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Original Article

Dewfall and its Geo-ecological Implication for Biological Surface Crusts in Desert Sand Dunes (North-western Negev, Israel)

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Abstract: Dew is an important water source for biological soil crusts and lichens in arid and semi-arid ecosystems. These crusts influencing the ecosystem processes resulting in a patchy ecotope and vegetation distribution. Microclimatic boundary conditions for nocturnal wetting were determinate. Maximum activity of crust is reached a few hours later after dewfall starts when cumulative dewfall exceeds 0.1 mm at dew point temperature differences around 0 K. Different microclimatic approaches were applied to estimated dewfall amounts in sand dunes of the north-western Negev. The annual dewfall amounts obtained from the zeroplane model was 26 mm a-1 and from the load cell 33 mm a-1, whereas other models mostly overestimated the dew amounts. However, spatial differences in shading after sunrise could explain the crust pattern in the sand dunes.

Key Words: Dew, Fog, Desert microclimate, Lichens, Cyanobacterial crusts, Nizzana

1. Introduction

Fog and dew and their geoecologcial implications for biological systems and ecosystem processes has been a relatively neglected topic in arid ecological research and desert meteorology, but plays an important role for ecological and physiological processes in such arid regions. In the hyper-arid coastal deserts of the Namib, Succulent Karoo and Atacama fog formed over the Benguela current respectively Humboldt Current and moving inland. In contrast to fog dew is a phenomenon where water vapour condenses on a substrate and transforms into liquid water once the saturation pressure at the temperature of the substrate is lower than the saturation pressure at air temperature (Beysens, 1995), the main source of vapour in a natural environment is the air itself. Once vapour is deposited on the surface by downward flux it will be thermodynamically stable only in case of droplet formation which requires large thermal fluctuations to overcome the cost in free energy of forming a liquid-vapour interface, i.e. to overcome interfacial tension (Beysens, 1995).

Characteristic for these arid ecosystems, especially in the costal fog deserts, are a large cover of lichens and soil crusts. These poikilohydric plants are able to use dew and fog for activation of their physiological activity. The importance of dewfall as a water source for higher desert plants has been discussed controversially (Waisel 1958, von Willert *et al.* 1992). Biological soil crusts, which are building up by cyanobacteria, green algaes, mosses, lichens and fungi, are important communities in various arid and semi-arid regions and have an important impact on the geo-ecological processes (Belnap and Lange, 2001; Veste, 2005). The biological crust influencing (i) infiltration and other hydrological conditions, (ii) stabilizing topsoil and reducing soil erosion, and (iii) enhancing the nitrogen pools by biological nitrogen fixation (Yair, 1990; Littmann *et al.*, 2000; Veste *et al.*, 2001a; Russow *et al.* 2005). These results in a small-scale ecotope and vegetation mosaic influenced by the spatial pattern of soil crusts. The spatial distribution within the ecosystem must be influence by micro scale ecological processes, e.g. dewfall and evaporation from the surface. Several studies regarding dewfall were already conducted in Israel. However, most of these studies used artificial surfaces to measure the dew deposition (e.g. Zangvil, 1996; Kidron, 2000).

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In the ecological context there is an important need to get a better understanding of the microclimatic processes of dew formation and their interaction with biological systems to improve the modeling of dewfall. In this paper, we investigated the microclimatic boundary conditions leading to activity of the biological soil crusts in the sandy northwestern Negev in Israel. For determination of dew amounts we apply a modified Zeroplane model approach (Littmann and Veste, 2006) and compare them some of the most common approaches to compute dewfall from field measurements. The spatial dewfall pattern will be related to the ecosystem processes and ecotope pattern.

2. Materials and methods

2.1. Study area

The study was carried out in the sand dunes of the northwestern Negev (**Fig. 1**). These sand dunes are the eastern most of the sand field covering the northern part of the Sinai Peninsula and the northwestern Negev (Veste *et al*., 2005). The local climate is determined by a sharp gradient from the Mediterranean coast to the arid climate of the Negev. The rainfall season is limited to the winter season (October to March) and average annual rainfall decreases from around 170 mm at the northern edge of the sand field to 90 mm near Nizzana. At the Nizzana test site (SD; 34° 23´ E, 30° 56´ N), the sand dunes form linear dunes with east-west direction, whereas north of the Nahal Nizzana the dune are a combination of barchanoid and longitudinal dunes. Linear dunes with heights of 8 - 18 m with east-west directions are typical forms in the southern parts of the Negev dune field. The northern experimental site (ND) is located about 33 km inland (34° 20´ E, 31° 2´ N) and is characterized by stable sand dunes. The interdunes at this site are covered by the soil lichens *Fulgensia fulgens*, *Squamarina cartilaginea*, *S. lentigera*, *Diploschistes diacapsis* and *Collema tenax* var. *vulgare* (Büdel and Veste, in press).

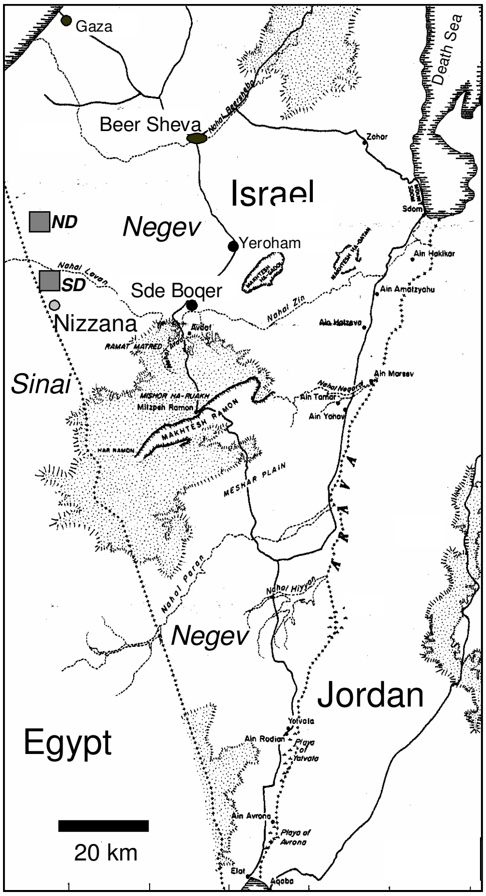

Fig. 1. Location map of the investigation sites in the north-western Negev, Israel: ND stands for Northern dunes, and SD stands for Southern dunes near Nizzana.

2.2. Microclimatic measurements

Microclimatological measurements were carried out at two recording stations on the encrusted parts of dune slopes facing north and south from July 1995 to June 1996. A field experiments to interrelated microclimatic boundary conditions and crustal activity was carried out in March 1999. Measured parameters (hourly means) relevant for this investigation were recorded at a mast supplied with a Campbell CRX10 data logger (Campbell Scientific, Logan, Utah, USA). Air temperatures at 2 m and 0.2 m above ground were measured with Pt 100 thermo resistor probes (Campbell 107-L, Campbell Scientific, Logan, Utah, USA), specific humidity was calculated from relative humidity measurements at 2 m (Vaisala HMP35C-L, Helsinki, Finland), net radiation data at 2 m were gathered using a REBS Q7.1-L net radiometer, soil heat flux at a depth of 0.15 m was measured with a heat flux plate (REBS HFT3-L). The leaf wetness sensor is a plate with an electronic resistance and indicates either a dry (no signal) or a wet surface (max. electronic signal). As relative humidity was measured only on the north-facing slope, we will present results for this site except of the eddy correlation technique, which was applied to the opposite slope where also vertical wind speed was available. This conventional arrangement of field measurement necessarily leads to some inaccuracy that has influence on the model results once gradients within an air column are introduced, e.g. critical temperature and vapour gradients between 0.2 m and the actual surface have to be neglected. However, most of the methods applied here derive data from standard measurement heights. As we had to compute relative humidity at 0.2 m from humidity data at 2 m without an additional measurement at the lower height, all results including gradients will possibly underestimate the actual situation to a certain extent.

2.3. Physiological activity of biological crusts

A modulated fluorescence system (Heinz Walz GmbH, Effeltrich, Germany) with a 6 mm diameter standard fiber optic was used for the measurements of the lichen activity (see Schroeter *et al*., 1992, Veste *et al*., 2001b). During the night a red light beam (680 nm, < 1 µmol m^{-2} s^{-1}) was operated continuously so that both the steady state fluorescence yield Fo in the night and F during daytime could be determined. The steady state fluorescence yield Fo signals were converted in relative units (maximum signal = 1.0). The fiber optic was fixed at a distance of 10 mm from the surface of lichen and at an angle of 60° to avoid shading the lichen thallus. The measuring system was placed 2 m away from the microclimatic installation in the northern dune system (ND, Fig. 1).

2.4. Microclimatic approaches

Several flux models were developed to describe evapotranspiration in various ecosystems. E.g. the Thornthwaite-Holzman method expresses upward flux in terms of positive values while negative values denote a downward vapour flux. It is this turbulent downward flux that leads water vapour to the surface where it may condensate under the appropriate boundary conditions. In this way, turbulent downward movements may be considered potential dewfall. However, any approach to infer dew deposition from evapotranspiration models depends on how condensation at the surface is considered potentially possible.

Thornthwaite and Holzman's (1942) gradient-flux model considers evapotranspiration as the upward (positive) flux of specific humidity between two heights above ground enforced by the vertical wind shear:

$$ETP = \rho k \frac{(q_{z1} - q_{z2})(u_{z2} - u_{z1})}{\ln\left(\frac{z_2}{z_1}\right)^2} \quad \text{(Eq. 1)}$$

where u_1 and u_2 is the horizontal wind speed at heights z_1 and z_2; q_1 and q_2 are the respective specific humidities; ρ is the air density, and k is the Karman constant.

A direct method to estimate both evapotranspiration and dewfall is the eddy correlation technique (Swinbank, 1955; Jacobs *et al*., 1998, 1999). Based on the fact that turbulent exchange of properties in the boundary layer is triggered by the friction velocity, surface roughness, and gustiness in terms of Reynold's covariance which considers the actual wind speed as the sum of mean wind speed over a period of time and the momentary variance, the approach correlates the vertical fluctuations of wind speed and specific humidity over a certain period.

$$ETP = \overline{(\rho \omega)' q'} \quad \text{(Eq. 2)}$$

where $(\bar{\rho}\ \bar{\omega})$ is the mean vertical flux of mass, $(\bar{\rho}\ \bar{\omega})'$ and are the respective mean deviations of mass flux and specific humidity from their means over the observation period. Positive fluxes indicating evapotranspiration (ETP), whereas negative values reflect downwards fluxes.

Including the energy budget and a parameterisation of turbulent flux, the Bowen Ratio Energy Balance (BREB) technique originally presented by Sverdrup (1936) method is widely accepted as the physically most sophisticated approach:

$$L_v E = -\frac{R_n + G}{1 + \gamma \frac{\partial t / \partial z}{\partial e / \partial z}} \quad \text{(Eq. 3)}$$

where L_v is the latent heat of vaporization (2.45 x 10^6 J kg^{-1}), E is the evapotranspiration, and Lv E equals the flux of latent heat (latent enthalpy). γ is the psychrometric constant (0.66 hPa K^{-1}), $\partial t / \partial z$ is the temperature difference between two heights of measurement, and $\partial e / \partial z$ the respective vapour pressure difference. L_vE/L_v converts the flux of latent heat into actual evapotranspiration values (mm), which become negative at night in case of negative net radiation. Such cases, however, represent potential dewfall as the method does not consider a downward vapour flux but the water equivalent of latent heat in the air column between the two measurement heights.

2.5. Zeroplane model

The application of the different microclimatic approaches for the estimation of vertical atmospheric water fluxes is problematic. The Thornwaite-Holzman approach and Bowen

Ratio Energy Balance overestimated the evapotranspiration (Littmann and Veste, 2006). A major problem is the fact that vapour fluctuations resulting from evaporative or condensation processes will be much larger just above the active surface (soil or vegetation) as compared to a larger height. Using the mean specific humidity of the air column under consideration or its gradient between two heights without including air movement, as a purely empirical boundary condition will underestimate dewfall while including wind speed or wind shear as in Thornthwaite-Holzman will lead to overestimations. The eddy-correlation technique will fail in reliefed terrain when topography enforces turbulent vapour flux. Especially in arid regions advective disturbance affecting the vertical fluxes and its influence cannot be neglected easily. E.g. in the study region sea breezes transport moist air towards the inland during the summer half year (Littmann, 1997) and decrease the actual evapotranspiration as compared to the water equivalent of the radiative budget.

A modified Zeroplane model was developed for the calculation of the evapotranspiration in drylands, which give more realistic values for the evapotranspiration at the study sites (Littmann and Veste, 2006). Advective disturbances were excluded and anomalous wind speeds relative to the mean monthly diurnal course of wind speed (roughly 10% of the overall series) were not considered following the anomaly filter:

$$(u_{Tn} - \bar{u}_{Tn}) > \sigma_n \qquad \text{(Eq. 4)}$$

where u_{Tn} is the actual data n at time T of the series, $\bar{u}_{Tn}$ is the mean at time T, and σ_n is the standard deviation.

The Zeroplane model was developed in three steps and can also applied for dewfall (for details Littmann and Veste 2006):

1. Vapour fluctuations with in the plane of observation are expressed as a differential series of mean hourly values of specific humidity dq. In the case of evapotranspiration they are positive. Decreasing of dq reflects a condensational removal of water vapour from the near ground air or constantly decreasing of water vapour input from the vegetation or soil.
2. Dewfall is the total amount removed from the plane. The removal (or increase in the case of evapotranspiration) within the mean time series sets in relative to the prior fluctuation. This is represented by the cumulated dq series and is consistent with the principle of continuity as cumulated dq also represents the flux of latent heat through the plane.
3. For the determination of dewfall the cumulated series is integrated following (Peixoto, 1973):

$$D = z\rho_w \int_{T=1}^{T=24} f(\bar{q}_{(1..24)} - \bar{q}_1) < 0 \qquad \text{(Eq. 5)}$$

where D = is the dewfall, z is the unit height of the plane (1 m) and ρ_w is the density of water vapour. The functions of the differential specific humidity series (which show a start value of zero at 0:00 local time, thus we called "Zero plane") because it is identical with the cumulated *d*q series. Nighttime negative values (D < 0) following (Eq. 5) are subject to the critical dewpoint filter. A perfect coincidence of hours where condensation is physically possible and those where the model output indicated a decrease in specific humidity was found.

2.6. Dewfall

Dewfall amounts were additionally measured by means of an acrylic condensation plate (20×20 cm) attached to a pressure transforming load cell calibrated daily. One plate was covered with Styrofoam underneath to prevent condensation on the lower face of the plate. Nightly dewfall was calculated as the individual daily range, which excludes any influence of contamination of the plate, by sand or dust. The use of hourly means removes in fact the oscillating noise of plate movements induced by wind.

2.7. Dewfall

A shading model was applied on a digital terrain model using Surfer 4.0 (Golden Software, USA).

3. Results

3.1. Boundary conditions for dew formation

Dewfall occurs when a high-pressure ridge in the mid-troposphere causes a subsidence inversion layer to form. This provides clear skies and light winds, conducive to good surface radiational cooling, and hence dew. **Fig. 2** shows typical microclimatic boundary conditions for the dew formation in the sand dune system. After sunset, terrestrial radiation from the surface leads to large negative net radiation values over 2 to 3 hours and a respective radiative cooling of

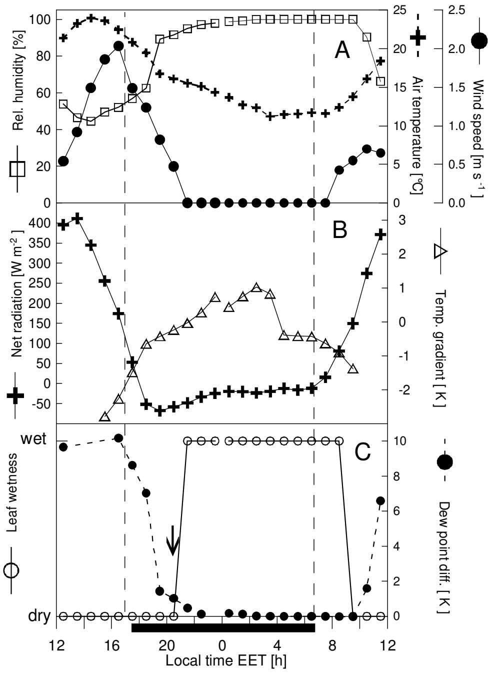

Fig. 2. Typical microclimatic boundary conditions for dew formation: Air temperature, specific humidity, net radiation, air temperature gradient between 2 m and 20 cm height, wind speed at 20 cm height, net radiation, dewpoint difference and leaf wetness sensor signal (March 08/09, 1999).

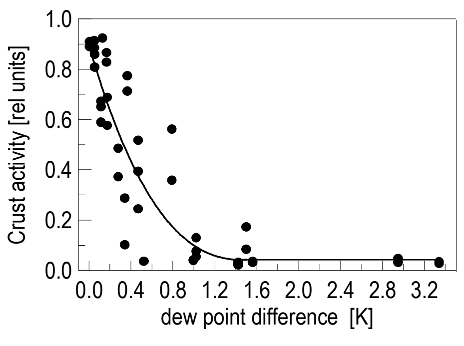

Fig. 3. Relation between differences between air and dewpoint temperature and physiologcial activity (relative units, max. steady state fluorescence yield Fo = 1.0) of soil crust lichens measured be the means of chlorophyll fluorescence during the night.

the near-ground air layer. The temperature gradient between 2 m and 20 cm height becomes >0 K, indication the build up of an extremely stable layer near ground. Once the stable layer

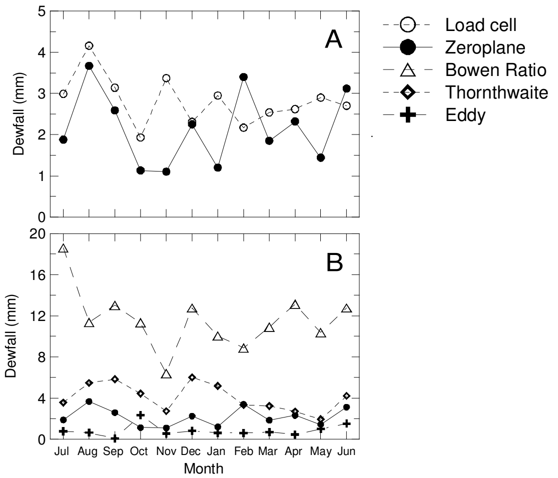

Fig. 4. Comparison of modeled monthly dewfall by the Zeroplane model and measured with a load cell (A) and after Bowen Ration Energy Balance (Sverdrup model), Thornthwaite-Holzman and Eddy correlation technique (B).

is developed, wind speeds at 0.2 m decrease towards 0 m s^{-1} (**Fig. 2B**) and even at the height of 2 m we found only calm situations or extremely light veering winds. Under such conditions the difference between air and dewpoint temperature drops below 1.0 K and the leaf wetness sensors simultaneously indicate dew condensation (**Fig 2C**). This process is consistent with a slightly increase of specific humidity, which was caused by downward vapour flux in a converging near-ground boundary layer (**Fig. 2A**).

3.2. Crust activity

Under these specific microclimatic conditions the condensation of dew leads to a re-wetting of the soil lichens crusts during the nights. Maximum physiological activity of the lichens always occurred when the dewpoint temperature difference reached 0 K (**Fig. 3**). In this situation the leaf wetness sensors indicated a cumulative wetting. However, between the response of the leaf wetness sensor to condensation and reaching the physiological response of the soil lichens very often a time delay could be observed. This indicates that a minimum dew amount had to be accumulating on the soil surface. However, in the interdunes fog was not leading to crustal activity, while moving over the soil surface without any wetting the crust surface.

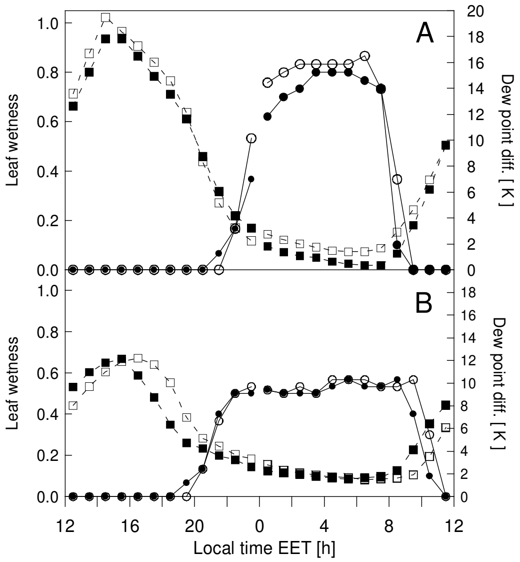

Fig. 5. Average of dewpoint differences (squares) and leaf wetness sensors (circles) on south-(closed symbols) and north-facing (open symbols) slopes in summer 1995 (A) and winter 1995/96 (B).

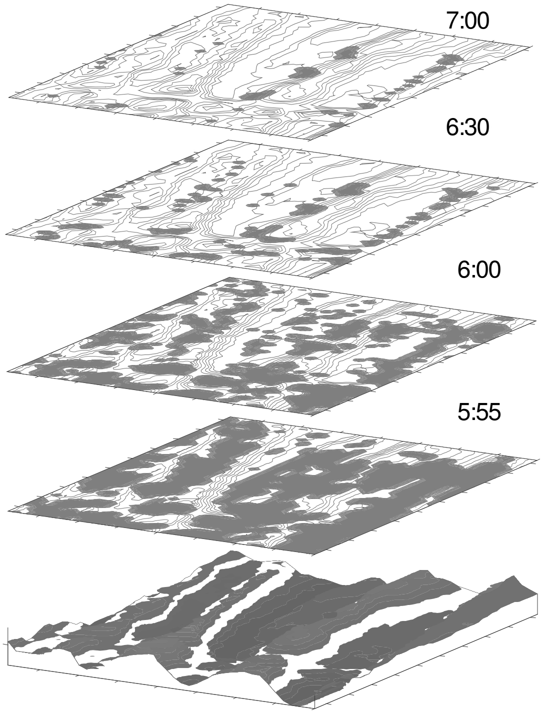

Fig. 6. Shading in March after sunrise (5:55 a.m.) until 7:00 a.m. at the Nizzana test site (SD): At 7 a.m. only the concave parts at the north-facing slopes and some parts of the dune base are shaded.

3.3. Determination dewfall

The annual dewfall in the sand dune system of Nizzana for the period from June 1996 to July 1997 was 26 mm a^{-1} using the Zeroplan approach. The monthly variations are shown in **Fig. 4.** Confirmatory measurement using the mean load cell data from two acrylic plates (**Fig. 4A**) was approximately 33 mm (two months had to interpolated due to lack of data) and is well within the range of the Zeroplane model output although the series do not show good correlation over the winter months.

The results of the different microclimatic flux approaches results in a large variation of the dew amounts. Most of the approaches showed higher values (**Fig. 4B**). The Sverdrup model (Bown Ratio Energy Balance) results in the highest monthly amounts. Also the Thornthwaite-Holzman model output is comparatively high. Both model results should be considered quasi-potential dewfall as they imply a homogeneous condensational flux of vapour over the considered air column irrespective of the application of the dewpoint filter. Only in case of small specific humidity gradients as in February to May, the Thornthwaite-Holzman model may give results comparable to gradient-free approaches. In our case, the eddy correlation technique modelled extremely low dewfall, which (different to its results for evapo-transpiration) is an effect of very low nighttime vertical wind speed in a stable layer near the surface. It is only months with nighttime convection induced by khamsinic depressions (May, October) when this model results in dewfall amounts comparable to those produced by the Zeroplane model.

3.4. Spatial variation

During summertime no differences in dew formation between south and north-facing slopes could be detected (**Fig. 5A**). However, in winter, the dune morphology leads to longer shading time of nearly 1 - 2 hours on north-facing slopes (**Fig. 5B**). This is even more pronounced in concave parts and at the dune bases (**Fig. 6**). Shading by shrubs leads even to a reduced soil evaporation and longer crust on the micro-scale, which is could be not included in shading model.

4. Discussions

It can be conclude from the field experiments that the activity of soil crusts can be clearly related to nocturnal dewfall and associated microclimatic processes in a desert environment. The combination of the in situ measurements of cryptogamic activity by chlorophyll fluorescence and its interrelations with the complex microclimatic boundary conditions will provide a tool for the calculation of long-term activity in different habitats. In this field experiment, it is clearly shown that a calculation of lichen activity only based on the dewpoint differences or even on leaf wetness sensors will not provide enough information to explain differences in crust development. Information about total amounts of precipitation inputs by dewfall, fog and rainfall are needed to get a better understanding of soil crust development in drylands. A cumulative dewfall event of ≥ 0.1 mm per night is required to reach the maximum physiological activity as it could be shown and laboratory and field conditions (Lange *et al*., 1992; Veste *et al*., 2001). This amount is also efficient to photosynthesise for 1 - 2 hours after sunrise. Average dewfall amounts per event vary between 0.06 - 0.12 mm, but strong dewfall in the Negev desert has been reported to reach values between 0.2 and 0.32 mm (Evenari *et al*., 1982; Zangvil, 1996; Kidron, 2000). The annual dewfall amounts obtained from the Zeroplane model (26 mm a^{-1}) and from the load cell (33 mm a^{-1}) are comparable to those measured by Evenari *et al.* (1982) at Avdat using Duvdevani wooden blocks (30 mm) and by Zangvil (1996) at Sede Boqer, 40 km southeast of our experimental site, using a modified Hiltner dew balance (18 mm). In this way, our model and measurement results are quite consistent with other data from the region while the other methods provide much too high (139 mm for the Bowen ratio), fairly high (48 mm for Thornthwaite-Holzman) or fairly low (10 mm for the eddy correlation technique) results. However, we did not find a pronounced annual course of dewfall as indicated by Zangvil (1996) with maxima in August to October and December and January and minima in April and November. August, September and December show also high values in our series but also February, April and June. It seems that even during frequent nighttime khamsinic situations as in April 1996 the near-ground vapour pressure was high enough to enable considerable dewfall. Extrapolating results from an extensive short-term field campaign at the Nizzana test site in September 1997, Jacobs *et al.* (1998) found dewfall around 5 mm for this month. However, they used microlysimeters on a hygroscopic substrate, which may yield a certain overestimate while their modelled results of 0.1 to 0.2 mm of dewfall per night (Jacobs, *et al*., 1999) come close to our findings.

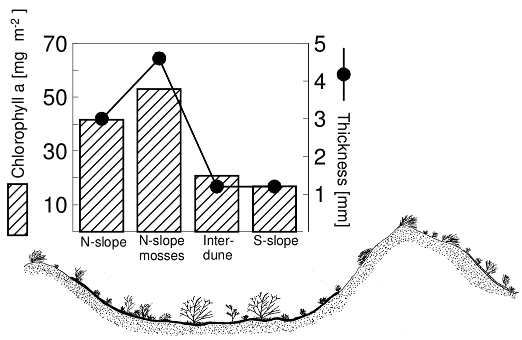

Fig. 7. Crust thickness and chlorophyll a content across the linear dune system in Nizzana from the north to the south facing slope: The thickest crust are found at the dune base and deep holes at the north-facing dune slope (after Veste *et al.* 2001b).

Small scale differences in shading, espially in winter time, seems to be an important factor determining crustal thickness and distribution in the Nizzana test site by prevention evaporation and drying out. On north-facings slopes the biological crusts are ticker (average 4.2 mm) as compared to interdunes and south-facing slopes (average 1.7 mm) (**Fig. 7**).

Locally around shrubs and at the dune base mosses occur in a high cover. In interdune of the northern dunes (ND, Fig. 1) soil lichens crusts covering large area (Veste *et al*., 2001a). It has to be pointed out that the organisms in the crust reaction different to the moisture sources. Cyanobacterial lichens and biological soil crusts are able to photosynthesised at high water content than lichens with green algae as symbiontes. Here high water content can even preventing photosynthesis by covering the lichens with a water film and reducing CO_2 diffusion (e.g. Lange *et al*., 1997). For this lichens rainfall is less efficient falling in high amounts. Therefore, it cannot drawn a generalizing conclusions of the importance of the various moisture source for the all-biological soil crust types.

This mosaic distribution of the crust types has drastic implications for geo-ecological processes within the dune system. Infiltration rates decrease with increasing crust thickness and crust biomass (Yair, 1990, 2001; Littmann *et al*., 2000) leading to run-off and re-distribution of water in the ecosystem. Surface stability is another factor determined by the soil crusts influencing the vegetation pattern in the sand dune system (Veste, *et al*. 2005). Crust distribution is well interrelated to the biomass distribution. In the concave holes the highest vegetation cover and thick biological crusts very often with mosses occurs together. This interrelation could be also shown by a spatial biomass model for the test site in Nizzana (Littmann and Veste, 2005). However, on the large scale the biological surface crust is counteractive the rainfall gradient from the northern margin (170 mm a^{-1}) the dry southern parts in Nizzana (90 mm a^{-1}) by limiting infiltration (Littmann *et al*., 2000). Unexpected, vegetation cover along the climatic gradient is negative correlated with rainfall amounts and is not changing in the interdunes (Littmann and Veste 2005; Veste *et al*. 2005). In the case, the biological crusts have an indirected effect on the vegetation patterns via the hydrological processes.

Also the physiological processes of the biological nitrogen fixation depending on a sufficient photosynthetic activity of the biological crusts. In general biological crusts containing cyanobacteria and lichens with symbiontic cyanobacteria are in several arid and semi-arid ecosystems the major paths of nitrogen input into the ecosystem and providing N for plant growth in this nutrient-poor environment (Russow *et al*., 2005; Veste, 2005).

5. Conclusions

The presented result underline the importance of dewfall for biological systems, but shows also some of the major feedback mechanisms of the microclimate on the complex ecosystems processes in desert sand dunes. The applied Zeroplane model result in realistic dewfall amounts for the arid sand dunes ecosystems with 26 mm yr^{-1}, which is an important water source for the biological crusts. Physiological activity of soil crusts lichens could be reached only after nocturnal dewfall of 0.1 mm and specific microclimatic boundary conditions supporting the nocturnal dewfall. The combination of such in situ measurements of physiological activity by chlorophyll fluorescence and its interrelations with the complex microclimatic boundary conditions will provide a tool to investigate the importance of dewfall for biological crusts in different ecosystems. However, small-scale pattern of microclimatic boundary conditions determined biological soil crusts types and have finally a major impact in ecosystem structure and functioning.

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