	53.1
576	silty sand	5.2	42.0	16.6		8.0
577	sand+snail shells	4.9	46.6	13.9		
578	sand	4.9	57.2	6.2	4.8±0.7	
579	sand with cc stains and 1-2 mm nodules	4.3	64.4		5.6±0.6	
580	sand with cc stains and 1-2 mm nodules	3.1	52.2	14.8		

17. Sekher VI, South		Ex.				
1	Uniform structureless sand, roots, slight consistency, vertical crack, burrows, 3% snailshells <1mm	0.5	39.1	8.9	3.2±0.5	38.6
2	Below Natufian layer, sand	0.8	38.3	9.1	11.9±1.4	0.5
3	dry loose sand	1.5	38.4	10.4	13.7±1.3	

18. Sekher VI, North	Ex.					
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5	horizontally bedded sand	1.6	40.6	10.2	2.9±0.1	39.4
6	10 cm Natufian layer. wavy, cc nodules, lag deposits	1.8	41.7	8.1	3.7±0.3	3.0
7	sand	2.0	36.0	7.2	12.5±0.5	

SOUTHERN CORRIDOR

19. Halamish west VLD crest

Ex.

611	sand	1.0	41.6			39.4
612	sand	1.7	37.3			3.1

20. Halamish East VLD crest

A

581	slightly moist sand	1.0	60.2	10.7		63.0
582	moist sand	1.8	59.0			6.6
583	moister sand	3.0	56.5	15.5		
584	"	4.0	64.8	9.6		
585	"	4.8	63.7	7.5		
586	"	6.0	59.3			
587	moister sand	7.0	78.7			
588	extremely moist sand	8.0	56.4			
589	extremely moist sand	9.0	64.2			
632	moist sand "brown sugar"	9.4	66.9	20.9	13.2±1.8	

21. Halamish VLD flank

Ex.

618	dune sand	1.2	61.1	14.9	23.3±3.4	61.1
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22. Halamish Interdune

Ex.

625	fine sand	1.3	35.6	26.8	14.3±1.8	37.9
626	fine sand	2.3	40.2	27.1	19.1±2.9	3.2

23. Beer Malka Transverse dune

Ex.

1	coarse sand	3.0	44.3	5.9	17.5±3.3	47.0
4	coarse sand	3.0	49.7	8.0	12.2±2.1	3.8

24. Nizzana Reservoir

Ex.

516	fine horizontally bedded sand	2.3	53.9	30.0	14.7±1.7	57.5
518	sand	3.7	61.1	25.9	20.4±2.4	5.0

25. Mitvakh VLD

Ex.

200	sand	9.3	62.0	21.0	14.3±0.8	62.0
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26. Mitvakh VLD crest East

A

671	slightly moist sand	3.6	44.5			42.5
672	loose sand	5.5	42.3			1.9
673	loose sand	6.5	40.8			

27. Shunera West VLD flank

Ex.

601	active sand	3.0	50.1			64.5
602	clean sand	8.0	63.4			10.6

603	sand with carbonate stains	10.5	73.3		
604	sand	12.7	71.4		
28. Besor terrace		Ex.+A			
Fluvial dune terrace					
635	dry sand	1.1	52.0		55.3
637	moist sand	2.1	56.8		6.5
638	sand	3.0	55.8		
639	sand few fines, slightly moist	3.5	54.6	23.7	12.3±0.9
640	dark sand , slightly moist	4.4	61.2		
641b	dark sand, few peds, slightly moist, white auger smears	4.6	43.3		
641a	sand	4.5	63.4		
RAMON CIRQUE, KURNUB GROUP – HATIRA FM SANDSTONES					
RAM1	red		98.1		40.1
RAM2	pink		10.2		34.9
RAM3	purple		60.2		
RAM4	green yellow		6.8		
RAM5	white		42.9		
RAM6	orange		22.5		
JURASSIC SANDSTONE					
RAM10	brown		137.1		71.7
RAM11	brown		47.1		57.3
RAM12	brown		30.8		
NORTHERN SINAI SANDS					
A20	Wadi Ghazala dune sand	no data	44.6		
A24	Bardawil Sabkha dune sand	no data	44.5		
A30	Al Arish dune sand	no data	31.3		
A22	Gebel Hamir (base?) dune sand	no data	31.0		
A4	Bir Hasana barchan dune	no data	27.3		
A38	Bardawil Sabkha sand	no data	26.2		
A1	Bir Hasana barchan dune	no data	25.9		
A26	Bir Gafgafa dune sand	no data	24.0		
A32	Al Arish dune sand	no data	21.6		
A16	Wadi Kharadein channel fluvial-reworked? sand	no data	21.2		
A39	Wadi Al Arish (by A-A airport) bank sand	no data	21.0		
A46	Gebel Libni jnct. rippled and laminated fluvial-reworked? sand	no data	18.9		
A34	Al Arish beach sand, heavy minerals	no data	18.2		
A67	(Bir) Maqdaba (Al-Arish?) wadi bank horizontal stratified fluvial-reworked? sand	no data	16.6		
A62	Al Arish (Wadi ?) channel bank fluvial-reworked? sand	no data	12.1		

4.4.2 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 decimate peds. Immediately prior to measurement, the sand samples were 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. Measurements were taken from four directions for each sample to avoid a 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.

4.4.3 Spectroscopic indices

The redness index [(RI), $RI = R^2/(B \cdot G^3)$] was found to be a favorable index for the quantitative spectral measurement of sand rubification (Ben-Dor et al., 2006; Levin et al., 2007) and has been applied to the Negev dunes by Tsoar et al. (2008) (Table 4.2). RI values correlated to extractable iron oxide after Ben-Dor et al. (2006) and Tsoar et al. (2008) ($R^2 = 0.89, 0.67$, respectively) for Israeli coastal and Negev sand coatings, 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 4.2). The dimensionless redness indices represent a ratio that indicates relative degree of redness. Continuum removal (CR) transformation of the NW Negev sands spectra showed a distinct absorption at 498 nm (Wenkert, 2006), close to that of goethite (485 nm) (spectral library, Grove et al., 1992), indicating the spectral potential to map this mineral. We chose the specific R, G, and B bands after Ben-Dor et al. (2006), though both RI results using the Ben-Dor et al. (2006) and Tsoar et al. (2008) bands are positively correlated ($R^2 = 0.94$).

4.4.4 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, leading to the rehabilitation of biogenic crusts and vegetation (Meir and Tsoar, 1996; Karnieli and Tsoar, 1995; Tsoar, 2008; Tsoar et al., 2008). By 1987, developing Negev dune vegetation and crust covers are presumed to have already created a bias in the Wenkart (2006) ferric index analysis based on Landsat imagery. To compare the Sinai results to those of the Negev and to minimize the effect of the biogenic crust on Negev surface reflectance, the 1987 dates were nevertheless chosen because they were closest (earliest) to the land cover change that began in 1982. Another image taken in August 2003 was examined for control and is mainly applicable to the relatively bare Sinai sands.

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 A).

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. An R^2 correlation of 90% was found between the ASD Fieldspec spectrometer RI and the resampled bands (Table 4.2). Regional redness index maps of northeastern Sinai and of the NW Negev sands were processed using the RGB bands (Table 4.2).

Table 4.2. Redness index band data from previous studies and the Landsat TM images, and their inter-relationship and relationship to Fe mass of sand-grain coatings.

Source	R-Band	G-Band	B-Band	R^2 of RI vs Fe mass extracted by dithionite- citrate-bicarbonate (DCB)	R^2
Tsoar et al., 2008	640	510	460	0.67	
Ben Dor et al., 2006	693	556	477	0.89	0.94 between Tsoar et al. (2008) and Ben Dor et al. (2006) RGB bands
Landsat 5 TM images (Row 175, images 38, 39)	Band 3; 614- 704 nm	Band 2; 500- 624 nm	Band 1; 435- 520 nm		0.90 between ASD Fieldspec spectrometer derived RI to RI by resampling the ASD spectra to Landsat bands

4.4.5 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).

4.4.6 Sedimentology

To better understand other sedimentological factors that may promote or hinder sand-coating redness, we investigated the link 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 sections were measured for moisture content by oven-drying.

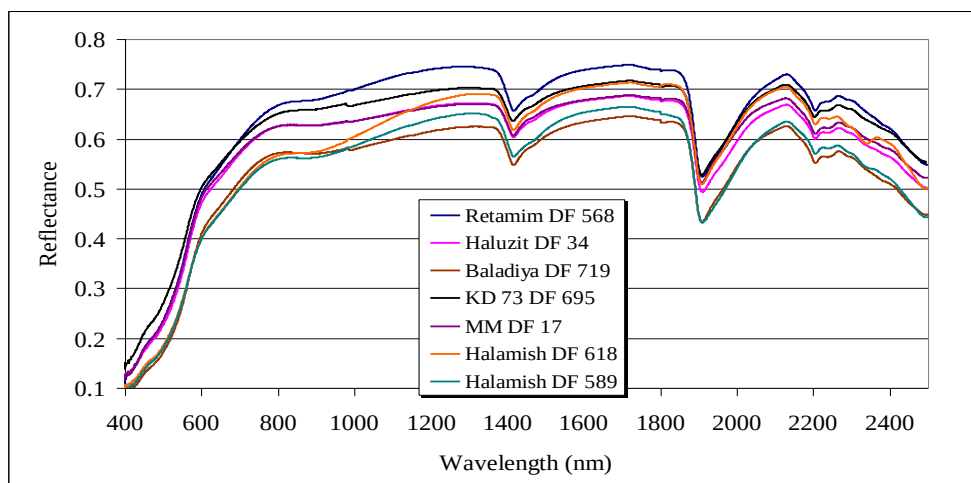


Figure 4.2. The reflectance spectra of sand samples from different parts of the NW Negev dunefield. The samples display different RI values. For sample details and section location, see Table 1 and Figure 4.5.

4.5 Results

4.5.1 Redness index properties

The reflectance spectra of seven samples from sections throughout the Negev dunefield at depths ranging from 2 - 10 meters are similar with respect to spectral features, although slight differences can be noted in the average reflectance and brightness (Fig. 4.2). RI and sedimentological results are presented in Table 4.1. 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 usually displayed relative sand color, as observed in the sampling plates, and corresponding RI down their vertical profile (Figs. 4.3 and 4.4; Table 4.1). Therefore, to obtain a representative RI value for each section, an average RI value and standard deviation of each section was calculated. This enabled the spatial trends of the RI to be mapped. Standard deviations for the RI of sections did not usually exceed ~15 % (Table 4.1; Fig. 4.5).

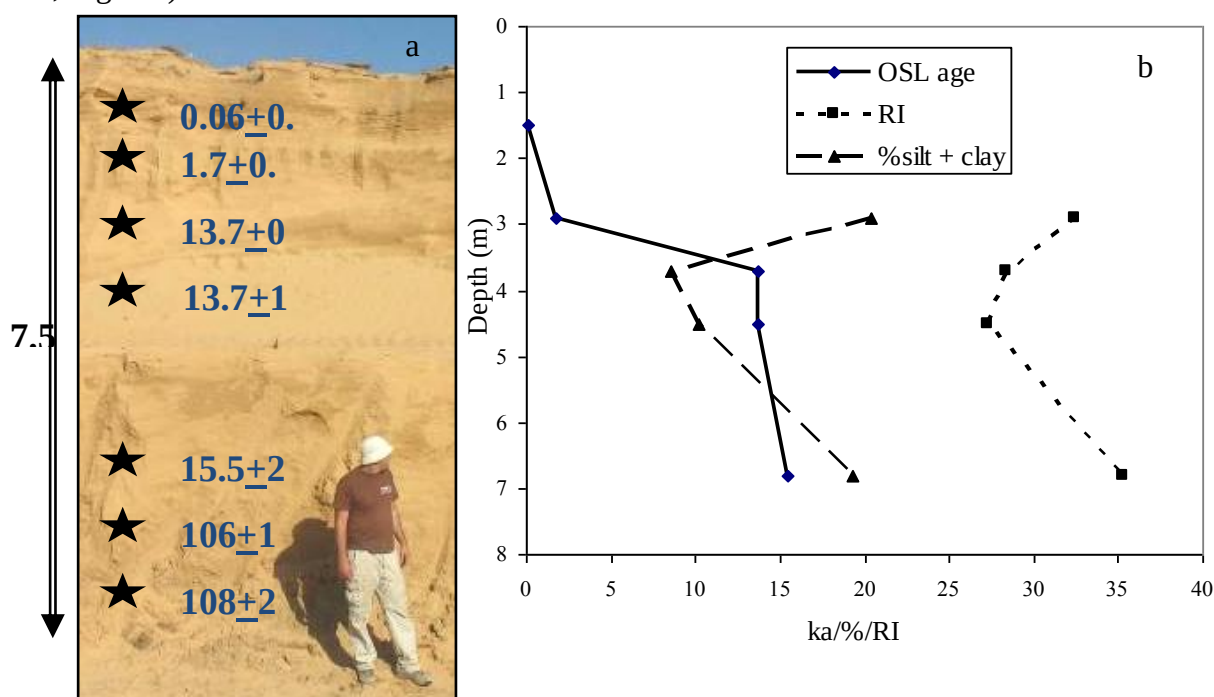


Figure 4.3. a. The Haluzit 1 section exposing a vegetated linear dune (VLD) axis. The OSL samples were defined based on dune stratigraphy. Observe the similarity of the sand color. The bottom two ages are from calcic sandy palaeosols and not dune sand. b. A depth profile of the OSL age, RI, and fine content of the section sands.

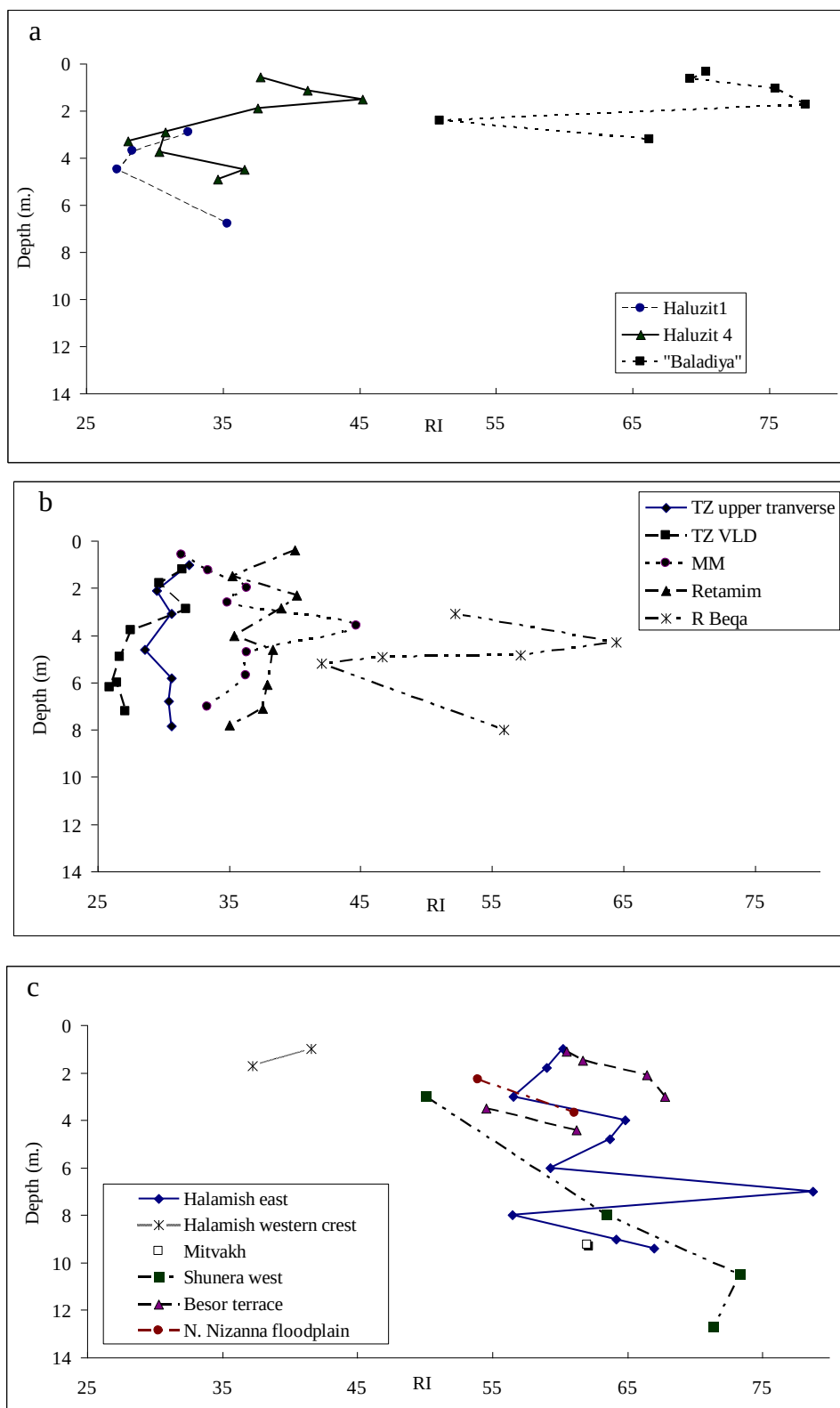


Figure 4.4. A plot of the changes in RI in relation to sand and dune section depth for selected sand and dune sections in sand/dune encroachment corridors: a. Northern corridor. b. Central corridor. c. Southern corridor. For section location, see Table 4.1 and Figure 4.5.

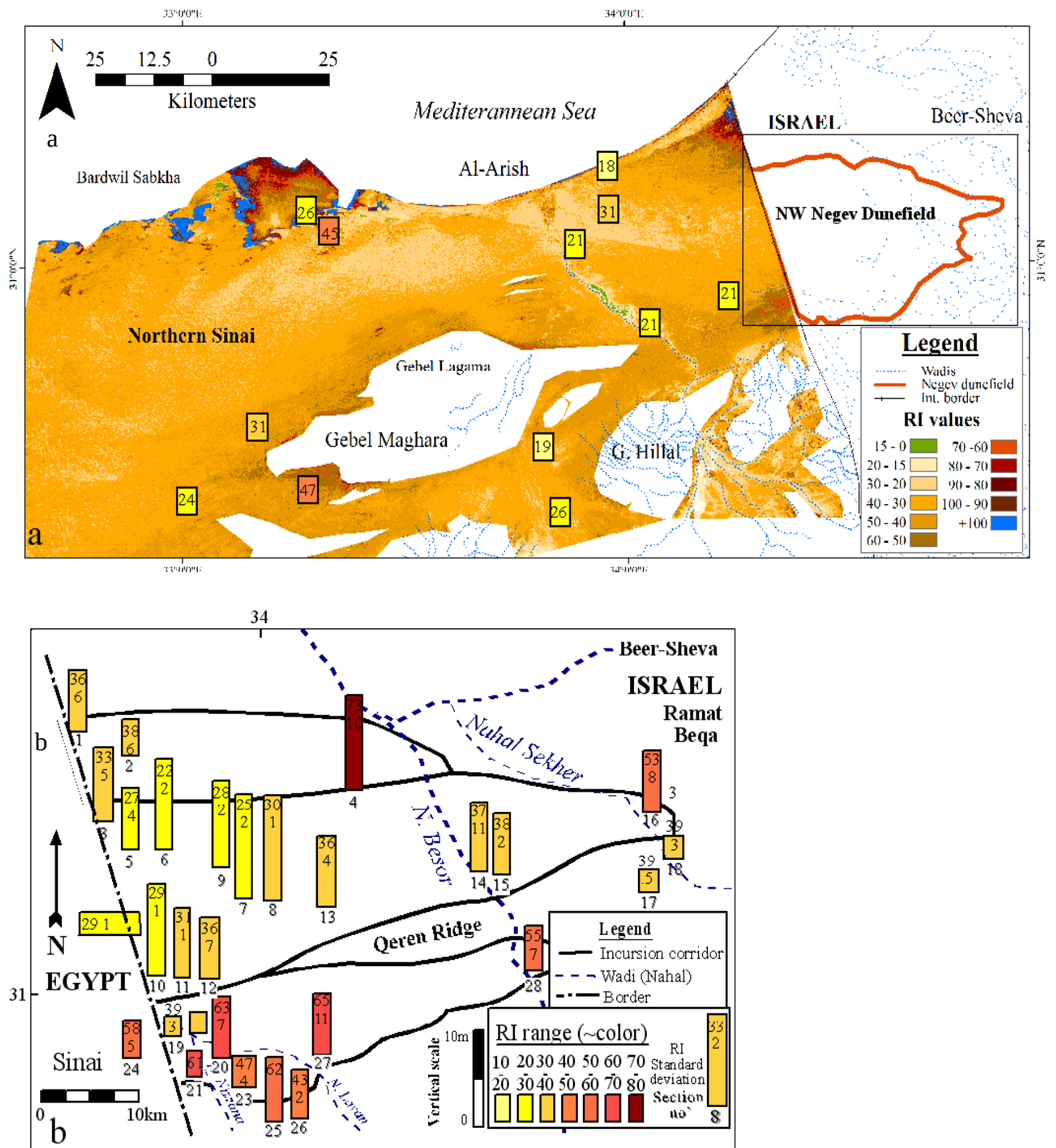


Figure 4.5. a. Remotely sensed RI values of the Sinai dunes derived from Landsat images taken in June, 1987. The colored boxes mark the location and spectrally measured RI of the Sinai sand samples. Note by the southwest corner of the Negev dunefield the redder Sinai sands close to the border.

b. Map of the encroachment corridors and average RI and standard deviation for the sampled dune and sand sections of the Negev. The Negev dunefield section numbers (in grey) correspond to Table 4.1.

The RI values of the Negev sands do not correlate with their OSL age (Fig. 4.6). In each section, OSL ages are naturally more mature with depth (Table 4.1; i.e., Fig. 4.3). 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. 4.3 and 4.4).

Spatial changes in the RI show that the southernmost encroachment corridor is significantly redder than the central corridor, which is the least red (Fig. 4.5), 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. 4.1b; section 14 in Fig. 54.b; Table 4.1), are also slightly redder (Roskin et al., 2011b). An outstanding section is Baladiya (Fig. 4.1b ; section 4 in Fig. 4.5b), in the northeast corner of the dunefield, which shows the highest RI values in the entire NW Negev dunefield (Table 4.1; Fig. 4.5).

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) (Table 4.1).

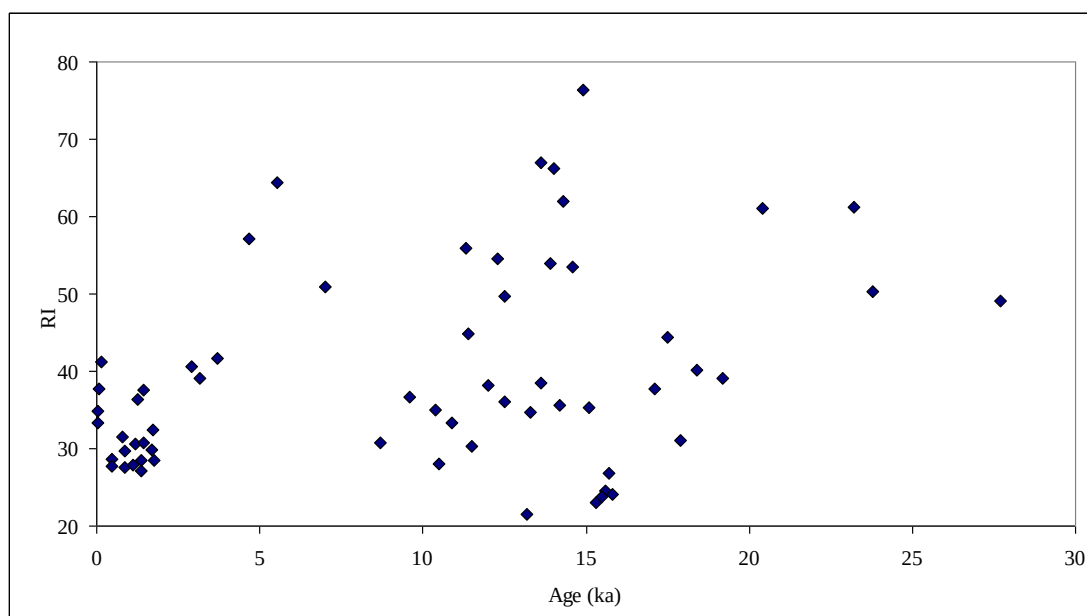


Figure 4.6. Scatter-plot showing the Negev OSL sand ages and their non-correlative RI values.

4.5.2 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 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 - 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.

4.5.3 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. 4.1b; section 6 in Fig. 4.5b; Table 4.1) 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. 4.5). 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 were also retrieved at Gebel Libni in the south and Wadi Khareidin located approximately 30 km west of the Egypt-Israel border where higher RI values are identified for the Negev and by remote sensing (Fig. 4.5). 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. 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. 4.5; Table 4.1). 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.

4.6. Discussion

4.6.1 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 et al., 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. 4.1 and 4.5), 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 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, laterite and precipitate iron on surrounding quartz grains.

Sample DF 83 of the Haluzit 1 section (Fig. 4.1b; section 3 in Fig. 4.5b; Table 4.1) of a sandy palaeosol dune substrate dated to 106 ± 19 ka (Fig. 4.3), 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 (Table 4.1). 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 (Littman, 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 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). Various color degrees 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 or contact with carbonate fines seems to suggest that the red color of the sand grains was inherited before their deposition in the Negev.

4.6.2 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. 4.1 and 4.5; Table 4.1). 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. 4.1), 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). These include the Baladiya section (Fig. 4.1b; section 4 in Fig. 4.5b; Table 4.1) 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. 4.1b; section 14 in Fig. 4.5b) 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 IRSL 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 Muleshoe 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, 2011b). These are additional indicators of the negligible changes in sand properties and sand-grain coatings during transport.

4.6.3 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. 4.5). 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 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 upwind RI similarity to remotely sensed Sinai surface sands (Fig. 4.5). 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 playas 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 (Farg, 1955), whereas Gebel Maghara's Jurassic sandstones are iron-oxide brown (Barakat, 1956). Lower Cretaceous outcrops at various northern Sinai mountain bases are yellow, pink, and purple (Farg, 1955). These descriptions are similar to the Jurassic and Lower Cretaceous sandstones sampled in the Ramon Cirque, suggesting that the sampled sands are reliable analogues.

The varying values of the Ramon Cirque sand RI are both below and above those found in the NW Negev dunefield. The Jurassic sands show two different values (Table 4.1). While the RI values of 30.8 and 47.1 are similar to the NW Negev encroachment corridor values, the RI value of 137 is the highest measured. The measured Lower Cretaceous samples are medium-size sand that are mainly attributed to a fluvial environment. Limited PSD analysis of the Lower Cretaceous from the northern Negev "Big erosional cirque" sandstone showed a polymodal distribution ranging from 70 to 370 μm (Weinberger, 1980). This differs from the Negev's fine-grained aeolian sands. The coarser grain-size makes part of these sands less prone to aeolian saltation from Gebel Maghara toward the Negev though smaller grain size fractions are possible winnowed sources. However, the grain size in the Negev's southern corridor is not relatively larger than those in the other corridors. The considerable grain-size and color differences between the Jurassic and Late Cretaceous Gebel Maghara sand grains and the erg sands make it unlikely that the former provide a significant source for the southern corridor Negev sands. Therefore, the reasons for, or origin of, the intensified redness of the southern corridor sands are probably due to their source, i.e., the Nile Delta.

4.6.4 Nile Delta sand-grain coatings

Thick, Late Pleistocene (generally > 12 radiocarbon ka BP) fluvial sand facies (Coutellier and Stanley, 1987; Stanley and Chen, 1991) from the Nile Delta have been suggested as a plausible source of the Sinai erg sands that stretch into the NW Negev (Roskin et al., 2011a). We suggest, based mainly on the extensive work of Stanley's team in the Nile Delta, that the Delta is also the site where the Sinai erg sand grains rubified. Fluvial abrasion of sand-grains down the long course of the Nile down to the Delta probably eliminated previous grain-coatings while the mineralogical suites and the coatings of the Delta sands contain supporting data for in-situ Delta rubification.

In the northeastern Nile Delta by the southern periphery of Lake Manzala, several Late Pleistocene sand sections have been found to contain opaque minerals such as magnetite in addition to amphiboles, pyroclastics, and epidote (Stanley et al., 1988). These heavy minerals may have been winnowed out during sand transport. At 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 m 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 (Frihy 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).

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 lower 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 lower Nile Delta are advancing eastward toward the Sinai (Misak 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 et al., 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 lower and central Delta may have been transported eastward into the NW Sinai. Between 30 - 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 upper 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.

4.7 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 filmy quartz grain coatings may be positively correlated to the depositional age of the sand. Based on full dune and interdune sand throughout the NW Negev dunefield, 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.

4.8. Acknowledgments

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Chapter 5: Palaeoclimate interpretations of Late Pleistocene vegetated linear dune mobilization episodes: evidence from the northwestern Negev dunefield, Israel

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5.0 Abstract

The vegetated linear dune (VLD) field of the northwestern (NW) Negev Desert, situated at the downwind eastern end of the northern Sinai - NW Negev Erg, constitutes an ideal setting for dating and interpreting its Late Quaternary dune encroachment episodes. This study builds upon the results of Roskin et al. (Age, origin and climatic controls on vegetated linear dunes in the northwestern Negev Desert (Israel), *Quaternary Science Reviews* 30 (2011), 1649-1674) that presented the stratigraphy of 35 sections and 97 optically stimulated luminescence (OSL) ages from the NW Negev dunefield. Here we refine our analysis of the Negev Late Pleistocene dune mobilizations and stabilizations and interpret their palaeoclimatic controls in light of regional and global sediment records and proxies.

While initial dune encroachment into, and stabilization in, the NW Negev took place during the Last Glacial Maximum (LGM) at ~23–18 ka, spatial and statistical analyses of the OSL dataset suggest that since the LGM, Negev dune activity was concentrated in two significant mobilization-stabilization episodes: a main episode at ~16–13.7 ka and a minor one at ~12.4–11.6 ka when the dunes reached their maximum spatial extent and stabilized. These episodes include rapid dune encroachment and accretion events and coincide with the Heinrich 1 and Younger Dryas cold events, respectively. The Late Pleistocene sand-transporting winds were characterized by a westerly direction that resulted in west-east VLD elongation.

Dune mobilizations may have occurred in response to wintertime East Mediterranean cyclonic systems that brought storms of rainfall and strong winds. The rapid dune mobilization events and their concurrence with the Heinrich 1 and Younger Dryas cold events suggest a more global control. Despite the rainfall, the elongating VLDs were probably sparsely vegetated because of the high wind power; their stabilization resulted from a decrease in storminess, with the onset of a more arid Holocene climate.

Other global low-latitude dune mobilizations and stabilizations are concentrated at the end of the Late Pleistocene, leading us to suggest that these were also controlled mainly by global cold-events and subsequent changes in windiness.

The recurring discontinuous aeolian sedimentation pattern found in OSL-dated VLDs provides new and important chronological and sedimentological insight into prominent dune mobilization and stabilization processes. The suggested link between global drops in cold-event windiness and low-latitude dune stabilization episodes emphasizes the prevalence of

winds over aridity regarding major dune mobilizations for low-latitude dunes, even if vegetated.

5.1 Introduction

5.1.1 Dunes as palaeoclimate records

The study of global palaeoclimate change has relied mainly on temperature and rainfall fluctuations as depicted in ice, marine and lacustrine cores, traditionally the proxies of choice due to their high resolution, sensitivity and continuity, and their ability to represent expansive parts of Earth. On land, speleothems have become the leading terrestrial palaeoclimate proxies as they comprise direct, specific environmental signals of in-situ rainfall (Enzel et al., 2008). They are excellent proxies of palaeo-temperatures, source of water vapor, and rainfall amounts (Bar-Matthews et al., 1999). High resolution dating of global and local windiness, however, has been less studied. Based on continuous dust records in ice, marine and lacustrine cores that show changes in global dustiness, recent studies have suggested that these changes were driven by global changes in wind gustiness (McGee et al., 2010). Terrestrial aeolian loess deposits are relatively continuous palaeoclimate proxies of glaciogenic and desert dust transport and deposition (Chen et al., 2003; Muhs et al., 2008), but their palaeoclimatic interpretations are complex due to varying grain-size distributions, mineralogy from mixed sources, varying transport distances and post-depositional processes (Kohfeld and Harrison, 2001) such as pedogenesis (Jacobs and Mason, 2007).

Dunes cover approximately one-third of the regions of Earth defined as arid (Lancaster, 2007). Geologically young landforms, dunes have been regarded as an important terrestrial source of information on palaeoclimates (e.g. Sarnthein, 1978; Lancaster, 2008). In general, active inland dunes have been used as indicators of arid conditions (e.g. Sarnthein, 1978; Munyikwa, 2005; Hesse and Simpson, 2006; Lomax et al., 2011). In accordance with the assumption that dune mobilization is induced by threshold decrease in precipitation and increased evaporation, this paradigm was shown to be true for mid-latitude dunes (Muhs and Holliday, 1995). Earlier models suggested that dunes are controlled both by precipitation and by windiness (Lancaster, 1988), but more recent models show that dune activity is primarily the result of wind power, and as such, even in humid climates with annual precipitation well-exceeding 1000 mm, exposed dunes can be mobilized (Tsoar, 2005; Chase, 2009; Tsoar et al.,

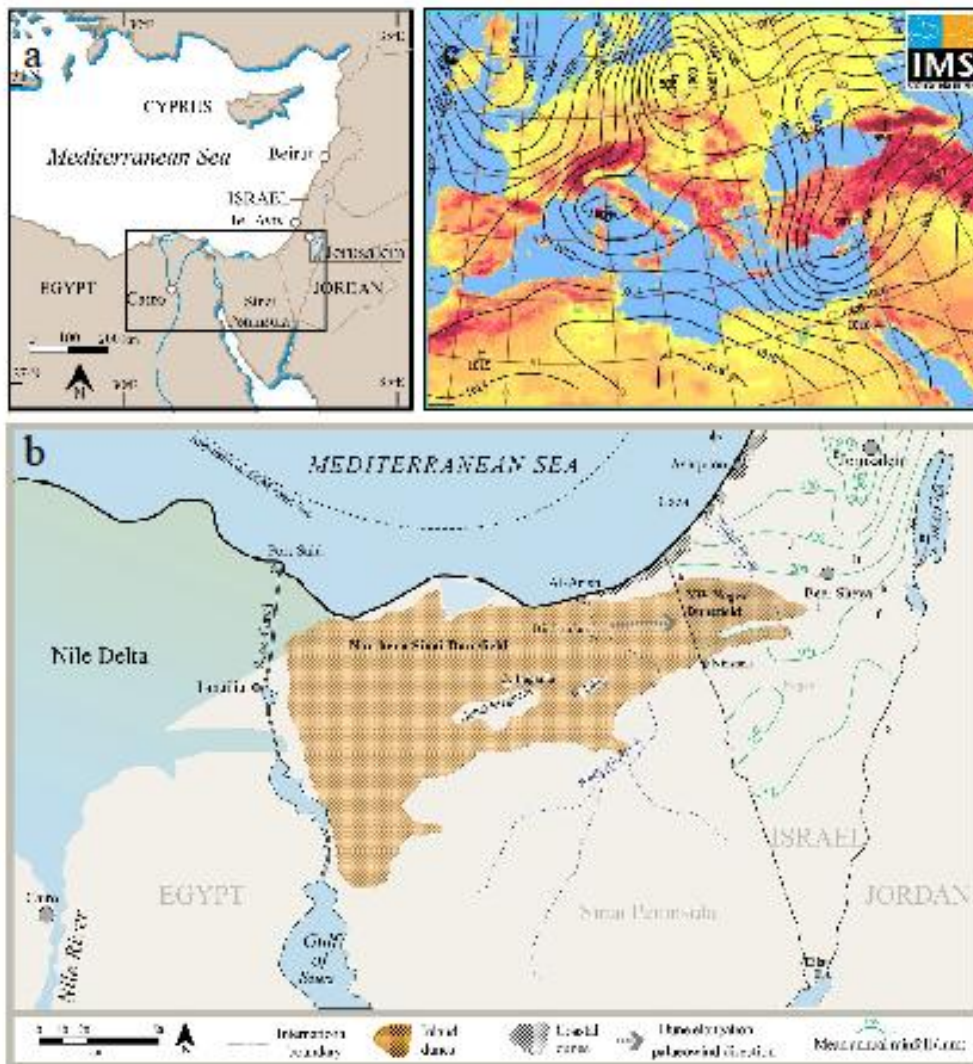


Figure 5.1. Regional map and location of regional sediment records.

a. Regional location map. Figure 1b is depicted in the black rectangle.

b. The Sinai-Negev Erg, extending from the Nile Delta eastwards into the northwestern Negev (NW). The arid Negev dunefield, situated south-southeast of the Eastern Mediterranean Sea, is currently still under the climatic control of wintertime cyclonic storms. Further south, the Negev is extremely arid as indicated by the isohyets (after Amit et al., 2006; Enzel et al., 2008). Note how only the central part or encroachment corridor (see Figure 2) of the NW Negev dunefield has up-dune continuity to the west into northern Sinai (see arrow), indicating that this part was spatially and temporally accessible to a larger sand supply. It can also explain why this part extends the furthest to the eastern limits of the dunefield south of Beer-Sheva. The locations of sediment records are as follows: a. Tzavoa Cave (Vaks et al., 2006); b. Lake Lisan (Dead Sea) (Bartov et al., 2003; Stein et al., 2010); c. Soreq Cave (Bar-Matthews et al., 1999; Bar-Matthews et al., 2003); d. Dead Sea Rift western escarpment cave speleothems (Lisker et al., 2010); e. SL112 EM core (Hamann et al., 2008), located approximately 100 km north of the edge of the map; f. Redeposited Hazeva Formation sand, (Dody et al., 2008); g. Wadi Faynan sand (McLaren et al., 2004); h. Hura loess section (Crouvi et al., 2008); i. Ramat Beqa loess section (Crouvi et al., 2008); j. Ruhama loess palaeosol section (Wieder et al., 2008); k. Qerem Shalom coastal aeolian sand and loess palaeosol section (Zilberman et al., 2007).

c. Synoptic map (Israel Meteorological Service;

http://www.ims.gov.il/IMSEng/All_tahazit/SynopticMaps.htm) of the Mediterranean region during the December 12th 2010 storm. Note the northerly track of the Cyprus Low that is centered over Cyprus and southern Turkey. This cyclonic storm delivered over 100 mm of precipitation to northern Israel while the northern Negev, in this case located south of the Low, received only strong southwestern to western winds.

2009; Yizhaq et al., 2007, 2009). Palaeoclimate studies of dune bodies, however, have failed to fully demonstrate the connection between wind power and dune activity.

Advances in optically stimulated luminescence (OSL) single aliquot regenerative-dose (SAR) protocols (Murray and Wintle, 2000) and sand drilling techniques (Munyikwa et al., 2011) increased the quality and quantity of OSL-based age estimates and enabled better chronological control of episodes of dunes and dunefield activity (e.g. Fitzsimmons et al., 2007; Telfer and Thomas, 2007). Despite the greater accuracy of OSL in dating dunes, however, the reliability of OSL ages in representing episodes of dune activity and their palaeoclimatic significance has been questioned on several grounds. OSL dating cannot pinpoint the onset of dune activation (Nanson et al., 1992; Telfer and Thomas, 2007; Fitzsimmons and Telfer, 2008), only their stabilization. Sampling does not always include a full dune section due to technical limitations (Bateman et al., 2003), and thus, sampling has often been neither systematic nor of sufficient spatial resolution (Telfer et al., 2010). Moreover, due to their dynamic and erosive characteristics, dunes constitute discontinuous records (Telfer and Thomas, 2007), and distinguishing between episodic and continual sedimentation is not always possible (Bateman et al., 2003; Telfer and Thomas, 2007; Chase, 2009).

Despite the common use of a probability density function (PDF) graph to display the distribution of the full age dataset, there is no universally accepted method to present OSL age distributions that will maximize palaeoclimatic interpretation. This ambiguity hindered the full understanding of the significance of the ages in southern Africa (Stone and Thomas, 2008). The relationship between linear dune formation dynamics, dune internal structure and the chronological and interpretation of luminescence ages for an entire dune field as a basis for palaeoclimatic reconstruction has not been fully addressed. On this line, this study will investigate the palaeoclimate of the northwestern (NW) Negev dunes (Fig. 5.1) by firstly improving the chronostratigraphic interpretation and temporal and spatial dune OSL chronologies presented by Roskin et al. (2011).

5.1.2 Episodes of northwestern Negev dunefield activity

The NW Negev dunefield, Israel (latitude 30° N, longitude 33° E) is a marginal low-latitude dunefield at the downwind end of the northern Sinai - NW Negev Erg (Sinai-Negev Erg) situated south of the Eastern Mediterranean (EM) Sea (Figs. 1a & 1b). The extensive northern

Sinai dunes upwind of the Negev dunefield hold ample sand reserves for the NW Negev, therefore, the encroachment of sand and stabilization of the Negev dunes are attributed mainly to climate change, i.e., windiness, and the processes are usually not supply-limited (Roskin et al., 2011).

The 40-km (from north-south) wide dunefield of $\sim 1,300 \text{ km}^2$ (Tsoar et al., 2008) protrudes into the northern Negev Desert and is surrounded by Late Pleistocene aeolian loess deposits (Zilberman, 1991; Crouvi et al., 2008). It is positioned along the southern part of the main wintertime Mediterranean Cyprus Low (a migratory, low altitude, cold surface low in the EM accompanied by a cold air trough in the middle latitudes) cyclone tracks (Fig. 5.1c). The dunefield receives approximately 150 mm to 80 mm annual average rainfall between October and April in the north and in the south, respectively. The annual average relative deviation of the rainfall by Nizzana in the southwestern dunefield is $\sim 40\%$ ($n=30$ yr), characteristic of arid lands (Z. Siegel pers. comm.). Thus, the dunefield is along a desert fringe of the Levant between the semi-arid and hyper arid climate belts. Droughts affect perennial shrub survivability (Siegal, 2009), which in some cases may affect dune dynamics (Tsoar, 2005; Yizhaq et al., 2009). The dunefield consists of stabilized VLDs (Tsoar and Moller, 1986; Tsoar et al., 2008) aligned in a general west-east direction and whose ground cover comprises 5–20% perennial shrubs while biogenic crusts stabilize the dune flanks and in some cases the crest (Danin et al., 1989; Tsoar et al., 2008; Siegal, 2009).

A spatially dense sampling and OSL-dating campaign from dune and interdune sections down to the dune substrate (Roskin et al., 2011) found sporadic aeolian sand deposition between 116–30 ka that had stabilized and developed into a sequence of calcic sandy to silty loamy palaeosols and that pre-date dune encroachment. Aeolian sand cover sufficient to form dunes only began to accumulate at ~ 23 ka and unconformably overlies these palaeosols, suggesting intense erosion at some localities preceding dune deposition, perhaps due to sandblasting. Following the Last Glacial Maximum (LGM; 22–18 ka), dunes covered the current boundaries of the dunefield along three west-east encroachment corridors between 18–11.5 ka (Figs. 5.2 and 5.3). This massive encroachment caused dune-damming of wadis and consequent formation of short-term standing-water bodies that deposited fine sediment units (Roskin et al., 2010b; Roskin et al., 2011). Additional remobilization episodes in the NW Negev were dated to the late Holocene (~ 2 –0.8 ka) and to modern times (150–8 yrs) (Roskin et al., 2011).

Two post-LGM peaks in dune activity were suggested for the Negev dunes, based on radiocarbon dates (calibrated after Goring-Morris et al., 2009) for cultural entities; the Middle Epipalaeolithic Mushabian and Geometric Kebaran cultures (~18–15.1 ka) and the Late Epipalaeolithic Harifian culture (~12.8–11.6 ka) (Goring-Morris and Goldberg, 1990). Roskin et al. (2011) identified a rapid encroachment event at ~16-14 ka that covered the western and central Negev dunefield. Only toward the end of this encroachment episode, dunes reached the far eastern parts of the dunefield, beyond Nahal Sekher in the central encroachment corridor and Nahal Besor in the southern corridor (Fig. 5.2). However, a distinction of two events during 18-11.5 ka is not apparent in the PDF plots of OSL ages in Roskin et al. (2011), nor have the age distributions been fully analyzed in regard to the VLD stratigraphy and dynamics.

5.1.3 Northern Negev Late Pleistocene palaeoclimate interpretation

Interpretation of the northern Negev Late Pleistocene palaeoclimate is complex, being situated along a fluctuating rainfall climate gradient. Evaluations of palaeoclimate changes relied mainly on sedimentological records that traditionally were used to examine wet-dry transitions, as summarized by Zilberman (1991). 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). On the other hand, the invasion of sand dunes from the Sinai into the NW Negev around the same time was interpreted to have occurred in arid (Magaritz and Enzel, 1990; Zilberman, 1991, 1993) and hyper-arid environments whose annual rainfall is assumed to have been less than 50 mm (Goring-Morris and Goldberg, 1990). Little attention, however, has been given to Late Pleistocene wind intensity changes in the Levant. Ben-David (2003) suggested that wind directions, despite increased Holocene aridity, have not changed since the Late Pleistocene dune encroachment. Enzel et al. (2008) proposed in a general fashion that more frequent, persistent and much intensified W-SW Late Pleistocene winds are needed to mobilize the Negev dunes. Ben David (2003) claimed that the correlation between Negev dune progression and an accurate palaeoclimatic picture is problematic. This calls on reviewing the Negev palaeoclimate by analyzing dune records with other types of sediment records and proxies from the Negev and from the same EM synoptic regime.

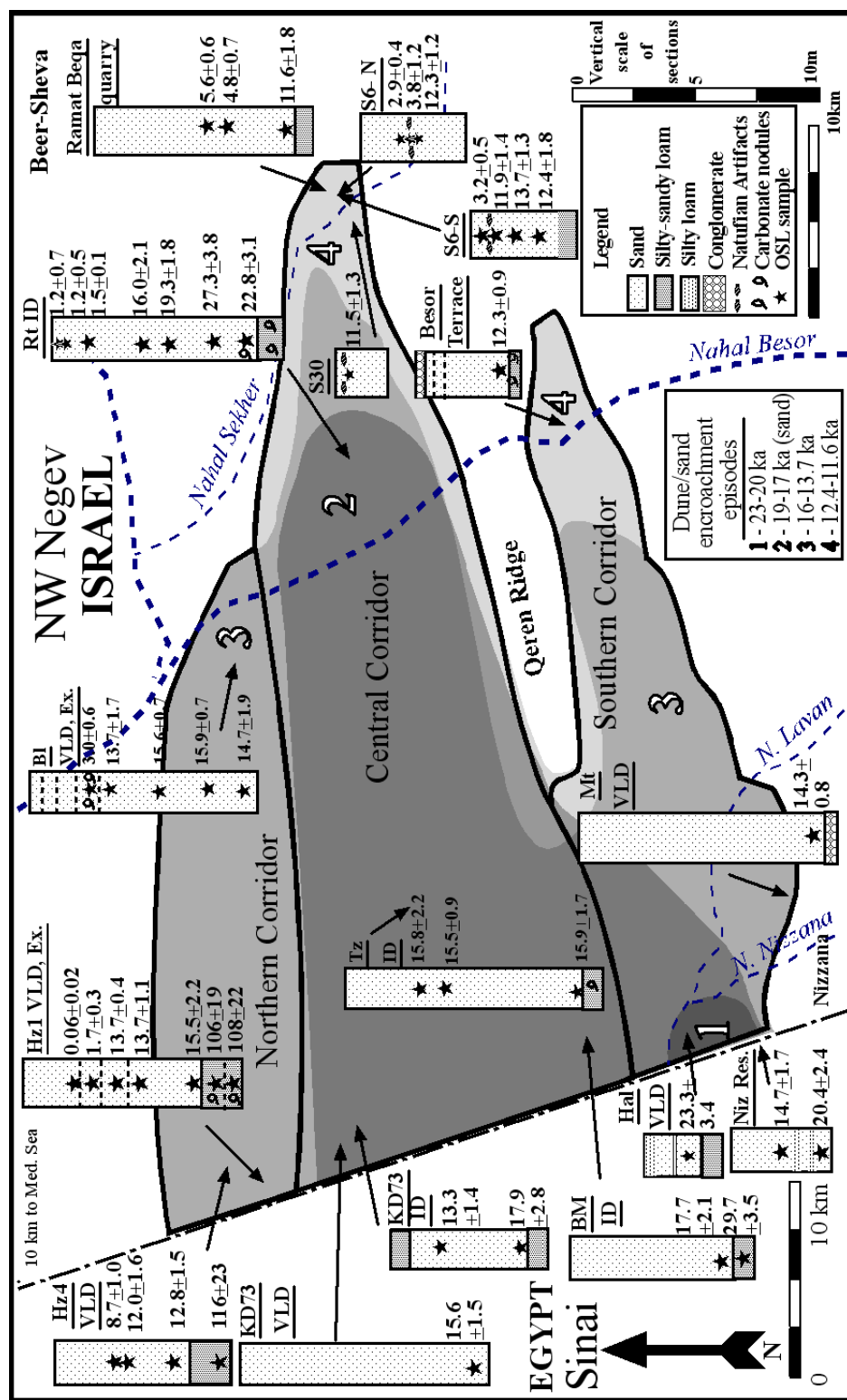


Figure 5.2. Late Pleistocene development stages of the NW Negev dunefield.

The map presents the three NW Negev dunefield encroachment corridors and the progressive stages (numbered gray shades) of dune encroachments [modified from Roskin et al. (2011)], demonstrating that the dunefield has not expanded during the Holocene. The logs display full VLD and interdune (ID) aeolian sand sections. The basal ages of the sections constitute a reliable and uniform dataset for classifying the mobilization-stabilization episodes. Note that stage 3 is not found at the eastern-most part of the central and southern corridors. These consist of only the later second episode, depicted on the map as stage 4.

Note that the northwest section HZ1-VLD and the northeast section BM-VLD have similar ages, suggesting rapid encroachment at ~15 ka. In contrast, the eastern dunefield in the other corridors was found to be younger. If this younger episode took place in the northern corridor, the sands may have been washed out by Nahal Besor.

The Late Pleistocene in the Levant is accepted to have been rainier based on high-resolution speleothem analyses and the Lake Lisan level records. Soreq Cave, located 65 km north of Beer-Sheva in the Judean Mountains (within a Mediterranean climate), currently receives an average of ~550 mm rainfall annually (Fig. 5.1). The $\delta^{18}\text{O}$ values in speleothem carbonates (whose "amount effect" interpretation has been questioned among others by Enzel et al., 2008 and Stein et al., 2010) reflect a Late Pleistocene with stable, cooler temperatures and a rainier climate in the EM region (Bar-Matthews et al., 1999; Bar-Matthews et al., 2003).

Tzavoa Cave, located ~30 km northeast of the Negev dunefield eastern edge, is at similar latitude and within a comparable isohyet range (150–160 mm) as the northern dunefield (Fig. 5.1). Based on several speleothems from arid regions of Israel, the estimated annual minimum rainfall required for speleothem formation in arid regions is double, roughly 300 mm (Vaks et al., 2006). Tzavoa speleothem growth intervals, dated by U-Th, indicate that indeed, the northern Negev Late Pleistocene annual rainfall was at least 300 mm between ~80 ka and 14–13 ka (Vaks et al., 2006). Tzavoa speleothem Sr concentrations, reflecting a relative dusty and windy environment, were generally higher during 38–14 ka (Vaks, 2008), the same time span initially suggested for the Sinai dunes (Goldberg, 1977; Roskin et al., 2011). This semi-arid Late Pleistocene northern Negev climate raises questions regarding the conditions conducive to Negev dune encroachment.

Water levels of Lake Lisan, the large, Late Pleistocene precursor of the Dead Sea (Fig. 5.1), have been accepted as a regional rain gauge (Enzel et al., 2008) and interpretations of changes in Lisan levels provided a regional palaeoclimatic synthesis with an emphasis on rainfall in the rainier Mediterranean climate zone north of the NW Negev dunefield. It may be noted that the Lisan levels were also strongly controlled by high evaporation rates. Studies have shown that at 29–25 ka, Lake Lisan rose 120 m, and after 25 ka it dropped 260 meters to the Lake's Holocene level (Bartov et al., 2003; Stein et al., 2010) suggesting a long-term though fluctuating Late Pleistocene aridification. Enzel et al. (2008) explained the formation of Negev dunes that coincided with the rise in Lake Lisan at 29–25 ka within a common general palaeoclimate scenario, whereby increased rainfall in the north is accompanied by stronger winds in the south. Could such a scenario also hold for the post-25 ka Lake Lisan decline, which coincides with the post-23 ka dune encroachments into the NW Negev (Roskin et al., 2011)?

Some fluctuations in the isotope record of Soreq Cave and the water levels of Lake Lisan coincide with the North Atlantic Heinrich 1 (H1) and Younger Dryas (YD) cold events (Bar-

Matthews et al., 1999; Stein et al., 2010). The palaeoclimate interpretations however, markedly differ: the H1 generally implies an arid environment while the YD implies a cold and rainy climate in the Mediterranean zone of the Levant (Bar-Matthews et al., 1999; Stein et al., 2010).

It has been proposed that fluctuations in the Late Pleistocene climates of the northern Negev and of central and northern Israel were controlled by winter storms generated in the EM Sea that brought rain and wind at magnitudes and frequencies higher than those of today (Enzel et al., 2008). This model was also used to explain Late Pleistocene Negev loess deposition and loess-generating Sinai-Negev dune migration that is suggested to have begun before 100 ka (Crouvi et al., 2008; Enzel et al., 2008; Crouvi et al., 2010; Amit et al., 2011). According to this model (Enzel et al., 2008), the Mediterranean Sea, as today, was the main control of Late Pleistocene rainfall in the Mediterranean climatic zone east of the EM and in the northern Negev. The model positively correlates the frequent and intense southerly Cyprus Low EM tracks to more frequent rainfall and windiness, as observed today in Beer Sheva (Dayan et al., 2008), which subsequently led to greater dust deposition. The southern Negev, over 50 km south of the NW Negev dunefield, is suggested to have been under a stable, extremely arid climate regime throughout the Late Pleistocene (Amit et al., 2006; Enzel et al., 2008). This implies that the general regional EM palaeoclimatic configuration proposed to have been over the northern Negev has also not changed drastically since the Late Pleistocene (Enzel et al., 2008). During the LGM, however, sea-level dropped by ~100–130 m and the Sinai coastline receded northward some 50 km (Goring-Morris and Goldberg, 1990; Butzer, 1997; Enzel et al., 2008) (Fig. 5.1a). This is suggested to have caused increased aridity and desertification in the northern Negev, although the dunefield still received cyclonic winds (Enzel et al., 2008).

According to Enzel et al. (2008), throughout the Late Pleistocene, aeolian sand, along with winter cyclonic rainfall, was transported toward the Negev by strong southwesterly and westerly winds. As the strongest EM winds are generated by a deep and cold Cyprus Low (Enzel et al., 2008), this can suggest that palaeoclimatic evidence of increased regional EM rainfall can correspond to increased regional windiness. Therefore, rainfall amounts interpreted from speleothem records in the dunefield region and further north that are derived from the EM cyclonic storm tracks may be a general proxy for past windiness, an assumption that will be further tested in this paper. Interpretations of the detailed chronologies and structure of the NW Negev VLDs are proposed to investigate the relevance of the EM palaeoclimate model and global climate change to Negev dune mobilization episodes.

5.1.4 LGM – Holocene transition climate changes

Windiness fluctuations during the LGM – Holocene transitional period were derived by drastic climate changes (Oldfield, 2005; Stein et al., 2010). A strong thermal gradient between the Late Pleistocene glacial cover and the equator has been suggested as the reason for a turbulent LGM atmosphere, which decreased at the beginning of the Holocene (Ditlevsen et al., 1996; McGee et al., 2010). This gradient inflicted windiness and consequently increased dust emissions in both hemispheres that subsided at differing rates following the LGM (McGee et al., 2010). However, the connection between global post-LGM - Holocene dune mobilization and stabilization and dust proxies, and this reduction in atmospheric turbulence, has not been established (Munyikwa, 2005; Chase, 2009).

Indeed, global environmental changes (and their consequences) from the LGM until the onset of the Holocene (post-LGM – Holocene transition) were extraordinary in every respect. These oscillations, such as the North Atlantic cold events, were most prominent in the mid to high latitudes (Shakun and Carlson, 2010). As such, the imprint of these events is difficult to detect at lower latitudes (Stein et al., 2010) though Roberts et al. (2008) identified isotopic shifts specifically in lakes around the entire Mediterranean basin during the H1 and YD, best explained by regional aridity. Based on PDFs of luminescence ages (>50), here replotted into histograms (Fig. 5.3), low-latitude dunes indicate that there was substantial dune mobilization since the LGM that stabilized by the Holocene in the Strzelecki Desert in Australia (Fitzsimmons et al., 2007), the Southern Kalahari and southern Africa (Chase and Thomas, 2006; Telfer and Thomas, 2007; Telfer, 2011), and the NW Negev (Roskin et al., 2011). This apparent post-LGM global synchronicity of dune ages (Roskin et al., 2011) has remained unexplained in a global palaeoclimatic context.

The relationship between global climate change in terms of windiness, dustiness, and dune activity has profound palaeoclimate and future climate implications. Evidence that aeolian dust input was twenty-five to thirty-fold higher in the polar regions during times of peak glaciations than during times of peak interglaciations (Broecker, 2000; Lambert et al., 2008) dictates that we should improve our understanding of this aspect of climate change (Broecker, 2000). The coincidence of similar dust-flux changes at the high and low latitudes of both hemispheres has been suggested to be the result of dust-driving global gustiness, i.e., relatively rare high-speed wind events (Winckler et al., 2008; McGee et al., 2010). Whether

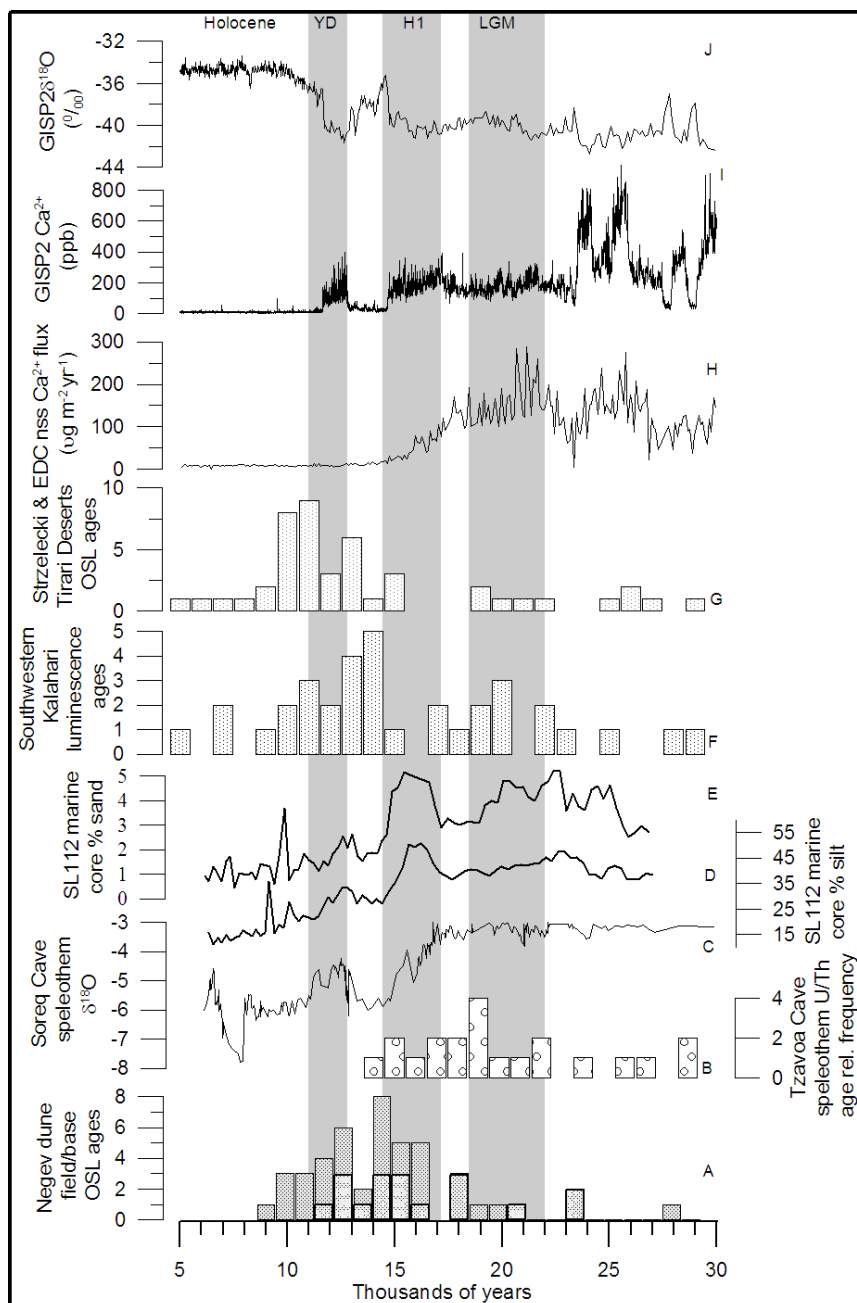


Figure 5.3. Late Pleistocene Negev and global dunefield age histograms and regional and global palaeoclimate records.

Stacked graphs comparing the NW Negev dunefield OSL age distribution with regional and global proxies. Light grey bars mark cold events corresponding to the Last Glacial Maximum (LGM), Heinrich 1 (H1) and Younger Dryas (YD).

Graphs from bottom: **(A)** Histogram of the NW Negev dunes OSL ages. The dark-lined bins with lighter insets show the ages for the dune bases. **(B)** Histogram of the Tzavoa cave speleothem Th-U ages. Higher age frequency indicates higher annual rainfall (after Vaks et al., 2006). **(C)** Plot of $\delta^{18}\text{O}$ values from the Soreq cave speleothem dated by Th-U (Bar-Matthews et al., 2003). **(D & E)** Records from the GeoTu SL112 marine sediment core west of Haifa, Israel (Hamann et al., 2008) showing the percents of aeolian silt and fine sand, respectively, based on end-member analysis. The sharp and short early Holocene peaks of silt and sand are attributed by Hamann et al. (2008) to a Nilotic source and not to a regional aeolian event. **(F)** Histogram of compiled southwest Kalahari linear dune luminescence ages (after Telfer and Thomas, 2007). **(G)** Histogram of Australian linear dunes (after Fitzsimmons et al., 2007). **(H)** Non sea-salt Ca^{2+} fluxes (Rothlisberger et al., 2008) from EPICA DOME C (EDC) in Antarctica. **(I)** Ca^{2+} (Mayewski et al., 1997) from Greenland's GISP2. **(J)** $\delta^{18}\text{O}$ from Greenland's GISP2 (Grootes et al., 1993).

terrestrial proxies are suitable for detecting these rapid changes, however, has been questioned (Allen et al., 1999). ,

5.1.5 Study goals

Using improved modeling of VLD elongation and accretion dynamics to refine the OSL chronological framework of the dunefield, this study aims to interpret the environmental and palaeoclimatic controls on the Late Pleistocene dune mobilization episodes in the NW Negev. We hypothesize that VLD mobilization and stabilization were in response to global changes in windiness. This study is based on 97 OSL ages and stratigraphy from 35 exposed or drilled dune sections of the NW Negev dunefield presented in Roskin et al. (2011), of which 15 were dated down to the base of the dune section. The resulting dataset for the ages of the dune base should resolve previous criticism of dune luminescence ages and PDFs (Telfer and Thomas, 2007; Stone and Thomas, 2008; Telfer et al., 2010) by creating a consistent and reliable spatial chronology of the early dune encroachment episodes.

The Late Pleistocene dune ages cited in Roskin et al. (2011) are analyzed spatially and statistically by running Mann-Whitney tests to investigate age grouping. The resulting dune mobilization-stabilization episodes are interpreted by comparison with regional dated aeolian and lacustrine sediments, speleothems and high latitude ice-core dust fluxes and isotope records. We then discuss the relationship between the Negev and global luminescence-dated dune mobilization and stabilization episodes that occurred from the LGM until the Holocene in the contexts of global and rapid climate changes, dust-driving windiness and gustiness (McGee et al., 2010).

5.2 Northwestern Negev dune encroachment episodes

Spatial analyses of the sand and dune sections of the main NW Negev dune encroachment, dated to ~18–11.5 ka (Roskin et al., 2011) divide the dunefield into two key Late Pleistocene dune encroachment episodes, at ~16–13.7 ka and at ~12.4–11.6 ka (Figs. 3, 4 & 5). The first episode covered the northern and western parts of the dunefield and includes an initial stage of basal sand cover that occurred at ~18–16 ka. The second episode ages are identified mainly in the eastern part of the dunefield (Fig. 5.2).

To clarify the spatial distinction between the two age groups, we also statistically analyzed the grouping using the Mann-Whitney test (Fig. 5.4). This test was chosen because it is a non-parametric statistical hypothesis test that compares two unpaired groups. The statistical p value answers the question: if the populations have the same median, what is the chance that random sampling would result in medians as far apart as observed. Two dune encroachment episode groups of western dunefield ages of ~18–13.7 ka and eastern ages of ~12.4–11.6 ka were analyzed from two OSL datasets; the entire dataset, and the dune base dataset that uniformly represents initial dune encroachment ages, despite its limited number of ages ($n=17$). The results indicate that the two episode groups are distinct. The Mann-Whitney value for the full dataset was 0.00 and $p<0.001$, while that for the dune base dataset was 0.00 and $p<0.01$. The low p value indicates that a significant difference exists between the two episode-representing groups.

During the earlier dune encroachment episode (~18–13.7 ka), sand sections of 2–10 m thick accumulated throughout the western part of the dunefield, mainly from ~16 ka (Fig. 5.2) onwards. To the east, dune and sand thickness decreases as evident in the Rt ID section where only 1–2 meters of sand accumulated. During this episode, sand was transported over ~85% of the dunefield. The substantial thickness and spatial cover of the sand led us to define the first episode of the main encroachment as the *main episode*. Its time span is similar to that of the H1 cold event (Fig. 5.3).

During the *second episode* (~12.4–11.6 ka), the dunes and sands elongated and stabilized in their easternmost configuration (Fig. 5.2). The dunes dating to this episode have indistinct undulating morphologies and are usually no more than 5 m thick. This episode exhibits less intense accretion, lower sand transport distances, and a shorter duration relative to the main episode. The time span of this episode is similar to that of the YD cold event (Fig. 5.3). The sands probably originated from dunes deposited during the main episode in the western dunefield. A single 13.7 ± 1.3 ka age in the Sekher 6-S section (Fig. 5.2) may indicate small sand quantities already deposited in the east at the end of the main episode. The Hz4 dune section in the west with ages of 12.8 ± 1.5 ka and 12.0 ± 1.6 ka may be an example of a western dune being reworked during the second episode. This finding is also supported by the similar redness intensity of the sand grains along the transport paths, regardless of grain ages (Roskin et al., 2010a) and similar grain size distributions (Roskin et al., 2011). Aside the Hz4 section, the paucity of units dated to ~12.4–11.6 ka in the west is probably due to later erosion of this

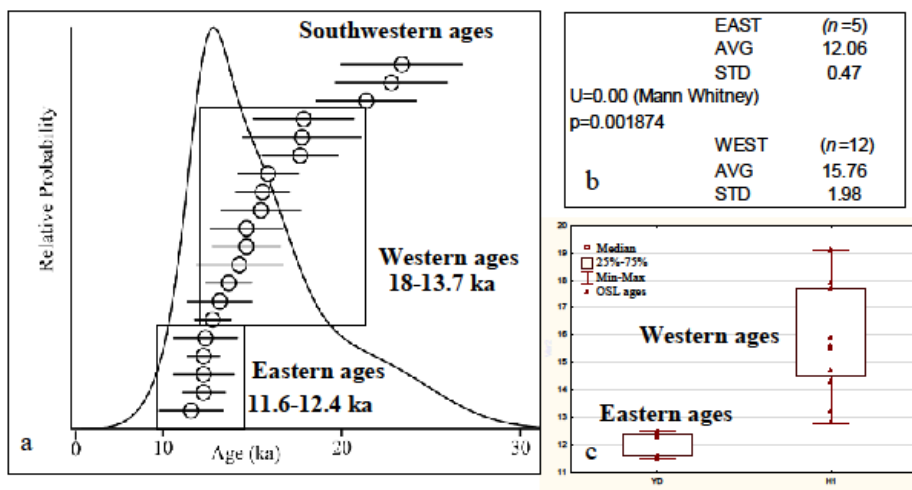


Figure 5.4. Analysis of the two NW Negev encroachment episodes.

a. A probability density function graph of the NW Negev dune base ages (data after Roskin et al., 2011) with rectangles delimiting the west and east dunefield age groups. Ages that were found only in the southwest corner of the dunefield were not included. b. Mann Whitney test data and results. c. Box and whisker display of the two dune encroachment groups.

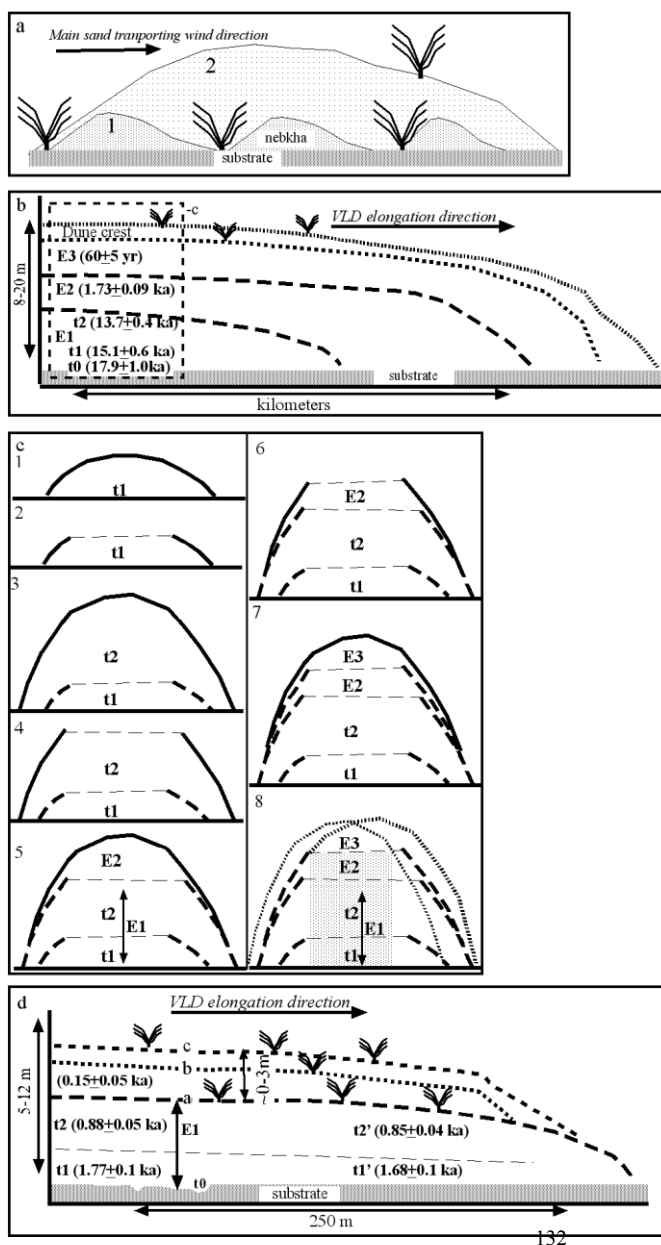


Figure 5.5 (previous page). VLD formation and stratigraphy.

a. Incipient VLD formation scheme. During a high wind-speed episode, a VLD forms/elongates via the formation of a nebkha that develops into a dune on its lee side. More shrubs clutch to the lee dune to form another nebkha, which continues the dune on its lee side (1), and so on and so forth. This process continues until the sand supply is exhausted or the wind speed relaxes below sand transport thresholds. This culminates in a VLD mobilization-stabilization episode – E1 (2).

b. The VLD internal structure along its axis. The dashed box depicts what has been observed in an exposed OSL-dated (cross) section (Hz1) perpendicular to a VLD axis (Roskin et al., 2011) (for location, see Fig. 2). E1-E3 units represent dune formation episodes that include accretion and elongation. t0 marks sand pockets from incipient sand or dune sand sedimentation that was not eroded by later mobilizations as found south of the Hz1 section. t1 and t2 mark samples taken from the base and top of the main dune mobilization-stabilization episode unit. The units may only be identifiable in an exposure by an unconformity based on bedding or by relict features from vegetated surfaces (organic material, snail-shells) that may represent a certain time interval when vegetation was sustained and its detritus was possibly slightly reworked and re-deposited. The VLD crest (dotted lines) undergoes intermittent activation from crosswinds that slightly erode and redeposit sand.

c. The VLD cross-section. Fig. 5c-7 portrays the axis section (rasterized) that is preserved while the dotted lines (c-8) represent the changing dune crest configurations due to crosswinds. Note how the active crest does not fully erode the main episodic units delimited by horizontal dashed lines.

d. This figure portrays dune elongation and accretion ages from the drilled late Holocene BM section (ages after Roskin et al., 2011), which is suggested to have a local sand source and comprises a field laboratory for understanding VLD formation. t1 marks the lower part of the episodic E1 unit and, based on OSL ages, it dates to the age of the primary nebkha and the lee dune deposition. This is followed by continued mobilization of sand that is deposited beyond t1', forming an initial elongated VLD structure. The time elapsed between t1 and t1' that indicates sand transport/dune elongation rate/time is regarded here as an "event" of dune accretion/elongation. The t2-t2' segment date an additional event from episode E1. Its age is naturally younger than the t1–t1' event. Thus, the t1 and t2 event ages represent separate VLD accretion and elongation events separated by ~850 years. When scaled to the Late Pleistocene Negev ages, both events can exemplify how the main encroachment episode (16-13.7 ka) section (Fig. 5b) accumulated.

Dashed-line b represents the truncated t2 unit top. Line c represents the current dune surface, reactivated only several years ago. Unit a-b represents local dune activation that has not been reworked so deeply since 150 a.&

relatively thin, episodic unit during the late Holocene remobilization, a process that is explained in the following section.

5.3 Negev vegetated linear dune dynamics, structure and chronology

5.3.1 Vegetated linear dune formation

Large proportions of the low to mid-latitude dune bodies in Australia (Nanson et al., 1992; Fitzsimmons et al., 2007; Cohen et al., 2010), southern Africa (Telfer and Thomas, 2007), and South America (Tripaldi and Forman, 2007; Tripaldi et al., 2011) comprise vegetated linear dunes (VLDs) that are currently stable in regions with low wind power. Accordingly, in past climates and environments, the dunes, either with or without vegetative cover, were active until stabilizing at their current position. Accordingly, a better understanding of the processes of VLD formation and elongation, whether the VLD is partly or fully vegetated, is needed. Hollands et al. (2006) suggested that linear dunes in the northwestern Simpson Desert, Australia mainly accrete by wind rifting and are not elongating. Recently, Telfer (2011), by densely OSL-dating full dune-sections spaced along the 600 m elongating tip of a simple linear dune in the south-western Kalahari has clearly demonstrated episodic sand accretion and dune elongation.

Vegetation cover is assumed to be the main reason behind VLD formation, for which several theories have been proposed. In contrast to the sinuous elongation that occurs with linear, unvegetated seif dunes, VLDs are thought to lengthen along straight lines and approximately in the direction of the prevailing wind (Tsoar and Moller, 1986; Tsoar, 1989; Tsoar et al., 2008). One simple explanation perceives VLDs as seif dunes that formed during the Late Pleistocene and that were subsequently stabilized during the Holocene, when the climate became more humid and less windy (Lancaster, 1994; Lancaster, 1995). But this approach cannot explain how the tuning fork pattern (Tsoar et al., 2008) or Y-junctions (Kocurek and Ewing, 2005), common in VLDs but missing from seifs (Tsoar and Moller, 1986; Tsoar et al., 2008), is formed. This coalescence, though not clearly understood, has been attributed to deflection by cross-winds of the extreme of the dune ridge during the elongation process in order to preserve dune spacing (i.e. Tsoar et al., 2008 and references within). It seems, therefore, that VLDs have always been vegetated to some degree, though probably more sparsely during colder periods when wind power was greater (Hesse and Simpson, 2006; Hollands et al., 2006; Cohen et al., 2010).

We propose that VLDs were formed under conditions of high wind power that prevailed during the Late Pleistocene but that are virtually unknown today. The transport of sufficient amounts of sand by high wind power is understood to lead to the formation of large nebkhas behind shrubs, followed by the development of lee (shadow) dunes behind each nebkha. The lee dune then connects to the nebkha downwind of it to form an incipient, elongated VLD (after Tsoar, 1989) (Fig. 5a). Consequently, parallel to the incipient VLDs, swales develop into interdune corridors where sand deposition is limited (Allgaier, 2008a).

The vertical accretion of sediment, in part by the coalescence of new nebkhas alongside and upon the VLD axis, is thought to occur along with dune elongation (Fig. 5.5). But strong cross winds can erode the upper dune surface, causing the depth of erosion and sedimentation on the dune crest and slopes to vary (Fig. 5.5b). In the southwest (SW) part of the Negev dunefield north of Nizzana (Figs. 1 and 2), annual summer sand erosion and winter storm sand deposition were found to amount and affect the upper 25 cm of the dune crest (Allgaier, 2008b). This explains how perennial shrubs on VLDs gradually adapt themselves to winds and consequent sedimentation and/or erosion. If the dune surface is held quasi-active, it will also not undergo pedogenesis, as described for the SW Negev dunefield (Blume et al., 1995) and observed by Roskin et al. (2011) for a majority of the dunefield.

5.3.2 Negev VLD mobilization-stabilization episodes

The densely dated NW Negev VLDs that include fully exposed and dated dune sections (Hz1, Fig. 5.2) helps clarify and model the episodic chronology of VLD mobilization. The model explains the significance of VLD OSL ages and VLD development and is further based on GPR profiles (Tsoar et al., 2010) and aided by concepts from papers by Kocurek (1998), Allgaier (2008a and b), Stone and Thomas (2008) and Tsoar et al. (2008).

Interpreting dune chronostratigraphy requires a familiarization with the terms linear dune activation, mobilization, elongation, and stabilization. A linear dune can be defined as *active* when it undergoes erosion and the deposition of sand while dune elongation is negligible. Negev dune activation usually entails crosswind-driven to and fro movements of the crest amounting to a thickness of several meters (Allgaier 2008b; Roskin et al, 2011) as discussed by Kocurek, (1998) and Telfer et al. (2010) and demonstrated by Telfer (2011). The luminescence signal of the upper dune is reset (Fig. 5.5) but this does not involve substantial

dune elongation. In climates with less than 80 mm annual precipitation, dune crests can be active at low wind powers (Tsoar, 2008).

VLD elongation, however, involves the transport of sand along the dune axis and its deposition at the dune nose, where the lengthening occurs. During elongation, sand is highly mobilized as it accretes on the dune axis and partially on the dune slopes. Sediment supply is a prerequisite for VLD elongation (mobilization) and subsequent dune encroachment, while dune (sand) erosivity and erodibility (Chase and Brewer, 2009) control local dune activation. Dune stabilization is a passive situation recognized as the absence of either dune elongation or limited lateral crest movement (activation). Vegetation and a biogenic crust cover usually develop on stabilized dunes (Tsoar et al., 2008).

Previously dune time-series have generally been interpreted in terms of "dunefield activity" (Fitzsimmons et al., 2007) or "dune accumulation" (Stone and Thomas, 2008). Both of these interpretations do not differentiate between mobilization or stabilization, since no account can be taken of erosive periods that are absent in the dune record. Because OSL ages represent the burial age of the sample, the specific location and OSL age represent the end of accretion/sedimentation (i.e., stabilization) of a mobilized VLD. A series of OSL ages along a section located in the dune axis (Fig. 5.5b; t1, t2), provide burial ages of sand during an accretive episode that occurred while the dunes were continuously mobilized. A basal age (Fig. 5.5b; t1) may generally date the initiation of sand accretion, while the upper age (Fig. 5.5b; t2) marks a later time of deposition immediately prior to stabilization. A t2 sample can also mark a later event of dune crest reworking. If this pattern is spatially repetitive throughout the dunefield, as found in the NW Negev (Fig. 5.5c), OSL age clustering, mainly between these bottom (t1) and top (t2) "end" event ages, can be perceived as representing what we define as a dune *mobilization-stabilization* episode. As concluded by Telfer et al. (2010), periods of enhanced activity are readily preserved.

The luminescence age taken at t1 specifically marks the burial time of sand at t1. However, the age can be interpreted to generally date (within errors) the range of the t1–t1' event that includes burial during mobilization (t1) and, due to elongation, burial shortly thereafter at t1'. As this mobilization event ended somewhere further down the dune in space and time, we can assume that both the t1 and t1' ages, taken along a partial segment of a dune axis, are somewhat older than the end (stabilization) event (t2) of the mobilization episode that probably occurred with maximum VLD elongation.

On a single VLD scale, this pattern was identified for Late Holocene dune movement at the BM site in the southwestern part of the Negev dunefield when dating two VLD axis sections 260 m apart (Fig. 5.5d). The OSL ages, recalculated using the Central Age Model (CAM) show that the t1 and t1' ages are 1.77 ± 0.1 ka and 1.68 ± 0.1 ka and the t2 and t2' ages are 0.88 ± 0.05 ka and 0.85 ± 0.04 ka, respectively. The ages, though within errors, appear to be slightly younger downwind and the two age sets mark two distinct and rapid mobilization and elongation episodes. The t2-t2' elongation, occurred in a maximum time-span of 130 years (including errors) so that dune elongation was surely greater than 2m/a. This pattern can be observed along the densely-dated linear dune in the south-western Kalahari for Late Pleistocene and Holocene ages (Telfer, 2011).

On a Late Pleistocene scale and at a dune encroachment corridor, the t1-t1' age differences are larger. The Hz1 section basal t1 age of 15.5 ± 2.2 ka is similar to the basal t1' ages 25 km downwind at section Bl (15.9 ± 0.7 ka and 14.7 ± 1.9 ka) (Fig. 5.2). These units are part of one main *mobilization-stabilization* episode. The upper t2 type ages, such as 13.7 ± 1.1 ka and 13.7 ± 0.4 ka in the Hz1 section and 13.7 ± 1.7 ka in the Bl section, date the stabilization. However, when sampling is based on drilling at 1-2 m intervals, both t1 and t2 ages may not be identified. In such cases and when additional chronological indicators are absent, any of the four (t1–t2') event ages can be taken to generally represent the whole E1 episode, and in the case of the Hz1 and Bl section, therefore, would fall into the span of the main episode, dated to ~16–13.7 ka.

To varying degrees, active dune surfaces can slightly truncate the depositional surface of the previous episode (Fig. 5.5b). Thin mobilization-stabilization event units may even be fully reworked by intermittent and possibly long-term local surficial activity, which can erase them from the chronostratigraphic record. For example, in the Hz1 section the second mobilization episode (12.4–11.6 ka) is not always apparent. Remnants of a minor dune *mobilization-stabilization* episode may be evident in sporadic ages of ~3 ka, found for example at the S6 section (2.9 ± 0.1 ka) and at the upper Bl dune in the east (3.0 ± 0.6 ka) (Fig. 5.2). Basal charcoal layers overlying pedogenic sand in a quarry at the southeast dunefield were dated by radiocarbon to ~3 and ~2 ka (Zilberman, 1991). At Nahal Lavan dune-dammed sediments began to accumulate by ~3.5 ka (Ben-David, 2003). Most likely this short episode was decimated in most places by the late Holocene VLD remobilization and dune crest activity.

The outcome of such episodic erosion and deposition is that initial sand arrival ages will be lacking and stabilization ages may be either post-dated by later dune crest activation or

truncated by later dune activity or mobilization. Accordingly, the OSL age clusters from the Negev dunes presented by Roskin et al. (2011) as PDF and histogram (Fig. 5.3) reliably represent the main VLD mobilization-stabilization episodes, with the older ages within each cluster usually representing initial encroachment and mobilization and the younger ages represent the time of dune stabilization. This may also explain the coincidence of the Negev age groups with the H1 and YD cold events, where the older sediments were buried and preserved due to increased windiness, and the younger ages mark stabilization due to drops in windiness at the end of the cold-events.

5.3.3 Rapid accretion and elongation

The thick aeolian sand and dune sections in the western Negev dunefield record a short-term, high wind-power event that is suggested to have formed the majority of the sand record of the Late Pleistocene main mobilization-stabilization episode (Fig. 5.2; see sections Hz1 VLD, Bl VLD, KD 73, Tz ID) (Roskin et al., 2011). Based on the 16-15 ka ages of the Tz and Bl sections, it seems that around ~15.8 ka there was a rapid event (Roskin et al., 2011). The nearly identical ages of ~13.7 ka for the end of the main mobilization-stabilization event at different parts of the dunefield are striking and indicate a joint abrupt stabilization, probably due to a sharp drop in wind power. Furthermore, the late Holocene section at BM-VLD, which formed under a different environmental regime (Roskin et al., 2011) (Fig. 5.5d), also suggest rapid mobilization and accretion, such that two mobilization-stabilization episodes (at ~1.7 ka and at ~0.85 ka), less than 1000 years apart, caused VLD accretions of 4–5 meters.

Varying sand thicknesses for the same mobilization-stabilization episode between the different encroachment corridors (in the west; Fig. 5.2) can be explained by differences in wind power, erosion or sediment supply. Sediment supply is probably the dominant factor: the central encroachment corridor, with the thickest sand accumulation and the easternmost extent, is connected directly upwind to the substantial dune field in Sinai with ample sand supply, whereas the northern and southern corridors, which are less extensive, partially continue into Sinai (Fig. 5.1). Wind speed and direction of the prevailing sand transporting storms were probably similar throughout the northern and central dunefield; the quite uniform west-east VLD orientations and OSL ages support this. Additionally, the regional extent of cyclonic wind properties can be inferred from current EM storms. The December 10–12th 2010 cyclonic storm of a northern EM Cyprus Low track led to over 100 mm of precipitation

in northern Israel (Fig. 5.1c). Throughout the Negev, far beyond the boundaries of the dunefield, strong and quite uniform, mainly southwesterly and also westerly winds persisted.

We suggest that VLD buildup was a rapid process. If dune accretion was a gradual sedimentological process, each small-scale sand deposition would be either eroded or reset luminescence-wise during dune activity, or prone to bioturbation during dune stabilization (Bateman et al., 2007). Assuming constant mobilization rates, the rapid TZ section dune accretion of 10 m in several hundreds of years reflects an average accretion rate of several cm/a while Southern Australia linear dunes were found to slowly accrete at only 1 cm/100a (Lomax et al., 2011). These accretion rates do not seem sustainable. Accordingly, the VLD mobilization-stabilization sections do not necessarily result from a gradual sedimentation process, but are rather sporadic, intense responses to extreme windiness events. This scenario emphasizes the non-linear dune-scape response to forcing events (Telfer et al., 2010). Most likely, between the two dune-building events, despite the Late Pleistocene climate suggested to have been stormier and rainier, the Negev dunefield was not fully mobilized.

Elongation rates of a single VLD such as for fresher late Holocene ages at the BM site (Fig. 5.5d), can support our assumption of rapid accretion during short mobilization-stabilization episodes. Based on the Late Pleistocene ages, we estimated that for the second episode at ~12.4–11.6 ka, average sand transport rates along the VLD elongation direction were approximately several m/a. An elongation rate of 20–30 m/a along the northern encroachment corridor is feasible (Fig. 5.2) for the main episode, which lends further support to the intensity of this episode. These dune elongation rates, however, are perceived as gross averages.

5.3.4 The change in Negev sand-transporting wind orientations since the Late Pleistocene

The west-east orientation of the Negev VLDs dated to the Late Pleistocene suggests a uniform west-east dune-elongation palaeo-wind direction that slightly differs from the current main sand-transporting wind directions. Since their formation, the dunes have not undergone noticeable lateral movement (Ben-David, 2003; Roskin et al., 2011). Late Holocene transverse dunes in the central encroachment corridor with eastern-facing slip-faces (Roskin et al., 2011) also imply westerly winds needed for dune encroachment.

Linear dunes in eastern Sinai are devoid of vegetation because of grazing and wood gathering activities by Bedouin nomads since the 18th–19th centuries up to recent times (Tsoar

1995) (Fig. 5.6a). As a result, small transverse dunelets were formed on the VLDs, as well as partially in the NW Negev dunes that were periodically exposed to grazing (Fig. 5.6c) (Tsoar et al., 2008). These dunelets were formed by the strongest current southwest winds in the area (Figs. 6b & d). Dune crests and slopes that date back 150 years (Roskin et al., 2011) support this scenario. In the same manner, when dunes formed during the Late Pleistocene, the strongest dominant wind was from the west and not the southwest as it is today.

Unpublished wind data from the Shivta Meteorological Site, ~8 km south of the dunefield, shows that the southwestern winds usually exceed the sand transport threshold and comprise 42% of all of the winds (at various categories of durations) (after Enzel et al., 2008). Summer northwesterly winds are less intense but more frequent (Fig. 5.6b & 5.6d). Winds typically encountered today, however, do not cause substantial VLD elongation nor do they rework the dune crest beneath depths of ~3 m. These observations indicate that the Late Pleistocene sand-transporting winds were not only more vigorous, but they also consistently blew from the west. This hints that the Late Pleistocene dune-transporting winds were possibly part of a different synoptic regime than the winds of today.

5.4. Regional palaeoclimate records

Palaeoclimatic interpretations of dune ages in themselves is often insufficient and therefore must be compared with and complemented by additional proxy data (Fitzsimmons et al., 2007; Lomax et al., 2011). To place the Late Pleistocene Negev dune mobilization-stabilization episodes within a regional context, we assembled a large set of regional Late Pleistocene palaeoclimate sediment records, mainly from Israel and within 100-km of the NW Negev dunefield. These include rainfall and isotope records from cave speleothems, Lake Lisan level records, and aeolian loess and sand records. We also addressed the timing of prehistoric findings in the NW Negev within the framework of the encroachment episodes.

5.4.1 Speleothems and Lake Lisan records

Tzavoa Cave speleothem records indicate that in the Northern Negev, the LGM was a relatively dry period of the Late Pleistocene (Fig. 5.3). During that time, the dunefield was farther from the Cyprus Low tracks (Enzel et al., 2008), as a result received less rain, but was still affected by the winds that accompanied the cyclones, which had mean radiuses of

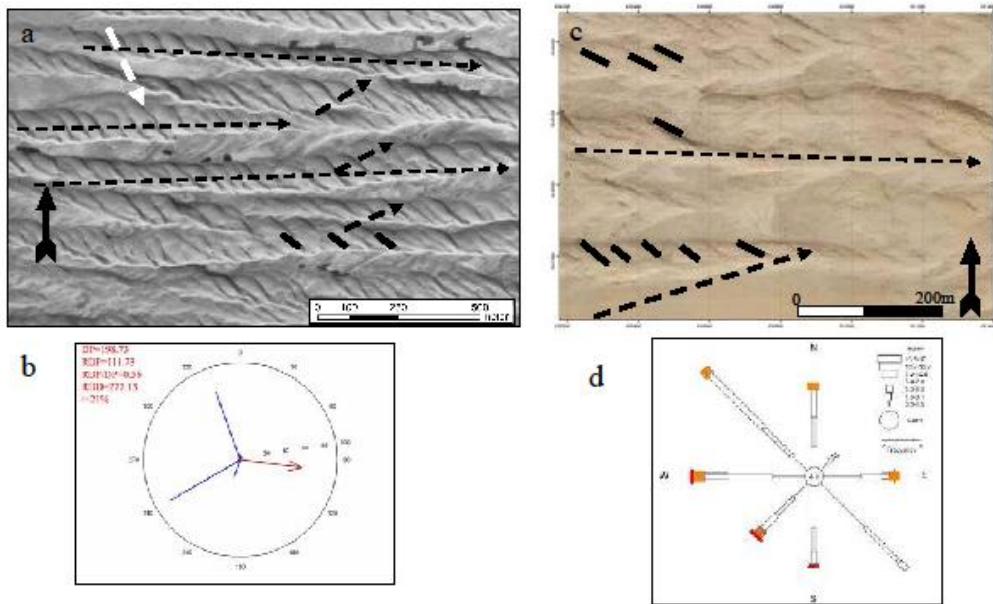


Figure 5.6. Superimposed dunelets on Sinai and Negev linear dunes.

- a. Aerial photograph of braided linear west-east oriented (thin dashed arrow) dunes in northeast Sinai (30° 51' 18.36"N, 34° 12' 09.9"E) approximately 10 km west-north-west of Nizzana (Fig. 1). Northeast facing dunelets are clearly apparent (double-line) on their crests in accordance with southwest sand-transporting winds (thick black arrow) as measured for Bir Lahfan (Fig. 6b - below).
- b. Wind rose diagram collected by Bir Lahfan, Sinai (for location see Fig. 1) showing strong southwest dunelet-forming component. Being located only 37 km from the Mediterranean coast, the strong northeast component is due to sea-breeze.
- c. Braided Negev VLD by the BM site. Symbols are as in Fig. 6a.
- d. Wind rose diagram (after Sharon and Margalit, 2002) of winds 15 meters above the surface at the Halamish ecosystem study site 5 km north of Nizzana and 5 km south of the BM site (Fig. 2).

~400–600 km (Campins et al., 2010) as observed for the December 10–12th 2010 storm (Fig. 5.1c). LGM ages are found only in the southwest corner of the Negev dunefield, perhaps because the ample sand supply required for dune buildup simply had not yet reached the NW Negev from Northern Sinai, despite the prevalence of strong winds which may have already transported sand into Sinai. The higher speleothem growth rates and high Sr concentrations in the Tzavoa Cave at 19–17 ka, towards the end of the LGM, indicate a relative windy and dusty environment (Vaks, 2008). This may be correlated to several ages of 18–17 ka from dune bases that are suggested to have preserved the incipient stage of the main episode (Roskin et al., 2011).

The rapid western Negev dunefield accretion event at ~15.8 ka is not contemporaneous with increased speleothem growth (i.e. increased rainfall) in the Tzavoa Cave record or with speleothem deposition periods in the arid Judean Desert west of the Dead Sea (Lisker et al., 2010) that indicate a drop in rainfall at ~16–15 ka. However, the Tzavoa speleothem terminal age of ~14–13 ka coincides with dune stabilization ages of 13.7–13.3 ka of the main episode. This implies that the main mobilization-stabilization episode occurred in a significantly rainier climate than the second episode. The Late Pleistocene $\delta^{18}\text{O}$ record from the Soreq cave (Fig. 5.3) is level throughout the LGM, followed by gradual warming that precedes a short cooling event at ~15.5 ka, after which there is a sharp warming that peaked at ~14.5 ka, (Fig. 5.3). The ~15.5 ka short cooling event is in agreement with the ages of the rapid dune mobilization-stabilization event at ~15.8 ka and the sharp re-warming age is similar to the drop in rainfall at 14–13 ka in Tzavoa Cave.

When compared to the fluctuation of Lake Lisan levels (Stein et al., 2010), it appears that the rapid drop in lake levels at 17.4–16 ka cal BP roughly coincides with the initial deposition of dune bases at 19–17 ka. A brief rise in lake levels at ~16–15 ka, indicating a more moist climate in the north, agrees with the rapid ~15.8 ka dune mobilization event. The return to aridity at 14.6–13.2 ka cal BP coincides with the major dunes stabilization at ~13.7–13.3 ka. The association of rapid Lake Lisan level fluctuation at 15–14.6 ka cal BP with the North Atlantic Heinrich 1 event, led Bartov et al. (2003) and Stein et al. (2010) to suggest that the following period in the Levant was colder and generally drier.

The Lake Lisan curve (Bartov et al., 2003; Stein et al., 2010) and the Soreq ^{18}O records identify a short, cool spell associated with the YD, with a short increase in rainfall, coeval with the second, less intense dune encroachment episode (Fig. 5.3). Speleothems in arid

regions (Vaks et al., 2006; Lisker et al., 2010) do not register this event. Several post-YD dune ages (10.7 – 9.3 ka) (Fig. 5.3) seem to indicate that there was a gradual subsidence in wind until the Holocene.

The partial correlation between rainfall-induced speleothem deposition and Lake Lisan fluctuations and the Negev dune mobilizations and stabilizations suggest that the Negev dunes are not solely controlled by windiness that occurred during periods of intensified rainfall.

5.4.2 Aeolian sand records

Regional aeolian chronologies coincide with the NW Negev main mobilization-stabilization episode ages of ~16–13.7 ka. The SL112 EM marine core, retrieved ~20 km east of Haifa and at water depth of 892 m, contained Late Pleistocene and Holocene fine sand, silt, and clay fractions (Hamann et al., 2008). The silt fraction sizes underwent end-member modeling to distinguish between aeolian, marine, and Nile sediments. The end member silt fraction modal grain size was 40 μm , similar to North African and Israeli dust storm grain sizes (Crouvi et al., 2008, 2009), and thus, those sediments were interpreted to be aeolian. Although fine sand and silt deposition was high during the LGM, an equivalent, outstanding episode that ended rapidly is recorded at 17–14.5 ka, based on accelerator mass spectrometry (AMS) radiocarbon dating (Hamann et al., 2008) (Fig. 5.3). The timing is similar to the Negev main dune mobilization-stabilization episode. The core's record of rapidly ending sedimentation is similar to the information gleaned from the Negev VLDs. The jumps in both sand and silt percentages are explained by Hamann et al. (2008) as "Heinrich-equivalent events". Sedimentological changes in the YD cold event, however, are only slightly visible in the SL112 core (Hamann et al., 2008) (Fig. 5.3). This again suggests the less intense regional effect of the YD relative to the H1 events.

Other studies of Mediterranean and North Africa dust and sand records at latitudes similar to those of the Sinai-Negev Erg also suggest regional-global windiness that began at the end of the LGM and continued until just before the start of the Holocene. Aeolian quartz/clay, Si/Al, and Zr/Al ratios in a piston core in the EM, roughly-dated to the LGM (23–18 ka) are interpreted to reflect a more arid climate, with increased wind speeds and greater dust transport (Calvert and Fontugne, 2001). Another piston core off the NW African margin shows low-latitude dustiness, i.e. windiness, synchronous with high-latitude Heinrich events (Julliena et al., 2007). Fe concentrations in a Nile Delta core represent aeolian dust input and

shows a major reduction in (Saharan) dust at 14.6 ka (Revel et al., 2010). Fine aeolian sand was also found to be deposited in the Atlantic off the NW coast of Africa during the H1 (Moreno et al., 2002; Mulitza et al., 2010). Along with the Hamann et al. (2008) SL112 core (Fig. 5.3), these finds indicate an extremely windy episode that was capable of transporting fine sand into the sea which co-occurred with the Negev main dune mobilization-stabilization episode.

Small dunefields, east and southeast of the study area, exhibit similar stabilization ages. Thirty-five km east of the southern part of the Negev dunefield, coarse-grain sand, reworked from the Miocene Hazeva Fm., was dated by OSL to 14.1 ± 1.2 ka (Dody et al., 2008). At Wadi Faynan in the southern Dead Sea Rift Valley, OSL age clustering suggests that aeolian activity subsided at ~ 13.7 ka (McLaren et al., 2004), which closely fits the stabilization ages of the Negev main mobilization-stabilization episode (Fig. 2). Similarities in dune stabilization ages between the NW Negev and these neighboring areas provides another indicator of a regional decrease in wind strength around ~ 14 ka, whereas OSL ages from these areas that match the second (YD) Negev episode are absent from the record. Although this may be due to the paucity of data, it may also suggest that this less intense episode did not significantly affect sand bodies located farther away from the EM.

5.4.3 Northern Negev loess records

Despite their dynamic and spatial similarities (Crouvi et al., 2010), the comparisons between NW Negev loess and dune chronostratigraphies are complex. Ben-David (2003) initially suggested that the NW Negev Late Pleistocene aeolian (dust) loess deposition and (sand) dune activity were synchronous events that reflect a windy environment. But sand and loess have different particle size distributions, which control their inherent dynamic thresholds and characteristics (Pye and Tsoar, 2009), making deposits of these aeolian sediments unique recorders of past, though differing, wind properties. Sand grains saltate whereas loess silt and clay particles are transported by suspension, and they require higher threshold friction velocity and lower transport speed than sand grains for erosion and transportation, respectively (Tsoar and Pye, 1987). Sand and dune bodies mobilize and stabilize in direct response to increases and decreases, respectively, in wind power, whereas the depositional behavior of loess remains to be elucidated, as it is aided by additional factors like rain (Kohfeld and Harrison, 2001) and vegetation (Pye and Tsoar, 1987). Hence, sand is eroded at

relatively lower wind velocities while the saltating sand can entrain silt and clay into suspension. Therefore, where dust is being transported, dune mobilization is expected.

Past windiness and particularly gustiness is inferred from dust influx. Gustiness is defined as short-term high-speed wind events that carry significant relative proportions of dust, the transport of which is non-linearly correlated to wind speed (Gillette, 1974; McGee et al., 2010). As dust deposition flux is controlled not only by windiness but also by the extent of moisture in the environment (Harrison et al., 2001), it may not be exactly correlated with peak windiness. On the other hand, sharp drops in dust-flux can signal a drop in windiness and in this scenario, dunes will probably stabilize. Moreover, Crouvi et al. (2009, 2010) and Enzel et al. (2010) suggested that the Negev loess, similar to other global loess fields adjacent to dunefields, is mainly formed by sand abrasion. Accordingly, the Negev primary loess deposits should somehow reflect both dune encroachment and stabilization events and the ages of the two sediments should be quite synchronous. However, the loess ages in the Negev indicate major and continuous deposition since ~95 ka (Crouvi et al. 2008; Crouvi et al. 2009), long before the Negev dune encroachments at 23-11.5 ka. Only the termination of loess deposition and sand stabilization ages are synchronous, at ~11.5-10 ka.

In northern Sinai, radiocarbon dates of late Upper Paleolithic sites of ~35–30 ka cal BC in basal climbing dune (ramp) sections, along with later ages similar to the Negev's main encroachment episode (Goldberg, 1977; Goring-Morris and Goldberg, 1990), while increasing the age range of dune stabilization, are still much younger than the ages of the Negev loess. The younger Sinai (~35 ka) and even younger Negev dunefield may have formed this late due to a limited sand supply from the presumed source, the Nile Delta region (Roskin et al., 2011).

The similar Negev upper loess and dune mobilization and stabilization ages may suggest synchrony between dune encroachment and massive loess-building dust deposition. The OSL-dated Hura loess section, 30 km northeast of the dunefield's eastern fringe, is the only loess section that shows rapid loess accumulation at 18.9 ± 1.3 – 17.3 ± 1.3 ka, eventually ending around 10.7 ± 1.3 ka (Crouvi et al., 2008). This may be associated with the main dune mobilization-stabilization episode, when the bulk of dune sand reached the NW Negev environs.

Other ages from northern Negev upper palaeosols (Fig. 5.1b) correlate with the main dune episode, though they may not represent the terminus of loess sedimentation. In the Ruhama

section, 30 km northeast of the dunefield, the upper loess palaeosol OSL age ranges between 14.8 and 12.4 ka (Wieder et al., 2008). At Qerem Shalom, immediately north of the study area, calcic sandy palaeosols were dated by OSL to 14.5 ± 2.3 and 13.4 ± 1.7 ka (Zilberman et al., 2007). This regional drop in loess deposition coincide with the onset of the more arid Holocene (Baruch and Goring-Morris, 1997; Vaks et al., 2006; Lisker et al., 2010), suggesting a corresponding drop in wind power (Tsoar and Pye, 1987), which also explains dune stabilization. Thus, despite being both aeolian sediments deposited in the same climatic and geographic contexts, the only full agreement between the current Negev loess chronology and the NW Negev dune ages is the simultaneous sand stabilization and the cessation of loess deposition by ~ 10 ka.

The late Holocene (2-0.8 ka) brief Negev dune encroachment and reactivation episode (Roskin et al., 2011) is also unrelated to any recorded Holocene loess deposition; this raises further questions about the suggested loess-dune chronological inter-relationship (Enzel et al.; 2008; Crouvi et al., 2010; Enzel et al.; 2010) and addresses the need to investigate additional palaeoclimatic controls on the Negev dunes mobilizations and stabilizations.

5.4.4 Northern Negev prehistoric sites

Early to Late Epipalaeolithic (~ 22 –9.6 ka cal BC) site density of the arid southern Levant (Sinai and Negev) compiled by Goring-Morris et al. (2009), peaks during the Older Dryas (associated with the H1) and the YD periods. This spans the Negev dune ages and is the only record that covers all of the Negev dune mobilization-stabilization episodes from the LGM until the Holocene. Increased human presence during a period of increased windiness seems incongruous. However, the prehistoric sites in the southern dunefield portray short-lived, seasonal camps situated mainly by paludal sediments. These sites were usually inhabited for less than 2 weeks (Goring-Morris and Goldberg, 1990) and are associated with dune-dammed standing water bodies dating to the same time (Roskin et al., 2010b). The clustering of prehistoric sites and seasonal movements between the Negev Highlands and dunefield lowlands may indicate a human response to seasonal availability of water. The sites were possibly inhabited during the less windy season at times when the dunes were stable and maintained the seasonal water bodies.

5.4.5 Summary: Late Pleistocene VLD mobilization-stabilization environment

Based on estimated doubled Late Pleistocene rainfall (Vaks et al., 2006), we suggest that the Negev dunes were vegetated in the past. Despite the higher rainfall, vegetation cover is not expected to have exceeded 20–30% in dunes, as observed for Israeli coastal dunes, and modeled by Yizhaq et al. (2009). Wind erosion stress, limited soil moisture and rapid infiltration due to the sand grain size limit vegetation proliferation on dunes (Tsoar, 1997). During mobilization episodes, higher crest wind speeds meant that the sparse dune vegetation was concentrated mainly on the dune flanks, helping to preserve the VLD morphology. Dune dynamics modeling in terms of vegetation, rainfall, and drift potential proposes that in a climate that receives 300 mm rain per year, dunes with 20–30% vegetative cover can be active when drift potentials (DP) [terminology from Fryberger (1979)] exceed ~400 (Yizhaq et al., 2007; Yizhaq et al., 2009). The Late Pleistocene DPs were probably significantly higher, considering that current DPs in northwestern Sinai have been calculated to be ~1000 (Roskin et al., 2011). Thus the Negev dunes were probably VLDs since their establishment, according to the formation models proposed above.

Regional aeolian sand records show abundant indicators of aeolian activity during the main Negev dune mobilization-stabilization episode, while the second episode stabilization ages resemble the upper loess sections dated to the onset of the Holocene. Accordingly, the Negev dunes stabilized due to the drop in windiness that, in some cases, was contemporaneous with a drop in rainfall, as understood from speleothem and Lake Lisan records. While these regional proxies indicate that the H1 and YD North Atlantic cold-events may have affected the EM differently in terms of temperature and rain, our findings suggest that during these cold-events also dune-driving windiness increased. Accordingly, we emphasize that dune stabilization events which are caused by decreased windiness (event t2 on Fig. 5.5) can be important markers for identifying climate change (Lomax et al., 2011).

5.5 The global palaeoclimate connection

5.5.1 Coincidence of GISP H1 and YD dust fluxes with Negev dune mobilization-stabilization episodes

The timing of NW Negev dunefield mobilization-stabilization episodes is comparable to the North Atlantic H1 and the YD cold events. These events are identified in the Greenland Ice Sheet Project (GISP) ice core records, such as $\delta^{18}\text{O}$ changes and dust-flux (Fig. 5.3). Antarctic EPICA Dome C (EDC) cores show sharp post-LGM drops in $\delta^{18}\text{O}$ and dust-flux at 14.6 ka,

however Greenland is closer to the Negev than Antarctica, which makes the GISP records more relevant for the Negev data. The interpreted GISP $\delta^{18}\text{O}$ reveals a temperature increase at ~ 14.5 ka, corresponding to the Bølling-Allerød warming, an event that separates the H1 and the YD (Liu et al., 2009) (Fig. 5.3). The YD, which was more pronounced in the Northern Hemisphere, was found to have ended abruptly (Steffensen et al., 2008). The Mediterranean in general and the EM region in particular were only partially affected by these North Atlantic climate systems (Allen et al., 1999; Bar-Matthews et al., 1999; Fletcher et al., 2010; Flocas et al., 2010; Stein et al., 2010).

Specifically, the Negev VLD stabilization ages coincide well with two declines in GISP ice-core dust Ca^{2+} flux between the LGM and Holocene at 14.5 ka and 11.6 ka that match the ends of the H1 and YD events, respectively. Although the debate continues over the possible sources of GISP and EPICA dust records, it is generally accepted that the high-latitude cold event high dust concentrations must have originated in mid- to low-latitude regions (such as the Sahara or the Sinai-Negev Erg) which were prone to dust entrainment by strong winds (McGee et al., 2010). Accordingly, these winds were also capable of generating regional dune mobility. This lends further support to our classification of two encroachment episodes (Figs. 2, 3 and 4) and suggests a connection between Negev dune mobility and global windiness. The sharp decline in dust-flux to negligible values also fits our interpretation that dune mobilization in the Negev was halted abruptly due to sharp decrease in wind power.

5.5.2. Post-LGM - Holocene global luminescence-dated dune mobilization and stabilization

The relationship between windy global cold events and dune mobilization and stabilization can be tested on other dunefields. Late Pleistocene linear dune activity younger than 35 ka has been identified in many deserts in the (low) latitudinal range but no clear-cut relationships were found with glacial/deglaciation transitions (Munyikwa, 2005; Lancaster, 2007; Singhvi and Porat, 2008). This implies that dune systems are not solely or immediately controlled by global glacial-interglacial fluctuations or by the subsequent climatic impact of those fluctuations, as initially proposed (Sarnthein, 1978).

Similar to the Negev VLDs, major luminescence age datasets of mainly vegetated linear dunes from the Southern Hemisphere shows a peak of ages following the LGM (Roskin et al., 2011) (Fig. 5.3). Episodes of dune activity also preceding the LGM were identified for the

Southern Hemisphere dunefields (Telfer and Thomas, 2007; Fitzsimmons et al., 2007; Lomax et al., 2011). This accentuates the palaeoclimatic significance of the large age clusters post-dating the LGM, while acknowledging the possibility that these later mobilization episodes are better preserved. Studies with fewer ages also support this find. Based on 11 OSL ages from two sections, Tripaldi et al. (2011) identifies for the southern fringe of the Medanos de los Naranjos dunefield in Argentina one long dry windy interval extending from 23–13 ka or two intervals from 23–20 ka and 16–13 ka. Luminescence dating of North African dunes produced similar age clustering: Fourteen OSL ages from the Great Sand Sea of Egypt, though sampled from depths of only 2–5 m, range from 22.8 to 11 ka (Besler, 2008). Nineteen OSL ages from the Western Sahara desert of Mauritania also cluster around 25–15 ka and 13–10 ka (Lancaster et al., 2002). All these dunes stabilized by ~10 ka, probably reflecting a general global decline in windiness. Radiocarbon dates, mainly from interdunal sediments, suggest major dune activity in the Sahara at ~18–12 ka (Nicholson and Flohn, 1980), and Swezey (2001) also identified aeolian activity in different parts of the Sahara since ~25 ka, with a distinct stabilization at 11 ka.

Dunefields in the Southern Hemisphere show increased dune mobilization and stabilization at 14–10 ka, later than the EDC drop in dust-flux around 14.6 ka. This continued post-dust dune mobilization in the Southern Hemisphere may reflect a drop in windiness, such that the wind lacked sufficient power for dust entrainment but was still able to transport sand. Furthermore, dune mobilization and stabilization in southeastern Australia coincides with the Antarctic cold reversal (ACR) at ~14.5–12.5 ka (Fitzsimmons et al., 2007; Lomax et al., 2011). The Kalahari age dataset between 14–10 ka (Telfer and Thomas, 2007) has a slightly different age distribution compared to the Australian age distribution (after Fitzsimmons et al., 2007) (Fig. 5.3) that may be attributed to local climatic factors and lagged response.

The strong correlation between Northern Hemisphere global and regional Late Pleistocene dust flux, and consequently between windiness, colder climates and dune mobilization, suggests that when annual average rainfall is equal to or above 100 mm, globally-controlled powerful, sand-transporting winds are a major factor in dune transport. Although arid conditions enable dune activation, this alone cannot trigger major dune elongation. Furthermore, VLD elongation in association with vegetation that functions as a wind obstacle, requiring stronger wind power than for bare (vegetation-free) dunes (Tsoar, 1989), further strengthens the connection between Late Pleistocene VLD elongation and globally enhanced windiness.

5.5.3. Post-LGM – Holocene palaeoclimatic control of NW Negev windiness

General atmospheric circulation during the glacial period through the LGM is accepted as having been more vigorous than in the Holocene, largely because of enhanced temperature gradient between the upper latitudes and the tropics (Ditlevsen et al., 1996). During the LGM-Holocene transition, cold temperatures still reigned in the upper latitudes, causing sharp baroclinic temperature gradients, though global circulation and windiness, including in the EM, subsided (Delmonte et al., 2004). The temperature gradient during glacial times and the North Atlantic (H1 and YD) cold events is suggested to have generally increased global low-latitude windiness (Swezey, 2001) along with rain (Hemming, 2004) and EM cyclogenesis (Enzel et al., 2008).

In order to resolve the effect of past temperature gradients on EM cyclogenesis we can harness studies on the current EM climate. Analogous to the baroclinic conditions that currently launch EM cyclonic storms (Enzel et al., 2008; Campins et al., 2010), the sharp baroclinity of the Late Pleistocene may have intensified EM cyclonic storm characteristics, which were further intensified in the winter and spring (Denton et al., 2005) when the temperature gradient between the high and the low latitudes became even steeper. Analysis of the last 40 years of EM cyclonic tracks has shown a positive connection between a decrease in winter storm track frequency and size to a decrease in baroclinity (Flocas et al., 2010) suggesting that an increase in baroclinity will yield higher storm frequency. Enhanced baroclinity has also been positively correlated with Beer-Sheva dust deposition, which, in turn, has been positively correlated to probably colder and rainier EM cyclonic winters (Dayan et al., 2008; Enzel et al., 2008). These findings may be analogous to times of significantly increased EM baroclinity, such as during glacial periods and North Atlantic cold events, implying that cyclone tracks were then larger and more frequent.

In addition, westerly (originating from the Atlantic and western Mediterranean) and also southwesterly (passing over North Africa) EM cyclones were found to have high velocities (>5 m/s; Flocas et al., 2010). Accordingly, Late Pleistocene dune-encroaching windiness may have been of a more westerly direction which fits the Late Pleistocene northern Negev dune orientation.

To summarize, the increased windiness of the Late Pleistocene may have been part of the recognized typical EM wintertime cyclonic mechanism, of events that today usually last ~24

hours, but in a few, rare cases approach 72 hours (Campins et al., 2010). The Late Pleistocene EM cyclonic mechanism may also have been more enduring at both the daily and the annual levels, enabling dune mobilization. This climate was punctuated by exceptionally strong wind events mainly during global cold-events, unrecognized in the modern record that explain the stratigraphic evidence of rapid accretion and transport rates and pulses.

The proposed direct control by global cold events on Negev dune formation may help approximate climate change scenarios for the Negev dunefield. Expected global warming and possible drought, as observed during the last decade and a half in the Negev dunefield (Seigal, 2009), if prolonged, will probably enable a certain small-scale activation of the Negev dunes due to vegetation dieback. However, global warming will probably include a decrease in baroclinity, EM cyclogenesis and consequent wind power (Pryor et al., 2009), thereby limiting dune mobilization and preventing dune encroachment into new territory.

5.6 Conclusions

By building on the data of Roskin et al. (2011), this paper: 1. Presents a model for VLD elongation, stabilization and accretion. 2. Improves the spatial and temporal resolution of the dune mobilization episodes in the NW Negev. 3. Temporally and conceptually ties the Negev dune mobilization-stabilization events to Northern Hemisphere cold events. 4. Shows the connection between Late Pleistocene windiness and drops in windiness to global dune activity and stabilization.

The NW Negev dunefield OSL-dated record is easily accessible due to the relatively high preservation of VLDs at the eastern end of the Sinai-Negev Erg. Here the major dune mobilization-stabilization episodes cause dune elongation and accumulation. The exposed, drilled and dated VLD sections together with previous modeling enabled us to suggest a general model of VLD formation. Combined with statistics and spatial analyses, this model enabled us to identify two post-LGM Negev dune mobilization-stabilization episodes: a main one at ~16–13.7 ka, and a minor one at ~12.4–11.6 ka, when the dunes reached their maximum spatial extent and stabilized. These episodes included rapid dune encroachment events and coincided with the H1 and YD cold events. In contrast to current southwestern sand-transporting winds, the Late Pleistocene dune-mobilizing winds were characterized by a strong western direction that dictated west-east VLD elongation.

Greenland ice core dust-flux records fluctuating during cold-events are controlled by changes in interpreted global wind power during the LGM-Holocene transition, at the same times that the Negev dunes mobilized and stabilized. Among others, Australian and southern African vegetated dunes, abundantly dated by OSL, also contain indications of significant post-LGM to Holocene dune mobilization and stabilization in accordance with Antarctic dust records. We therefore suggest that some of the post-LGM global dune mobilizations were controlled mainly by global cold-events and subsequent changes in windiness. The low and mid-latitude global windiness that entrained dust also transported dunes and drops in windiness caused dune stabilization.

The Late Pleistocene EM cyclonic systems were probably deeper, more frequent, and longer-lasting than those of today, thereby generally enhancing sand-transporting windiness in the Sinai and Negev. Sufficient amounts of rainfall allowed the dunes to support vegetation which in turn demanded increased windiness for sand transport. The H1 and YD cold-events further increased windiness in the EM, boosting the Negev dune mobilizations and stabilizations. Stabilization of Negev VLDs was brought about by a decrease in regional storminess with the onset of a more arid Holocene climate.

The discontinuous aeolian sedimentation pattern found in VLDs provides new important chronological control on prominent dune mobilization-stabilization episodes. The suggested link between global drops in wind power following cold-events and low-latitude dune stabilization emphasizes the prevalence of winds over aridity regarding major dune mobilizations for low-latitude dunes, even if they are vegetated.

The anticipated global warming and the increased potential for severe drought conditions, based on observations of the last decade and a half in the Negev dunefield, could eventually enable a certain level of Negev dune activation. However, the counteracting absence of a cold and windy climate will limit dune elongation and subsequent considerable dune encroachment.

This work highlights the importance of luminescence dating of VLDs down to their base and across their internal structure, as well as their contribution to understanding regional and global palaeoclimate changes. It also laid the groundwork for future, comprehensive research in different ergs and dunefields (including northern Sinai) to improve the temporal resolution and palaeoclimatic implications of late glacial global aeolian activity.

6. SUMMARY

6.1 Synopsis

6.1.1 Overview

The study presents a large and spatially dense optically stimulated luminescence (OSL) age database of the NW Negev dunefield, which it supports with fully documented dune sections, sedimentological and spectroscopic data, geomorphic attributes, and remotely sensed spatial patterns. OSL age clustering reveals the main NW Negev dunefield development stages from the age of the underlying substrate to the current dune activation modes. That data helps us understand and model the vegetated linear dune elongation and accretion dynamics, which combined with the spatial and statistical analysis, allows us to refine the timing and intensity of the Negev dune mobilizations. This enables us to perform a palaeoclimatic analysis focusing on heightened periods of windiness in the past.

This information allows us to identity the timing and characteristics of the palaeoclimatic and paleoenvironmental controls of the NW Negev dunefield mobilizations which were useful for tracing aspects of the origins and timing of the incipient Sinai-Negev erg history.

The Negev dune mobilization-stabilization events are temporally and conceptually tied to Northern Hemisphere cold events and the study proposes a connection between Late Pleistocene windiness / drops in windiness and global dune activity and stabilization.

The evolutionary account of the Late Pleistocene Negev dunefield is an invaluable asset for interpreting historical late Holocene and recent Negev dune mobilizations and reactivations as well as for assessing future dune and dunefield responses to climate change scenarios. Several methodologies and concepts were found to challenge accepted notions: Ground penetrating-radar (GPR) profiles of Late Pleistocene vegetated linear dunes showed no differentiation between chronostratigraphic units. Dune sand redness intensity was not found to correlate with age.

The methodologies, sedimentological, and chronological data in this study combined with the proposed interpretations that it offers can serve as a "tool-kit" for future palaeoclimate research into arid aeolian sand and specifically, vegetated linear dunes (VLDs). These issues which comprise the main scientific input of this study are briefly reviewed in the following sections.

6.1.2 Methods for dune studies

6.1.2.1 Exposed VLD stratigraphy

OSL ages were obtained from a variety of landforms relevant to understanding landscape evolution and for correlating the ages with previous studies that targeted mainly fluvial and interdune sections. Aside from the VLDs, other landforms in the study area were: 1) mature palaeosol substrates beneath the dunes; 2) interdune (ID) fluvial/aeolian sediments; and 3) transverse dunes. Stratigraphy of exposed dune sections perpendicular to the dunes axis combined with OSL dating of discerned sand units was invaluable for analyzing VLD dynamics and chronostratigraphy.

6.1.2.2 OSL age performance

OSL ages, determined using the single aliquot regenerative dose (SAR) protocol was found to be a reliable age estimator for the Negev sands and sediments. All sand samples displayed a strong initial OSL signal and rapid decay. The relatively homogeneous nature of the sands produced similar, low dose rates, reflecting the high quartz content. De distributions were usually normal and most of the OSL samples had over-dispersion values below 20%. Samples with several tailing aliquots of higher and/or lower De values were attributed to contamination by bioturbation, a minute quantity of underlying older sand, subtle differences in beta micro-dosimetry and possible differences in sand grain exposure to light during saltation.

In order to organize previously published and unpublished ages into a single chronological framework, the ages from past luminescence protocols and published ^{14}C dates were compared to the OSL ages in this study in places where similar units were dated. In most cases, a high correspondence was found between the ages even though they were not sampled at the same time or at the same location/stratigraphic section. This multiple suite of concordant ages obtained via different dating methods is a positive evidence of the reliability and significance of the OSL ages, and places the NW Negev dunefield in a well-defined chronological framework.

6.1.2.3 Sand age - redness index ratio

Throughout the NW Negev dunefield, the spectrally-measured redness index of full dune and interdune sand does not correlated with the OSL depositional age found for the sand. These results disagree with the initial hypothesis, challenging prevailing assumptions. Though there is no supporting evidence, it cannot be ruled out that Negev sands that have remained in-situ

since the Late Pleistocene may have undergone pedogenetic processes and rubification shortly after their deposition in a Late Pleistocene climate rainier than today's and have not changed since then. However, the Sinai sands have similar RI values to the Negev sands, suggesting that the iron-oxide coating of the sand grains is an earlier diagenetic characteristic of the sands.

6.1.2.4 GPR applicability for VLDs

The GPR reflection profiles of the dune cross-sections and playas penetrated 5-10 meters but were an unreliable tool for sampling-based identification of sand units which came mainly from Late Pleistocene dunes. Vague horizontal contacts found in the dune axis may have been caused by bioturbation of bounding surfaces and homogeneous grain-size. Bedding sets of the upper dune crests and slopes were distinguished by GPR.

Although the GPR interpretation of playas identified reflection units, ground-truthing drilling showed that the playa stratigraphy usually consisted of a main silt unit several decimeters thick, close to the surface, overlying a sandy sequence. The GPR interpretation may have been biased by thin silty units inducing reflection.

6.1.3 Evolution of the NW Negev dunefield

The NW Negev dunefield's OSL-dated record is relatively accessible due to the high preservation of VLDs. In this case, the major dune mobilization-stabilization episodes led to dune elongation and accumulation. The sedimentary archive of the Late Pleistocene Negev VLDs reveals that different VLD morphologies usually have similar chronostratigraphy, thus allowing the dunefield's evolution to be mapped.

OSL ages support earlier works (Zilberman, 1991, 1992; Ben-David, 2003) which contended that sand deposits had been in the NW Negev since ~100 ka. The sands are part of calcic sandy to silty palaeosols that were exposed at the surface until ~30 ka and then eroded and buried by the encroaching dunes. These palaeosols point to a relative stability in the Negev prior to the LGM and provide invaluable markers that dunes were not present in the Negev before ~30 ka. Archaeological evidence also indicates a transition from sand and loess-loam surface sediments to dune encroachment around ~25 ka (Goring-Morris and Goldberg, 1990). OSL probability density functions (PDFs) delimit the main episodes of aeolian sand transport and deposition into the Negev, while spatial, stratigraphic, and statistical analysis complements the age clustering, promoting an inclusive understanding of the Negev dunefield's evolution. The OSL age distribution revealed three age clusters; 24-10 ka, 2-0.8 ka, and 150-10 years which are

consistent with the chronostratigraphic units of the VLD axis and point to the main dune encroachment and remobilization episodes.

At ~24 ka, initial evidence of dune encroachment was only found in the southwestern dunefield. The main NW Negev dune encroachment, dated to ~18–11.6 ka, can be divided into two important Late Pleistocene dune encroachment episodes: ~16–13.7 ka (first episode) and ~12.4–11.6 ka (second episode). The first episode covered the northern and western parts of the dunefield and encompassed several mobilization events. An initial event of basal sand cover occurred at ~18–16 ka. Later, sections of 2–10 m thick accumulated throughout the western part of the dunefield, mainly from ~16 ka onwards; dune and sand thickness decrease in the east. During this episode, sand was transported over ~85% of the dunefield. Based on the substantial thickness of the sand and its spatial cover, the first episode of the main encroachment was defined as the main episode. Its time span resembles the Heinrich 1 cold event dated by ^{14}C to be in the range of 16.8 ka (Hemming, 2004) to ~14ka (Vidal et al., 1999) and suggests that the post-LGM period left a strong aeolian imprint on the region.

During the second episode (~12.4–11.6 ka), the easternmost configuration of the dunefield was shaped and dunes and sands became elongated and stabilized. The dunes dating to this episode have indistinct undulating morphologies and are usually less than 5 m thick. This episode involved less intense accretion, lower sand transport distances, and a shorter duration than the main episode. The sands probably originated from dunes deposited during the main episode in the western dunefield. In the western field, the paucity of units dated to ~12.4–11.6 ka was probably due to the later erosion of this relatively thin, episodic unit during the late Holocene remobilization. The time span of this episode resembles the Younger Dryas cold event.

Massive dune encroachment in the southern dunefield caused the damming of fluvial systems, and formed shallow standing-water bodies that deposited light-colored, often sandless loam units that supported short-term prehistoric camps. Spectrally mapped by the supervised classification of a Landsat TM (2003) image of the spectral enhancement of a mineral composite, exposed surfaces of standing-water deposits show similarities between the southwestern Negev dunefield and northeastern Sinai in the vicinity of Wadi Al-Arish. It is therefore suggested that Wadi Al-Arish was also blocked during the main dune encroachment, causing extensive upstream and interdune flooding and deposition of fines.

Based on chronostratigraphy and intermittent OSL age clustering between 2 and 0.8 ka this study provides the first identification of the distinct late Holocene dune remobilization episode. This episode contains unusual spatial characteristics and dune superposition and probably

developed under windy conditions, albeit less intense than the Late Pleistocene encroachment episode.

The abundant late Holocene palaeoclimatic stratigraphic, archaeological and historic data in the Negev has furnished the debate on climate change and whether increased aridity in fact induced the collapse of the Northern Negev's Byzantine towns and extensive agricultural infrastructure (Issar et al., 1989; Rubin, 1990; Avni et al., 2006). The present study supplies new evidence for the discussion, suggesting that drought followed by strong winds, probably triggered the later Holocene dune mobilization. The ruins of the Byzantine city of Halussa (Haluzza) are covered by 1-2 meters of sand, and historical letters attest to sand incursion that decimated its grape vines (Meyerson, 1994). Although by this time humans had long occupied the northern Negev, the dune thickness argues against possible reactivation solely due to the anthropogenic effect brought on by the decimation of the region's stabilizing biogenic crust and vegetation cover. The coeval formation of vegetated-linear and transverse dune types may be a consequence of strong west-east winds which elongate the VLD's (Tsoar et al., 2008). Byzantine sites along the northern Sinai and southern Mediterranean coast of Israel have been covered by several meters of sand (Neev et al., 1987), which may imply that stronger winds from Mediterranean winter storms, affected Israel, including the northern Negev.

Intermittent sand activity and stabilization in the last 150 years ago has reactivated dune crests and slopes but without causing dune elongation. The OSL ages are consistent with anthropogenic land-use changes. Six OSL ages date to ~68-40 years, probably relating to increased dune activity due to Bedouin presence and grazing that gradually ended in the early 1950's.

6.1.4 VLD dynamics

The study proves that the Negev VLDS to be accreting and elongating duneforms. This is based on OSL dating of exposed sections that displayed discrete sedimentary units, with 12 sections dated by at least 3 ages. OSL ages essentially represent the burial age of the sample. While the location of the OSL age in a dune section is generally accepted to represent the end of dune mobilization, this study extends the significance of OSL ages in regard to VLDs.

A series of OSL ages along the defined sedimentary unit of a VLD axis provided burial ages of sand during an accretive event that occurred while the dunes were continuously being mobilized. The basal age of a defined sand unit generally dates the initiation of sand accretion, whereas the age of the upper unit marks a later time of deposition immediately prior to

stabilization. If this pattern is spatially repetitive throughout the dunefield, as found in the NW Negev, we can perceive OSL age clustering, mainly between the bottom (initial – mobilization) and top (end – stabilization) event ages, as representing what we define as a *dune mobilization-stabilization event*. If the several units are of similar ages, that sequence represents a dune mobilization-stabilization episode composed of several events.

The bottom of mobilization-stabilization dune units might be lost, and the top slightly truncated, with missing ages. To varying degrees, active dunes can slightly truncate the upper unit and depositional surface of the previous mobilization-stabilization event/episode. Thus, stabilization ages may be postdated by later dune crest activation and re-deposition or truncated by later dune mobilization events/episodes.

As discontinuous landforms, linear dunes can be considered problematic for palaeoclimate reconstruction (Telfer and Thomas, 2007; Chase, 2009). Nonetheless, the discontinuous aeolian sedimentation pattern in VLDs provides important new chronological control of prominent dune mobilization-stabilization episodes by recording the major dune building events and episodes. Thin mobilization-stabilization units may have been fully reworked by intermittent and possibly long-term local surficial activity, which could have erased them from the chronostratigraphic record. Therefore, VLDs are prime recorders of dominant periods of enhanced windiness.

Furthermore, following the line that increase in dust during the Late Pleistocene is due to (short-term) gustiness (McGee et al., 2010), it is suggested that the thick aeolian sand and dune sections recorded short-term, high wind-power events that formed the majority of the sand record during the main encroachment at 16-13.7 ka. Around ~15.8 ka there was a rapid accretive event that probably involved dune elongation. Different sections of the dunefield revealed almost identical ages of ~13.7 ka, marking the end of the main mobilization-stabilization episode and show a joint abrupt stabilization, probably due to a sharp drop in wind power.

Other measurements also supported the rapid-event windiness hypothesis presented here. If dune accretion was a gradual sedimentological process, each small-scale sand deposition would be eroded or reset luminescence-wise during dune activity, or else prone to bioturbation during dune stabilization (Bateman et al., 2007). Gross averaged sand transport rates along the direction of the VLD elongation were approximately several m/yr to several tens of m/yr, lending further support to the suggestion of rapid mobilizations of the Negev dunes.

6.1.5 Relationship of Negev dune mobilization ages to loess depositional ages account

A chronological comparison and interpretation of the partial, dynamic, and spatial similarities between the NW Negev dunes and three documented northern Negev primary hilltop loess deposits encompassing the NW Negev dunefield recently reported by Crouvi et al. (2008, 2009) is inevitable. Around the NW Negev dunes, considerable variance has been found in deposition amounts and rates for sections of Negev hilltop loess, which according to Crouvi et al. (2008), are connected to the Sinai-Negev dune mobilizations. They postulate that the silt grains comprising most of the loess are not reworked from the dunes but are generated probably through active eolian abrasion of the sand grains under past climates characterized by intensified winds. It is reported that Negev loess deposition included several depositional phases, the first beginning already around ~200 ka near Har Harif (Crouvi et al., 2008), long before the Negev dune encroachments. The loess ages in the Negev indicate significant dust deposition after ~95 ka (Crouvi et al. 2008, 2009), long before the Negev dune encroachments at 23-11.6 ka. During this period, loess sections recorded relatively thick, though varied, dust deposition. The only apparently synchronous occurrences were the termination of loess deposition (Crouvi et al., 2008) and Negev dune stabilization at around ~11.6-10 ka, which may be the results of a sharp (global) drop in windiness at the end of the Younger Dryas. So, based on the similarity between the ages of the Negev upper loess and the NW Negev dune stabilization, can we offer any insight regarding the chronological and palaeoclimatic similarities between the older Sinai and Negev sand mobilizations and loess-building dust depositions? One could attribute the earlier loess deposits to low-latitude windiness such as earlier glacial and (Heinrich) cold-events, during which time sands and dunes may have been activated further west. A sand loess sequence activated during global cold-events since the Heinrich 5 event has been suggested for the Mu Us sand field in China (Zhou et al., 2009). However, for lack of chronological records regarding dune mobilization in Sinai, the Nile Delta and the Egyptian desert west of the Delta this cannot be confirmed here. It should also be noted that dune radiocarbon dates and OSL ages southwest of the Delta in the Great Sand Sea of Egypt usually do not exceed the LGM (Bubenzer et al., 2007; Besler, 2008).

There is no report of loess deposition related to the late Holocene (2-0.8 ka) Negev dune encroachment and reactivation episode. The relatively limited scale of this episode might have incorporated limited dust deposition, which was possibly later and more-easily eroded from the upper sediment record. In several interdunes, dust sections overlay thick sand deposits and may mark a dust deposition slightly postdating sand stabilization and/or additional late Holocene dust deposition.

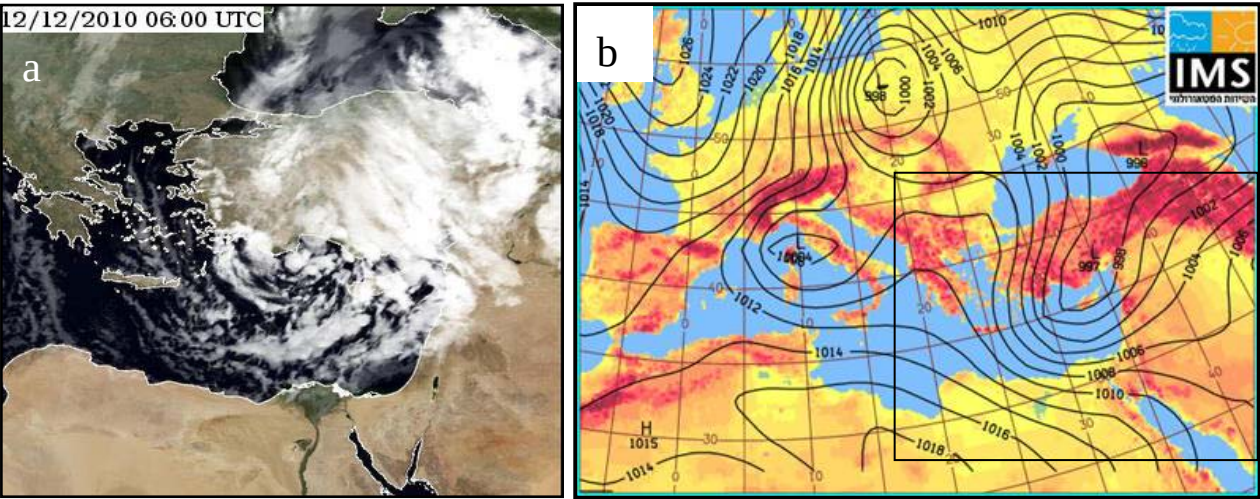


Figure 6.1 The December 10-12th, 2010 Eastern Mediterranean cyclone storm.

a. Image of the cyclone over the East Mediterranean Sea..

b. Synoptic map of the Mediterranean region during the December 12th 2010 storm. Note the northerly track of the Cyprus Low that is centered over Cyprus and southern Turkey.

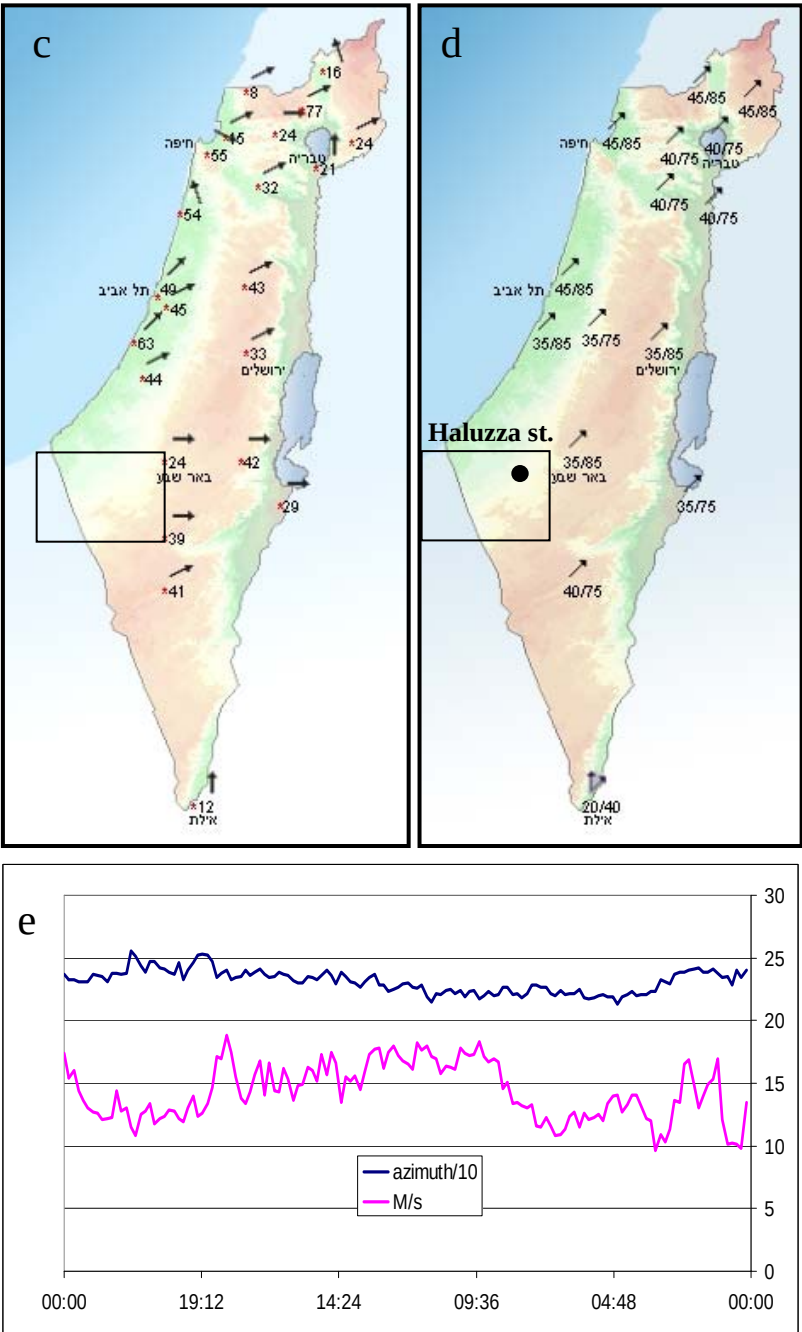
c. Israel wind map 11.12.10.

d. Israel wind map 12.12.10. Note increased speeds of southwesterly wind directions measured at the Haluzza metrological station.

e. Northwestern Negev wind direction and speed at the Haluzza station (for location see d) on 12.12.10.

Data acquired from the Synoptic map Israel Meteorological Service

http://www.ims.gov.il/IMSEng/All_tahazit/SynopticMaps.htm)



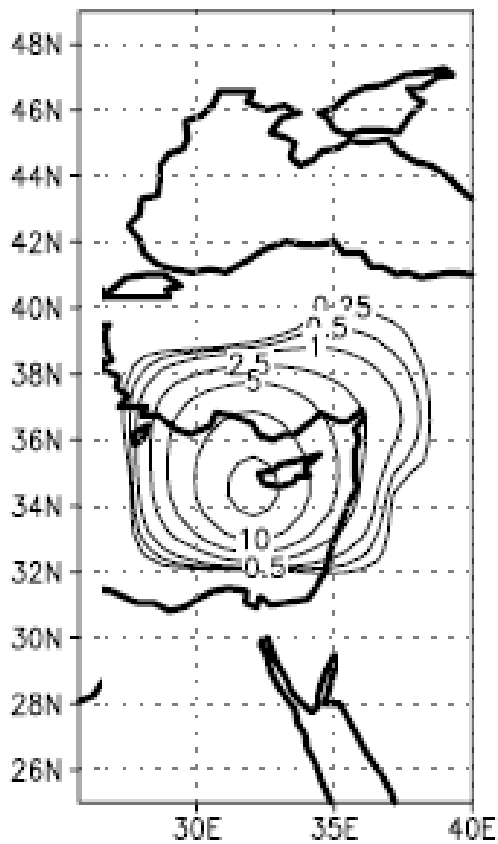


Figure 6.2 Annual Cyprus cyclone track density (after Campins et al., 2010). Note the strong gradient north of northern Sinai.

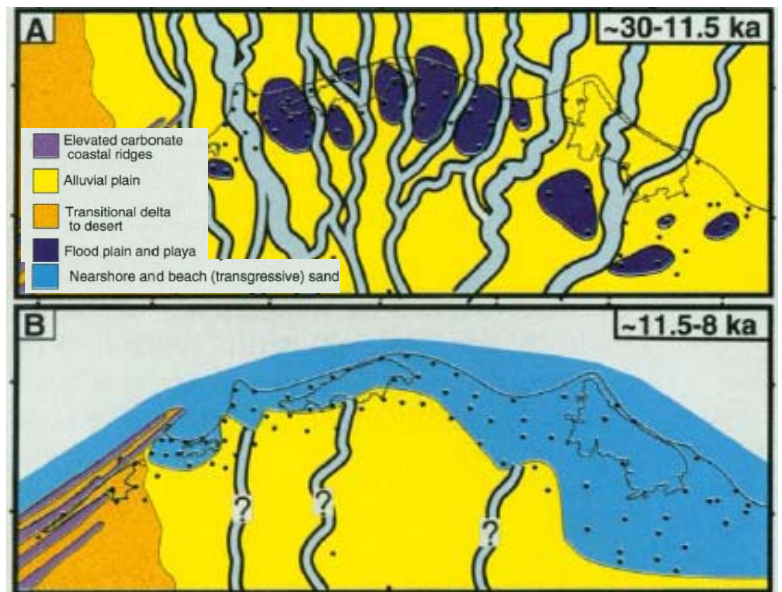


Figure 6.3. Time slice palaeogeographic maps of the northern Nile Delta (Stanley and Warne, 1993). The red line marks the current coast-line. Note how the 30-11.5 ka age fits the full age span of Negev dune activity.

Indeed, further research is required to define the links between dune mobilization and downwind loess deposition and their corresponding palaeoclimates and palaeoenvironments.

6.1.6 Negev dune-driving palaeoclimates

As suggested by Enzel et al. (2008), the Late Pleistocene Eastern Mediterranean Sea cyclonic systems were probably similar to the current Cyprus low pressure synoptic conditions that produce winter storms, rain, and windiness, in Israel, including and north of, the northern Negev (Fig. 6.1). The current routes of the Cyprus Lows run mainly in the center and northern parts of the EM (Figs. 6.1 and 6.2) (Campins et al., 2010), their radius averages around 400-600 km (Flocas et al., 2010), extending over northern Sinai, and their duration rarely exceeds 3 days (Campins et al., 2010). Today, when these systems move across the EM, a high pressure gradient is formed over the northern Sinai region generating strong SW-W orientated winds. Based on the orientations of the VLDs, we can infer strong westerly palaeowinds that differ from the current SW-W sand-transporting dominating winds, suggesting that Late Pleistocene winds also differed slightly in direction

General atmospheric circulation during the glacial period through the LGM is accepted as having been more vigorous than in the Holocene, largely because of enhanced temperature gradient between the upper latitudes and the tropics (Ditlevsen et al., 1996). During the LGM-Holocene transition, cold temperatures still reigned in the upper latitudes, causing sharp baroclinic temperature gradients, though global circulation and windiness, including in the EM, subsided (Delmonte et al., 2004). Analogous to the baroclinic conditions that launch EM cyclonic storms today (Enzel et al., 2008; Campins et al., 2010), the sharp baroclinity in the EM during the Late Pleistocene may have intensified EM cyclonic storm characteristics. These were probably further intensified during winter and spring (Denton et al., 2005) when the temperature gradient between the high and low latitudes became even steeper. Analysis of the last 40 years of EM cyclonic tracks has shown a positive connection between a decrease in winter storm track frequency and size, and a decrease in baroclinity (Flocas et al., 2010) suggesting that increased baroclinity yields higher storm frequency. Enhanced baroclinity has also been positively correlated with Beer-Sheva dust deposition, which, in turn, has been positively correlated to colder and rainier EM cyclonic winters (Dayan et al., 2008; Enzel et al., 2008). These findings may relate to periods of significantly increased EM baroclinity, such as glacial periods and North Atlantic cold events, implying that cyclone tracks were then larger, deeper, more frequent, and longer-lasting.

Generally, adequate Late Pleistocene rainfall allowed the dunes to support a denser vegetation cover than today's, so that for sand transport to occur, increased windiness was required. The receding northern Sinai coastline during the LGM distanced the cyclonic systems from the Sinai-Negev erg (after Enzel et al., 2008) and probably caused a drop in rainfall and consequent vegetation cover. This led to increased dune sand erosivity and probably more sand erodibility and transport due to the strong winds. This climate was punctuated by exceptionally strong wind events (gustiness; see McGee et al., 2010), mainly during the global cold-events, which are not recognized in the modern climate record, explaining the stratigraphic evidence of rapid accretion, transport rates, and pulses.

Stabilization of the Negev VLDs resulted from a decrease in regional storminess and accompanying windiness at the end of the Younger Dryas, and the onset of a more arid Holocene climate.

6.1.7 Evolution of the Sinai-Negev erg

It is suggested that Northern Sinai dune encroachment is chronologically similar, though slightly earlier, than the Late Pleistocene NW Negev encroachments. With reference to wind data for the Sinai and to the modeling of dune activity under differing wind regimes and vegetation (Yizhaq et al., 2009), I wish to suggest that in some scenarios, during the wetter Late Pleistocene which sustained a higher vegetative cover, Negev dune erosivity was not a prerequisite for sand to encroach from Sinai into the NW Negev. Rather, I propose that strong winds and sand availability were the primary factors underlying sand transport across northern Sinai and into and onto the NW Negev.

Differential sand supply from Northern Sinai can also explain the morphologic differences between the dunes of the three NW Negev encroachment corridors. This is also expressed by the lack of a regional correlation between sand age and sand thickness. Western parts of the central corridor boast maximum sand thicknesses that taper off to the north, east and south.

Based upon limited Sinai radiocarbon ages from the Gebel Maghara vicinity (Goldberg, 1977), this research suggests that the sand supply that eventually reached the Negev only began to appear in Northern Sinai after ~35-40 ka, as a result of ample Nilotic sand supply that was probably available to erosion during the last glacial period when global and Mediterranean Sea levels dropped (Fig. 6.3). Late Pleistocene to current Nile Delta sand grain stain intensity and mineralogy values (Stanley et al., 1988; Stanley and Chen, 1991; Stanley and Warne, 1993)

offer supporting (though partial) evidence that Nile Delta sands may be the main, already-red source of sand for the Sinai-Negev Erg.

However, due to the paucity of the data we cannot rule out that dune mobilization which was confined to the Sinai took place long before the above mentioned ages.

6.1.8 Global dune-driving windiness

Greenland ice core dust-flux records that fluctuated during cold-events were controlled by changes in interpreted global wind power during the LGM-Holocene transition at the same times that Negev dunes were mobilized and stabilized. The Australian and southern African vegetated dunes, also densely dated by OSL, contain indications of major post-LGM to Holocene dune mobilization and stabilization events. I therefore suggest that some of the post-LGM global dune mobilizations were controlled mainly by cold-events and subsequent changes in windiness. The low and mid-latitude global windiness that entrained dust that was deposited in higher latitudes also transported dunes, and drops in windiness caused dune stabilization.

The suggested link between falls in global wind power following cold-events and low-latitude dune stabilization strongly points to the prevalence of winds rather than aridity as responsible for major dune mobilizations in low-latitude dunes, even though these may have been vegetated.

The anticipated global warming and the increased potential for severe drought conditions based on observations of the last decade and a half in the Negev dunefield (Siegal, 2009; Siegal et al., in prep.) might eventually bring about a certain level of Negev dune activation. However, global warming will probably include a decrease in baroclinity, EM cyclogenesis and consequent wind power (Pryor et al., 2009) thus limiting dune mobilization and preventing dune encroachment into new territory. Therefore, the counterbalancing factor of a warm and windy climate will act to limit dune elongation and any considerable dune encroachment.

6.2 Overview of research contribution

6.2.1 General

My Ph.D. research has involved intensive field work and laboratory analyses on the aeolian sediments of the northwestern Negev Desert. I have gained new insights into sedimentological research methods for sand and dunes and the internal structure and formation dynamics of vegetated linear dunes. An unprecedented number (97) and density of optically stimulated

luminescence (OSL) sand ages leaves us with a better understanding of quartz grain luminescence properties and capabilities. Aided by dense spatial and vertical OSL ages, age clustering was identified in probability density functions (PDF), representing Late Pleistocene and Holocene periods when dunes encroached into the Negev. Exposed stratigraphic sections permit the refinement of the PDF interpretation. These sections have supported the development of a conceptual regional palaeoclimate model which emphasizes sand-transporting windiness, a climate feature that is generally overlooked in palaeoclimatic studies, explaining dune encroachments and stabilizations. Also the palaeoclimatic model suggests a global palaeoclimatic interpretation of worldwide dune mobilization and stabilization between the Last Glacial Maximum (LGM) and the Holocene.

The results also improved the (OSL) chronology to the Epipalaeolithic to recent archaeological survey and research of the northern Negev and for the first time enable us to date the periods when standing-water deposits synchronously developed with dune encroachments as a result of dune-damming. The results and modeling have improved our differentiation between anthropogenic and natural influences on the upper dune section. The learned controls of past dune mobilization allowed a conceptual forecast to be made of potential dune mobilization and consequent dustiness relevant to future climate change and global warming.

The data collected in this research allowed me recently to become a contributing member of the Sand Seas and Dunefields of the World Digital Quaternary Atlas <http://inquadunesatlas.dri.edu/>

6.2.2 Details

6.2.2.1 Research Methods

- a. Sands with minimal moisture content may usually be hand-drilled and sampled easily for OSL and sedimentology to depths of 10-12 meters using Dormer drilling equipment.
- b. For the first time, systematically dated dune bases were found to be unequivocally beneficial for dating initial dune encroachment.
- c. OSL ages measured by the single aliquot regenerative (SAR) protocol can date extremely recent (up to 8-10 years) landform changes such as surficial dune activity. The recent ages were correlated to periodic anthropogenic and domestic land-use changes in the last 150 years.

- d. Ground penetrating radar (GPR) which is believed to be a promising tool for mapping shallow sediments, failed to provide reliable imaging of the main internal VLD structure where palaeosols are absent and there is a uniform fine sand and grain-size distribution.

6.2.2.2 Late Quaternary landscape evolution and palaeoclimate implications

- a. Both Late Pleistocene and Holocene Negev dune mobilization occurred over short time spans of 1-2 kyr and were probably characterized by rapid accretion and elongation events subjected to wind duration and intensities that are unrecognized today.
- b. Accordingly, apart from crestal activity, the Negev dunes were generally stable in the Late Pleistocene and Holocene.
- c. The correlation between the Late Pleistocene Negev dune encroachment episodes and the Northern Hemisphere cold-events marks the regional climatic extent of these cold events, and more uniquely, the reason for windy palaeoclimates at low-latitudes in the past.
- d. Using global dune age data and dust flux data in ice cores, low-latitude vegetated linear dunes with accretive stratigraphy can be unique recorders of significant changes in past global windiness, especially since the LGM.
- e. First time evidence of an intermittent late Holocene 2-0.8 ka dune mobilization episode highlights the impact of intense anthropogenic activity on sand erosivity and the potential for future dune mobilization due to short-term climate fluctuation.
- f. Standing-water bodies dated to Late Pleistocene dune encroachment and late Holocene remobilization were formed due to massive dune-damming.

6.2.2.3 Vegetated linear dune structure, dynamics, and sedimentology

- a. The present research mapped the internal structure of the VLD and showed for the first time that this is different from the internal structure of seif dunes.
- b. The dune *mobilization-stabilization* model defines the capabilities and limitations of OSL dating for sand accretion and dune elongation in preserved VLDs.
- c. For the first time, the intensity of spectrally measured Negev sand grain redness was found not to increase with age, as previously suggested. This implies inherited sand redness.
- d. Unimodal grain-size distribution peaks plotted along the sand transport paths do not indicate a decrease in grain-size downwind.

6.3 Future research

The present research highlights the importance of luminescence dating of VLDs down to their base and across their internal structure, and the contribution of VLDs to understanding regional and global palaeoclimate changes. It also lays the groundwork for future comprehensive research of different ergs and dunefields (including northern Sinai) to improve the temporal resolution and palaeoclimatic implications of late glacial global aeolian activity

To complete the story and palaeoclimatic interpretation of the Sinai-Negev erg and present the future environmental implications, it is important to conduct a research program similar to the present one, for the northern Sinai dunefield, and to implement a comprehensive, goal-oriented sampling scheme.

The contribution of sand and dune mobilization to dust formation and downwind loess (soil) deposition is a hot subject in the aeolian and sedimentological scientific community. It is a crucial factor in understanding dunefield-derived dust generation and loess formation and geologic, historic, and current dust characteristics in arid and semi-arid environments. Based on the data collected in this study, I have developed hypotheses that I plan to pursue.

The correlation between Late Pleistocene glacial and cold-events and Negev dune mobilization may also have left an imprint on loess accumulation rates in the joint form of close-range deposition of dune mobilization products and deposition of distal dust. These hypotheses can be tested for the Sinai-Negev erg and downwind aeolian loess deposits.

Dune-dammed standing water bodies may provide possible positive feedback regarding climate change in the NW Negev. By mapping the spatial and volumetric characteristics and analyzing the quality, genesis, endurance, failure, and infiltration of the Negev standing-water bodies, we can assess their contribution to life sustenance and aquifer refill.

I would also like to compare the 2-0.8 ka VLD mobilization episode of the Negev with the (mainly) parabolic dune and sand mobilization ages along the Israeli coast, which are reported to cover Roman-Byzantine sites.

These future research topics encompass various methodologies and sedimentologies relating to palaeoclimates from the Pleistocene to the present. The results will invariably improve our understanding of past and future climate change—mainly with regard to increased windiness and the conditions that form, modify, entrain, transport, and deposit sand and dust particles, and their environmental impact on the air, and on the earth's surface and sub-surface.

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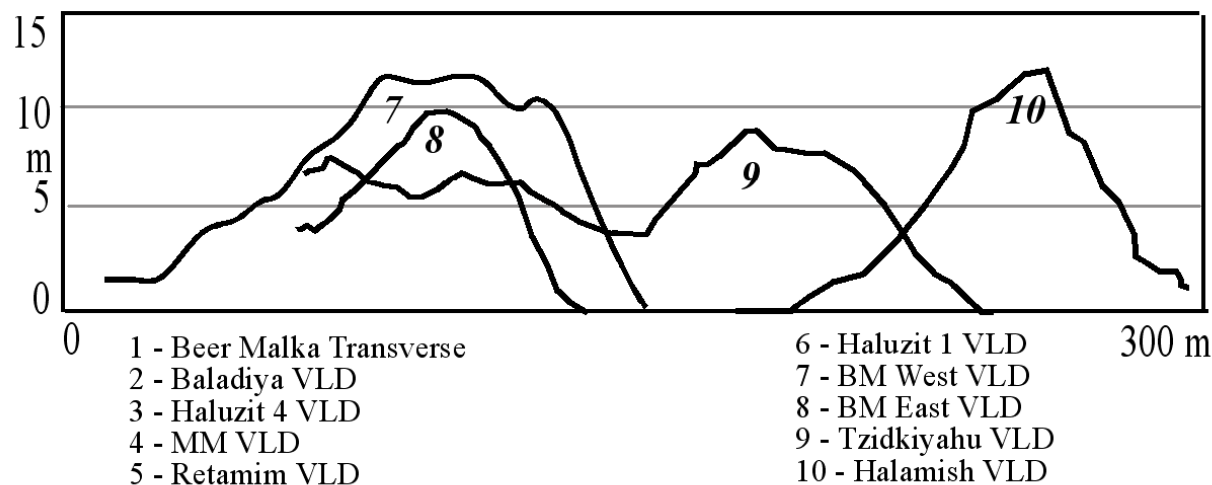
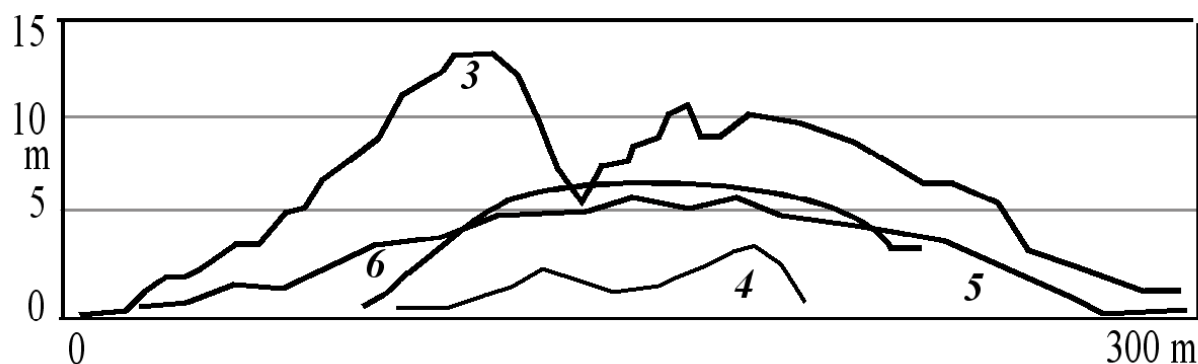
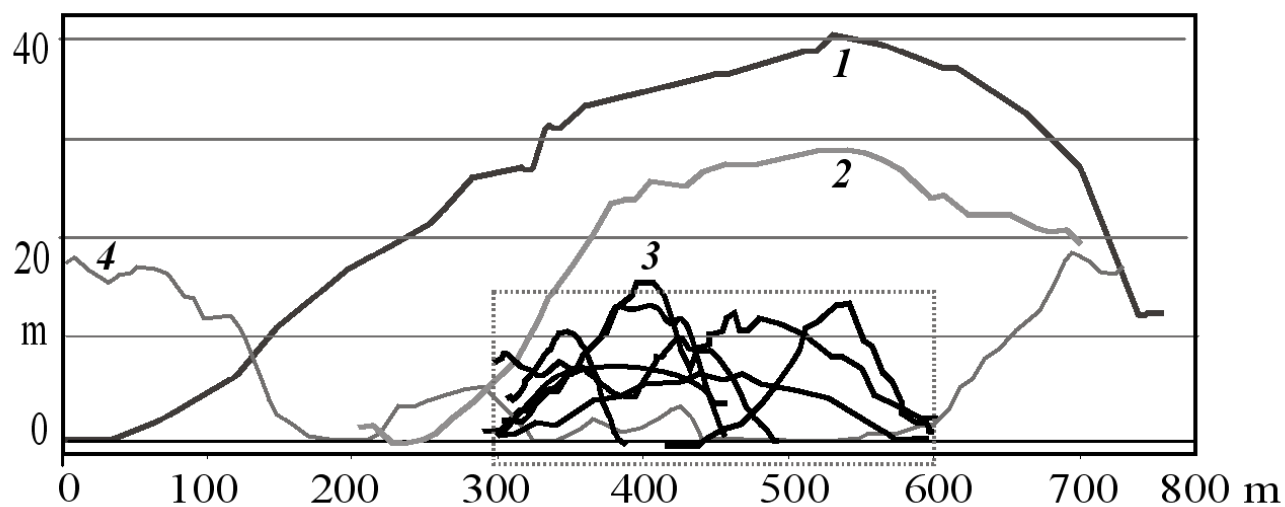
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APPENDICES

Appendix A. Topographic and ground-penetrating radar sections

Appendix B. Sedimentology data

Appendix C. Remote sensing procedures



A.1 Topographic cross-sections of the Negev dunes, sampled for OSL.. For dune location see figure 2.3. Note the different geometries; height, width, slope angle and face and general height/width ratio. Dunes 5 and 7-9 were chosen for GPR profiling. One of the reasons was their differing cross-section geometries.

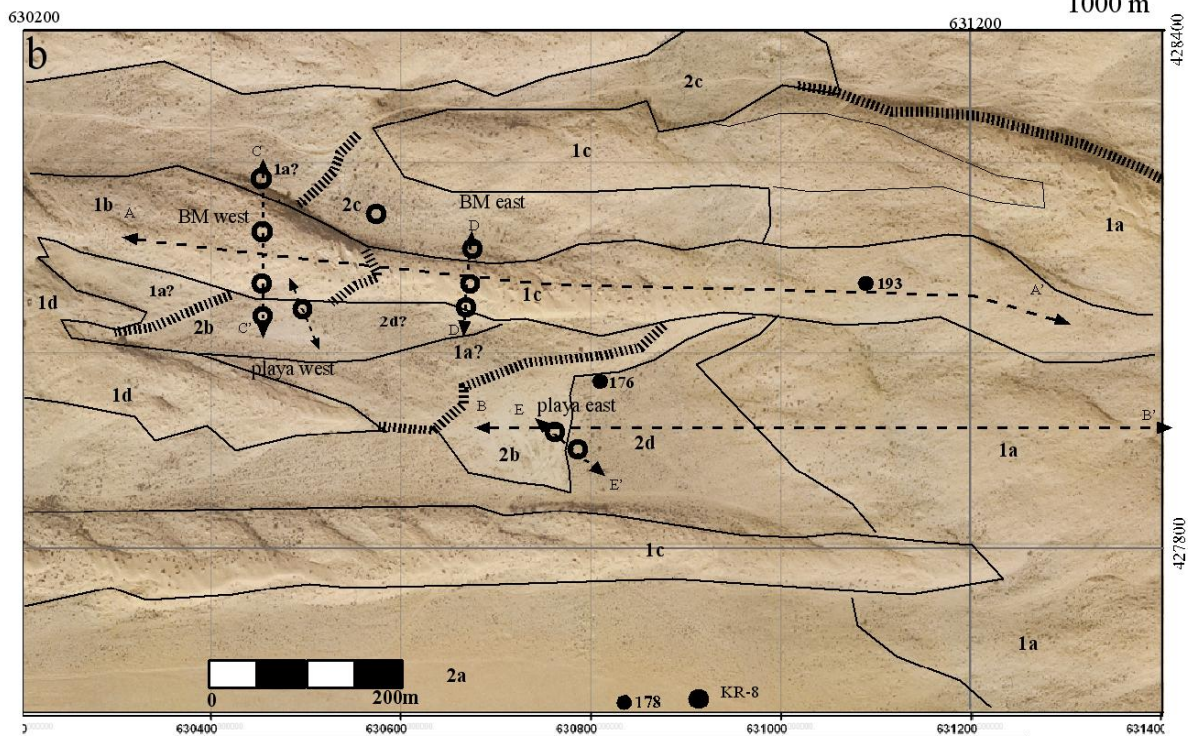
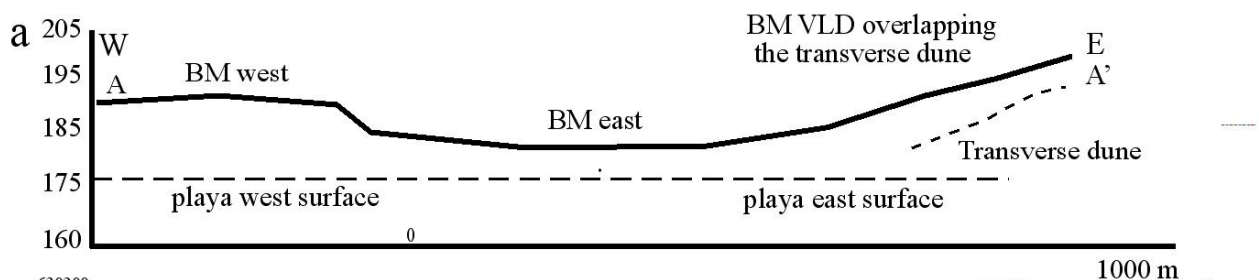
A.2. Sampling sites and sections data, where OSL sampling took place.

No '	Name	Site Description	Latitude (N)	Longitude (E)	Sampling method	Max. sampling depth (m)
1	Haluzit 4	Stable, broad VLD crest.	31° 8'46.77"N	34°19'45.83"E	3.5 m ex. +A	6.9
2	Haluzit 4 Hothouse ID	Interdune sand.	31° 9'46.77"N	34°19'15.83"E	Ex.	2.4
3	Haluzit 1 west	VLD axis.	31° 8'28.35"N	34°18'35.76"E	Ex.	8.5
4	Baladiya	Broad high VLD.	31° 9'46.22"N	34°31'51.28"E	3.5 m ex. +A	10
5	KD 73 ID	Interdune depression beneath transverse dune.	31° 05'06.41"N	34°19'22.13"E	A	6.6
6	KD 73	VLD crest.	31° 05'05.05"N	34°19'23.59"E	A	9.2
7	Tzidkiyahu ID	Interdune depression at base of transverse dune slip-face.	31° 2'38.20"N	34°21'44.48"E	A	11.2
8	Tzidkiyahu upper	Transverse VLD interdune.	31° 2'38.20"N	34°21'44.48"E	A	8
9	Tzidkiyahu VLD	VLD crest.	31° 2'38.20"N	34°21'44.48"E	A	7.2
10	BM western playa	Interdune playa.	30°58'40.99"N	34°21'56.62"E	Ex.	1
11	BM west flank	VLD southern flank.	30°58'38.95"N	34°21'56.57"E	A	5
12	BM west	VLD crest.	30°58'40.99"N	34°21'56.62"E	A	10
13	BM west N flank	Steep (30 ⁰) northern slip face.	30°58'40.99"N	34°21'56.62"E	A	5
14	BM east S flank	Southern stabilized VLD flank.	30°58'39.43"N	34°22'2.25"E	Ah	2.5
15	BM east	Active, flat VLD crest.	30°58'39.43"N	34°22'2.25"E	A	7.6
16	BM east N flank	Northern stabilized steep VLD flank.	30°58'39.43"N	34°22'2.25"E	Ah	2.5
17	BM ID	Interdune depression beneath transverse dune slip face.	30°58'40.56"N	34°22'1.00"E	A	7
18	BM eastern playa (W)	Interdune 160 m west of transverse dune with silty surface.	30°58'33.52"N	34°22'9.05"E	A	5
19	BM eastern playa (E)	Interdune 150 m west of transverse dune with sandy surface.	30°58'33.52"N	34°22'9.06"E	A	11.5
20	Halamish East	Broad VLD crest 100 m west of VLD nose plunge into Nahal Nizzana.	30°56'38.77"N	34°23'38.89"E	A	10
21	Halamish dune flank	Trench exposing dune flank base contact with fluvial sediments.	30°56'26.81"N	34°23'25.38"E	Ex.	1.2
22	Halamish ID West	Stratified sediment layers.	30°56'21.84"N	34°22'25.81"E	Ex.	2.5
23	Halamish ID (East)	Stratified sediment layers.	30°56'26.81"N	34°23'25.68"E	Ex.	2.5
24	Beer Malka	Transverse dune stoss slope contact with floodplain.	30°55'49.82"N	34°24'17.64"E	Ex.	5
25	Nizzana reservoir	Sand piedmont upon N. Nizzana floodplain.	30°53'42.29"N	34°24'49.07"E	Ex.	3.7
26	Mitvakh	Stabilized fossilized VLD axis upon Plio-Pleistocene terrace.	30°53'58.16"N	34°27'37.01"E	Ex.	9.25
27	Besor terrace	Elevated pebble covered dune sand terrace.	30°59'58.77"N	34°25'6.11"E	A	5
28	MM	Low active VLD axis.	31° 05'31.91"N	34°26'6.93"E	Ex. + A	7.5
29	MM ID	Loamy interdune.	31° 05'31.41"N	34°26'6.93"E	Ex.	1
30	Retamim ID	Interdune - northern toe of dune.	31° 05'35.27"N	34°41'52.07"E	A	8.2
31	Retamim	Broad active VLD crest.	31° 05'35.27"N	34°41'52.07"E	A	8
32	Nahal Sekher XXX	Exposure of shallow Harifian site.	31° 05'48.53"N	34°45'30.07"E	Ex.	1.5
33	Nahal Sekher VI	Excavation of Natufian site.	31° 05'49.35"N	34°49'41.07"E	Ex.	2.6
34	Ramat Beqa	Sand quarry 12 m. wall beneath active sand.	31° 6'11.66"N	34°49'47.95"E	Ex.	8

A.3 Site data, description and main results for the ground-penetrating radar profiles.

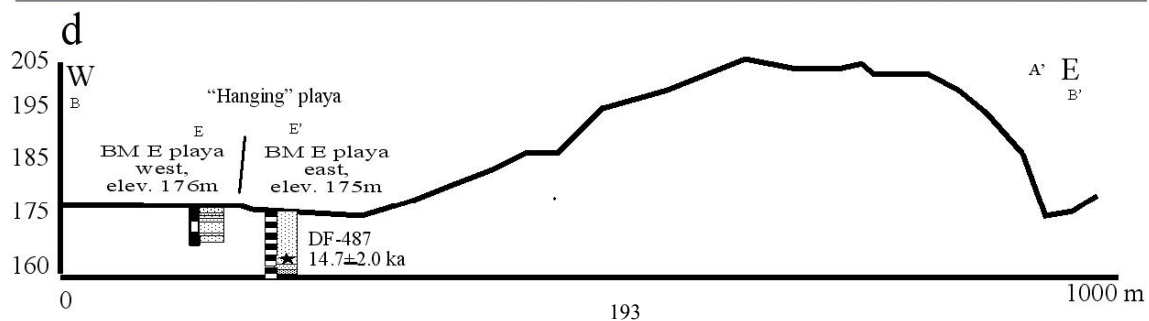
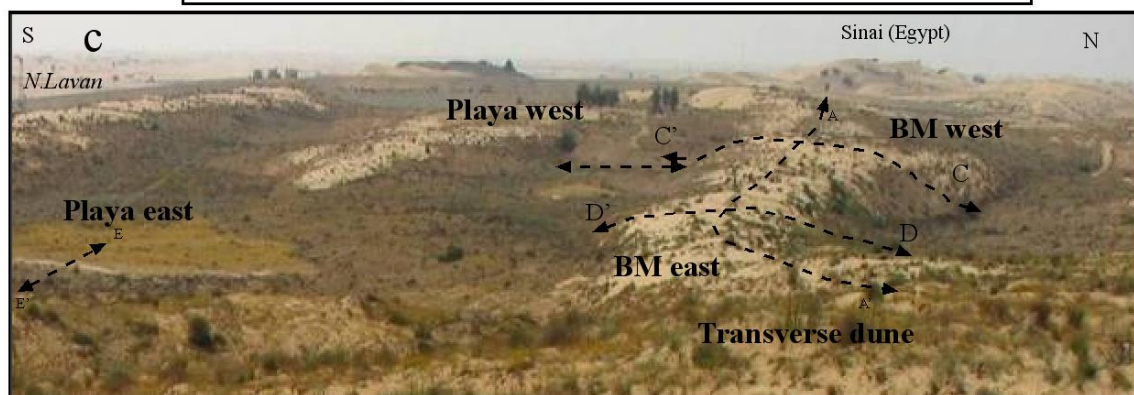
Research site-section	Site/profile Description	Latitude (N) Longitude (E)	Frequency MHz	Process and display method	Length (m)	Display and Interpretation Unit-identification	OSL ages	Remarks and radar-stratigraphy interpretation
BM-west	Asymmetrical shaped VLD with flat crest.	30°58'40.99"N 34°21'56.62"E	100	Migration, amplification, instantaneous frequency wiggle	150	10 m penetration. Discontinuous reflections throughout the VLD define radar facies. Dune slopes: Radar facies of planar-wavy reflections dip parallel to sub parallel to the surface slope angle (7-33°). The facies' contacts with neighboring ones vary; parallel, oblique and tangential. Dune axis: Radar facies of planar-wavy reflections dip parallel to sub parallel to the horizontal-sub-horizontal surface slope angle. The facies' (reflection) contacts are generally horizontal along the axis and oblique and tangential with slope facies. Dune-slope-base upper facies overlays playa facies.	2.2 ka – 10 yr	Radar facies contacts were merged into generally continuous and parallel main reflection contacts. Northern (recent) steep slope is tangentially differentiated. The southern slope has cut and fill structures.
BM-east	Asymmetrical shaped VLD with flat crest.	30°58'39.43"N 34°22'2.25"E	100 and 200	Migration and amplification of changes in depth + wiggle (200 MHz)	70	8-10 m penetration. Discontinuous reflections throughout the VLD define radar facies. Dune slopes: Radar facies of planar-wavy reflections dip parallel to sub parallel to the surface slope angle (7-20°). The facies' contacts with neighboring ones vary; parallel and oblique with truncated downlapping-toplapping topset reflection geometries on the southern slope. Dune axis: Radar facies of generally planar reflections dip parallel to sub parallel to the horizontal-sub-horizontal surface slope angle. 200 MHz image show up to 8 reflection surfaces. The facies' (reflection) contacts are generally horizontal along the axis and extend towards the slopes Horizontal VLD-axis sand units.	1.67 ka – 0.15 ka	Radar facies cover large proportions of the dune cross-section other than the southern slope and were merged into generally continuous and parallel main reflection contacts.
BM Playa west	Interdune playa.	30°58'40.99"N 34°21'56.62"E	100	Migration and bandpass filter; wiggle	50		undated	Drill de-validated GPR interpretation.
BM Playa east (west profile)	Interdune 160 m west of transverse dune with silty surface.	30°58'33.52"N 34°22'9.05"E	100	Migration and bandpass filter; wiggle	30	2-3 m penetration. Horizontal reflections. Distinct 2 m deep bounding surface. 191	Undated and uncorrected topographically	Calcic loam palaeosol at 8 m is not identified. The shallow bounding surface may be due to silt presence down to 1 meter.

BM Playa east (east profile)	Interdune 150 m west of transverse dune with sandy surface.	30°58'33.52"N 34°22'9.06"E	100	Migration and bandpass filter; wiggle	30		14.7 ka at base. Uncorrected topographically	.
Tzidkiyahu Transverse	West-East transverse /barchanoid VLD and sand interdune (ID) depression .	31° 2'38.20"N 34°21'44.48"E	100	Migration and amplification of changes in depth	120	Upper section: 8 m penetration. 3 distinct units/2 reflections parallel to the surface slope. Slip face: 5 m penetration (not reaching ID continuation). Reflection surfaces of foresets parallel to 30° slope. Upper facies is a continuation of the upper section while the toplapping ID upper unit. Interdune (ID): 10 m penetration and 3 horizontally bounding facies.	1.2-1.4 transverse; 50-70 yr slip face; 15.5-15.9 ka ID	Calcic loam palaeosol at 10 m is not identified. Survey was conducted from west to east.
Tzidkiyahu VLD, transverse dune (N-S) transect)	North-south transverse/bar chanoid interdune top between two VLD crests.	31° 2'38.20"N 34°21'44.48"E	100	Migration and amplification of changes in depth; wiggle	190	Poor and partial 4-6 m penetration with reflections following the duneslope. Vague dune axis horizontal surfaces.	1.4 ka VLD	
Retamim VLD	Asymmetric and broad VLD.	31° 5'35.27"N 34°41'52.07"E	100	Migration and amplification of changes in depth	170	Poor and partial 4-8 m penetration with vague reflections following duneslope. Possible cut and fill axis unit. Possible wedge of ID sand overlapped by colluvial dune base sand.	10.4 ka VLD; 27-1.2 ka ID	Calcic loam palaeosol at 8 m depth of interdune deprsion base is not identified.

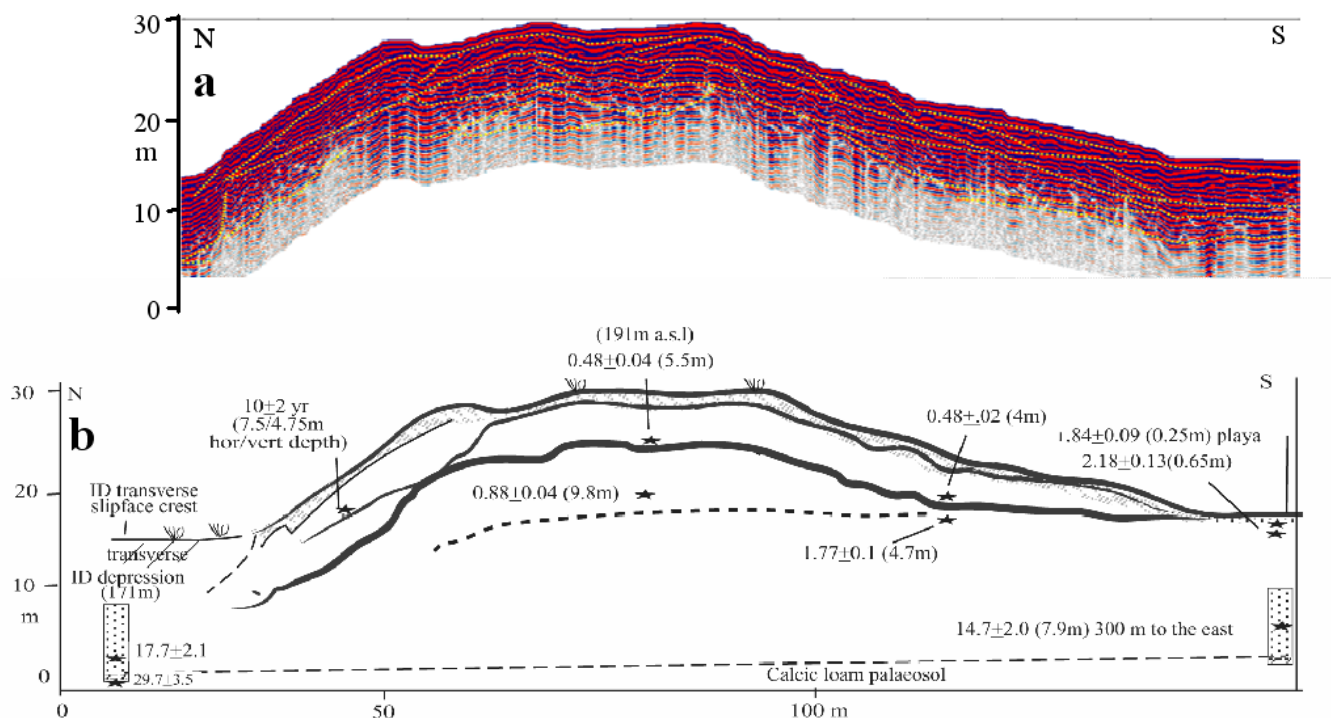


LEGEND

- | | | |
|--------------------------|----------------------------|---------------------|
| 1a Transverse dune slope | 2a Nahal Lavan flood plain | Tranverse slip face |
| 1b Broad VLD | 2b Interdune playa | Unit boundary |
| 1c Narrow/Straight VLD | 2c Interdune depression | GPR profile |
| 1d Short /meanering VLD | 2d Vegetated sandsheet | Manual drillhole |

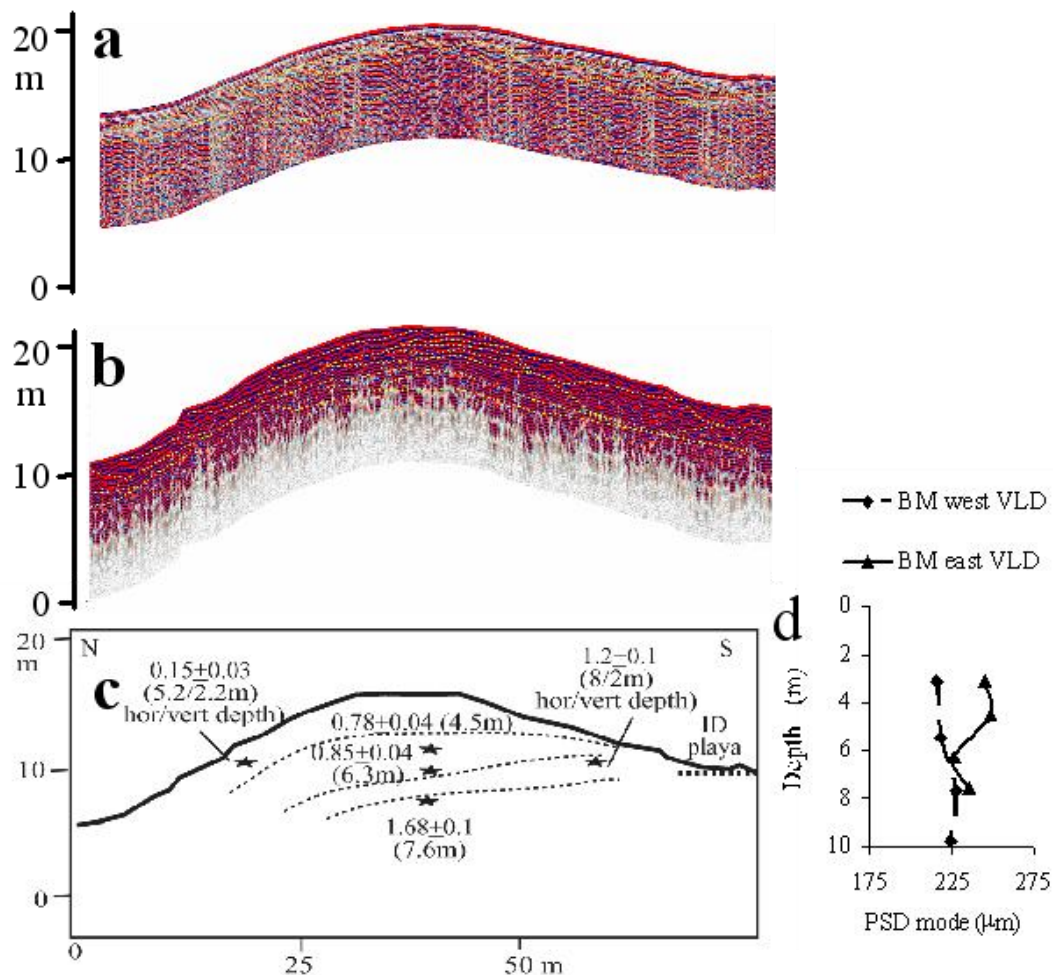


- A.4 The BM site (previous page) . This site was the most-widely drilled and GPR-profiled. This was due to the assumption that the ancient watercourse of Nahal Lavan in the floodplain valley south of the site (where drill KR-8 is in the figure) ran through the site west of the large BM transverse dune (c), to be later diverted by the dune encroachment. However, no evidence was found for this assumption.
- Longitudinal topographic section of the BM VLD, densely OSL-dated and profiled.
 - Geomorphic map of the BM site.
 - Photograph looking west over the BM site taken from the BM transverse dune. Taken in February 2008, the site is covered with perennial vegetation, let alone the active dune crests. The section lines correspond to figure A.3b.
 - Topographic section of the BM east playa and the large transverse dune to its east. The transverse dunes that often accumulate downwind of fluvial sand deposits and the playa deposits comprised supporting data for the hypothesis that the BM site was part of the ancient and diverted (Nahal) Lavan wadi.

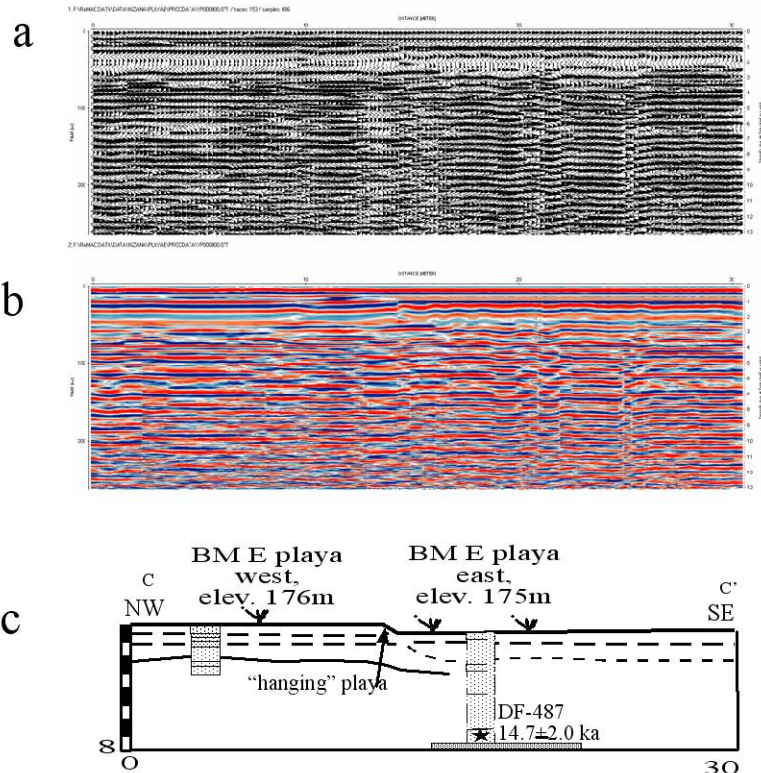


A.5 BM VLD west

- Initial interpretation of the BM VLD west profile. The yellow lines mark interpretations of contacts (see table A.2 for details).
- Main sedimentary contacts and OSL ages of the BM VLD west profile. Note the thick, steep and young northern face, the correlation between the playa and dune base ages and the young VLD ages. The 0.48 ka age was identified in the same unit both in the dune axis and in the dune slope. Note the interdune aeolian sand fill whose base, dated to the late Pleistocene, overlays calcic loam palaeosols.

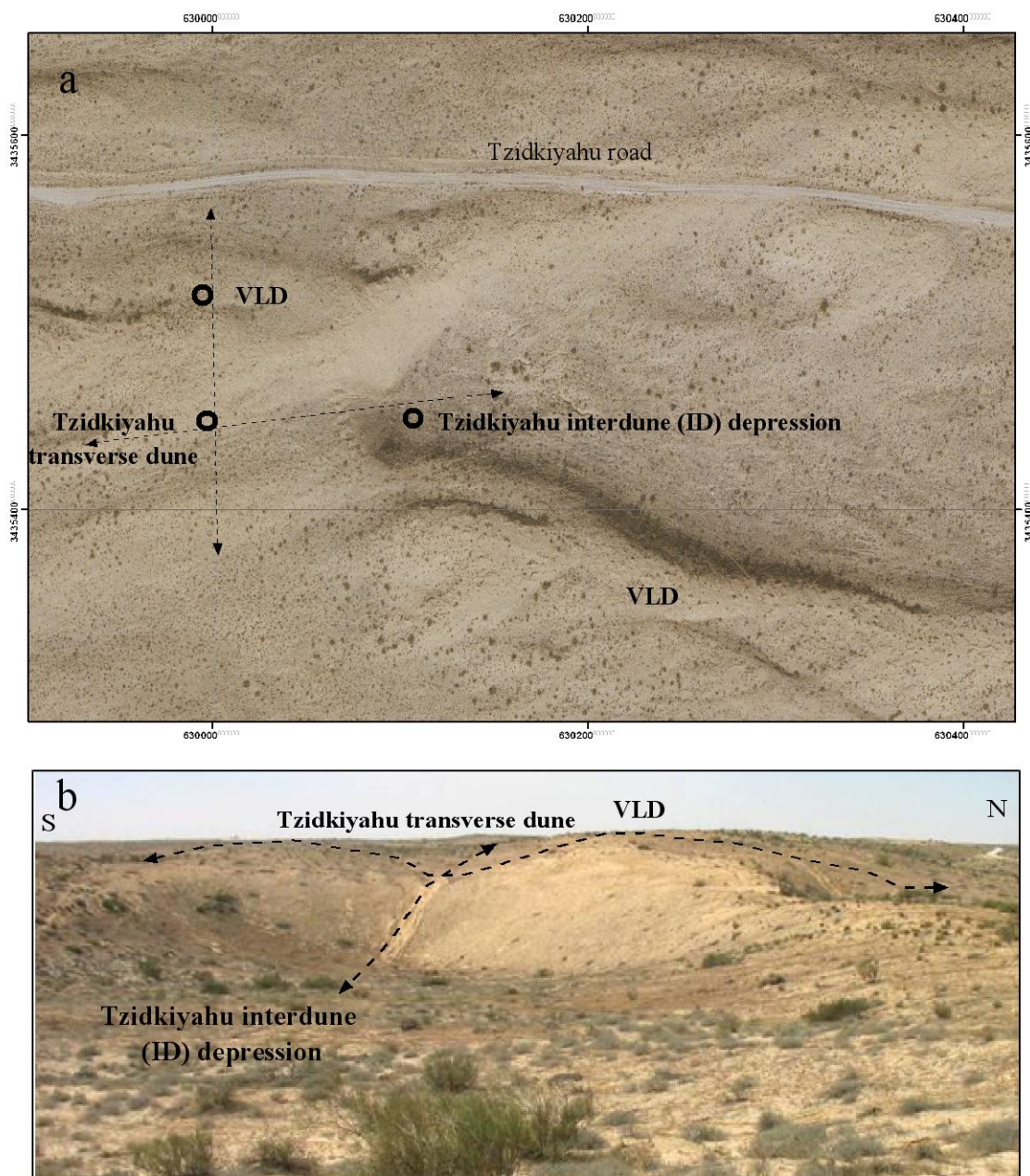


A.6 BM VLD east GPR profile. GPR processed image (see A2 for details) for 100 (a) and 200 (b) MHz antennae respectively, and (yellow line) interpretations of contacts. The latter reveals continuous contacts but shallower penetration. Note the complexity of the interpretation that is merged into figure A.6c along with OSL ages. The profile shows sediment facies along the slopes and accretion of sand in the dune axis with horizontal contacts. This was observed at Haluzit (see chapters 3 and 5 for details on analysis of VLD stratigraphy). d. Particle-size distribution mode vs depth of the BM VLD west and east profiles.



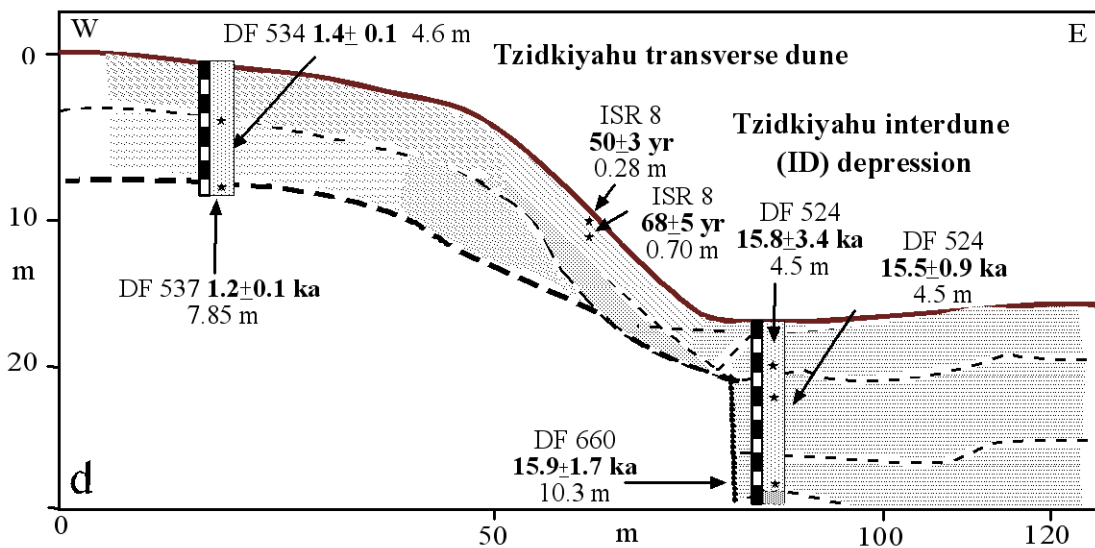
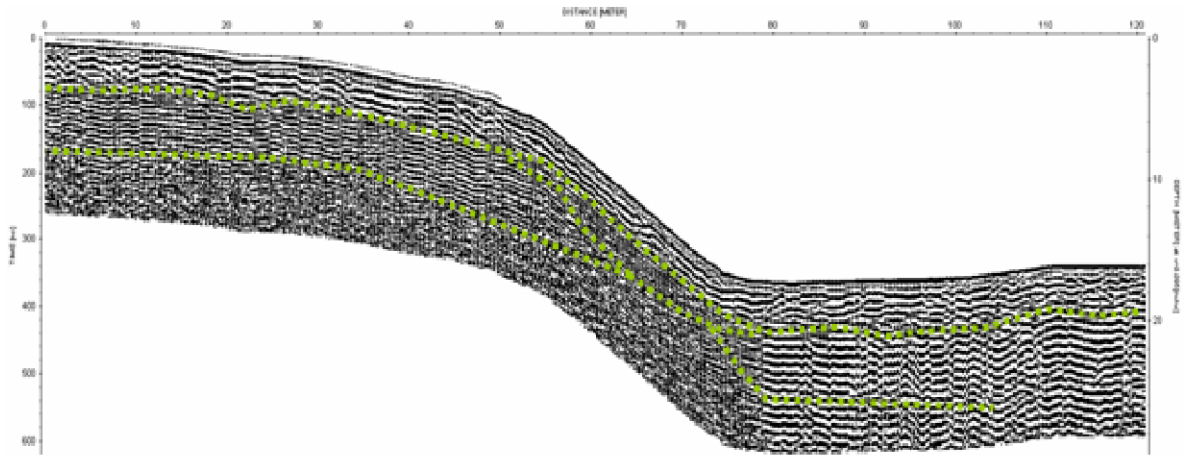
A.7 BM playa east GPR profile. This playa includes two parts (see fig. A.4a & c). The western part has a cover of light-colored silt with sparse cover of bushes. The eastern section is sandy and has approximately a 20% cover of bushes.

- Display of black-filled wiggles. Contiguous wiggles trace curves, generally with a black fill of the positive area, are one common display type of the amplitudes intensity field. This type of plot is also known in seismic reflection as variable area (VAR).
- Migration processing and bandpass filtering. Migration attempts to remove diffractions, distortions and out-of-line reflections. This makes the reflection profile resemble the geological structure. Bandpass filters the noise and preserves the character of the primary reflections (after Neal, 2004).
- Subsurface GPR interpretation based upon two drills. Note that the eastern drill did not identify any silts. The clear reflection surface in the processed images may be due to surface silt cover, and slight textural changes of silty sand to sand between the upper sediments.



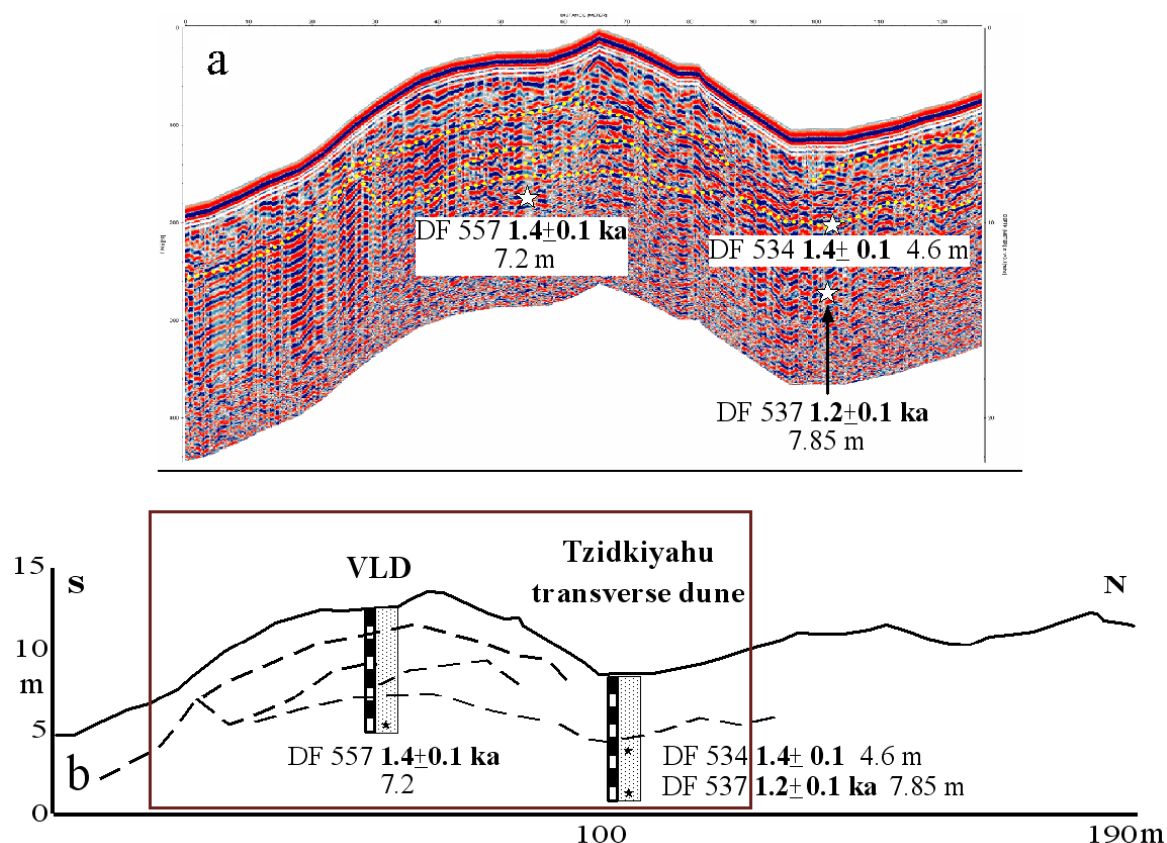
A.8 Tzidkiyahu site.

- Orthophoto of the Tzidkiyahu site. Here three drills and two GPR profiles were conducted.
- Photograph of the Tzidkiyahu site (looking west) and impressive eastern-facing transverse dune structure slip-face.



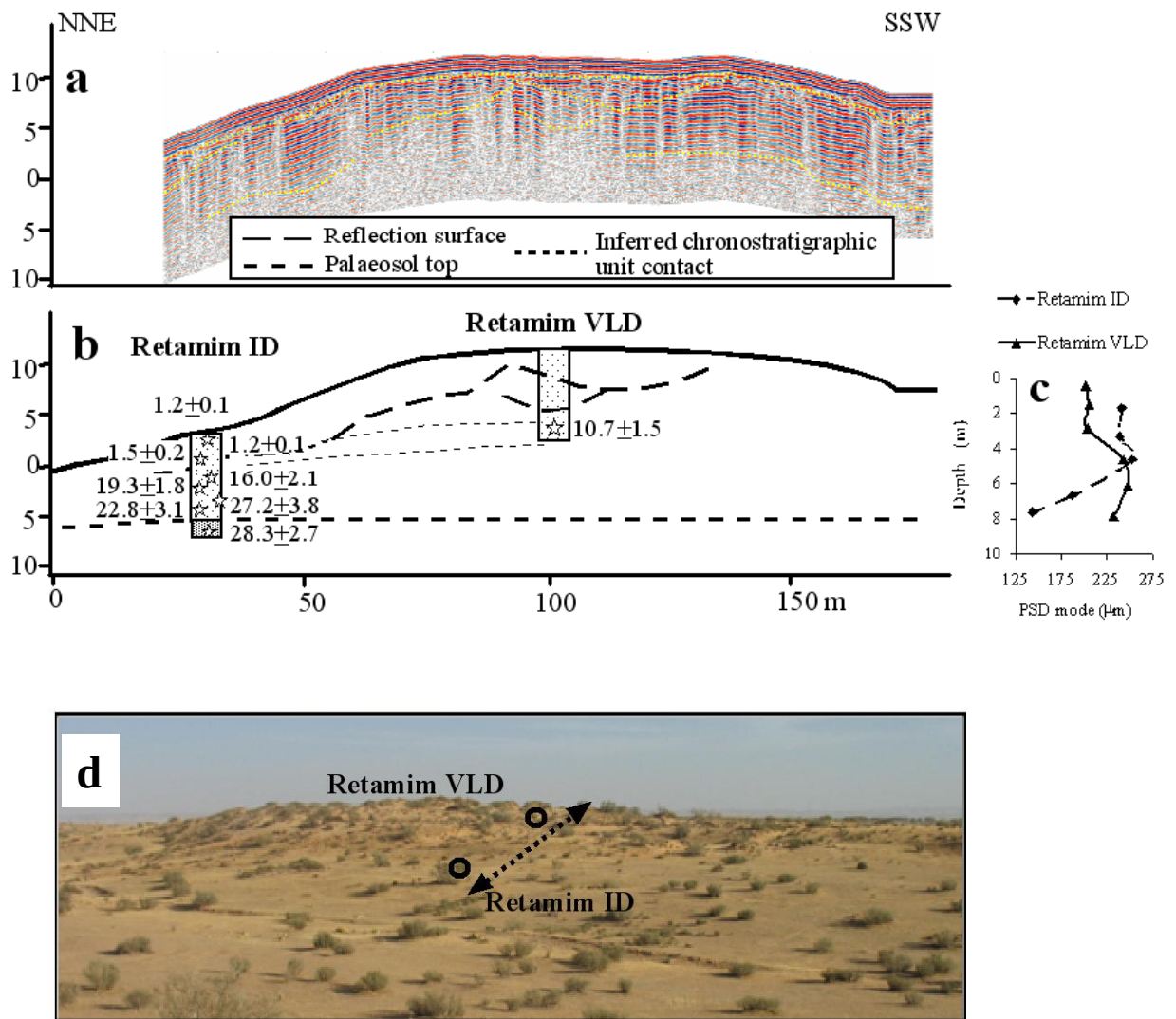
A.9. Tzidkiyahu site west-east GPR profiles.

- West-east GPR wiggly-display profile of the late Holocene transverse dune structure, its slip face and late Pleistocene aeolian sand that probably also underlay the transverse structure. Slip face penetration was very limited.
- Interpretation enhanced by OSL ages of the GPR display. At 7.8 meters depth and in the lower unit, an OSL age of 1.2 ± 0.1 ka was determined and sand at 4.6 meters in the middle unit was dated to 1.4 ± 0.1 ka. The slight age inversion which is barely within errors, may indicate rapid accretion events at successive or similar times. The upper radar facies unit of the transverse dune is probably recent, as indicated from the recent ages of 50 ± 3 yr and 68 ± 5 yr dated within the upper slip face unit. OSL ages for each unit of the aeolian sand reveal surprising and closely-fit similarity 15.9 ± 1.7 ka, 15.5 ± 0.9 ka and 15.8 ± 3.4 ka composing prime evidence of the rapidity of dune accretion. The calcic palaeosol substrate at 9.8 m does not induce identifiable reflection.



A.10. Tzidkiyahu site GPR profiles.

- North-south GPR (migration) profile of late Holocene VLD and south of it forming an interdune, the upper part of the transverse dune fill. The reflections of this profile are more vague than for the BM VLD profiles though the general pattern is similar. The transverse (interdune) section reveals poor reflections compared to the east-west profile that runs parallel to the dune advance direction. The similar (1.4 ka) ages at similar depths, possibly beneath the same reflection surface may suggest that a planar depositional surface formed then, possibly of a transverse like structure and later was superimposed by VLDs such as those dated to 0.5 and 0.9 ka at the BM site.
- North-south topographic profile and drill data. The dashed black box marks the upper figure A9a.



A.11 Retamim site GPR profiles.

- North-south GPR profile showing limited penetration and vague discontinuous contact surfaces.
- North-south topographic profile interpretation and drill data. The dashed line marks a suggested contact based upon OSL ages but not discerned in the GPR profile. The calcic palaeosol substrate at 7.2 m does not induce identifiable reflection.
- Particle-size distribution mode vs depth of the two Retamim profiles.
- Photograph of the VLD and the GPR profile northern part of the profile. Note the broad VLD morphology that differs from the VLDs updune by the border while is similar to dunes in the northwestern part of the dunefield.

B.1 Particle size analysis results.

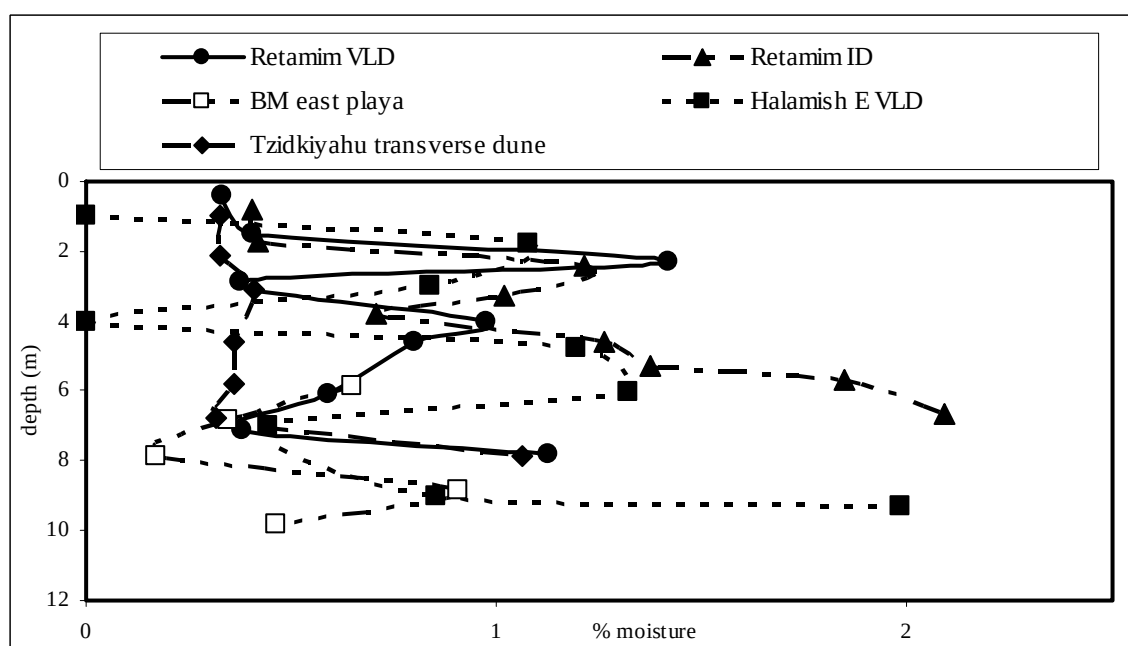
Incurson path/site/ sample	Setting	Depth (m.)	Sand (%)	Silt (%)	Clay (%)	Fines (Silt+ Clay) (%)	Mode (%)	Kurtosis	Skewness
Northern									
<u>Haluzit4</u>									
32	D	1.15	86.8	11.3	1.9	13.2	145.9	.129	.479
34	D	1.9	85.2	12.5	2.2	14.8	140.5	-	-
35	D	3.3	87.4	10.8	1.8	12.6	140.0	-	-
304	D	4.9	86.1	12.2	1.7	13.9	131.0	.395	.514
306	CS	6	20.2	60.2	19.6	79.8	14.0	-	-
308	CS	6.9	34.0	44.7	21.3	66.0	3.9	4.922	2.146
<u>Haluzit4 Hothouse</u>									
41	ID	0.6	80.9	16.4	2.7	19.1	130.0	-	-
42	ID	1.8	78.5	19.3	2.2	21.5	117.0	1.875	1.025
<u>Haluzit1</u>									
802	D	2.9	79.6	18.0	2.4	20.4	123.1	-	-
803	D	3.7	91.4	7.6	1.0	8.6	167.2	-	-
804	D	4.5	89.8	8.6	1.6	10.2	144.9	-	-
81	D	6.8	80.7	17.1	2.2	19.3	111.7	.145	.2
82	CS	7.2	77.4	20.2	2.4	22.6	119.6	-	-
83	CS	7.5	72.6	22.3	5.1	27.4	167.7	-.222	.402
85	CS	8	42.0	46.5	11.4	58.0	111.6	.906	1.19
<u>Baladiya</u>									
72	D	0.65	76.0	21.0	3.1	24.1	139.7	-.029	.577
74	D	1.75	86.6	11.2	2.2	13.4	150.4	-.088	.325
75	C	2.4	75.9	20.3	3.8	24.2	131.3	.014	.588
76	D	3.2	82.1	15.5	2.4	17.9	171.7	-.002	.585
714	D	5.7	91.2	7.0	1.7	8.7	202.0	.045	.136
719	D	9.8	79.7	17.8	2.5	20.3	150.0	.039	.608
720	D	10.25	73.9	23.0	3.1	26.1	152.0	-.186	.55
721	CS	0.3	68.0	28.2	3.8	32.0	118.0	.331	.796
Central									
KD 73 ID depression									
680	ID	1	59.2	33.4	7.4	40.8	198.2	.426	.997
681	ID	2	82.9	14.3	2.8	17.1	195.5	-.107	.497
682	ID	3	86.8	11.5	1.8	13.2	211.0	.805	.838
683	ID	4	94.0	5.4	0.6	6.0	331.0	-.168	.364
684	ID	4.5	90.6	8.4	1.0	9.4	189.0	.286	.685
685	ID	6	86.0	12.3	1.7	14.0	265.1	-.261	.369
687	CS	6.2	40.6	44.7	14.7	59.4	78.3	4.345	2.13

<u>KD 73</u>									
690	D	1.5	88.6	9.8	1.6	11.4	172.0	.167	.6
692	D	4.5	93.0	6.0	1.0	7.0	162.5	.322	.138
695	D	9.2	94.0	5.3	0.8	6.1	213.0	.107	.244
<u>MM</u>									
11	D	1.25	89.4	9.0	1.6	10.6	145.5	.052	-.031
13	D	2.6	90.9	7.9	1.2	9.1	175.7		
16	D	5.7	90.7	7.7	1.6	9.3	164.8	.08	.181
17	D	7	86.5	11.8	1.7	13.5	147.2	-.087	.243
18	ID	1.05	65.0	31.2	3.8	35.0	106.7	1.46	1.144
<u>Retamim dune base</u>									
541	ID	1.7	89.7	9.1	1.2	10.3	241.1	-.2	.344
543	ID	3.3	94.2	5.1	0.7	5.8	239.3	.26	.251
545	ID	4.6	96.5	2.9	0.7	3.5	252.3	.937	-.024
548	ID	6.65	91.7	6.4	1.9	8.3	186.7	.166	.03
700	ID	7.6	82.9	13.7	3.4	17.1	143.8	-.121	.433
701	ID	7.65	91.1	7.2	1.7	9.0	218.0	.368	-.161
702	ID-CS	7.75	73.6	22.5	4.0	26.5	99.8	1.531	.861
704	CS	7.95	63.7	31.8	4.5	36.3	94.3	-1.48	.45
<u>Retamim VLD</u>									
560	D	0.4	87.0	11.7	1.4	13.1	201.7	-.173	.472
561	D	1.5	87.6	11.2	1.2	12.4	205.7	-.192	.484
563	D	2.85	91.2	7.7	1.2	8.8	203.0	-.008	.254
565	D	4.6	93.4	5.9	0.7	6.6	243.0	.514	-.044
566	D	6.1	94.9	4.5	0.7	5.1	248.0	.43	.182
568	D	7.85	95.6	3.7	0.7	4.4	232.0	.863	-.114
<u>Ramat Beqa</u>									
580	AS	3.1	91.1	7.6	1.3	8.9	203.0	-.112	.205
579	AS	4.3	85.3	13.1	<u>1.7</u>	14.8	188.3	-.343	.22
578	AS	4.85	93.0	5.1	<u>1.1</u>	6.2	211.1	.556	-.161
577	AS	4.9	86.1	11.7	2.3	13.9	224.6	-.383	.172
576	AS	5.2	83.4	14.1	2.6	16.6	207.8	-.368	.23
575	AS	8	78.1	19.2	2.7	21.9	132.2	-.063	.51
<u>Nahal Sekher VI</u>									
NS-1	AS	0.5	91.0	7.8	1.1	8.9	192.0	.088	-.085
NS-2	AS	0.75	90.9	7.7	1.4	9.1	191.0	.197	-.066
NS-3	AS	1.5	89.6	8.9	1.5	10.4	205.6	.118	-.076
NS-4	AS	2.65	91.8	7.2	1.0	8.2	199.0	.377	-.207
NS-5	AS	1.6	89.9	8.4	1.8	10.2	189.7	-.21	.211

NS-6	AS	1.8	91.8	6.7	1.4	8.1	194.3	.147	-.004
NS-7	AS	2	92.8	6.1	1.1	7.2	191.6	.312	-.004
NS-10	FID	0.3	65.0	29.0	6.0	35.0	164.0	-.55	.52
NS-11	AS	0.45	89.6	8.6	1.8	10.4	196.5	.222	-.23
<u>Tzidkiyahu</u> <u>VLD</u>									
550	D	1.2	93.6	5.4	1.0	6.4	185.0	.092	.227
554	D	3.8	95.2	4.0	0.8	4.8	189.9	.121	.274
557	D	7.2	95.6	3.9	0.5	4.4	203.4	.615	-.16
<u>Tzidkiyahu</u> <u>Transverse</u>									
534	D	4.6	94.7	4.6	0.7	5.3	187.8	.338	.14
537	D	7.85	90.8	7.9	1.3	9.2	207.5	-.055	.404
<u>Tzidkiyahu</u> <u>ID depression</u>									
521	ID	2.5	94.4	4.9	0.7	5.6	203.0	.67	-.089
524	ID	4.5	93.0	6.2	0.8	7.0	201.0	.471	-.141
660	ID	10.25	96.0	3.5	0.5	4.0	231.5	.234	.424
662	CS	11.1	54.2	31.1	14.7	45.8	205.5	.366	1.039
<u>BM west</u>									
508	D	3.1	92.8	6.3	1.0	7.2	215.6	-	-
509	D	5.5	93.7	5.5	0.8	6.3	218.3	.16	.066
510	D	7.7	87.3	11.2	1.5	12.7	227.1	-	-
511	D	9.8	86.9	11.6	1.5	13.1	224.3	-.304	.339
110	D	4	91.0	8.0	1.0	9.0	245.2	-.106	.187
111	D	4.7	86.5	11.8	1.7	13.5	250.5	.023	.25
<u>BM east</u>									
512	D	3.1	93.4	6.0	0.7	6.6	245.0	.166	.157
513	D	4.5	94.0	5.4	0.0	5.4	248.7	-.01	.307
514	D	6.3	83.9	14.4	1.7	16.2	226.4	-.214	.556
515	D	7.6	85.4	12.9	1.6	14.6	235.4	-.275	.464
<u>BM</u> <u>depression</u>									
505	ID	5	90.3	8.8	1.0	9.7	219.3	-.114	.305
506	ID	6.5	91.3	7.4	1.3	8.7	217.3	.501	-.163
507	CS	6.85	44.6	39.8	15.6	55.4	93.7	4.989	2.081
<u>BM east</u> <u>playa</u>									

501	D	5.1	92.0	7.0	0.7	7.7	230.0	.23	.017
485	D	5.85	92.7	6.6	0.7	7.3	239.0	.302	.108
487	D	7.85	92.1	7.0	0.9	7.9	273.5	-	-
489	S	9.85	66.0	28.1	5.9	34.0	201.0	-.176	.738
491	S	10.4	71.7	23.6	4.7	28.3	216.3	-	-
493	S	11.3	63.8	30.3	5.9	36.2	120.6	1.327	1.235
Southern									
<u>Halamish trench</u>									
617	FID	1.2	0.0	68.0	32.0	100.0	4.2	8.528	2.448
618	D	1.2	85.2	12.9	2.0	14.9	242.3	-.269	.392
631	FID	2.2	9.0	70.0	21.0	91.0	4.6	42.375	5.61
<u>Halamish ID</u>									
620	FID	0.1	0.0	74.1	25.9	100.0	4.0	3.615	1.811
621	FID	0.35	58.3	38.2	3.6	41.7	84.9	-.041	.563
622	FID	0.45	0.0	82.0	18.0	100.0	21.0	1.437	1.365
623	FID	0.7	56.9	38.6	4.5	43.1	86.0	-.405	.363
624	FID	1	0.0	72.5	28.5	101.0	4.0	5.173	2.12
625	FID	1.25	73.2	22.9	3.9	26.8	174.0	-.308	.439
626	FID	2.3	72.9	22.9	4.2	27.1	184.2	.973	1.079
<u>Halamish East</u>									
581	D	1	89.3	9.0	1.7	10.7	218.0	-.17	.375
583	D	3	90.4	8.4	1.2	9.6	172.0	-.068	.509
585	D	4.8	92.5	6.3	1.2	7.5	230.9	-.068	.245
632	D	9.4	79.1	18.1	2.8	20.9	188.0	.372	.886
633	FID	9.5	58.2	36.1	5.7	41.8	76.6	6.136	2.196
<u>Nizzana Reservoir</u>									
516	AS	2.3	70.0	25.9	4.1	30.0	93.6		
518	AS	3.7	74.1	22.3	3.6	25.9	123.7	.744	.957
<u>Beer Malka</u>									
1	D	3	94.1	5.3	0.6	5.9	419.0	1.338	-.53
4	D	3	92.0	7.3	0.7	8.0	285.3	.556	-.387
<u>Mitvakh</u>									
202	D	0.5	82.4	15.2	2.5	17.6	213.0	-.308	.439
201	C	5	84.0	14.0	2.4	16.4	217.0	-.349	.373
200	D	9.25	79.1	17.8	3.2	21.0	105.4	-.307	.517
<u>Besor terrace</u>									
639	AS	3.5	76.3	20.7	3.0	23.7	189.0	-.315	.491

B.2 Sand moisture profiles for selected dune and interdune sand sections.



Sample No. (DF)	Peak Height Quartz 20.8°	Peak Height Plagioclase 27.9°	Peak Height Calcite 29.4°
3	12.9	0.8	2
13	12.3	4.7	0.3
16	11.6	2.5	0.6
17	15.3	2.4	0.8
32	15.8	4.9	1.2
34	21.6	4.6	1.4
35	9.8	2.8	1.2
42	14.2	3.9	1.9
74	11.2	3.5	0.8
75	11.4	3.1	1.2
76	14.8	5	0.8
111	17.7	5.9	1.8
200	13.3	1.3	5.6
522	14.2	3	0.7
526	18.2	2.6	0
530	16.3	1.9	0.5
534	16.8	2.4	0
537	18.6	1.6	0.6
554	21.3	5.5	0.6
557	24.1	4.6	0
565	13.9	1.2	0.6
568	11.3	2.7	0.4
571	19.9	2.8	9.6
578	19.5	2.9	0.9
580	15.8	4.1	1.1
586	15.3	1	3
589	12.9	4.8	5.1
602	15.2	1.3	3.5
604	19.9	3.5	4.6
1004	11.3	1.5	1.1
506	14.7	0.8	2.3
513	12.8	1.8	1.5
515	17.8	2.9	3.4
509	14.9	3.2	1.2
511	10.1	4.6	1.5

B.3 X-Ray diffraction (XRD) mineralogy results of the NW Negev dunefield sands.

C.1 Pre-Processing of Landsat Images – Radiometric and Atmospheric Corrections.

Images were corrected using a Dark Object Subtraction (DOS) method following Equation 1 (Chavez, 1996, Song et al., 2001):

$$\text{Eq. A1.} \quad \rho = \frac{\pi d^2 (L_{sat} - L_p)}{T_v (E_0 \cos(\theta_z) T_z + E_{down})}$$

where:

ρ – Surface reflectance

d^2 – Sun-earth distance in astronomical units (Chander et al., 2009)

L_{sat} – At-satellite radiance

L_p – The path radiance

E_0 – Exoatmospheric solar irradiance (Chander et al., 2009)

T_v – Atmospheric transmittance from the target toward the sensor.

T_z – Atmospheric transmittance in the illumination direction

E_{down} – Downwelling diffuse irradiance (estimation explanation below).

θ_z – Solar zenith angle (calculated for the solar elevation given in the image header file)

Due to the atmospheric scattering effects, the dark object is not absolutely dark. Assuming 1% surface reflectance for the dark objects (Song et al., 2001, Chavez Jr., 1988), the path radiance is estimated as

$$\text{Eq. A2.} \quad L_p = \left(\frac{L_{max\lambda} - L_{min\lambda}}{Q_{cal\max} - Q_{cal\min}} \right) \times DN_{min} + L_{min} - 0.01 (E_0 \cos(\theta_z) T_z + E_{down}) T_v / d^2 \pi$$

Where:

$L_{min\lambda}$ – Spectral at-sensor radiance that is scaled to $Q_{cal\min}$ [W/(m²sr μm)]

$L_{max\lambda}$ – Spectral at-sensor radiance that is scaled to $Q_{cal\max}$ [W/(m²sr μm)]

$Q_{cal\ min}$ – Minimum quantized calibrated pixel value corresponding to $L_{min\ \lambda}$ [DN]

$Q_{cal\ max}$ – Maximum quantized calibrated pixel value corresponding to $L_{max\ \lambda}$ [DN]

DN_{min} Selection was conducted according to Mauz (http://arsc.arid.arizona.edu/resources/image_processing/landsat/minimum-dn.html).

Several versions of the DOS models based on Equation 1 are reported in the literature and summarized in Table 1:

Table A1: Parameters comparison of DOS models.

Method	T_v	T_z	E_{down}
Basic DOS	1	1	0
COST model (Chavez, 1996)	1	$\cos(\theta_z)$	0
DOS3 model (Song et al., 2001)	$e^{-\tau_r/\cos(\theta_v)}$	$e^{-\tau_r/\cos(\theta_z)}$	Rayleigh (6S)
Vogelmann-DOS3 model (Paolini et al., 2006)	1	$e^{-\tau_r/\cos(\theta_z)}$	0

DOS1 assumes no atmospheric transmittance loss (T_v and T_z to be unity), and no diffuse downward radiation at the surface (E_{down} to be zero) in equation 1 (Chavez, 1988). DOS2 approximates T_z by $\cos(\theta_z)$ for TM 1–4, and unity for TM 5 and 7. Chavez (1996) showed that, for most acceptable images with atmosphere optical depth between 0.08 and 0.3, and solar zenith angle between 30° and 55°, transmittance in the illumination direction can be approximated, to a first order, by the cosine of solar zenith angle. DOS3 computes T_v and T_z assuming Rayleigh scattering only, that is, no aerosols (Song et al., 2001). The optical thickness for Rayleigh scattering (τ_r) is estimated in Equation 3 (Kaufman, 1989) as:

$$\text{Eq. A3.} \quad \tau_r = 0.008569\lambda^{-4} (1 + 0.0113\lambda^{-2} + 0.00013\lambda^{-4})$$

E_{down} for Rayleigh atmosphere was estimated by 6S (Vermote et al., 1997) for zero aerosol optical depth at 550nm.

For the current research DOS3 was chosen based on Song et al. (2001) who compared several absolute atmospheric correction methods and a relative radiometric normalization. They

conducted their evaluation based on classification and change detection tests how well a multi-temporal dataset was brought to a common scale by an atmospheric correction, and not how absolutely accurate the obtained surface reflectance was. According to their findings the best accuracies were obtained using the DOS3 model (although the basic DOS and a relative radiometric normalization were also acceptable).

Further support is given by Paolini et al. (2006) who compared a DOS correction to a relative radiometric normalization method using a similar classification accuracy assessment approach. Their recommendation is to use the DOS model when certain data is available because it leads to consistent results and is less complex and less time consuming comparing to the relative radiometric normalization. The required data includes sensor characteristics, illumination and observation geometry, estimations of atmospheric components (i.e. path radiance and molecular absorption), and the relationship between gain and offsets of the sensors. Since this data is available for all Landsat scenes it poses no restriction. Further, it is more recommended for use with a long time series since as the total number of images increases, this also increases the difficulty in finding common Pseudo-Invariant Features necessary for the normalization (Paolini et al., 2006).

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C.2 Resampling ASD RI to Landsat TM5 spectral resolution.

The Landsat 7 spectral response file was used to calculate a weighted average for one sample.

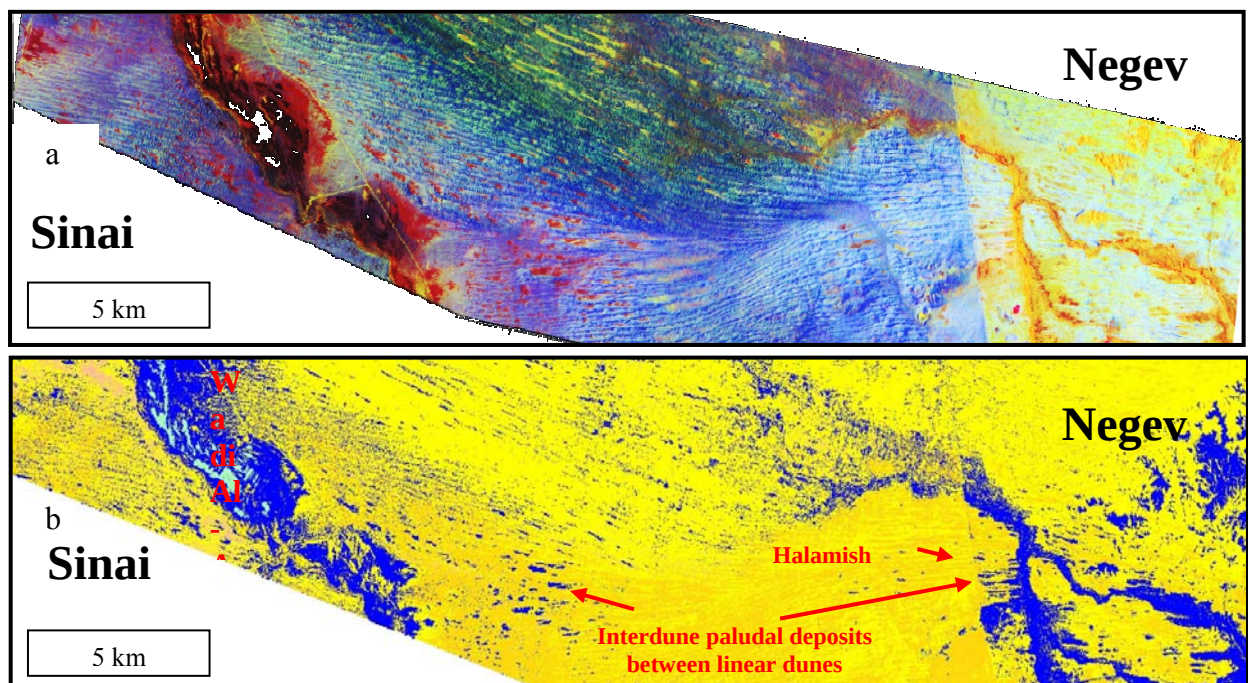
The result was compared it to ENVI built-in resampling. The results significantly differed.

A numerical integral instead of average (in excel - using trapezoid area calculations) was also calculated though it seems that it is not suited for reflectance units (opposed to radiance. Thus

ENVI resampling was chosen. ENVI help: "...ENVI assumes critical sampling and uses a

Gaussian model with an FWHM equal to the band spacings..." - meaning averaging assumed a step function for ASD bands and ENVI assumes Gaussian.

C.3 Landsat TM5 (2003) of the southeastern dunes of northern Sinai and the southwestern part of the NW Negev dunefield. a. mineral composite spectral enhancement (band5/band7, band5/band4, band3/band1). Paludal deposits are shown in yellow and red. b. Thematic map. The Negev paludal loams (in blue) are highly calcareous suspended detritus originating from upstream highland carbonate strata. This, along with their flat appearance and contrast with surrounding quartz dunes (yellows) gives them a distinct spectral-mineralogical signature in the Negev and along W. Al-Arish. Previously attributed to neotectonics (Kusky and El-Baz, 2000), the Sinai loams are suggested to have been deposited due to dune damming during the late Pleistocene main dune incursion.



ולאו דוקא באקלים צחיח. כמויות המשקעים הרבות אפשרו מחד גיסא, תנאים מוטבים להתפתחות צמחיה בדיונות ומאידך גיסא, הרוחות שנלוו לסערות הגבירו את הסחיפה וצמצמו את הכיסוי של הצמחייה. העדר ממצא אודות פלישת דיונות לפני כ-24 א"ש בנגב כנראה נובע מהעובדה שהדיונות טרם התקדמו עד הנגב מסיני, למרות שבתקופות קרחוניות קדומות סביר שהיה משטר אקלימי ורוחות מספקות להנעת חול ופלישת דיונות.

דיונות אורכיות עם צמחיה שתוארכו בשיטות לומינסנציה, בעיקר מההמיספרה הדרומית פעילות בעיקר משיא הגלציאל האחרון (LGM) ומתייצבות סמוך לתחילת ההולוקן. בדומה לסינכרוניות של דיונות הנגב לאירועי הקור של ההינריך 1 והיאנגר דריאס, גילי לומינסנציה העולמית של דיונות אורכיות עקביות עם הנפילה החדה של כמות האבק בגלעיני קרח של שתי ההמיספרות. אי לכך, מוצע שהשינויים בהספקי הרוח בין שיא הגלציאל לתחילת ההולוקן קבעו הן את פעילות הדיונות האורכיות עם צמחייה והן את התייצבותן.

בהתבסס על הקשר בין פעילות דיונרית גלובלית בקווי רוחב נמוכים ועל רוחות בעלות אנרגיות גבוהות הקשורות לתקופות הקרחוניות ואירועי קור גלובליים, לא נצפה בפלישה מחודשת של דיונות הנגב בהיקף של הפלייסטוקן המאוחר בעקבות בצורות, וגם לא של דיונות דומות באזורים אחרים. כיום, סערות חורף המגיעות מהים התיכון והפוקדות את צפון הנגב מפעילות רק את חלקן העליון של הדיונות כפי שנמדד, תוארך וגם נצפה בחתכי המח"ק. אי לכך, סערות חורף חזקות עשויות לגרום להתארכות דיונרית מועטת של יחידת חול דקה.

תבנית ההשקעה הבלתי-רציפה של חול בדיונות אורכיות בעלות צמחייה בסוף הרביעון מספקת תובנות חדשות וחשובות אודות תהליכי ניידות והתייצבות דיונות וממחישה את רגישות הדיונות בשולי רצועת המדבריות לשינויי אקלים ואספקת סדימנט. הקשר המוצע בין הפחתה ברוחות חזקות בסוף תקופות של אירועים קרים גלובליים והתייצבות דיונות בקווי הרוחב הנמוכים מדגיש את מרכזיות השפעת הרוחות על אפיזודות ניידות ופלישת דיונות גם כשהם בעלי כיסוי של צמחייה.

אפיזודת ניידות דיונות בסוף ההולוקן (2-0.8 א"ש) כללה לסירוגין, פלישת דיונות אורכיות ורוחביות בחלקו המערבי של הפרוזדור המרכזי וניידות חול על פני כל שטח שדה הדיונות תוך הצטברות שכבה של חול בעובי של 1-2 מטרים במזרח שדה הדיונות. פעילות איאולית זה כללה הפעלה מחדש של חולות קדומים יותר. מכיוון שתקופה זו תואמת לתקופה הרומית, הביזנטית והערבית הקדומה, יתכן שהדיונות נחשפו לארוזיה ולפעילות איאולית בשל לחץ האדם שהביא להסרה של הצמחייה והקרומים המייצבים. חתכי הדיונות העבים בחלקו המערבי של הפרוזדור המרכזי ומספר תתי-קבוצות של הגילים מצביעים ככל הנראה על מספר אירועים של עוצמות רוח גבוהות שאפשרו הסעת חול משמעותית וזמינות חול מרובה.

חילופין של יציבות ופעילות איאולית של חול ב-150 השנים האחרונות הפעיל מחדש מדרונות ורכסי הדיונות של הנגב אך הדיונות לא התארכו. הגילים מתקבצים לתקופות בהם שימושי הקרקע הוגברו כגון: פעילות דיונות מוגברת עקב נוכחות בדואים ורעייה שהופסקה בתחילת שנות ה-50, שהחלה מחדש בשנת 1967 והופסקה שוב בעקבות פינוי סיני כחלק ממימוש הסכם השלום עם מצרים ב-1982.

אדמומיות החול, המוגדרת ע"י אינדקס ההתאדמות (Redness Index; RI) $(RI=R^2/(B \cdot G^3))$, המשקפת את כמות ציפויי תחמוצות ברזל על גרגרי הקוורץ, לא השתנתה לאורך אזורי הפלישה השונים ולעומק חתכי הדיונות והחולות הבין-דיוניים. לא נמצא מתאם בין עוצמת ה-RI לגילי הלומינסנציה של החול. עבור רוב החתכים הסטרטיגרפיים, ערכי ה-RI והמאפיינים הסדימנטולוגיים של החול היו די אחידים.

שתי תרחישים מוצעים להתאדמות החול. אחת, התאדמות החול התרחשה במהירות לאחר השקעתה בסוף הפלייסטוקן באקלים שלפי עבודות קודמות, היה לח יותר (כ-300-350 מ"מ/שנה משקעים). מכיון שאין עדויות וקשה למצוא הסבר לתרחיש זה, נראה שהתפתחות ציפויי גרגרי החול וכתוצאה מכך התאדמות החול היתה מינימלית מאז השקעת החול בנגב. לפי ניתוח דוגמאות חול מצפון סיני, חישה מרחוק ועבודות קודמות, מוצע שתכונות ה-RI של גרגרי החול הם ירושה ממקורות ים החולות בסיני או בדלתת הנילוס. מוצע שציפויי גרגרי החול הם תופעות דיאגנטיות וצבעם היה אדום במידה דומה מאז יציאתם האיאולית בסוף הפלייסטוקן מחלקה המרכזי והעליון של דלתת הנילוס. ממצאים אלו יחד עם גילי פחמן של יחידות חול של הדלתא מסוף הפלייסטוקן ואתרים פרה-היסטוריים מצפון סיני, מספקים את המידע הזמין אך החלקי שמצביע על כך שזמינות חול דיוני בצפון סיני הותנע בעקבות חשיפת חולות נחליים של הדלתה, בעיקר עקב ירידת מפלס ים התיכון בתקופות קרחוניות.

פעילות הדיונות בפלייסטוקן המאוחר הן תוצאה של הספקי רוח גבוהים בהרבה מאלה של היום בהתאם למודלים קיימים. משקעים יחד עם עצמת רוחות, הוליכו אבק עד חול דק לצפון הנגב. בהינתן זמינות חול, מודל אקלימי זה אינו מסביר את האפיזודות והאירועים הקצרים של פלישת חול לנגב שנמצאו בו-זמנית עם אירועי הקור של ההיינריך 1 והיאנגר דריאס. בדומה לגלציאל האחרון, אירועי קור אלו גרמו להשקעת אבק מוגברת בגלעיני קרח של ההמיספרה הצפונית. השקעה מוגברת זו מוסברת על-ידי קצבי סחיפה מוגברים מרצועת המדבריות העולמית בקווי רוחב הנמוכים. זאת ע"י עוצמות רוח מוגברות עקב גרדיאנטים מרידיאנליים מועצמים. עוצמות הרוח המוגברות אירעו בין השאר במסגרת סערות חורפיות מעל מזרח ים התיכון במשך ימים מספר. מכיוון שעוצמות הרוח הדרושות לסחיפת אבק גדולות מאלו להנעת חול בסלטציה, אנו מציעים שדיונות הנגב הגיבו והתארכו בעקבות הספקי רוח גבוהים בקווי הרוחב הנמוכים. כיווני הדיונות מצביעים על רוחות מערביות אשר שונים מהרכיב הדרום-מערבי הדומיננטי כיום. ממצא זה עשוי להצביע על כך שכיווני הרוחות בוני-הדיונות בסוף הפלייסטוקן היו מעט שונים מאלו של היום. יוצא דיונות צפון-מערב הנגב התארכו באקלים לח בעל הספקי רוח גבוהים בהרבה מזה של היום.

גילי הלומינסנציה (OSL) שנקבעו בפרוטוקול ה-SAR נמצאו אמינים להערכת גילי החולות והסדימנטים של צפון-מערב הנגב. בדיקות סטנדרטיות של התנהגות הלומינסנציה של גרגרי הקוורץ שכללו שיקום מנת הקרינה, פיזור-יתר, וחישוב יחסי המחזוריות של הפרוטוקול, הראו שהסדימנטים מתאימים לפרוטוקול ה-SAR.

התפלגות גילי ה-OSL של החול מצביעים על שלש קבוצות גיל: 10-24 אלף שנה (א"ש) לפני זמנינו (ל"ז), 0.8-2.0 א"ש ל"ז ו-10-150 שנים ל"ז. קבוצות גיל אלו מתאימות לחלוקה הכרונו-סטרטיגרפית של חתכי צירי הדיונות ומיצגים את האפיזודות המרכזיות של ניידות ופלישות דיונות אורכיות (VLD). תיארוך חתכים חשופים בעזרת OSL הניב תובנות אודות דינמיקות וקצבי התעבות והתארכות דיונות הנגב. צפיפות מרחבית (בשדה הדיונות) ואנכית (בחתכים) של הגילים אפשר תיארוך אמין של אפיזודות הפלישות. חולות בעמקים הבין-דיוניים עברו מעט התעבות והצטברות חול והדיונות עברו מעט תנועה לטרלית מאז היווצרותן. הדיונות האורכיות עם צמחייה צוברות לרוב חול לאורך צירן באפיזודות ניידות חול מרכזיות. בתהליך זה, תחילה מוסר חול מחלקה העליון של הדיונה ואחרי-כן מושקע חול, דבר הבונה חתך סטרטיגרפי הכולל מספר יחידות חול עם מגעים אופקיים.

כל אפיזודה של ניידות חול עשויה להכיל מספר אירועים קצרים יותר. בכל אפיזודה, נוצרת יחידת סטרטיגרפיה המייצגת זמן מתחילת ניידות החול ועד התייצבות החול. ליתר דיוק, החולות התחתונים של כל יחידה מאוחרות קמעא מזמן תחילת אירוע הפלישה והקבורה של החול, וגג היחידה לרוב צעיר במקצת מגיל ההתייצבות. בין האפיזודות, הדיונות לא מתארכות, אך פעילות חול מקומית הכוללת הסעה, השקעה והסרה מתרחשת, בעיקר ברכסי הדיונות. אלה, משפיעים על הכיסוי הצמחי והקרום הביולוגי ומאפסים את סיגנל הלומינסנציה של גרגרי החול.

בראייה גלובלית, שדה הדיונות של צפון-מערב הנגב כבר לפני כ-100 א"ש ל"ז אך הדיונות הראשונות פלשו לפינה הדרום-מערבית של שדה הדיונות רק בכ-24 א"ש ל"ז. אפיזודת הפלישה המרכזית התרחשה בין 18 ל-11.6 א"ש. מרבית הגילים מתרכזים ב-13.7-16 א"ש ו-11.6-12.4 א"ש ל"ז. תקופות אלו שוות-זמן עם אירועים של ירידה בטמפרטורה בהמיספרה הצפונית המוכרים כאירועי קור (cold-events) "היינריך 1" (Heinrich 1) וה"יאנגר דריאס" (Younger Dryas), בהתאמה. באפיזודה של 13.7-16 א"ש מרבית החול והדיונות פלשו והושקעו בצפון-מערב הנגב. גופי חול מאסיביים שקעו בחלק המרכזי. ב-11.6-12.4 א"ש הדיונות הגיעו לקצה המזרחי של שדה הדיונות. קצבי התקדמות דיונות נמצאו שונות בין החלקים והאפיזודות: כ-25 מ"שנה בחלק הצפוני ב-13.7-16 א"ש ועד כ-5-10 מ"שנה ב-11.6-12.4 א"ש במזרח שדה הדיונות. הערכות אלה מתייחסות לטווחי השגיאה של גילי הלומינסנציה ואי לכך, על-פי סטרטיגרפיה, נראה שהאפיזודות כללו אירועי ניידות קצרים ומהירים.

פלישת דיונות בדרום שדה הדיונות חסמה את מרבית הוואדיות, דבר שגרם בעקבות שטפונות, להתפתחות גופי מים עומדים אשר השקיעו משקעי חול, ובעיקר סילט וחרסית בהירים. בין יחידות אלו לבסיס הדיונות התגלו אתרים ובעיקר חניונים אפי-פליאוליתיים, דבר המצביע על קשר בין גופי המים לפעילות האדם הקדמון. מיפוי פני השטח מבחינה ספקטרלית בעזרת מיון מונחה (supervised classification) של הדמאות לנדסט (Landsat) (משנת 2003) אשר עובדו בעזרת אינדקס מינרלים, מצביע על כך שקיים דמיון בין משקעים אלו בנגב לבין סביבות ואדי-אל-עריש בצפון סיני. אי לכך מוצע שואדי אל-עריש נחסם בעקבות אפיזודת הפלישה המרכזית שהגיעה לנגב, דבר שגרם כבנגב, להתפשטות גופי מים עומדים במעלה הוואדי ובין הדיונות האורכיות השוכנות לצידי הוואדי, ולהשקעת משקעים.

תקציר

המחקר על אקלימי העבר (פליאו-אקלים) מבוסס על גלעיני קרח, גלעיני משקעים ימיים ואגמיים ומשקעי מערות מתמקד במשקעים וטמפרטורה. דיונות משמשות ארכיב פליאו-אקלימי ייחודי גם לאנרגיות הרוחות, דהיינו הספקי הרוח. שדות דיונות (חוליות) המורכבות בעיקר מדיונות אורכיות המכוסות צמחיה (Vegetated Linear Dune - VLD) יושבות בקווי הרוחב המרכזיים (מרכז אסיה) והנמוכים בעיקר של ההמיספירה הדרומית (באוסטרליה, אפריקה הדרומית ודרום אמריקה). הדיונות לרוב בעלות כיסוי חלקי של צמחיה וקרומים ביוגניים ומתארכות בצורה קווית. בין הדיונות ישנם פרוזדורים (עמקים) בין-דיוניים. דיונות אלו אמנם יציבות היום, אך בעבר היו פעילות כאשר הכיסוי הצמחי היה דליל.

דיונות אורכיות מיוצבות המכסות שטח של 1,300 ק"מ² יוצרות את שדה הדיונות של צפון-מערב הנגב. שדה זה מהווה את הקצה המזרחי של ים החולות צפון סיני – צפון-מערב הנגב המתפרס מאזור דלתת הנילוס דרך צפון סיני עד רמת חובב בצפון הנגב, במקביל לרוחות הסערה השליטות ממערב ודרום-מערב.

מחקר זה מזהה את המאפיינים העיתיים והמרחביים של פלישת הדיונות מצפון סיני לצפון-מערב הנגב, התייצבות הדיונות, הסביבות ואקלימי העבר שהניעו והפסיקו פלישות אלו. בנוסף נחקרו גם היבטים סדימנטולוגיים: היחס בין מורפולוגיית הדיונות וצבע חול הדיונות לגיל החול בשיטת הלומינסנציה (Optically Stimulated Luminescence - OSL), משמעות צבע החול וישום מכ"ם חודר-קרקע (מח"ק GPR); למיפוי המבנה הפנימי של הדיונה.

במסגרת בניית תכנית הדגימה, השדה מופה בשיטות חישה מרחוק וחישה מקרוב. בעזרת מיפוי וניתוח סריגי גבהים נמצאה שונות בגיאומטריית החתך טופוגרפי של הדיונות. פרוזדורים (עמקים) בין-דיוניים נמצאו חוליים עד לעומק של כ-10 מטר. בהתאם לנ"ל, שדה הדיונות סווג לכתריסר יחידות דיוניות - גיאומורפיות. אלו מוזגו לשלשה פרוזדורי פלישת דיונות מקבילים: צפוני, מרכזי ודרומי, שבכולן הדיונות בכיוון כללי מערב-מזרח. אתרי דיגום עיקריים נבחרו בקפידה בקצוות המערביים והמזרחיים של השדה. זאת לצורך איסוף מקיף של מידע אודות מאפייני החול והדיונות ולהערכת גילי וקצבי התקדמות הדיונות. פרופילי מח"ק באורך מצטבר של כמעט 1 ק"מ של חתכי דיונות ומשקעים בין-דיוניים שנאספו בעזרת אנטנה 100 MHz, חדרו ושיקפו 4-8 מ' בלבד של תת-הקרקע ואף לא שקפו את תשתית הדיונות. אי לכך, מח"ק נמצא כאמצעי לא יעיל לזהות יחידות חול עבור דיגום עלי ידי קידוחים.

בסך הכל נותחו ונדגמו לאנליזות כ-40 חתכים מכ-20 אתרים. אלו כללו קידוחים ידניים של דיונות והפרוזדורים בין-דיוניים וחתכים חשופים שהניבו מידע רב אודות הסטרטיגרפיה הדיונית. מהם נבדקו עשרה חתכים בין-דיוניים שכללו חילופין של חול איאולי עם משקעי מים-עומדים ומשקעים נחליים. מעל 300 דוגמאות נאספו ונמדדו ספקטרוסקופית לצורך כימות עוצמת התאדמות החול; 118 דוגמאות נבדקו עבור התפלגות גודל-גרגר, 97 דוגמאות תוארכו בשיטת הלומינסנציה (OSL), ו-36 דוגמאות נבדקו מבחינה מינרלוגית.

תשתית הדיונות, הינן קרקעות קלציות עתיקות בהרכב סיין חולי (20-72% חול) עד סילטי. דיונות הנגב עצמן נעדרי קרקעות עתיקות (פליאוסולים) ורק בשתי דיונות נמצאו תצבירי קלציום קרבוט. חולות הדיונות והפרוזדורים הבין-דיוניים הינן בעלי התפלגות חד-שיאית. רכיב החול הינו 75-95%, רכיב הסילט הינו 5-20% ורכיב החרסית לרוב קטן מ-4%. התפלגות גדלי הגרגר של החול בחתכים המערביים דומה לאלו שבמזרח. הדיונות עשירות בקוורץ עם מעט פלגיוקלז. הרכב החול של הפרוזדור המרכזי טהור מהחול מצפון ומדרום ומכיל פחות דקים (סילט וחרסית), קלציט ופלגיוקלז.

עבודה זו מוקדשת לאבי מורי
מיכאל רסקין זכרונו לברכה (הת"ש – התשס"ה)



משרד האנרגיה והמים
המכון הגיאולוגי

העיתויים והמשמעויות הסביבתיות והפליאו-אקלימיות של פלישות הדיונות לצפון-מערב הנגב בסוף הרביעון

יואל רסקין

עבודת זו הוגשה לסנאט של אוניברסיטת בן-גוריון בנגב
לקבלת תואר " דוקטור לפילוסופיה "

העבודה נעשתה בהדרכתם של :

ד"ר נעמי פורת, המכון הגיאולוגי, ירושלים

פרופ' חיים צוער, המחלקה לגיאוגרפיה ופיתוח סביבתי, אוניברסיטת בן-גוריון בנגב

פרופ' דן ג. בלומברג, המחלקה לגיאוגרפיה ופיתוח סביבתי, אוניברסיטת בן-גוריון בנגב