Aeolian sand movement in an arid linear dune ecosystem, Nizzana, Western Negev, Israel

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Axel Allgaier M.A.

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1.Gutachter: Prof. Dr. Roland Baumhauer
2.Gutacher: Prof. Dr. Barbara Sponholz
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Contents

List of Figures

List of Tables

Zusammenfassung

Bewachsene Lineardünen dominieren weite Teile von ariden und semiariden Sandgebieten in denen bimodale Windsysteme und ein eingeschränktes Angebot von durch Wind transportierbarem Material aufeinander treffen. Beispiele sind die Kalahari im südlichen Afrika und die Simpson-Strzelecki Wüste in Australien. Wurde das Vorhandensein von Bewuchs auf Lineardünen ursprünglich als Zeichen von Inaktivität angesehen, so setzte sich nach eingehenden Untersuchungen die Ansicht durch, daß eine einfache Unterscheidung zwischen aktiven und inaktiven Dünen anhand dieses Merkmals nicht möglich ist. Allerdings besteht ein Zusammenhang zwischen dem Grad der Aktivität, der Form der Dünen und der Art und Dichte des Bewuchses. Wird der natürliche Zustand beispielsweise durch Beweidung verändert, die Vegetationsdichte also reduziert, kann eine Intensivierung äolischer Prozesse zu Veränderungen der Dünenform führen. Im umgekehrten Fall, also bei einer Zunahme der Vegetationsdichte, kommt es ebenfalls zu einer Anpassung der äolischen Prozesse an die neue Situation.

In einer dreijährigen Studie wurden die aktuellen äolischen Transportprozesse in einem vormals beweideten Lineardünengebiet nahe Nizzana an der israelisch-ägyptischen Grenze untersucht. Das Untersuchungsgebiet unterlag im Zeitraum von 1967 bis 1982 grenzüberschreitend starkem Beweidungsdruck, welcher zur vollkommenen Zerstörung der natürlichen Vegetation führte. Als Folge kam es zu verstärkter äolischer Aktivität und signifikanten Veränderungen der Morphologie der Dünen. Nach dem Ende der Beweidung auf israelischer Seite erfolgte dort eine rasche Rückkehr der Vegetation in den Dünengassen und an den Fußbereichen der Dünen. Darüberhinaus wurde auch eine Abnahme der offensichtlich aktiven Bereiche auf den Dünenrücken beobachtet. Die Situation auf ägyptischem Gebiet westlich der Grenze blieb bis heute unverändert. Die vorliegende Untersuchung zielt darauf ab, die veränderte äolische Morphodynamik östlich der Grenze zu erfassen. Der Schwerpunkt liegt dabei auf der Untersuchung der räumlichen und zeitlichen Verteilung des äolischen Sandtransportes sowie der diesen beeinflußenden Faktoren Morphologie, Oberflächenbeschaffenheit und Vegetation.

Für die Studie wurden an vier repräsentativen Standorten insgesamt fünf Meßstationen eingerichtet. Die Meßstellen befanden sich in einem Dünenkorridor, auf einer degradierten Düne mit unruhigem Kleinrelief sowie auf einer exponierten Düne mit minimalem Bewuchs. Alle Stationen waren mit Geräten zur Erfassung der Windgeschwindigkeit in drei verschiedenen Höhen (0.24 m, 0.65 m, 1.77 m) ausgerüstet. Die Windrichtung wurde jeweils in 2.40 m Höhe gemessen. An vier Stationen wurden akustische Sensoren zur Erfassung von Sandbewegung eingesetzt. Die Informationen über Wind und Sandbewegung wurden mit Hilfe von Datenloggern kontinuierlich aufgezeichnet. Sandfallen dienten an allen Meßstellen zur Ermittlung der transportierten Sandmenge. Am Gipfelbereich der exponierten Düne wurden auf einer Fläche von etwa 2670 m^2 mit 310 Erosionsmeßstäben in regelmäßiger Anordnung

Veränderungen der Form und des Volumens beobachtet.

Die Lage der Stationen in Korridor, auf degradierter und exponierter Düne dienten zur Untersuchung des Einflußes der Morphologie auf das lokale Windfeld und damit die Intensität und Richtung der Sandbewegung. Um die Einflüsse unterschiedlicher Oberflächenbeschaffenheit und Vegetationsbedeckung zu untersuchen, wurden im Dünenkorridor drei Stationen an zwei Standorten errichtet. Auf einem Rechteck von etwa 1000 m^2 wurde eine offene Sandfläche ohne Vegetation und Oberflächenkruste geschaffen. Hier konnte Sandbewegung ohne den Einfluß von zusätzlicher Oberflächenrauhigkeit beobachtet werden. Gleichzeitig konnte im direkten Vergleich mit der exponierten Düne die Auswirkung der unterschiedlichen Position im Gelände ermittelt werden. Am zweiten Standort wurden im Abstand von etwa 20 m zwei weitgehend identische Stationen aufgebaut. Die Erste verblieb im natürlichen Zustand: Die vorhandene biogene Oberflächenkruste und die Gefäßpflanzen wurden nicht verändert. An der zweiten Station wurde zunächst die Oberflächenkruste entfernt, die Gefäßpflanzen jedoch nicht verändert. Hierdurch war es einerseits möglich, im direkten Vergleich mit der vegetationslosen Station den Einfluß der Vegetation auf den Sandtransport im Korridor zu beobachten. Andererseits konnten die zusätzlichen Auswirkungen der Oberflächenkruste ermittelt werden. Im Januar 1999 wurde die Vegetationsbedeckung halbiert, um die Auswirkungen einer geringeren Pflanzendichte zu beobachten. Der Einfluß der biogenen Oberflächenkruste ohne Gefäßpflanzen wurde durch eine Windkanalversuch mit einer ungestörten Krustenprobe untersucht.

Die aktuellen Windverhältnisse im Untersuchungsgebiet, die zu Sandtransport führen, wurden aus langjährigen Meßreihen des AERC ermittelt. Diese wurden mit den während der Meßperioden 1997/98 und 1998/99 an den fünf Stationen aufgezeichneten Daten abgeglichen. Zur Ermittlung der Auswirkungen der Morphologie auf Windgeschwindigkeit und -richtung wurden das Verhältnis der Geschwindigkeit und die Differenz der Richtung zwischen den an der exponierten Düne gemessenen und den an den übrigen Stationen ermittelten Werten bestimmt.

Um längerfristige, dauerhafte Veränderungen im Untersuchungsgebiet zu erkennen wurden Luftbilder aus verschiedenen Jahren seit 1982 analysiert. Zusätzlich wurden Bodenaufnahmen aus früheren Jahren mit aktuellen Aufnahmen verglichen. Zur Erfassung aktueller Veränderungen wurden die Erosionsmeßstäbe auf der exponierten Düne zunächst dreidimensional vermessen und ihre Länge über der Oberfläche anschließend regelmäßig ermittelt. Dies geschah während der Wintermonate wöchentlich, bzw. nach jedem Sturm. In den Sommern 1998 und 1999 wurde die Länge der Stäbe monatlich gemessen.

Aus den langjährigen Winddaten wurde ein bimodales Windsystem sichtbar, in welchem die vorherrschende Richtung der sandtransportierenden Winde in Abhängigkeit von der Jahreszeit wechselt. Während im Winter südwestliche Winde dominieren, die an epsiodisch auftretenden Tiefdruckgebiete gebunden sind, herrschen im Sommer nordwestliche Winde vor. Diese verdanken ihre Entstehung einem Land-Seewindsystem, welches auf der Temperaturdifferenz zwischen Mittelmeer und Negev beruht. Bedingt durch die stabilen atmosphärischen Verhältnisse während dieser Zeit (Mai bis November), weisen die Sommerwinde eine große Stetigkeit in Stärke und Richtung auf. Dadurch werden die geringeren Geschwindigkeiten im Vergleich zu den Winterwinden kompensiert. Die während der Meßperiode gemessenen Daten bestätigen diese Ergebnisse.

Die Windrichtung hat Auswirkungen auf das Verhältnis zwischen der Windgeschwindigkeit an exponierten Stellen auf den Dünen und in den Dünenkorridoren. Während bei dünenparallelen Winden die Relation in Bodennähe, das heißt in 0.24 m über der Oberfläche, von Korridor zu Dünengipfel bei 0.8 lag, sank sie für senkrecht zum Dünenverlauf wehende Winde auf die Hälfte. In Abhängigkeit von der der Anströmungsrichtung der Lineardünen wird der Luftstrom im Lee unterschiedlich stark abgelenkt. Minimale Ablenkung erfolgt bei parallel zur Düne wehenden Winden, sowie bei einem Anströmwinkel von 90°. Ein weiterer Faktor für den Betrag der Ablenkung ist die Distanz zur windwärtigen Düne. Je kleiner der Abstand wird, desto größer ist die Ablenkung. Die Ablenkung erfolgt immer in Richtung zur Düne.

Der Einfluß der Lineardünen auf den Sandtransport in den Korridoren setzt sich zusammen aus der Abschirmung und der Ablenkung von Winden. Durch die starke Verlangsamung der Windgeschwindigkeit bei großen Anströmwinkeln führen nur aus Richtungen annähernd parallel zum Dünenverlauf wehende Winde oder sehr starke Stürme zu Sandbewegung in den Korridoren. Da der Wind während zyklonaler Winterstürme kontinuierlich von südlichen auf westliche Richtungen dreht, nimmt die Windgeschwindigkeit in den Korridoren im Verlauf eines solchen Sturms in der Regel zu, auch wenn die Windgeschwindigkeit über den Dünen gleich bleibt oder leicht abnimmt. Dadurch und durch die zusätzliche Ablenkung der Strömung findet Sandtransport in den Korridoren überwiegend parallel zu den Lineardünen statt. Die geringe Richtungsvarianz, der große Anströmwinkel und die geringe Geschwindigkeit der Seebrise verhindert Sandtransport in den Korridoren während des Sommers.

Regen, auch in kleinsten Mengen, unterbindet Sandtransport während des Niederschlagsereignisses. Die Transportraten erreichen jedoch ein bis drei Stunden nach dem Ende des Niederschlags wieder dieselben Werte wie vorher. In Verbindung mit der geringen Niederschlagshäufigkeit bedeutet dies, daß die Auswirkungen auf die äolischen Transportprozesse nur gering sind.

Die natürlich vorhandene Gefäßpflanzenvegetation bedeckt im untersuchten Korridor 17 Prozent der Oberfläche. Dadurch wird bei durchschnittlichen Winterstürmen die transportierte Masse auf ein Prozent der an offenen Flächen gemessenen Werte reduziert. Bei außergewöhnlich starken Stürmen erhöht sich die Menge auf bis zu acht Prozent. Ähnliche Auswirkungen hat eine Reduzierung der Vegetation um die Hälfte. Auf exponierten Dünen ist die Vegetation in isolierten Bulten konzentriert. Diese bewirken eine kleinräumige Veränderung des Windfeldes. Während es im Lee der Vegetationsflecken zu Ablagerung von Material kommt, findet an den Seiten verstärkte Abtragung statt. Da die Hauptwindrichtung jahreszeitlich wechselt, "wachsen" die Bulten durch Abtragung an allen Seiten. Bei degradierten Dünen findet sich Vegetation hauptsächlich auf Bulten, die durch ihre Größe und Häufigkeit die dazwischenliegenden Flächen stark abschirmen. Durchschnittliche Stürme führen hier nur zu geringer Sandbewegung in den Bereichen zwischen den Bulten. Während Starkwindereignissen wurde jedoch die Zerstörung einiger Bulten beobachtet.

Die mikrophytische Oberflächenkruste, die alle Flächen in den Korridoren und weite Teile der Fußhänge der Dünen bedeckt, verringert die Sandbewegung auf Werte, die mit den in der Studie verwendeten Geräte nicht quantifiziert werden können. An der unveränderten Station im Korridor wurde während des Meßzeitraums keine Saltation registriert. Der Versuch im Windkanal ergab, daß die derzeit im Untersuchungsgebiet auftretenden Winde nicht geeignet sind eine etablierte, ungestörte Kruste zu zerstören.

Auf exponierten Dünnenkämmen mit geringer Vegetationsbedeckung findet äolischer Sandtransport mit wenigen Ausnahmen täglich statt. In den Wintermonaten bilden sich während Wetterlagen mit geringen regionalen Luftdruckunterschieden an einigen Tagen keine ausreichend starken lokalen Winde aus. Die Form des Sandkörpers passt sich dem jahreszeitlichen Wechsel der Hauptwindrichtung an. Im Winter bildet sich ein ausgeprägter, sinusförmig verlaufender Grat, verursacht durch die große Richtungsvarianz der sandtransportierenden Winde. Im Sommer schaffen Winde mit geringer Richtungsvarianz ein konvexes Profil mit nach Süden gerichteten Rutschhängen. Der Winter zeichnet sich durch eine ausgeglichene Massenbilanz bei hohem Umsatz aus. Im Sommer ist die Bilanz stark negativ, so daß es insgesamt zu einer Verringerung des Volumens kommt. Diese Erkenntnis aus den direkten Messungen deckt sich mit den Analysen der Luft- und Bodenbilder, die eine deutliche und kontinuierliche Abnahme der Dünenhöhe und der offensichtlich aktiven Bereiche seit 1982 ergaben.

Fünfzehn Jahre nach Beendung der Beweidung beschränkt sich äolische Sandbewegung im Lineardünenfeld Sde Hallamish auf die Gipfelbereiche von Dünen. Jedoch gibt es auch hier große Unterschiede im Grad der Aktivität. Die aktivsten Bereiche sind hohe, exponierte Dünenkämme. Hier findet ganzjährig äolischer Sandtransport statt und die Vegetationsdichte ist innerhalb des Untersuchungsgebietes am geringsten. Je niedriger die Dünen werden und je mehr man sich der Grenze nähert, desto dichter wird die Vegetation. Die Dünenkorridore und der überwiegende Teil der Fußhänge können als inaktiv betrachtet werden, die dort unter ungestörten Bedingungen stattfindende äolische Sandbewegung führt nicht zu signifikanten Änderungen der Morphologie. Mittelfristig wird die Aktivität auch auf den aktuell mobilen Dünenbereichen nachlassen, da durch fehlenden Sandnachschub das zu bewegende Material fehlt. Eine vollständige Stabilisierung der Dünen durch Ausbildung einer biogenen Kruste auch auf den Dünenkämmen erscheint jedoch unter den gegebenen Bedingungen unwahrscheinlich, da auch niedrige, degenerierte Dünen aktuell offene Sandflächen aufweisen, auf denen episodisch Sandbewegung stattfindet.

Summary

Vegetated linear dunes dominate large parts of arid and semiarid sand areas where bimodal wind systems and reduced availability of transportable material are combined. Examples are the Kalahari in southern Africa and the Simpson Strzelecki desert in Australia. While the presence of vegetation on linear dunes was originally regarded as indication of inactivity, after detailed investigations the opinion became generally accepted that a simple distinction between active and inactive dunes is not possible on this basis. However, a connection between the degree of the aeolian activity, the form of the dunes and the kind and density of the vegetation exists. If the natural condition is changed for example by grazing which reduces vegetation density, intensified aeolian processes can cause changes of the dune morphology. In the reverse case, ie with an increase of the vegetation density, an adjustment of the aeolian dynamics to the new situation will also take place.

In a three-year study the current aeolian transportation processes were examined in a linear dune area previously used for grazing near Nizzana at the Israeli-Egyptian border. The research area was subject to heavy grazing across the border, which led to the total destruction of the natural vegetation in the period of 1967 to 1982. As a consequence, intensified aeolian activity and significant changes of the morphology of the dunes were observed. After the end of the grazingg on the Israeli side, a rapid return of the vegetation in the interdune corridors and on the footslopes of the dunes took place. In addition also a reduction of obviously active areas on the dune crests was observed. The situation on Egyptian territory west the border remained unchanged until today. This study is aimed at understanding the changed aeolian morphodynamics east the border. The emphasis was placed on the investigation of the spatial and temporal distribution of aeolian sand transport as well as on the influencing factors morphology, surface condition and vegetation.

Five measuring stations were set up at four representative locations for the study. The measuring stations were located in an interdune corridor, on a degraded dune with uneven small relief and on an exposed dune with minimum vegetation. All stations were equipped with devices for the measurment of the wind velocity in three different heights (0.24 m, 0.65 m, 1.77 m). The wind direction was measured in 2.40 m height. At four stations acoustic sensors were used for the detection of sand movement. The information about wind and sand movement was recorded continuously by dataloggers. Sand traps were used at all stations to determine the transported sand quantity. Changes of the form and the volume were monitored at the crestal area of the exposed dune on a surface of approximately 2670 m^2 with 310 erosion pins in a regular arrangement.

The situation of the stations in the interdune corridor, on the degraded and the exposed dune served for the investigation of the influence of the morphology on the local wind field and thus the intensity and direction of sand movement. In order to examine the influences of different surface conditions and vegetation cover, three stations were established at two locations in the interdune corridor. An open sand surface without vegetation and surface crust was created on a rectangle of approximately 1000 m^2. Here sand movement could be observed without the influence of additional surface roughness. At the same time the effect of the different position within the area could be determined in direct comparison with the exposed dune. At the second location two stations were set up approximately 20 m apart which were to a large extent identical. One remained in its natural condition: the existing biogenous surface crust and the vascular plants were not changed. At the other station the surface crust was removed, but the vascular plants were left unchanged. Through this arrangement it was on one hand possible to observe the influence of the vegetation on sand transport in the interdune corridor in the direct comparison with the vegetationless station. On the other hand the additional effects of the surface crust could be determined. In January 1999 the vegetation cover was reduced by 50 per cent at the station without crust, in order to observe the effects of a smaller plant density. To determine the effects of a surface crust without further vascular vegetaion, a windtunnel experiment with an undisturbed crust sample was conducted.

Current sand transporting wind conditions in the research area were determined from long-term measurements of the AERC. These were matched with the data recorded at the five stations during the measuring periods 1997/98 and 1998/99. For the determination of the effects of the morphology on wind velocity and wind direction the relation of velocity and the direction difference between the values of the exposed dune and those of the remaining stations were calculated.

In order to recognise long-term, lasting changes in the research area aerial photographs were analysed taken in different years since 1982. In addition ground photographs from earlier years were compared with more recent pictures. For the monitoring of current changes the erosion pins on the exposed dune were surveyed three-dimensional and afterwards their length above the surface was regularly determined. This was done weekly or after each storm during the winter months. In the summers 1998 and 1999 the length of the pins was measured monthly.

The long-term wind data revealed a bimodal wind system, in which the prevailing direction of the sand-transporting winds changes seasonally. The winter is dominated by southwesterly winds, which are connected to episodic passages of low pressure cells. Northwesterly winds prevail in the summer. Their origin is a land-sea wind system, which is based on the temperature difference between the Mediterranean and the Negev. Due to stable atmospheric conditions during this time (May until November), the summer winds exhibit a high steadiness of magnitude and direction. Thus the lower windspeeds compared with the winter winds are compensated. The data recorded during the measuring period confirm these results.

The wind direction has effects on the relationship between the wind velocity in exposed places on the dunes and the interdune corridors. During dune-parallel winds the relation at ground level (0.24 m) between interdune corridor and dune crest was 0.8 but it sank to half for winds blowing perpendicular to the dune.

The deflection of air flow in the lee of the linear dunes is dependent on the incident flow direction. Minimum deflection takes place with winds blowing parallel to the dune, as well as with an angle of attack of 90°. A further factor for the amount of the deflection is the distance from the upwind dune. Smaller distances lead to more intense deflection. The deflection is always towards the dune.

The influence of the linear dunes on sand transport in the interdune corridors is based on the combination of sheltering and flow deflection. The dune ridges protect the interdune corridors against winds with high incident angles towards the dune ridge orientation. Only winds blowing parallel to the dunes or very strong storms lead to sand movement in the corridors. Since the wind direction during cyclonic winter storms changes continuously from southern to western directions, the wind velocity in the interdune corridors usually increases in the course of such a storm, even if the wind velocity above the dunes remains steady or decreases slightly. This effect and the flow deflection restricts sand transport in the interdune corridors to dune-parallel directions. The small direction variance, the high angle of attack and the low magnitude of the sea-beeze prevent sand transport in the interdune corridors during the summer.

Rain, also in smallest quantities, prevents sand transport during the precipitation event. However, the transportation rates reach the same values as before the rain until one to three hours after the end of the precipitation. In connection with the low frequency this means that the effects of moisture on aeolian sand transport are only small.

The naturally existing vascular vegetation covers 17 Prozent of the surface in the examined interdune corridor. This cover reduces the transported mass during average winter storms to one per cent of the value measured at open surfaces. Exceptionally strong storms cause significantly higher transport. A reduction of the vegetation to half of the initial value has similar effects. On exposed dunes the vegetation is concentrated in isolated hummocks. These cause a small-scale change of the wind field. While accumulation of material was observed in the lee of these obstacles, intensified deflation takes place at the sides. Since the main wind direction changes seasonally, the hummocks 'grow' by deflation on all sides. At degraded dunes vegetation is mainly located on hummocks, which by their size and frequency shield the intermediate surfaces to a large extent from sand-transporting winds. Average storms lead only to little sand movement within the areas between the hummocks. During strong wind events the destruction of several hummocks was observed.

The microphytic surface crust, which covers all surfaces in the interdune corridors and large parts of the foot slopes of the dunes, reduces the sand movement to values, which cannot be quantified with the devices used in the study. Saltation was not determined at the undisturbed station in the interdune corridor during the measuring period. The wind tunnel experiment showed that the present winds in the research area are not capable of destroying an established, undisturbed crust.

On exposed dune crests with low vegetation cover aeolian sand transport takes place daily with few exceptions. In the winter months, during weather conditions with small regional air pressure differences, no sufficiently strong local winds develop. The morphology of the sand body adapts to the seasonal change of the main wind direction. In the winter a pronounced, sinuous crestline is formed, caused by the high direction variance of the sand-transporting wind. In the summer winds with small direction variance create a convex profile with southfacing slip faces. The winter is characterised by a balanced mass budget with a high turnover of material. In the summer the balance is strongly negative, so that a net decrease of the volume occurs. This result from the direct measurements is in accordance with the analyses of the air and ground pictures, which showed a distinct and continuous decrease of the dune height and the obviously active areas since 1982.

Fifteen years after termination of grazing aeolian sand movement at the linear dune field of Sde Hallamish is limited to the crestal areas of dunes. However, there are also significant differences in the degree of the activity in these areas. The most active areas are high, exposed dunecrests. Aeolian sand transport takes place all year round and the vegetation density is lowest within the research area. The lower the dunes become and the closer one gets to the border, the denser becomes the vegetation. The interdune corridors and the major part of the foot slopes can be regarded as inactive, the aeolian sand movement taking place there under undisturbed conditions does not lead to significant changes of the morphology. In medium-term the activity is expected to diminish on the currently mobile dune ranges too, since a lack of sand supply leads to a lack of material to be moved. However, a complete stabilisation of the dunes by the formation of a biogenous crust also on the dune crests appears unlikely under the current conditions, since even low, degenerated dunes exhibit open sand surfaces, on which episodic sand movement takes place.

Introduction

Aeolian sand movement is considered to be the limiting factor for vegetation growth on sandy soils in semiarid areas, much more than moisture or precipitation (Tsoar, 1990). If in such an environment with minimal precipitation, a well-developed vegetation cover is disturbed over a large area and during several consecutive years, for example by grazing, the result can be increasing aeolian activity and a negative feedback, leading to the destruction of the remaining vegetation by windblown sand. Such a scenario was described for the linear dunes of Sde Hallamish/Nizzana by Tsoar & Møller (1986). A reversal of these changes was observed after the cessation of grazing activity in the area: a quick recovery of the vegetation took place, again changing the appearance of the dunes. This study focusses on the aeolian dynamics within this linear dune ecosystem after more than 15 years of undisturbed development and on the factors controlling this development.

1.1 Dune forms, sediment supply and wind regime

'The formation of desert sand dunes is dependent on the availability of suitable sediment and winds strong enough to entrain and transport it' (Thomas & Tsoar, 1990, p.471). This means that active desert dunes will appear in an environment where a sufficient amount of sand-sized sediment is met by winds strong enough to initiate and maintain particle movement (Thomas & Tsoar, 1990; Livingstone & Thomas, 1993). Although the prerequisites for dune formation are known, the genesis of dunes is not well understood, but once a dune form has been established it interacts with the airflow and leads to its perpetuation until the environmental conditions change significantly (Thomas & Tsoar, 1990, p.472). An interdependence between the form of the obstacle and the airflow will develop. The relation between dune form and airflow will ideally be a dynamic equilibrium. The dune will adjust to the current airflow by the movement of material (Lancaster, 1987, p.519). 'Fundamentally, the equilibrium morphology, particularly the size, of desert dunes is a function of their sediment budget, which may be expressed as the balance between erosion and deposition, or sand supply and removal at each point on the dune, summed for the dune as a whole' (Lancaster, 1985). When an equilibrium between sand supply and wind energy is established in a uni-directional wind regime, the form may migrate, but it will keep its shape (Tsoar, 1985, p.56). The shape will be such that it offers the least resistance to the airflow. When limited sand supply is combined with a bimodal wind regime, linear dunes are formed (Wasson & Hyde, 1983). If the sand-moving wind direction changes seasonally, the mobile areas of the dune ridges adjust to the prevailing wind. The dynamics of non-vegetated linear dunes (*seif*) within such a wind regime have been investigated in detail by Tsoar (1978; 1983). He also offered a model for the transformation of *barchan* dunes into linear dunes through the elongation of one of its horns once they move into an area with such a wind regime (Tsoar, 1984).

Linear dune landscapes generally consist of three major features (Lancaster, 1982): the dune crests as the uppermost parts, the footslopes or plinths which form the base of the dune and the interdune corridors, separating the ridges from each other. *Vegetated linear dunes* are common to wide areas of the Australian deserts (Ash & Wasson, 1983; Mabbutt & Wooding, 1983) and the Kalahari desert (Lancaster, 1981; Thomas, 1988). Similar in size are *seif* dunes, yet they are devoid of vegetation and have a sharp, sinuous crest line. It has been argued that *vegetated linear dunes* and *seif* are two dune forms differing in their genesis (Tsoar, 1989). On the other hand, several authors (Thomas, 1988; Tsoar & Møller, 1986) report that *vegetated linear dunes* can develop *seif*-like features if the vegetation cover is destroyed. Therefore Livingstone & Thomas (1993) suggest considering *vegetated linear dunes* and *seif* not as 'discrete fundamental dune types' but as morphological variations of the same caused by differing environmental conditions.

Linear dunes are features which are mainly elongated along their main axis. Sand is transported along the dune ridge by the wind component parallel to the ridge. Oblique winds are deflected on the downwind side of the crest, depending on the angle of incidence, dune shape and atmospheric stability (Tsoar, 1983; Livingstone, 1988; Sweet & Kocurek, 1990). This deflection is important for the shaping of the active dune parts (Tsoar, 1983). According to Tsoar (1983) angles of incidence $\leq 40°$ will lead to higher flow velocity on the lee side than on the dune crest, while at angles $\geq 40°$ the airflow in the lee of the crest is slowed down. Flow patterns over the dune ridges at Nizzana have been investigated by Sharon *et al.* (2002). The authors conclude that in addition to the angle of incidence, and dune shape, 'background wind velocity' is a major factor causing flow variation over linear dunes, while atmospheric stability does not have a significant influence (Sharon *et al.*, 2002, p.887). In addition to the elongation of linear dunes, Hesp *et al.* (1989) and Rubin (1990) found evidence for a lateral movement of linear dunes and the asymmetrical cross section of many linear dunes also points to a limited movement in lateral direction. Recent studies of the sediment record at Nizzana (Rendell *et al.*, 1993; Harrison & Yair, 1998) have revealed that the dunes have been in place since at least the late Pleistocene and that their lateral movement has been negligible.

1.2 The concept of dune activity and the role of vegetation for aeolian sand transport

What amount of aeolian sand movement is necessary to call a dune 'active'? This is an important question, as 'inactive' dunes have frequently been used as signs for a drier or windier environment during the period of their emplacement and thus as proof of climatic change (Ash & Wasson, 1983; Thomas & Shaw, 1991; Thomas, 1992; Livingstone & Thomas, 1993). In order to use presently inactive dunes as an indicator for a drier and/or windier paleoclimate it is necessary to be sure about their state in relation to the present wind regime (Livingstone, 1988) and other environmental conditions such as vegetation and sand supply.

Ash & Wasson (1983) noted that in linear dune fields only the crestal areas of the dune ridges show signs of mobility like ripple marks and active slip faces, while the plinth and the interdune corridors appear stable. Livingstone (1989) showed that dune activity increases towards the dune crest. He suggests that the increase of wind velocity with dune height causes the higher activity. Airflow is accelerated when it has to pass an obstacle of significant lateral dimension. The properties influencing the amplification factor are the height of the obstacle and the length of the windward slope. A detailed overview of concepts is given by Tsoar (1985, p.48ff). Independent of the obstacle´s profile, the maximum near surface flow velocity is reached at the crest (Frank & Kocurek, 1996a). The effect of flow acceleration or speed-up on the windward flanks of linear dunes similar to those at Sde Hallamish has been described by Lancaster (1985, p.582).

Ash & Wasson (1983) as well as Livingstone (1988) state that it is not possible to draw a sharp line to separate 'active' from 'inactive' dunes. This is especially true for linear dunes, as these do not migrate like *barchans* (Livingstone & Thomas, 1993). Important variables are windiness, rainfall and vegetation cover (Ash & Wasson, 1983, p.22). Sand supply is also important in connection with windiness: 'A supply of sediment, generally sand-sized, is essential for continued dune activity' (Livingstone & Thomas, 1993, p.97). High windspeeds will not lead to active dunes if no sand is available for dunebuilding processes. In the context of this study, 'active' is defined as morphological significant aeolian sand transport processes acting on an annual basis (Livingstone & Thomas, 1993, p.97). This includes all surfaces of the research site, ie dune ridges and interdune corridors.

Vegetation is present in all sandy deserts except for the most arid or overgrazed regions (Ash & Wasson, 1983). Tsoar & Møller (1986) consider vegetation as normal for all areas receiving more than 50 mm of annual rainfall, so its simple presence cannot be regarded as a sign of inactivity of a dune landscape. Nonetheless, vegetation increases the aerodynamic roughness of a surface and thereby extracts energy from the airflow, reducing shear stress at the soil surface. It is therefore important to determine the degree of protection against wind erosion offered by a given vegetation cover compared to an open sand surface under the same meteorological conditions.

Stockton & Gillette (1990) and Musick & Gillette (1990) have been investigating the relationship between plant cover and erodibility of surfaces, concentrating on the influence of vegetation density on the partition of shear stress between vegetation and soil surface. Lancaster & Baas (1998) undertook field studies over areas with different vegetation cover.

Wolfe & Nickling (1993) focus on the general effect of sparse vegetation within an environment prone to wind erosion. They distinguish three main effects: (a) cover of the surface, (b) momentum extraction from the airflow and (c) trapping of soil particles already in motion. In their work they stress that dead vegetation also plays an important role in the protection of a surface against wind erosion, a point also made by Wiggs *et al.* (1995).

Thomas & Tsoar (1990) discuss the role of vegetation for desert dunes. They distinguish three groups of interaction between vegetation and dune geomorphology: surface stabilization, accretion focus and determination of dune morphology. They conclude that the presence of vegetation alone - even if the vegetation cover is nearly complete - is not a sufficient attribute to call a dune field 'inactive'. This contrasts earlier theories which considered a dense vegetation cover on desert dunes to be a sign of inactivity. It has been argued that such currently vegetated dunes were formed during periods with a more arid climate than present (Lancaster, 1981; Sarntheim, 1978). Meanwhile, this simple assumption has been doubted (Livingstone, 1989) and several recent field studies have proved this point (Wiggs *et al.*, 1995; Wiggs *et al.*, 1996; Bullard *et al.*, 1997). On the other hand Wiggs (1993) mentions a possible threshold cover of 14 per cent for certain areas of the Kalahari desert, above which no noteworthy aeolian sand transport occurs. However, he does not claim this to be a universal value. The value is in accordance with the findings of Marshall (1973), who concludes that wind erosion will increase rapidly if the vegetation cover drops below 15 per cent on level alluvial sand surfaces. Yet Ash & Wasson (1983, p.20) report sand movement on dune crests with a ground cover of 35 per cent . Wolfe & Nickling (1993, p.56f) state that a surface cover of 40 per cent will lead to 'skimming flow', which means that the surface is not influenced by the flow above the roughness elements, ie vegetation. If the cover is below 40 per cent, the wakes of the roughness elements do not overlap completely and therefore leave 'voids' in which the forces of the wind can act directly on the surface. Also Lee (1991) notes that high densities of roughness elements lead to a smooth aerodynamic surface above the elements. Apart from the percentage of surface cover by vegetation, its nature, its distribution and the actual wind regime in the area under observation have to be considered (Thomas & Tsoar, 1990, p.478).

The vegetation communities within a dunefield are not uniform but vary across the dune profile (Thomas & Tsoar, 1990). For Nizzana this has been investigated by Tielbörger (1997). The influence of a single obstacle and of different roughness element concentrations is discussed by Wolfe &

Nickling (1993, *p.56f*) and Ash & Wasson (1983). An isolated obstacle may lead to increased velocities - and thereby to increased erosion - as the flow is streamlined around it (Ash & Wasson, 1983; Wolfe & Nickling, 1993; Thomas & Tsoar, 1990). This effect has been observed for vegetation cover of up to 25 per cent when plant diameter is approximately equal to height. The effect is most pronounced for a cover of 5 - 10 per cent (Ash & Wasson, 1983, p.19). The same obstacle may act as an accretion focus for moving sand on its lee side due to the reduced wind velocity in that area (Hesp, 1981). Deposition of material might also occur on the windward side, depending on the size of the plant and its porosity to the wind (Thomas & Tsoar, 1990). The influence of single obstacles is only local, resulting in only small coppice dunes of nebkas. However, increasing sand supply might lead to the development of 'real' dunes from these focus points, reducing the influence of the plants as the dune body grows (Thomas & Tsoar, 1990). For all these effects, the plant does not have to be alive. If it retains its shape after death, it will influence the wind field in much the same way as if it was alive (Ash & Wasson, 1983).

The influence of vegetation on the dune morphology itself is achieved in the case of linear dunes by the reduction of the influence of secondary winds by increasing vegetation cover (Thomas & Tsoar, 1990, p.482ff). A well-developed vegetation cover on a linear ridge leaves only the strongest winds able to move sand, while decreasing vegetation cover enables winds of lower magnitude to act on the dune body. The effect of decreasing vegetation on linear ridges has been shown for the research area by Tsoar & Møller (1986): After the destruction of vegetation secondary ridges were superimposed onto the existing dunes.

While the majority of geomorphological studies focuses on the effect of a given vegetation cover on the processes acting, these processes do have an effect on the plants. Several studies at the research site have focussed on this aspect (Kadmon, 1994; Kadmon & Leschner, 1995; Tielbörger & Kadmon, 1995; Tielbörger, 1997; Prasse, 1999). Vulnerability is highest in the early stage of the life of a plant. After germination has taken place, the survival of the seedlings depends to a large extent on the mobility of the surface, because shifting sand can cover vegetation, deflation may expose roots and saltating grains can damage the plants (Kadmon, 1994; Tielbörger, 1997; Kadmon & Leschner, 1995).

In addition to the role played by vascular vegetation, the influence of cryptogamic soil crusts on deflation has been the focus of several studies mainly in drylands used or suitable for grazing (West, 1990; Harper & Marble, 1988; Belnap, 1995; Belnap & Gillette, 1998; Leys & Eldridge, 1998). The resistance of such crusts against wind erosion has been tested in wind tunnel studies (McKenna Neuman *et al.*, 1996; McKenna Neuman & Maxwell, 1999; McKenna Neuman & Maxwell, 2002). The results of these studies show that intact microphytic crusts inhibit deflation of sandy soils, while a destruction (eg trampling by grazing animals) leads to a sharp increase of aeolian transport rates (Leys & Eldridge, 1998). Their role in protecting soils against erosion by wind and water, thus increasing landscape stability, has been acknowledged (Eldridge & Greene, 1994). The authors also stress the importance of algal crusts for the initial stabilization of a surface after a disturbance (Eldridge & Greene, 1994, p.405). Microbial crusts are considered vital for the stability of sandy soils in rangelands (Leys & Eldridge, 1998).

1.3 Investigating sand transport: models and reality

A considerable amount of literature deals with the question of how to determine or calculate aeolian sand transport from wind data, as this standard parameter is determined at most meteorological sites. Iversen & Rasmussen (1999) compare several of the published equations in their report on wind tunnel experiments using different slopes and grain sizes to establish a relationship between shear stress and aeolian mass transport. Their findings establish a good relationship between wind speed, grain size and slope angle for uniform sands. However, real world data sets with non-uniform grain size distributions do not fit their proposed model. Data scatter caused by uncertainties inevitable in field experiments (McKenna Neuman *et al.*, 1997) leads to large differences in calculated values of mass transport of almost an order of magnitude (Iversen & Rasmussen, 1999).

Bauer *et al.* (1996) argue that most, if not all, existing models for aeolian sand transport fall short of being able to predict actual transport rates. They found that even in relatively simple environments such as flat sandy beaches the factors influencing sand transport are so numerous and interdependent that it is impossible to predict aeolian sand transport based on simple physical equations. In order to obtain valid data for a specific site, it is therefore necessary to measure wind speed and sand transport on the spot in the field.

The standard parameter for the characterisation of the erosivity of wind is the shear velocity u_*. Most equations for aeolian sand transport use u_* as the main variable responsible for the actual transport. It can be determined by using vertical wind profiles and the von Karman-Prandtl logarithmic velocity profile law which can be written as

$$\frac{\overline{U}}{u_*} = (\frac{1}{k})ln(\frac{z}{z_0}) \tag{1.1}$$

where $\overline{U}$ is velocity at height z above the surface. The empirical *von-Karman* constant k is commonly taken as 0.4 (Pye & Tsoar, 1990, p.31). The roughness length z_0 is a property of the surface and can also be determined using the vertical wind velocity profile.

When large roughness elements such as shrubs or other dense vegetation patches are present, the reference plane is no longer the surface but a layer at height d, the *zero plane displacement*. In this case equation 1.1 is modified to

$$\frac{\overline{U}}{u_*} = (\frac{1}{k}) ln(\frac{(z-d)}{z_0}) \tag{1.2}$$

From a purely physical point of view, the use of u_* may be correct. Yet under field conditions it appears impossible to measure flow velocity within the inner surface layer (HUNT *et al.*, 1988) with conventional instrumentation like cup anemometers (BAUER *et al.*, 1992). BAUER *et al.* (1996, *p.648*) point out the limitations of equation 1.1, as for an estimation of u_*, flow has to be steady and uniform and the profile in the lower part of the boundary layer must be log-linear.

Over dunes, the situation becomes even more complicated. PYE & TSOAR (1990, *p.43*) point out that 'the near-surface wind profile over a hill deviates from the ideal logarithmic profile', making the accurate calculation of shear velocities difficult. Studies on *barchan* dunes have shown that values of u_* decrease up the stoss slope while transport rates increase (LANCASTER *et al.*, 1996; MCKENNA NEUMAN *et al.*, 1997). The authors conclude that 'conventional wind profiles derived from anemometry on dunes do not measure the part of the boundary layer that is significant for sediment transport' (LANCASTER *et al.*, 1996, p.62). In other recent researches (MULLIGAN, 1988; BUTTERFIELD, 1991; BURKINSHAW *et al.*, 1993; FRANK & KOCUREK, 1996A) wind profiles on the stoss slopes of dunes have been found to be not log-linear and therefore not useable to determine u_*. In another approach, BURKINSHAW & RUST (1993) report that wind speed measured at 6 cm above the surface appears to reflect the shear stress at the surface, while measurements at 140 cm are not suitable for an estimation.

STEERK *et al.* (1998) question the concept of shear stress for the description and calculation of aeolian mass transport altogether. Their results indicate that the horizontal, streamwise velocity component of the airflow close to the surface is the important and governing factor for saltation. Similar results were obtained by other authors (MCKENNA NEUMAN *et al.*, 1997; WALKER, 1999). Therefore it appears to be sufficient to measure the horizontal windspeed at a height close to the surface in order to be able to estimate sand transport for a given wind at a given site.

Sand flux in the field is usually determined by sand traps. Various types are described by PYE & TSOAR (1990, *p.324ff*). An assessment of three sand trap designs (*Leatherman*-trap (LEATHERMAN, 1978), *Fryberger*-trap (FRYBERGER *et al.*, 1984) *Ames*-trap (GREELEY *et al.*, 1982)) in respect to their trapping efficiency has been done by GREELEY *et al.* (1996). Their results show that the *Leatherman*- and *Ames*-traps catch an average of 30 per cent of the transported material, while the more sophisticated *Fryberger*-trap overcollects. In order to catch the bulk of transported material, it is important to use traps which collect from the surface upwards, as the mass flux of particles decreases exponentially with height (ANDERSON & HAFF, 1988). Transport rates have been found to be 100 times higher at 1 cm height than at

10 cm (Sorensen, 1985). Chepil & Milne (1939) found that 57 per cent of saltating grains travel below 5 cm, while Butterfield (1999) states that 79 per cent move below 1.8 cm above the bed. Sorensen (1985) found that the height h of the flight path of saltating grains is dependent on the shear stress u_*.

1.4 Aim of the study

For the area of Sde Hallamish near Nizzana, Tsoar & Møller (1986) have shown that intensive grazing of flocks of sheep and goats within an arid linear dune ecosystem leads to an increase of dune activity and significantly changes dune morphology. This study investigates the situation at the site 15 years after the cessation of grazing activities. It concentrates on answering the following questions:

1. What is the nature of winds causing sand movement within the linear dune ecosystem?
2. How do morphology, surface condition and vegetation cover affect aeolian sand transport? What is the relative influence of each of these parameters?
3. What is the actual spatial distribution of aeolian sand transport within the linear dune system? Does the current condition represent a 'steady state' resp. an 'equilibrium'?

To answer these questions a variety of methods was used. Basic wind parameters were monitored over prolonged periods of time at different sites. Sand transport was recorded parallel to the wind data. Experimental changes of surface conditions were carried out at the field site. Wind tunnel experiments were conducted to evaluate surface parameters. Monitoring of an active dune part was carried out to identify changes of morphology. Aerial and ground photographs taken at different times since 1982 have been analysed for differences.

1.5 The research area

The research has been carried out in the Sde Hallamish sands near Nizzana. This site was established in 1988 and is maintained by the *Arid Ecosystems Research Centre* (AERC) at the Hebrew University of Jerusalem. Prior to its designation as a permanent research area, intensive grazing of flocks of sheep and goats had destroyed the vegetation cover. Grazing was stopped in 1982 and access to the area was restricted on the Israeli side due to the proximity to the border with Egypt. This circumstance offered the opportunity to evaluate the recovery of a heavily disturbed dune ecosystem and to determine its actual state after more than 15 years of undisturbed development.

1.5.1 Geographical setting

Sde Hallamish/Nizzana is located in the western Negev desert, adjacent to the border between Israel and Egypt (see figures 1.1, p.18 and 1.2, p.19). The area represents the eastern extension of the Sinai continental sand field. Mean altitude is 200 m a.s.l. with a local minimum of 190 m and a maximum of 212 m. It consists of several parallel linear dune ridges and adjacent interdune corridors running W-E. A dry river, Nahal Nizzana, running perpendicular to the dunes, terminates the ridges approximately 2 km east of the border. The anticlines of the Negev highlands rise towards the south, preventing aeolian movement of sand in that direction. Roads running parallel to each other on both sides of the international border were built in 1982, following the peace treaty between Egypt and Israel. The roads are cut through the dune ridges or built on dams where they cross interdune corridors.

1.5.2 Climate

The Negev is part of the deserts associated with the subtropical high pressure belt. Its northern part lies at the fringe of this zone and is therefore influenced by the seasonal changes of its location. The Atlas of Israel (1985, p.12) assigns Nizzana to climate type BSh (dry hot steppe) or BWh (dry hot desert) according to the classification of Köppen. The local climate of the research site is to a large extent controlled by the proximity of the Mediterranean and the resulting diurnal and seasonal temperature differences. Meteorological data, especially on wind and precipitation have been collected continuously since the establishment of the AERC research station in 1988.

Wind The wind regime at Sde Hallamish is governed by the seasonal shift of regional pressure systems. Summer in the Middle East is characterised by stable air masses caused by the development of a thermodynamic high pressure area in the upper atmosphere. This leads to a regular diurnal wind regime at the research site: a light breeze from SE prevails during night-time, caused by the rapid cooling of the land surface of the Negev and the proximity of the warm water body of the Mediterranean. In the daytime, heating of the desert surface causes a reversal of airflow, leading to the establishment of a northerly breeze in the early afternoon hours. This breeze regularly reaches wind speeds above saltation threshold at the dune crests where it causes sand movement.

In winter, night-time conditions are similar to summer, thus the prevailing wind direction is SE. Wind speed of this breeze is low (Littman, 1997). In contrast to summer, no regular reversal of flow direction occurs. Instead the magnitude of the SE-breeze may increase during daytime to values above saltation threshold if stable atmospheric conditions are present. Such conditions are shortlived as due to the southward shift of the subtropical high pressure zone in winter, low pressure cells pass the eastern Mediterranean. High magnitude SW-winds are associated with their passage. Out of an average of 25

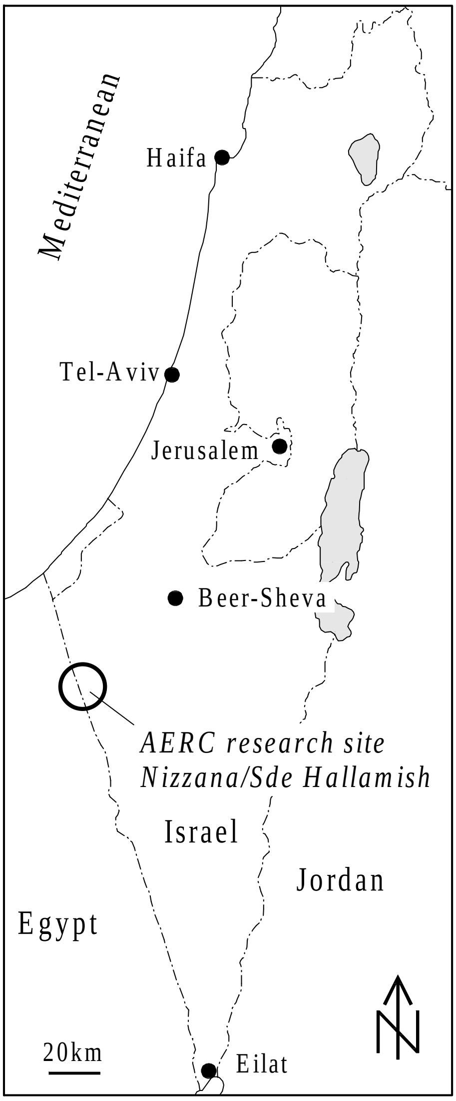

Figure 1.1: Location of the research site.

cyclones in the eastern Mediterranean during winter, about half actually reach the coast of Israel (Karmon, 1994, p.23).

Precipitation Rainfall is restricted to the winter months from November until May (figure 3.37, p.86) and is usually associated with the passage of frontal depressions. Average annual precipitation is about 90 mm with a high relative standard deviation of 68 per cent (Tsoar & Møller, 1986). During all three winters of the study period, rainfall was below average.

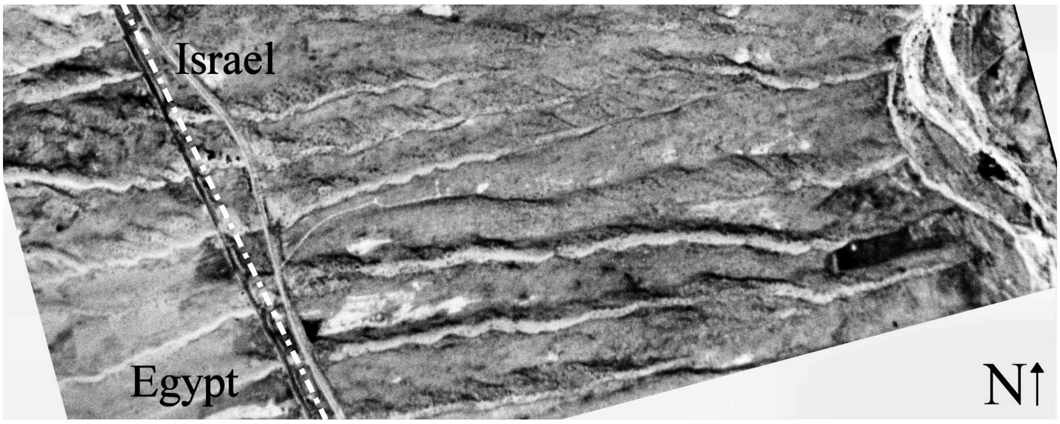

Figure 1.2: Aerial view of the research site Sde Hallamish/Nizzana. Picture taken in 1992. Border alignment approximate. Scale and elevation data is provided in figure 2.1, p.23.

1.5.3 Morphology

Vegetated linear dune ridges are the most prominent feature of the research area. The three major features making up linear dune landscapes, namely crest, foot slopes or plinths and interdune corridors (LANCAS TER, 1982) are found in Nizzana. Recent changes in land use (TSOAR & MØLLER, 1986) have led to a modification of most of the ridges into *braided linear dunes* (PYE & TSOAR, 1990, p.209), so that deviations from the ideal morphology of vegetated linear ridges are widespread.

Slip faces, blow-out areas and ripple marks are common on the *dune crests*. Several ridges at the research site contain segments with *seif*-like, sharp edged crest lines. However, compared with the Egyptian side, these features are rare on the Israeli side. The crestal area is about 20 m wide on average with a low perennial vegetation cover between 12 to 17 per cent (TIELBÖRGER, 1997, p.264). The vegetation is concentrated in isolated spots along the centerline of the ridge and at the transition zone between the active crest and the flanks of the ridge. Slope angles on the dune crest are low, apart from slip faces and the flanks of vegetation covered hummocks.

The *plinths* of the dunes are separated from the crest by a knickpoint. The mostly crust-covered, concave slopes connect the crests with the interdune corridors. Vegetation cover is higher than at the crests, while sand mobility is very limited. At Sde Hallamish, dune ridges are to a large extend asymmetrical. Therefore the extent of the plinth area varies considerably. Many sections of ridges have plinths only on one side, while on the opposite side the crest borders directly on the interdune corridor.

The *interdune corridors* are generally flat, level areas with only small undulations. The gradient increases towards the plinth of the ridges, where in most cases no clear boundary can be drawn. The sandy surface is covered by a microphytic surface crust of varying strength and thickness. Runoff processes lead to local erosion and the incision of small rills and gullies (YAIR, 1990). Fine grained sediment (clay) is deposited in local depressions at the end of these gullies. Other patches of fine grained material have been identified as remnants of previous fluvial processes (HARRISON & YAIR, 1998) caused by flooding from Nahal Nizzana.

The dunes at the research site are aligned roughly west-east (86/266°). Local deviations from this general trend reach up to 10° with implications for the cross sections of these parts. While generally the northfacing slopes are steeper, ridges with a deviation towards more northerly directions display steeper southfacing slopes. Dune heights reach up to 15 m. The distance between ridges ranges from 150 m to 200 m. Several ridges converge at Y-junctions, a typical feature of vegetated linear dunes (PYE & TSOAR, 1990, p.208). The majority of the ridges originate on the eastern bank of a S-N-running wadi approximately 5 km west of the border on Egyptian territory. All ridges are terminated at Nahal Nizzana, 2.5 km east of the border. Episodic flooding of this wadi currently prevents the dune ridges from extending further eastward.

1.5.4 Vegetation

The vegetation at Nizzana can be divided into a macrophytic and a microphytic part. The macrophytes are dominated by perennial shrubs and cover the interdune corridors and increasing areas of the dune ridges. A microphytic crust (WEST, 1990) composed of algae and fungi covers most surfaces in the interdune corridors and stabilised dune ridges.

The vascular vegetation can be divided into seven distinct plant communities (TIELBÖRGER, 1997). These communities correspond largely with the morphologic features, which indicates a connection between biotic and abiotic properties at the site. Perennial shrubs dominate the species rich interdune corridors, where vegetation cover is estimated to be 21 per cent (TIELBÖRGER, 1997, p.270). Cover at the plinths is generally higher than in the interdune corridors. The dune crests are dominated by a perennial grass (*Stipagrostis scoparia*) and the semishrub *Heliotropium digynum* (TIELBÖRGER, 1997, p.264).

A very distinct feature of the research area is the presence of a microphytic surface crust (YAIR, 1990; VERRECCHIA *et al.*, 1995; PRASSE, 1999). It is composed of algae and fungi which form a hard crust of varying thickness and strength when dry. Five types of crust were distinguished by KIDRON (1995). Dominating are cyanobacteria, mainly *Microcoleus sociatus* (LANGE *et al.*, 1992), mosses and lichens play a secondary role. The floristic composition shows six species of cyanobacteria-, two green algae- and two moss species (KIDRON, 1995). Wetting of the crust causes swelling of the organic parts and the clay minerals. The crust is present in all interdune corridors as well as on the foot slopes of the dune ridges. It covers the surface of these areas almost entirely. In general, strength and thickness gradually decreases towards the crests of the dune ridges. Where the dune ridges are degraded, patches of the crust are also found on the crest area.

1.5.5 Land use history - the human impact

Artefacts found in the research area show that the area has been inhabited since the Paleolithicum (Goring-Morris & Goldberg, 1990). Relicts of later periods indicate a continuous agricultural use until present times. During the past 50 years, the linear dunes of the Sde Hallamish site have undergone several considerable changes in land use, caused mainly by political developments. Since the end of Israel's war of Independence in 1949, most of the Bedouin living in the Israeli Negev have either been forced into Egypt or have been relocated by Israeli authorities to live around the town of Beer-Sheva. Until 1967 the closed border prevented the use of the sand dunes of Sde Hallamish for grazing purposes on the Israeli side. At the same time, the increased population density in the Egyptian part of the Sinai led to severe overgrazing of the sand dune area. Following the occupation of the Sinai peninsula by Israeli forces during the war of June 1967, the situation changed. During the period 1967 until 1982, Bedouin nomads were able to graze their flocks in the research area, as the former border was not enforced anymore (Tsoar & Møller, 1986; Meir & Tsoar, 1996). The existence of a well at the confluence of Nahal Nizzana and Nahal Levan east of the dune field made the area especially attractive. Aerial photographs taken during this period show a complete destruction of the vegetation and a distinctive change of dune morphology within 10 years (Tsoar & Møller, 1986).

After the establishment of the international border between Egypt and Israel, following the peace treaty of 1981, access to the area for the purpose of grazing was again denied on the Israeli side. The cessation of grazing led to a quick recovery of vegetation. Aerial photographs of 1984 already show a distinctive color difference in the interdune corridors between the Israeli and the Egyptian side of the border, where grazing continues. In 1986, West (1990, *p.204*) reports a well-developed microphytic cover on the Israeli side. Since 1982 until today, no agricultural use has been made of the research area. Until recently the area has been only accessed for scientific purpose. During the period of the study increasing disturbance was observed, caused by army vehicles and private off-road vehicle traffic.

Methods

Surface properties, vegetation and morphology each have a distinguished influence on aeolian sand movement. Attempts have been made to single out the specific influence through the choice of location for the measurement sites, treatment of natural vegetation or surface cover and the setup of the research equipment at the sites. The locations as shown in figure 2.1, p.23 were chosen based on previous geomorphological and botanical studies conducted in the research area of Nizzana (Tsoar & Møller, 1986; Yair, 1990; Allgaier, 1993; Tielbörger, 1997). Site A was established in an interdune corridor. Vegetation and surface crust was removed to create an open sand surface with no restricting influence on sand movement other than the position within the corridor. Site B has been set up in a similar morphologic position in order to make vegetation or surface cover the distinghuishable difference to site A. One part (B1) has been left unchanged in its natural condition with a crust covered surface and a vegetation cover of around 17 per cent. On the other part (B2) the surface crust has been removed to investigate its effect on sand movement. In a later stage vegetation cover was reduced from 17 to 9 per cent. At site C the processes on a partly stabilized dune ridge without mobile sand bodies have been observed. Site D was set up on an exposed dune crest with open, rippled sand surfaces and widely distributed accumulations with slipfaces, ie morphologic features indicating high aeolian activity. Here, the influence of the topographic position, mainly the exposition of the surfaces to sand transporting winds and the availability of loose sand for transport could be determined.

Table 2.1: Data acquisition periods at the measurement sites.

	A		B		C		D	
period	begin	end	begin	end	begin	end	begin	end
I (1997)[a]	-	-	20/1	22/5	-	-	13/1	23/3
II (1997/98)	24/11	13/4	7/11	13/4	29/12	13/4	10/11	10/5
III 1998/99)[b]	23/11	25/4	24/11	25/4	24/11	25/4	27/11	25/4

[a] no use of *saltiphones*
[b] measurement of surface change at site D until 13/10

Data was collected during the winter and early summer of three consecutive years (1997-1999, table 2.1). All sites were equipped with an array of three anemometers to measure wind speed at different heights above the surface and a wind vane to determine wind direction. Sand traps were installed to determine sediment transport. In addition, saltation sensors were installed at sites A, B and D. The resistance of the microphytic crust against erosion by wind has been tested experimentally in the wind tunnel of Ben–Gurion University.

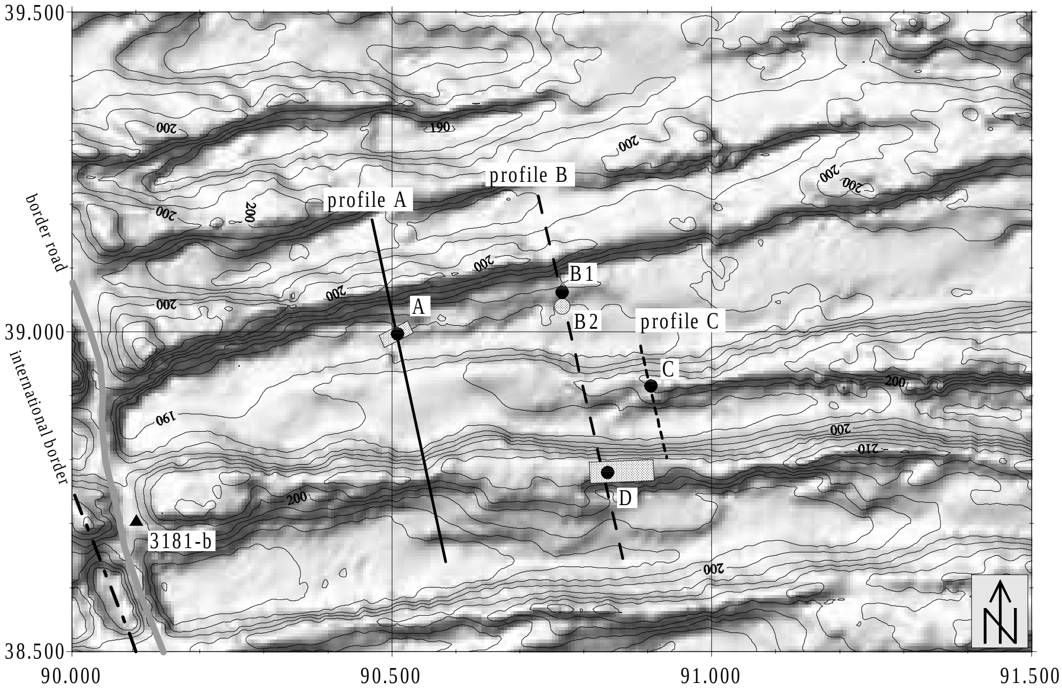

Figure 2.1: Map of the research site Nizzana/Sde Hallamish. Based on elevation data of 1989 provided by the AERC. Isoline distance is 2 m.

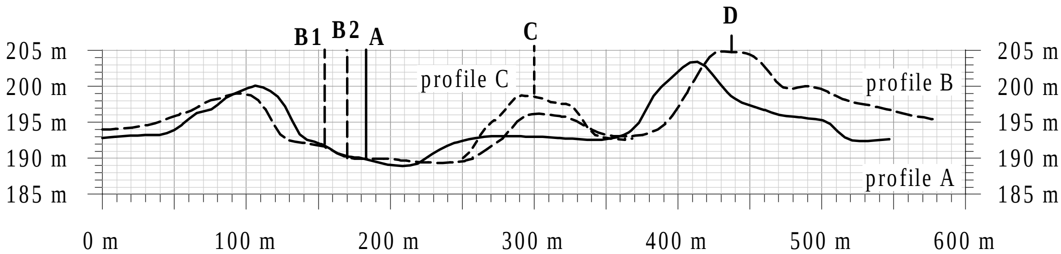

Figure 2.2: Cross sections at the sites of measurement (see figure 2.1). Based on elevation data of 1989 provided by the AERC. Vertical exaggeration 1:5.

2.1 Site selection and description

The most apparent differences in surface structure and surface cover are found between the dune crests and the interdune corridors. The active state of many crestal areas is obvious, documented by the presence of ripple marks, open sand surfaces and active slip faces. In the corridors, most surfaces are crust covered, so that if the loose sand present in patchy thin layers is moved, it does not leave evidence after transport. To investigate the flow conditions and transport processes in the interdune corridors, three sites were set up. Two more sites were placed on dune ridges of apparently differing grades of activity.

2.1.1 Interdune corridor sites

The sites were situated in areas which PRASSE (1999, p.40f) calls 'flache Dünentäler'. The surface is covered by a microphytic crust with an average thickness of approximatly 1 mm, thin layers of loose sand appear in patches. The vegetation has been described as *Echiochilon fruticosum - Thymelaea hirsuta* community (TIELBÖRGER, 1997). Annual vegetation is concentrated under the canopy of perennials or in their immediate vicinity. Patches of annuals appear mainly in areas where the crust had been disturbed previously (PRASSE, 1999). The sites were set up to represent different possible states of surface condition. Site A resembled the condition as found prior to 1982, when the surface crust had been destroyed by the trampling of grazing animals and the edible vegetation had been consumed. Aeolian sand transport is not hindered by surface obstacles or cementing, therefore available wind energy is the restricting factor. Site B1 represented the current interdune surface condition after 15 years of undisturbed development, while at B2 the surface crust was taken away to evaluate its influence on aeolian activity. In a later stage, vegetation density at site B2 was reduced to half the natural value to determine the impact on sand transport.

2.1.1.1 Site A: no vegetation, no surface crust

Site A served to determine the influence of vegetation and the influence of morphologic position on aeolian sand transport. To achieve this, all vegetation and the microphytic crust was removed from the surface. Through comparison with data recorded at sites B1 and B2 (see below) it was possible to determine the degree of surface protection offered by different vegetation cover under similar flow conditions. By comparing wind and sand transport data of sites A and D (dune crest), the influence of morphologic position could be evaluated, as the surface condition resembled the dune crest (open sand). At the site, the corridor is 90 m wide. Height difference to the adjacent dune in the north is 10 m, the neighbouring ridge in the south is a subdued feature. Its crest reaches only about 3 m above the surface of the corridor at the position of the instruments of site A. With the upper anemometer being located 1.77 m above the surface, this leaves a height difference of roughly 1.3 m (see figures 2.1, p.23 and 2.2, p.23). For wind directions normal to dune ridge orientation, the effect of different 'obstacle' heights could thus be determined.

The cleared area was of rectangular shape with a total area of 1027 m^2, its long axis oriented along the expected wind directions of the major winter storms (see fig. 2.3, p.25). Six sand traps facing SW were deployed: Four (A/1 - A/4) located in the centre part of the rectangle, close to the anemometer

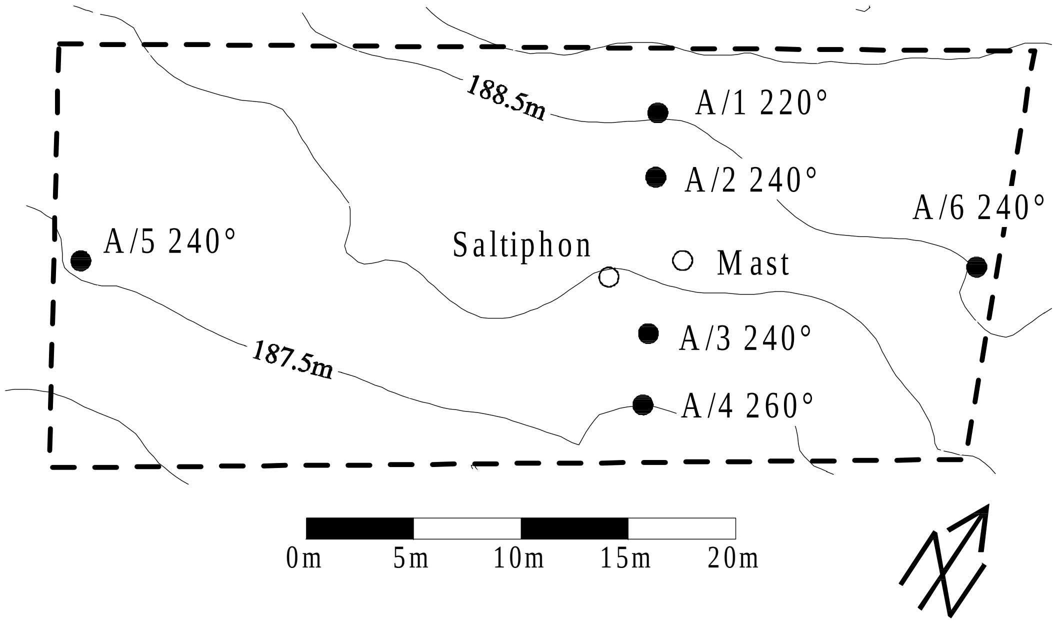

Figure 2.3: Setup of site A.

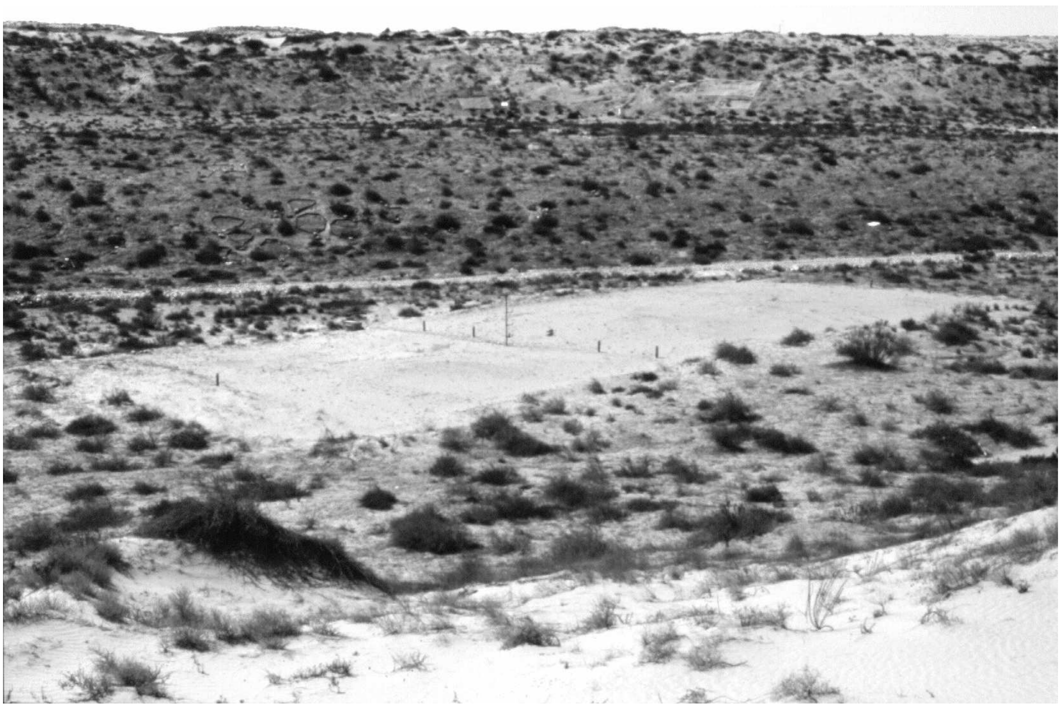

Figure 2.4: View of site A from NE, 17/3/1999.

mast and the saltiphone. Trap A/5 was placed at the western (upwind) end of the test area to record material coming into the area, while trap A/6 at the eastern (downwind) end measured material transported out of the site.

2.1.1.2 Site B: natural vegetation cover, with and without microphytic surface crust

Site B was used to investigate the influence of vegetation, both macrophytic and microphytic, on sand movement. It has therefore been divided into two sub-sites. They will be referred to as B1 and B2.

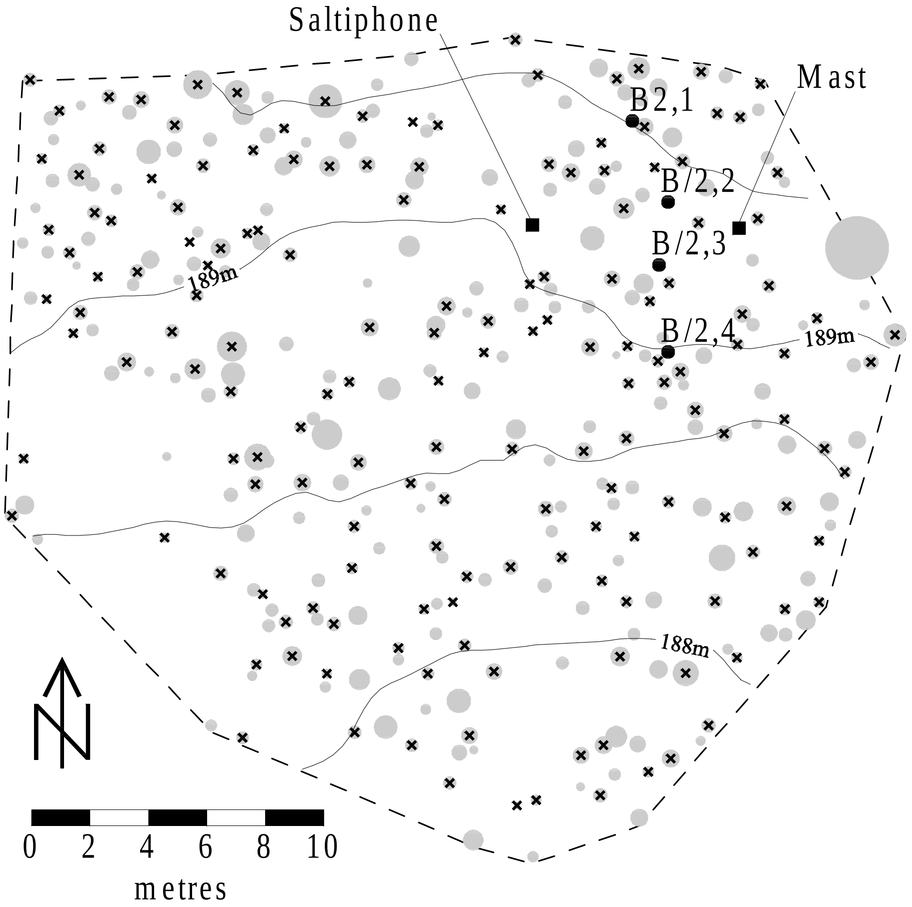

Figure 2.5: Setup of site B2. Grey circles show the planar area of the vegetation patches. Plants marked with 'x' were removed on 23/1/99.

Site B1 was used to determine the aeolian activity in the interdune corridor under present undisturbed natural conditions. The surface crust and the vascular vegetation were left undisturbed throughout the duration of the project. Crust cover was 100 per cent of the surface area not occupied by macrophytic vegetation. Natural perennial vegetation cover was approx. 17 per cent. The mean vegetation height at sites B1 and B2 was 0.37 m. Four traps facing W,

Figure 2.6: View of site B2 from SE after reduction of plant cover to 9.5 per cent, 18/4/1999. The mast of B1 is visible in the upper left corner of the image.

NW, NE and S were installed within a 6 m wide circle around the mast holding the anemometers. The site was accessed via a narrow path east of the traps in order to minmize disturbance of the surface crust and ensuing mobilization of material.

Site B2 was set up approximately 20 m south of B1 in order to investigate the influence of the microphytic surface crust on aeolian transport. The crust of the SW-sector was destroyed within a radius of 25 m from the anemometer mast by raking the surface. Vascular vegetation cover was similar to B1 until 23/1/99. On this date the natural perennial vegetation cover was reduced to 9.5 per cent in order to evaluate the effect of reduced cover on sand transport rates. The four traps at B2 were installed to face the high magnitude winter winds from the western sector.

Vegetation patch positions and sizes at site B2 were determined in January 1998, using an electronic theodelite (total station). Figure 2.5, p.26 shows the area and set-up of site B2. The area of the circles represents the actual planar cover of the plants. Except for two individuals of *Retama raetam* all vegetation patches are broadest at their base on the surface so that they will affect the airflow immediatly above the surface most. The plants that were removed in January 1999 were chosen randomly from the list created during the EDM-survey. The total area S_{B2} without surface crust was 615 m^2. The sum of the vegetation patch sizes was 108 m^2, assuming circular shapes of the 326 (n) patches. Calculated vegetation cover was 17.6 per cent. Vertical projected area A' was 59 m^2. Usually A' is determined as the product of the average element height and diameter, assuming a cylindrical shape. Observation of the plant patches of site B2 led to the conclusion that the use of a rounded shape would better represent their actual appearance. Therefore A' has been determined as

$$A' = \sum \frac{\pi r^2}{2} \tag{2.1}$$

where r is the mean of patch height and patch radius. Based on the above information the roughness element concentration L_c can be calculated. It is defined by Wolfe & Nickling (1993, p.57) as

$$L_c = \frac{nA'}{S_{B2}} \tag{2.2}$$

At site B2 L_c of natural vegetation was 0.096 and 0.052 after the removal of half of the vegetation patches. Based on published values for flow regimes and associated roughness element concentrations (Wolfe & Nickling, 1993, p.57), the undisturbed vegetation at site B should result in *wake interference flow*, where the wake zones overlap, thus reducing the area prone to be influenced by the airflow. After the reduction of vegetation site B2 fell into the *isolated-roughness flow*-class, where each vegetation patch develops its own 'wake and separation region' (Wolfe & Nickling, 1993).

The morphological situation is similar to site A, but as the corridor chosen for the measurement sites widens towards the east, it is 120 m wide at site B, compared to 90 m at site A. The height difference towards the northern dune is smaller at site B, being 7.5 m for B1, resp. 8.5 m for B2. For the southern dune values are 4 m resp. 5 m for B1 and B2.

2.1.2 Dune ridge sites

Within linear dune ecosystems, the dune crests are considered to be the most active part. To assess the aeolian activity on different dune crests in comparison to the interdune corridors, two sites have been set up on two dunes which differ in vegetation cover and surface structure.

The majority of the dune crests at Nizzana appear to be partly stabilized. The crests are dominated by vegetated hummocks of up to 1.5 m height, the total plant cover of these areas has been estimated to be above 20 per cent (Tielbörger, 1997). Open sand surfaces are restricted to areas between the hummocks. These open areas are level, ripple marks are indicators of aeolian processes.

Open, non-vegetated sand surfaces are found only at the highest and most exposed dune crests at the research site. These areas are usually about 30 m wide, with generally low angled surfaces, except where slip faces developed. Signs of aeolian activity, such as ripples and slip faces, are common throughout the crest, while vegetation cover is sparse. Overall vegetation cover is estimated to reach a maximum of 10 per cent.

2.1.2.1 Site C: partly stabilized dune crest

The location has been chosen to estimate aeolian activity on a dune ridge which is partly stabilized. Hummocks covered by vegetation and microphytic crust dominate the crest area. Signs of destruction like protruding plant roots due to erosion at their bases are common.

Figure 2.7: View of site C and vicinity from W, 10/5/1998.

The ridge has an asymmetric cross section with a steep, high northfacing slope and a gentle, low southfacing slope. The adjacent interdune corridor to the south is narrow, its surface consists in general of weakly cemented sand. The border between the northern slope and corridor is marked by a sharp knickpoint. Northern slope and corridor surfaces are crust covered. The crust disappears at the upper knickpoint where the slope borders on the former crest area. The vegetation at this site has been characterised as *Moltkiopsis ciliata - Convolvulus lanatus* community with a mean plant cover of 22 per cent (Tielbörger, 1997, p.267). The appearance of the ridge changes continuously from W to E as its height increases. On its western part, south of site A, it is completely covered by the microphytic crust, while 500 m east of site C it resembles the dune crest at site D (see below).

The anemometer array and four sand traps were placed in a north-south oriented 'channel' of bare sand. This orientation appeared to be common on the dune ridge. Two traps at the northern side of the crest faced NW, two traps in the centre part of the crest faced S and W.

2.1.2.2 Site D: active dune crest

The location of site D (figure 2.1, p.23) was chosen to represent a highly active, exposed dune section. On the scale of surface activity within the research area, it was to be at the opposite end compared to site B1. The possible influence of vegetation, surface conditions or obstacles on airflow or sand transport was smallest at the site. Thus wind speed, wind direction and saltation recordings taken at site D have been used as reference values, against which the data of the other sites has been compared. To determine seasonal changes of morphology and the sediment budget of active dune parts, surface changes were monitored at site D in addition to the measurements of wind parameters and sand transport.

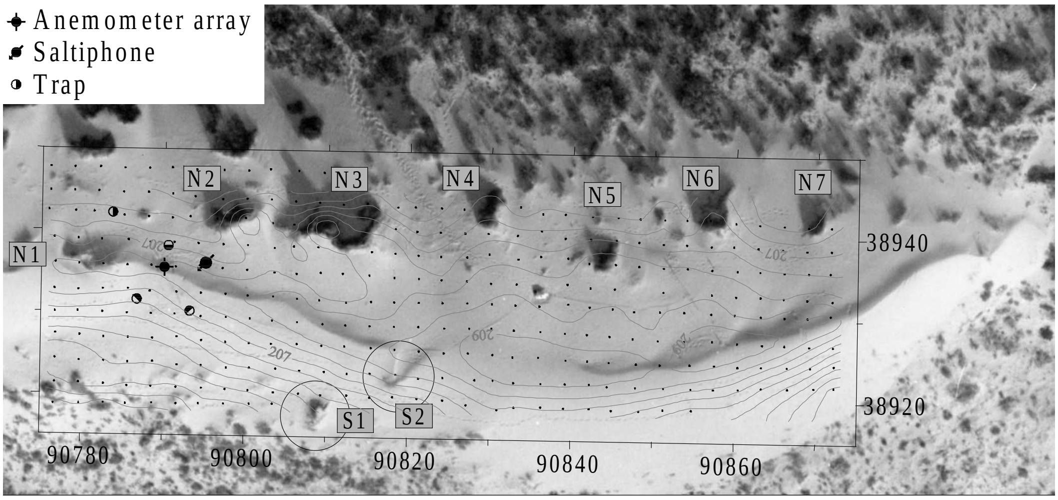

Figure 2.8: Setup of site D. Isolines show topography on 24/11/1997, the aerial photograph was taken on 22/11/1997. Dots mark the positions of erosion pins.

Dune height at the site was approximately 12 m. The dune crest consists of an area with no vegetation at the centreline in its eastern part and a western part with three well developed vegetated hummocks, situated close to or at the centre of the crest. Plants are concentrated at hummocks up to 2 m high. These are commonly occupied by *Stipagrostis scoparia* or *Cornulaca monacantha*. The vegetation protects the underlying sand from erosion and a thin microbiotic crust is often found under the plants. Annual vegetation is present only on or in the immediate vicinity of the hillocks (Prasse, 1999). Tielbörger (1997) describes the vegetation as *Stipagrostis scoparia - Helitropium digynum* community. To ease description, the vegetation patches at the site have been numbered N1 to N7 north of the crestline and S1, S2 on the southern side of the ridge (see figure 2.8).

A distinct knickpoint marks the border between the crest and the lower part of the ridge (figure 2.10, p.31). The slope bordering at the crest is straight and steep, up to the angle of repose of sand. The lower part of the ridge shows a concave curvature towards the interdune corridor. The length of the

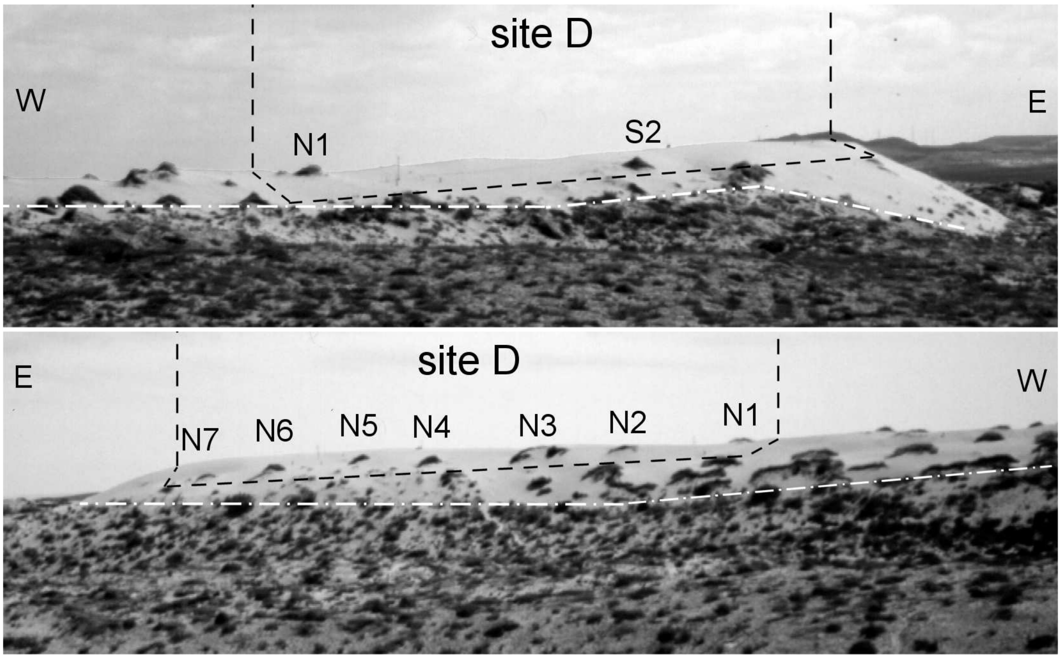

Figure 2.9: Views of site D from SW (upper picture) and NW (lower picture), 21/4/98. The white line marks the border between crusted and non-crusted surfaces.

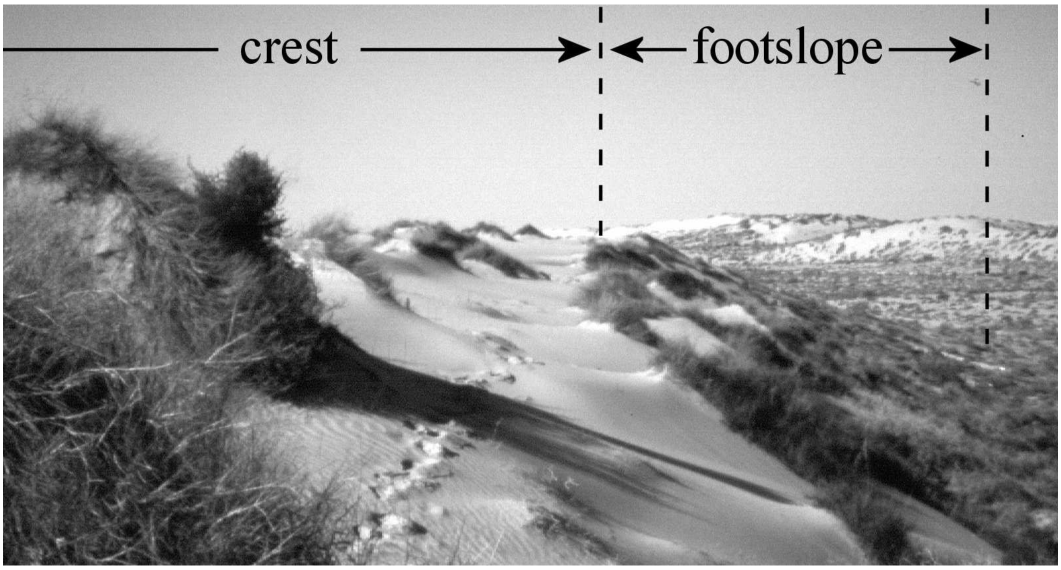

Figure 2.10: Transition area between crest and footslope on the northern slope at site D. Picture taken at N4 (see figure 2.8, p.30) looking W. Note vegetated hummocks along the knickpoint and the centerline of the dune crest.

Figure 2.11: Instrumentation at site D, December 1998. Ripples and scour marks indicate recent sand transport by northerly winds. Also note accumulated organic litter in the lee of hummock N1 visible to left of the anemometer array.

lower slope differs between the northern and the southern side of the ridge because the surface of the southern interdune corridor is inclined towards the ridge (figure 2.2, p.23). Vegetation cover increases downslope, reaching its maximum at the transition zone between dune and corridor surface. A thin surface crust develops in areas not affected by avalanching of sediment. The white line in figure 2.9 depicts the approximate border between crusted and non-crusted areas at the dune. At the eastern termination of the ridge section, no footslope exists, a convex slope of loose sand connects interdune corridor and dune crest. Thus avalanching sand from the dune crest can reach the interdune corridor.

The instrumentation is shown in figure 2.11. A set of four (1998/99: five) traps was placed at approx. 5 m distance from the mast. The high mobility of the surface led to repeated failure of parts of the instrumentation, so that the continuity of the data records is not as good as at the interdune sites. Most common causes for failures were filling of the traps and covering of the *saltiphone* and/or the lowermost anemometer by accumulation of sand. Deflation caused tipping of the *saltiphone* and excavation of sand traps.

To estimate the input of material to the dune ridge from less vegetated areas on Egyptian territory, a sand trap was placed immediately east of the border road, close to the trigonometric point 3181-b (see figure 2.1, p.23).

2.2 Instrumentation

To investigate aeolian transport processes it is essential to know wind speed and direction during the actual transport event. In order to obtain continuous data with sufficient temporal resolution, data loggers were used to automatically record wind speed, wind direction and saltation. Due to the distance between the measurement sites, three loggers had to be used. One for site A, one for sites B and C and a third one for site D. Logger time was synchronized prior to installation and was checked during the weekly visits to the site.

2.2.1 Wind speed

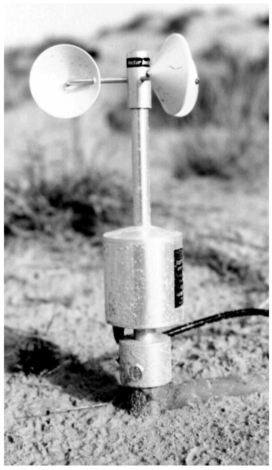

Figure 2.12: *Porton A100*

Long term recordings of wind speed and direction have been provided by the AERC. These data sets contain hourly mean wind speed and direction and were used to determine the general wind regime of the study site. Detailed recordings of wind speed were taken using cup anemometers (*Porton A100*, manufactured by *Vector Instruments*). The instruments create a voltage which is proportional to the wind speed. Their threshold speed is 0.15 m s^{-1}, which is well below the threshold speed for sand movement. Three anemometers were attached to a mast at 0.24 m, 0.65 m and 1.77 m above the surface. This spacing allows for an estimate of u_* by simply using the wind velocity difference between two heights, where the upper height is 2.718 times the height of the lower (Bagnold, 1941, p.51). Apart from the possibility to determine u_*, the measurement of wind speeds at different heights was considered vital, as it provides information about the influence of small scale morphology and vegetation cover on the near surface wind speed (sites B and C). Wind velocity was measured in 10 second intervals. These measurements were averaged over two minutes. The length of the recording intervals was chosen in order to minimise the effect of averaging on the apparent threshold speed for aeolian transport (Stout, 1998) while enabling the system to run at least one week before the memory of the data logger would be filled. Because of the uncertainties connected with the measurement and determination of the parameters necessary for the calculation of shear stress (see section 1.3, p.14), I decided to use the horizontal wind velocity measured at 0.24 m above the surface ($u_{0.24}$) as the independent variable for the induction of aeolian sand movement.

2.2.2 Wind direction

Wind direction was determined at each site at a height of 2.30 m by wind vanes attached to the top of the masts holding the anemometers. Direction readings were not averaged, actual values were stored parallel to the wind speed data.

2.2.3 Saltation sensors

Accoustic saltation sensors (*saltiphones*) have been used to determine periods and magnitude of saltation. The *saltiphone* produced by *Eijkelkamp* consists of a microphone housed in a horizontal steel tube with a diameter of 5 cm. Wind vanes keep the tube in line with the wind direction. The microphone is connected to an electronic filter which relays only signals caused by impacting sand grains. These signals are converted into impulses, which are recorded by a data logger. The maximum count rate is 500 pulses per second. A more detailed technical description of the *saltiphone* is given by Spaan & van den Abeele (1991).

The *saltiphones* were set up at a distance of approximately 3 m from the anemometer arrays to reduce interference. The centre of the microphone membrane was set to a height of 4.5 cm above the sand surface in order to measure saltation as close to the surface as possible. A free space of 2 cm below the protective steel tube proved to be sufficient in most cases to allow for the free rotation of the saltiphone. The instruments were connected to the station's data logger. The total number of impacts within the recording interval of the wind speed sensors were written to the data logger.

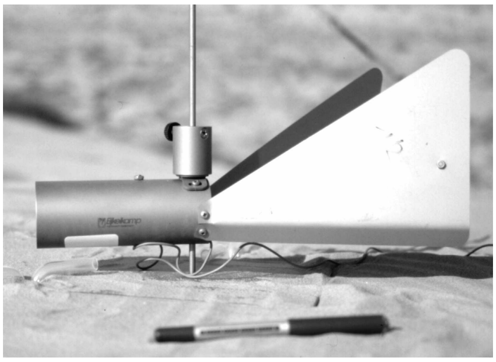

Figure 2.13: *Saltiphone*, wind direction from left.

2.2.4 Sand traps

Sand traps were installed to detect sand movement and to determine the amount of material moved in connection with the *saltiphones*. The traps used were of the *Leatherman*-type (Leatherman, 1978; Rosen, 1979). This type of sand trap was chosen on account of its simplicity and sturdiness. It has no moving parts that can be jammed or damaged during prolonged times of exposure to adverse climatic conditions. Its main disadvantage is that it does not rotate towards the wind. Therefore full efficiency is only reached when the airflow is in the direction of the opening. Wind tunnel experiments (see below, p.36) showed that trap efficiency for oblique winds can be calculated if the angle between trap opening and wind direction is known.

The traps used were built of PVC tubes of 100 cm length and 11 cm outer diameter. When emplaced, the lower half of the trap is buried and serves as a collector for the trapped sediment. The upper half provides two openings flush with the surface. A narrow one serving as the entrance for moving sediment and a wide one to provide throughflow of air. The narrow, windward opening measures 48 cm x 6 cm. A mesh of 0.063 mm covers the wide opening which spans half of the circumference of the tube. The volume of the collector is 6283 cm^3, based on an inner diameter of 10 cm and a height of 50 cm. To prevent scouring at the base of the traps, aluminium disks (2 mm thick, 30 cm diameter) were placed around the traps. The size of the disks proved to be sufficient under all conditions at sites A, B and C, while at site D strong winds caused scouring despite the protection (see figure 2.14). Prior to the installation of the traps tubes with an inner diameter of 12 cm were buried in the ground to provide a stable hole in the non-cohesive dune sand to ease handling and adjusting. The void between the trap and the outer tube was sealed with strips of rubber foam.

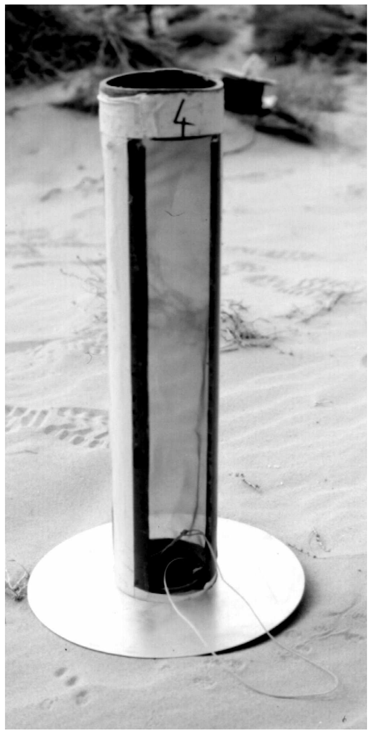

Figure 2.14: *Leatherman* trap fitted with aluminium disk to prevent scouring at the base.

Four *Leatherman* sand traps were placed in the vicinity of each of the anemometer masts, facing towards the expected main wind directions. The traps were checked for sediment at least once a week during the winter season. Small amounts of sediment were emptied into plastic bags and taken to the laboratory to be dried and weighed. Larger amounts were determined by measuring the sand level in the tube. The mass was calculated by assuming a specific weight of the sand of 1.6 g cm^{-3} (Pye & Tsoar, 1990, p.287). This value was verified by the comparison of the calculated and weighed mass of several samples.

The height of the sand trap is sufficient to trap the majority of the saltating sand particles as the mean height of the trajectories does not exceed 5 cm (Pye & Tsoar, 1990, p.105). Saltation heights on loose surfaces like sand are even smaller (Bagnold, 1941, p.36) and the distribution with height is highly skewed (Pye & Tsoar, 1990, p.107). The trap also catches particles moving along the surface (creep). The capacity of the collector proved to be sufficient for the majority of the sand transporting events at the sites A, B and C. At site D high transport rates and changes of morphology led to rapid filling of the traps or burial by avalanching sand from advancing slip faces. Collectors at the site were observed being filled within less than 10 minutes during heavy storms. Sand transport at site A during the exceptional storm in March 1998 also exceeded trap capacity (see 3.4.1.2, p.62).

Trap efficiency for oblique winds As has been mentioned above, the traps do not rotate to face actual wind direction and therefore their trapping efficiency varies. In order to determine the correlation between trap opening and wind direction, experiments were carried out in the wind tunnel at Ben–Gurion University. A trap as used in the field was placed at the downwind end of the test section of the wind tunnel. Dune sand taken from the research area was fed into the flow at a constant rate at the upwind end of the test section. The carrying capacity of the flow was adjusted to exceed input rate of sand. Several runs were conducted with the trap opening directly facing the flow direction to determine the amount collected at ideal conditions (direction difference 0°). During successive runs with unchanged feeding rate and flow velocity the opening of the trap was rotated in steps of 10° until a direction difference of 90° relative to the flow direction. Repetitions were made in both directions (left and right in flow direction) in order to detect possible helical effects within the wind tunnel. Amounts collected at angles other than 0° were calculated as fractions of the amount collected at 0°. Thus the dependency of wind direction on trap efficiency could be determined independent of the actual conditions in the wind tunnel compared to field conditions.

A linear correlation was established. Based on the experiments, trap efficiency E in per cent depending on wind direction can be written as

$$E = 100 - 1,24x \quad \text{for} \quad x < 80 \tag{2.3}$$

where x is the difference in degrees (°) between measured wind direction and orientation of the trap opening.

2.3 Wind speed differences

Wind speed differences have been calculated between site D and sites A, B and C. Values have been determined for $u_{0.24}$ ($S_{0.24}$) and $u_{1.77}$ ($S_{1.77}$), as the behaviour of accelerated airflow has been shown to be dependent of height above the surface. In addition, surface roughness differs at the sites and an influence on near surface wind speed was expected. The calculation is based on all datasets where wind data was recorded at all three sites and where $u_{0.24}$ at site D was above 4 m^{-1}.

Published studies on wind speed differences mainly focus on the upslope increase of wind speed on the windward flank of dunes. Lancaster (1985, p.582) uses the 'speed-up factor' S which is defined by Mason & Sykes (1979) as

$$S = \frac{V_{crest}}{V_{base}} \tag{2.4}$$

where V_{crest} and V_{base} are the wind speeds measured at the dune crest, resp. at the dune base. The term *speed-up factor* for S implies a positive acceleration of a given wind speed. Due to the nature of the research subject of this study – parallel linear dunes – the 'true' undisturbed wind speed could not be determined over a flat, open surface upwind of the research site. Therefore the – generally higher – wind speed measured at the dune crest (site D) was taken as the reference wind speed and was set into relation to the data obtained at sites A, B and C, situated at lower locations. Thus, in the following, the *speed ratio* S_z is defined as

$$S_z = \frac{u_{zX}}{u_{zD}} \tag{2.5}$$

that is, the ratio between the wind speed u_{zX} at the sites in the interdune corridor (A and B) resp. at the low ridge (site C) and the wind speed at the dune crest (u_{zD}) at corresponding heights z above the surface.

2.4 Wind direction differences

The deflection of sand moving winds will alter sand transport direction. To determine the amount of deflection δ_W at sites A, B and C, direction values measured at these sites have been subtracted from values measured at site D. The direction recorded at site D is assumed to represent the direction of the wind not influenced by the linear dune ridge as it has been measured at 2.3 m above the highest part of the ridge. The database contains all data sets where wind speeds at 1.77 m above the surface ($u_{1.77}$) exceeded 2 m s^{-1} at site D, 0.5 m s^{-1} at sites A, B and C and δ_W was below 90°. It is assumed that differences of more than 90° are either caused by wakes or no continuous flow within the dune field due to low wind speeds.

2.5 Calculation of sand flux

Sand flux q at sites A, B and D was calculated based on data obtained by the *saltiphones* and the sand traps. The calculation procedure was developed after efforts to calibrate the saltation sensors through experiments in the wind tunnel failed. The method makes use of the results of Greeley *et al.* (1996), includes the trap efficiency E as determined in the wind tunnel experiments (see p.36) and the data records of saltation, wind direction and mass of material caught in the sand traps.

In a first step, all relevant data records have been filtered out of all recorded data sets. Relevant data records are those where the *saltiphones* recorded impacts and were data of sand mass from the sand traps is available for the same period. The trap efficiency E for each trap was then calculated for each record based on equation 2.3, p.36. If the direction difference x between air flow and trap opening is 0°, 30 per cent of all grains recorded by the saltation sensor are considered to end up in the trap (Greeley *et al.*, 1996, p.48):

$$I_{it} = I_{rec} * 0,3 \quad \text{for} \quad x = 0 \tag{2.6}$$

where I_{it} are impacts being collected in the trap and I_{rec} are the recorded impacts. For all datasets where $0° < x < 80°$ equation 2.6 has to be rewritten as

$$I_{it} = I_{rec} * 0,3 * E \tag{2.7}$$

in order to take the reduced trap efficiency into account. As the data records contain data of a fixed time interval, I_{it} provides the rate of material entering the trap if the mass of one impact I_{it} is known. To determine the mass equivalent m_I for one measured impact I_{it}, the mass of the sediment sample of each trap m_S is divided by the sum of the I_{it} of the sampling period of the trap.

$$m_I = \frac{m_S}{\sum I_{it}} \tag{2.8}$$

Multiplied by the number of impacts I_{rec} of the data record we get the sand flux q (mass distance^{-1} time^{-1}).

$$q = \frac{m_I * I_{rec}}{d} \tag{2.9}$$

where d is the distance as measured perpendicular to wind direction.

2.6 Resistance of the microphytic surface crust against aeolian erosion

In order to test the resistance of the microphytic surface crust against deflation, an undisturbed sample was taken to the wind tunnel at Ben–Gurion University (Pye & Tsoar, 1990, p.320). It was to be tested if wind speeds common at the site are high enough to cause a destruction of the crust.

To obtain a large undisturbed patch of surface crust, a wooden frame was built to fit the size of the test section of the wind tunnel (width 70 cm, length 90 cm). In December 1997 the frame was taken to the field site where it was placed in the interdune corridor in the vicinity of site B. It was installed so that the upper rim of the frame was level with the surrounding surface. The sand dug out for the depression was filled in the frame. Care was taken that the broken remnants of the crust remained on the top of the sediment filled in. The sediment was then moistened to initiate crust growth and the frame remained in place afterwards until October 1999. The crust grew back quickly and its appearence was similar to the adjacent undisturbed surface. Germinating seedlings were removed in order to keep a 'clean' yet undisturbed crust surface.

For the test run in the wind tunnel, the windvanes of two *saltiphones* were dismantled. The remaining steel tubes containing the microphones were placed at the downwind end of the frame. In this arrangement, the centres of the microphones were located approximately 2.5 cm above the sand surface. Flow velocity was measured with three pitot tubes at 0.032 m, 0.105 m and 0.23 m above the surface which were connected to pressure transducers. Data was recorded in intervals of 1 s using a SQUIRREL data-logger. The shear stress u_* was calculated based on the measured wind profile.

2.7 Changes of morphology

Significant changes in the morphology of the linear dune ridges of Sde Hallamish had been noticed on aerial photographs after land use changes. The study of TSOAR & MØLLER (1986) showed that the destruction of vegetation led to an increase of dune height accompanied by a change of dune morphology. After the re-establishment of the border between Israel and Egypt in 1982 aerial photographs showed indications of a quick regrowth of vegetation in the interdune corridors and on the footslopes of the dunes as early as 1984. A further increase of vegetation cover and a reduction of obviously active areas at the dune crests between 1992 and 1999 were observed on additional aerial photographs taken during that period. It was now to be tested if this increase of vegetation would lead to a decrease in dune height ie a reversal of the process described by TSOAR & MØLLER (1986).

The changes of morphology were investigated at different temporal and spatial scales: direct measurements of daily and seasonal changes were undertaken at the apparently active dune crest at site D (see p.30). Long term developments were analyzed using ground and aerial photographs of different years.

2.7.1 Direct measurements

To monitor and quantify changes of morphology at site D in detail, an array of 310 erosion pins was used. These metal pins (diameter 2 mm, length 900 mm) were staked out in 11 rows of up to 33 pins roughly parallel to the ridge (figure 2.8, p.30). The horizontal distance of the pins was approximately 3 m and their initial length above the surface was set to 400 mm. Based on fixed reference points in the interdune corridor, 3D-positions of the pins were determined using a theodelite (EDM). The coordinates of the reference points were determined with a differential GPS based on the known coordinates of the trigonometric point 3181-b situated next to the border road (see figure 2.1, p.23). To determine surface changes, the length of the pins above the surface was measured with a tape measure at least once a week during winter. Additional measurements were taken after storms as soon as possible to determine the change caused by single events. A total of 91 data sets have been acquired (20 measurements 22/2/97-7/6/97, 40 measurements 7/11/97-10/5/98, 26 measurements 24/11/98-5/5/99). Similar techniques were used successfully on a large Namib dune (Livingstone, 1989), on transverse coastal dunes (Burkinshaw & Rust, 1993) and on vegetated linear ridges in the Kalahari (Wiggs *et al.*, 1995) to evaluate dune mobility. Prior to this study, summer was regarded as a period of low aeolian activity due to the moderate wind speeds recorded. Therefore detailed measurements were only planned for winter and spring. The situation found in autumn 1998 showed that considerable changes take place at the crest during summer. As a result, during the summer months of 1999 five additional measurements of the pins were taken in four-week intervals to provide more detailed information about aeolian activity during this period of NW-winds.

The absolute height of the surface at the pin was calculated through continuous addition of the differences of the length of the pins based on the EDM survey. The calculated heights were used to interpolate surfaces and to generate maps of zones of erosion and deposition. The grid width used for the interpolation is one metre. To compute the volume changes of the dune ridge, the calculated surfaces have been subtracted from each other. All calculations have been done using the software package SURFER.

2.7.2 Long term changes

Direct measurements of processes can cover only small portions of the whole area and also only a short period of time. Therefore ground and aerial photographs of the research site have been used to investigate the visible changes in surface cover and morphology. The earliest available ground pictures were taken in 1993. Aerial pictures from seven different years between 1982 and 1999 were available. The first set shows the initial condition of the area after total destruction of the vegetation immediately after the closing of the border. This state represents the consequences of severe grazing as described by Tsoar & Møller (1986). The last set was taken after the end of the measurements in December 1999. The remaining five sets cover the years 1989, 1992, 1994, 1997 and 1998.

Results

3.1 Wind regime

The basic parameters to describe aeolian sand transport are windspeed and wind direction. Knowing the wind regime of an area allows for a rough estimation of overall aeolian activity, yet to gain a detailed insight into actual processes it is essential to use detailed wind data obtained at different morphologic locations. To determine the regime of sand moving winds and to compare different data sets, TSOAR (1983) offers a method to calculate the *effective wind*:

$$e = t(\frac{\overline{u}}{u_t})^3 \tag{3.1}$$

where $\overline{u}$ is the average velocity above threshold velocity, t is the time (in minutes) for $\overline{u}$, u_t is the threshold velocity for sand movement and e the number of equivalent reference wind minutes. Using equation 3.1 the long term data of the AERC and the wind data obtained during the study period have been transformed to *effective wind* for sectors of 10°.

3.1.1 Long-term data

Wind speed and wind direction measurements taken regularly at the site by the AERC have been used to characterize the local wind regime. The database contained hourly windspeed and direction data from 1991 to 1996. The data was measured 15 m above the interdune surface at a position approximately 70 m east of site A, close to the southfacing plinth of the adjacent dune. The height difference between the interdune surface at the position of the meteorological mast of the AERC and the dune crest was 9 m. Saltation at the dune crests was assumed to start at a threshold u_t of 5 m s^{-1}. A wind regime with two main sand moving directions emerges (figure 3.1, p.42): NW and WSW-winds. The NW-winds prevail during summer, while WSW-winds were recorded during winter. The resultant wind direction for $u_t = 5$ m s^{-1} is 267°, which is parallel to dune ridge orientation.

Based on the AERC´s data, the drift potential (DP) of the area as defined by FRYBERGER & DEAN (1979) is 108 vector units, the resultant drift potential (RDP) is 75 (TSOAR, *pers. comm.*). This value of RDP marks low energy wind environments. The mobility index (M) according to LANCASTER (1988) is 500, which defines fully active dunes.

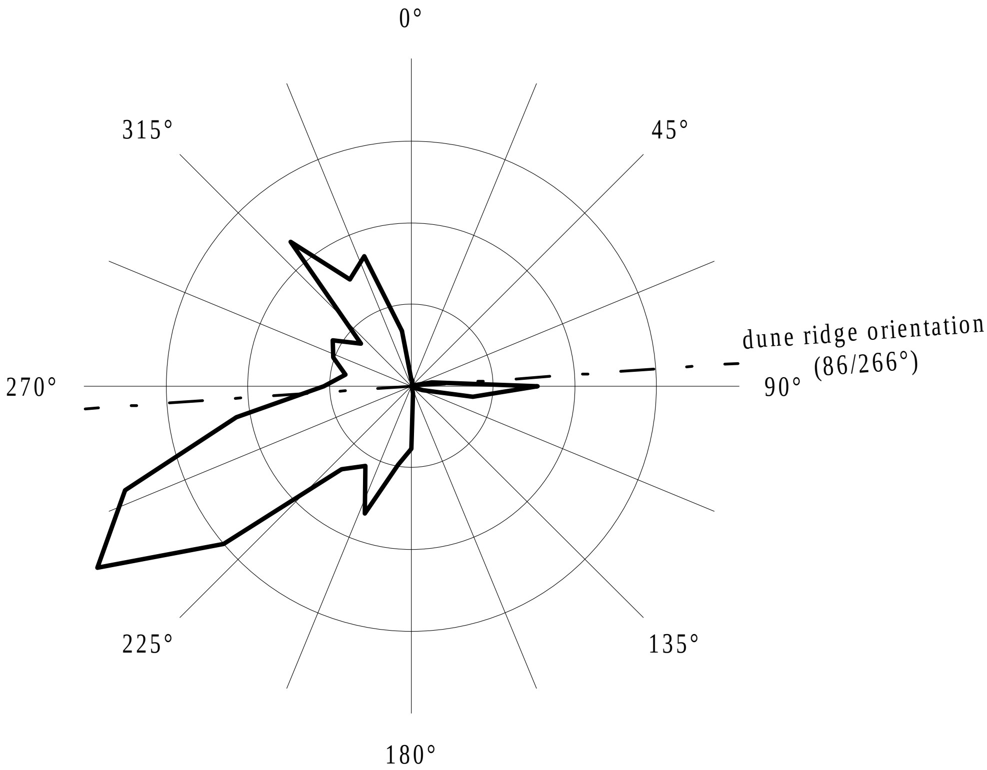

Figure 3.1: Effective wind e for Nizzana. Calculated with an assumed threshold velocity u_t of 5 m s^{-1}. Based on hourly wind recodings 1991 - 1996 recorded at 15 m above the interdune corridor surface. Data provided by the AERC.

3.1.2 Measurements 1997-1999

The *effective wind* for data recorded between January 1997 and May 1999 at sites A, B, C and D is shown in figure 3.2. It has been calculated assuming a u_t of 4 m s^{-1} measured at 0.24 m above the surface and averaged over 2 minutes. This value of u_t is based on the field measurements during the research period, which showed that continuous saltation regularly started as soon as $u_{0.24}$ surpassed 4 m s^{-1}. The lack of recordings taken during summer causes a predominance of southerly to southwesterly winds at the dune crest (site D) in contrast to the AERC-results (figure 3.1). NW to N winds typical for spring and summer are also visible in the graph, but much less pronounced than for long-term, all-year data. The results for the interdune sites A and B indicate that the effective wind e is almost unidirectional, parallel to dune ridge orientation. Resultant wind direction at these sites is approximately 250°. It will be shown below that during the summer months wind speed in the interdune corridor does not surpass u_t. Therefore the resultant wind can be regarded as representative for the whole year. On both dune sites, the low ridge at site C and at the dune crest of site D, resultant direction is close to 220°. In contrast to the interdune sites, this value is not to be regarded as

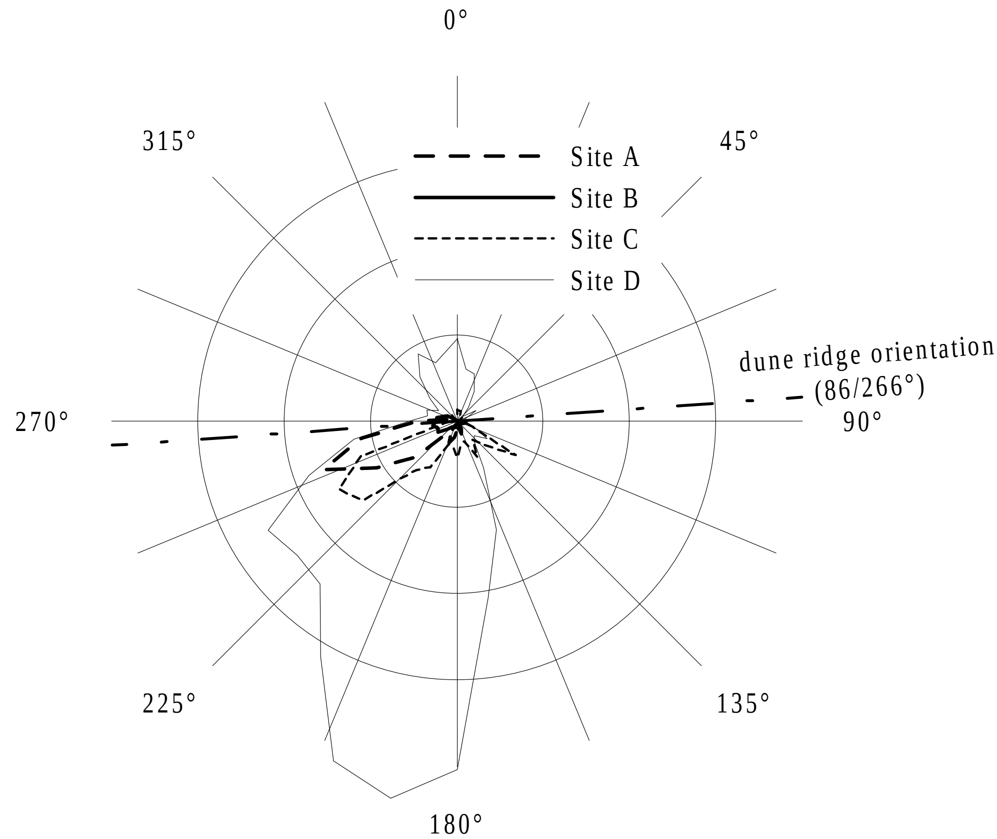

Figure 3.2: Effective wind e for sites A, B, C and D based on measurements at 0.24 m above the surface during the study period. $u_t = 4$ m s^{-1}. Resultant wind directions: 249° (A), 250° (B), 217° (C), 221° (D).

representative for the whole year due to the incomplete dataset, lacking data for the summer months.

In the datasets recorded during the study period, three typical weather situations causing winds above saltation threshold ($u_{0.24} \geq 4$ m s^{-1}) were observed which are characterized as follows:

- summer day
- winter day, stable atmosphere
- winter, passage of low pressure cell

The first two situations are of diurnal nature, following a very regular scheme connected to daytime hours and sunshine conditions. The passage of low pressure cells is independent of daytime, yet confined to the winter season.

Summer day A typical daily cycle consists of a clockwise 360° rotation of wind direction with maximum wind speeds during NW to N winds in the afternoon. Wind speed during the early morning is very low at the dune crest and

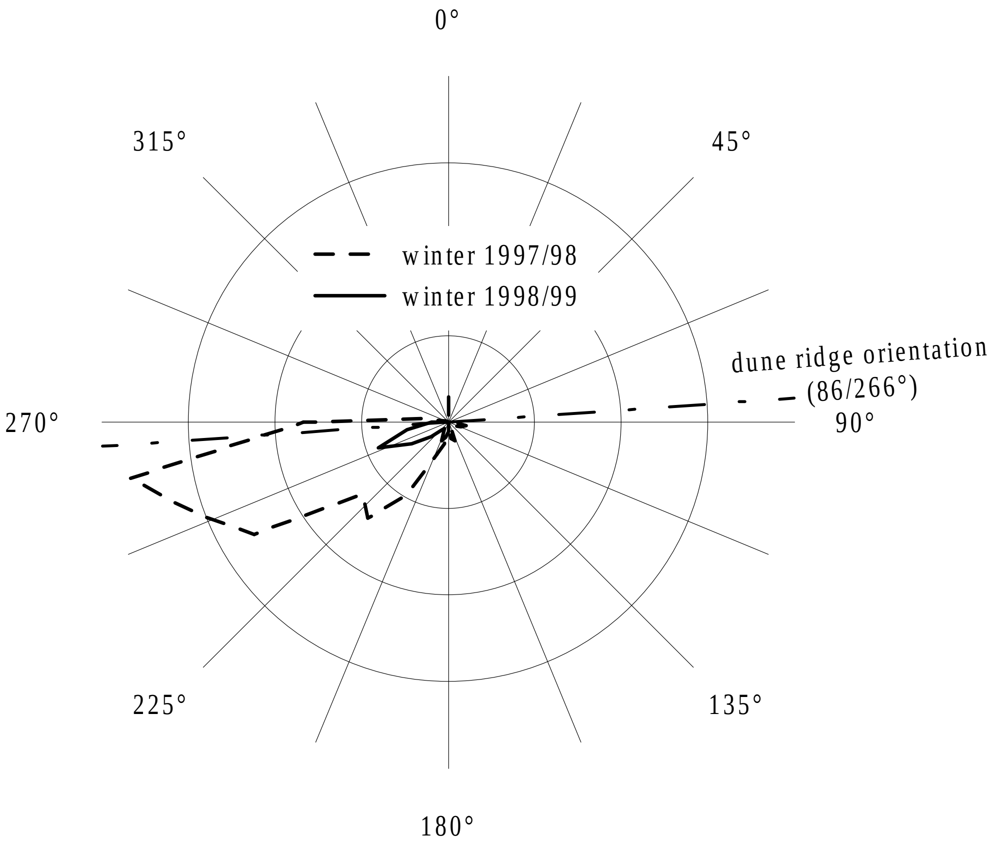

Figure 3.3: Effective wind e for site A for two winters, based on measurements at 0.24 m above the surface during the study period. $u_t = 4$ m s^{-1}.

regularly below instrument threshold (0.15 m s^{-1}) in the interdune corridors. Southerly to southeasterly directions prevail. Starting at sunrise, as the air above the Negev is heated up, wind speed increases gradually and wind direction changes continuously towards N. Saltation threshold u_t ($u_{0.24} \geq 4$ m s^{-1}) at the dune crest is reached in the early afternoon, wind direction at that time has changed to NW. Daily maxima of 6 m s^{-1} are common, 8 m s^{-1} are exceptional (see p.79). Wind speed declines towards sunset, falling below u_t. Wind direction changes to NE during the first half of the night and later to SE. Towards summer, u_t at the crest is reached earlier and earlier and the duration of the event is prolonged. Once established, magnitude and direction of the breeze remain almost constant. In the late afternoon a gradual drop of wind speed is later followed by a change of wind direction. Wind speeds above 4 m s^{-1} were not recorded in the interdune corridors during summer days. The described pattern is repeated each summer day with low variance of wind speed and direction (based on own and AERC data).

Winter day, stable atmosphere As far as wind direction is concerned, cloudless winter days combined with small regional pressure gradients lead to the development of a similar daily wind scheme as observed in summer. Low magnitude SE winds prevail during the night and early morning hours. In contrast to summer days, wind speed increases during the second half of the night prior to sunrise. Saltation threshold u_t at the dune crest is reached early after sunrise when wind direction has changed to S. Maximum wind speed is reached during the morning, wind direction changes continuously clockwise to SW. Wind speed declines during the afternoon while direction changes towards N and reaches its minimum shortly after sunset. As in summer, wind speed in the interdune corridors remained below 4 m s^{-1} during these events.

Passage of low pressure cells High magnitude winter winds out of the SW sector are connected to cyclones passing the eastern Mediterranean from W to E. The highest wind speed values were recorded at the dune crest during winds from S and SW (15/3/98, see p.62). On average these cyclones dominate the weather for about 5 days (Karmon, 1994, p.23), the storm phase lasting approximately 12 hours. The scheme of these cyclonic storms is such that during the initial phase wind speed rises sharply and reaches its peak usually within an hour after the onset of the storm. Initial wind direction is usually SSE, changing gradually to S. In an ideal case, a quick direction change to SW during the event marks the passage of the warm front and a further change to NW shows the passage of the cold front. The saltation threshold at the dune crest is passed immediately at the beginning of the storm. In the interdune corridor, the saltation threshold speed is surpassed only when wind direction is parallel to the dune ridge orientation, except during very high magnitude events (15/3/98, p.62, 17/2/99, p.59). In such cases wind speeds above u_t have been recorded in the corridor during southerly wind directions. Episodic high magnitude sand transporting events during winter span a wide range of magnitude. Average and extreme events are described in detail in section 3.4.1.1. The episodic nature of cyclonic storms causes an inter–annual variation of the intensity of sand movement during winter.

3.2 Wind speed differences

Mean values of speed ratio S_z are shown for the directions of the main sand transporting winds (SE to N) in figure 3.4. The results are based on all datasets where $u_{0.24} \geq 4$ m s^{-1} at site D and corresponding u_z (z: 0.24 m or 1.77 m) at sites A, B and C was above 0 m s^{-1}. For directions other than these directions, the number of datasets was too low to calculate mean values.

3.2.1 Interdune corridor, bare surface (site A)

Because the surface conditions at site A were similar to those at site D, values of S_z show the influence of the neighbouring dune ridges on wind speed in an interdune corridor. Maxima for S_z at both heights were determined for wind

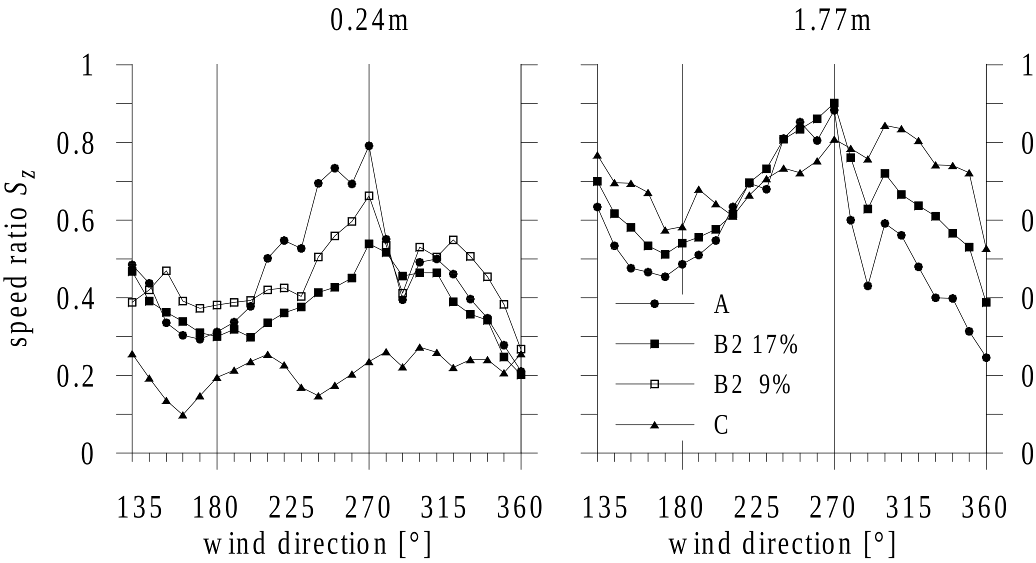

Figure 3.4: Relation of speed ratio S_z and wind direction for $u_{0.24}$ and $u_{1.77}$.

directions parallel to dune orientation (270°). The value of $S_{0.24}$ is 0.79, which means that under ideal conditions near surface wind speed in the interdune corridors is approximately 80 per cent of dune crest values. Higher above the surface, the difference is lower, as $S_{1.77}$ is 0.88. Towards southerly directions a steep decrease was measured between 240° and 200°, values of $S_{0.24}$ reached a minimum of 0.29 at 170° and increased again to 0.49 for winds from SE. A similar trend with an equal minimum was determined for $S_{1.77}$, but the decrease is less sharp and absolute values remain above $S_{0.24}$. The decrease of S_z towards northerly directions is sharp towards a local low at 290° ($S_{0.24} = 0.39$, $S_{1.77} = 0.43$). A local maximum at 310° ($S_{0.24} = 0.5$) is followed by a steep decline towards the minimum for northern winds, which is 0.21 for $S_{0.24}$ at 0°/360°. As for southerly winds, values of $S_{1.77}$ are always above those of $S_{0.24}$.

3.2.2 Interdune corridor, vegetation (site B)

Site B differs from site D in two ways: morphologic position and vegetation cover. Thus the results of S_z at site B show the combined influence of morphologic position and vegetation cover. Differences between site B and site A are vegetation cover and width of the corrdior.

As the mean vegetation height at the site was 0.37 m, $u_{0.24}$ was measured within the vegetation canopy. Values of $S_{0.24}$ were determined for two vegetation densities, 17 per cent and 9 per cent. The changes of vegetation cover led to differences of $S_{0.24}$. Values for 17 per cent cover remain below those of site A for all directions except between 150° and 170°. For 9 per cent cover $S_{0.24}$ is generally higher than for 17 per cent, values are above those of site A for most northerly winds and between 150° to 200°. Values of $S_{1.77}$ for winds between 210° and 270° are almost identical with corresponding values at site A, which indicates that vegetation cover does not influence airflow at that height. Higher values than at site A were determined for wind directions between 130° and 200° and all winds with a northern component. The overall pattern of velocity differences at site B for both heights is similar to that determined at site A, including the local minimum for WNW-winds.

3.2.3 Dune crest, partly stabilized (site C)

At this site, $u_{0.24}$ was measured within the zone where the small scale morphology – hummocks of approximately 1 m height – influences airflow, resulting in a maximum value of 0.27 at 310° and a minimum of 0.09 at 160°. The distribution of peaks and lows differs from the interdune corridor sites and does not appear to be directly connected to dune ridge orientation. The values of $S_{1.77}$, based on velocity measurements taken above the roughness elements, show a pattern similar to the interdune sites A and B, yet absolute values are generally higher than at these sites. An exception are winds between 210° and 270°, were values at site C are lower. Contrasting the situation at sites A and B, $S_{1.77}$ for winds out of the NW-sector was higher than for southerly winds with the same angle towards general dune ridge orientation. Maximum was determined for 300° (0.84). Values remain above 0.7 between 270° and 350°. Lows of $S_{1.77}$ are located at 170° (0.57) and 360° (0.52), ie wind directions normal to dune orientation.

The results of the speed ratio calculations show that wind speed in areas lower than the highest dune ridges is dependent on wind direction relative to dune orientation. The influence is higher for airflow close to the surface. A comparison of dune parallel winds (270°) between sites A and D reveals that the height difference between dune crest and interdune corridor is responsible for a 21 per cent reduction of $u_{0.24}$ in the corridor. Natural vegetation cover of 17 per cent at site B2 leads to a reduction of $u_{0.24}$ by 46 per cent for the same winds. A reduction of the vegetation cover at site B2 to 9 per cent causes a reduction of only 34 per cent. According to the results, dune parallel winds must reach the minimum of 1.3 $u_{t0.24}$ at the dune crest in order to cause sand movement over non-vegetated surfaces in the interdune corridors. Over vegetated surfaces 1.8 $u_{t0.24}$ is required. For wind directions other than between 210° and 280°, $u_{0.24}$ is reduced to less than 50 per cent over vegetation free surfaces in interdune corridors.

3.3 Wind direction differences

Figures 3.5, 3.6 and 3.7 show the deviation of wind direction (δ_W) at sites A, B2 and C from the value determined at site D.

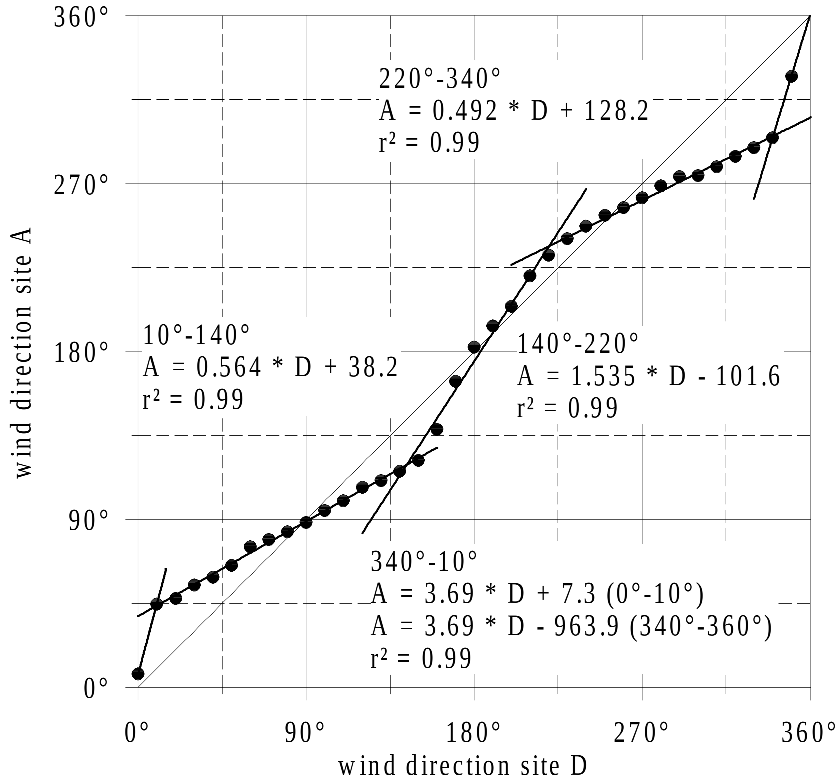

Figure 3.5: Wind direction differences (δ_W) site D - site A.

3.3.1 Interdune corridor, bare surface (site A)

Four sectors with a distinct deflection pattern were recognized for this site: 10°-140°, 140°-230°, 230°-340°, 340°-10° (figure 3.5). No deflection was recorded for wind directions normal (0°, 180°) and parallel (90°, 260°) to the dune orientation. The sectors of winds from dune-parallel directions are wider (120° and 130°) than those of winds normal to dune orientation (30° and 80°). In general δ_W for winds with a northern component is higher than for winds with a southern component. Highest values of δ_W were determined for winds (at site D) from 340° (δ_W -44°), 10° (δ_W +33°), 150° (δ_W -27°) and 220° (δ_W +16°). Winds from SW were least deflected. All deflections were directed towards the ridges.

3.3.2 Interdune corridor, vegetation (site B)

The pattern of deflection is similar to that of site A, but in general δ_W is smaller (figure 3.6, p.49). An exception are winds from SW, where δ_W at site B is higher. The sectors are identical to those determined for site A. No or minimal alteration of wind direction was calculated for similar winds as for site A (0°, 90°, 170°, 270°). The difference between winds with northern or southern components is less marked than at site A, but still present. Maximum

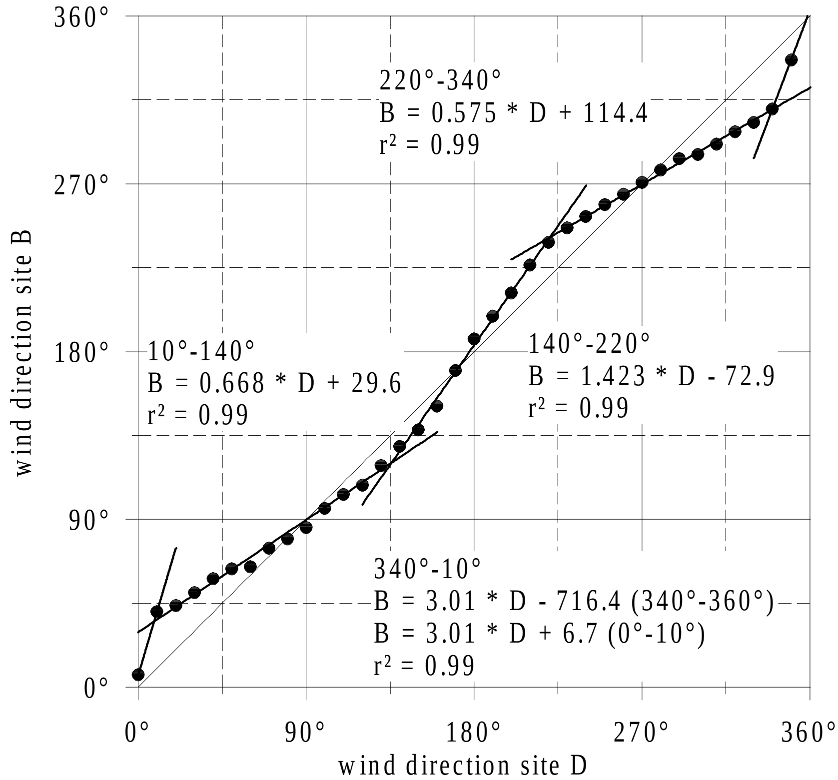

Figure 3.6: Wind direction differences (δ_W) site D - site B.

δ_W was determined for winds (at site D) from 340° (δ_W -30°), 10° (δ_W +26°), 220° (δ_W +20°) and 130° (δ_W -13).

3.3.3 Dune crest, partly stabilized (site C)

The overall pattern of deflection was similar to that of sites A and B, but the values of δ_W at site C were significantly lower than those determined for the interdune corridor sites (figure 3.7, p.50). Especially for winds with a southern component, deflection is only minimal. Highest values of δ_W were determined for NE-winds. No deflection was determined for directions normal (0°, 180°) and parallel (90°, 270°) to dune orientation. Linear regressions have been found for all directions except NW to N, where no trend was recognized.

The results of measured direction differences show that strong linear relationships exist between wind direction at the reference site D and wind direction at the interdune sites A and B. At the low dune ridge of site C only small deviations were determined. In general, flow is deflected towards a more dune-parallel direction. Minimal δ_W was detected for winds which are either parallel or perpendicular to the orientation of the dune ridges. Highest values were observed for NNW-winds at sites A and B. The amount of deflection decreases as corridor width increases.

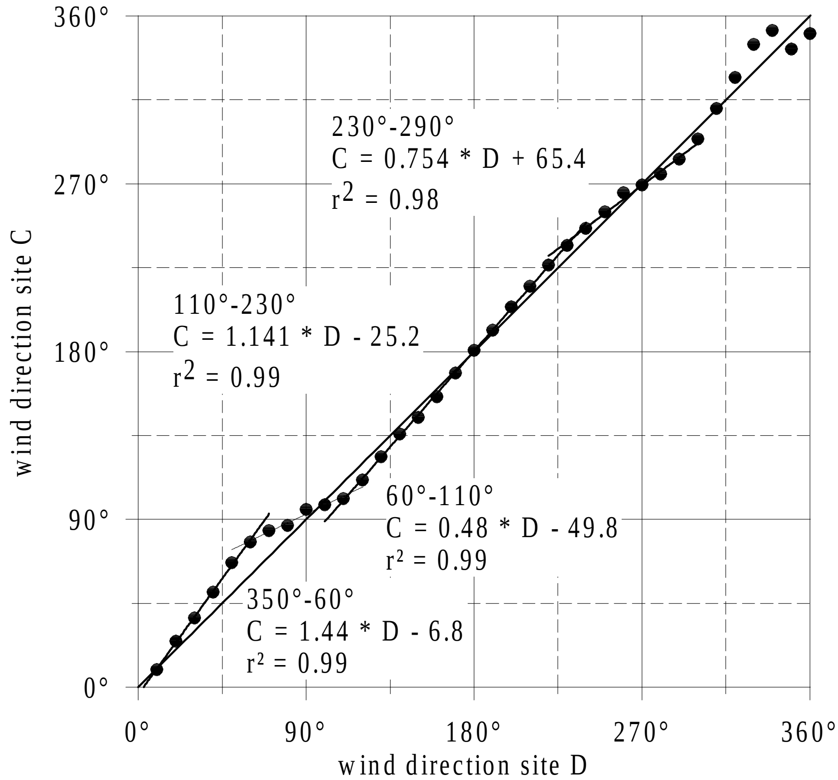

Figure 3.7: Wind direction differences (δ_W) site D - site C.

3.4 Influences on sand transport

In the following sections, the influence of dune morphology, rainfall, vegetation and surface crusting on sand transport will be exemplified using data from several representative events. The sand transport data of 14 events is shown in figures 3.8, 3.9, 3.10 and 3.11. Out of these, 10 are winter storms of which 5 were accompanied by rainfall (13/12/97, 5/1, 16/3, 17/3 and 19/3/98). The four summer events were recorded in April (21/4/98, 19/4, 21/4 and 24/4/99). Wind speed and direction data of the events are shown in the specific subsections.

3.4.1 Dune morphology

As has been shown above, wind speed and wind direction are influenced by the upwind morphology. For wind directions oblique to dune ridge orientation, the speed ratio $S_{0.24}$ between the dune crest and the interdune corridor is below 0.8 (figure 3.4, p.46). Oblique winds are deflected towards more dune parallel directions. The influence of morphology on sand transport can be shown by direct comparison of flux rates and transported mass determined at sites A and D during an event. The surfaces of both sites are similar: they are free of mycrophytic crust and vascular vegetation, the latter being concentrated in

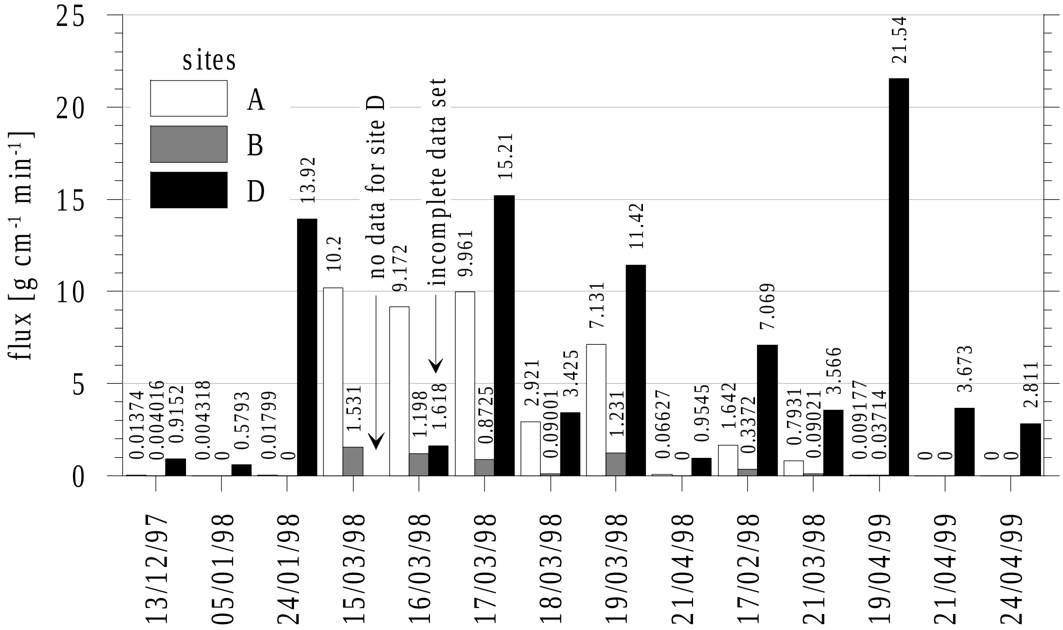

Figure 3.8: Mean sand flux at sites A, B and D for selected events.

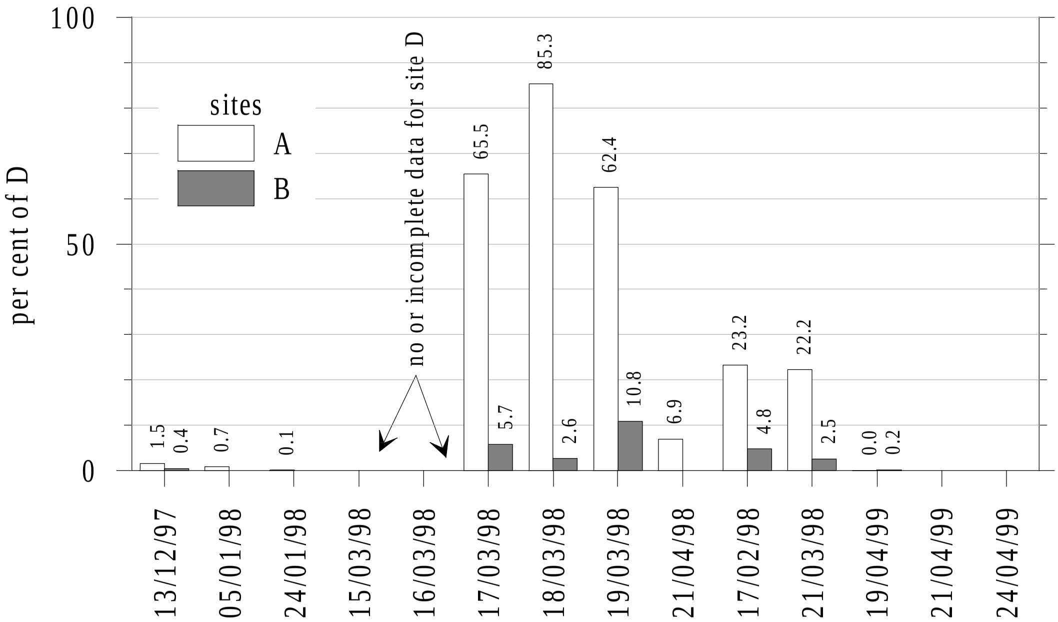

Figure 3.9: Mean sand flux at sites A and B in per cent of mean flux at site D.

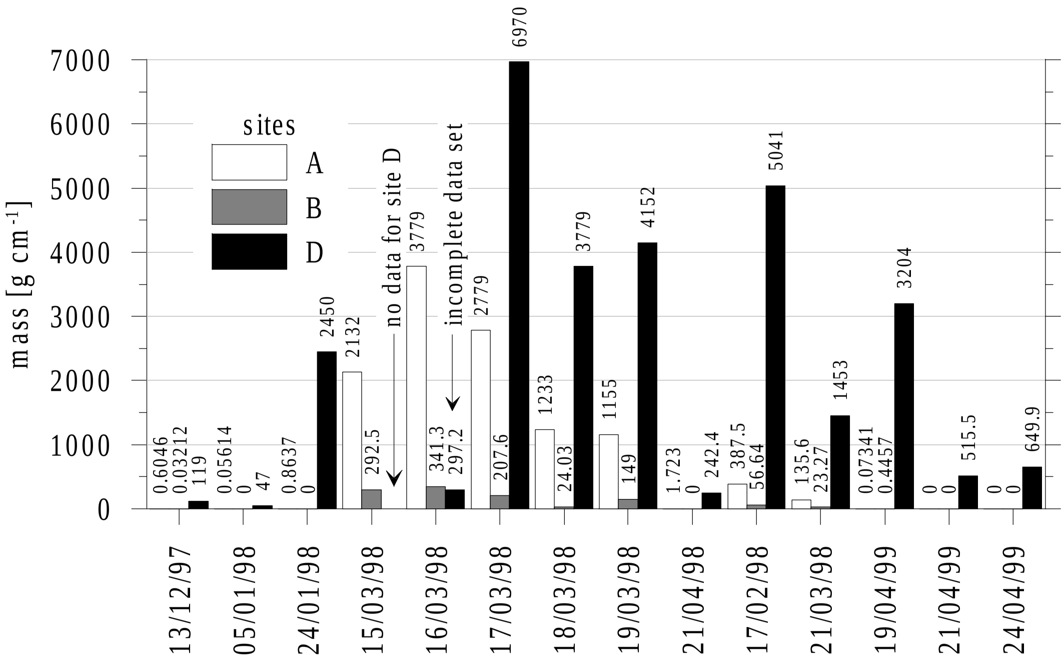

Figure 3.10: Transported mass at sites A, B2 and D for selected events.

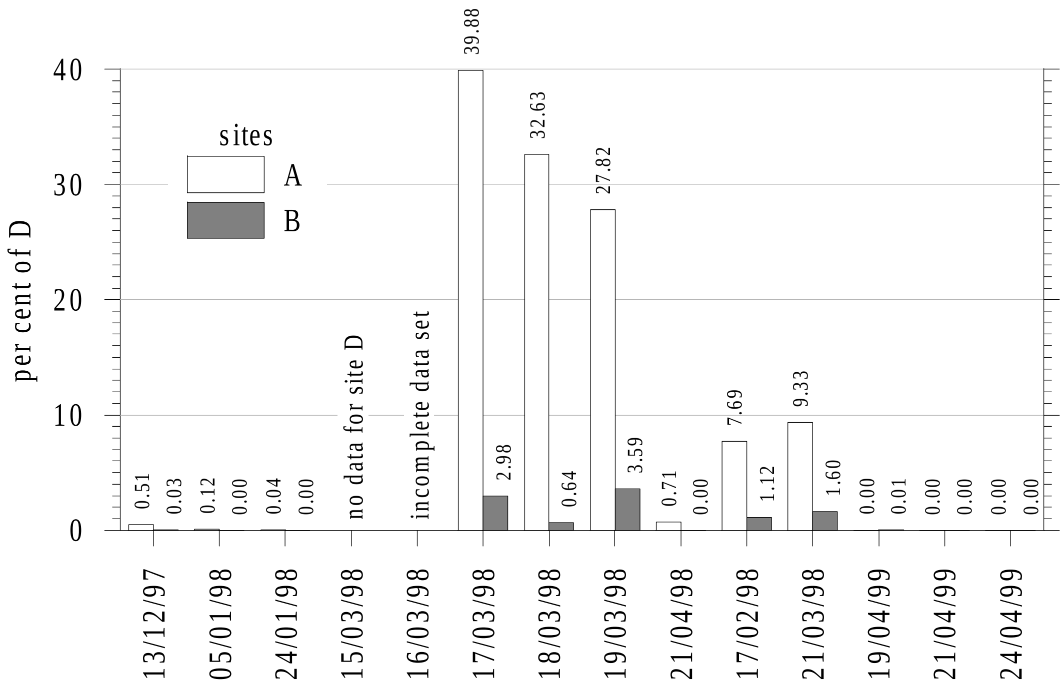

Figure 3.11: Transported mass at sites A and B2 in relation to site D.

clumps at site D, which leaves large areas open to undisturbed airflow. The sites differ in their location: site D is exposed to winds from all directions, while at site A all but dune parallel winds are modified by upwind obstacles.

Sand movement in the interdune corridor was only detected during winter storms and occurred only when sand transport at the dune crest had been recorded, too. Saltation was recorded at sites A and B2, where the microphytic surface crust had been destroyed and removed. Its frequency was found to depend mainly on the direction of the undisturbed airflow above dune level. No sand movement was recorded at these sites during winds attributed to the diurnal summer regime. The total amount of material transported at sites A and B2 was calculated, based on *saltiphone* measurements and trapped material for the periods between November 1997 and May 1999 shown in table 2.1, p.22. The results in $g^{-1}cm^{-1}$ and direction intervals of 10° are shown in figure 3.12, p.54. The line represents the mean value for all traps at the site (locations see figures 2.3, p.25 and 2.5, p.26), while the bars represent the range between maximum and minimum value. The dominating transport direction is WSW, which is slightly oblique to dune ridge orientation. A secondary peak has been determined for winds from SSE. Both directions are associated with winter storms. The main transport occurred during winter 1997/98, when all of the high magnitude storms ($u_{0.24} \geq 10$ m s^{-1}) of the research period were recorded. Eight events causing more than 100 g cm^{-1} of sand transport at site A were recorded. Of these, four exceeded 1000 g cm^{-1}. During the following winter 1998/99, recorded storms were of lower magnitude with no exceptional events such as 15/3 to 19/3/98 (see 3.4.1.2). Only six events with more than 100 g cm^{-1} were measured, the maximum of sand tranported being 625 g cm^{-1} during one event. Total transport in this winter was 2319 g cm^{-1}, which is 18.1 per cent of the amount in the previous winter. The relation of e for these two periods is similar, values for 1998/99 amount to only 18.1 per cent of 1997/98 (figure 3.3, p.44). Thus the possibility that the lower transport of the second winter is a result of diminished sand supply within the disturbed area of site A can be ruled out. Resultant wind direction was similar for both years, 252° in 1997/98 and 248° in 1998/99. The result also shows that e is a valid tool to characterize the sand transport capability of a known wind regime.

Site C was not equipped with a saltation sensor, yet the results of wind speed measurements and sand trap data allows a comparison with data of the other sites. Thus an estimation of the magnitude of sand transport was possible. The amount of transportable material at the site was low. Open and level surfaces are restricted to the areas between hummocks covered by vegetation and surface crust. The orientation of ripple marks, used as indicators for sand transport, showed that transport directions were restricted to the orientation of the axis of the hollows between the hummocks. Near surface wind speed $u_{0.24}$ in these hollows was dependent on wind direction above the roughness elements. As the anemometer array was situated in a narrow gap which was oriented approximately 0°/180°, highest values of $u_{0.24}$ were mea-

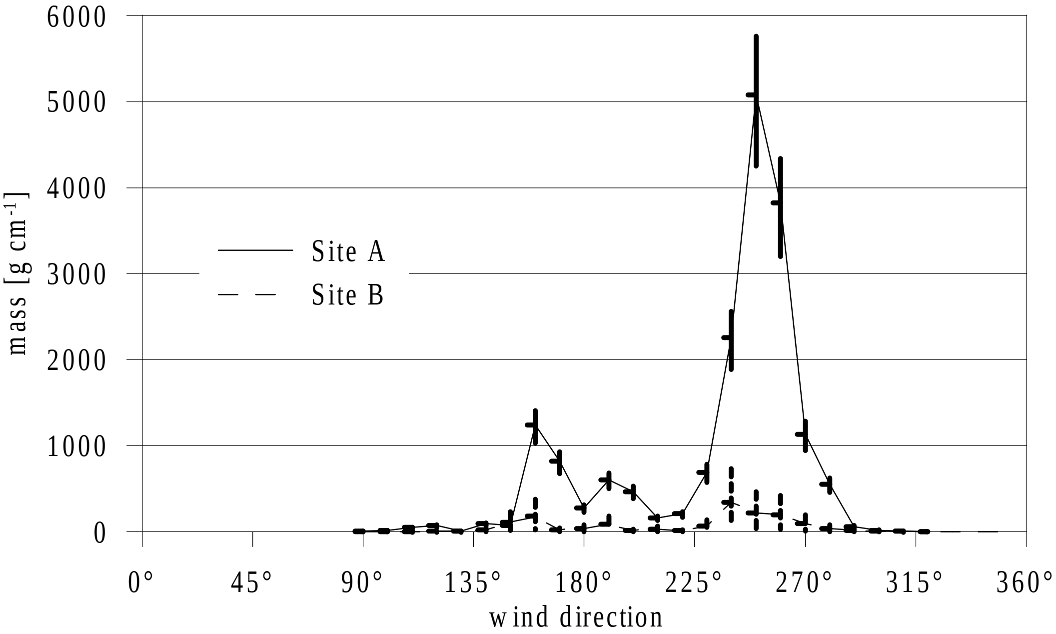

Figure 3.12: Transported mass at the interdune corridor sites A and B2 between 20/11/97 and 25/4/99.

sured during southerly winds in the initial phase of cyclonic winter storms, velocities being similar to site A. In all cases of cyclonic storms, $u_{0.24}$ at site C decreased below values measured at the interdune corridor sites A and B after wind direction had changed to SW or W. During the steady northerly winds of summer, $u_{0.24}$ at site C was similar to the values determined at site A. On several occasions, however, it remained below instrument threshold, despite values of up to 3 m s^{-1} at the interdune sites. Major modifications of small scale morphology were only observed during the exceptional storm in March 1998, during which several hummocks were destroyed by wind action. During average winter storms, erosion has been observed at the bases of hummocks, causing subsequent avalanching of the weakly cemented sand into the gap, adding transportable material to the open surfaces.

Sand transport at site D was observed and recorded during all seasons. During summer, when no wind speed and saltation measurements were taken, changes detected through the erosion pin measurements served as indicators for sand transport. The active areas at the dune crest experienced movement from all directions. The expected predominance of transport from S to SW during winter and NW during summer – based on the calculation of effective wind e (figures 3.1, 3.2, pp.42, 43) – could be verified through *saltiphone* recordings, measurement of trapped mass and erosion pin data. Other than at the interdune corridor sites, the data on sand flux for site D is not complete for several storm events. The highly mobile sand at the site covered the *saltiphone* on several occasions, preventing recordings. Sand traps were found filled completely after an event, making it impossible to calculate flux, as the time when the trap was full is unknown. Traps have been observed to fill up during high magnitude events within minutes. In addition, traps were sometimes buried or excavated by shifting sand. Flux data for site D is therefore available only for a limited number of events. The limitations of the sand traps and the *saltiphone* in the active environment of the dune crest prevented the determination of the total amount of sand moved at site D during the study period. Transported mass was estimated based on erosion pin measurements used to record morphologic changes. The results of these measurements are shown in chapter 3.5.

The sand trap installed upwind of site D at the gap of the border road yielded transport amounts similar to those determined at the undisturbed corridor site B1. No significant input of material from an upwind source can thus be expected.

3.4.1.1 Winter winds

Two of the selected events represent 'ideal' cyclonic storms as described on page 45. These were recorded on 17/2/99 (see p.59) and 21/3/99 (see p.56). The magnitude of the storms differs considerably. Higher wind speeds than on 17/2/99 were only recorded during the event in mid-March 1998, while moderate wind speeds were measured on 21/3/99. This is reflected in the average flux rate, which was 7.0 g cm^{-1} min^{-1} on 17/2/99 and thus twice the value of 21/3/99. Nonetheless, the relation between the dune crest and the interdune corridor is similar, the flux reaching 23.2 per cent, resp. 22.2 per cent at site A. The transported mass at the dune crest was 1400 g cm^{-1} during the average event and 5000 g cm^{-1} during the high magnitude storm. At site A, 9.3 per cent, resp. 7.7 per cent of these values were measured.

Winter storms which do not fit into the scheme of an 'ideal' cyclone led to different results. During the exceptional storm in mid-March 1998 (see p.62) flux rates up to 15.2 g cm^{-1} min^{-1} at the dune crest and 2.9 g cm^{-1} min^{-1} to 10.2 g cm^{-1} min^{-1} in the interdune corridor were determined. The rates at the dune crest were most likely even higher, as data is missing for 15/3/98 when the highest wind speeds of the research period were measured. The flux between the crest and the interdune corridor differed less than during 'ideal' events, reaching values between 62 and 85 per cent at site A. The transported mass at site D is estimated to have exceeded 7000 g cm^{-1} during the first phase of the event. This estimation is based on the measured value of 17/3/98. Wind speed during the first two days of the storm was even higher than on 17/3, but *saltiphone* and sand trap data for these days are not available as the installations were damaged by the storm. For site A the transported mass ranged between 27 and 40 per cent of site D values.

Very high flux rates (13.9 g cm^{-1} min^{-1}) were determined at site D during the event of 24/1/98 (see p.70). The transported mass was 2450 g cm^{-1} and thus above the values for 21/3/99. Flux rates at site A reached only 0.1 per cent of this value, the transported mass was negligible (0.04 g cm^{-1}). The wind direction during maximum wind speed of 10 m s^{-1} was perpendicular to dune orientation.

Diurnal southerly winds during periods of stable atmospheric conditions in winter (see p.73) caused minor sand transport at the dune crest. Maximum wind speeds at site D remained mostly below 6 m s^{-1} and did not exceed 8 m s^{-1}. As maxima were reached during southerly directions, where $S_{0.24}$ is below 0.5, no sand movement occured during these events in the interdune corridors.

3.4.1.2 Examples

Cyclonic storm 21/3/99 This event is an example of a cyclonic storm of average magnitude. The event was the last winter storm of the 1998/99 season and occurred after a three-week period dominated by diurnal NW-winds typical for the summer wind regime. Figure 3.13, p.57 shows the passage of a low pressure cell. Increasing southerly winds at the beginning of the storm shortly after 02:00, and a change to SW as the warm front passes between 09:00 and 11:00. Winds from SW to W in the warm sector from 11:00 to 17:00 and finally the quick change to N after 17:00, at the passage of the cold front. Saltation has been measured at sites A, B2 and D. Trapped sand at site C strongly suggests sand movement at that site, despite the low recorded values of $u_{0.24}$. The instrument at site C was probably faulty, as the figures for $u_{0.65}$ and $u_{1.77}$ at the site are within the normal expected range for such an event.

Windspeed before the storm was low at the dune crest and close to instrument threshold in the interdune corridor. The wind direction from SE is typical for nighttime conditions at Sde Hallamish. As the direction changed to S, windspeed increased at all sites. Sand movement at site D starts as $u_{0.24}$ passes 4 m s^{-1}. At that time windspeed in the corridor is just above threshold of the anemometers at site A, while at site B $u_{0.24}$ reached a value of 1.5 m s^{-1}. Sand flux at the dune crest remains low until $u_{0.24}$ reached 6 m s^{-1} at 07:30. At 08:00 the highest continuous sand flux values were measured. The gustiness of the wind increased after 10:00, following the change of wind direction to SW connected to the passage of the warm front. At the same time average windspeed and sand flux decreased. A low of about 4.5 m s^{-1} was reached at 14:00, followed by a rise to 7 m s^{-1} by 18:30. After this peak, the passage of the cold front led to a significant drop of wind speed, causing a short break in saltation. Afterwards wind speed rose again and sand movement continued, now caused by NW-winds. Sand flux q was now higher than during equivalent SW-winds. Sand movement ended as $u_{0.24}$ dropped below 4 m s^{-1}.

Sand movement in the interdune corridor started after 10:00 as $u_{0.24}$ reached 4 m s^{-1} at sites A and B. The change of wind direction from normal to almost parallel to dune ridge orientation was accompanied by an increase of $u_{0.24}$ at sites A and B. As can be seen in figure 3.14, p.58, this increase of speed coincided with an almost linear increase of speed ratio S_z at both heights. During the first phase of the storm, S_z for sites A and B rose as wind direction

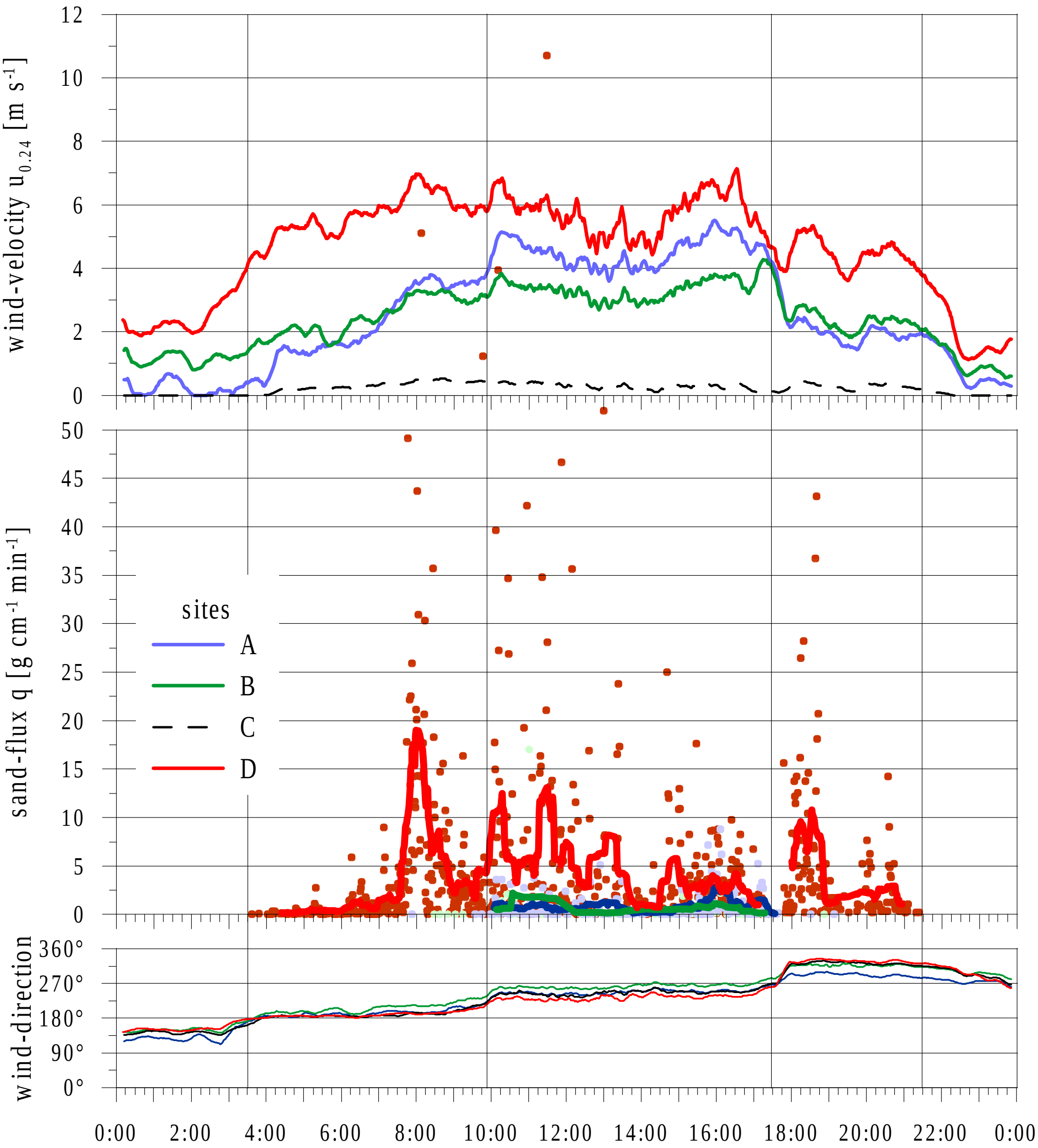

Figure 3.13: Wind conditions and sand transport during 21/3/99.

changed continuously from S to SSW. Here, the increase was sharper for site A than for site B. Both wind speed $u_{0.24}$ and speed ratio $S_{0.24}$ at site A were lower than at site B, except during the period when wind direction was parallel to the dune ridges. The graphs of $S_{1.77}$ for the two sites show a similar pattern. Values at site B were higher for all winds approaching the dunes at high angles, while they were identical for dune parallel winds. This effect is observed during most winter storms and is also reflected in the general pattern of *speed ratio* values as shown earlier (figure 3.4, p.46). It is interpreted to be a result of the greater width of the corridor in combination with the presence of vegetation at site B.

At site C, the minimum of $S_{1.77}$ was determined during the short period

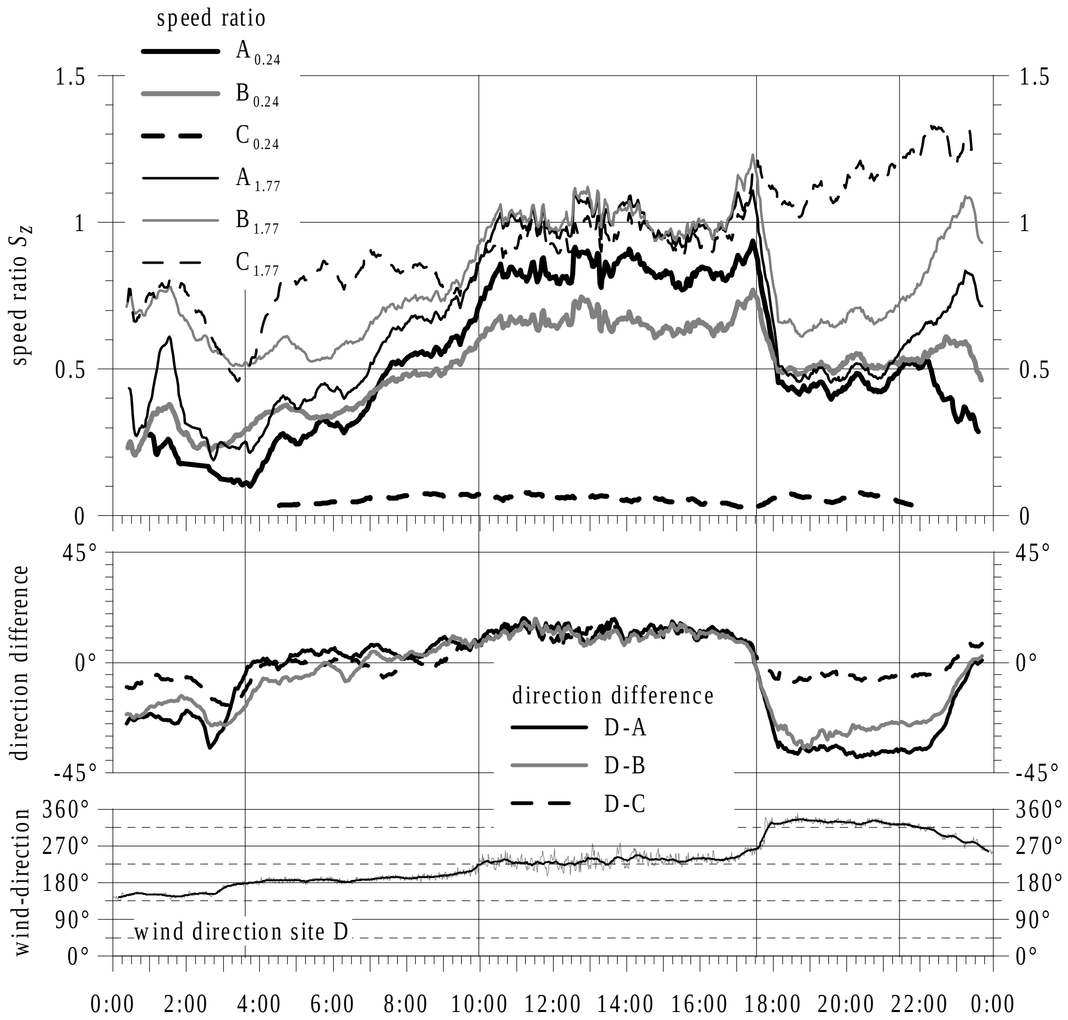

Figure 3.14: Speed and direction differences 21/3/99.

when wind direction was exactly normal to dune orientation. Values for $S_{0.24}$ were not determined at site C due to faulty velocity readings. Within the warm sector, as wind direction was SW, $S_{1.77}$ was near 1 at all sites. During this period, $S_{0.24}$ was above 0.8 at site A and above 0.6 at B. The highest values of S_z for all heights of measurement were measured during the passage of the cold front, as wind direction was parallel to the corridor when the change from SW to NW occurred. After this change a very sharp drop of S_z was observed at all sites except site C.

Only minor deflection was observed in the interdune corridor as long as wind direction was normal or parallel to the dune ridge, whereas deflection towards a dune parallel direction was recorded when wind direction was oblique to ridge orientation. The reversal of deflection can be seen very clearly as the wind changed from SW to NW in the early evening. At site C, the trend was similar to that at the corridor sites, but as expected, absolute values of δ_W were considerably smaller during oblique winds.

High magnitude cyclonic storm 17/2/1999 This storm serves as an example of a high magnitude event. It lasted 12 hours and contained all typical elements of a cyclonic storm. During the storm, wind speeds ($u_{0.24}$) up to 11.9 m s^{-1} were recorded at site D. Higher values have only been measured during the storm in mid-March 1998 (see p.62). Saltation thresholds were surpassed at all field sites.

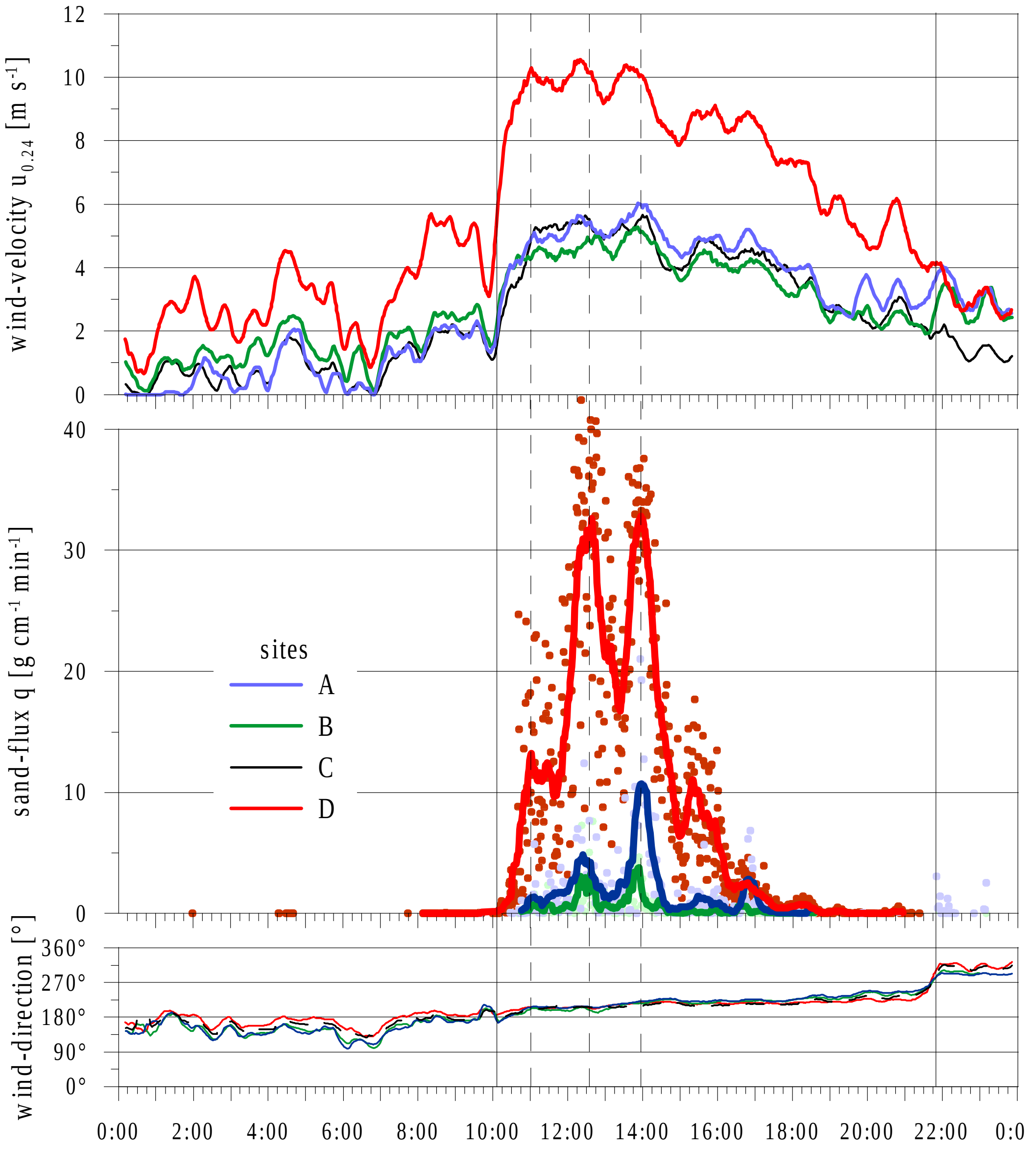

Figure 3.15: Wind conditions and sand transport during 17/2/99.

Until 07:00 low and unsteady winds were recorded at all sites. They were well below u_t at sites A, B and C, while single isolated impacts were recorded at site D as $u_{0.24}$ reached 4 m s^{-1}. Wind directions were S to SE and E, typical for nighttime and morning conditions. After 07:00 wind speed began to rise and wind direction steadied on S. At site D $u_{0.24}$ reached 6 m s^{-1} shortly

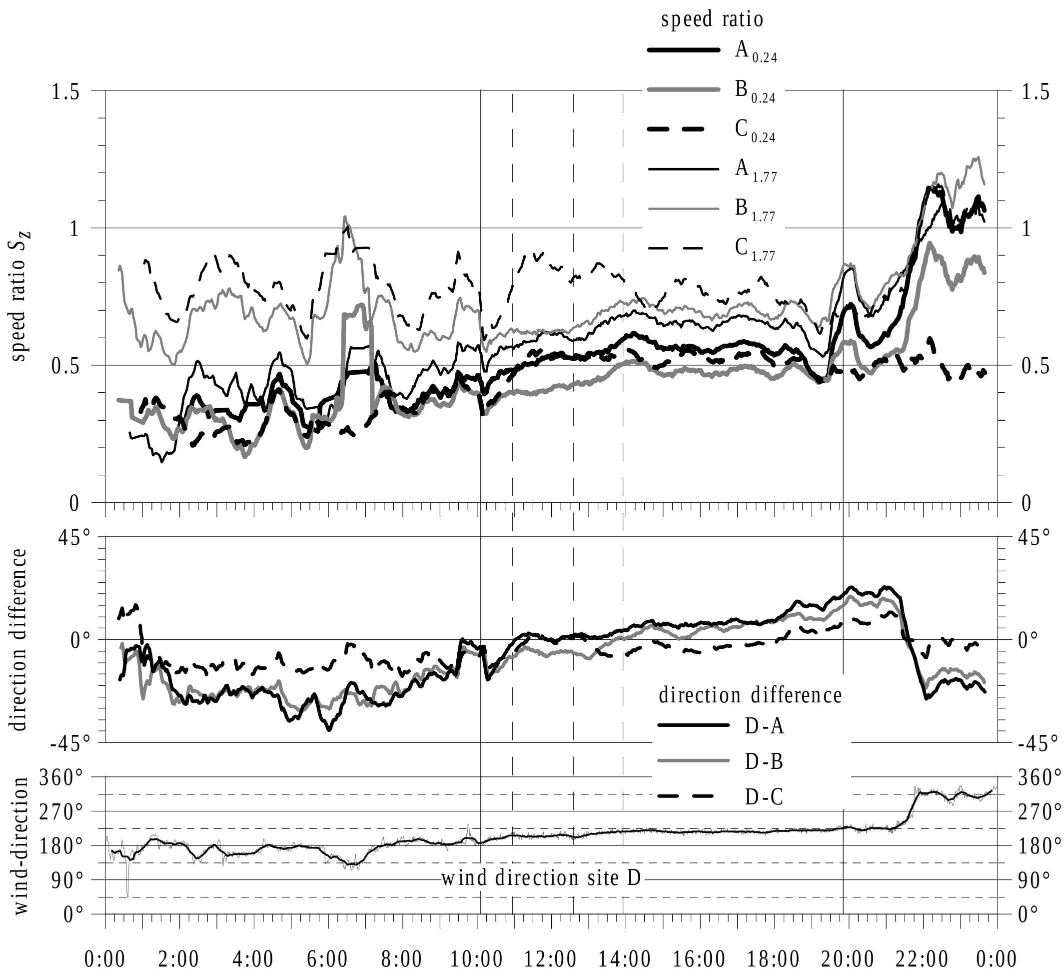

Figure 3.16: Speed and direction differences during the storm on 17/2/99.

after 08:00, sand flux remaining marginal. The onset of the storm at about 10:00 was marked by a short drop of wind speed below 4 m s^{-1} which was followed by a sharp rise towards a first peak at site D at 11:00, where mean $u_{0.24}$ reached 10 m s^{-1}. At the same time saltation threshold u_t was surpassed at the interdune corridor sites. Two more peaks of wind speed occurred at 12:30 and 14:00 with $u_{0.24}$ reaching 10.5 m s^{-1}. After the second peak the wind speed fell gradually. Wind direction changed from S to SW between 10:00 and 14:00. Sand movement ceased after 18:30 in the interdune corridor when $u_{0.24}$ at site A fell below 4 m s^{-1}. Saltation was recorded at site D until 22:00. Afterwards the end of the storm is marked by a significant, sudden change of wind direction to NW and a drop of $u_{0.24}$ at site D below 4 m s^{-1} .

Unlike during storm events of average magnitude (see p.56), sand transport in the interdune already started in the initial phase of the storm, when wind direction was still normal to dune orientation. Maximum wind speeds at the interdune corridor sites A and B were measured at 14:00, during the third peak, reaching 6 m s^{-1} (A) and 5.5 m s^{-1} (B). As in the previous example, the change of wind direction during the main phase of the storm was accompanied by an increase of the speed ratio S_z for the interdune sites, thus indicating the influence of the upwind obstacles. When saltation started, $S_{0.24}$ at site A was below 0.5 while shortly after the third peak it had increased to above 0.6. The development of $S_{0.24}$ at site B was parallel to site A, but the values were generally lower (0.3 resp. 0.5). Values of $S_{1.77}$ are higher at site B, increasing to above 0.7 after the third peak, again a result of the greater width of the corridor at the site compared to site A. At site C values of $S_{0.24}$ remained almost constant at 0.5 after the first peak until after the end of the storm. In contrast to the interdune sites, $S_{1.77}$ at site C reached its maximum of 0.9 shortly after the first peak of the storm. Values declined afterwards, but remained above interdune corridor numbers until the end of the storm. At all sites $S_{0.24}$ was lower than $S_{1.77}$. During the night following the storm, after the direction change to northerly winds, $S_{1.77}$ was close to 1 at all sites, indicating an undisturbed flow over the dune ridges (figure 3.16, p.60).

Table 3.1: Sand flux and wind direction during 17/2/99. Values for the peaks are maxima during one recording interval of two minutes. Mean values were determined for the saltation period of each site.

	flux q					direction		
	g cm^{-1} min^{-1}			per cent of D		degrees		
peak	A	B	D	A	B	A	B	D
I	5.7	1.5	24.1	23.7	6.2	214	209	209
II	12.4	7.2	42.4	29.2	17.1	203	198	203
III	20.9	7.4	37.6	55.6	19.7	219	214	219
mean	2.0	0.4	8.0	23.2	4.8			

The sand transport rate at site D was closely correlated to $u_{0.24}$ (figure 3.15, p.59). Wind speed remained high (mean $u_{0.24}$ 10 m s^{-1}) from 11:00 - 14:00. Afterwards a stepped decrease was recorded. The highest values of wind speed $u_{0.24}$ and sand flux q at site D were measured during the second peak. Figures at the third peak were lower. In contrast to this, $u_{0.24}$ and q in the interdune corridor sites increased during the storm and reached their maximum at the third peak (table 3.1). The increase was both absolute and relative. Between the second and the third peak, wind direction had changed 16 degrees clockwise, thus lowering the angle of attack between airflow and dune orientation. Direction differences at all sites developed according to the general trend shown in figures 3.5, p.48, 3.6, p.49 and 3.7, p.50 where minima were determined for winds perpendicular and parallel to dune orientation.

Very high magnitude cyclonic storm 15/3-19/3/98 A stormy period began in the afternoon of 15/3/98 and lasted until the 19th. As it was preceeded and followed by days of very low wind speeds, it is treated as one storm which can be divided into three distinct phases. During the storm, the highest wind speeds of the study period were recorded. The recurrence interval of storms of similar magnitude is >10 years (Tsoar, pers. comm.). Sand movement occurred at all locations. The storm, whose initial phase lasted from the early afternoon of the 15/3 until the evening of 16/3 without intermission, showed the limitations of the measuring equipment. At site D, the capacity of the *Leatherman* collectors was exhausted within less than 15 minutes during the first period of extremely high wind speeds on the 15/3. The *saltiphone* at the site was buried during the first night, while the mast holding the anemometers was tilted due to deflation around one of the anchoring poles.

First phase, 15-16/3/98 (figures 3.17, p.63 and 3.18, p.64) A period of very low wind speed preceeded the onset of the storm. The storm started shortly after 16:30 with a sudden rise of wind speed, wind direction was SE and changed soon after towards S. Intensive sand movement was recorded at site D, while in the interdune corridor saltation was detected only at site A as $u_{0.24}$ remained mainly below 4 m s^{-}1 until 18:45. As wind direction at that time was perpendicular to dune orientation, $S_{0.24}$ was below 0.4 at sites A and B during the intitial storm phase. Therefore $u_{0.24}$ in the interdune corridor passed 4 m s^{-}1, necessary for continuous sand movement, only after wind speed at the dune crest had reached values above 12 m s^{-1}. Wind speed at site D remained on a high level between 20:00 and 22:00. Values of S_z increased while wind direction gradually changed clockwise from SE to S. Between 22:00 and 02:00 wind speed gradually receded at a rate similar to the rise on the previous evening. At site D $u_{0.24}$ dropped to 6 m s^{-}1 and to 2.3 m s^{-}1 at sites A and B. Saltation stopped at the interdune sites after $u_{0.24}$ fell below 4 m s^{-}1. As in the previous examples, differences of corridor width are reflected in the differences of values of S_z. The values of $S_{1.77}$ are higher at site B, those of $S_{0.24}$ are almost identical at both sites despite the vegetation cover at site B.

Saltation data and thus calculated flux q at site D is unreliable during the night, as the *saltiphone* recording the data was covered by shifting sands. Therefore the sand flux values until 02:00 must be considered as minimum values. In addition, the maximum recording rate of the instrument was exceeded on several occasions prior to burial. As all sand traps were filled up during the night, no exact data on transported mass is available for site D after the initial 20 minutes of the storm in the evening of the 15th until 11:00 of the 16th, when the traps were emptied and the *saltiphone* was reset. However, as $u_{0.24}$ did not drop below 6 m s^{-1} at the site, continuous sand movement must be assumed.

The passage of the warm front between 02:00 and 03:00 of the 16th led to a rapid direction change to WSW, that is dune parallel winds. Wind direction was steady afterwards until 13:30. The direction change was accompanied by

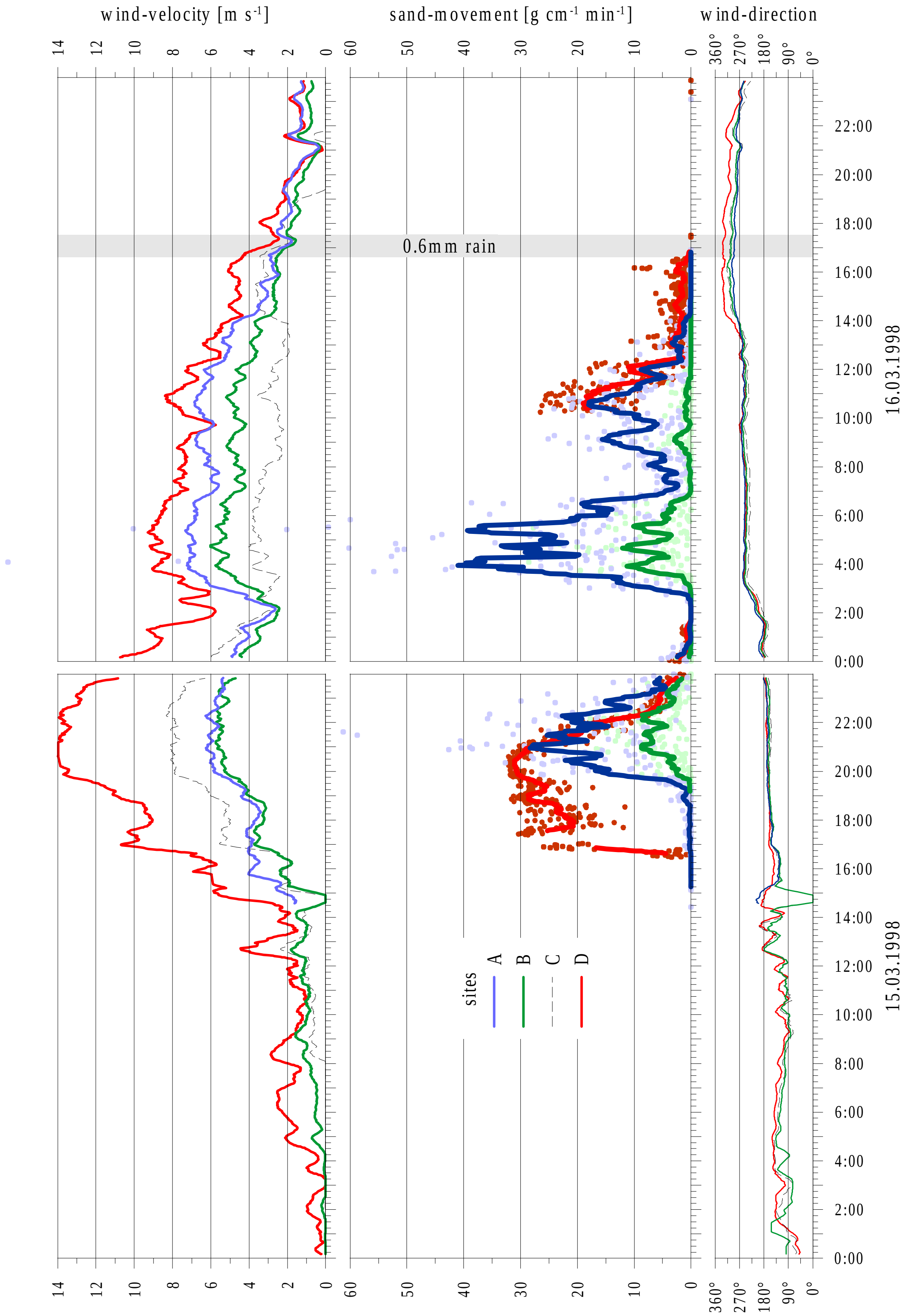

Figure 3.17: Wind conditions and sand transport 15/3 and 16/3/98.

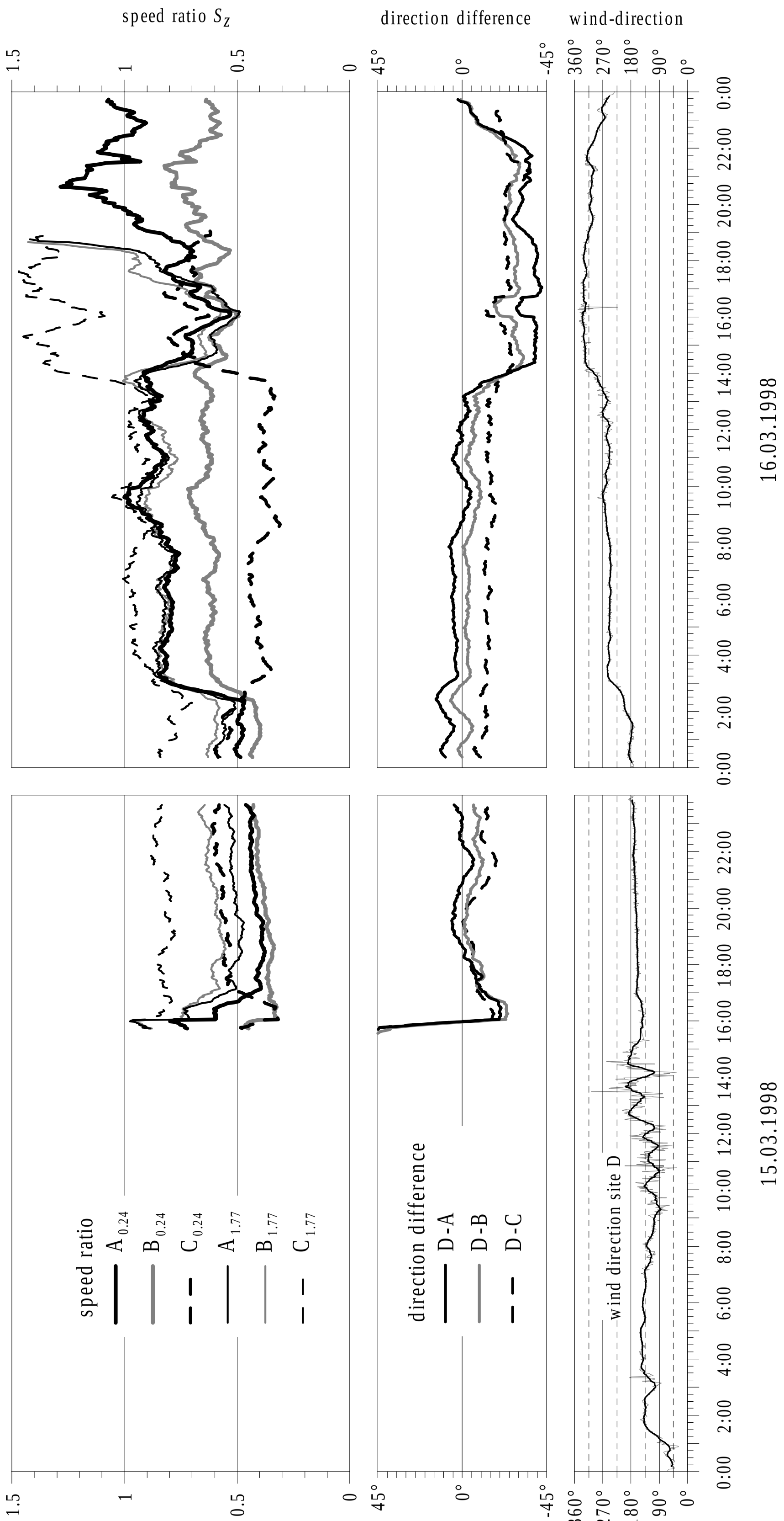

Figure 3.18: Speed and direction differences 15/3 and 16/3/98.

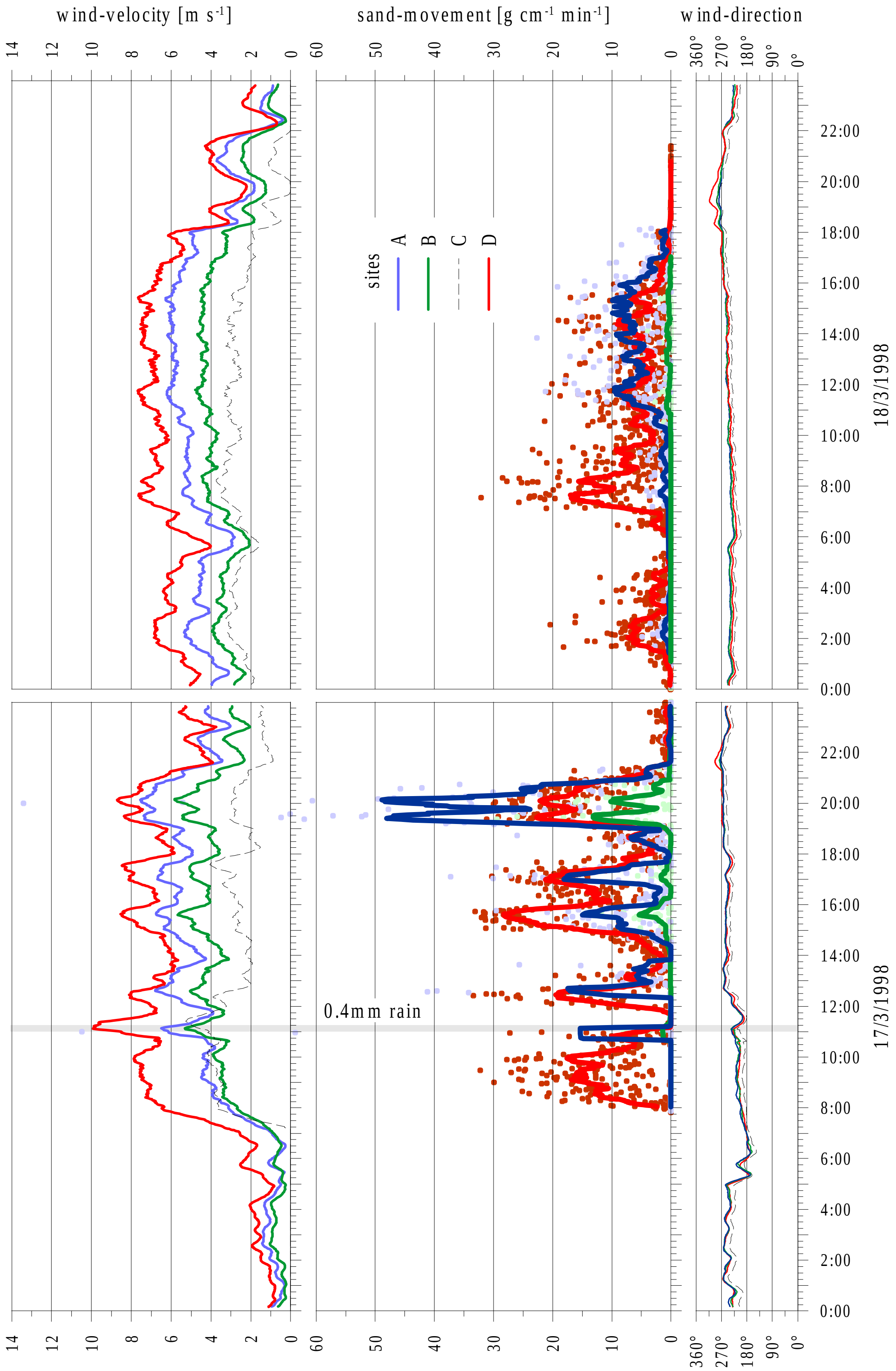

Figure 3.19: Wind conditions and sand transport 17/3 and 18/3/98.

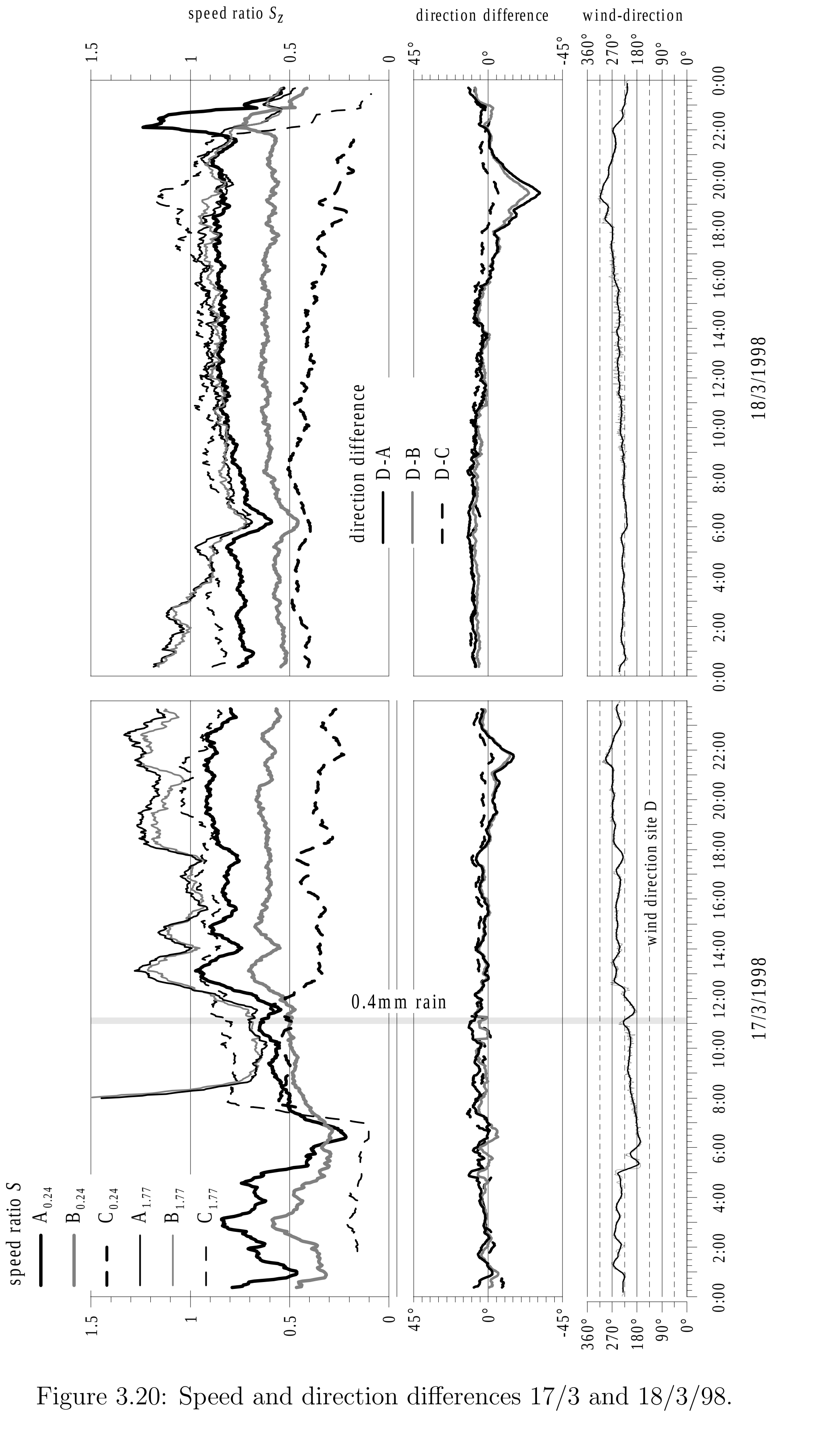

Figure 3.20: Speed and direction differences 17/3 and 18/3/98.

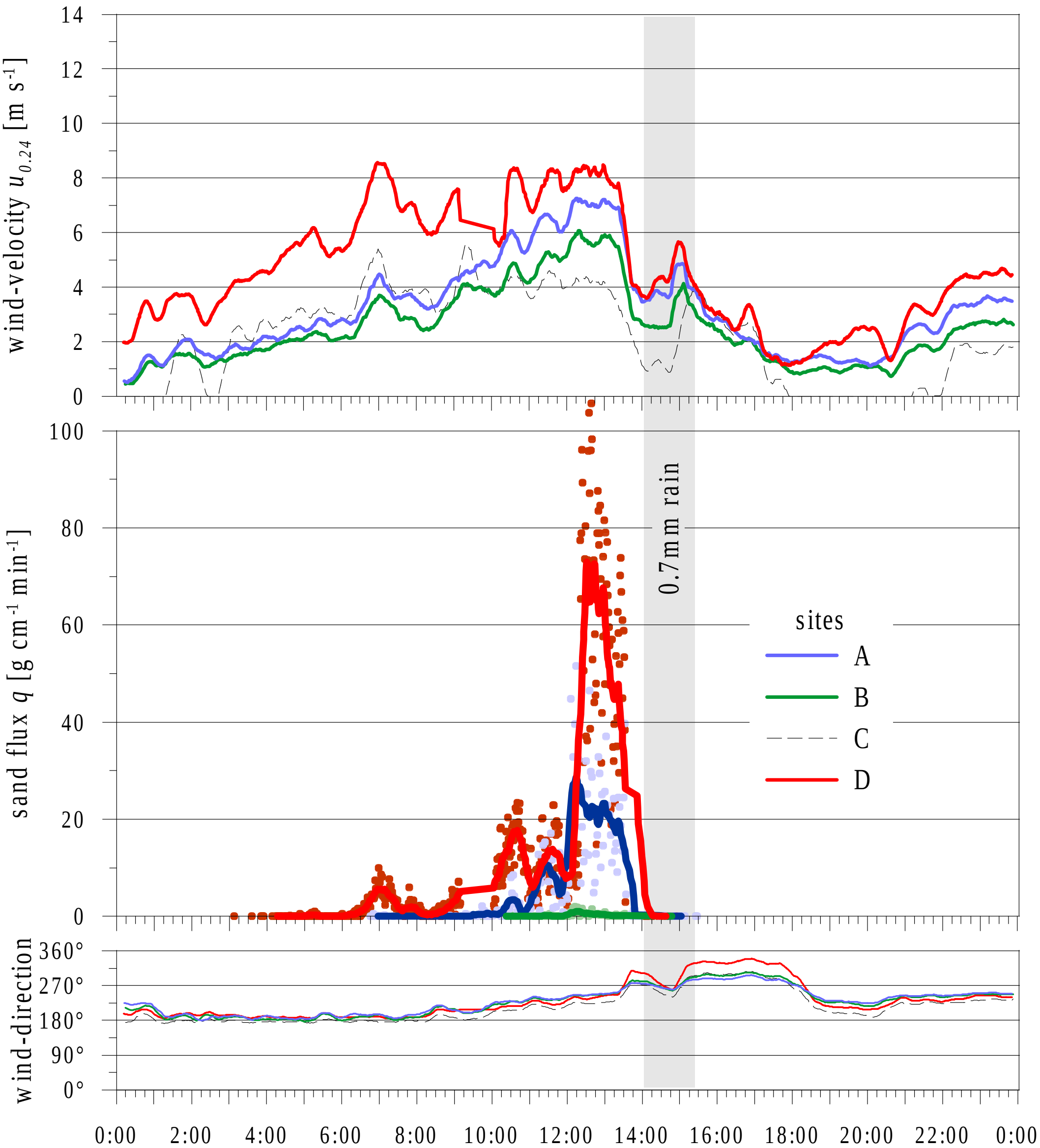

Figure 3.21: Wind conditions and sand transport during 19/3/98.

an increase of wind speed at site D and a sharp rise of S_z for the interdune sites. At site A $S_{0.24}$ and $S_{1.77}$ reached between 0.8 and 1.0, while at site B $S_{0.24}$ remained below 0.7. Values of $S_{1.77}$ at site B were similar to site A. At site C $S_{1.77}$ also rose to values close to 1.0, while at the same time $S_{0.24}$ dropped to values of 0.3. During this period S_z at site A was independent of measuring height. After the change of wind direction, wind speed at site A was higher than before, although wind speed at site D was lower. At site B $u_{0.24}$ reached similar heights as before the change, but the difference towards site A had increased. Continuous saltation was recorded at sites A and B and must be expected at site D as $u_{0.24}$ remained above 4 m $s^{-}1$. Saltation data for site D

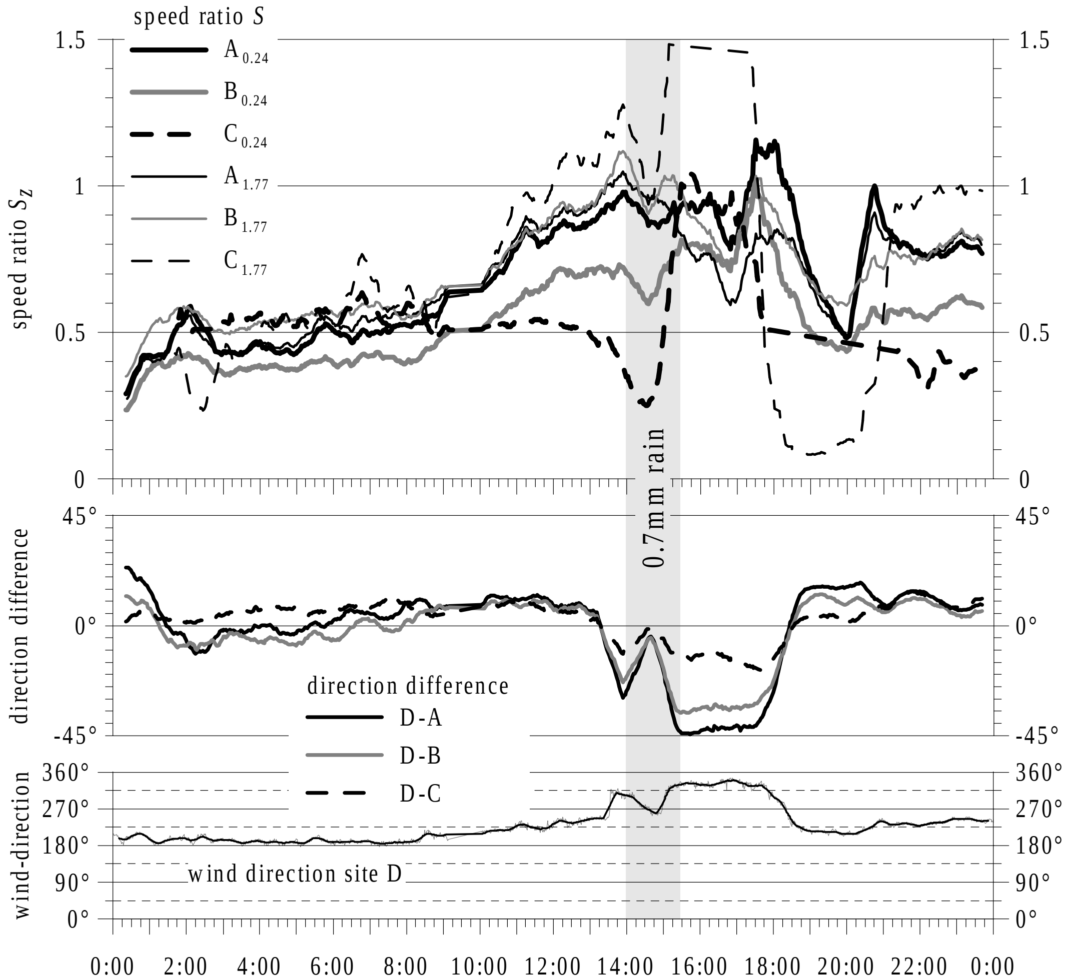

Figure 3.22: Speed and direction differences 19/3/98.

is available only after 11:00. At site C the recorded $u_{0.24}$ of less than 4 m s^{-}1 makes saltation unlikely.

Between 13:30 and 14:30 wind direction changed to NNW, increasing the angle between airflow and dune ridges. This change was accompanied by a significant drop of $u_{0.24}$ and q at the interdune corridor sites. $S_{0.24}$ at sites A and B fell to minimum values of 0.5. At site C $S_{0.24}$ rose sharply to 0.8, while $S_{1.77}$ reached values above 1.4. Wind direction difference at sites A and B increased. Wind speed had decreased gradually since 10:00. No saltation was measured at any site where $u_{0.24}$ dropped below 4 m s^{-}1. At 16:30, 0.6 mm of rain were recorded. This event coincided with the drop of $u_{0.24}$ below 4 m s^{-}1 at site D, thus terminating the first storm phase.

Second phase, 17-18/3/98 (figures 3.19, p.65 and 3.20, p.66) This phase of the storm was characterised by moderate windspeeds, which led to significant saltation mainly at sites A and D. At sites B and C $u_{0.24}$ remained close to and below saltation threshold resulting in low flux, respectively no sand transport at all. Wind speed between the first and the second storm phase remained well below threshold velocity at all sites. Direction was unsteady, mainly SW. Variation of S_z was very high during the low windspeed period. The second storm phase started with a quick rise of wind speed at 6:00. Wind direction changed to S. Saltation started at 8:00 at site D as $u_{0.24}$ reached 6 m s^{-1}. $S_{0.24}$ at that time was 0.5 at site A and 0.4 at site B. Wind speed in the interdune corridors did not pass the threshold for continuous saltation. At 11:00 a short rainfall event of 0.4 mm was accompanied by gusts leading to saltation at site A. During the gusts maximum $u_{0.24}$ of 15 m s^{-1} were recorded at site D. After the rainfall, a break in saltation was recorded at all sites. Sand movement commenced at site D at 11:30, at site A at 12:30. Only single impacts were recorded at site B until 14:30. Wind direction changed after the rain event clockwise to W within one hour. This change led to a rapid increase of S_z at all sites and heights, except at site C. There $S_{0.24}$ dropped from 0.5 to 0.3, remaining below 0.5 until the end of the second storm phase. At sites A and B values of $S_{1.77}$ remained above 1.0 until 03:00 of the 18th.

Wind direction remained constant (WSW) after 13:00 until 18:00 while wind velocity varied. Three distinctive peaks show up between 14:00 and 22:00 in the sand flux records and are visible in the speed graphs as well. After 18:00 wind direction changed to W, $S_{0.24}$ increased at sites A and B, while values at site C dropped from 0.4 to 0.3 and below. The sand flux at site A surpassed the flux at site D during peak velocities between 19:00 and 21:00, $S_{0.24}$ at that time was close to 0.9 at site A. After 22:00 wind direction changed to WSW, accompanied by a general drop of wind speed. A decrease of S_z at the interdune sites and an increase of $S_{0.24}$ at site C was observed in this period. Intermittent sand transport was recorded at site D afterwards as $u_{0.24}$ remained above 4 m s^{-1}, while no saltation was recorded in the interdune corridor. The wind speed rose again during the early hours of the 18th, increasing the flux at site D and leading to saltation at site A around 02:00. At 6:00 a drop of wind speed below u_t at all sites was followed by a change of wind direction to SW. This event shows up clearly in the graph of S_z (fig. 3.20, p.66). Afterwards wind speed increased and saltation was recorded at site D. At sites A and B sand movement remained sporadic until 11:00. When wind direction approached WSW, sand movement intensified at site A as $S_{0.24}$ rose to above 0.8. Between 11:00 and 18:00 S_z was constant at site A. Afterwards wind speed dropped below threshold at all sites. Movement at site B remained low as wind speed remained close to 4 m s^{-1}. $S_{0.24}$ at site B was below 0.7 throughout this phase. $S_{1.77}$ at all sites was close to 0.9 throughout all periods during which wind direction was parallel to the orientation of the dune ridges.

Throughout both days, $S_{0.24}$ at site C was below 0.5, resulting in $u_{0.24}$ of less than 4 m s^{-1}. Amounts of trapped sediment indicated only little sediment movement at the site during this phase of the storm. The second phase ended at 18:00 of the 18th, when wind speeds dropped below u_t. No sand movement was recorded during the night until 03:00, when the third phase of the storm began.

Third phase, 19/3/98 (figures 3.4.1.2, p.67 and 3.22, p.68) This final phase of the storm was characterised by moderate wind speeds at all sites, but exceptional high flux at site D. Wind speed increased from 2:30, threshold speed was passed at site D at 3:00, wind direction was S. A first peak of $u_{0.24}$ was recorded at 7:00 with velocities above 8 m s^{-1} at site D and above 4 m s^{-1} at site A, where the first saltating grains were recorded. After this peak, wind speed decreased at site D until 10:00. Until this time wind direction had remained S. It began to change continuously afterwards towards W. Parallel to this change, $S_{0.24}$ increased from values of 0.5 (site A) and 0.4 (site B) to 0.95 (A) and 0.7 (B) until 14:00. In the course of this development, wind speed in the interdune corridor increased until 12:00, leading to saltation at sites A and B. Wind speed dropped quickly below saltation threshold u_t after 13:30. After the onset of rain at 14:00 (0.7 mm until 15:30) $u_{0.24}$ remained below 4 m s^{-1} at all sites during and after the rainfall except for a short peak at 15:00. No more sand movement was recorded at any site afterwards.

Storm 24/1/1998 This storm is one of several consecutive storms during the second half of January 1998. Sand movement is restricted to site D at the dune ridge. Movement of single grains was detected at site A in the interdune corridor, where $u_{0.24}$ had reached values of 5 m s^{-1}. At site B within the vegetation canopy $u_{0.24}$ remained below 4 m s^{-1} and no saltation was detected. At site C $u_{0.24}$ surpassed 4 m s^{-1} during the peak of the storm, but decreased to values below those measured at sites A and B in the course of the event as wind direction changed. Wind direction during the main phase of the event was from S, gradually changing to SW and W afterwards.

Wind speed values in the interdune corridor remained below 2 m s^{-1} after 2:00 until 6:00. No difference was measured between $u_{0.24}$ at sites A and B. At site C wind speed was close to instrument threshold until 6:00 with a single exception at 4:00, when a small peak is visible in the graph of site D. At 6:00 saltation started at site D as wind speed exceeded 5 m s^{-1}. Wind speed at site D increased to 10 m s^{-1} until 10:00, sand flux rose with wind speed. Sand movement ended shortly after 16:00 and wind speed dropped to 0 at 17:30 at all sites.

The development of the speed ratio S_z and wind deflection during the storm is shown in figure 3.24, p.72. Values of S_z at both interdune sites reached their maximum between 13:00 and 15:00. The increase of S_z at these sites occurred parallel to the change of wind direction from S to W at site D on the dune crest. Maxima were reached when wind direction was parallel to dune ridge orientation. At site C only $S_{0.24}$ could be determined as $u_{1.77}$ was not measured during this storm. The peak of 0.5 was reached during winds from 180°, values declined as wind direction shifted towards W.

The values of direction difference between the dune crest and the interdune corridor corresponded with calculated average values during the main phase of the storm. Deflection was highest when wind direction at site D was oblique to dune ridge orientation. Minima were recorded during periods of wind direction

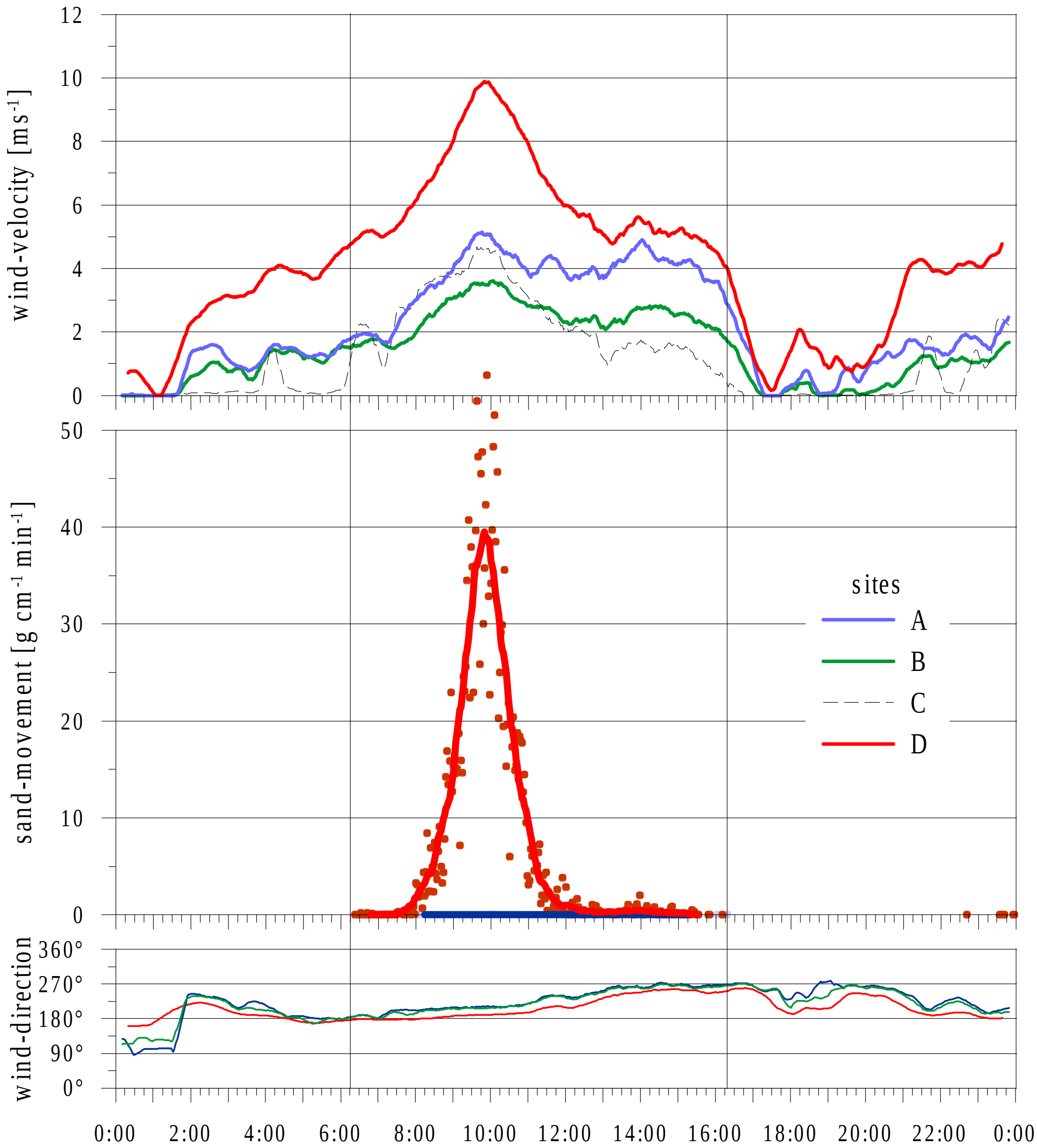

Figure 3.23: Wind conditions and sand transport during 24/1/98.

perpendicular or parallel to the ridges. Direction difference for site C could not be determined due to missing direction data.

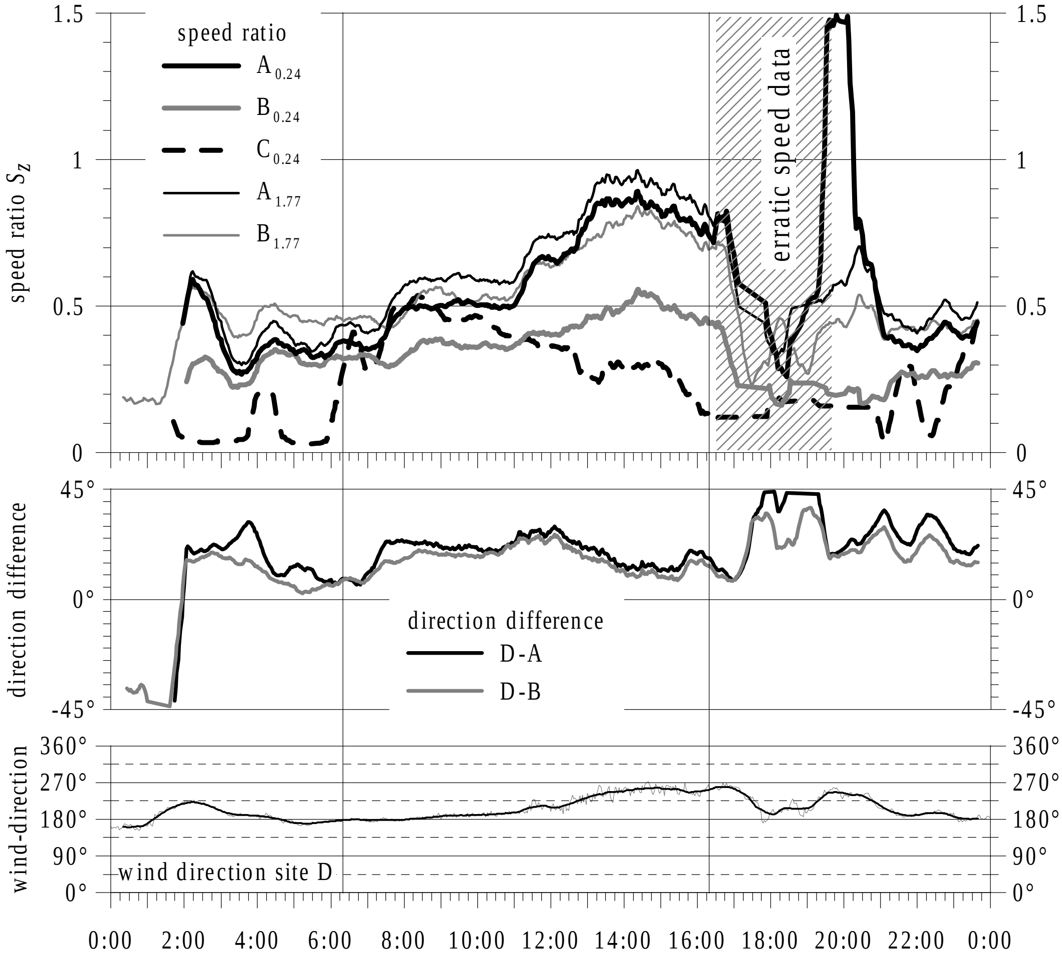

Figure 3.24: Speed and direction differences during the storm of 24/1/98.

Diurnal winter winds, January 1998 Cyclonic storms were identified as the major cause for sand movement during winter. Yet when regional pressure gradients were low, local wind systems developed, reaching wind speeds above u_t at the dune crest at site D.

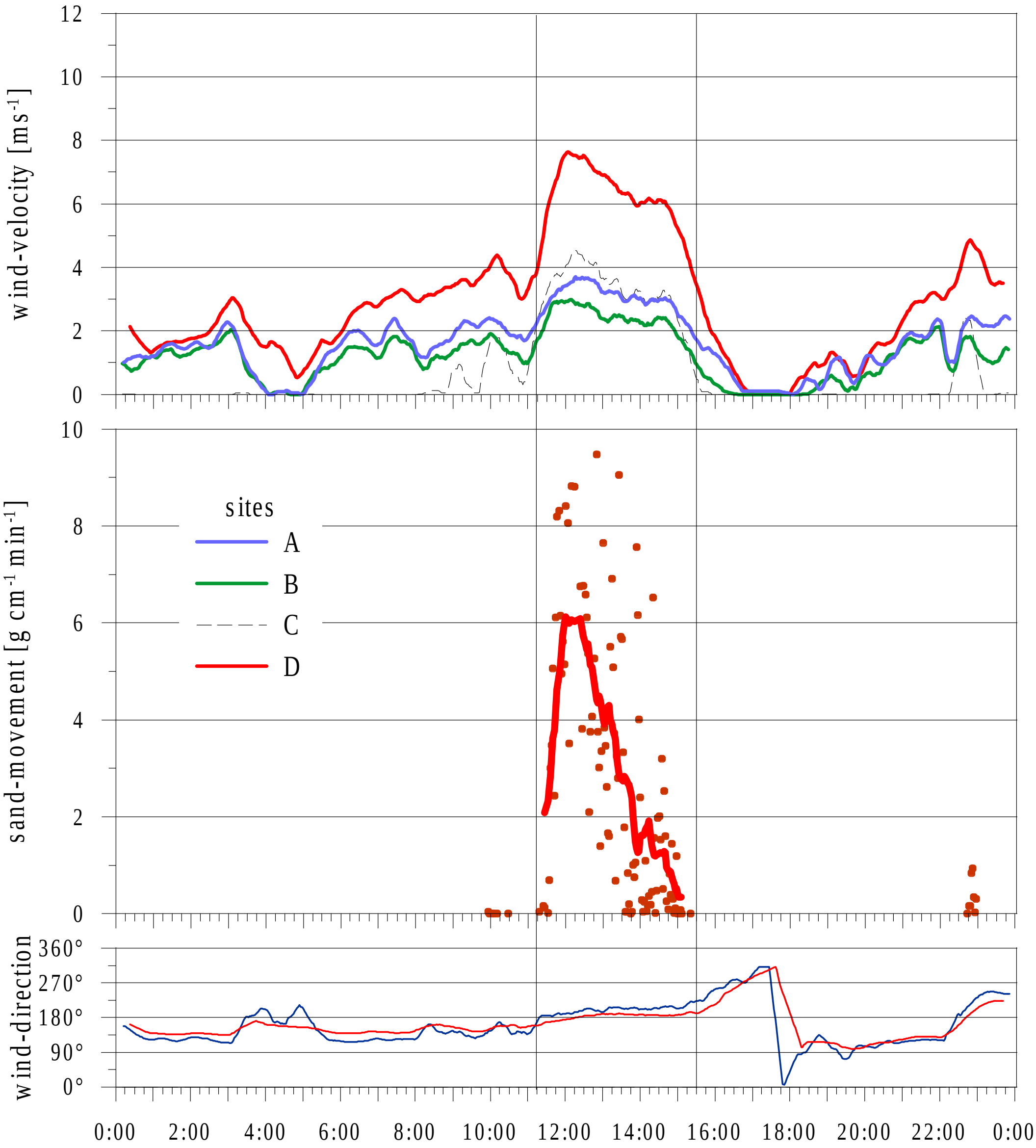

Figure 3.25: Wind conditions and sand transport during 17/1/98.

An example for such a development is the second half of January 1998, when a shallow low over the SE-Mediterranean permitted the development of the diurnal wind system during a period of 11 days. During this period wind speed increased beginning in the early morning hours around 04:00 and reached its maximum by late morning/noon. Saltation started a s early as 08:00, lasting four to six hours. Wind speeds at site D were only moderate, reaching up to 8 m s^{-1}. The rise of wind speed was accompanied by a continuous clockwise

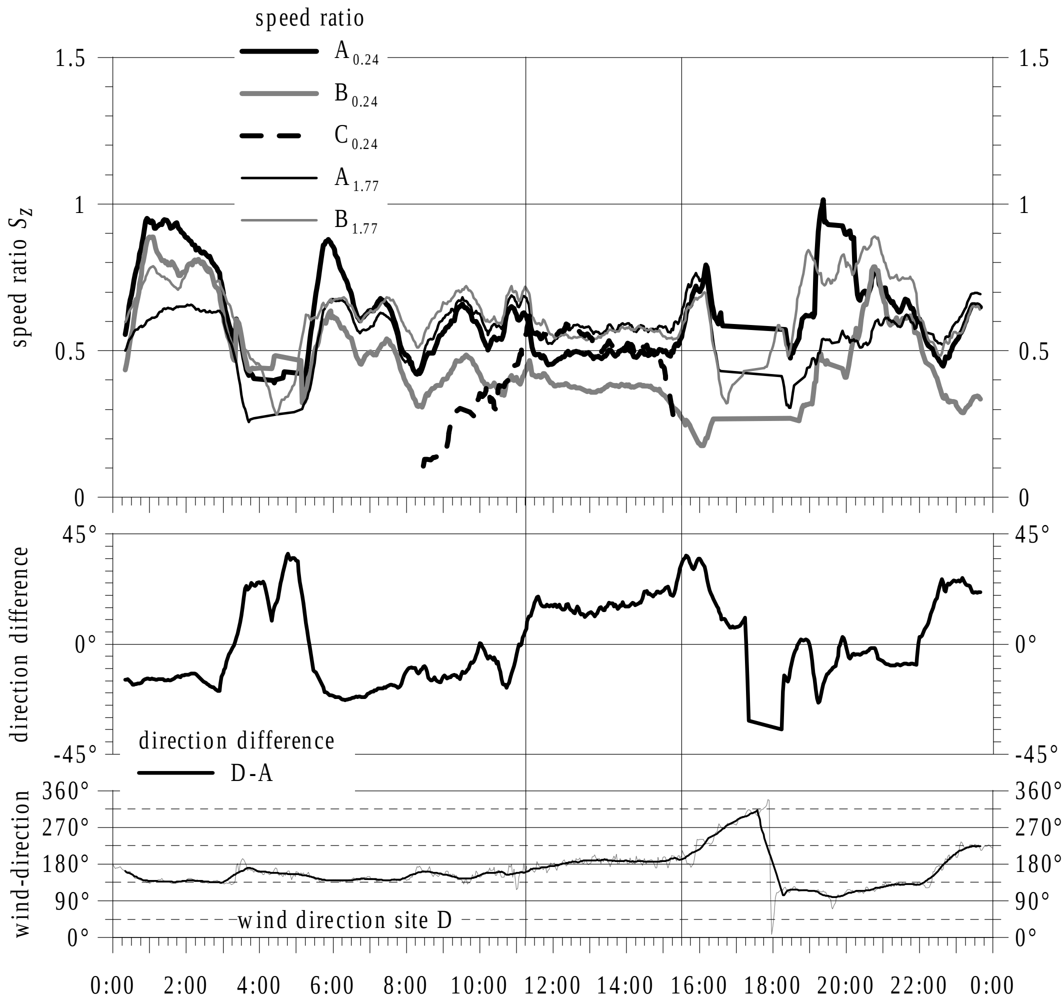

Figure 3.26: Speed and direction differences 17/1/98. Missing direction data for sites B and C.

change of wind direction from SE during the early morning to NW in the evening and on to NE during the night. Figures 3.25 to 3.30 show wind and sand transport data of three consecutive days. Direction data is missing for sites B and C. These events are examples of sand transport by southerly winds within the diurnal wind regime. Similar patterns were recorded on 26/1, 28/1 and 30/1/1999.

The change of wind direction is reflected in the graphs of direction difference and speed ratio S_z. Deflection between sites A and D reached minima for winds parallel or perpendicular to dune orientation, shown very clearly in figure 3.28, p.76. A linear increase of S_z at all heights parallel to the change of wind direction from S to WSW was determined for the interdune sites. Except for site B2, where $S_{0.24}$ reached only a maximum of 0.6, S_z values were above 0.9 for dune parallel winds (270°). At site C, $S_{0.24}$ reached values of 0.65 during winds at site D from 180°. For dune parallel winds values dropped to 0.1 and below. Values of up to 0.4 were determined for northerly directions. Similar

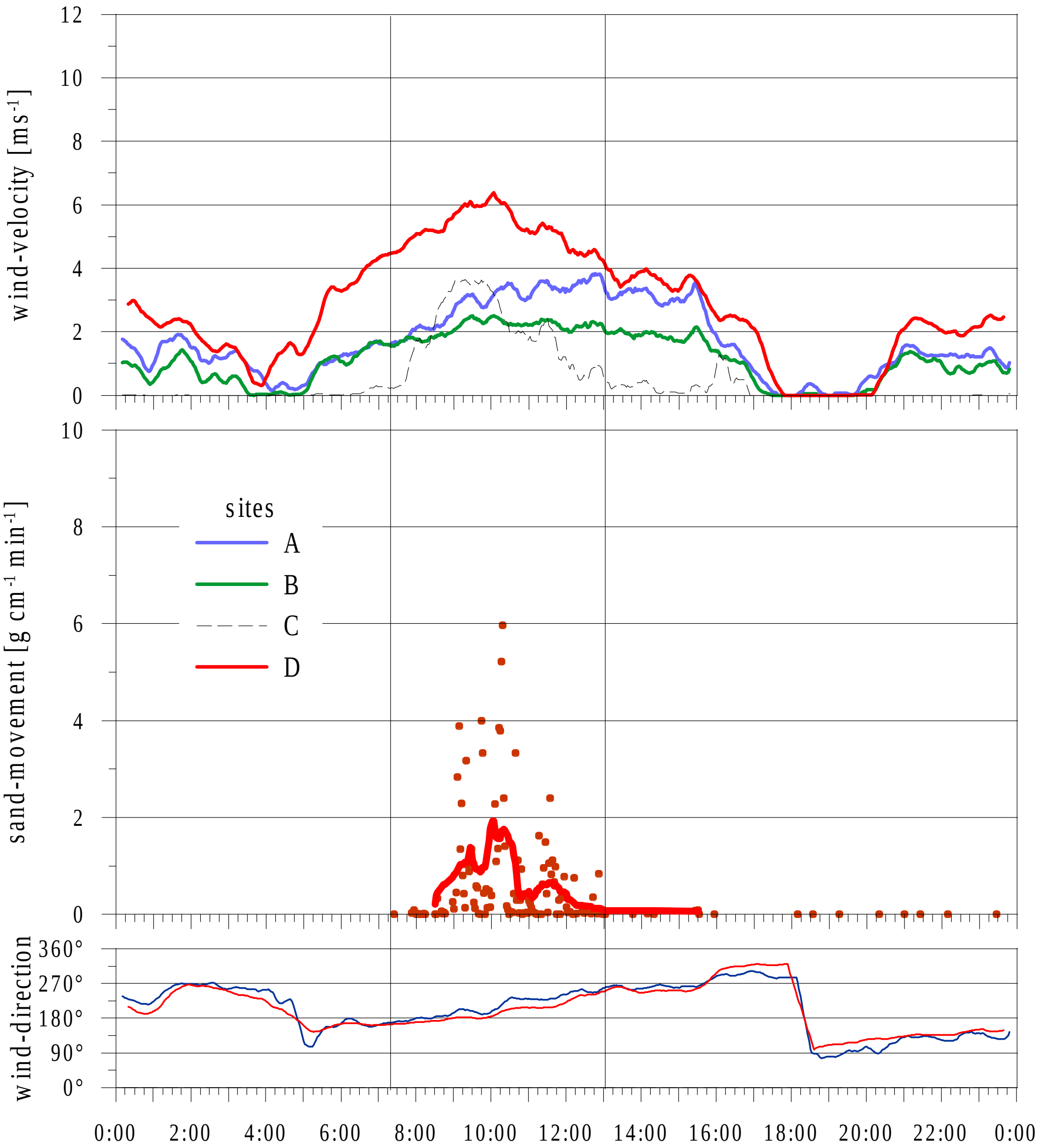

Figure 3.27: Wind conditions and sand transport during 18/1/98.

behaviour was observed on 17/1 and 19/1 (figures 3.26, p.74 and 3.30, p.78).

The periods of sand transport on 17/1 and 19/1 were characterized by little variation of wind direction and thus little variation of S_z at the interdune sites. Wind speed $u_{0.24}$ at site A reached values close to 4 m s^{-1}, but did not lead to continuous sand transport. Short gusts in the evening of the 19/1 (fig. 3.29, p.77) led to isolated saltation at site A. Within the vegetation at site B $u_{0.24}$ remained below the values of site A during the whole period, thus no saltation was detected. At site C $u_{0.24}$ reached maxima during southerly winds, while during other wind directions $u_{0.24}$ was close to instrument threshold. As 4 m s^{-1} were surpassed at noon on the 17/1 (3.25, p.73) it is likely that saltation occurred.

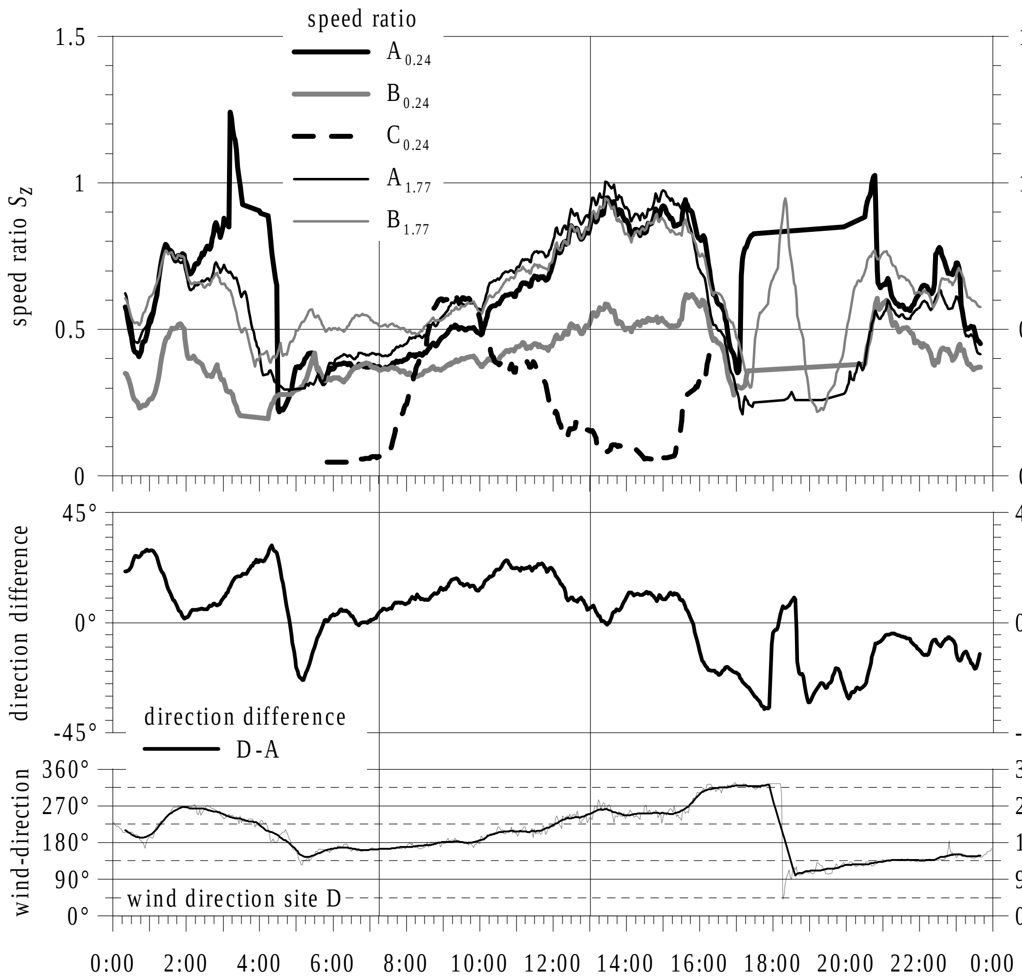

Figure 3.28: Speed and direction differences 18/1/98. Missing direction data for sites B and C.

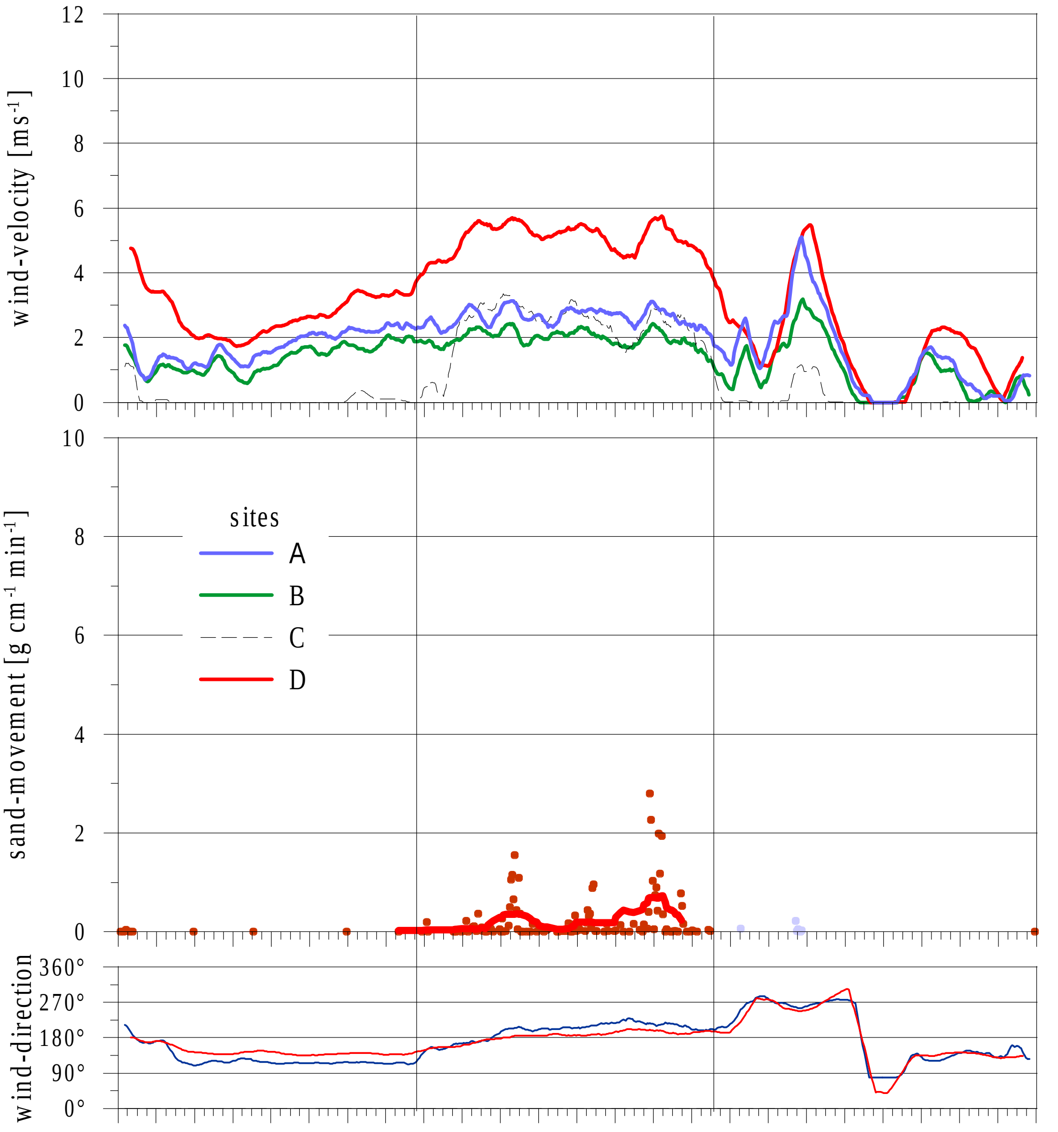

Figure 3.29: Wind conditions and sand transport during 19/1/98.

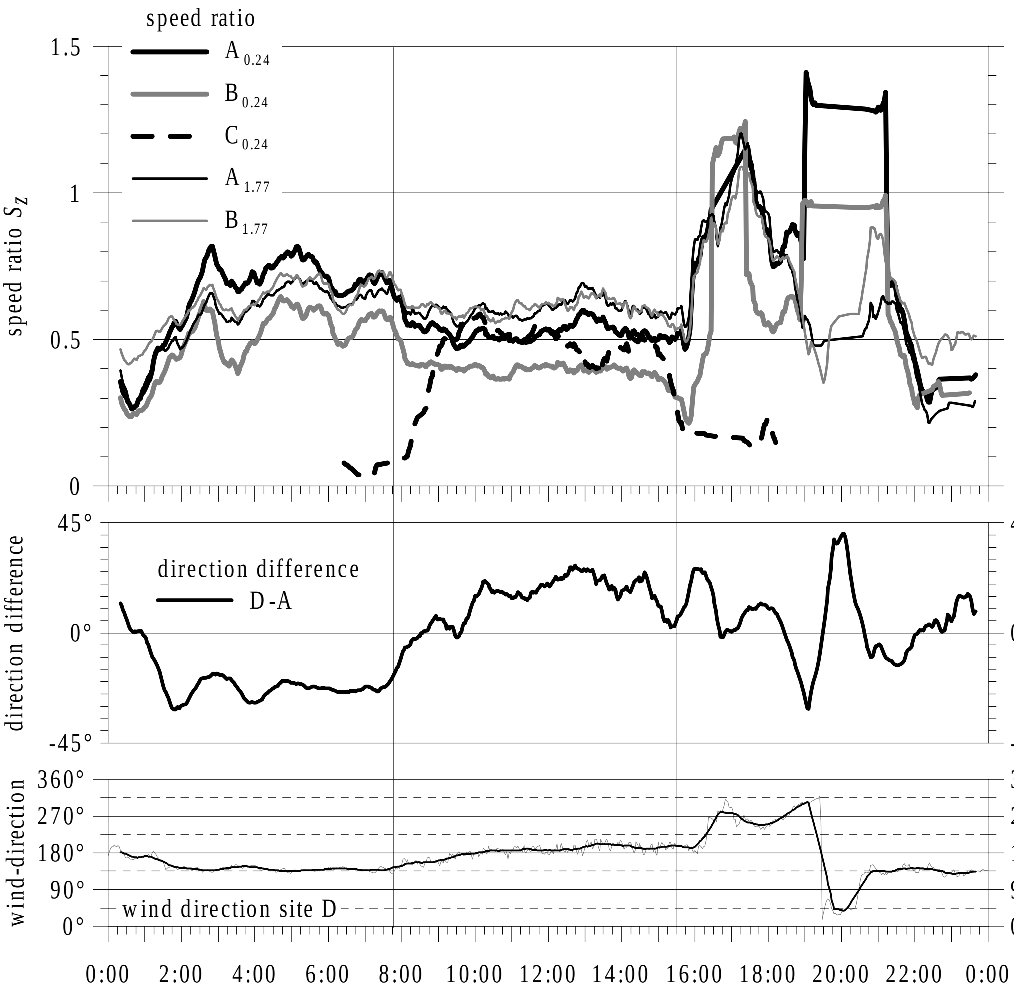

Figure 3.30: Speed and direction differences 19/1/98. Missing direction data for sites B and C.

3.4.1.3 Summer winds

During the summer wind regime, u_t was reached and surpassed daily at the dune crest. Winds above saltation threshold were from NW to N. During none of the sea-breeze events recorded in the research period did wind speed $u_{0.24}$ in the interdune corridor reach 4.0 m s^{-1}. The high angle between dune and wind direction resulted in low values of $S_{0.24}$ for the interdune corridor sites, in general below 0.5 (see figure 3.4, p.46). Under these circumstances, based on a threshold speed u_t of 4 m s^{-1}, the required minimum $u_{0.24}$ at the dune crest to cause movement in the corridors is $\geq$8 m s^{-1}. Yet, the highest recorded values of $u_{0.24}$ at site D were between 6 to 8 m s^{-1}. It is therefore assumed that during summer no significant saltation occurred at sites A and B in the interdune corridor caused by the sea-breeze. As the figures in table 3.10, p.109 show, the events presented as examples represent a high magnitude summer event (19/4) and two events of average strength. During an average early summer day (April) between 300 and 600 g cm^{-1} are moved by NW-winds at the active dune crest. Regular daily sand movement during summer is expected to be of similar intensity. Aeolian activity at site C appears unlikely during summer, as $u_{0.24}$ at the site did not surpass values measured at the interdune corridor sites.

3.4.1.4 Examples

19/4/99 This first example of diurnal summer winds is exceptional in wind speed and sand flux at the dune ridge site D, yet the general pattern of the development during the day is typical for early summer. Mean sand flux exceeds all winter storms (see figure 3.8 p.51), and transported mass is above average winter storms (figure 3.10 p.52). These exceptional values may be caused by the fact that the location of the *saltiphone* coincided with the area of maximum change (see figure 3.56, p.108). At the steep windward slope, flow acceleration of the northerly winds led to increased erosion. The reason for the unusally high wind speeds is seen in the exceptionally high temperatures in the Negev on that day. Despite the high wind speeds at the dune crest, no saltation was recorded in the interdune corridor.

The situation at site D on 19/4/99 was such that the mast holding the anemometers was positioned at the southfacing slipface of the dune´s crestline. This slipface had been built up by earlier northerly winds, reversing the crestline created by the SW-storms of winter. The lowermost anemometer (0.24 m) had been covered during the previous day, while the rotor of the instrument originally mounted at 0.65 m was now about 0.2 m above the surface of the crest. Its data was therefore used for the calculation of $S_{0.24}$. Subsequently the height of the uppermost anemometer of site D was reduced by about 0.4 m. A comparison with its counterparts at 1.77 m in the interdune corridor to determine $S_{1.77}$ therefore yielded higher values than usual, up to 1.5 during SW to W winds.

The wind speed and wind direction graphs in figure 3.31, p.80 show the calm

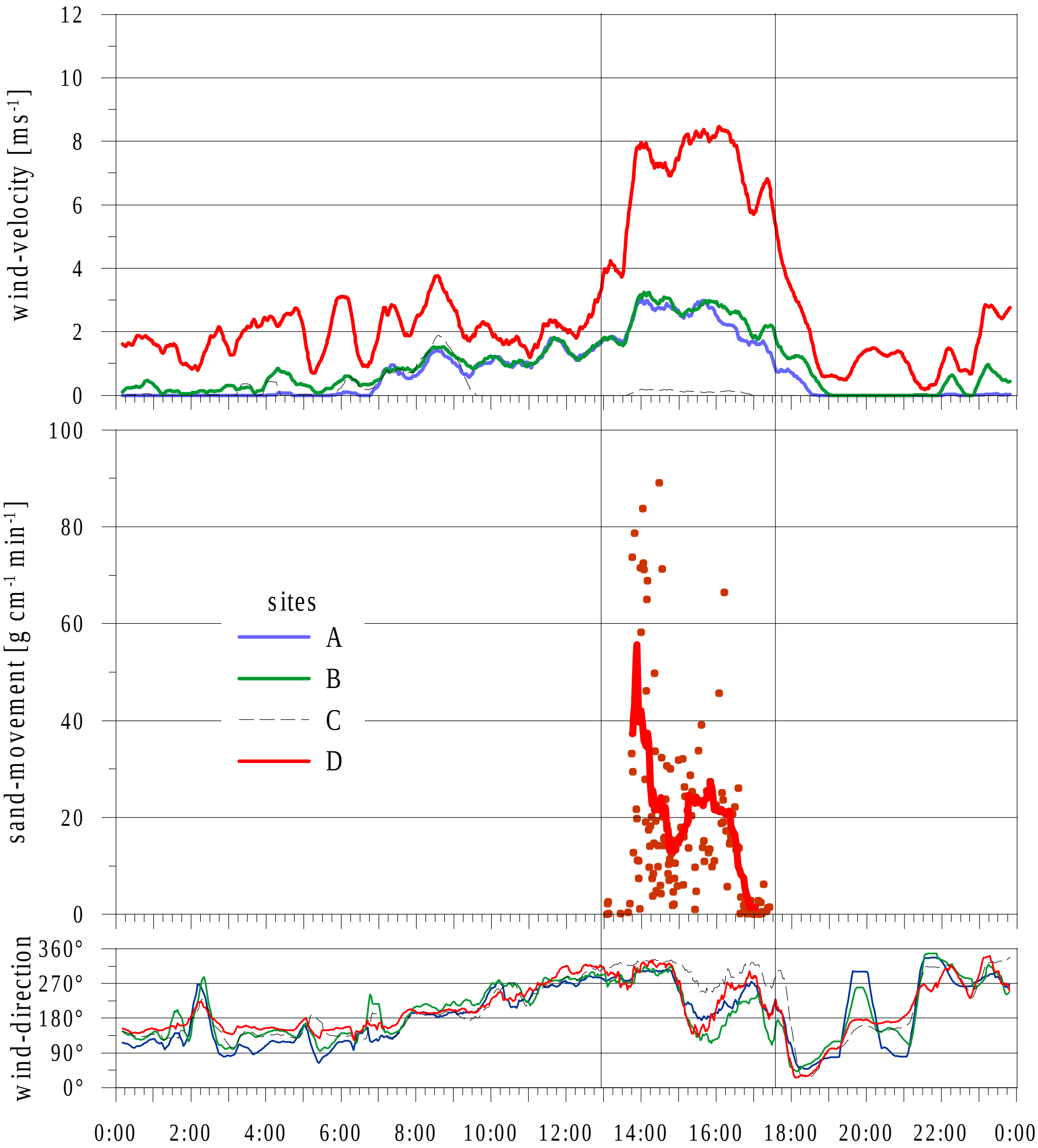

Figure 3.31: Wind conditions and sand transport during 19/4/99.

in the interdune corridor and low $u_{0.24}$ at the dune crest for SE-winds during night-time. As wind speed increased during the early morning hours, wind direction changed from SE to S and after 10:00 gradually moved to northerly directions. The change of wind direction from S to W is accompanied by an increase of the speed ratio S_z, which rose to 1.5 ($S_{1.77}$) and above 0.8 for ($S_{0.24}$) at both interdune sites during dune-parallel wind conditions (figure 3.32, p.81). Wind speed $u_{0.24}$ at site C did not pass instrument threshold except when wind direction was S or N, therefore $S_{0.24}$ has not been determined. Values of $S_{1.77}$ indicate that $u_{1.77}$ at site C is similar to sites A and B during the morning and considerably higher than in the interdune corridor during maximum speed at site D.

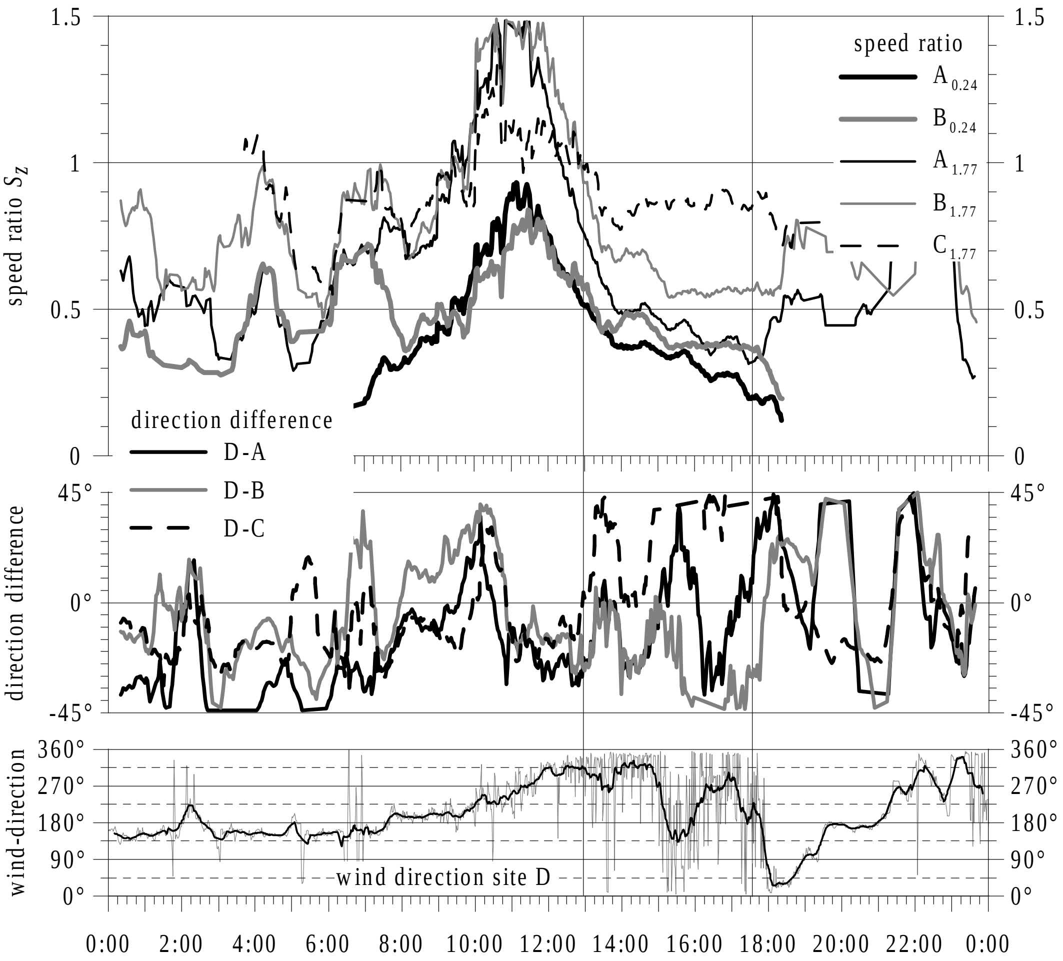

Figure 3.32: Speed and direction differences 19/4/99.

Initial saltation was detected at 13:00 at the dune crest (site D), when $u_{0.24}$ surpassed 4.0 m s^{-1}. Average wind speed soon afterwards rose sharply to 8.0 m s^{-1} – gusts reaching 10.0 m s^{-1} – and remained at that value until 16:30, leading to high flux rates q. During the period of sand movement, wind direction changed towards NE. The apparent SE winds visible in the direction graph in figure 3.31 are a result of averaging of direction values. Actual values in 3.32 show the variation of wind direction during that period. Parallel to the change of wind direction S_z decreased. The decrease was especially pronounced at site A, $S_{0.24}$ reaching a minimum of 0.2 while at site B values remained above 0.3. Values of $S_{1.77}$ at sites B and C stabilized during the second half of the event at 0.55 (B) and 0.83 (C). At site A, $S_{1.77}$ was lower than at site B during this period, thereby demonstrating the influence of corridor width on the speed ratio. Wind speed $u_{0.24}$ in the interdune corridor did not surpass u_t, but remained at about 3.0 m s^{-1}. As $u_{0.24}$ at site D dropped below 4.0 m s^{-1} around 17:30, saltation stopped. After 19:00 $u_{0.24}$ in the interdune corridor fell below instrument threshold, while at the dune crest continuous air movement was recorded throughout the night.

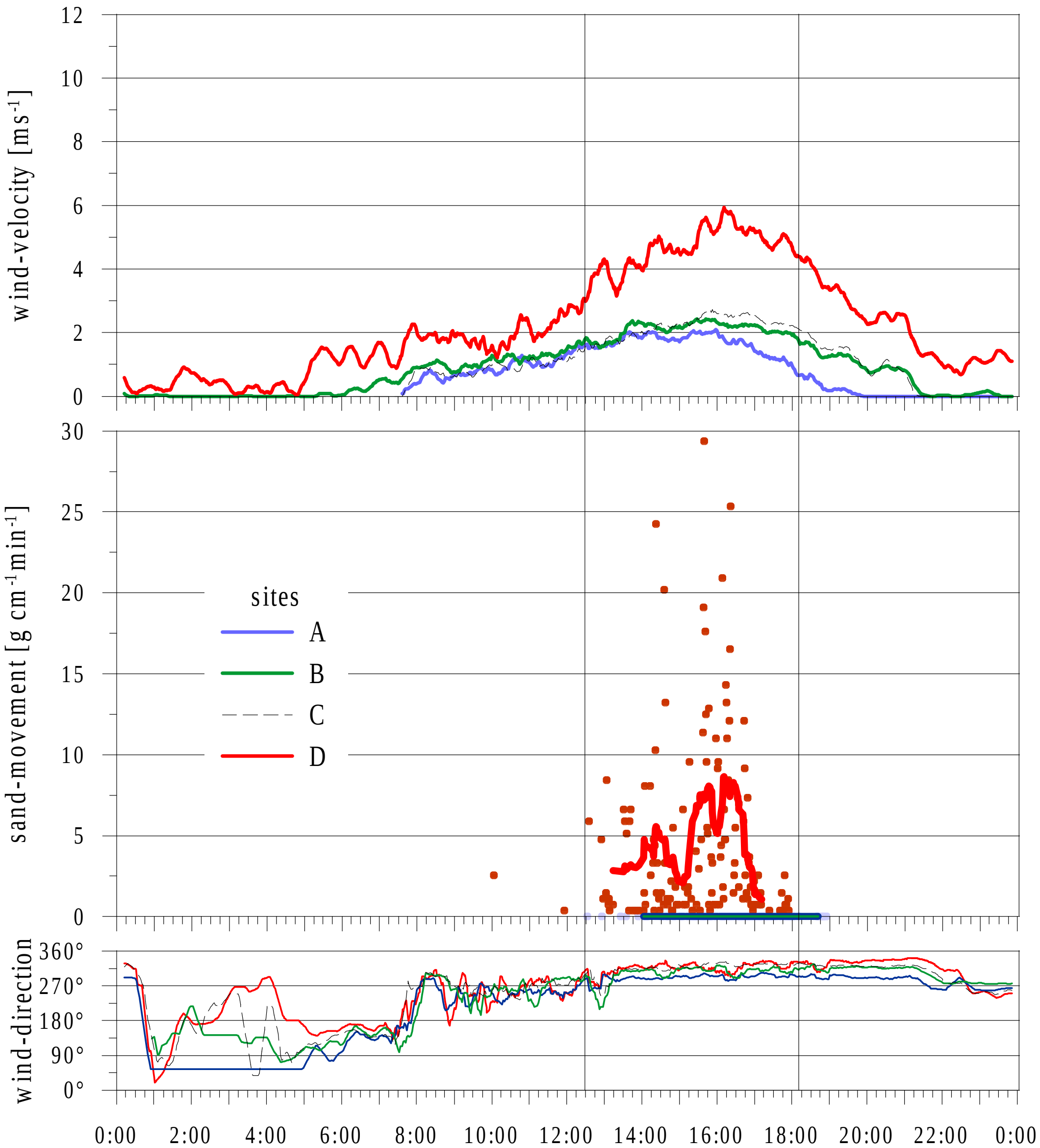

Figure 3.33: Wind conditions and sand transport during 21/4/99.

21/4/1999 This example shows the daily cycle of wind and sand transport of early summer. During the calm night air movement did not surpass instrument threshold (0.15 m s^{-1}) in the interdune corridors, whereas on the dune crest low winds were measured. Wind speed increased after 05:00, air movement was also measured at site B in the interdune corridor. At site A and on the low ridge of site C airflow passed instrument threshold at 07:30, accompanied by a rapid change of wind direction from SE to W. A steady increase of wind speed was measured in the corridor up to the maximum of 2 m s^{-1} around 13:30. At site C and at the dune crest, wind speed increased further, reaching its maximum of 3 m s^{-1} (site C) and 6 m s^{-1} (site D) at 16:00. Development of wind speed at site B was parallel to site A until 13:30. Afterwards, as wind direction changed further clockwise, it increased further, following the graph

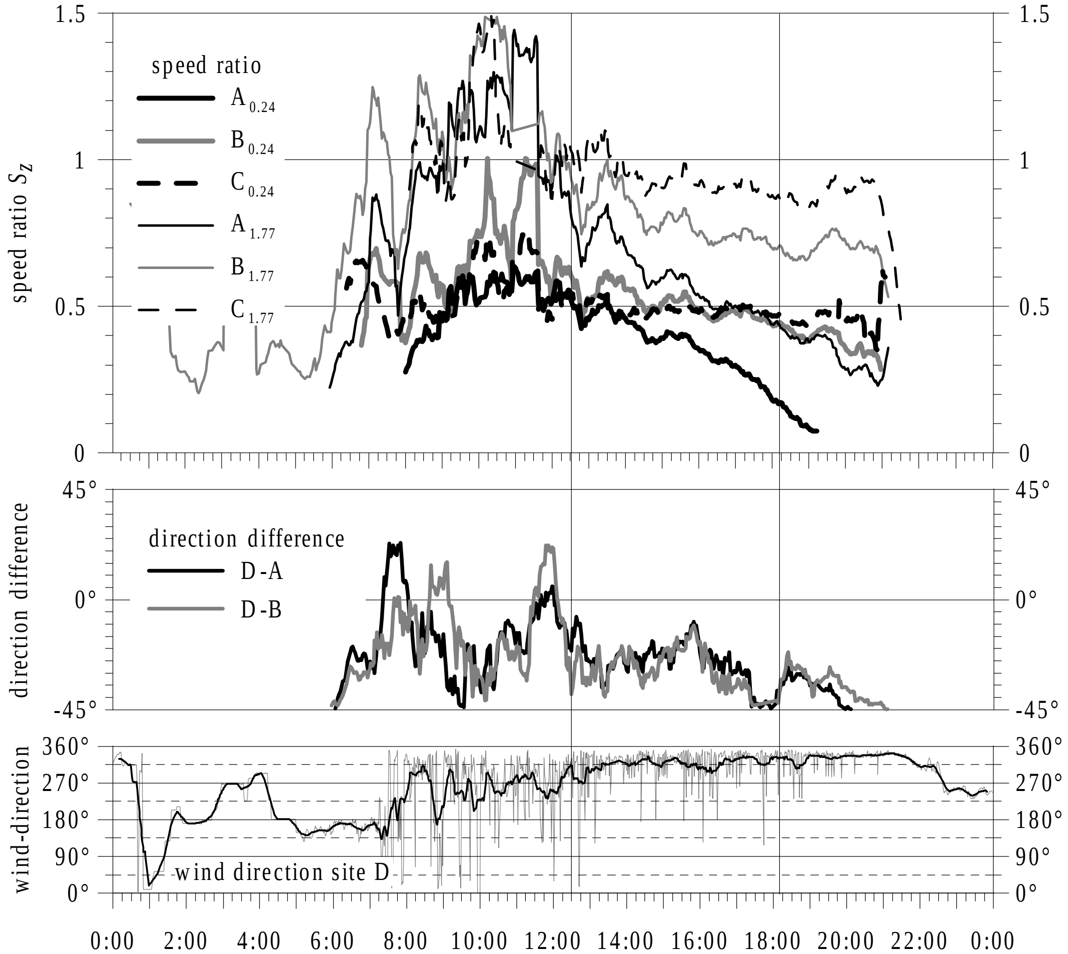

Figure 3.34: Speed and direction differences 21/4/99.

of site C. After 16:00 wind velocity decreased gradually and $u_{0.24}$ fell below instrument threshold at 20:00 at site A, and at 21:00 at sites B and C. At site D air movement was registered all night.

Saltation was recorded at site D between 12:30 and 18:00, as long as wind speed $u_{0.24}$ remained above 4.0 m s^{-1}. The speed ratio $S_{0.24}$ at site A remained below 0.6 throughout the day, even when wind direction was parallel to dune orientation. Values of $S_{1.77}$ were considerably higher and reached 1.4 during dune parallel wind directions. Because of the reduced height of anemometer D3 (see above), these values of $S_{1.77}$ are considered to be too high. S_z at all heights diverged between sites A and B as wind direction changed towards more northerly directions. The decrease of S_z caused by the shift of wind direction was steeper at site A than at site B. Despite the vegetation cover at site B, values of $S_{0.24}$ were higher than at site A. The same results were determined for $S_{1.77}$. These results are in accordance with the general trend as shown earlier in chapter 3.2 and figure 3.4, p.46. At site C S_z at both heights remained almost constant during northerly winds. Values of $S_{1.77}$ varied between 1.1 and 0.8.

Wind direction changed gradually from unsteady westerly directions to NW and N at site D. An easterly component was present during the morning. More westerly values were recorded in the interdune corridor. After 13:00 wind direction variation was reduced at all sites. Deviation from wind direction at sites A and B is high for SW and NW winds. Direction readings from site C were faulty and could not be used for calculating devialtion.

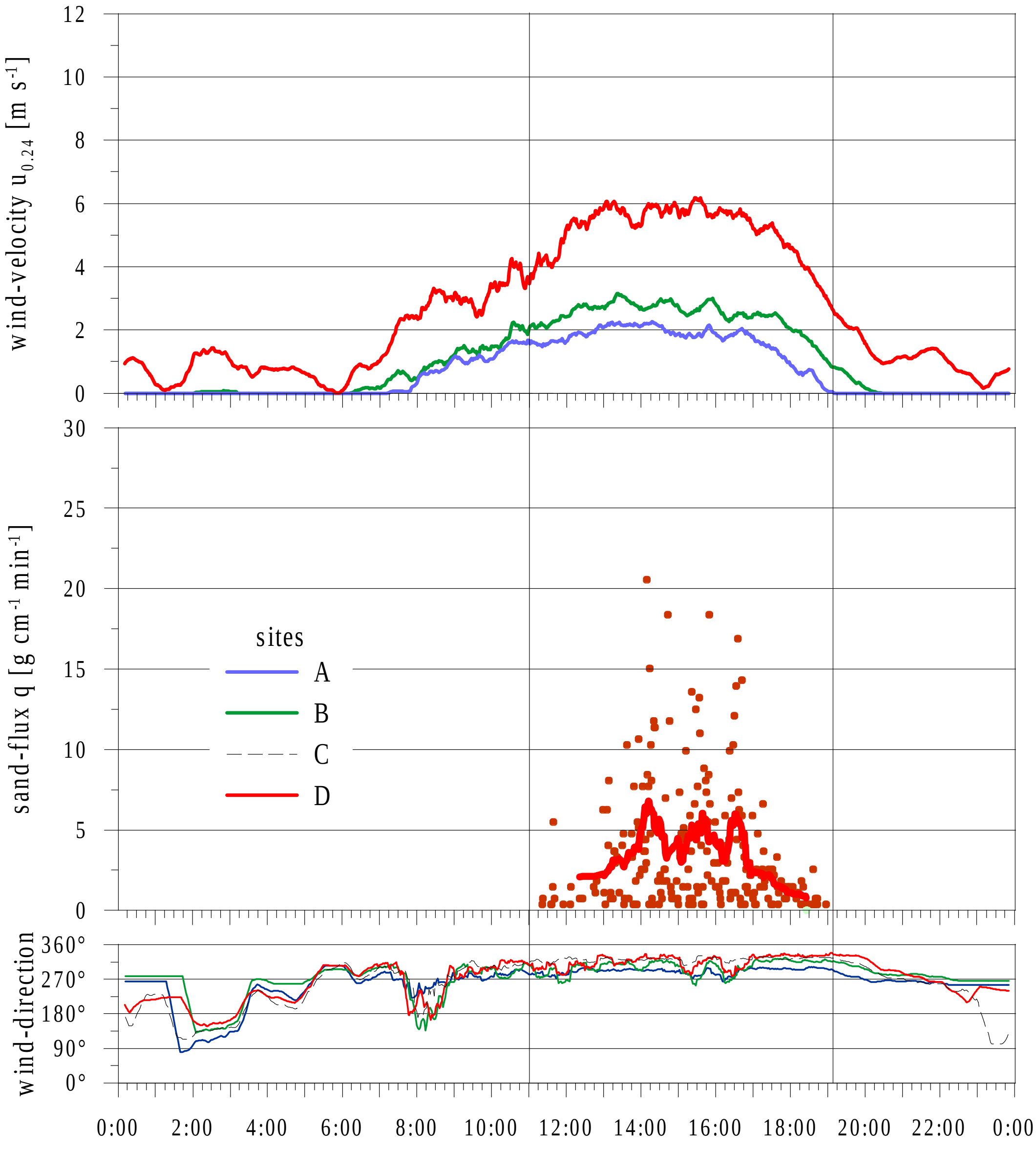

Figure 3.35: Wind conditions and sand transport during 24/4/99.

24/4/1999 The third example was recorded three days after the previous event. Its windspeed graph (figure 3.35) shows the uniformity of the diurnal wind system at Sde Hallamish during early summer. Saltation was again recorded only at site D, where the onset of movement coincided with the in-

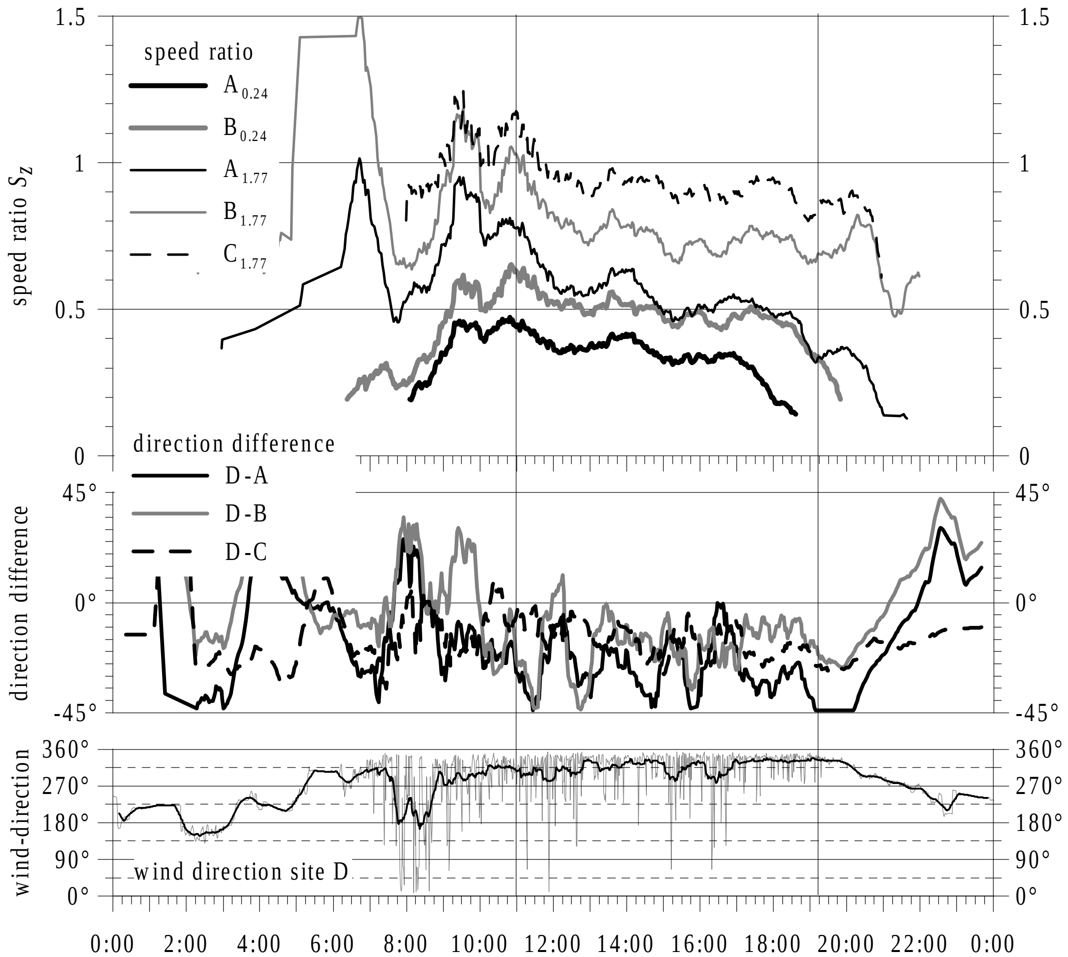

Figure 3.36: Speed and direction differences 24/4/99.

crease of windspeed $u_{0.24}$ to values above 4.0 m s^{-1} and ended when $u_{0.24}$ fell below 3.5 m s^{-1}. Maximum windspeed at the interdune corridor sites was 2 m s^{-1} (site A) and 3 m s^{-1} at site B. At site C instrument failure caused the loss of data for $u_{0.24}$. Maximum $u_{0.65}$ at site C was 4 m s^{-1}. The difference between the events of 21/4 and 24/4 lay in the duration of the fully developed sea-breeze. On the 24th maximum wind speed was reached earlier and remained virtually constant for almost four hours, while on the 21st wind speed increased constantly to the maximum and decreased in the same way after a single, short peak. During the presence of the sea-breeze, a slow but constant change of wind direction towards N was recorded. This change occurred on both days, but on the 21st it was faster, documented by the steeper decrease of S_z on the 21st (compare figures 3.34 and 3.36). Otherwise the development of S_z followed the same pattern as during the previous examples.

3.4.2 Rainfall and surface moisture

Surface moisture increases the cohesion between sand grains and thus increases the necessary force for the initiation of saltation. This in turn leads to higher threshold speed u_t. Data on rainfall amount and intensity have been provided by the *AERC*. The installed equipment detected rainfall of 0.1 mm. The data collected by the rain recorders has been synchronised with recorded wind data to determine the effects on sand movement. The rainfall events at the research site are associated with frontal depressions and are usually of short duration.

Rainfall during the research period was below average. The winter 1998/99 has been the driest since the beginning of rain measurements at the site in 1988, with a total of 29.4 mm (11 events) compared to 74.8 mm (43 events) in the previous winter and 75.4 mm in 1996/97 (53 events) (AERC-DATA).

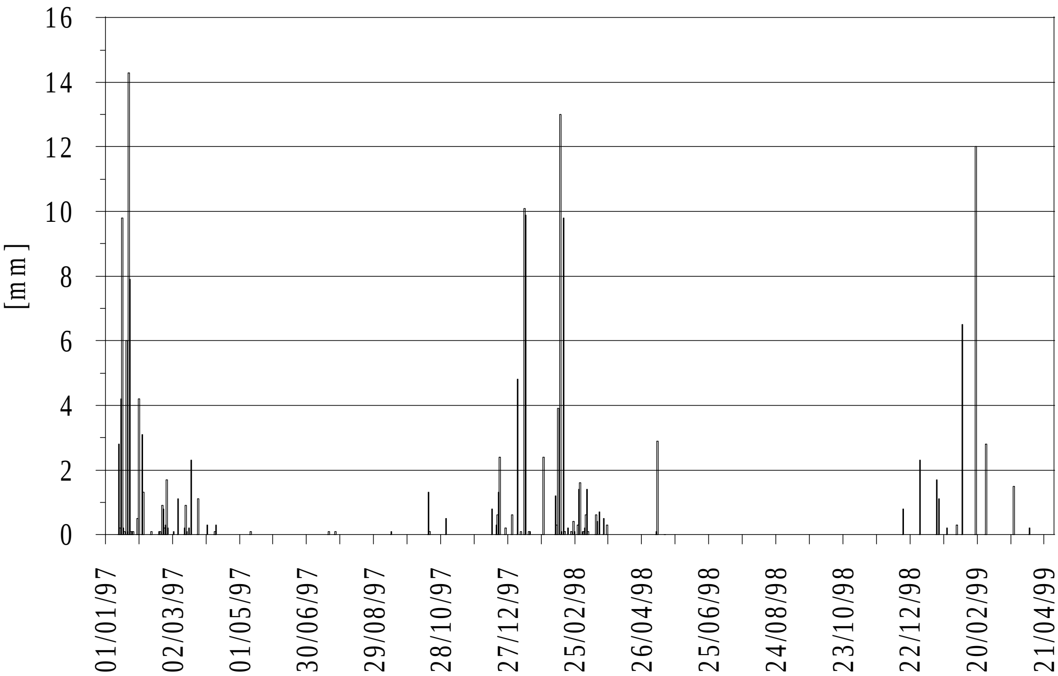

Figure 3.37: Daily rainfall at the research site January 1997 - May 1999 (data provided by AERC).

3.4.2.1 Low intensity rain

During cyclonic storms of average magnitude, very low amounts of precipitation (0.1 mm) led to a significant reduction of saltation. An example is the event of 13/12/97. As shown in figure 3.38, p.87, the 0.5 mm of rain which fell within 1:45 hours reduced sand flux at the dune crest to isolated grain movements for a period of 4.5 hours. The surface had been dry prior to the storm, as the previous rain had fallen about five weeks before on 2/11/97. Saltation came to a halt immediatly after the onset of rain, when 0.2 mm fell within 11 minutes. Wind speed ($u_{0.24}$) throughout the event remained above 4 m s^{-1}

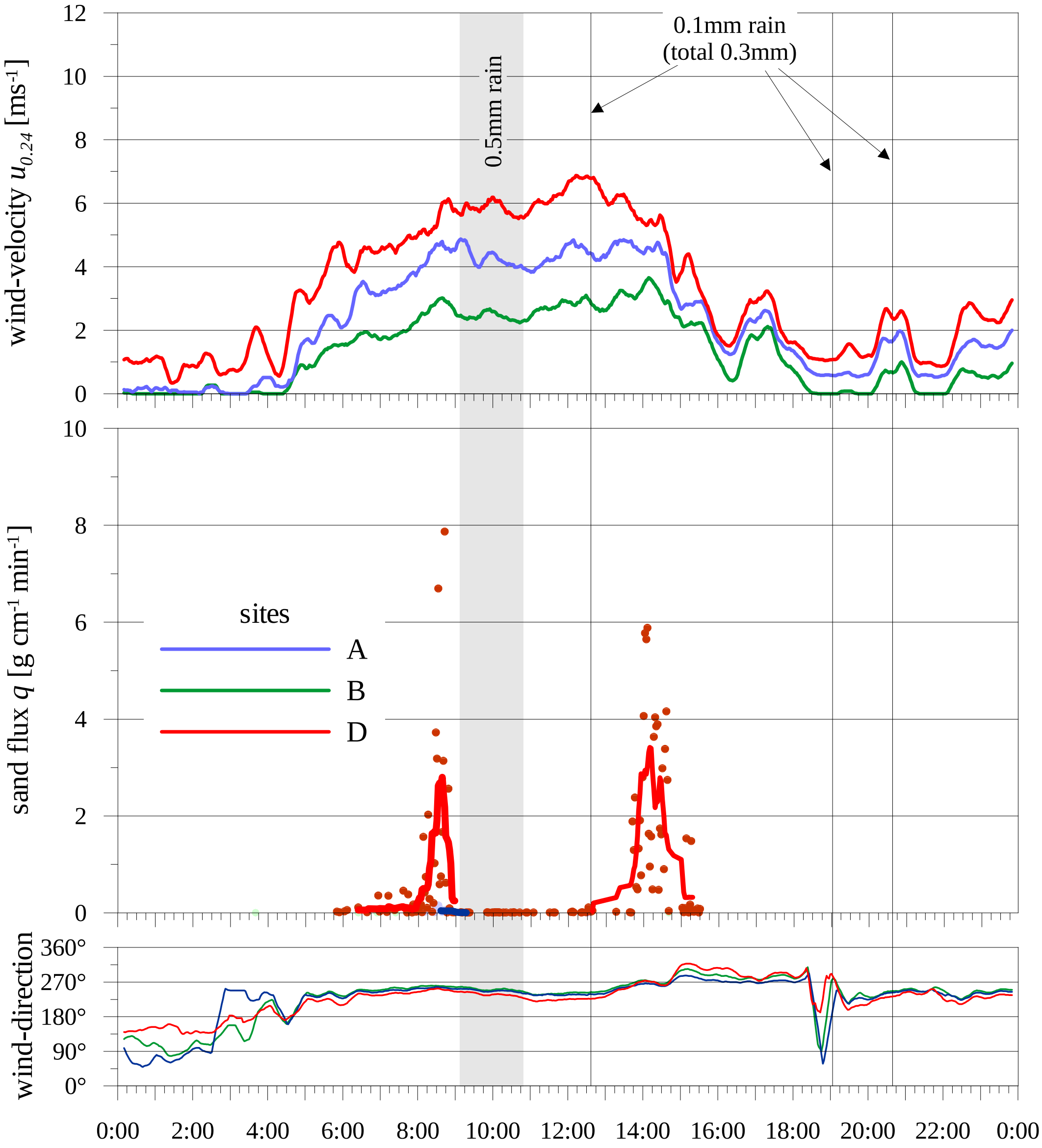

Figure 3.38: Wind conditions, sand transport and rainfall during 13/12/97.

at sites A and D, which under dry conditions would lead to saltation. Speeds of 6 m s^{-1} as measured at site D caused flux of about 20 g cm^{-1} min^{-1} during 'dry' storms. Flux returned to 'dry' values at site D about one hour after the last recorded rainfall. Saltation stopped after 15:30 when wind speed dropped below 4 m s^{-1}. Sand movement in the interdune corridor was only recorded at site A prior to rainfall although wind speed remained above 4 m s^{-1} until 15:00.

During high magnitude events results were similar. After the onset of rain, saltation was halted completely, but resumed after a short time and flux reached 'dry' values again. An example for such a situation is the storm of 17/3/98 (see figure 3.19, p.65). Here rainfall of 0.4 mm which fell within

10 minutes was sufficient to stop saltation for 30 minutes, wind speeds reaching 6 to 10 m s^{-1}. Afterwards saltation continued unaffected at the dune crest and the interdune sites. In addition to the precipitation during the period of wind speed above u_t on 17/3, 0.6 mm fell in the evening of the previous day (figure 3.17, p.63). This rain event did not affect sand movement on the following day. Saltation started as soon as $u_{0.24}$ reached 4 m s^{-1} in the morning of 17/3 at sites A, B and D.

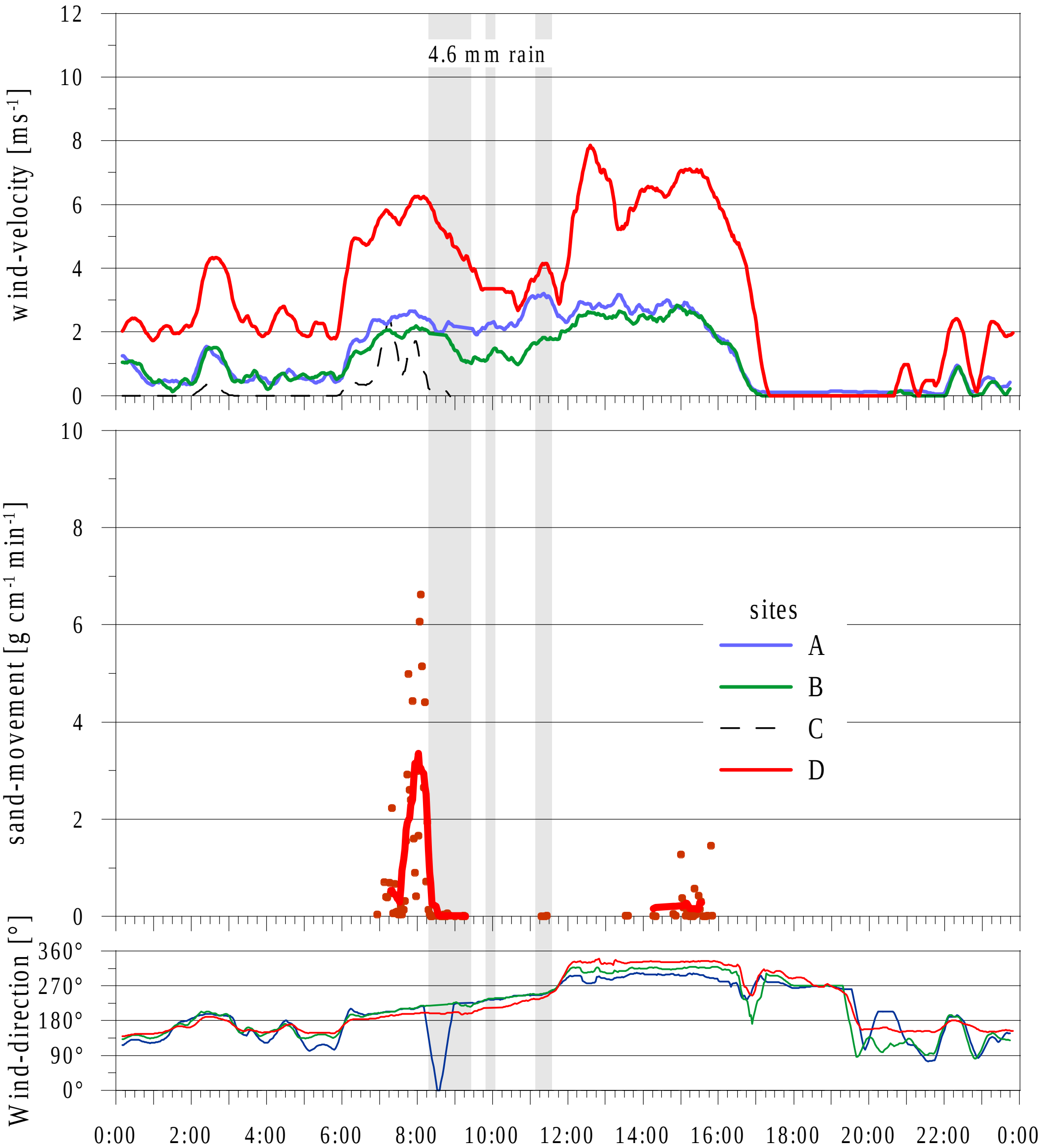

Figure 3.39: Wind conditions, sand transport and rainfall during 5/1/98.

An example of the effect of high amounts of rain falling during a storm is the event recorded on 5/1/98 (figure 3.39). During the morning 4.8 mm of rain fell between 08:20 and 11:33. Only five rain events exceeded this amount within the research period (see table 3.2, p.90). During the storm, sand movement was

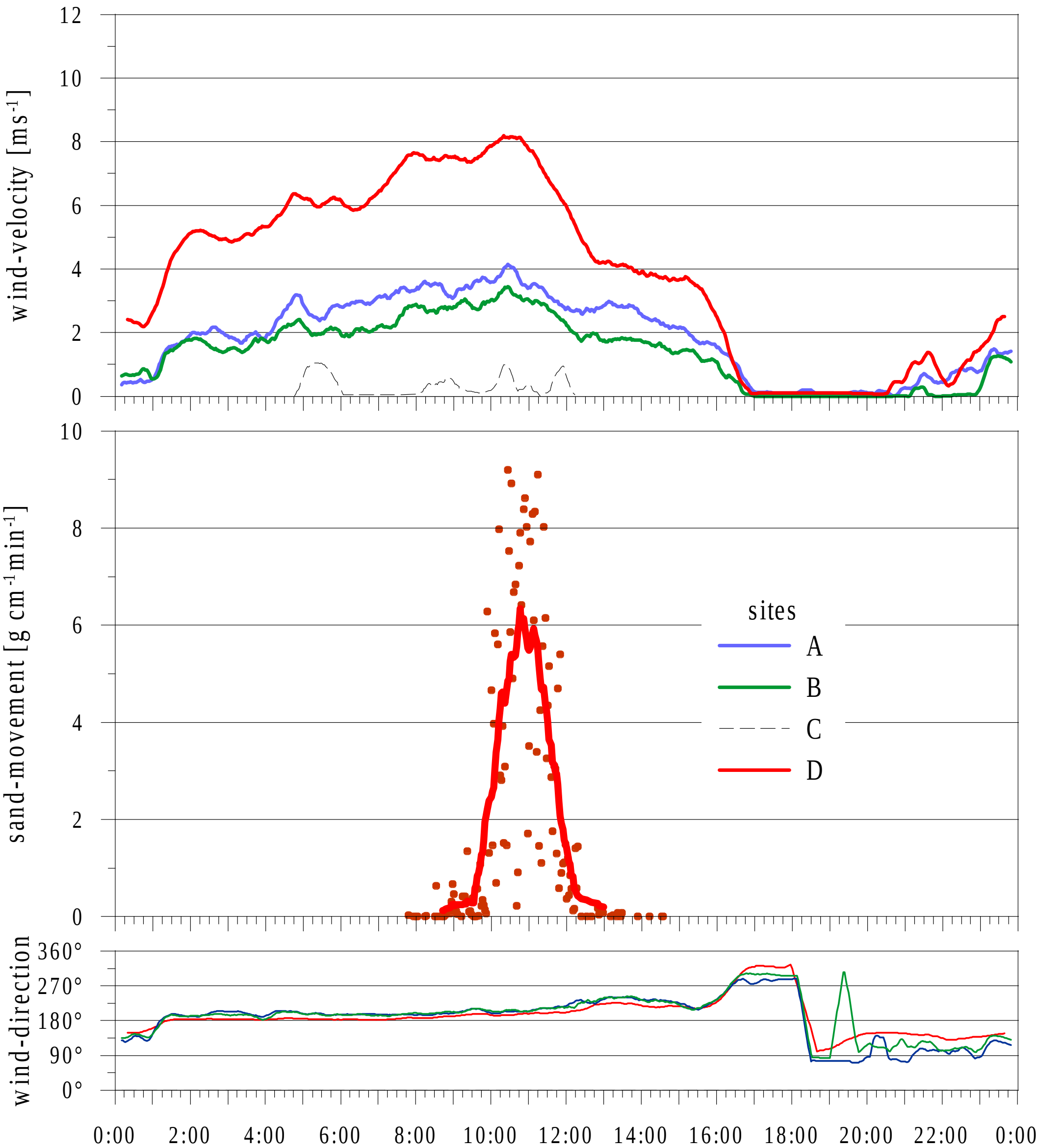

Figure 3.40: Wind conditions and sand transport during 6/1/98.

recorded at site D only. At the interdune sites $u_{0.24}$ did not exceed 4.0 m s^{-1} during the day. Winds were from SSW, reaching a first maximum of 6.6 m s^{-1} at 08:07. This coincided with the maximum of sand transport, immediately before the onset of rainfall. Continuous saltation stopped when rainfall started. Sporadic saltation was recorded during rainfall at site D until $u_{0.24}$ fell below 4.0 m s^{-1}. Sand movement resumed around 15:00, but flux rates remained below pre-rain values despite higher wind speeds. Saltation ended when $u_{0.24}$ dropped below 6.0 m s^{-1}. The high amount of rainfall on 5/1/98 influenced sand transport on the following day. Although $u_{0.24}$ at site D rose above 4.0 m s^{-1} in the early morning hours of 6/1/98, no sand movement was detected until 08:00. At that time $u_{0.24}$ had reached 7.8 m s^{-1}. Normal, 'dry', flux was only measured after 10:00. Saltation continued until $u_{0.24}$ fell below 4.0 m s^{-1}.

The results show that the wetting of the surface by rainfall stops saltation, but that sand movement resumes after a short period even after considerable amounts of precipitation. Observation in the field during rain events revealed a quick drying of the uppermost sand layer over non-crusted surfaces after the end of rainfall. Especially if high wind speeds prevailed. Four days after the event of 5/1/98, on 9/1/98, the uppermost sand layer (0.5 cm) at site D was dry, while below sand was wet up to a depth of 12 cm.

3.4.2.2 High intensity rain

While most of the rain events during the research period yielded less than 3.0 mm (figure 3.37, p.86), rainfall during several exceptional storms reached up to 20.0 mm and intensities of up to 36 mm h^{-1}. After such high intensity rain events, traps at the interdune corridor sites contained sand sized sediment although no wind speed above saltation threshold u_t had been measured since the previous emptying and no saltation had been detected. However, sand grains were found sticking to the outer and inner sides of the sand traps as well as to the anemometers and masts. At the dune crest, sand aggregates were observed in the lee of slipfaces, which disintegrated on drying. Traces of drop impacts were observed on non-crusted surfaces in the interdune corridor. These observations are seen as indicators for rainsplash as an additional cause of ‘aeolian’ sediment movement.

Table 3.2: High intensity rainfall events (depth $\geq$ 3.0 mm) November 1997 - April 1999 sorted by depth.

	precipitation mm	max. intensity mm h^{-1}
11-12/01/98	20,0	18
12/2/98	13,0	18
19-20/2/99	12,0	5
15/2/98	9,8	36
7/2/99	6,5	7
5/1/98	4,8	24
10/2/98	3,9	30

data source: AERC, 1999

Trapped mass during five high intensity rainfall events is shown in table 3.3, p.91. These events are among the seven with the highest total rainfall during the research period (table 3.2). The event on 19-20/2/99 was not used, because sand traps already contained material from the high magnitude storm of 17/2/99. Despite the high intensity on 5/1/98 no measurable amount of sand was transported. During 11/2/98 short gusts above u_t led to saltation at site A,

Table 3.3: Trapped sediment after high intensity rain events at the interdune corridor sites A and B. Values of 5/2/99 for comparison with an average 'dry' event.

	11-12/1/98	10/2/98	12-15/2/98[a]	7/2/99	5/2/99[b]
A/1	33,17	-	7,60	4,05	114,46
A/2	46,46	20,57	10,58	3,10	418,22
A/3	102,65	-	5,86	1,77	1382,30
A/4	94,26	15,75	20,15	1,81	1759,29
A/5	12,84	-	11,46	1,64	0,85
A/6	74,48	11,57	6,98	3,05	879,65
mean[c]	70,2	9,58	10,23	2,76	910,78
B1/1	9,27	14,1	5,83	1,65	1,16
B1/2	12,01	9,7	6,59	1,23	0,97
B1/3	0,76	0,8	2,42	0,77	0,77
B1/4	8,99	13,1	4,75	4,7	-
mean	7,76	9,43	4,9	2,1	0,73
B2/1	8,49	-	6,33	1,35	94,49
B2/2	-	-	7,15	3,39	35,95
B2/3	10,23	-	7,21	1,25	153,59
B2/4	2,65	-	5,59	1,75	82,29
mean	5,34	-	6,57	1,94	91,58

[a] two events
[b] no rain
[c] excluding trap A/5 due to surface conditions similar to site B1

which explains the high amounts of material trapped. The predominance of winds from the W sector is responsible for the generally low amounts recorded at trap B1/3, which was oriented NE.

The highest amount of precipitation during a single storm was measured on 11-12/1/98. The effects of the high intensity rainfall on sand movement is shown in figure 3.41, p.92. Continuous saltation was detected at the dune crest (site D) until the first rain. Single impacts were recorded during the periods of rainfall. These became sparser in the course of the storm. Rainfall was in most cases accompanied by a drop of $u_{0.24}$, yet in general wind speed at site D remained above u_t for dry conditions. Wind speed in the interdune corridor reached u_t only during short peaks. At site A very low flux was detected during 12 minutes prior to the first rainfall, despite values of $u_{0.24}$ above 5 m s^{-1}. Later, wind speeds of 7 m s^{-1} caused slightly higher flux prior to the third rain event.

The data show that rainsplash causes sand transport independent of the surface condition. At site B1, where the surface was crust covered, transport

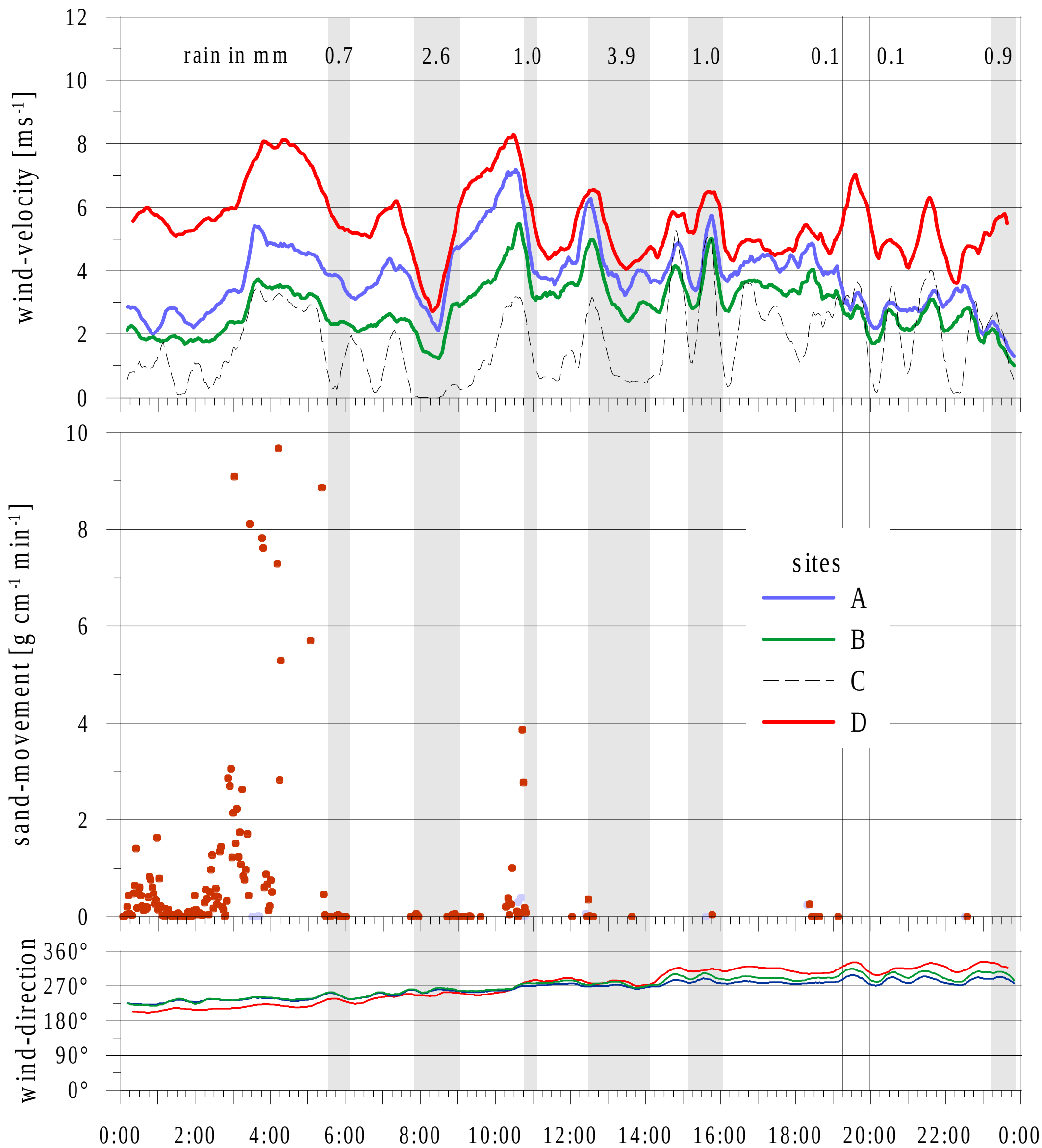

Figure 3.41: Wind conditions, sand transport and rainfall during 11/1/98.

by rain splash exceeds transport by wind during 'dry' storms (table 3.3). The relative amount of total sand transport caused by rainsplash for sites A and B is shown in table 3.4, p.93. It shows that over crust covered areas, rainsplash is responsible for more than 20 per cent of the total amount of material moved. While over non-crusted surfaces the mass is similar, the quota is negligible.

Table 3.4: Transport modes at sites A and B in per cent of total mass caught in the respective trap between November 1997 and April 1999.

site A	A/1	A/2	A/3	A/4	A/5[a]	A/6	mean[b]
splash[c]	0,2	0,2	0,2	0,3	36,1	0,2	0,2
tempest[d]	76,7	68,7	57,9	60,3	47,8	61,0	64,9
sum	76,9	68,9	58,1	60,6	83,9	61,2	65,2

site B1	B1/1	B1/2	B1/3	B1/4	mean
splash[c]	40,9	12,9	16,6	26,3	24,2
tempest[d]	36,5	76,5	40,2	61,7	53,7
sum	77,3	89,4	56,9	88,0	77,9

site B2	B2/1	B2/2	B2/3	B2/4	mean
splash[c]	0,8	2,2	1,4	1,4	1,5
tempest[d]	41,3	39,2	19,4	37,5	34,4
sum	42,1	41,4	20,8	35,8	36,8

[a]upwind surface conditions similar to B1
[b]excluding A/5
[c]5 events on 11-12/1/98, 10/2/98, 12/2/98, 15/2/98, 7/2/99
[d]4 events on 12/12/97, 15-16/3/98, 19/4/98, 17/2/99

3.4.3 Vascular Vegetation

The influence of vascular vegetation on sand movement in the interdune corridor is reflected in the results of site A and site B2. At both sites, the surface crust had been removed in order to rule out any influence of this feature. Both sites were set up in the centre of a corridor. Sand transport in the interdune corridors was restricted to dune parallel winds for the majority of storm events. For these winds, no influence of the height of adjacent dunes or the width of the corridor was found in the analysis of S_z values. Thus the significant difference between the two sites is the vegetation cover. The total amount of sand transported per unit width at sites A and B2 is shown in table 3.5. Values are based on *saltiphone* recordings and material caught in the respective traps.

Table 3.5: Measured total sand transport at the interdune corridor sites A and B2 during the research period.

	before 23/1/99		after 23/1/99	
trap	kg cm^{-1}	% of total	kg cm^{-1}	% of total
A/1	14.43		1.41	
A/2	16.83		1.64	
A/3	19.53		1.90	
A/4	15.95		1.55	
A/6	19.30		1.88	
A (Ø)	17.21	91.1	1.68	8.9
	vegetation cover 17%		vegetation cover 9%	
B2/1	2.54		0.59	
B2/2	0.21		0.05	
B2/3	1.71		0.40	
B2/4	0.31		0.07	
B2 (Ø)	1.19	81.6	0.27	18.4
% of A	6.9		16.1	

The position of a trap within the plot at site A (see figure 2.3, p.25) had a significant influence on the amount of sand transport it recorded. Transported material increased from the edges of the cleared area towards the center and towards the downwind (NE) end. Traps A/3 and A/6 recorded similar amounts, which indicates that the upwind (SW) fetch of A/3 was large enough to allow for the saturation of the airstream during saltation events. Thus the results of traps A/3 and A/6 are regarded as representative for the non-vegetated interdune corridors prior to 1982. Trap A/5 at the upwind end of the plot has been omitted as the amount caught during the research period was too low to be used for a calculation of flux values. The 'raw' mass collected in this trap amounts to 0.15 per cent of the mass caught at traps A/3 and A/6.

The differences between the traps at site B2 are more pronounced than at site A. The reason is seen in the position of the traps relative to vegetation patches, as their distribution (see figure 2.5, p.26) plays an important role for the small scale variations of sand flux. The mean value calculated from the results of all four traps at site B2 is regarded as the representative sand flux at vegetated interdune corridor sites.

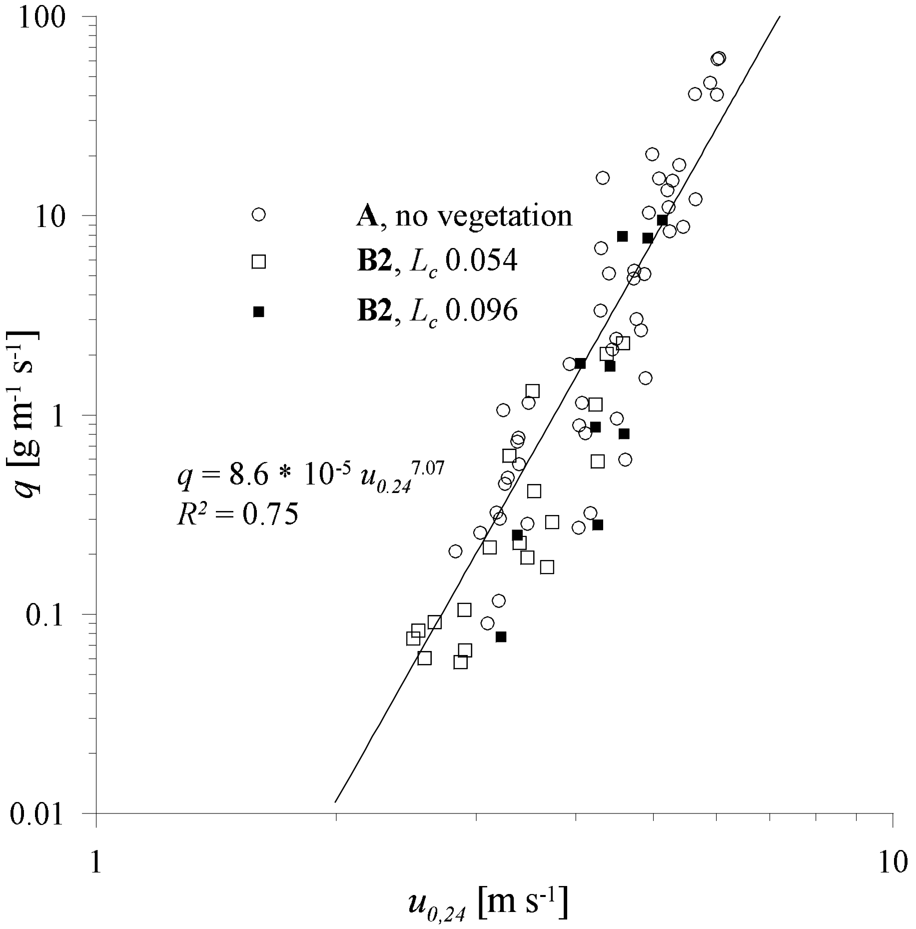

Figure 3.42: Relation between measured mean sediment flux (q) and mean windspeed ($u_{0.24}$) during saltation events at the interdune corridor sites A and B2.

Sand flux q is dependent on near surface windspeed. For transport events during the research period the relation between $u_{0.24}$ and average sand flux at sites A and B2 is shown in figure 3.42. It shows that if wind speed is measured within the vegetation canopy, the relation is independent of vegetation density.

The reduction of sand transport over vegetated surfaces is caused by the extraction of momentum from the airflow. To quantifiy this effect, the wind speed at different heights over the vegetationless surface of site A was set in relation to the corresponding values of site B2. The results are shown in figure 3.43, p.96 for different vegetation densities during saltation. Mean speed has been calculated at each site for the actual period during which saltation was recorded. Above the vegetation canopy, at 1.18 m and 1.77 m above the sur-

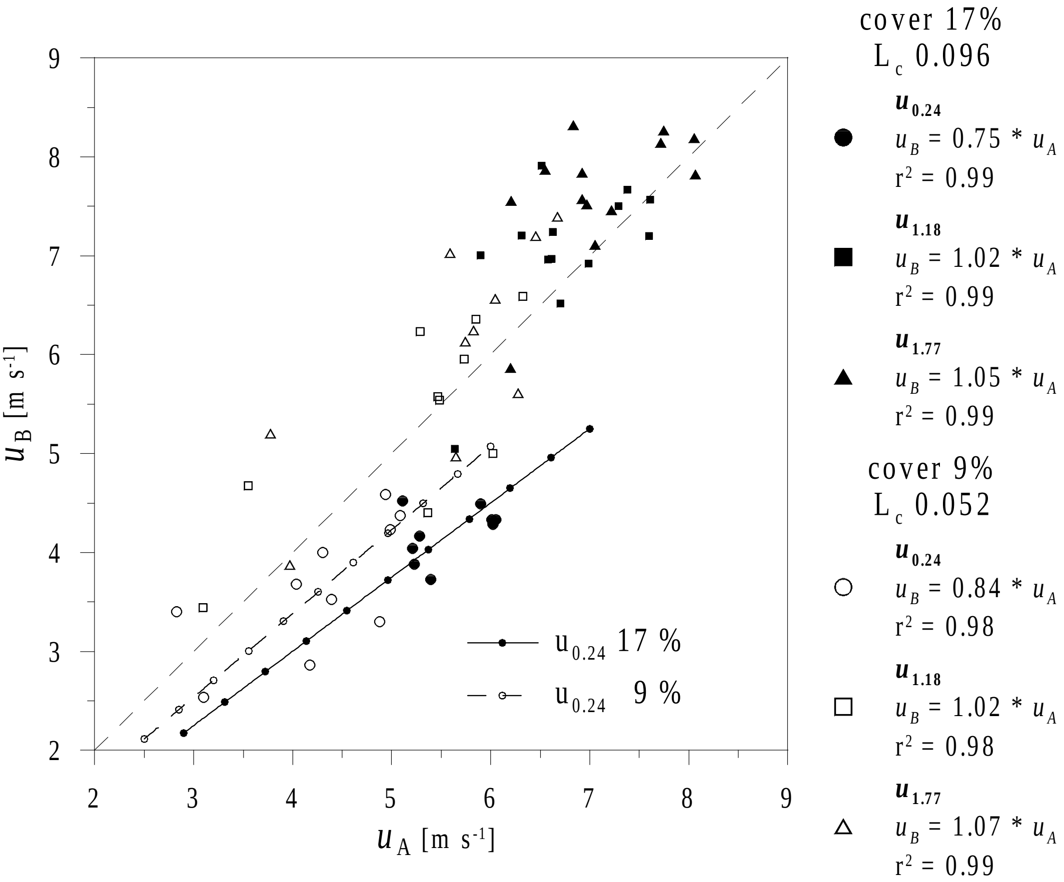

Figure 3.43: Relation of average wind speed during saltation at the interdune corridor sites A and B2.

face, the relation between sites A and B2 is close to 1, independent of vegetation cover at site B2. Wind speed being generally higher at site B2. Within the vegetation canopy at 0.24 m above the surface, the relation is strongly dependent on vegetation cover. Natural vegetation cover of 17 per cent (L_c 0.096) reduced average speed to 75 per cent of windspeed over non-vegetated surfaces. When average $u_{0.24}$ at site A was below 5 m s^{-1} no sand movement was recorded at the vegetated site B2. After the vegetation cover at B2 had been reduced to 9 per cent (L_c 0.052) average $u_{0.24}$ of 3 m s^{-1} at site A was sufficient to cause saltation at site B2. Wind speed within the vegetation canopy was reduced to 84 per cent of the free flow velocity.

As flux q is dependent on speed u, lower wind speeds lead to lower flux and thus to lower total transport. Figure 3.44, p.97 shows the effect of different vegetation cover on the transported mass per unit width during selected events. Under 17 per cent cover, transported mass at site B2 is about 1.0 per cent of the transported mass at site A during average winter storms. The value is significantly higher during exceptional events (15/3-17/3/98, 19/3/98, 16/4/98) reaching 8.5 per cent. A reduction of the vegetation cover to 9 per cent led

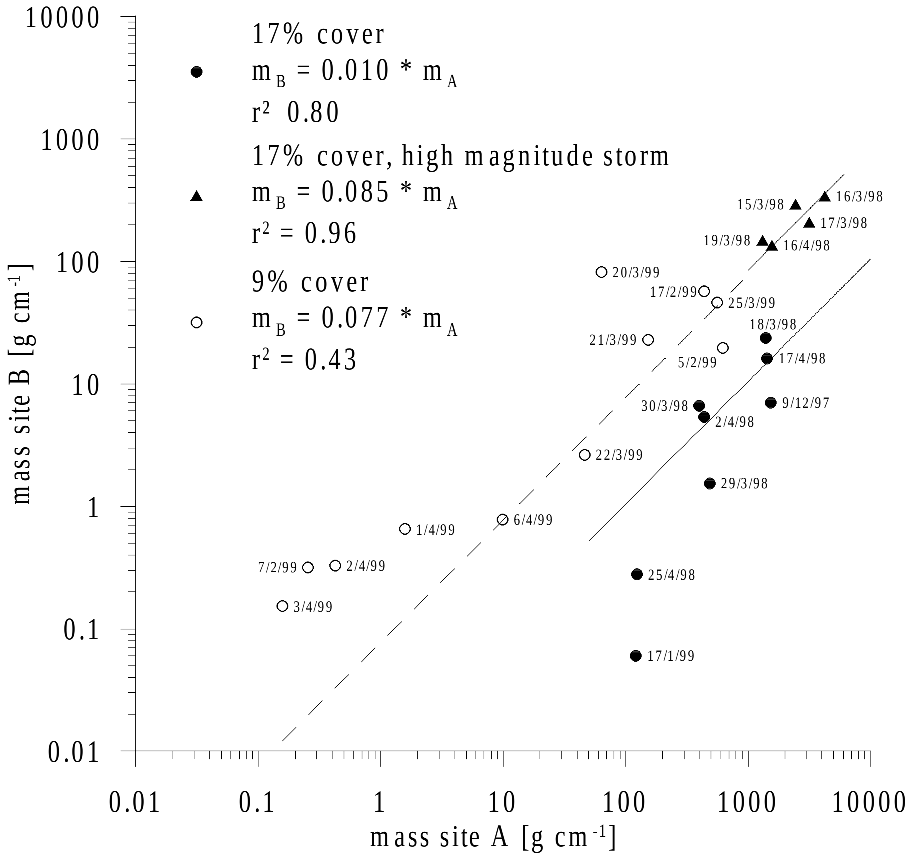

Figure 3.44: Relation of transported mass at sites A and B2.

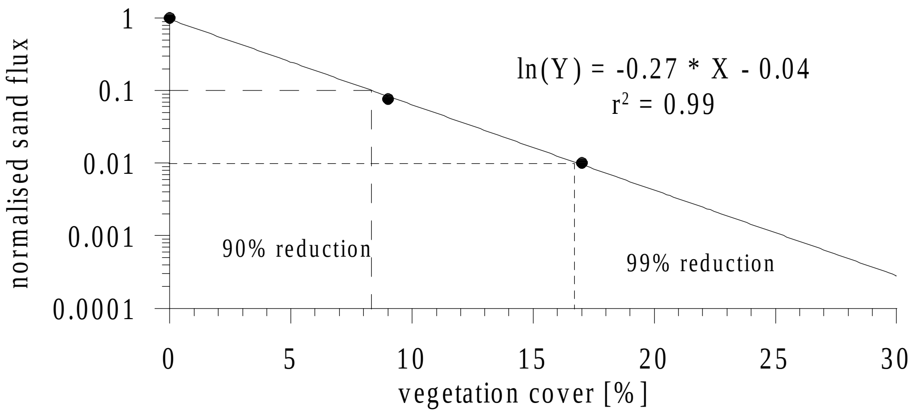

Figure 3.45: Relation between plant cover and sand transport in the interdune corridor at sites A and B2.

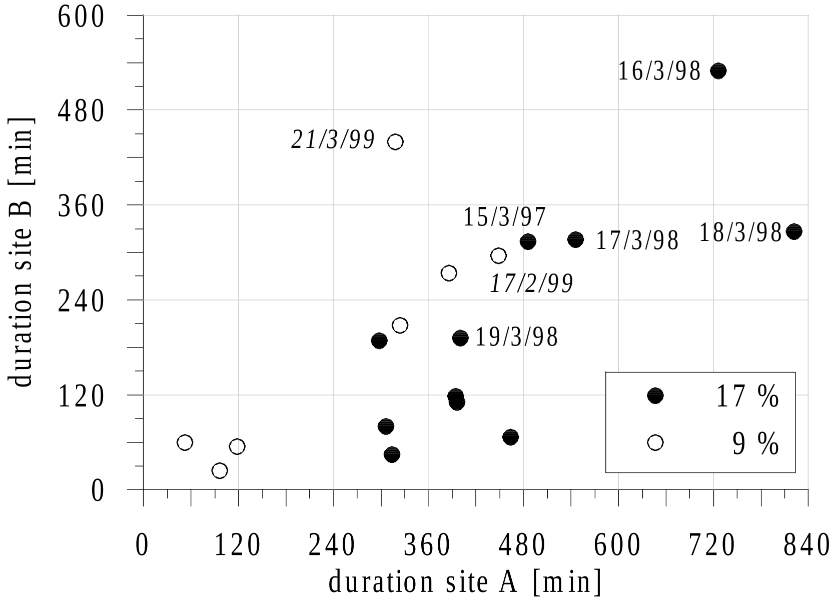

Figure 3.46: Duration of saltation events at sites A and B2 for differing vegetation cover.

to an increase of the average transported mass at site B2 to 7.7 per cent of the transported mass at site A. Exceptional high magnitude events as those during March 1998 have not been recorded after the reduction of the vegetation cover. Therefore no measured flux data of such events under reduced cover is available. Assuming the same relation for flux between average storms and exceptional storms as under 17 per cent cover, transport at site B2 would reach above 65 per cent of site A under the reduced cover. Using the results of site A (0 per cent cover, 100 per cent transport) and site B2 during 'normal' cyclonic storms, the effect of vegetation cover on sand transport in the interdune corridor is well described by an exponential function as shown in figure 3.45, p.97.

Decreasing vegetation cover led to an increase of the duration of saltation events at site B2 compared to the duration of events at site A (figure 3.46, p.98). In general, events at site A were longer than at site B2, an exception is the event recorded on 21/3/99 (see section 3.4.1.2, p.56) and a short event recorded in early spring on 1/4/99. The ratio during the exceptional events in March 1998, under 17 per cent cover is similar to the values for average cyclonic storms under reduced vegetation cover.

Results of site B1 and trap A/5 indicate that sand transport over undisturbed, crust covered interdune surfaces is negligible (see also below, p.99). Therefore all transported material at sites A and B2 originated within the disturbed area. For site A the amount of deflation has been calculated based on the results of traps A/3 and A/6. Based on the length of the cleared rectangle of 47 m and the measured volume of approximately 13.400 cm^3 cm^{-1}

at trap A/6 during a period of 521 days, an average deflation of 1.96 cm per year was estimated for non-vegetated, non-crusted interdune corridor surfaces. Considering that trap A/3 with a smaller fetch of approximately 32 m yielded the same volume as trap A/6, estimated average deflation is 2.86 cm per year. Corresponding values for site B2 have been calculated by relating mean values of site B2 to the average of traps A/3 and A/6 (table 3.5, p.94). Average transported mass during average storms at site B2 prior to 23/1/99 was 0.9 per cent of traps A3 and A6, thus a deflation of 0.025 cm a^{-1} can be expected for 17 per cent vegetation cover. Taking the high magnitude events of spring 1998 into account, deflation at site B2 rises to 0.18 cm a^{-1}, that is 6.2 per cent of site A Under 9 per cent cover average transport rose to 12.2 per cent of site A, resulting in 0.35 cm a^{-1} deflation.

The material eroded from the disturbed surface at site A was deposited in the undisturbed area immediately E of the test site. Vegetation cover and structure was similar to the situation at sites B1 and B2. A mound of loose sand was formed just outside the cleared rectangle, partly covering existing vegetation.

3.4.4 Microphytic surface crust

The influence of the microphytic crust on sand transport in the interdune corridor was studied in the field by setting up two similar sites at a close distance of which one was left in its natural condition (B1) while the crust was removed on the other (B2).

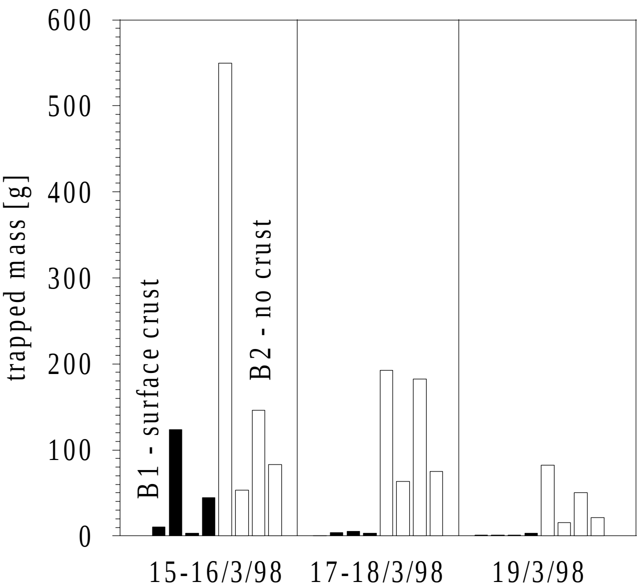

Figure 3.47: Sediment trapped at sites B1 and B2 the storm 15-19/3/98. Values have not been corrected for efficiency differences caused by varying wind directions.

Between November 1997 January 1999 the traps facing W and NW at site B1 collected on average 19 per cent of corresponding traps at site B2. Between January 1999 and April 1999 this value sank to 2 per cent, caused by the increase of sand movement at site B2 due to reduced vegetation cover. Compared to open sand, the combination of vascular vegetation and surface crust led to a reduction of sand movement to 0.29 per cent. Converted to deflation values this means that the possible annual lowering of crust covered interdune surfaces is less than 0.1 mm. The results are based on the mass trapped at site B1 by the *Leatherman*-traps and not on calculated flux, as no saltation was ever recorded by the *saltiphone* installed at the site.

Exceptional amounts of material were trapped at site B1 during the first phase of the storm of 15-19/03/98. Intensive sand transport was recorded in the interdune corridor during the storm at sites A and B2. The temporal distribution of transport at site B1 is remarkable: during the second and third phase of the event, only negligible amounts of sand were trapped at site B1, while the values at sites A and B2 remained high and linked to $u_{0.24}$. Due to malfunction of the *saltiphone*, the flux q at B1 could not be determined, therefore figure 3.47, p.99 shows the amount of sediment trapped. As the principal setup of the sand traps was similar at both sites, the comparison of the values allows an estimation of the intensity of sand movement.

Figure 3.48: Surface crust sample after two years of development, 13/10/99. Note the motorcycle track running across.

The influence of surface crusting on sand transport was determined ex-

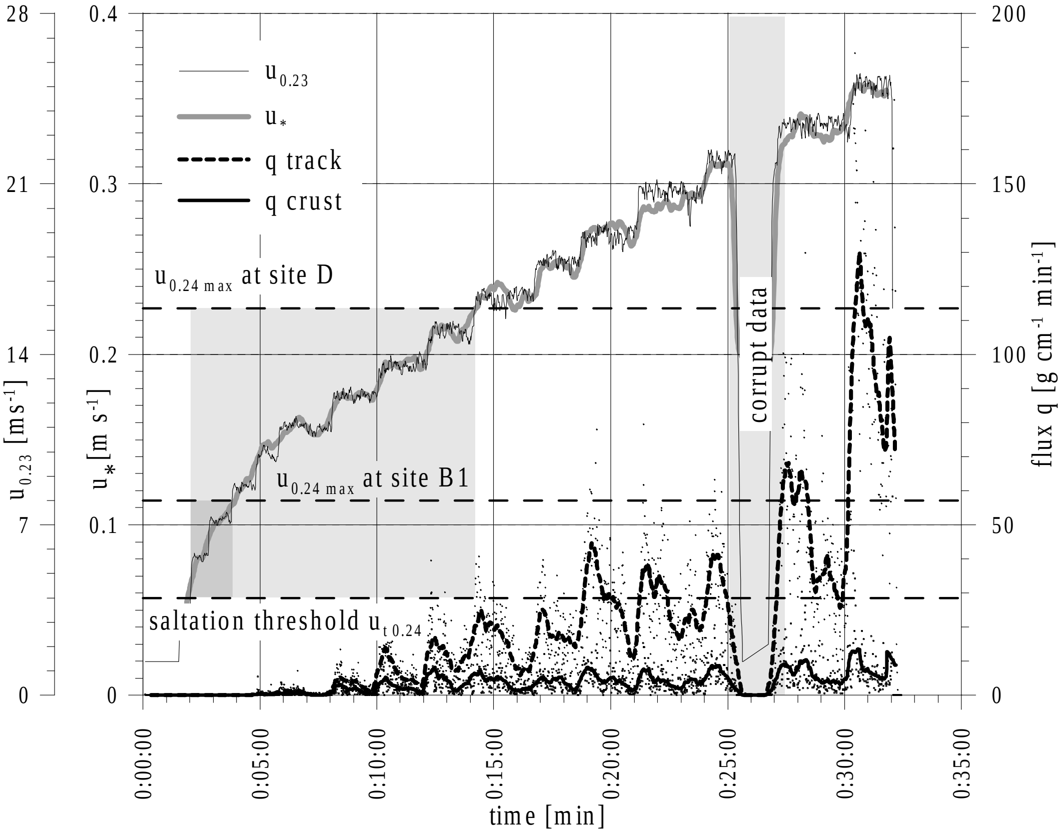

Figure 3.49: Wind tunnel experiment on surface crust strength. For comparison, u_* and $u_{0.23}$ values are plotted. Saltation threshold speed $u_{t0.24}$ determined in the field and measured maximum values of $u_{0.24}$ are indicated. The large grey rectangle envelops conditions measured at site D, the small grey rectangle depicts the range of $u_{0.24}$ measured at site B1.

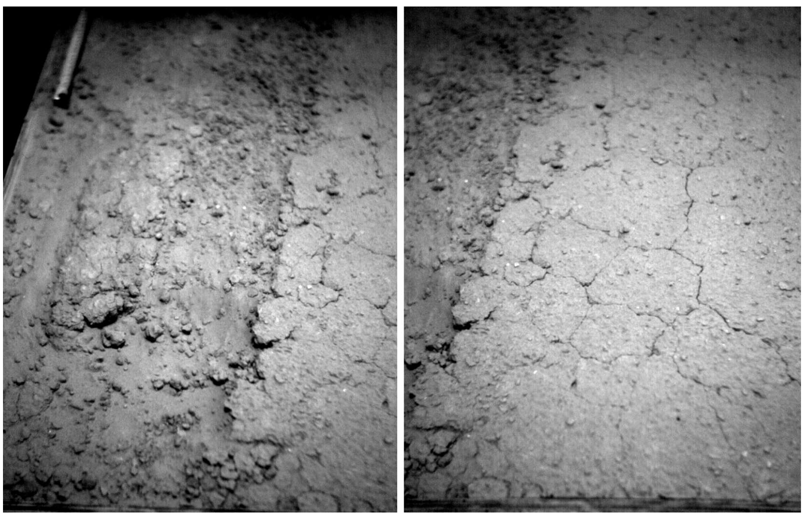

Figure 3.50: The surface crust sample after the experiment. The images show the area displayed in the upper part of figure 3.48.

perimentally conducted in the windtunnel at Ben–Gurion University with an undisturbed crust sample (see section 2.6, p.38). The sample remained undisturbed until sometime during summer 1999, when an unknown motorcyclist managed to leave his track in the frame (see figure 3.48, p.100). What on first sight appeared to be the end of the experiment turned out to be an advantage for the test run in the wind tunnel. As the track was approximately parallel to the long axis of the tray, it was possible to set up instrumentation in such a way that measurements could be undertaken on the disturbed and the undisturbed part of the surface at the same time. The condition of the surface was such, that on the undisturbed part a pattern of cracks caused by the shrinking of the crust upon drying was visible. Small quantities of loose sand were present in these cracks which had a width of approximately 1 mm. The boundary between the wooden tray and the sand surface had been disturbed during the transport from the field to the wind tunnel and was a source of loose sand during the test run. The tyre-track contained aggregates of broken crust and otherwise weakly cemented sand as well as loose sand grains.

During the experiment, windspeed was increased in steps until the frame became airborne and finished the experiment. This sudden end of the experiment occurred when flow velocity had reached 25 m s^{-1}, a value which exceeded maximum windspeed measured in the field during the study period by 10 m s^{-1}. Saltation started at $u_* = 0.14$ or $u_{0.23}$ 10 m s^{-1}. Impacts of saltating grains were recorded at the disturbed and the undisturbed side. Observation during the run showed a sweeping of the small cracks of the undisturbed crust surface. A low but constant number of impacts was recorded on the undisturbed side. On the disturbed side, the loose material in the tyre track was removed first. When $u_* \geq 0.2$, impact numbers increased significantly compared to the undisturbed side. During later stages small but constant amounts of material were eroded from the sides of the track. Aggregates were also moved. During the experiment near surface horizontal flow $u_{0.23}$ and u_* are proportional.

Until the adjustment of the flow in the wind tunnel after the change of the fan speed, both u_* and impact rate dropped. Sharp increases of the impact rate followed the increase of airspeed, respectively u_*. The impact rate over the undisturbed crust remained almost constant over the whole period of the experiment. After the adjustment of the airflow to a steady state, saltation ceases over the undisturbed crust. Over the disturbed area, the impact rate increases slowly up to $u_* = 0.36$, while it rises sharply afterwards. This sharp rise marks the point when pieces of crust were detatched from the sides of the tyre track and larger aggregates inside the track started to move. The impact rates display a regular pattern: A peak during the increase of shear stress and a lower plateau when the flow velocity is steady. The state of the crust sample after the experiment is shown in figure 3.50, p.101. On the undisturbed side the cracks in the crust are visible due to the removal of loose particles. Otherwise the crust remains intact. The motorcycle track has been cleared of all loose material. Signs of erosion on areas bordering the disturbance are visible.

3.5 Changes of morphology at an active dune crest

The effect of aeolian sand movement on the morphology of an active dune crest has been analyzed at different time scales. Measurements of surface change were, whenever possible, carried out immediately after a storm in order to be able to assign the changes to a specific event. To investigate the changes associated with either the summer or winter wind regime, the results have been analyzed for the season. Longer term changes have been identified based on photographs.

3.5.1 Daily changes

In the following, changes between two measurements of the erosion pins at site D are shown. These show ideally the changes caused by a single event or the combined results of consecutive similar events.

3.5.1.1 Average winter storms, 15 - 22/3/99

The changes at site D shown in figure 3.51 were caused by different events between 15/3 and 22/3/1999. They were preceded by a prolonged period dominated by weak NW sea-breezes.

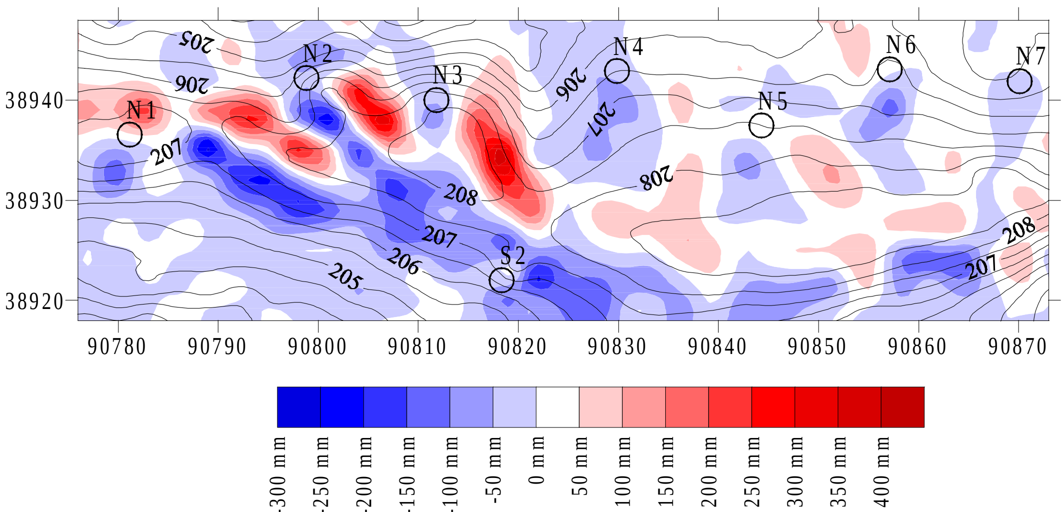

Figure 3.51: Changes at site D between 15/3 and 22/3/99.

The deflation and deposition pattern indicates the predominance of winds from the SW quadrant. The main areas of deflation were located at steeper, SW-oriented slopes and at the upper parts of the dune crest in the western part of the area. These were areas of deposition during the earlier sea-breeze events. In the eastern part, deflation occurred mainly along the upper part of the southfacing slope. Northerly winds as observed during the final phase of the event on 21/3 (see figure 3.13, p.57) caused the deepening of a NW-SW oriented blowout south of N2 and N4. Deposition was concentrated in leeside positions (relative to SW winds!) in the western part, either in the lee of vegetation patches (N1, N2, N3) or at the slipfaces of the crestline. Only minor changes were detected at the eastern part of the area, where vegetation patches were smaller and were located further away from the crestline. The sediment balance of the period was positive.

Table 3.6: Volume changes site D, 15/3 - 22/3/99.

	volume (m^3)	area (m^2)
accumulation	+65.8	1465.41
deflation	-63.6	1446.77
net change	+2.2	

3.5.1.2 High magnitude cyclonic storm, 17/2/99

The changes shown in figure 3.52 were caused by a single high magnitude event (see figures 3.15, p.59 and 3.16, p.60). Main sand transport was caused by winds from S to SW, which is reflected in the deflation and accumulation pattern.

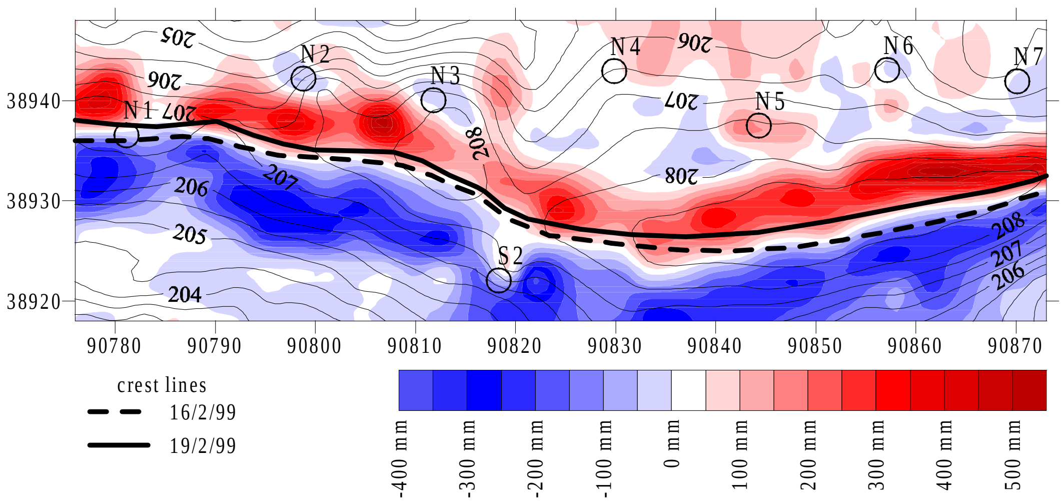

Figure 3.52: Changes at site D between 16/2 and 19/2/99.

Changes of morphology caused by the storm were most significant along the crestline. The crestline was moved northwards. Deflation of up to 400 mm occurred on steep south-facing slopes, average deflation was 113 mm. Areas of low inclination, like the SW–corner of the site, underwent only minor modification. Downwind of plant patch S1, situated on the windward slope, no change

Table 3.7: Volume changes site D, 16/2 - 19/2/99.

	volume (m^3)	area (m^2)
accumulation	+177.3	1705.91
deflation	-140.3	1204.1
net change	+36.9	

was detected, while increased deflation was measured on its eastern side. Accumulation was concentrated immediately north of the crestline, mainly in the form of a slipface along the crestline. Other preferred areas of deposition were the lee sides of plant patches N2, N3 and N4, while at N5 deposition was determined on all sides. The depositional area between N4 and N6 was characterised by scattered sparse vegetation. The mean deposition was 104 mm. On the lee side of the ridge, beyond the crestline, wide areas remained unchanged. The sediment balance of the event was positive (see table 3.7).

3.5.1.3 Very high magnitude cyclonic storm, 15/3 - 19/3/98

Typical for cyclonic storms, the main sand transport was caused by winds from the SW quadrant (see figures 3.17-3.22, p.63ff). Other than during the previous example, a significant amount of transport was caused by dune parallel winds.

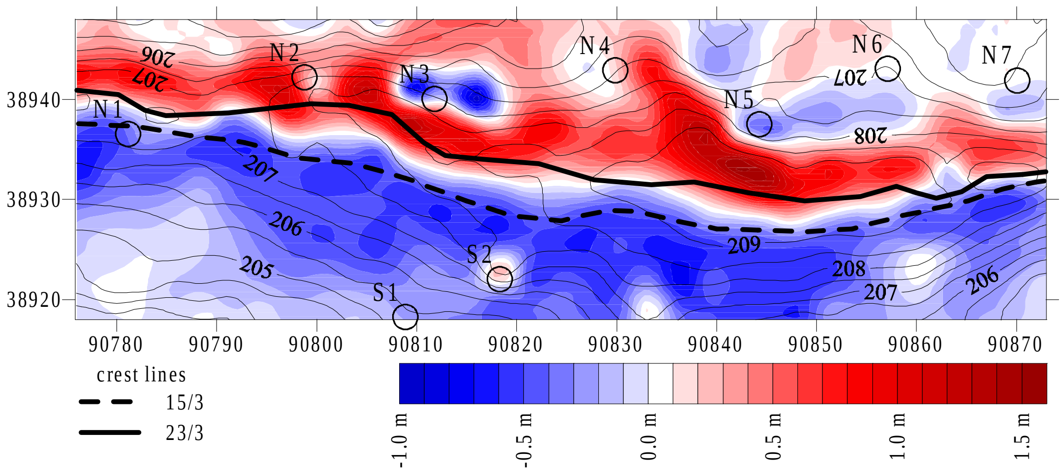

Figure 3.53: Changes at site D between 15/3 and 23/3/98.

Deflation of up to 1.0 m occurred mainly on southfacing slopes along the crest, areas most exposed to the S and SW winds which made up the majority of the storm. Deposition was recorded immediately north of the crest, maximum values reaching 1.5 m. The crestline was moved up to 3 m north. Leeside deflation was measured along the base of the crest in the eastern part of the site. Two areas of only minor deflation were detected: first, the SW part of the site, an area below 206 m and with little inclination, situated immediately upslope of a knickpoint which is marked by denser vegetation. Second, the NE edge, situated in the shadow of the crest and two vegetation patches. In general, the vicinity of vegetation patches was less affected by surface change. An exception was N3, where the slopes of the exposed hummock were eroded. In addition, patch S1 did not survive the storm, it was destroyed completely.

Table 3.8: Volume changes at site D, 15/3 - 23/3/98.

	volume (m^3)	area (m^2)
accumulation	+519.6	1323.0
deflation	-520.1	1629.0
net change	-0.5	

Erosion and deposition of material within the monitored area was balanced. 520 m^3 were eroded and the same amount was deposited. However, during the first two phases the sediment budget was negative, as 146 m^3 of material were missing prior to the final phase on the 19th. This volume was regained by the 23rd during the third phase of the storm and four low magnitude events on the days following.

3.5.1.4 Low intensity winter winds, 12 - 28/1/98

The influence of low magnitude winter winds was restricted to the leeward vicinity of the dune crest and the upper windward slope. Local diurnal winds affected mainly the SE corner of the area (Figure 3.54, p.107). The most intensive deflation was measured at the upper part of the steepest slopes facing S to SE in the eastern part of the monitored area, reflecting the main wind directions during transport. Material was accumulated immediately downwind on the N-facing slipfaces of the crest. Changes were only minor towards the western part of the site, where the angle of the sand transporting winds towards the crest was low. No change was recorded in the NE region and in the SW corner. Accumulation and deflation (table 3.9, p.106) between 12/1 and 23/1/98 amounted to about 30 per cent of the volume caused by a single storm event on 17/2/99 (see figure 3.52, p.104 and table 3.7, p.105). The sediment balance of the period was positive, 7.1 m^3 were added to the site.

Table 3.9: Volume changes site D, 12/1 - 28/1/98.

	12/1-23/1/98		23/1-28/1/98	
	volume (m^3)	area (m^2)	volume (m^3)	area (m^2)
accumulation	+60.0	1125.49	+78.3	1248.5
deflation	-52.9	1789.26	-88.8	1661.5
net change	+7.1		-10.5	

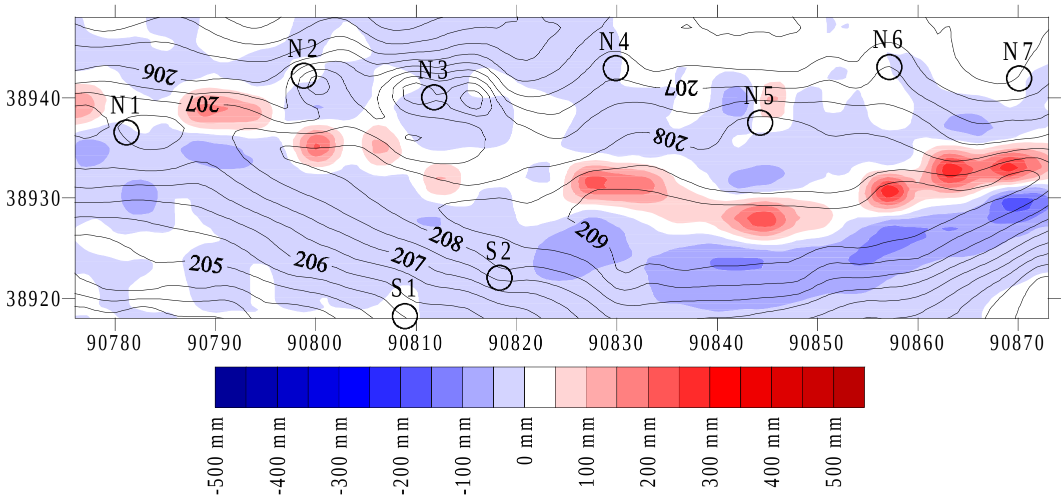

Figure 3.54: Changes at site D between 12/1 and 23/1/98.

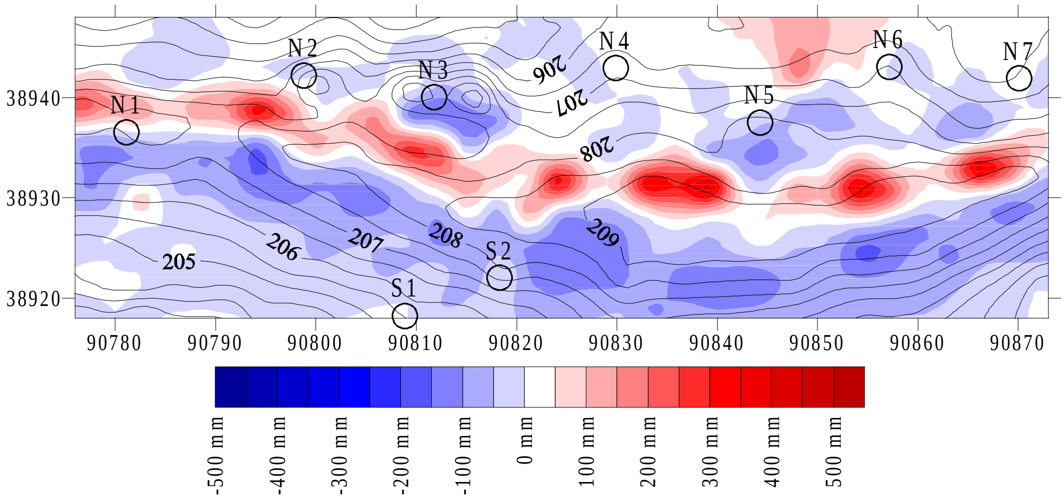

Figure 3.55: Changes at site D between 23/1 and 28/1/98.

Changes between 23/1 and 28/1/1998 resulted from four similar events on the 23, 24, 25 and 28/3. The deposition areas were concentrated along the northern side of the dune crest. The main areas of deflation were the upper parts of the southfacing slopes along the crestline. The distribution along the crestline was more balanced than during the preceeding days, caused by longer periods of sand movement and a wider directional range of the respective winds. Additional deflation occurred north of the crestline, concentrated between hummock N3 and the depositional slipface to the south of it. Secondary leeside flow parallel to the dune is held responsible for this phenomenon. Again, northfacing slopes located further away from the crest and the SW corner of the area were not affected by changes. Sediment balance was negative, 10.5 m^3 were lost during the period.

3.5.1.5 Summer sea-breeze

Changes during summer were more regular than during winter, because the wind regime was steadier. Two examples of changes caused by the sea-breeze are shown to illustrate their nature.

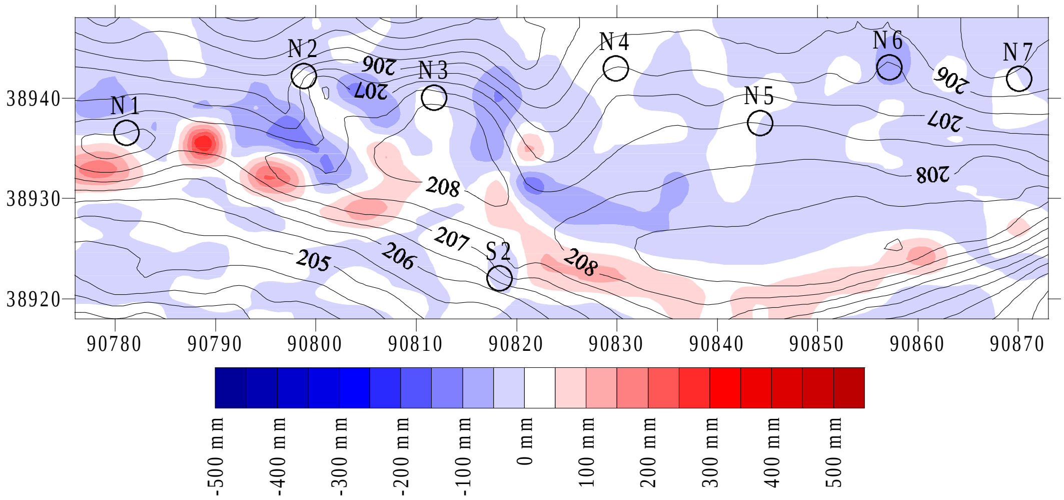

Figure 3.56: Changes at site D between 18/4 and 20/4/99.

Figure 3.56 shows the changes of two days during early summer. The events on 18/4 and 19/4 (see p.79) caused a net loss of 3.1 m^3 of sand. Erosion was concentrated at the gaps between hummocks N1, N2 and N3 as well as the eastern flank of N3. Material was deposited at the upper part of the southfacing slope of the crestline. Only minor changes were observed on low sloping surfaces south of the crestline and over wide areas of the northfacing part of the monitored area. This pattern was repeated with litte variation until the end of May 1999.

In the course of summer, the deposition zones were moved southward due to the regular wind pattern. The situation in mid–summer is shown in figure 3.57, p.109. During one month, 79.9 m^3 of material were moved out of the monitored area. This equals a daily loss of 2.8 m^3. The distribution of deflation areas at the northern slope is governed by the vegetation patches. Isolines show the disection of the upper slopes. The highest amounts of deflation were measured between the vegetation covered hummocks. The zones most affected in the western part of the site are elongated in NW–SE direction. In the eastern part, where the gradient of the northfacing slope was lower, deflation was less severe. Values were generally lower on the lower parts of the slope. Deposition was measured southward of all hummocks on the northern slope. The main areas of deposition remained, however, the southfacing slipfaces (see figure 3.65, p.115). In the eastern part accumulation is concentrated downwind of the gap betwen N1 and N2. Part of the deposition (east of S2) occurred outside of the pin array.

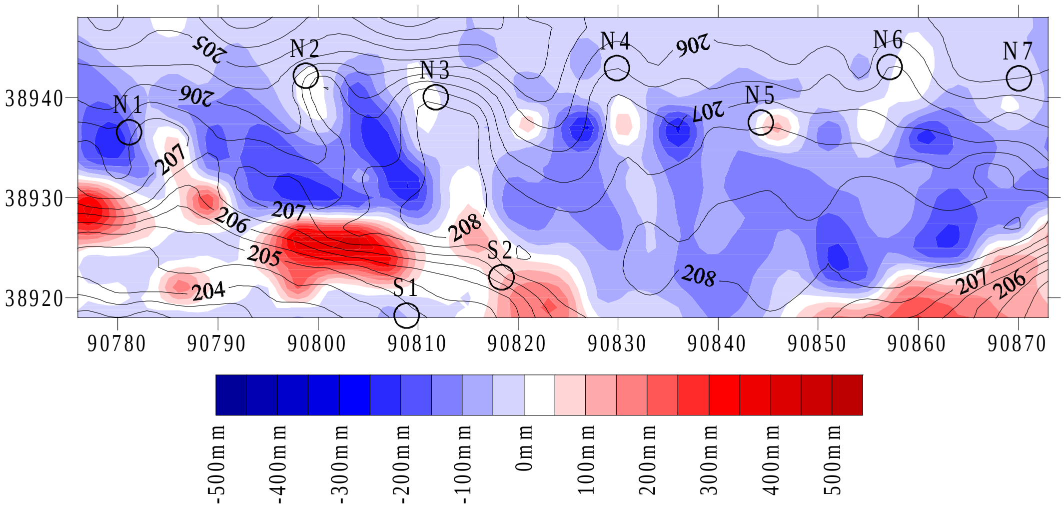

Figure 3.57: Changes at site D between 27/6 and 26/7/99.

Table 3.10: Changes during sea-breeze events in summer 1999.

		total volume m^3			rate m^3 d^{-1}		
	days	accum.	defl.	balance	accum.	defl.	change
18-20/4	2	36.6	39.7	-3.1	18.3	19.9	-1.6
20-25/4	5	38.2	42.8	-4.6	7.6	8.5	-0.9
25/4-5/5	10	92.2	91.6	0.6	9.2	9.2	0.1
5-31/5	26	49.2	153.4	-104.2	1.9	5.9	-4.0
31/5-27/6	27	244.8	309.7	-64.9	9.1	11.5	-2.4
27/6-26/7	29	93.9	173.8	-79.9	3.2	6.0	-2.8
26/7-1/9	37	110.5	148.8	-38.3	3.0	4.0	-1.0
1/9-13/10	42	69.2	152.2	-83.0	1.6	3.6	-2.0
total	178			-377.4			
average					*6.7*	*8.6*	*-1.9*

3.5.2 Seasonal changes during 1997 up to and including 1999

Through aerial photography and ground observation it was known that the active areas at the dune crests undergo periodic changes. These are caused by the seasonal change of the direction of the dominant sand transporting wind (SW in winter, NW during summer). To associate changes of morphology based on erosion pin measurements with the wind regime, seasons were determined based on wind recordings. The beginning of summer is thus marked by the last recorded SW-storm in spring, while the first SW-storm in autumn determines the beginning of winter. The seasons of the research period are shown in table 3.11, p.110.

Table 3.11: Periods of wind regimes during 1997 - 1999.

date	regime	
20/2/97	winter	begin of measurements
18/4/97	summer	
17/11/97	winter	
19/4/98	summer	
14/12/98	winter	
27/3/99	summer	
13/10/99		end of measurements

3.5.2.1 Initial situation 20/2/97 (late winter/early spring)

The erosion pins were put into place on 20/2/97. The situation on this day is shown in figure 3.58. The photograph on p.111 shows the appearance of the dune crest one week earlier. At this time several cyclonic storms had already affected the area. Therefore a crestline with northfacing slipfaces was present at the centre of the crest. The crest is more pronounced on the eastern part of the site, while in the western part, where larger vegetation covered hummocks were present, its course is less clear. All plant patches are isolated features.

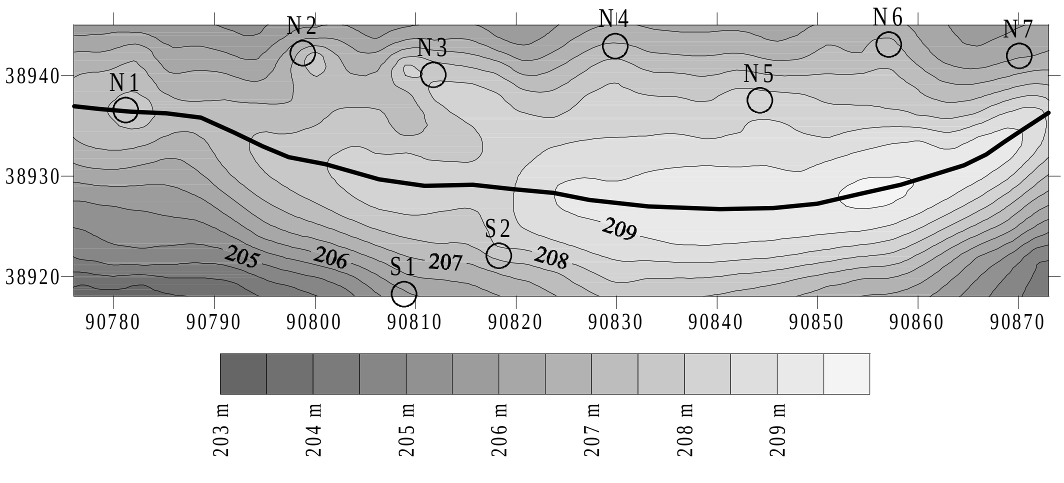

Figure 3.58: Initial situation site D, 20/2/97. The black line indicates the position of the crestline.

3.5.2.2 Spring 1997

Due to the start of measurements in the middle of the season, the results show only part of the changes during winter 1996/97. Cyclonic storms which affected the area after the setup of the erosion pins led to a northward shift of the dune crest with an exception at the eastern end of the monitored area, where a southward movement was detected. The main areas of deflation were

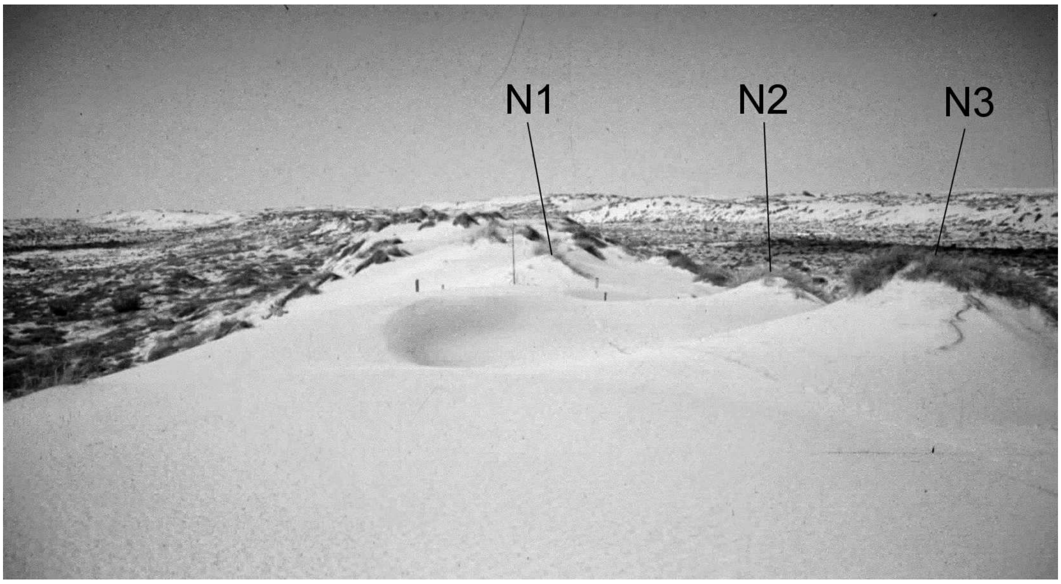

Figure 3.59: Site D, 11/2/97, situation before the beginning of pin measurements. View westward from 90825/38930.

the upper areas of slopes facing SW. Significant erosion occurred also around hummock N3 north of the crestline and at the northfacing slope at the eastern end of the dune crest. Zones of deflation were detected around the majority of vegetation patches. An exception was N1, situated at the crestline. Here deposition occurred north of the vegetation. Deposition was observed along the centre axis of the dune, where a crestline with north-facing slipfaces had formed. Between N1, N2 and N3 the zone of deposition extended onto the northfacing slope. The southward shift of the crest at the eastern end of the ridge was caused by deposition on the SE-facing slipface south of the crestline. Areas at the northern fringe of the crest were affected only marginally by wind action. The sediment balance for the season is negative. During the two-month

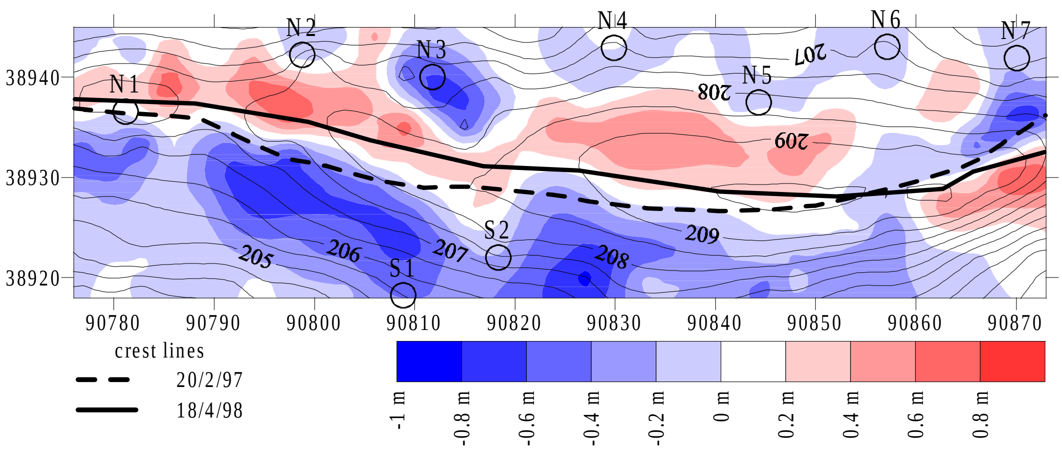

Figure 3.60: Changes at site D during spring 1997.

period 93 m^3 of material were removed from the area under observation.

3.5.2.3 Summer 1997

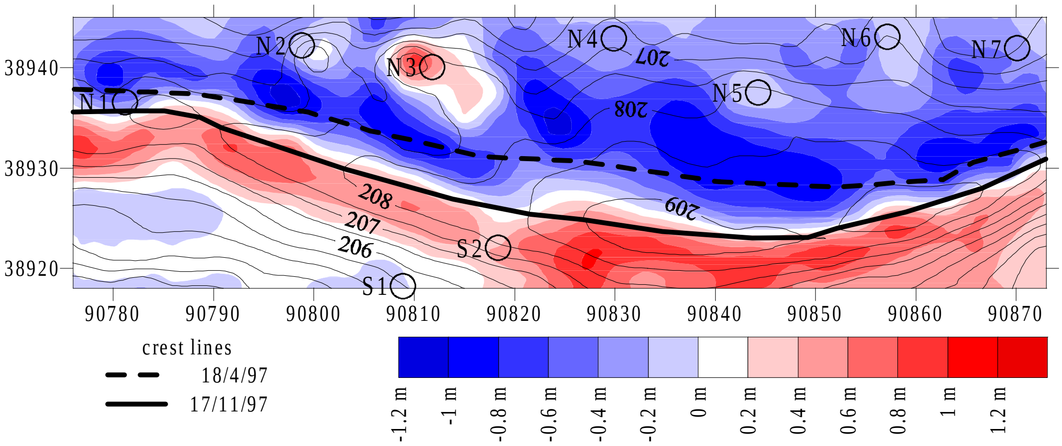

Figure 3.61: Changes at site D during summer 1997.

During summer 1997 the main areas of deflation were located on the northern side of the ridge. Increased values were determined in elongated zones oriented NW-SE south of the gaps between the plant covered hummocks. The amount of deflation increases with height. The SW-corner of the monitored area, which was sheltered by the crestline towards winds from the north and vegetation patches on its southern border, underwent only minor deflation. Vegetation patches on the northfacing slope acted as accretion focus (N3) or reduced deflation in their immediate surroundings (N2, N4-N7). Deposition occurred only on southfacing slopes. The sinuous shaped area of deposition shows the position of the slip faces caused by the northerly winds. As during the preceding winter, the sediment balance was negative. At the end of summer the volume of the sand body was 317 m^3 lower than at the beginning.

3.5.2.4 Winter 1997/98

The situation at the beginning of winter 1997/98 is shown in figure 2.8, p.30. The aerial photograph was taken after the first winter storm and shows clearly the initial crestline and results of southerly winds in the lee of S2, where accumulation was caused by the vegetation.

The distribution of deflation and deposition during winter 1997/98 (figure 3.62, p.113) was a mirror image of the changes during the preceding summer (figure 3.61). The most intensive deflation occurred on all slopes facing southerly directions, the slip-faces of the summer winds. Little protection was offered by litter and dead plant remnants at S2, while S1 was destroyed during the storm on 15/3/98 (see 3.4.1.2, p.62). Most plant patches on the northern side experienced erosion at their base, mainly caused by high magnitude

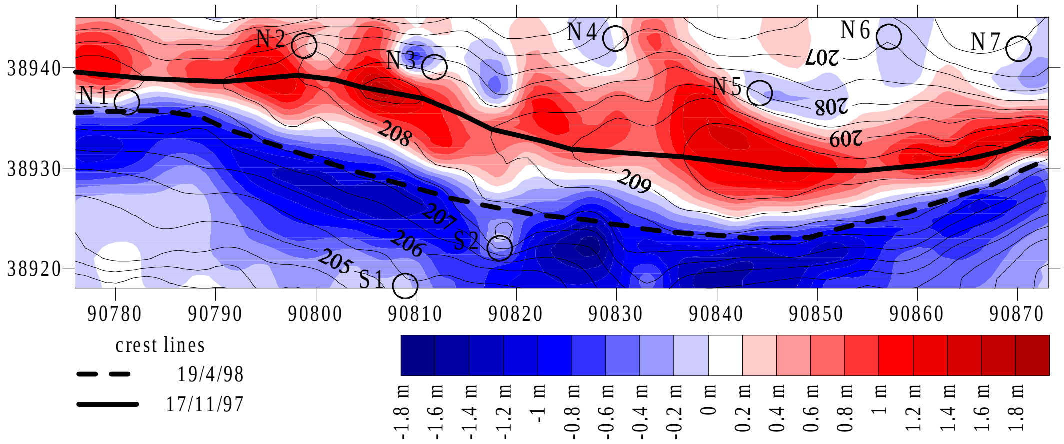

Figure 3.62: Changes at site D during winter 1998.

storms. Especially hummock N3 lost material on its lower slopes, while the feature itself remained unchanged. The NE corner as well as the SW part of the monitored area underwent little change. The NE part was protected by the establishing crestline and the existing vegetation patches. In the SW part, the vegetation at the border between plinth and crest provided protection. Accumulation was concentrated along the crestline, but also occurred on the northern crest area and on the footslopes beyond the monitored area. Areas of highest accumulation values coincide with the deflation centers of the preceding summer.

Figure 3.63: Crestline during winter, view westward, 12/1/98.

Figure 3.63 shows the sharp-edged northfacing crestline in mid-January. By the end of winter the crestline had moved up to N2, visible at the right corner of the image. The major event of this season was the high magnitude storm in early spring which caused substantial changes of the morphology of the dune crest (15/3-19/3/98 see section 3.4.1.2). Again, the sediment budget during the season was negative, but the total loss amounted to only 24 m^3.

3.5.2.5 Summer 1998

The winds during the summer of 1998 caused 1044 m^3 net loss. This amount was unexpected, as it was assumed that major changes would only be caused by the strong winds of the winter storms. Instead the rather low magnitude events of summer caused significant changes. The activity during summer led to the loss of 135 erosion pins through burial or deflation as no visits were made to the field site between May and November 1998. The missing pins were replaced and all pins were resurveyed on 8/12/98 to provide a new base for measurements.

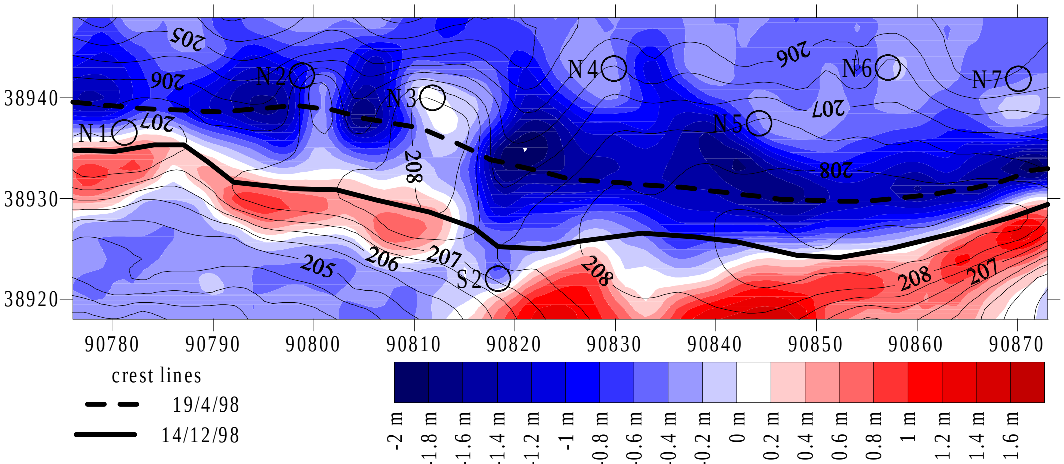

Figure 3.64: Changes at site D during summer 1998.

Deflation by the summer winds occurred on all slopes facing N but reached its maximum values along the centre of the ridge where most of the deposition had taken place during the heavy storms in spring (see figure 3.53, p.105). In the gaps between the hummocks N1 to N3 blowouts reached a depth of up to 1.8 m and in the area east of N3 deflation was up to 2.0 m. Apart from the northern slopes, deflation was also observed at the SW-corner, in the lee of the crest. Deposition was concentrated on the southfacing slopes where slipfaces as shown in figure 3.65, p.115 were established. Considerable amounts of material appear to have been deposited on the slip face reaching to the corridor surface at the southern fringe of the monitored area and beyond. Here the dune ridge advances over a crusted surface of the interdune corridor which borders directly on the active zone (see figure 2.9, p.31). Direct monitoring of this process with erosion pins was restricted to the uppermost part, since repeated access to the

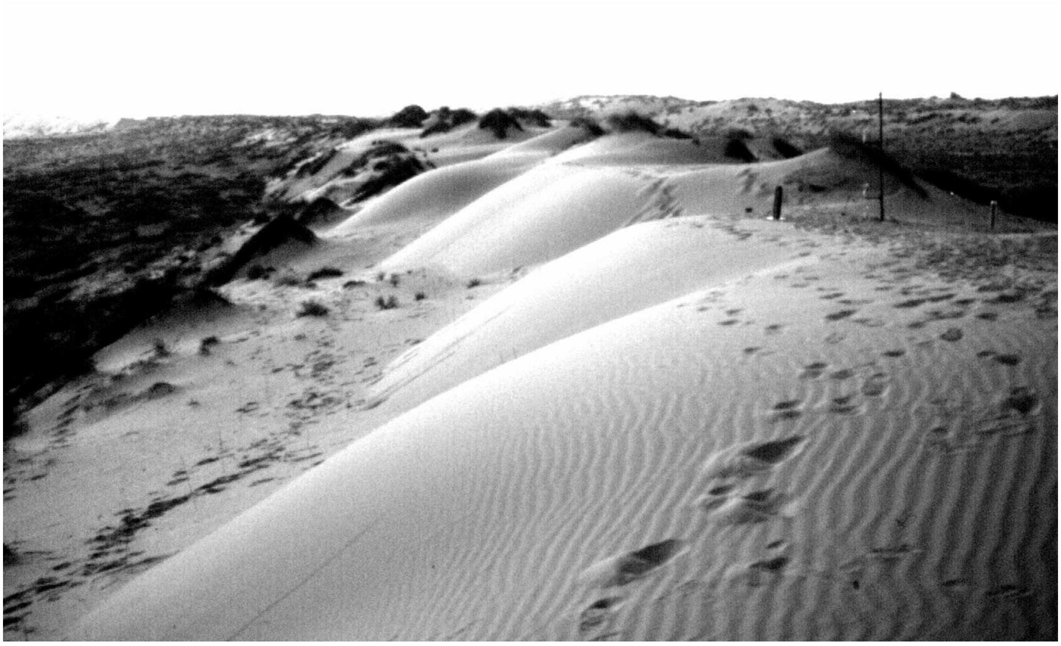

Figure 3.65: Situation at site D at the end of summer, 23/11/98. View W from 90810/38930. South-facing slipfaces advancing over the deflation area in the SW part of site D. See figure 3.68, p.117 for comparison with conditions at the beginning of summer.

slipface itself would have caused avalanching of sand and thereby would have influenced the measurements beyond tolerable limits.

Lower amounts of deflation were measured in the vicinity of the vegetation patches on the northern part of the ridge. Patches N4 to N7 sufferd from erosion at their bases, where roots were unearthed. Hummock N3 remained connected with the main sand body at the crestline.

3.5.2.6 Winter 1998/99

The winter of 1998/99 was characterised by a scarcity of heavy storms and very little rainfall. Dune volume between December 1998 and March 1999 was an increased by 13 m^3. This was the only time during the study period, that the net volume change of the dune crest area was positive. Deflation zones were concentrated on southfacing slopes, mainly at the areas of the slipfaces of the previous summer. Material was deposited in the former blowout areas between hummocks N1, N2 and N3 as well as immediately N of the crestline. Wide areas of the northern part of the site remained unchanged. The same applies to the gently sloped SW-corner between the crestline and the southern rim of the active zone.

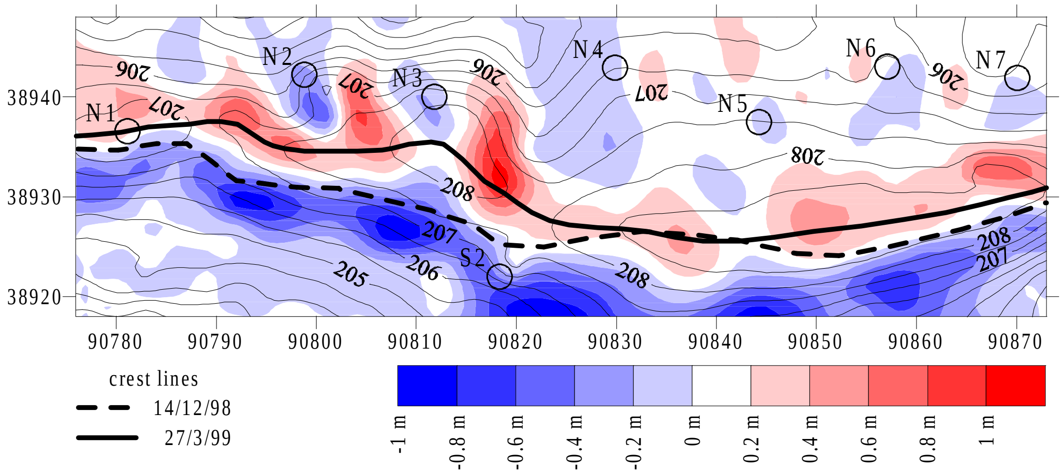

Figure 3.66: Changes at site D during winter 1998/99.

3.5.2.7 Summer 1999

The change of wind regime occurred in late March. The pins were measured weekly until 5/5/99. Afterwards, based on the experience of the previous summer, regular measurements of the erosion pins were carried out every four weeks until October. Due to these field visits it was possible to monitor the changes occurring during summer in more detail (see table 3.10, p.109). It also prevented the loss of pins.

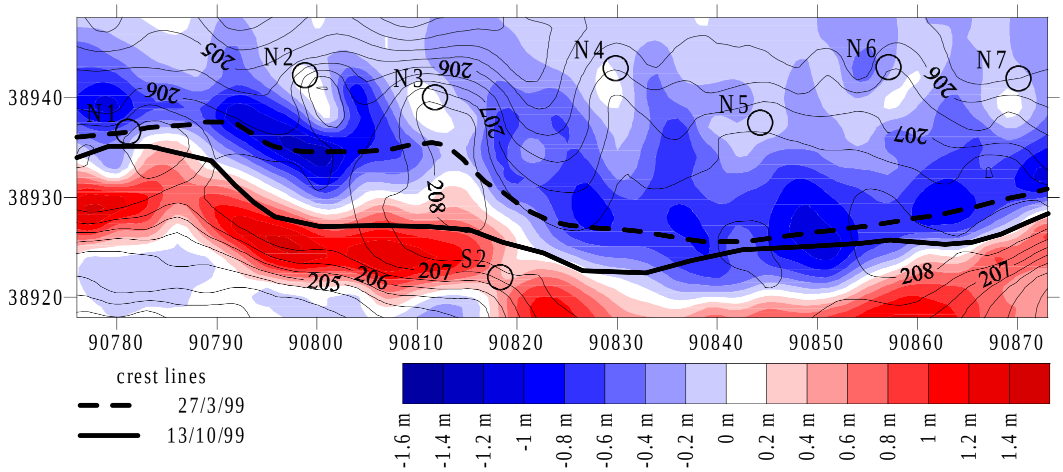

Figure 3.67: Changes at site D during summer 1999.

Little change was observed until early April, as during March the sea breeze developed only in the late afternoon to strengths just above u_t at the crestline. The lower parts of the active zone were not affected by these events. Distortions like footprints made during pin measurements were preserved in these areas for periods of more than a week. The destruction of the winter forms began when the sea breeze developed regularly in the early afternoon to strengths well

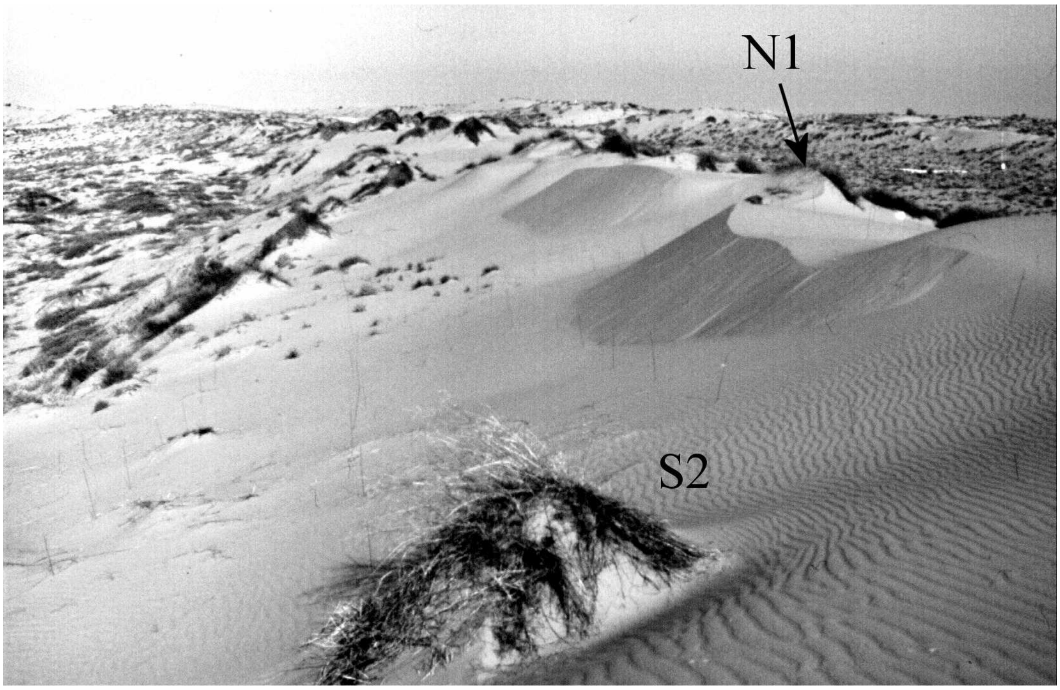

Figure 3.68: Site D at the beginning of summer, 5/5/99. Southfacing slip-faces mark areas of deposition during sea-breeze. See figure 3.65, p.115 for comparison with end of summer conditions.

above saltation threshold. Sand was eroded on the northern and centre part of the crest. The areas between the plant covered hillocks, especially between N1, N2 and N3, underwent severe deflation. The resulting blow-out areas were oriented parallel to the main wind direction, NNW. The vegetation patches themselves on the northern side remained largely unchanged. Deposition took place south of the brink, where slip faces had been formed. At the eastern part of the area, the slip face extended to the interdune corridor and the dune advanced several centimetres across the crust covered surface of the corridor. In the western half, the slip faces covered a blow-out area which underwent continuous erosion during the previous winters.

The result of the summer winds was similar to the previous year (figure 3.64, p.114), but the areas affected by deflation were more restricted to the immediate crest. The zones of deposition are located further south, indicating a net movement of the dune crest. Unfortunately this put the main slipface in the eastern part outside the monitored area. The loss of material amounted to 377.4 m^3 within 178 days, which means that on an average day 1.9 m^3 of sand were eroded from the crest.

3.5.2.8 Total changes, February 1997 - October 1999

The volumes of the total and seasonal changes are shown in figure 3.70, p.118. The calculation of the changes during the measurement period is based on volumes calculated with the pin-coordinates of 20/2/97 and 13/10/99 (965 days). Between these two dates a loss of 1803 m^3 was observed over a surface area of 2668 m^2. The total loss per unit area was 0.67 m^3/m^2. This equals an average loss of 0.26 m^3/m^2 per year, respectively a reduction of average dune height of 0.26 m a^{-1}. This is nine times the amount determined for bare sand surfaces in an interdune corridor (site A, see p.98). The observation revealed a negative linear trend for the total volume of the monitored dune crest area which can be written as

$$V = -1.87 * d \tag{3.2}$$

where V is the volume in m^3 and d is the number of days. Correlation is strong ($r^2 = 0.97$). Average elevation of the monitored area was 207.37 m asl. on 20/2/97, and 206.67 m asl. on 13/10/99. Main losses occurred during summer when northerly winds prevailed. During the winters 1997/98 and 1998/99 the sediment budget of the monitored area has been almost balanced.

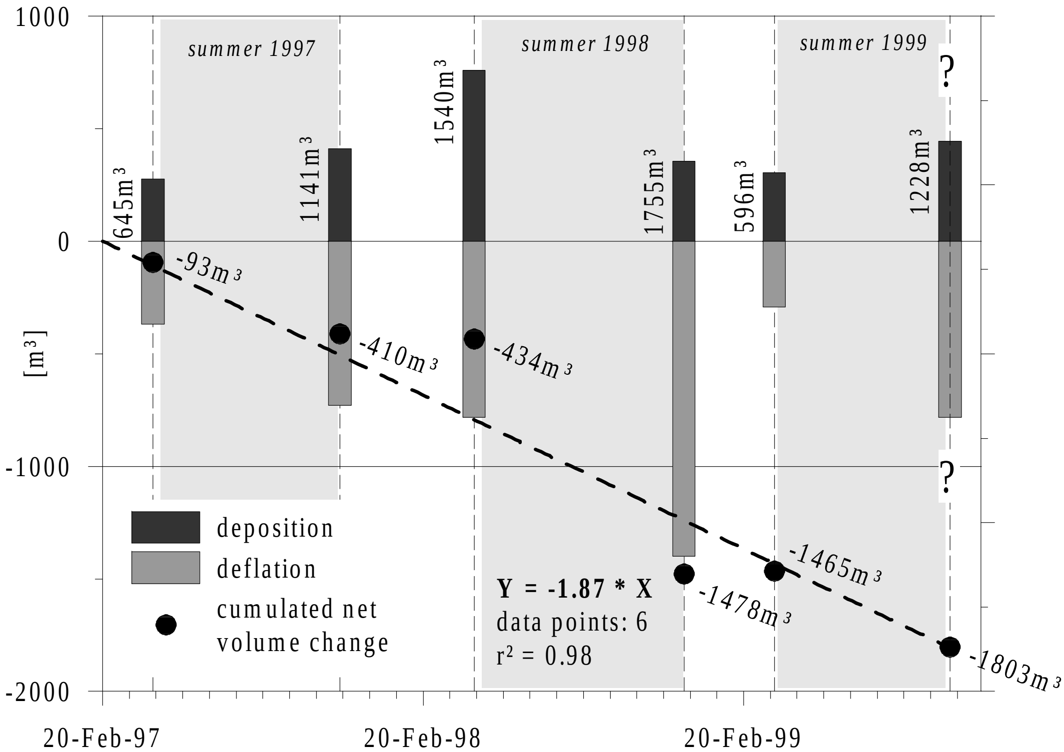

Figure 3.70: Volume balance and aeolian activity at site D between February 1997 and October 1999. The figures at the bars indicate the sum of deflation and accumulation. The end of measurements does not coincide with the end of summer 1999.

The spatial distribution of accumulation and deflation is shown in figure 3.71, p.119. The highest values of deflation were measured east of 90815 along the center of the dune crest and along a N-S oriented area at 90823. These were the areas with the lowest vegetation cover. No vegetation was present

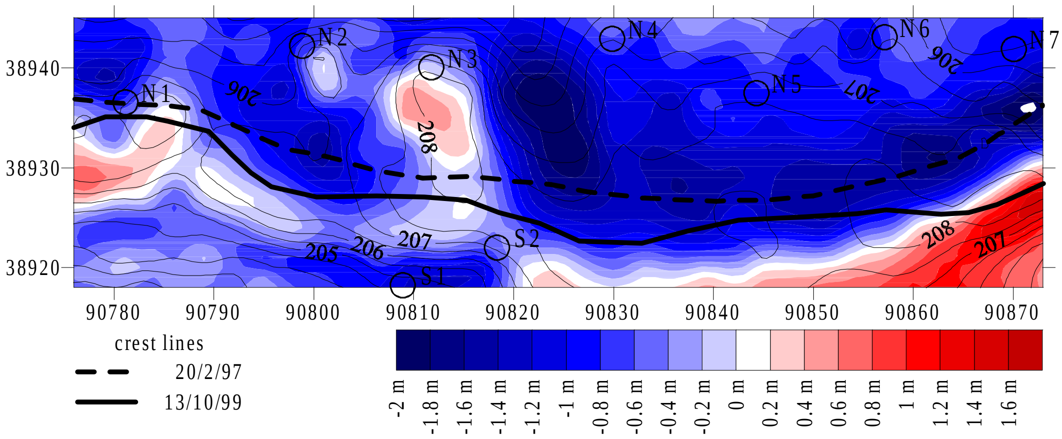

Figure 3.71: Changes at site D, February 1997 - October 1999.

east of 90820 along the crestline, as it was the most active area. West of 90815 the hummocks N1 to N3 were covered with vegetation, offering protection for the surrounding and leading to deposition or reduced deflation of sediment. The accumulation on the S and SE fringe of the monitored area indicate a shift of the dune ridge towards the south.

The rate of change has been calculated for each season based on the total volume of moved material (deflation and deposition), surface area and duration of the respective period (figure 3.69). During summer, the intensity of sand movement was lower than during winter, but due to the longer duration, more sediment was moved (figure 3.70, p.118). The variation is higher for winter, intensity during winter 1999 being only half of spring 1997. The number of high magnitude storms in late winter 1999 was lower than during the previous seasons. The value of spring 1997 may be biased as measurements for the first part of winter are missing.

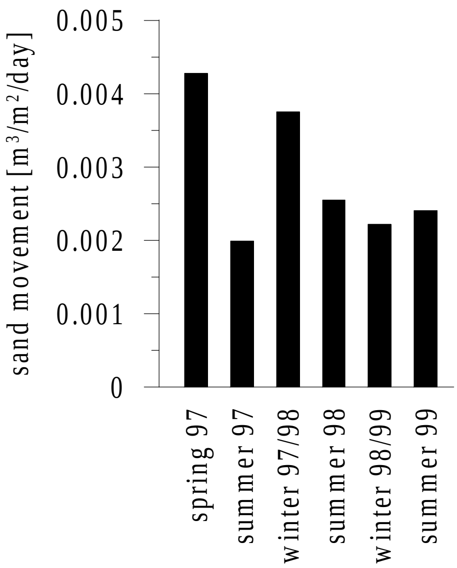

Figure 3.69: Seasonal differences of average daily sand movement.

The position of the crestline at the end of each season is shown in figure 3.72, p.120. Sand movement during winter resulted in a sharp, sinuous crestline, while sand movement during summer built up lobes with southfacing slipfaces, mainly downwind of gaps between plant patches. The amount of lateral change of crest position was dependent on the activity in the preceding season. Crest position at the beginning of measurements in February 1997 was within the centre of the winter zone. Activity during summer 1997 was low

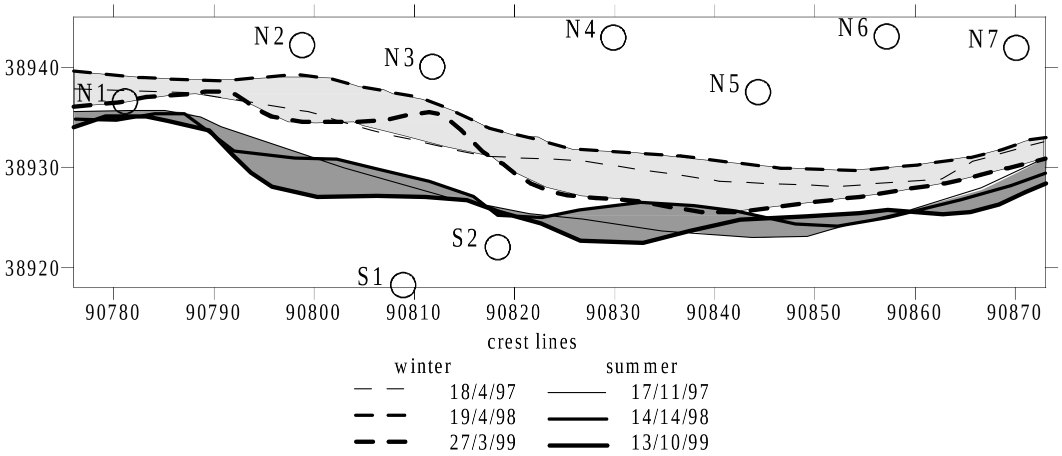

Figure 3.72: Position of the crestlines at the end of each season.

compared to the following summers, therefore the movement of the crestline was only moderate. During winter 1997/98 several heavy storms caused the farthest northward movement during the research period and led to a connection of hummock N3 with the main sand body (figure 3.62, p.113). This connection remained intact until the end of measurements although during summer 1998 both absolute and relative sand movement was exceptionally high (figures 3.64, p.114, 3.70, p.118, 3.69, p.119). The following winter of 1998/99 was short and with the exception of 17/2/99 (see pp.59 and 104) saw only low magnitude cyclonic storms. This resulted in only minor northward movement of the crestline, especially in the center part of the area, which is situated downwind (for SW winds) of plant patch S2 and has a steep south-facing slope (figure 3.66, p.116). At the more exposed slopes in the western part as well as at the eastern end of the area, crestline position in February 1999 resembled the situation of April 1997. During summer 1999 the crestline reached its southernmost position. High values of advancement were recorded downwind of the deflation zones of the western and centre part, where the deposition of the material took place within the monitored area.

Exceptionally low variation of the position of the crest was observed in the vicinity of plant patches N1 and S2 and at the eastern end of the dune ridge. The crestline remained close to N1 during the whole observation period, the plant patch acting as an ‘anchor’. The influence of S2 was restricted to summer. The winter position of the crestline at the eastern end of the ridge was the same throughout the duration of the project. Based on the results of three years of observation it appears that the crestline alternates seasonally between two distinct zones (shaded area in figure 3.72). The results do not allow the assessment of possible net lateral movement of the dune.

3.5.3 Changes 1989 - 1999

The monitoring of surface changes indicate that active parts of the dune ridges are subjected to continued deflation. In order to put these results into a wider temporal and spatial context within the linear dune area of Nizzana, ground and aerial photographs taken during the 10-year period between 1989 and 1999 have been compared with the focus on the development of open sand surfaces and vegetation.

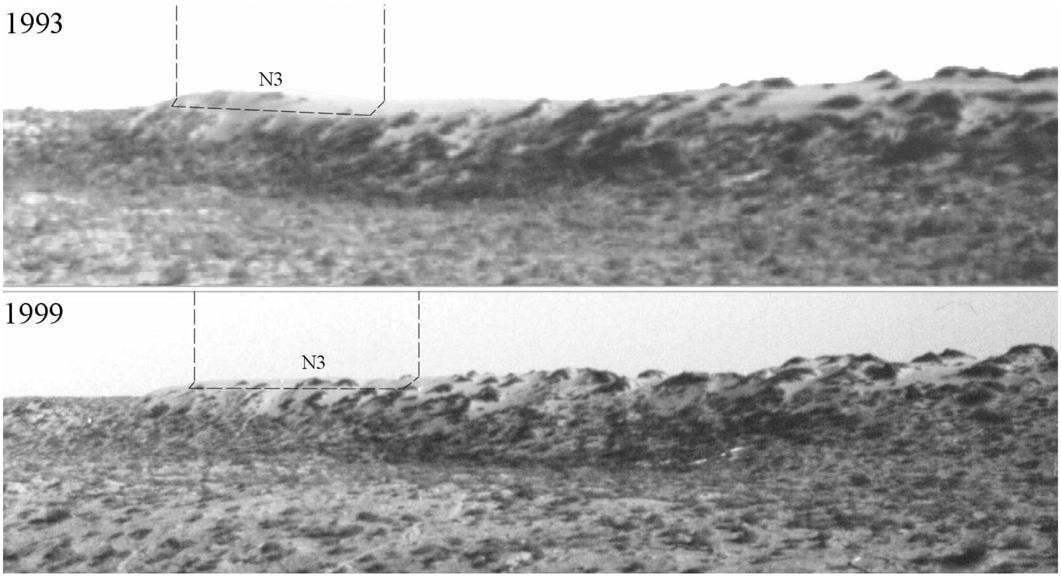

Figure 3.73: Differences between March 1993 and April 1999, area of site D viewed from NW. Vegetation patch N3 is marked and the area monitored during the project period is delimited.

Changes at site D 1993 - 1999 The ground photographs taken from a similar perspective and in the same season in 1993 and 1999 (figure 3.73) show changes at the dune where site D was set up in 1997. The main visible changes since 1993 were an increase of vegetation cover at the dune crest and a reduction of dune height. In 1993, the eastern part of the dune crest was a continuous sand body with only a few patches of vegetation in the area of the later site D. Hummocks topped by vegetation, standing out of the main sand body were located at the dune crest only in the western part of the 1993 picture. The other vegetation appears to be only slightly elevated above the surrounding sand surface. In 1999, this area appears disected, large and deep blowout areas can be recognized between the hummocks. The former continuous crestline with isolated hummocks was replaced by a chaotic assemblance of hummocks. The significant reduction of dune height is most clearly visible at the eastern part. Hummocks N1 to N5 stand out clearly in 1999, while in 1993 only N2, N3 and N4 are visible as vegetation patches integrated into the main sand body. Site D in 1999 resembles the western area in the picture of

1993. In contrast to the crestal area, the footslopes do not appear to have undergone any significant change during this period.

Changes 1989 - 1998, based on aerial photographs In pictures taken in 1989 published in Littman & Gintz (2000, p.80) and Sharon *et al* (2002, p.872) the crests of both ridges are almost free of vegetation. Only on the southern dune, in an area close to the road, was vegetation present. The crests were continuous open sand bodies. In 1992, the area occupied by obviously vegetation free surfaces is reduced (figure 1.2, p.19). The continuous open sand body on the northern dune reached up to the wadi in 1989, in 1992 it was terminated just north of the later site B. Vegetation on the southern ridge increased mainly between the road and approximately 90600, south of site A. This transition area is also visible in the centre of figure 3.73, p.121. From that position on to the termination at Nahal Nizzana, the ridge crest appears only sparsely vegetated in 1992.

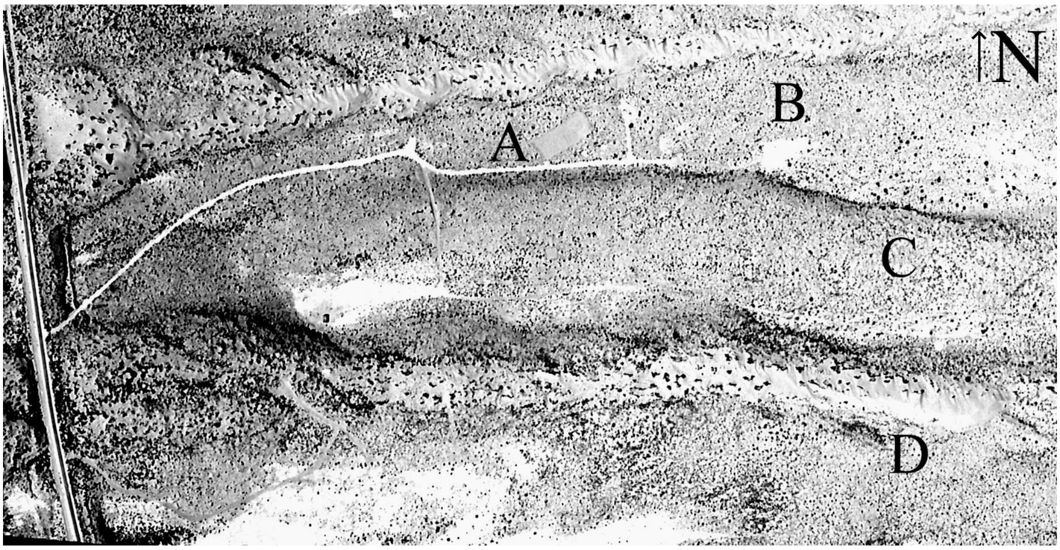

Figure 3.74: Aerial view of the research site in August 1998. Measuring sites are indicated. Note increasing vegetation density at the dune ridge westward of site D.

The aerial photograph in figure 3.74 shows the vicinity of the research plots in August 1998, ie during the second half of summer. Parts of the dune crests show signs of activity: South of vegetation patches accumulation areas stand out. These features are common over most of the northern dune, while on the southern dune they are restricted to the visible eastern half. Vegetation density on the southern dune increases towards the border road. The part of the ridge where site D had been set up, had the lowest visible vegetation cover, resembling the state of the whole ridge crest in 1989. The interpretation of the aerial photograph was confirmed through ground checks. Transect walks along the dune ridges revealed that open areas with ripple marks and slipfaces, ie signs of aeolian activity, increased towards the east as relative dune height increased. This longitudinal development repeated itself east of site D. Three more vegetation free areas similar to site D were identified on the image and in the field up to termination of the ridge at Nahal Nizzana.

The emerging ridge containing site C did not appear to have undergone substantial changes since 1989. Yet, it shows the dependency of activity on dune height even clearer than its southern neighbour. West of site C the surface colour indicated nearly total cover of the microphytic surface crust, similar to the interdune corridor surfaces. Light coloured areas increase towards the east, accompanied by an increase of relative dune height. Active, vegetation-free areas of a similar state as site D are visible at the termination of the ridge at the banks of Nahal Nizzana on all available aerial photographs.

3.6 Summary results

1. What is the nature of winds causing sand movement within the linear dune ecosystem?

   The research site has two main sand transporting wind directions, each dominating a distinctive season. The S to SW winds are restricted to winter and can be divided into two groups: Episodic high magnitude cyclonic storms and diurnal, low magnitude winds of local origin. Within the winter season between the beginning of November and the end of March about five to ten cyclonic storms are common. During these events, which may last several days, the initial wind direction of sand transporting winds is S to SE. As the low pressure cell passes over the area from W to E, wind direction changes gradually clockwise from S to W. During calm weather situations in winter the local wind system may develop to strengths causing sand transport. These events are characterized by a continuous clockwise wind direction change from NE winds during night-time to winds from N in the early evening, Saltation threshold is reached at the dune crests during the morning and the early afternoon hours from directions between 180° to 270°.

   Sand-transporting NW to N winds during daytime are a summer phenomenon caused by a local temperature gradient between the Mediterranean and the Negev desert. Their direction range is narrower than that of the winter sand transporting winds. They are of lower magnitude but are steadier. They are a diurnal feature and affect only the crestal parts of the dune ridges.

2. How do morphology, surface condition and vegetation cover affect aeolian sand transport? What is the relative influence of each of these parameters?

   Starting at interdune corridor level, wind speed and thus the intensity of sand movement increases with height. This means that under similar surface conditions the interdune corridors are the least active areas while dune crests are the most active zones. In addition to height, speed ratio S_z between dune crest and the interdune corridor is dependent on wind direction above the dune crests. S_z is lowest for winds normal to dune orientation and highest for dune-parallel winds. Winds with directions oblique to the dune trend are deflected towards more dune parallel directions in the interdune corridors. Differences between $u_{0.24}$ at the high dune crest of site D and the low dune of site C depend mainly on wind direction relative to the small-scale morphology of the low dune.

   Rainfall stops aeolian sand movement. Over moist surfaces no continuous saltation was measured. Saltation over non-crusted surfaces resumes within one to three hours after rainfall events if $u_{0.24}$ remained above u_t for dry sand. High intensity rainfall causes very low intensity sand transport through drop impact.

The continuous vascular vegetation cover in the interdune corridor reduces near-surface wind speed and thus reduces shear stress at the sand surface. Saltation threshold speed u_t within the vegetation canopy is reached during fewer events than over open areas (site A). Compared to open surfaces, the natural vegetation cover of 17 per cent reduces sand transport to 1.0 per cent during average storms and to 8.5 per cent during high magnitude storms. A reduction of the vegetation by half causes a more than seven-fold increase of sand transport during average storms to 7.7 per cent compared to open surfaces.

Hummocks topped by mainly dead vegetation on low dune ridges cause an increase of surface roughness and thus reduce near surface wind speed ($u_{0.24}$) in the areas between them. Surface activity at site C was nonetheless sufficient to prevent the establishment of a microphytic crust on sand surfaces between the vegetated hummocks. Sand transport at the site is highly dependent on wind direction. Only winds parallel to gaps reach speeds $u_{0.24}$ capable of causing significant saltation. During exceptional storms hummocks are subjected to destruction despite their crust and vegetation cover. Scouring at the bases of hummocks may ultimately lead to a destruction of the feature.

Vegetation patches at active dune crests protect their immediate environment from deflation and lead to lee-side deposition of material. Gaps between two patches on windward slopes are the focus of increased erosion. This in turn may lead to a destruction of the hummock, as its base becomes subject to scouring. Deposition of sand within the vegetation of a hummock was not observed. New vegetation patches were not established during the research period, germinating seedlings were regularly destroyed by surface activity.

An established microphytic surface crust like in the interdune corridors reduces aeolian deflation to amounts below values measurable with the devices used in the current study. Very weak crusts on extremly exposed surfaces as for example at hummocks at site C can be destroyed by exceptionally strong winds. Disturbances of the crust such as erosion rills, footprints or tyre tracks can also be starting points for further destruction by aeolian processes. When the crust is removed from the surface, transport rates in the interdune corridors under natural vegetation cover reach about 1 per cent of dune crest values. Complete destruction of vegetation in the interdune corridors leads to transport rates of approximately 10 per cent of dune crest values during winter storms.

3. What is the actual spatial distribution of aeolian sand transport within the linear dune system? Does the current condition represent a 'steady state' resp. an 'equilibrium'?

   Within an average year, aeolian sand transport is likely to occur on all surfaces without surface crusting. However, currently all interdune corridors and plinth surfaces are crust covered. Extensive, non-crusted surface areas are restricted to the crestal parts of the linear dune ridges. Thus aeolian sand tranport under current natural conditions is restricted to the dune crests. Saltation at the crests of dune ridges was observed during all seasons. The appearance of the active crest areas changes with the predominant wind of each season. Deflation was most severe on steep windward slopes, that is northfacing slopes during summer and southfacing slopes during winter. Deposition was detected in the lee of obstacles and the crestline. Activity at the lower parts of the crest, in a magnitude which would change the morphology of a ridge, was only observed during exceptional events. The sediment budget of highly active areas was neutral during winter, while during summer dune volume decreased, resulting in a net loss of volume within the research period. Indications of considerable previous loss of material at the dune crests can be found through the comparison of ground and aerial photographs taken between 1982 and 1999.

In the interdune corridors threshold speed u_t for loose sand was only reached during cyclonic storms, therefore aeolian sand transport at the test sites A and B2 was restricted to winter. Saltation over crust covered surfaces (B1) was observed during the initial phases of very high magnitude winter storms, when loose sand sized material present on the microphytic crust is moved. Aeolian erosion did not affect surfaces in the interdune corridors and on the plinths of dune ridges covered by the microbial crust. Destruction by wind action was observed at exposed crust covered features like vegetation bults at site C during high magnitude winter storms. Deposition of loose sand onto crust covered plinth surfaces was observed as a result of cyclonic storms.

As in the autumn 1999, significant aeolian sand transport processes are restricted to the uppermost parts of highest dune ridges in the area. Their sediment budget is negative. The major part of the ecosystem is not affected by the current wind regime as far as morphological significant sand transport is concerned. Interdune surfaces and footslopes of the dune ridges are stabilised by the combination of a microphytic crust which cements sand grains and a vascular vegetation cover which reduces near surface wind speed.

Discussion

4.1 Sand moving winds

A prerequisite for the development and maintenance of linear dunes is the existance of a bimodal wind regime and sufficient sand supply (Wasson & Hyde, 1983). The aspect of sand supply will be dealt with below (section 4.2.2, p.139). The bimodal nature of the wind regime at the research site has been ascertained by determining the effective wind e (Tsoar, 1983) based on long term data. Winds from the SW quadrant are the dominant force responsible for sand transport at dune top level. Second to this major component are winds from the NW quadrant. Calculated resultant effective wind direction is parallel to dune ridge orientation, thus all aeolian requirements for the formation of linear dunes are fulfilled, the existence of the dunes at the site can be explained with the general current wind regime.

The sand transporting NW winds are a summer phenomenon, while SW winds are restricted to the winter months (see also Tsoar, 1983). This seasonality of the wind regime explains the results of calculated e based on near surface wind speed measured in the course of the current research (figure 3.2, p.43). As the majority of datasets in this database was acquired during winter, the results for site D at the dune crest must be regarded as strongly biased towards SW directions. Setting aside this bias, the results of the current study are consistent with the findings of Littmann (1997) and data from nearby climatic stations published by Tsoar & Møller (1986).

The basic local wind regime is of diurnal nature during both summer and winter. A uniform daily cycle of wind direction and wind speed is dominant at the research site during stable atmospheric conditions over the eastern Mediterranean. This daily cycle is caused by the temperature gradient between the water body of the Mediterranean to the NW and the desert surface of the Negev in the SE. The direction of the gradient changes between daytime and night-time, causing a 360° rotation of wind direction within 24h. At night, southerly winds of low magnitude prevail during all seasons, as the bare surface of the desert cools down below the temperature of the water, thus causing a reversal of the airflow. The phase of the rotation is shifted seasonally due to the changing length of the days and the changing temperature of the water body and the land surface. The heating of the land surface to temperatures above those of the Mediterranean leads to NW winds in the afternoon. During the summer months, these winds surpass saltation threshold at dune crests like site D each day, causing sand transport and deflation on wide areas of the windward slopes of the crest area. The transported material is accumulated at southfacing slipfaces at the crest. The occurrence of this sand transporting sea–breeze is very predictable throughout summer. Its maximum measured near surface velocity ($u_{0.24}$) was 8 m s^{-1}. In winter, southerly winds dominate most parts of the day, as the heating of the desert's surface is slow. The southerly winds during the morning hours occasionally reach u_t and then cause sand transport at the uppermost parts of exposed dune ridges. A change to northerly directions occurs only in the late afternoon. These northerly winds do not reach saltation threshold velocity u_t. The variability of the speed and the direction of the sand transporting winds during summer is low due to the stable atmospheric condition throughout the Middle East. No cloud cover prevents surface heating. During winter the development and magnitude of a diurnal regime depends on the regional distribution of atmospheric pressure as well as on more local aspects, as for example cloud cover. The latter influences surface heating and thus the resulting temperature gradient. The diurnal winter system will only develop to full scale if cloud cover and regional pressure gradients are low. Yet in general, the magnitude of the resulting airflow is low, leading only to minor sand transport at the most exposed areas of dune crests.

Disturbances of the local wind system in winter are caused by the instability of the regional pressure system, as eastward moving cyclonic depressions affect the eastern Mediterranean. These cyclonic low pressure systems are episodic events and are a regular cause for high magnitude southwesterly winds. The major causes for sand movement during winter are thus not the locally developed winds, but episodic high magnitude storm events. Their frequency and magnitude varies from year to year. While during the winter 1997/98 storms of average magnitude were frequent and an exceptionally high magnitude event was recorded in March, the following winter saw only moderate events. This is reflected in the average daily sand movement (see figure 3.69, p.119). While the values for summer show little differences from year to year, values for the three observed winters vary considerably. However, the high intensity determined for spring 1997 should be viewed with caution, as it is based on data of only the second part of the winter, February until April. The high magnitude events in the following winters were all recorded in this late winter period, while between November and January only storms of average magnitude were observed in all three years. Therefore it appears likely that the high values of average sand movement in 1997 are a result of the incomplete data set. In addition to the differences in magnitude of sand moving winds, winter storms show also a greater variation of wind direction than summer winds. Within an event, winds exceeding u_t change from initially southerly directions to SW and W. This has implications on the way they interact with the dune morphology and on the spatial distribution of aeolian sand transport. These aspects will be discussed below in detail (Sections 4.2.1 and 4.3).

The assumption at the beginning of this study was, based on a wind regime with two dominant, seasonally changing wind directions (Tsoar & Møller, 1986), that the high magnitude winter storms would have the most pronounced influence on dune morphology and sediment budget of the active areas at the crests. As sand flux q is a power function of windspeed, the winter storms were expected to move large quantities of sand mainly at the dune crests. As saltation height increases with shear stress (Sorensen, 1985), high windspeeds would also cause high trajectories of saltating grains, thus leading to great transportation widths. These, in turn, would increase the probability of sand grains to be blown off the crest, although Walker (1999) found that the majority of saltating grains settles within saltation width in the lee of the crest. It was also assumed that winter storms cause sand movement in the interdune corridors, footslopes and low dune ridges whenever loose sand is available at disturbances of the surface crust cover. In contrast to the episodic events in winter, the steady but moderate diurnal sea-breeze during summer was expected to only modify the active dune areas, namely to reverse the direction of the slipfaces at the crest to attain a low resistance, steady state profile as described by Tsoar (1985). It was not expected to cause any significant volume change of the dune body. Because of its low magnitude and narrow direction variance, it was assumed that the sea-breeze would only affect the most exposed upper parts of the dune crest, while on the lower flanks of the crests, on the remaining dune body and in interdune corridors, no aeolian sand movement would take place.

The results of this study show that the high magnitude winter storms are indeed capable of inducing sand transport throughout the research site, as long as the surface is not covered by a microbial crust. They are regularly causing sand transport at dune ridges which are partly stabilised like site C, because there, other than in the interdune corridors, non-crusted areas are common. These areas are either concave blowouts exposed towards west or narrow channels of varying directions between crust covered hummocks, which were affected only when wind direction was parallel to their orientation. Under such conditions, sand flux and transported mass are considered to be similar to values determinesd at site A. At the active dune crests, the cyclonic winter storms move large quantities of material. They lead to a reshaping of the dune crest, resulting in a typical winter form (see below, section 4.2.1, p.130). Transport from the dune crest areas onto the lower footslopes was observed in the field during the exceptional storm of mid-March 1998. A thin layer of loose sand could be seen on crust covered surfaces of the footslopes and beyond after the storm. Similar effects were not observed afters torms of average magnitude.

However, in contrast to the initial assumption, the high magnitude events do not cause the most significant changes at the currently active dune crests as far as the sediment budget is concerned. While the exceptional storm of March 1998 moved considerable quantities of sand, the sediment budget at the crest remained neutral (table 3.8, p.106). The storm with the second highest magnitude on 17/2/99 even led to a positive budget. It appears that the potential of high magnitude storms to mobilise additional material in areas of the dune which are not vulnerable to wind speeds common during summer, makes up for the loss of material through transport onto the vegetated, stabilised lower slopes and into the interdune corridor. The sources for this material are seen in the upper parts of the footslopes, ie the transition zone between mobile crest and inactive plinth where the surface crust is thin or non-existent. In addition, dune parallel winds during a cyclonic storm contribute to a balanced budget by transporting sand along the crest, as long as an upwind sand source exists. Winter must thus be regarded as the season in which the sediment budget of active dune parts is maintained. The events during winter caused by the development of a local wind field affected only the upper parts of the dune crest and were thus of little significance for the long term development of these areas.

The low magnitude winds of summer lead to large quantities of moved sediment and cause a negative sediment budget at site D, as shown in figure 3.70, p.118. The lower intensity of sand movement during summer as compared to winter is balanced by the diurnal occurrence of sand moving winds (figure 3.69, p.119). Sand accumulated along the centre line of the dune crest during high magnitude winter winds from S to SW is transported southward during summer by NW-winds. At all areas below the dune crest – footslopes, low ridges, interdune corridors – the summer winds do not cause sand movement.

4.2 Factors influencing sand transport

4.2.1 Morphology

The influence of morphology on airflow and thus on sand transport is to be divided into two scales: on a small scale, the distribution of slip faces, local crest lines and obstacles like vegetation bults determines the development of the active dune parts. On a larger scale, the dune ridges are obstacles to an idealised free airflow over a plain surface, they modify this airflow and thereby influences the aeolian processes in the interdune areas, depending on wind speed and direction.

4.2.1.1 The small scale

At site D monitoring of the dune crest using erosion pins has shown that the distribution of zones of erosion and deposition is connected to slope exposition and gradient. The highest intensity of change was determined on either the steepest parts of the area under surveillance or at gradient changes. Although the linear ridges at Sde Hallamish are not reversing dunes like the ones investigated by Burkinshaw & Rust (1993), a reversal of the dune crest morphology is caused by the seasonal change of the wind regime. As in their study on the influence of dune slope on erosion rates under reversing winds for transverse coastal dunes, the highest erosion rates at site D were measured at the steepest parts of the windward slopes, caused by the compression of airflow as it is forced to change direction.

The SW-storms of cyclonic nature are responsible for the movement of sand from lower areas of the slope towards the dune crest. During their initial phase, wind direction being roughly 90° to dune ridge orientation, their effectivity in eroding sand at lower parts of the active, southfacing crest area is highest (Burkinshaw & Rust, 1993). This is because the gradient of a slope as 'seen' by the approaching wind varies, depending on wind direction. It is steepest for wind directions approaching the contour lines at 90°. Under such conditions, material is transported upslope and deposited immediately north of the crest where streamlines diverge and/or a separation vortex develops, depending on wind velocity and form of the crest (Sweet & Kocurek, 1990; Frank & Kocurek, 1996b). As wind direction changes gradually clockwise during a cyclonic storm, upslope transport is reduced, while transport along the dune axis gains importance. Airflow oblique to the dune ridge is deflected in the lee of the crest (Tsoar, 1983; Sweet & Kocurek, 1990), its speed depending on the angle of the wind towards the dune ridge. The lower the angle, the higher the speed of the deflected flow. For angles $\leq 40°$ the speed of the deflected airflow is higher than the speed at the crestline (Tsoar, 1983). Within the separation zone, sand transport direction is therefore oblique to incident direction (Walker, 1999). Upslope, ie northward, transport is halted when wind direction reaches dune ridge orientation. Sand transport is now parallel to the crest's orientation. Erosion zones were observed north of the slipfaces at the eastern part of site D during single storm events and as summary results during winter seasons, stressing the efficacy of lee side transport. The diurnal low energy southerly breezes and the NW-sea breeze developing irregularly during the transition period in spring modify the form of the crest during winter. Its triangular, *seif*-like appearance during winter is mainly a result of the wide directional variance of the cyclonic storms which act on both sides of the crest during an event (McKenna Neuman *et al.*, 1997). The increasing influence of the sea-breeze leads to an accentuation of the triangular form as long as SW-winds are still occurring in spring. After the last cyclonic storm of the season, a quick transition to the summer form follows. Exceptionally high flux rates at these parts, as for example on 19/4/99 (3.4.1.4, p.79) may be explained by the deposition structure of the sand caused by a preceding event with sand transport from the opposite direction: in the case of 19/4/99 northfacing slipfaces without layering, thus no bonding of the grains and high erodibility combined with the steep gradient of the slope, leading to maximum compression of airflow (Tsoar, 1985; Burkinshaw & Rust, 1993).

Figures 3.68, p.117 and 3.65, p.115 show the development of the crest during summer: at the beginning of summer crest and brink are not separated. High slipfaces are built up where sand transport is most intense, indicating flow separation in the lee of the crest. Reversed flow may occur on the lower part of the active crest as a result of its form (Sharon *et al.*, 2002). Other parts of the crest attained a rounded cross section (Sweet & Kocurek, 1990) during the first sea-breeze events, as the winter crestline was only weakly developed. Thus no separation of airflow occurs during NW winds. Evidence of this are the ripple marks in the foreground of figure 3.68, caused by northerly winds south of the gap between N3 and N4 (figure 2.8, p.30). The sharp edged crestline is transformed by the steady NW-sea breeze into a rounded convex profile with separated crest and brink in the course of summer, a form typical for dunes under unidirectional winds (Tsoar, 1985, p.58). Recent studies of reversing dunes have yielded similar results (McKenna Neuman *et al.*, 1997, p.1112). The continuation of the ripple marks visible in figure 3.65, p.115 onto the leeward slope indicate a good adjustment of the dune to the prevailing wind as no separation of the airflow from the surface appears to occur. This is seen as a result of the combination of moderate wind speeds during summer, their unidirctional nature and the ensuing well adjusted form of the dune crest in late summer. The lack of a separation cell in the lee of the part of the dune which is active during summer leads to a restriction of sand transport to oblique directions at high acute angles, as no flow lee side deflection occurs. Therefore longitudinal transport is minimal during summer.

4.2.1.2 The larger scale

On the larger scale of the dunefield of Sde Hallamish, the ridges in their present conditions can be idealised as fixed obstacles on a flat surface, influencing airflow as it passes over them. Speed and direction values in the interdune corridors differ from those sampled simultaneously at the dune crest.

Horizontal flow deflection Flow direction is altered by obstacles. Obstacles with a large lateral extension will deflect flows approaching the obstacle at low angles. High angles of attac (close to 90°) should not lead to flow deflection. In this case the flow direction is either maintained if the flow remains laminar or else turbulence develops on the lee side of the obstacle. Sharon *et al.* (2002) have shown this for the dune ridges at the research site. A very distinct pattern of flow deflection at the interdune sites is based on the wind direction above the crests of the dune ridges. On the lee side of the ridges oblique airflow is deflected to directions sub-parallel to the general trend of the linear dunes. The degree of deflection in the corridor depends on wind direction relative to the dune ridges. A strong linear dependency with minima for dune parallel and dune normal wind directions has been shown (figures 3.5, p.48 and 3.6, p.49).

The general pattern of deflection at both interdune corridor sites is similar, yet absolute values of δ_W are generally lower at site B than at site A. Differences of δ_W between the sites for specific directions can be explained by their position relative to the respective neighbouring upwind dune and its morphology. Although both sites are located in the same interdune corridor, the cross sections of the corridor and the dunes differ significantly (see figure 2.2, p.23) as the two neighbouring ridges diverge downwind.

At site A the deflection for northerly winds is considerably higher than for southerly winds. The main reason for this behaviour is seen in the differences of the height of the adjacent dunes. While the neighbouring dune to the north is 10 m high, the height of the ridge to the south is only 4 m. The latter dune emerges as a distinct form only east of the border road, increasing in height towards the east. The result is a stronger deflection of winds from SE compared to winds from SW. In the vicinity of site A the southern ridge is inactive and has a rounded cross section, while the northern ridge has an active crest with steep southfacing slip faces reaching to the interdune corridor. This 'sharper' profile is thought to cause the steep gradient at the change of the direction of deflection between 340° and 10° compared to the gradient at the similar position for southerly winds. The longitudinal changes of the northern dune are a contrast to those of the southern dune: its height decreases eastward. As a result, deflection values of NW winds are higher than those of NE winds. This decrease of height is also responsible for the lower values of δ_W at site B compared to site A, analoguous to the difference between SW and SE winds at site A. In contrast to site A, higher values of δ_W were determined for SW winds than for SE winds at site B. For SW winds even absolute values of δ_W are above those of site A. These differences are attributed to the differences in the cross section of the corridor and the local orientation of the southern dune ridge at the sites. The corridor is substantially wider at site B, whereby the distance between dune and anemometer array increases more towards the south, ie the array is not located in the centreline of the corridor. For this reason it was expected that deflection for SW winds would decrease. Yet dune height increased from 4 m at site A to 7 m at site B, thus it is a greater obstacle with potentially greater influence on the airflow. For winds from SE, deflection is indeed smaller although dune height increases further towards E. But the corridor also widens towards E and dune orientation SE of site B deviates from the general trend. Therefore the angle of attack at the ridge responsible for the deviation of winds with a SE component blowing towards site B is lower than for winds blowing towards site A.

The position of site C was chosen because it represented a 'degenerated' form of a formerly active linear dune ridge. With regard to the observed development of the dunes at Sde Hallamish between 1982 and 1997 it was to resemble a possible later, more stabilised stage of site D. It was thus expected that the deviation of wind directions determined at the site from those at site D would be marginal. The determined deviation was indeed small, but nonetheless a pattern similar to the interdune sites is recognisable in figure 3.7, p.50. For southerly winds, this can be explained by the presence of a higher dune to the south, which appears to affect airflow over the neighbouring, lower, dune. The deviation figures determined for winds with a northerly component between 270° and 90° were unexpected, because the next dune to the north was considered to be too far away and too low to have any influence on the wind regime above site C. The highest values of δ_W have been determined for NE winds for which the distance to the next upwind ridge is highest, approximately 400 m. Since an influence of the next upwind ridge was ruled out thus a wind direction at the foot of the ridge which is parallel to that determined at site D was anticipated, the deflection must be caused locally by the dune ridge. As the northfacing slope at site C is steep

($\geq$20°) it appears likely that flow deflection occurs in the lee of what remains of the dune crest (Tsoar, 1983). During several of the storms described by Sharon *et al.* (2002) flow reversals were observed on the crest of the same dune approximately 300 m W of site C. For NW winds the values of δ_W are inconsistent, which may be a consequence of vortices caused by the combination of the steep slope and the rough terrain upwind of site C.

In general the angle between dune ridge and wind direction for oblique winds with a speed $u_{0.24}$ higher than 4 m s^{-1} decreases downwind of the ridge. The value of δ_W depends on the incident angle of the wind towards the dune ridge, on the distance of the measuring site to the upwind dune and on the cross section of the ridge. The range of wind directions of sand moving winds is smaller in the interdune corridors than at the dune crests. As a result, sand transport, if occurring, is mainly parallel to dune ridge orientation. Transport of material from the dune to the corridor is unlikely, as values of δ_W are higher close to the ridge, where even reverse flow has been determined (Sharon *et al.*, 2002). On the other hand, sand from unprotected surfaces in the interdune corridors may be added to the ridges by oblique winds as suggested by Tsoar & Møller (1986) and Pye & Tsoar (1990, p.211).

Flow acceleration and sheltering effect Spatial differences of wind speed within the dunefield are the result of acceleration of airflow on windward slopes or deceleration on the lee side of transverse obstacles. The intensity of acceleration depends on slope length and gradient (Jackson & Hunt, 1975; Lancaster, 1985; Tsoar, 1985; Mulligan, 1988; Burkinshaw & Rust, 1993; Frank & Kocurek, 1996a; Lancaster *et al.*, 1996; McKenna Neuman *et al.*, 1997; Walker, 1999). Changes of slope and gradient in the research area were only observed at the active dune crests, where the effects on the pattern of erosion and deposition at site D are shown above. However, as the directions of the major sand transporting winds range from 135° to 0°/360°, slope and gradient of the linear ridges as 'seen' by the approaching wind vary. The dependency of the speed ratio S_z on wind direction at the research site is shown in figure 3.4, p.46. These results confirm the findings of Lancaster (1985) who found a strong relationship between S and the angle of incidence between wind and dune. As a result of the setup of the measurement sites, the determined values of S_z, although basically simply the reciprocal value of S, are not to be compared directly with the acceleration factor S of airflow up a specific slope as in the published studies, which deal with unidirectional flow up a slope of a single dune. The values of S_z in this study rather represent the average differences between wind speed at different positions within the dune field for the major sand transporting wind directions at Sde Hallamish. These differences of wind speed thus contain effects of upslope acceleration as well as effects of lee-side deceleration, which cannot be clearly distinguished based on the available data. However, as one aim of this study was to investigate and explain the distribution of aeolian processes within the linear dune field, the focus has been set on the differences of the near surface wind speed instead of the exact mechanisms causing them.

Based on the model of Jackson & Hunt (1975), which states that the 'fractional speed-up ratio' Δ_S equals two times the height of the obstacle divided by the slope length, its value tends to zero for airflow parallel to the long axis of linear dunes. Under such conditions, similar wind speed can be expected independent of the situation of the measuring station, given that the surface roughness is also similar. Indeed, the highest values of S_z were determined at all sites for wind directions parallel to dune ridge orientation. The fact that S_z at site A remains below 1 for dune parallel winds indicates that the gradient of flow velocity in the boundary layer above the ridges is steeper than in the interdune corridors. It appears that this steeper gradient affects mainly the near surface flow, as values of $S_{1.77}$ for both interdune sites (A and B) are nearly identical (0.9), independent of the differences of surface roughness, while $S_{0.24}$ at site A is lower (0.79) despite the similarity of surface conditions between sites A and D. This may be an indication that streamline compression affects near surface airflow over surfaces protruding into the boundary layer even if it is not deflected by a slope, ie if it is parallel to the orientation of a linear dune. For other than dune-parallel wind directions, the spread between the values of $S_{0.24}$ and $S_{1.77}$ has been described as the influence of the speed-up of the near surface airflow within the 'amplification layer' (Frank & Kocurek, 1996a) as it passes over the dune, stressing the importance of near-surface measurements in studies investigating sand transport processes. The direct influence of a possible separation cell caused by transverse winds in the lee of the northern dune can be ruled out for sites A and B2. Flow reattachment occurs about four dune heights downwind of the point of flow separation (Frank & Kocurek, 1996b) . Height difference at site A is 10 m, distance from the brink is 70 m. At site B height differences and distances are 7 m/50 m (B1) and 9 m/70 m (B2). Distances grow for oblique winds while the probability of the formation of a separation cell declines. Thus the interdune sites are outside a possible separation cell during all wind directions.

The pattern of speed variation is similar at both interdune sites, but the values of $S_{0.24}$ within the vegetation canopy at site B2 are significantly lower. The highest values of $S_{0.24}$ were 0.54 for dune parallel winds under 17 per cent vegetation cover. A comparison of this value with site A ($S_{0.24} = 0.79$) indicates that 54 per cent of the velocity difference between site D and B2 can be attributed to the influence of vegetation, while the remaining 46 per cent is due to the morphology, ie the position of the measuring site within the landscape. As for flow deflection, differences of the cross section of the corridor and the adjacent dunes play a significant role for the sheltering effect. As has been noted above, the dune to the south of sites A and B is considerably lower than the dune bordering on the corridor to the north (see also figure 2.2, p.23). In addition to the resulting differences in sheltering, the ridge of site D, where the reference speed was determined, has an asymmetric cross section. The gradient of the northfacing slope is steeper, leading to increased acceleration compared to the southfacing slope. Thus two factors effecting S_z are combined: for southerly winds, lower amplification of the reference airflow due to the lower gradient but less sheltering of the interdune corridor site due to the lower height of the upwind ridge. For northerly winds the relation is reversed. This superposition of factors leads to higher values of S_z at the interdune sites for all winds with a southern component as compared to those from northern directions with the same angle of incidence. The effect is more pronounced at 0.24 m than at 1.77 m at both sites and at site A more than at site B.

Values of $S_{1.77}$ at sites A and B for winds between 210° and 270° are equal. Thus windspeed at both sites is equal for approximately dune parallel winds and those with a SW component, independent of surface roughness. As $u_{1.77}$ was found to be independent from surface roughness, it may be used as an indicator for the intensity of sand transport over bare surfaces, as the relation between flux and $u_{0.24}$ is known. Looking now at the relation of $S_{1.77}$ for winds with a northerly component, or southerly winds with an obliquity greater than 60°, we find that values at site B2 are significantly higher than at site A. Hence if site B2 was without vegetation cover, sand transport would be higher than at site A. The main cause for the increase of S_z at site B2 relative to site A for oblique winds is the greater width of the interdune corridor at site B. The characteristics of the dependency of S_z on wind direction is not symmetric for northerly and southerly winds. While for winds from the southern sector values diverge only for angles of incidence $\geq$60°, no equality exists for winds from the northern sector, where the difference between site A and B2 increases from dune parallel winds (270°) to almost transverse directions (350°). This behaviour is best observed during the sea-breeze events of summer. Both, $S_{0.24}$ and $S_{1.77}$ are higher at site B2 during the periods of sand transport at the dune crest. All recorded sea-breeze events show this pattern. As the surface roughness at site B2 is higher and thus lower values of $S_{0.24}$ were expected, the reason for the differences between site A and site B is the combination of a reduced dune height of the northern dune and a greater distance to this obstacle at site B compared to site A.

At site C, although it is situated on a dune ridge, the graph of $S_{1.77}$ is similar to those of sites A and B as far as the positions of maxima and minima are concerned. However, the amplitude is lower and values are considerably higher than in the interdune corridor for high acute angles and transverse flow. In addition, the relation is reversed, values for northerly winds are higher than for southerly winds with identical obliquity. The vicinity of the high dune to the south causes lower values of $S_{1.77}$ than in the corridors for oblique winds from the SW sector, while the wide, open upwind area towards the north leads to high values for northerly winds. The fact that for dune parallel winds $S_{1.77}$ is below interdune corridor values is regarded to be a result of an increased surface roughness at the site, which places the anemometer in a lower part of the boundary layer. The roughness elements at site C are higher and wider than at site B and in contrast to the porous plants, the hummocks are massive. Within the roughness layer the behaviour of $S_{0.24}$ at site C differs considerably from the interdune corridor sites. Values of $S_{0.24}$ remain below 0.3 for all directions, which means that $u_{0.24}$ at the site remains below 30 per cent of dune crest speeds. The distribution of peaks and lows does not show a direct connection between $S_{0.24}$ and the general dune trend. It rather reflects the local distribution of hummocks and voids relative to the position of the anemometer array.

The influence of wind direction on wind speed at different morphologic positions is easily detected by regarding cyclonic storms passing over the area, because wind direction changes during the event. During the early phases of cyclonic storms high wind speeds and transport rates are recorded at site D. Winds are usually from S during this phase. The first gusts on the 15/3/98 (see figure 3.17, p.63) even had a SE component. The lowest values of S_z were determined during these southerly winds. Very high transport rates were usually observed at the exposed dune crest at the beginning of a storm, while in the interdune corridors wind speed and hence transport rates remained low. The maxima of wind speed and sand movement in the interdune corridors were recorded during later stages of the storm, after the wind direction had changed to more westerly directions. Especially during the event of 17/2/99, when at the dune top two almost identical peaks of wind speed and sand transport were recorded within a period of 1.5 hours. Between the peaks, wind direction at site D changed 20° clockwise. At site A this led to an increase of S_z from 0.52 to 0.61 and a twofold increase of sand transport (figure 3.16, p.60).

For oblique winds it has been shown by various authors (Frank & Kocurek, 1996b; Walker, 1999) that a recovery of disturbed flow occurs downwind of obstacles. It has also been stated that at the base of the upwind slope of dunes flow velocity is reduced (Tsoar, 1985). Highest δ_W for any given oblique wind direction leading to attached deflected flow is expected at the lower part of the lee slope, values decreasing with distance from the slope (Sweet & Kocurek, 1990). Similar results have been obtained by Walker (1999) for high acute angles at transverse dunes and Sharon *et al.* (2002) at the linear dunes of the present study. Similar to wind direction, wind speed recovers downwind of obstacles (Frank & Kocurek, 1996b; Walker, 1999). Although Walker (1999) doubts a complete recovery to undisturbed upwind flow conditions within an area of closely spaced dunes, it appears likely that within a linear dune field such as at Nizzana, a dynamic equilibrium will be established if wind speed and direction remain constant over a period of more than a few minutes or if the gradients of change are low.

Given such a 'stationary' situation, the results for the interdune sites in this study, determined near the centreline of the corridor, show average values of δ_W and S_z. Upwind of the sites δ_W is considered to be higher, while S_z should increase downwind and vice versa. The increase of wind speed increases the probability of sediment transport in the interdune corridors in downwind direction. Together with the lower δ_W the likelihood of sand transport from the corridor to the dune increases as corridor width increases. During winter storm events saltation transport has been observed at non-crusted interdune surfaces like sites A and B2, hence such areas are a source of material for dune building as suggested by Tsoar & Møller (1986). During summer, values of $S_{0.24}$ at the interdune corridor sites remain below 0.5 during the period of the highest wind speeds $u_{0.24}$ at site D, which in turn reached only 6 m s^{-1} during the majority of events. As the threshold velocity $u_{0.24t}$ was found to be 4 m s^{-1}, sand transport in the interdune corridors during summer requires a minimum $u_{0.24}$ of 8 m s^{-1} at site D. At Nizzana this is the maximal value during summer (Tsoar & Møller, 1986, p.89). Assuming the increase of S_z downwind of the measuring sites, significant sand transport during summer, even under ideal conditions (no vegetation, no surface crust) can be expected only over less than 50 per cent of the interdune corridor surface. A direct transport of material from one dune ridge to the next is very unlikely under current conditions. Transport distance in the lee of the dunes during transverse flow is considered to be not more than the characteristic saltation length for between 70 to 95 per cent of the total sediment moved (Walker, 1999). Of the remaining 5 to 30 per cent, the major part will settle on the dune slopes. These findings are in accordance with observations at Nizzana, where only during winter storms with exceptionally high wind speeds has sand been transported away from the crest to be deposited on the plinth and even in the interdune corridors. There increased roughness caused by the dense vegetation cover in addition to deceleration of airflow prevents further transport. The results support the thesis of Wiggs *et al.* (1995) that the limiting factor for sand transport in non-vegetated interdune corridors and on lower dune slopes is the available wind energy and not availability of material.

The results of the current study show that the influence of the linear ridges modifies the airflow in such a way that sand transport in the interdune corridors is significantly reduced compared to dune crest values. The difference of measured flux rates over surfaces of similar roughness (sites A and D) ranges within the results determined by McKenna Neumann *et al.* (1997), who found that transport increases by 1-2 orders of magnitude from the base to the top on the upwind slope of dunes. All oblique winds are modified by the dune ridges in two ways: first, the direction of airflow is changed downwind of the ridge to a more ridge parallel direction and second, the near surface speed is reduced. The amount of modification depends on the incident angle of the wind (Sweet & Kocurek, 1990). Winds with an angle of 90° to ridge orientation are not deflected, but velocity differences between crest and corridor are highest. Dune parallel winds are not deflected and speed difference is minimal. The W-E orientation of the dune ridges prevents all but dune parallel winds from acting on the surface in the interdune corridors without interference. Hence only westerly winds play a major role for sand transport in the corridors. The evaluation of total sand transport in the interdune corridor shows that during the measurement period directions were limited to a narrow range between 240° and 270° at the interdune corridor site A. Easterly winds of sufficient strength were not recorded during the study, but may play a role according to published wind records of the site (Sharon *et al.*, 2002, p.874).

4.2.2 Surface properties

The availability of transportable material is considered by Williams & Lee (1995) to be 'one of the most significant factors controlling saltation transport'. In other words: if the carrying capacity of the wind is higher than the amount of available material, one will observe either deflation of an erodible surface, that is a change of morphology, or, if the surface is protected, sand transport will be well below expected values. An increase of vegetation cover at the ridges is visible on aerial and ground photos taken since 1982. At the same time, dune height decreased. A loss of material was also directly measured at site D during the study period. Two factors are responsible for this development: the cutoff of the ridges from their upwind sand source and the re-stabilisation of the interdune corridors after the cessation of grazing.

The local sand source of the dunes is situated on Egyptian territory, several kilometres west of the border. The linear dunes of the research site can be traced on aerial photographs to sand ramps on the eastern bank of a S–N running wadi. Net transport on *vegetated linear dunes* is along the crest of the dune (Wopfner & Twidale, 1988; Pye & Tsoar, 1990). Sand accumulations observed after storms at the cuts of the patrol roads through the dune ridges, mainly on the Egyptian side, are regarded as proof of continued longitudinal sand transport. These accumulations are cleared by bulldozers as soon as the road is likely to become blocked. Usually the sand is simply moved to the area next to the road dam. On the Israeli side, only minor accumulation at the gaps has been registered, while on the windward side of the cuts signs of deflation of sand are common. The cemented sand of the dune core is exposed. Walking on these surfaces was possible without leaving footprints. The gaps in the ridge thus appear to trap most of the material transported along the dune and thereby cut the sediment supply of the dune ridges east of the border roads. Further dune parallel transport of material east of the border therefore leads to a depletion of material at the upwind end of the dune ridge. The decrease of vegetation cover and the evident increase of surface mobility as one follows the dune ridges eastward shows the lack of transportable sediment east of the border road (see figure 3.74, p.122).

In the interdune corridors surface properties such as soil moisture and the presence or absence of a surface crust play a major role in the way that the availability of transportable material is restricted. Tsoar & Møller (1986) claim that if interdune corridor surfaces are only sparsely vegetated, sand from the corridors is added to the dune ridge by oblique winds. Thus, apart from the major upwind sediment source, interdune areas can act as a secondary source if their surface is composed of loose sand. Wind and sand transport data of the interdune sites A and B show that although the main transport direction in the interdune corridors is parallel to the dune ridge orientation, a southern component is also present as shown in figure 3.12, p.54. This would lead to sand transport from the corridors to the ridges if open sand surfaces were present. As sand transport from a linear dune ridge into an adjacent interdune corridor is very limited due to lee side deflection of oblique winds (see above, p.132), it appears likely that on destruction of a protective layer on an interdune surface, material from these areas plays a major role in the increase of dune height as observed at the research site between 1967 and 1982 (TSOAR & MØLLER, 1986).

Looking onto Egyptian territory using recent aerial photographs (1999) and ground observation from across the border, the dunes do not appear to have changed since 1982. Vegetation cover is sparse. Grazing flocks of goats and sheep were observed during the study period in the interdune corridors and on the dune ridges, thus keeping the cover of vegetation and surface crust low. The active appearance of the ridges west of the border is considered to be a result of sufficient sediment supply from the major source and the ongoing destruction of any surface cover by grazing animals.

4.2.2.1 Soil moisture

Rainfall was recorded during several storms where u_t was surpassed prior to precipitation. When the intensity of rainfall was high enough to wet the upper sand layer, sand movement stopped. Pore water increases the cohesive forces between the sand grains, thereby increasing the necessary force for initiating sand movement (MCKENNA NEUMAN & NICKLING, 1989; FECAN *et al.*, 1999). If the storm continued after the rainfall, the upper layer dried up quickly and as soon as the cohesive force of the moisture had gone, the grains started moving again, with transport rates reaching those of dry conditions (see eg figure 3.38, p.87). At the same time moisture content in the underlying sand remains high after rainfall events. Due to the large pores of the sand water penetrates into the sand body. As an example, a total of 1.7 mm of rainfall led to a moist layer of 3 cm on 15th January 1999. During dry periods the uppermost layers dry up quickly. Rain has therfore only a very limited influence on the sand movement. In general the effect is restricted to the time of rainfall itself, while immediately afterwards the drying of the surface begins, enhanced by wind, which in turn leads to the resumption of sand movement. These findings correspond well with the conclusions of JACKSON & NORDSTROM (1997) who, in a study at a sandy beach environment, found that the drying via evaporation is a more important control of sand movement than the water content of the sand.

Evidence for rainsplash was found at the research site in different forms: sand grains were found sticking to instruments and sand traps contained sediment after rain events not accompanied by wind speeds above u_t. Moreover, recorded sporadic impacts during rainfall as on 5/1/98 (figure 3.39, p.88) or 13/12/97 (figure 3.38, p.87) are most likely caused by sand grains detached by rainsplash. The impact of raindrops as a cause for the signal is ruled out as the electronic filter of the *saltiphone* distinguishes between sand grains and raindrops (see (SPAAN & VAN DEN ABEELE, 1991). On open, non-crusted sand surfaces drop impacts leave small craters and aggregated sand grains after rain events with low wind speeds. At the dune crests evenly distributed aggregated moist sand was observed on the lee side of the crestline after high intensity rainstorms with winds above u_t for dry sand where during and immediately after the rainfall no continuous saltation was measured. Transport of these aggregates is most likely along the surface below 2.5 cm as otherwise the movement would have been registered by the *saltiphone*. In wetter environments, detachment of sand grains by impacting raindrops has been reported to increase sand flux during rainfall to values above dry conditions (van Dijk *et al.*, 1996). In contrast to the conditions on the Dutch coast, where this study was conducted, rain events are only few and of short duration in Nizzana, thus this effect was not measured.

Raindrop impacts also influence the surface structure of the soil crust. When the crust is wetted, it becomes soft and impacting raindrops can create a rough surface depending on the type of the crust (Prasse, 1999, p.109). Over crusted surfaces in the interdune corridors (site B1) material transported by rainsplash made up more than 20 per cent of the transported mass. The surface crust does not appear to protect the surface from dislodgement of grains by raindrop impact, as the mass of transported material was similar to splash transport over non-crusted surfaces at site B2 (see table 3.3, p.91). This contrasts with the significant protective role played by the crust under dry conditions. However, the morphological significance of rainsplash is negligible: the frequency of high intensity rainfall is low, the total transported mass during an event is small and the transport width is low. The latter effect is enhanced by the fact that during precipitation, wind speed was often observed to drop below saltation threshold for dry conditions.

The effect of rainfall on sand movement during the research period was restricted to the actual event. The drying of the uppermost sand layer led to a quick resumption of sand movement after precipitation. Keeping in mind that the years 1997 to 1999 experienced below average annual rainfall, it appears possible that in significantly wetter years, the higher number of rain events will have a more pronounced influence on sand movement. Either directly through increased cohesion between sand grains or indirectly through increased germination of plants due to the higher amounts of precipitation, which in turn leads to a decrease of near surface wind speed caused by the increased vegetation density.

4.2.2.2 Microphytic surface crust

Microphytic crusts are considered as the first colonisers of severely disturbed lands (Eldridge & Greene, 1994; McKenna Neuman *et al.*, 1996). Several different types of microphytic crust are observed and described for Nizzana (Yair, 1990; Verrecchia *et al.*, 1995; Kidron, 1995; Prasse, 1999). The crust has been listed as vegetation in the introduction chapter, which is justified because it is a biological feature in the area and not an abiotic surface crust formed for example from fines by raindrop impact. However, its appearance and the way it influences the aeolian activity in the research area is more that of a surface property. The crust does not act as a roughness element which considerably influences near surface wind speed as does the vascular vegetation. Independent of its individual appearance, all crust types described for Nizzana act in a similar way: they protect the underlying sand against the forces of the wind by armouring the surface.

Figure 4.1: *Leatherman* trap and surface crust after rain, 31/1/97. Patches of loose sand (light areas) on wet crust.

Field measurements showed that over crust covered surfaces sand transport is negligible. All traps at site B1 and trap A5 at the upwind end of site A caught only minor amounts of material. No saltation was ever detected by the *saltiphone* at site B1. The experiment on crust strength conducted in the wind tunnel (see 2.2.4, p.36) showed that mobilisation of a significant amount of sand in the interdune corridor is unlikely under present wind conditions. Wind speeds of 1.5 times the maximum $u_{0.24}$ at site B1 (8 m s^{-1}) measured during the research period were necessary to initiate saltation during the experiment, even at a disturbance caused by a motorcycle tire (see figure 3.49, p.101). The results of the experiment are in accordance with published data on the resistance of microbial crusts against deflation at high wind speeds of up to 19 m s^{-1} (McKenna Neuman *et al.*, 1996; McKenna Neuman & Maxwell, 1999) and the increase of susceptibility to erosion upon disturbance of the crust (Belnap & Gillette, 1998; Leys & Eldridge, 1998).

The wind tunnel experiment was undertaken with a young crust – two years of growth – which provides only little resistance against mechanical distur-

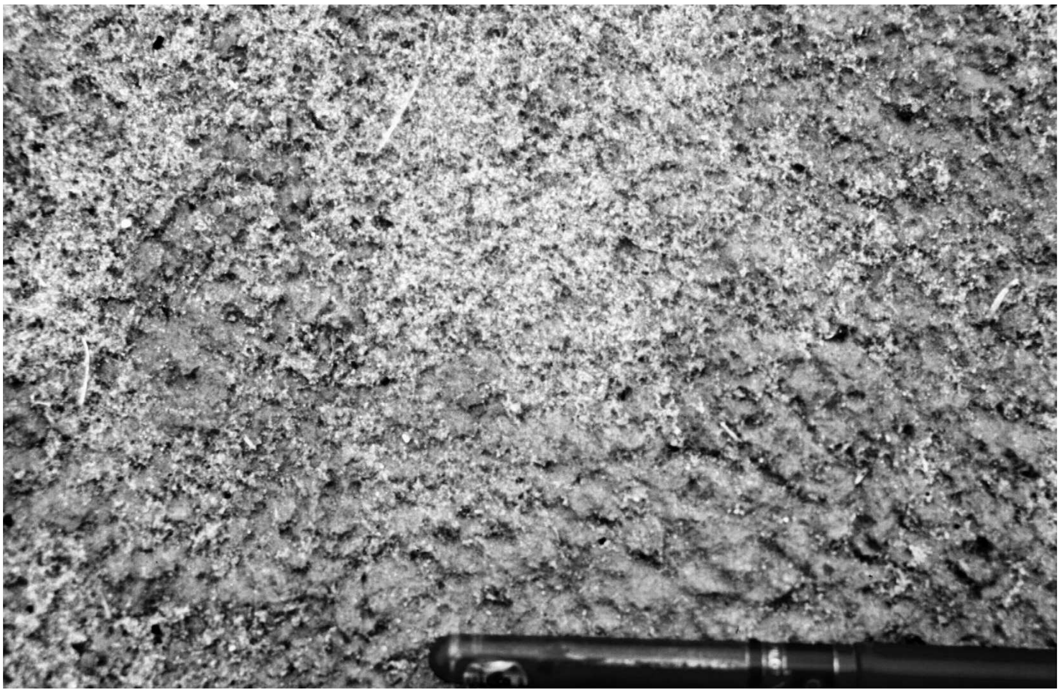

Figure 4.2: Loose sand particles on wet surface crust. Detailed view, 31/1/97. Rough structure of the surface caused by raindrop impacts. Loose sand grains moved by splash.

bance. Crust covered areas which are resistant against disturbance by walking are present at the research site (own observation and Prasse, 1999), therefore once a crust cover has been established, other factors in addition to wind are needed in order to destroy it. The wind tunnel experiment showed that disturbances are areas where erosional processes find a weak point for further destruction of neighbouring crust areas. Natural causes for disturbances of the crust are wildlife, such as burrowing animals or porcupines, which destroy the crust when digging for food. Abiotic factors such as high intensity rain events can lead to surface runoff on crust covered sandy surfaces (Yair, 1990; Kidron, 1995), causing erosion of rills and small gullies with subsequent depositional fans at their ends. Such small scale disturbances are the source of patches of loose sand present on the crust. Once the material has been loosened from the crust, it can be transported across the crust covered surfaces. In turn, such saltating particles are considered to cause a weakening of microbial surface crusts which could finally lead to their destruction (McKenna Neuman *et al.*, 1996; Rice *et al.*, 1997; McKenna Neuman & Maxwell, 1999; McKenna Neuman & Maxwell, 2002).

The results of the sand traps installed at site B1 prove that the amount of saltating particles over crust covered surfaces is very low. Given the low frequency of sand transporting events at the non-crusted sites (A, B2) at similar topographic positions, abrasion processes by saltating grains will not have a considerable weakening effect on the microphytic crust under present conditions, even if their amount would increase. In addition, the regrowth rate of the crust over disturbed areas is high, due to input of aeolian dust (Littman, 1997; Littmann & Gintz, 2000; Yair & Verrecchia, 2002). This input is also likely to counter any loss of loosened material. At the open sand surfaces of site A and B2 initial stages of crusts have been observed regularly after wetting of the surface by precipitation. Without repeated artificial destruction of the surface, a weak crust cover would have developed within a single winter season over the main part of these areas (Yair & Verrecchia, 2002). Even if a process of abrasion of the surface crust is currently taking place at Nizzana, its effects are countered by the regrowth rate of the crust.

Because the microbial crust covers all plinth and interdune corridor surfaces, these areas can, under present conditions, not be a source of sand for dune building. Even during the strongest event recorded during the study (March 1998), no evidence of significant deflation in the interdune corridors or at the plinths of the ridges was found. The importance of the microphytic crust for the protection of the surface against deflation increases during periods of drought, as it provides protection of the soil even when vascular vegetation cover is sparse (Eldridge & Greene, 1994). The microbial crust must be regarded as the most important element for the stabilisation of the sand surface.

4.2.3 Vascular vegetation

Aeolian sand transport requires open sand surfaces as a source of transportable material. Any vegetation cover reduces the open area and thus the amount of available material as has been shown earlier for the microphytic crust. Vascular vegetation also reduces near surface wind speed and thereby leads to a reduction of sand transport rates and total transported mass. The setup of the measurement sites of the study include areas of different vegetation cover, representing typical locations of the plant communities determined by Tielbörger (1997). Over large parts of the interdune corridors, the distribution of perennial plants can be regarded as homogeneous from a purely aerodynamic view, notwithstanding the fact that a heterogenety in the distribution of dominating species has been found (Tielbörger, 1997, p.270). At the active dune ridges vegetation is concentrated at isolated hummocks, representing single obstacles to airflow, which in turn cause inhomogenuous flow conditions.

In the interdune corridors, the present natural vegetation cover of 17 per cent at site B2 has been shown to be insufficient to stop sand transport completely if the underlying surface crust is removed. Considering only the transported mass per unit width, the presence of vascular vegetation at site B2 led to a reduction to 6.9 per cent of bare surface values (table 3.5, p.94). A more detailed view revealed that two exceptional events were responsible for the bulk (94 per cent at site B2) of the transported mass between November 1997 and January 1999. During average winter storms the vegetation at site B2 reduced

Figure 4.3: Hummock N1, seen from W, 12/1/98. Emerging winter crestline visible in the background.

transported mass to about 1.0 per cent compared to the non-vegetated interdune surface of site A. The storms of exceptional magnitude increased the transported mass to 8.5 per cent of site A. This exceeds the increase during average storms caused by a reduction of the vegetation cover. Average winter storms transport 7.7 per cent of the mass at site A if the vegetation cover at site B2 is only 9 per cent.

The steep increase caused by the removal of about 50 per cent of the vegetation is attributed to a change in the near surface flow regime. According to Wolfe & Nickling (1993) the roughness concentration L_c determines the flow regime. A vegetation cover of 17 per cent, corresponding to a value of 0.096, results in 'wake interference flow', where the wakes caused by the roughness elements are superimposed, leaving only isolated areas where the near surface flow may act unrestricted onto the surface. 'Isolated-roughness flow' is caused by a cover of less than 16 per cent, or $L_c \leq 0.082$, leading to fully developed, isolated wakes in the lee of obstacles and leaving a large percentage of the surface open to unobstructed flow. Thus the step from one flow regime to another caused a steep increase of sand transport. In addition Ash

Figure 4.4: Hummocks N1 (centre, front) and N2 (centre, left), seen from W, 19/10/99. Situation at the end of summer. Note the deep blowouts between the hummocks. See figure 4.3 for comparison.

& Wasson (1983) claim that a plant cover of 5 to 10 per cent with plants having a diameter equal to height causes acceleration of near surface wind speed to 130 per cent as flow is diverted around the obstacle. As the plant geometry at site B2 fulfills this requirement, increased flux may in part be a result of such an accelerated flow.

As the natural cover of 17 per cent was close to the threshold for different flow regimes, it appears likely that during exceptional events, as in March 1998 and February 1999, 'isolated roughness flow' developed, thus explaining the exceptional amounts of transported mass during these storms which exceeded values expected for 9 per cent cover. The general results of the influence of vegetation cover are in accordance with other published data: a rapid increase of sand transport was found by Marshall (1973) if cover drops below 15 per cent, while Wiggs (1993) claimed 14 per cent as a threshold, above which only negligible transport occurs. As in the study of Lancaster & Baas (1998), an exponential decrease of sand flux related to vegetation cover was determined for the interdune corridors of the research site. Although absolute values differ, most likely caused by the different sediment and vegetation, the general trend is similar.

Near surface horizontal flow velocity $u_{0.24}$ has been used as the variable determining the intensity of sand movement. Differences in vegetation cover did not lead to differences of threshold speed $u_{0.24t}$ above which saltation starts over non-crusted surfaces, yet actual values of $u_{0.24}$ are dependent on vegetation

Figure 4.5: Hummock N2, seen from E, 10/5/98. Taken after rainfall. Ripple marks accentuated by different moisture content. Note lee side accumulation.

cover. The influence of the vegetation cover on $u_{0.24}$ was well documented during the storm of 15/3 - 19/3/98 (section 3.4.1.2, p.62). During phases when saltation threshold velocity was surpassed at sites A and D and wind direction was sub-parallel to the dune ridges, $S_{0.24}$ was below 0.7 at site B, while at the same time it was above 0.8 at site A. During these periods values of $S_{1.77}$ at sites A and B were similar to $S_{0.24}$ at site A, therefore the difference of the near surface wind-speed must be caused by the vegetation at site B2. It was also shown for the period of measurements that $u_{0.24}$ increases if vegetation cover decreases (see figure 3.43, p.96), while wind-speed determined above the vegetation canopy did not change.

Values of mean sand flux during a saltation event and $u_{0.24}$, measured within the vegetation canopy, show a good correlation (see figure 3.42, p.95) which is independent of vegetation cover; at least up to a vegetation cover of 17 per cent. If the relation between $u_{0.24}$ over unvegetated surfaces and different vegetation cover for the same event is considered, it becomes evident that the gradient of the linear relation depends on the vegetation cover (figure 3.43, p.96). An influence on $u_{1.77}$ has not been found, which stresses the importance of measurements within the vegetation canopy. This also confirms the results of Lee (1991), who concluded that evenly spaced vegetation creates an aerodynamically smooth 'surface' above the tops of the vegetation.

The increase of transported mass as vegetation cover decreases is partly due to the increase of transport duration. Saltation threshold velocity is reached earlier during an event and events of lower magnitude cause sand transport as

Figure 4.6: Hummock N3, 20/2/97. Isolated feature on the northern fringe of the dune crest. Compare to figure 4.7, p.149.

cover decreases. Although no strong relationship has been found, a trend is recognisable. Under 17 per cent cover, no event shorter than 5 hours at site A caused sand movement at site B2, while under 9 per cent cover saltation was recorded during events of less than 1 hour duration at site A.

The flux rates determined at the interdune corridors have to be viewed with caution as all non-crusted sand surfaces were results of repeated artificial disturbance. No upwind sand source other than the surface within the limited disturbed experimental area exists, as the results at site A (trap A/1) and site B1 showed. Thus a depletion of transportable material must be expected, resulting in decreasing flux rates due to a lack of material. It is most likely because of the generally higher flux that this phenomenon is more pronounced at site A than at site B2. Despite regular use of a harrow to equalise the surface of site A, several storms uncovered underlying, weakly cemented sand layers, so that parts of the area were not covered by loose sand.

Even though considerable aeolian activity at the non-crusted interdune corridor sites was detected, regular manual destruction of vegetation was necessary to prevent the re-establishment (A) or an increase (B2) of vegetation cover after crust removal. In addition to germination and growth of vascular plants, microphytic crust growth was fast (see above, p.144). The mobility of the interdune surfaces, found to be important for the distribution and abundance of annual plants (Kadmon & Leschner, 1995), is currently too low to prevent the reestablishment of vegetation, be it microphytic or vascular.

The general flow regime at site C is governed by the small scale morphology

Figure 4.7: Hummock N3, 3/4/98. Northward movement of the crestline has incorporated the isolated feature of figure 4.6, p.148 into the dune crest.

(see above, p.136). This relief is a result of the distribution of vegetation during a more active stage of dune development. Since then, erosion of material between plant patches has led to the 'growth' of hummocks. Between these, vegetation cover is low, thus within the narrow gaps, 'isolated roughness flow' conditions must be assumed. The hummocks themselves are almost completly covered by organic material. A high percentage of the vegetation appeared to be dead during the study period, but still offering protection against erosion. Erosion was observed at the base of the hummocks. Very high wind speeds in March 1998 were able to destroy the organic cover of several hummocks and subsequently the features were eroded. Germination and survival of annual plants were noted, yet no increase of vegetation density or crust cover was observed. At the open sand surfaces germination was considerably less intensive than at the interdune corridor. This could be an indicator of higher mobility of the surface at site C, caused by the more exposed position, but reduced availability of moisture at the site may play a role, too (Kadmon & Leschner, 1995).

The distance between obstacles within the wind field is responsible for the effect of the obstacle on the flow. The distance between vegetated hummocks at site D varied between 5 m to 15 m. Such singular obstacles lead to an increase of flow velocity in their immediate vicinity unless their porosity is high (Ash & Wasson, 1983, p.18). The majority of the perennial vegetation at site D was situated on the upper part of mounds. At the level of saltation transport, airflow was forced around the obstacle. This effect led to increased erosion at

Figure 4.8: Hummock N5, seen from S, 2/1/98. The feature had been partly destroyed by wind action.

Figure 4.9: Hummock S2 (foreground), seen from E, 14/12/98.

the windward side and at the flanks of the mounds. Material was removed, creating hollows around the plant where roots became visible. In the lee of the hummocks deposition occurred due to a slowdown of flow.

Single porous plants or patches can also act as traps for saltating grains (Ash & Wasson, 1983, p.18). At site D this was observed at patches N1 (figure 4.3, p.145) and S2 (figure 4.9, p.150) during the early stages of the study. The bults of *Stipagrostis scoparia* were level with the surrounding surface, thus acting as porous obstacles. As deflation of the surrounding surface continued, the hummocks 'grew' and were transformed into non-porous obstacles, prone to erosion at their flanks (figures 4.4, p.146 and 3.68, p.117).

The hummocks located on the northern side of site D are ambivalent in their influence on dune morphology. On one hand, they act as accretion focus and protect the area in their lee, on the other hand, erosion is enhanced in the gaps betwenn the hummocks, especially during spring and summer. When the sea-breeze becomes the dominant sand transporting wind, the destruction of the winter crestline starts in the voids between the hummocks. Deep blowouts are formed with deposition areas immediately downwind (see 3.57 p.109). The areas least affected by deflation at the dune crest are characterised by the presence of small shrubs with a high porosity. At site D these where the NE and SW corner of the area monitored with erosion pins.

Vegetation cover and surface stability have been shown to be interdependent (Wiggs *et al.*, 1994; Kadmon & Leschner, 1995). A dense vegetation cover will protect a surface from the influence of the wind and in turn it will be more stable than a surface of the same material under identical wind conditions but less vegetation cover. The influence of the patchy vegetation at active dune crests is restricted to the vicinity of large, well established hummocks and the transition zones between crest and footslope. Germination was common at site D during calm periods after rainfall, as reported for similar dune crests by Kadmon (1994, p.139f). Yet the survival rate in the open, mobile areas of the dune crest was extremely low (Prasse, 1999). In the interdune corridor, observation at the sand accumulation east of site A confirmed the fact that desert dune sand is very favourable as a substrate, given that it does not move, gets enough water and nutrients (Tsoar, 1990; Prasse, 1999). Germination was significantly more intense than at the neighbouring, crust covered areas. Setting aside water and nutrients, surface stability is the key factor in Nizzana determining plant community distribution (Kadmon & Leschner, 1995; Tielbörger, 1997).

4.3 Current condition and spatial distribution of sand transport

During the current project, significant aeolian sand transport was restricted to non-crusted surfaces. Undisturbed crust-covered surfaces experienced only sporadic sand movement during exceptional events in winter. Amounts were too low for a calculation of sediment flux. At disturbed, ie non-crusted interdune corridor surfaces, saltation occurred regularly during all cyclonic winter storms. Apart from the surface conditions, the magnitude of sand transport in the interdune corridors depended much on the nature of the storm. The highest average flux at site A in relation to site D during an event was 85 per cent. This event on 18/3/98 (see figures 3.19, p.65 and 3.20, p.66) was characterised by only moderate wind speeds, but a long duration and a slow change of wind direction. Absolute flux at the dune crest reached average values of winter storms. The highest flux values in the corridor were reached during dune parallel flow. Earlier, when wind direction was perpendicular to dune orientation, flux in the corridor was extremely low. As a result transported mass at site A was only 32 per cent of site D. On the preceding day, wind speed was higher and wind direction during the main phase of the event had a strong SW component. Absolute flux values at the dune crest were more than three times higher than on the 18/3/98. Average flux at site A reached only 65 per cent of site D, but transported mass was 40 per cent of site D compared to 33 per cent on the 18/3/98. During average winter storms, average flux reaches approximately 20 per cent of site D in non-vegetated interdune corridors (site A) and between 2 per cent and 5 per cent if vegetation cover in the corridor is 17 per cent (site B2). Tranported mass amounts to up to 10 per cent of dune crest values at site A and approximately 1.5 per cent at site B2.

The distribution of the surface crust correlates with relative height. The area of open surfaces increases towards the crests of dunes. Sandy surfaces in the interdune corridors are completely covered by a microphytic crust except for local disturbances caused by human activity, wildlife or erosion by running water (Yair, 1990; Kidron, 1995). Reduction of crusted areas can also be correlated with an increase of surface activity. A similar dependency exists for vascular vegetation (Kadmon & Leschner, 1995). The lack of a surface crust on otherwise undisturbed surfaces must be regarded as an indicator for high surface mobility. Artificially disturbed surfaces at sites A and B2 showed initial stages of crust growth within a short period of time unless further manual disturbance was caused. Dune crests are the most unstable areas if the grade of surface stability is determined through an inspection of the presence and nature of surface crusts. Interdune corridor surfaces mark the opposite end of the scale. All results of the current study lead to the same conclusion, thus confirming earlier studies (Kadmon & Leschner, 1995). An explanation for this effect are the findings of McKenna Neuman *et al.* (1997), who report a gradual increase of shear stress between footslope and crest of a non-crusted dune, leading to an increase of sand flux q (see also p.134). This, in turn, aggravates the establishment of a surface crust or the germination and survival of vascular plants. In general, high and exposed places facing the tempest direction are most affected. At Nizzana the crests of high ridges – if viewed as a morphological unit – must be regarded as active, confirming the results of earlier studies on similar dunes (Ash & Wasson, 1983; Wiggs *et al.*, 1995). Aeolian processes cause considerable transport and surface change at these exposed surfaces. The upper flanks of active dune tops are the areas where most of the change takes place. The highest mobility was determined along the axis of the dune and in the vicinity of well established phytogenic hummocks. On the other end of the spectrum, crust covered interdune surfaces, in combination with a cover of vascular vegetation must be regarded as inactive. Although sand movement occurred over such surfaces, the amount was too low to be detected by the *saltiphone* and is therefore considered to be too low to have any signifance for the development of the morphology of the interdune surfaces. A negative effect on vegetation, either mechanical through grain impact or through the depletion of nutrients, is ruled out. The number of grains is too low to cause harm on plants and an input of nutrients through aeolian transport was shown in other studies at the site (Littman, 1997; Littmann & Gintz, 2000). The material moved over stabilised interdune surfaces originates in disturbances caused by wildlife activity (eg porcupines, gazelles, ants, bees). This kind of natural disturbance has not been observed to lead to a feedback process, where aeolian erosion, acting at the fringes of such disturbances, causes widespread crust destruction. Small openings in an otherwise undisturbed crust cover usually heal after a short time and do not provide an access point for wind induced erosion or deflation. Only where continued efforts were made to keep the crust from growing on the surface – like at sites A and B2 – aeolian sand transport could be observed in the interdune corridors.

The transition zone between dune crest and interdune corridor, the dune plinths or footslopes, was not included in this study, because the necessary regular access to check sand traps and other equipment would have altered the conditions by destroying the surface crust. Visual observation indicated that inactive and active parts can be distinguished by the presence or absence of a surface crust. At the upper parts of the footslopes, crust strength is considered to be too weak to withstand exceptional storms if located on the windward side. On the lee side, avalanching sand was observed to occasionally cover stabilised, crust covered surfaces. This phenomenon is widespread on the southfacing slopes, causing a sharper border between active and inactive surfaces than on the northfacing slopes. The more frequent and unidirectional sand transport during summer is held responsible for this phenomenon.

At already degenerated dune ridges sand transport is also confined to the uppermost areas. Open, non-crusted areas are highly correlated with local morphology. Crust cover is lowest in areas which could be identified as remnants of the secondary dunes identified by Tsoar & Møller (1986). In the shallow depressions separating these positive forms, a usually thick crust prevents aeolian sand movement. Within the positive forms, which are mainly oriented NW-SE, open sand, vulnerable to aeolian processes, is located between vegetation covered hummocks. Therefore aeolian sand transport is possible only on a low percentage of the dune ridge. The relation between crust covered and open areas changes in favour of the open areas as dune height increases. A gradual transition from an inactive, crust covered ridge to a dune with an active crest similar to site D could be observed in lateral direction from W to E at the dune of site C.

The results of this current study are in accordance with the findings of Ash & Wasson (1983), who claimed that 'significant movement only occurs on the crests' of linear dunes under similar climatic conditions as in Nizzana.

Conclusion

Aeolian sand transport at Nizzana is governed by two distinctive wind regimes: During winter, high magnitude winds of cyclonic nature dominate. Main wind directions are S to W, changing clockwise during an event. During periods of low atmospheric pressure gradients, autochtonuous diurnal southerly breezes of low magnitude develop. Summer is dominated by diurnal NW-winds of moderate strength, caused by the temperature difference between the land surface of the Negev desert and the water body of the Mediterranean. These winds show little variance of magnitude and direction.

The interdune corridors are inactive, ie no significant saltation transport of sand was observed as long as the microphytic surface crust remained undisturbed. Current wind velocities are unable to destroy an established crust cover. Mobility of open sand areas was found to be insufficient to prevent the regrowth of a surface crust or the germination and survival of vascular vegetation. Therefore if the natural condition of the surfaces is left unchanged, sand movement is restricted to the uppermost parts of dune ridges. The grade of activity is dependent on dune height. The highest parts of the ridges are the most active areas. The high magnitude storms of winter cause a high turnover of material at the active dune crests and affect all non-crusted areas of a ridge. The southerly component of the storms causes transport from the lower part of the crest to the uppermost central part. The sediment budget of the crests remains balanced.

The seasonal change of the main sand transporting wind direction changes the appearance of the active dune parts as they adjust to the prevailing sand transporting wind. During the transition periods of autumn and spring, the previous steady state form is destroyed and changed into the new form. The summer form is caused by high frequency – low magnitude events: a seabreeze from NW dominates sand transport once the duration of sunshine is long enough to cause sufficient heating of the desert surface above the water temperature of the Mediterranean to initiate the airflow. The breeze causes a convex assymmetric profile along the main wind direction at the active parts of the dune ridges. The interdune corridors are sheltered even from the strongest summer winds. Threshold flow velocity $u_{0.24t}$ is not surpassed in the interdune corridors during summer. The winter form is caused by low frequency – high magnitude events: the cyclonic storms associated with the passage of low pressure cells over the Eastern Mediterranean are of short duration, but due to their high wind speeds they are able to transport large quantities of material at the dune crests. During the storm season, a sharp edged, meandering crestline is formed along the centre of the dune crest. The high wind speeds in connection with the SW to W directions can cause sand transport in the interdune corridors if the microphytic surface crust is destroyed previously. The threshold velocity for bare sand is passed regularly in the interdune corridors during cyclonic storms. The periods between the passage of two low pressure cells are usually dominated by calm weather with very low wind speeds. If the temperature gradient between the cold desert surface and the warm Mediterranean is sufficiently steep, a SE-breeze may develop which can reach threshold speed at the uppermost parts of the dune ridges.

The surface of the interdune corridors is protected by the combined effects of an organic surface crust and perennial vascular vegetation. Their presence prevents aeolian sand movement. The corridors are also sheltered from all winds other then those parallel to the dune ridges. Only if the microphytic surface crust is disturbed or destroyed can aeolian processes act on the underlying sand. Vascular vegetation without underlying microphytic crust cover is not sufficient to halt saltation completely. Small scale disturbances of the microphytic crust do not lead to a feedback process resulting in large scale destruction as the regrowth rate is fast, while the occurrence of storms is limited. However, repeated destruction of the surface crust by off-road vehicle traffic, grazing of sheep or goats or intensive walking over large areas may lead to such a development. The frequency and magnitude of sand movement on the dune crests prevents the establishment of a dense vegetation cover. Seedlings are likely to be covered or uprooted by the shifting of sand. Vegetation spots already present are modified by erosional processes at their base. Deposition at their lee side is also common, but due to a lack of material, deflation prevails.

The stabilization of the linear dune area originated in the interdune corridors. The generally low wind speeds of the area in connection with the sheltering of the interdune corridor from all summer winds were a favourable environment for the quick establishment of a surface crust and vascular vegetation after the end of continuous destruction by grazing as common before 1982. Increasing surface cover caused further reduction of sand movement and ultimately leads to an inactive surface. Because airflow is accelerated at the windward slopes surface mobility increases with dune height. Therefore the upper part of the dune ridges are subjected to higher wind speeds. The higher mobility of the surface makes the dune crests less suited areas for the establishment of vegetation.

The gaps cut for the patrol roads along the border and their continued clearance on Egyptian territory currently restrict the eastward movement of sand. As linear dunes are mainly sand passing forms, the dunes east of the border are thus cut off from their main sand source. The active dune crests are depleted of sand through continuous eastward transport. In addition, every quantity of material, however small, which is transported off the crest onto the plinth or into the corridors will stay there under current conditions. It is very likely to be incorporated into the surface crust, where it will be fixed until mechanical destruction of the crust. This continued decrease of available material for aeolian transport leads to reduced mobility and thus increasing vegetation cover at the dune crests.

Under present day environmental conditions, a complete stabilization of the linear dunes at Nizzana is not expected, if one equates ‘complete stabilization’ with the absence of any aeolian sand movement. The equilibrium state of undisturbed vegetated linear dunes under similar climatic conditions (Kalahari, Australian desert) is such that limited aeolian sand movement at the dune crests occurs. However, a further reduction of very active areas such as the crest at site D is likely as currently the dunes east of the border lack an upwind source of transportable material.

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