The Role of Dew in the Water and Heat Balance of a Bare Soil in the Negev Desert

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Nurit

The product of real love
is as mysterious as the dew
and requires a full dark night to become
yet only minutes of light to disperse

So let us keep our red mystery for the night
our real love for the darkness
and protect the dew of our becoming together
against the evaporation of the light

Peter Cooper

Abstract

Water is the main limiting factor for biological activity in arid and semi-arid environments. Any source of water in these harsh environments is, thus, very important. Previous studies indicated that dew deposition could be equivalent to 10% or more of the total annual precipitation in the Negev desert. If this is the case it would undoubtedly be an important source of water for plants and small insects. Because plants cover only a very small fraction of the surface of deserts in general, and of the Negev in particular, the study of dew deposition on bare soil surfaces is essential.

During the night, the latent heat flux towards the soil surface is very small, and therefore the amounts of dew deposition are very small as well. This poses some very special technical measurement difficulties. Various methods for measuring dew are described in the literature, most of them using artificial condensing plates with varying physical properties. One of these methods is the Hiltner dew balance, which was used for several years in the Negev desert to continuously record dew deposition. This device is very convenient and simple to use, yet its adequacy has to be proven, since the energy balance of its condensation plate is different from that of the soil surface above which it is installed. To overcome these limitations, the examination of the adequacy of a micro-lysimeter for this purpose was proposed.

The deposition of dew and its subsequent evaporation represent a diurnal cycle of water content changes in the uppermost soil layers. This mechanism of dew deposition is, however, not the only possible mechanism by which such a diurnal cycle can be observed. A different possible mechanism is direct water vapor adsorption, a mechanism that has been put forward as being an important link in the water cycle of arid and semi-arid regions during the dry season. The magnitude and the frequency of this phenomenon were not thoroughly investigated. The relative contribution of each of these two mechanisms has never been studied.

Irrespective of the mechanism (dew deposition or direct vapor adsorption) by which water may be added to the soil it is reasonable to assume that only the uppermost layers of the soil will be affected, and thus a fraction or all of the added water will be evaporate during the following day. Such a diurnal cycle of water content means that there is a corresponding diurnal cycle of latent heat flux that may affect the pattern of radiant energy dissipation at the soil surface. The pattern of the energy partitioning at the soil surface plays an important role in global and meso-scale studies and is usually integrated as sub-models in global and meso-scale models. The radiant energy dissipation at the land-atmosphere interface above bare soil is determined by the moisture content of the soil surface. It is therefore reasonable that the quality of land surface models should be judged by the accuracy with which they compute the aforementioned soil water content.

The objectives of this research were therefore:

1. Examine the correlation between the amounts of dew deposition on artificial surfaces and the amounts deposit on bare soil.
2. Describe the daily pattern of changes in water content in the upper soil layers and identify the mechanisms by which water is added to the soil (dew deposition or direct adsorption) and their frequency.
3. Assess the relative magnitude of latent heat flux density throughout the dry season, and evaluate its importance in view of meso- and global-scale meteorological models.

The research was carried out at the Wadi Mashash Experimental Farm in the Northern Negev, Israel, over a bare loess soil. For implementing the first objective, measurements were undertaken during two measuring seasons (during the dry seasons of 2000 and 2001). During both periods the incoming and reflected short-wave radiation, net-radiation, wind speed, sensible and soil heat flux were measured at a meteorological station in the vicinity of which a Hiltner balance and a micro-

lysimeter with an undisturbed soil sample were placed. During the first period, the depth of the micro-lysimeter was 15 cm while at the second period it was 55 cm.

For implementing the second and the third objectives, nine 24-h field campaigns were carried out during the dry season of 2002. During each campaign, the 100mm topsoil was sampled hourly, and the gravimetric water content distribution at 10mm increments determined. Micro-meteorological measurements included incoming and reflected short-wave radiation, net-radiation, wind speed at four levels, dry- and wet-bulb temperatures at 1m height, and soil heat flux. In addition, dew deposition amounts were measured using a conventional Hiltner dew-balance and the changes in mass of an improved micro-lysimeter were recorded. The representativity of the micro-lysimeter was assessed by comparing its surface temperature to that of the surrounding surface using thermal images acquired on an hourly basis during several campaigns.

In order to correlate between the amounts of dew deposition on artificial surfaces and the amounts deposited on bare soil, a method for estimating the actual dew deposition on the soil surface was needed. As long as their heat balance is similar to that of the surrounding area, micro-lysimeters provide an absolute reference for latent heat fluxes. Micro-lysimeters have been used mainly to estimate evaporation from wetted soil surfaces and the maximum reported depth in the literature is 12 cm. A 15 cm deep micro-lysimeter was thoroughly tested and found unsuitable for our purpose as significant differences between the soil surface temperature of the micro-lysimeter and the surrounding soil were detected. These differences can only be due to the lack of deeper layers that contribute to the upward heat flux and were found to be large enough to affect the latent heat flux towards the soil surface. Indeed, a close inspection of soil temperature profiles indicated that only at a depth of 50cm is the temperature constant for daily intervals.

An improved micro-lysimeter, with a depth of 55 cm, was built to match these findings. The micro-lysimeter was again tested by comparing its surface

temperatures to that of the surrounding soil. The differences were significantly smaller than those measured with the previous micro-lysimeter, and their magnitude was such that their effect on vertical fluxes is negligible.

In order to define the depth to which a diurnal change in the water content occurs, a series of field campaigns was carried out, in which the gravimetric water content distribution at 10mm increments was determined. These campaigns were undertaken not only to shed light on the water content distribution in the soil but also to determine the seasonal distribution of the phenomenon. Indeed, the soil surface temperature did usually not drop below the estimated dew-point temperature of the air mass and it was, therefore, to be expected that no dew would deposit on the soil surface. Indeed, dew deposition on the soil surface was not visually observed during the various field campaigns and it was concluded that in the area in which this study was carried out, actual dew deposition on a bare soil surface is probably a rare occurrence.

Even though dew deposition was not observed, a clear daily cycle of water content of the uppermost soil was detected. The magnitude of these diurnal changes was very small and the average water contents throughout the entire measurement period very low: a maximum of approximately 2% and a minimum that ranged from 1 to 1.5%. The maximum was attained shortly before sunrise (5:00-6:00 local standard time) and the minimum close to 15:00. These results, corroborated by micro-lysimeter measurements, indicate that the process of water absorption is not restricted to the night but begins during the late afternoon.

The maximum daily water content of the uppermost 1 cm of the soil was found to be relatively constant during the various measurement days and independent of the prevailing meteorological conditions. The minimum water content was not constant and it, therefore, determines the total amount of water vapor that may be adsorbed. A strong and significant correlation between the total adsorption of water vapor by the soil during the "absorption period" and the total potential evaporation between sunrise and sunset of the previous day was found.

A good agreement between the total amount of water gain as measured with the soil samples and the micro-lysimeter was found. On the other hand, the Hiltner balance underestimated the total latent heat flux. These findings, together with the fact that during the nights no signs of dew deposition were observed on the soil surface, imply that dew deposition is not the phenomenon by which water is added to the soil. It was concluded that the main process responsible for the observed diurnal change in water content is the direct adsorption of water vapor by the soil rather than dew deposition.

The water content of the uppermost soil layer was found to be significantly and systematically lower than the wilting point. Most of the commonly used meteorological models would, therefore, assume no latent heat flux. However, latent heat flux density was ~20% of the net-radiation during the night and 10-15% of the net-radiation during the day. Hence models assuming that during the dry season there is no latent heat flux over deserts may lead to erroneous results.

As no dew deposition was detected on the soil surface, but was detected by the Hiltner dew balance artificial condensing plates of any kind cannot be used to evaluate dew deposition quantities on a bare soil surface. Although the magnitude of the latent heat fluxes are larger than those previously reported, the long-held belief that dew is of prime importance in this type of ecosystems needs to be reassessed.

The results of this study bring about new and important knowledge regarding the diurnal changes in soil water content and the energy partitioning over loess soil, under very dry conditions. This knowledge, if applied, can improve the performance of meso- and global-scale meteorological models.

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Chapter 1

Introduction

On Dew and Water Adsorption

– Literature Review

1.1. Introduction

As early as biblical times, dewfall was considered as a precious and life-supporting phenomenon. It was mentioned as a source of great fertility (Genesis 27:28; Deuteronomy 33:13; Zechariah 8:12) and its withdrawal was regarded as a curse from God (2 Samuel 1:21; 1 Kings 17:1). It was deemed as the symbol of wealth (2 Samuel 17:12; Psalms 110:3); and from its refreshing influence, as an emblem of brotherly love and harmony (Psalms 133:3), and of rich spiritual blessings (Hosea 14:5). A popular story dating from the 16^{th} century is that hundreds of liters of water were produced yearly by condensation of dew on a sealed sarcophagus located in the yard of the abbey of Arles-sur-Tech (France) (Beysens et al., 2001). The lack of potable water in certain arid areas has been at the heart of an interest in dew deposition and its possible use. Nikolayev *et al.* (1996) revised the idea of dew collection in the light of the basic physics of condensation and proposed a model for computing condensation rates on dew condensers. Awanou and Hazoume (1997) proposed two types of radiators to condense atmospheric humidity in hot and dry climates based on theoretical aspects of the psychrometric diagram. The possibility to store dew was investigated in the Canary Islands coasts (Hollermann and Zapp, 1991) and the prospect of condensing atmospheric moisture for use in small-scale irrigation was suggested (Alnaser and Barakat, 2000).

In agriculture, dew has a dual role (Wallin, 1967). When dew is formed during the night and the plant's leaves are covered by water drops, after sunrise the stomata open and the plant assimilates without restriction due to the very low stomatal resistance, thereby growing better (Slatyer, 1967). Despite the small amount of free liquid water involved in the process of dew deposition, it can play an important role in the recovery of the water content of a plant after extreme water loss (Went, 1955).

However, in other cases, the presence of dew on the leaves of agricultural crops has a negative effect, as it may lead to the spread of various plant diseases. An important factor of plant disease control and protection is duration of leaf surface wetness, as the spores of many pathogens need a film of water over the leaf tissue in order to germinate and infect the host (Pedro and Gillespie, 1982a, Morin et al., 1993). In particular, the development of bacteria and fungus has been found to be highly influenced by the presence of dew (Auld et al., 1988, Zuberer and Kenerley, 1993, Zhang and Watson, 1997). Thus, most models of crop foliar diseases include factors related to both pathogen biology and the environmental regulation of the presence of dew (Wilson et al., 1999).

The presence of dew is not only important in agricultural context but also has a potential effect on various remotely sensed parameters. Dew affects the soil surface albedo (Menenti et al., 1989, Minnis et al., 1997) as well as the albedo of plant canopies (Pinter-Jr, 1986, Fraser, 1994). In the passive microwave region (radar) Jackson and Moy (1999) found that dew is unlikely to have a significant effect on soil moisture signals. However, Ridley et al. (1996) and Wigneron et al. (1996) found that radar measurements are strongly sensitive to the presence of dew. Dew has a similar influence on the extraction of quantitative crop information from radar imagery (Wood et al., 2002).

The importance of dew is not restricted to its direct utilization. In natural ecosystems, dew serves as an important source of moisture for plants, biological soil crusts, insects, and small animals, especially in desert environments, where water resources are limited (Jacobs et al., 1999). Its importance is emphasized by the extensive research of this subject in the Negev Desert, Israel, since the beginnings of the 1950th onward (e.g., Neuman, 1956, Evenari et al., 1971, Garratt and Segal, 1988, Zangvil, 1996). Biological soil crusts can absorb dew water. Dew may have particular importance for the total period of potential net CO_2 uptake by these crusts, providing relatively long phases of photosynthesis from low moisture supply during the early morning (Evenari, 1985, Lange et al., 1992, Lange et al., 1998, Wilske et al., Submitted). It has been suggested that the germination of desert annual seeds may be enhanced if dew occurs frequently (Gutterman and Shem-Tov, 1997). Dew droplets that condense on plant canopies can provide moisture that helps them overcome the dry season (Willis, 1985). Its importance has been particularly examined in sandy areas (Zentay et al., 1985) and in enhancing the survival of heliophile species in the dry season (Jacobs et al., 2000a).

Moreover, dew is a source of water for insects and small animals. Examples are the *Diacamma Rugosum*, a common ant in India, which acquires a substantial fraction of its water requirements from dew (Moffett, 1985); and the snail *Trochoidea seetzenii,* which was found to use dew as a source of water (Degen et al., 1992, Shachak et al., 2002).

Dew has some geomorphologic effects as well: it contributes both to the chemical and mechanical weathering of rocks (Evenari et al., 1971); it enhances the development of karst formations in arid environments (Castellani and Dragoni, 1987); it has a role as a source of moisture in the stabilization process of sand dunes (Subramanian and Kesava-Rao, 1983); and it has an effect on the aeolian accumulation of natural atmospheric dust (Goossens and Offer, 1995).

Dew is primarily a physical phenomenon, which affects the energy balance at the soil-plant-atmosphere interface. As such, it should be taken into account whenever the exchange processes within plant canopy are analyzed (Jacobs et al., 1996) as it appears to be a major sink of available energy in the early morning (Pitacco et al., 1992) and to affect the actual canopy temperature (Bourque and Arp, 1994). On bare soil, dew depositing during the night usually evaporates during the following morning, and thus exhibits a diurnal cycle of the water content of the uppermost soil layer. This cycle involves exchange of latent heat flux between the soil and the atmosphere, thereby affecting the energy balance at the soil surface.

The energy balance at the soil surface plays an important role in regional and meso-scale studies. However, little detailed information exists on the daily course of energy balance components (Malek and Bingham, 1997). In order to circumvent this lack of information, the actual latent heat flux density may be related to an easily computed flux. For meso- and global- scale models the concept of moisture availability has been proposed and the actual evaporation is computed as a fraction (α) of the easily obtained potential evaporation. As α has been defined as depending on the soil water content of the uppermost soil layer (Carlson et al., 1984, Chen and Dudhia, 2001), changes in the water content resulting from dew deposition may affect these meteorological models. The degree of this effect and the order of the errors introduced by neglecting this phenomenon are unknown.

The above-mentioned studies underline the considerable importance of dew in arid and semi-arid ecosystems. Due to the fact that plants cover only a very small fraction of the surface of deserts, the study of dew deposition on bare soil surfaces is essential. However, it has been found that even though dew deposition was observed on plants and on artificial surfaces, no visible moistening of the bare soil surface was evident during the same nights and in the same location (Kidron et al., 2000, Wilske et al., Submitted). The correlation between dew amounts that deposit on artificial surfaces and the amounts of dew that deposit on the soil surface should therefore be questioned. Moreover, water vapor adsorption has been put forward as being an important link in the water cycle of arid and semi-arid regions (Danalatos et al., 1995, Kosmas et al., 1998, Kosmas et al., 2001). The link between this phenomenon and the deposition of dew (either on the soil surface or on artificial surfaces) has not been studied. In the subsequent sections of this chapter the physical background of both the phenomena of dew deposition and direct water vapor adsorption are described, followed by a critical review of the methods to assess their magnitude, by computational approaches and by direct measurements.

1.2. Physical Background

Water is held within the soil matrix by adsorption at surfaces of particles and/or by capillarity in the pores. It is not always possible to distinguish which of these two mechanisms controls water retention, and usually their combined effect is measured. However, the soil's reaction is influenced by these mechanisms. In particular, water attracted by reactive clay minerals (i.e., adsorption) will cause swelling but when water is attracted by capillarity into the pores of sandy soil, swelling does not occur (Marshall et al., 1996). The relative humidity within the pores has been suggested as the criterion determining which of the two mechanisms is dominant. At high relative humidity in the soil pores (>0.6), the retention curve is determined by capillarity, while at low relative humidity (<0.6) it is determined by physical adsorption (De Vries, 1958). The different formation mechanisms of these two processes are discussed in the following sections.

1.2.1. Dew

Dew is fundamentally a result of phase transition, in which water vapor is transformed into liquid when it comes in contact with a surface. The primary condition for the formation of dew is that the temperature of the surface on which condensation takes place is lower than or equal to the dew-point temperature. The two most critical aspects of phase transition, in the case of dew formation, are the nucleation of the liquid phase and the nature of the growth of the droplets (Beysens, 1995).

Nucleation is the formation of the smallest liquid drop that is thermodynamically stable. The nucleation rate depends upon the wetting properties of the surface, and specifically, upon the contact wetting angle. The contact angle is the angle formed at the interface between the liquid droplet and the surface, as illustrated in Figure 1.1. A contact angle of zero means a complete flattening of the drop and the complete wetting of the solid surface by the liquid. Theoretically, a contact angle of 180^{o} would mean a complete rejection of the liquid by the gas-covered solid, and the drop would retain its spherical shape. The wetting angle of pure water upon clean and smooth inorganic surfaces is generally zero, but if the surface is rough or coated with adsorbed surfactants of a hydrophobic nature, this angle can be considerably greater than zero (Hillel, 1971).

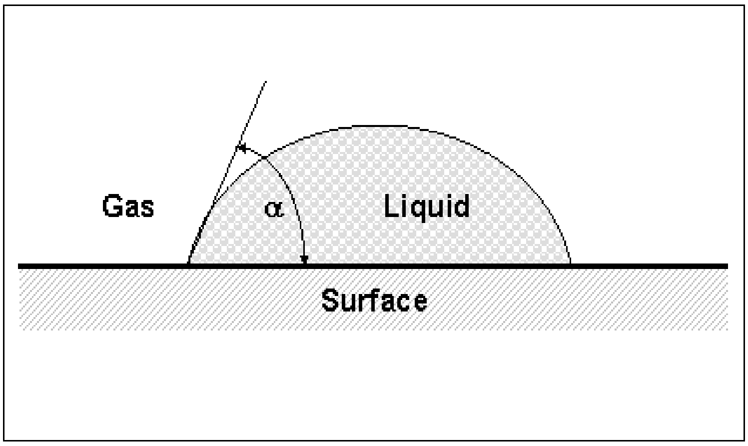


***Figure 1.1**: The contact angle (α) of a drop resting upon a plane solid surface (after Hillel, 1971)*

Once a droplet of water has nucleated on the surface, it begins to grow due to the formation of a concentration gradient of water molecules around the drop. After a coalescence of two drops, a new drop, with a volume equal to the sum of the two initial drops, is formed. This process lowers the surface energy, thus favoring the formation of a new single drop. The droplet growth pattern and the final spatial distribution of the drops on the surface is highly dependent on the surface properties (Beysens, 1995).

In natural ecosystems, several factors determine whether dew will be formed: radiation exchange between the earth's surface and the atmosphere; turbulent heat and water vapor transport in the lower internal boundary layer and within the plant canopy if present; and heat and vapor transport in the underlying soil (Atzema et al., 1990). However, natural atmospheric humidity condensation is associated with two opposite atmospheric conditions. On the one hand, in order for dew to deposit, a radiative cooling of the surface is needed. The rate of cooling depends on the emissivity of the sky, which, in turn, depends on the water vapor concentration in the air, i.e. the lower the vapor concentration, the higher the radiative cooling and thus the higher the rate of deposition. On the other hand, the process of dew deposition obviously requires water vapor, thus the higher the vapor concentration, the higher the rate of deposition.

1.2.2. Direct Adsorption

Adsorption is an interfacial phenomenon resulting from the differential forces of attraction or repulsion occurring among molecules or ions of different phases. Various types of adsorption can occur, depending on the phases involved: adsorption of gases on solids, of gases on liquid surfaces, and of liquids on solids. In the process of water vapor adsorption by soils, adsorption of gas on solid surface is involved.

The first quantitative discussion of the adsorption of gases on solids was given by Irving Langmuir. The equation that relates the amount of gas adsorbed on a surface to the pressure of the gas at constant temperature was defined as the adsorption isotherm, which is derived from a kinetic discussion of the condensation and evaporation of gas molecules at the surface. The Langmuir isotherm is based on the gradual coverage of a surface with adsorbed molecules, in which saturation occurs when the adsorbed layer is uniformly one molecule thick (Moore, 1963).

Two kinds of adsorption are usually distinguished – physical and chemical adsorption. Although there are instances in which it is difficult to definitely assign the adsorption to one of these types, in most cases the decision is not difficult. Chemical adsorption is the result of strong binding forces, comparable with those leading to the formation of chemical compounds, and can be regarded as the formation of a sort of surface compound. This process requires very high energy (80-400 Kj/mol) and is seldom reversible. In contrast, physical adsorption is the result of the operation of forces between the solid surface and the adsorbed material that are similar to the van der Waals forces between molecules. This process requires much less energy (~20 Kj/mol or less), and is quite readily reversible (Moore, 1963). Under natural conditions the available energy does not reach levels high enough to cause chemical adsorption. The process responsible to water vapor adsorption by the soil is physical adsorption (Hillel, 1998).

The magnitude of vapor adsorption on solid surfaces depends on the surface area. The larger the area, more sites exist to which the vapor can attach. In soils, clay particles have the largest surface area, and adsorption takes place mainly on these particles. Therefore, the amount of water adsorbed by the soil increase with the increase of clay content (Thomas, 1928). Moreover, due to the different surface areas of different types of clay, the clay type also affects the adsorption amount. Montmorillonite, for instance, adsorbs much more water than kaolinite at the same relative humidity, and illite has an intermediate position. The effects of clay properties on adsorption are illustrated in Figure 1.2 (Marshall et al., 1996). Under natural conditions, clay particles are never completely dry. A so-called "air-dry" soil is commonly found to have a mass wetness of several percent. The exact percentage depends on the above-mentioned characteristics of the soil, as well as on the humidity of the ambient air (Hillel, 1998).

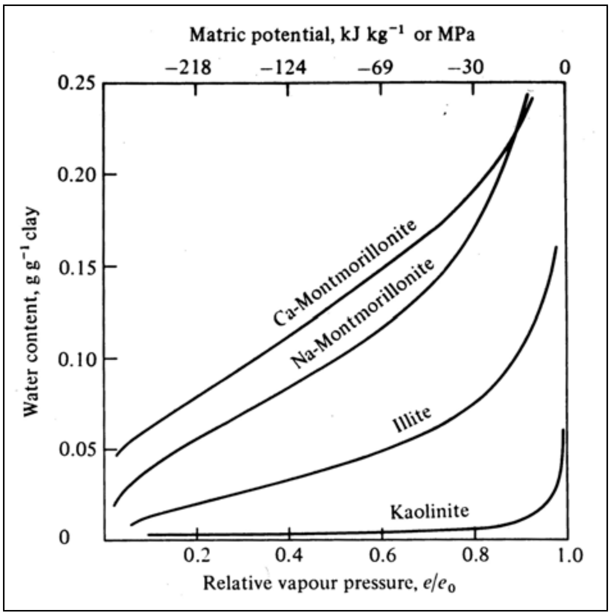


Figure 1.2: *Adsorption of water vapor by different clays (Source: Marshal et al., 1996)*

Regardless of the type of clay, the strength of clay-water adsorption is clearly greatest for the first layer of the water molecules. The second layer is attached to the first by hydrogen bonding, and the third to the second, and so forth, but the influence of the attractive force field of the clay surface diminishes with distance so that beyond a few molecular layers it apparently becomes vanishingly small (Hillel, 1998).

1.3. Quantitative Methods

Diurnal changes in the water content of the uppermost soil layer have been predicted by various models (e.g. Parlange et al., 1998) that are based on the theories that describe the coupled flow of energy and mass in the soil (e.g., Philip, 1957, De Vries, 1958, Milly, 1982, Milly, 1984). The role of water vapor transport was recognized and incorporated in these models. Under extremely dry conditions, water movement in the liquid phase becomes negligible and the change of water content at any given depth will be the result of water vapor movement and physical adsorption or desorption (Scanlon and Milly, 1994). A large number of publications have been dedicated to the problem of energy and mass transport within the soil (e.g., Bristow et al., 1986, Passerat et al., 1989, van de Griend and Owe, 1994, Yakirevich et al., 1997, Chiu-On, 1999). Recently, for example, Qin et al. (2002) presented a detailed model that linked energy fluxes in the atmospheric boundary layer to the coupled transport of mass and energy in a sand dune in the Negev desert. Their model is based on the above-mentioned theories, and does not distinguish between condensation of dew and vapor adsorption. However, models of this kind are not easy to implement and require a large number of variables.

Studies of dew have been largely devoted to two main aspects. The first is the duration of dew (how long the surface remains wet), as the duration of leaf wetness determines the risk for diseases, and thus is important in agronomic context as well as in remote sensing applications. The second is the amount of dew, (how much water the soil-plant system gains), important for ecologists, meteorologists and micro-meteorologists. The research works that contributed the most to the existing knowledge of these two aspects are summarized in Figure 1.3 (page 14), and are discussed in sections 1.3.1 and 1.3.2.

1.3.1. Dew Duration

During the late 70s and the 80s, growing awareness that leaf wetness influences the spread of plant diseases led to a number of attempts to model dew duration. A simple model that describes the link between the period during which the leaf is wet and the environment of a Cocoa pod was first proposed by Monteith and Butler (1979). Later, Butler (1980) suggested some revision of the above-mentioned model to allow the computation of air and dew-point temperature changes after dawn using common meteorological data. This model accounts for the effects of pod size, absorbed short wave radiation and wind speed on film thickness and wetness duration.

Thompson (1981) approached the problem from a different angle. His objective was to assess the commonly held belief that plants remain wet after rain or dew as long as the relative humidity remains above 90%, and he developed a multi-layer model of crop canopies. This model is based on the application of the Penman-Monteith equation (Monteith et al., 1965) to a number of horizontal layers of the crop. The results provided no justification for assuming that the 90% relative humidity criterion will give a satisfactory indication of the leaf wetness in an entire canopy. This criterion was found to be more successful for indicating the duration of leaf wetness during occasions of light dewfall with a dry soil surface, so that distillation is small and dew forms almost exclusively in the upper parts of the canopy. The aim of this model was, however, to quantify the duration of leaf wetness *after* dew has condensed, and not to predict the dew duration. In order to apply this model, variables of the Penman-Monteith equation should be available (e.g., radiation, soil heat flux, air temperature and humidity, and wind speed), most of which are not always available in meteorological stations.

Pedro and Gillespie (1982a) estimated dew duration from micrometeorological measurements for three different crop canopies by using an energy-balance technique combined with heat transfer theory applied to flat plates. Computed values of total dew duration on exposed leaves differed from observed durations by less than 30 minutes. Estimated dew duration on shaded leaves was less accurate, differing from observations by 60 minutes on average. These differences in accuracy were attributed to the use of diffuse solar radiation measured outside the canopy as an estimator of incoming radiation on shaded leaves. In an additional study, the adoption of an energy balance approach to estimate dew duration from standard weather station observations of wind speed, air and dew-point temperatures, and cloud cover was examined in three widely different canopies (Pedro and Gillespie, 1982b). Average errors of approximately 1 hour in dew duration were found in all three cases. This may represents the limit of accuracy for a practical, operational system since weather stations generally report at 1h intervals (Pedro and Gillespie, 1982b).

Janssen et al. (1991) developed a simple model to estimate the frequency and duration of dew using only data on cloud cover, relative humidity, air temperature, and wind velocity. This model ignores the heat flux and the heat storage within the canopy. The results proved to be generally in agreement with measurements of dew occurrence on the surface of an automatic sequential dew sampler.

Recently, Madeira et al. (2002) used energy balance analyses to estimate dew using two different methods to quantify the downward long wave radiation. The first used cloud cover and cloud altitude for calculating sky temperatures, whereas the second is based only on cloud cover for estimating apparent sky emissivity. These models correctly predicted dew occurrence in 91% and 88% of the cases, respectively.

Several approaches were used in all the above-described models in an attempt to relate different environmental factors to the duration of dew. However, each of the presented models was applied and examined in different environments and on different plants, and thus it is impossible to compare them. Results of sensitivity analysis of the dew simulation approach proposed by Pedro and Gillespie (1982b) indicate that the sensitivity of any variable depends to a large extent on the values of all other variables, and therefore varies both spatially and temporally (Scherm and Van Bruggen, 1993). It is reasonable to assume that these kinds of uncertainties would be found if similar sensitivity analysis was performed on the other models.

As has been mentioned, the duration of dew, or the presence of water droplets on the leaves, is an important factor in plant disease development and was therefore intensively studied. In studies that focus on bare soil this issue is of lesser importance. The amounts of dew, however, are very important, both in studies that focus on plants and those that focus on bare soil, as discussed in the next section.

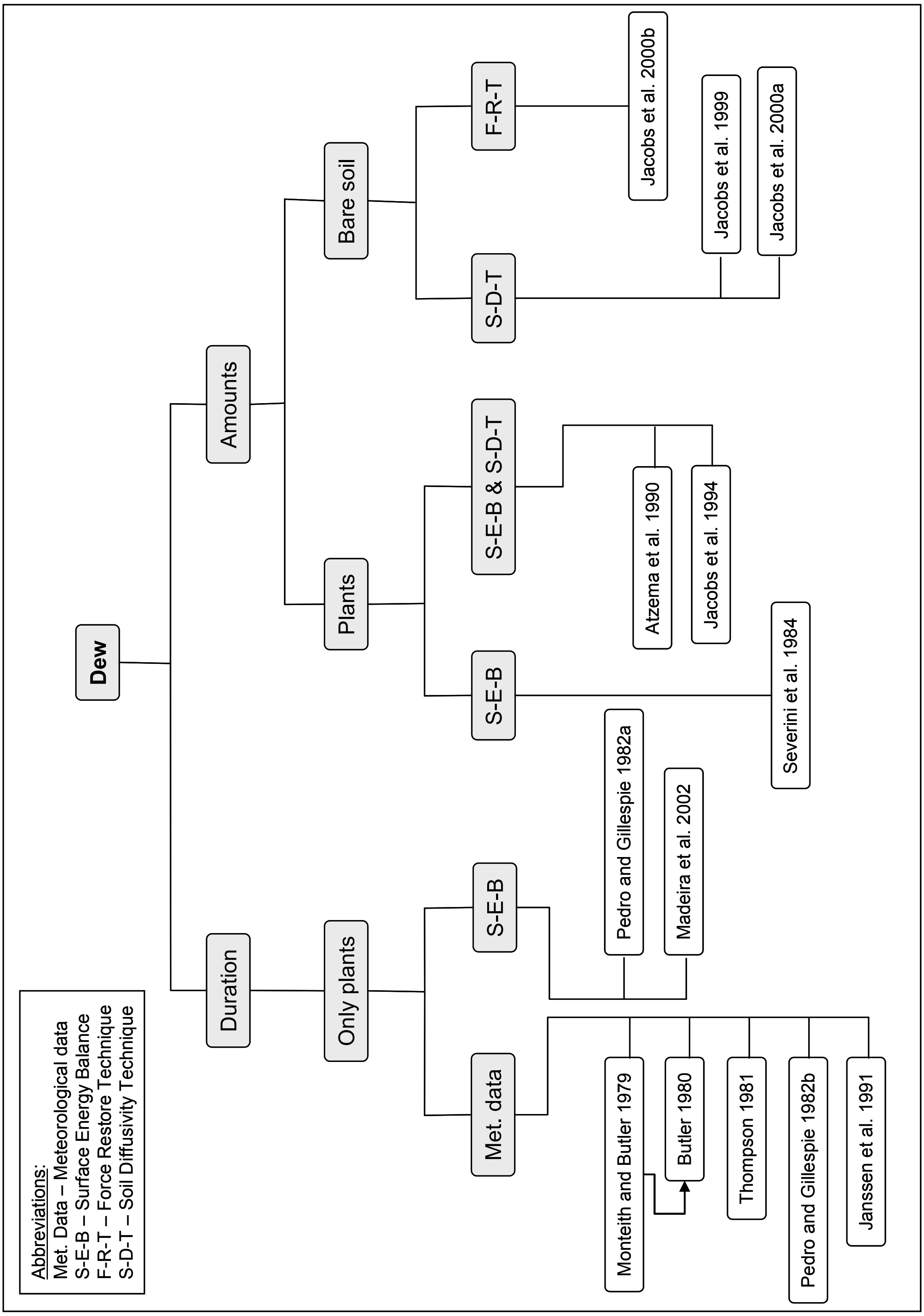


Figure 1.3: *Summary of the main dew models and computation methods.*

1.3.2. Dew amounts

Being a potential source of water in arid and semi-arid regions, it is important to quantify the amount of dew. As dew condensation is the opposite process of evaporation, it is possible to compute the amount of dew deposition using the conventional micro-meteorological methods developed for computing evaporation, provided that the sensitivity of the instruments is sufficient to detect the very small fluxes involved.

A few attempts have been made to compute the amounts of dew deposition. These can be classified according to the approach used for the computations and by the environment in which they were applied. Dew is an interface problem. It occurs on the boundary between the atmosphere and the condensing surface that, in natural environments, can be bare soils, rocks, or plant canopy. As such, it is possible to apply computations of transport processes from both sides of the boundary. Two main approaches have been used, one which uses the transport equations of energy and mass in the soil, while the other is based upon computations of latent heat flux in the atmosphere. Theoretically, these two different approaches should yield the same amounts.

On plants, several attempts to quantify the daily dew deposition have been undertaken. Their main objectives were to develop methods for computing these amounts. However, each of these methods was assessed by comparing the results to a different measurement method.

Severini et al. (1984) computed the daily course of dew deposition on grass and the subsequent evaporation by combining the energy balance equation with standard micrometeorological measurements. The computed fluxes were compared to those measured by a lysimeter with a special setup of 7 aluminum trays filled with grass and roots in order to achieve a better resolution of the measurement. The determination of values of actual evapotranspiration and dewfall were generally in good agreement with measurements.

Atzema et al. (1990) tried to achieve a better insight into the distribution of dew within a moderately tall crop canopy, considering two possible sources of moisture: the atmosphere for the dewfall and the moist soil for dew-rise. The amount of dewfall was estimated by measuring the above crop vapor flux towards the canopy through the Energy-Balance-Bowen-Ratio technique. Dew-rise was estimated by computing the moisture flux in the upper layer of the soil using the water transport equation (Philip and de Vries, 1957). Total water condensation was computed as the sum of both. The same approach of separating between dewfall and dew-rise, and combining these two methods was used by Jacobs et al. (1994). They attempted to describe the development of the free water profile within a crop canopy due to dew and the course of the drying process during the early morning, and compared the model results to the measured flux using artificial condensing plates.

Because plants cover only a very small fraction of the surface of deserts, and because dew serves as an essential source of water in desert regions, the study of dew deposition on bare soil surfaces is of interest. The main attempts to quantify the amounts of dew deposition on bare soil were made in a research site in Nizzana in the northwestern Negev desert of Israel, within the inter-dune of a linear dune system. Jacobs et al. (1999, 2000a) present a simple simulation model to describe the daily dewfall and the early drying processes. An important implicit assumption they made is that under dry conditions, soil water is mainly transported in the vapor phase. Based on this assumption, the transport equations (Philip and de Vries, 1957) were reduced to include only the two terms describing the movement of water vapor by moisture and temperature gradients. The final equation for computing dew deposition is similar to that obtained by Kondo et al. (1990) for evaporation.

Jacobs et al. (2000b) suggests a different approach for computing the amounts of dew condensation. A simple physical model including the force-restore technique for surface moisture was developed. This model simulates the accumulated dew amounts as well as the early morning drying process. Actual dew amounts were assessed using micro-lysimeters. Model simulations agree well with the observations. Their results suggest that the reduction of the vapor pressure in the soil pores under extremely dry soil conditions is essential for the dew process.

Models are not easy to implement, and require a large number of variables. Although several models have been proposed to compute the amounts of dew deposition, they were developed under specific conditions and no attempts to implement them under varying conditions have been undertaken. Moreover, they were validated by comparing their output to different methods of direct measurements. An accurate and sensitive method is clearly needed in order to evaluate these and future models. The available methods to directly measure dew deposition are critically reviewed in the next section.

1.4. Measurement methods

Various devices for measuring dew have been mentioned in the literature. Some of them are used to measure dew duration, others to measure dew amounts, and others are used to simulate both the duration and the amount of dew deposition. Duration measurements depend upon a response to moisture deposits by a change in weight, length or electrical resistance of the sensing equipment (Noffsinger, 1965). Pedro and Gillespie (1982a, b) used an electrical impedance grid that was found to mimic the wetness duration on nearby leaves within 15-30 minutes. In other studies, commercial leaf wetness sensors, made of a gold grid (5.5×5.5 cm) were used (Scherm and Van Bruggen, 1993, Kidron et al., 2000). As it is relatively simple to accurately detect the duration of dew, a number of devices have been developed. The commercially developed leaf wetness sensors are sufficiently accurate for most needs.

When a device to measure dew amounts is required, the situation is completely different. As early as in 1965, a review of methods and techniques for measuring dew revealed three main approaches for measuring dew: weighing lysimeters, weighing dew gages and Duvdevani blocks (Noffsinger, 1965). Of these three approaches, only the lysimeters have the capability to detect adsorption, as they weigh soil samples in which adsorption can take place. Artificial condensing surfaces (whether of dew gages or blocks) cannot adsorb vapor and thus cannot detect this process.

The Duvdevani dew gage is one of the first devices designed for measuring dew amounts (Duvdevani, 1947). A specially treated wooden block of 32x5x2.5 cm is exposed at sunset 1 meter above the ground. The dew distribution found in the early morning on the wood is compared with a series of photographs showing increasing steps of dew incidence. Each photograph has a numerical equivalent of dew deposits. These gages were used in Israel, introducing a countrywide network of dew observations (Gilead and Rosenan, 1954). Despite the ability to measure dew amounts using this gage, the observed dew amounts merely have a climatological meaning, as they are strongly affected by the thermal (i.e. capacity and conductivity) as well as the radiative (i.e., color and roughness) properties of the wood. Thus, the observed values can serve only as a mean for comparing between different sites, not for obtaining absolute values. Another disadvantage of this method is that it requires manual collection and evaluation of the blocks during the early morning. Nevertheless, it was later used in Rajasthan desert in India for assessing the contribution of dew to sand dune stabilization (Subramaniam, 1983).

An improved method for manual measurement of dew amounts was recently suggested (Kidron, 1998). The Cloth-Plate method consists of 10x10x0.2 cm glass plates attached to 10x10x0.5 cm plywood plates. A highly absorbent synthetic cloth (6x6 cm) is attached to the center of the glass plate. This method can be used for comparing dew deposition amounts at different locations and heights. This method was used in the Negev desert for studying the effect of elevation on dew and fog precipitation (Kidron, 1999), for assessing the contribution of dew to the development of biological crusts over sand dunes (Kidron et al., 2000), and for evaluating the condensation and evaporation patterns at three different habitats (Kidron, 2000). As these methods are based on manual collection in the early morning, the output of these measurements is the total dew amount per night. They cannot be used for evaluating the deposition rate or the dew duration.

In many cases, however, it might be desirable to measure both the duration and the amounts of dew deposition. In order to do so, a continuous measurement is needed. A dew recorder of the Kessler-Fuel type has been described (Noffsinger, 1965). It consists on a blackened aluminum plate of a flatted conical shape with its aperture directed upwards. The weight changes resulting from dew deposition are mechanically transmitted to the recording beam. A similar approach was used to design the Hiltner dew balance, with which long-term measurements of dew deposition were undertaken in the Negev Heights of Israel (Zangvil and Druian, 1980, Zangvil, 1996). The Hiltner balance is based on the continuous weighing of an artificial condensation plate that hangs from a beam 2 cm above the ground. This device is very convenient and simple to use, yet its adequacy has yet to be proven since the energy balance of its condensation plate is very different from that of the soil surface above which it is installed. Three factors are responsible: (1) An air gap effectively isolates the hanged condensation plate from the ground; (2) The properties of the material of which the condensation plate is made (a thin plastic plate) are very different from those of the soil; and (3) The dew condensing on the plate accumulates on it, while dew formed on the soil surface may infiltrate into the soil. The Hiltner dew balance can therefore, be considered as a “potential dew” gauge, the results of which are most likely correlated to atmospheric conditions. These limitations also apply to the other methods mentioned above. Different approaches must be used if the goal is to measure actual dew deposition on the soil surface.

One possible manner to evaluate actual condensation and evaporation from the soil is by determining the water movements in the upper soil layers. An electrical conductance soil-moisture meter was developed (Bunnenberg and Kuhn, 1980). It is based on measuring a disturbed soil sample under arid climatic conditions, i.e., conditions in which vapor movement predominates. Although it shows good performance, this device has two main drawbacks: it measures disturbed samples, thus it can not measure *in-situ*, and its presence changes the heat transport, thus affecting the condensation and evaporation fluxes.

An alternative manner of assessing the actual dew deposition rates and amounts is the use of lysimeters and micro-lysimeters (Rosenberg, 1969, Waggoner et al., 1969, Sudmayer et al., 1994, Jacobs et al., 2000a). The resolution and accuracy of the lysimeters should be very high, as the measured fluxes are extremely small. A detectable change of 0.01 mm was found to be insufficient, as in most cases the quantities observed were equal to or less than the device uncertainty (Rosenberg, 1969). Overcoming this technical limitation, lysimeters and micro-lysimeters can theoretically provide the absolute reference for latent heat fluxes as long as the soil and the heat balance of the sample are similar to those of the surrounding area. They can therefore be used to measure both dew deposition and direct water vapor adsorption. Figure 1.4 summarizes the main direct dew measurement methods.

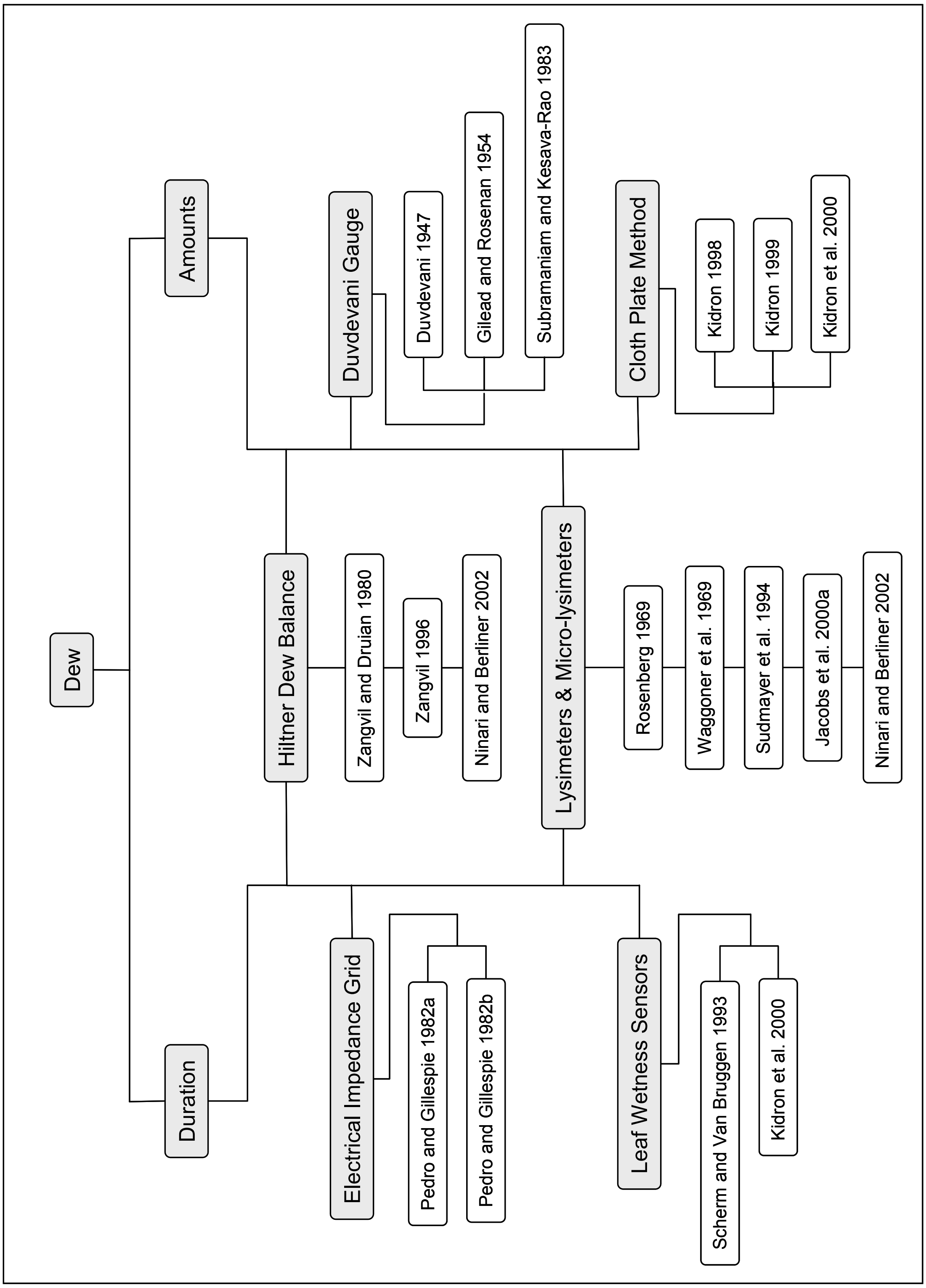


***Figure 1.4**: Summary of the main dew measurement methods.*

1.5. Summary and Objectives

Water is the main limiting factor for any biological activity in arid and semi-arid environments. Thus, any source of water in these harsh environments is very important. Being more than 10% of the total annual precipitation in the Negev desert (Zangvil, 1996), dew is definitely an important water source. Due to the fact that plants cover only a very small fraction of the surface of deserts, the study of dew deposition on bare soil surfaces is essential.

Although several models have been proposed to compute the amounts of dew deposition, they were developed under specific conditions and no attempts to implement them under varying conditions have been undertaken. Moreover, they were validated by comparing their output to different methods of direct measurements. An accurate and sensitive method is clearly needed in order to evaluate these and future models. No such method has been assessed to date. Moreover, as has been detailed above, the correlation between the amounts of dew that deposit on artificial surfaces and those that deposit on the soil surface should be questioned.

Furthermore, water vapor adsorption has been put forward as being an important link in the water cycle of arid and semi-arid regions during the dry season (Danalatos et al., 1995, Kosmas et al., 1998, Kosmas et al., 2001). The magnitude and the frequency of this phenomenon were not thoroughly investigated, and the link between this phenomenon and the deposition of dew (either on the soil surface or on artificial surfaces) has never been studied.

Lastly, both dew deposition and direct vapor adsorption result in water gain in the uppermost soil layer that is lost due to evaporation during the following day. This diurnal cycle of water content involves a diurnal cycle of latent heat flux that may affect the pattern of radiant energy dissipation in general and the energy partitioning at the soil surface in particular.

<u>The objectives of this research</u> were therefore to:

1. Examine the correlation between the amounts of dew deposition on artificial surfaces and the amounts deposit on bare soil.

2. Describe the daily pattern of changes in water content in the upper soil layers and identify the dominance and frequency of those mechanisms by which water is added to the soil (dew deposition or direct adsorption).

3. Assess the relative magnitude of latent heat flux density throughout the dry season, and evaluate its importance in view of meso- and global-scale meteorological models.

Chapter 2

Measurements

Quantifying Actual Dew Deposition on the Soil Surface

As published in *Atmospheric Research*:

Ninari N. and Berliner P.R., (2002). The role of dew in the water and heat balance of bare Loess soil in the Negev Desert: Quantifying the actual dew deposition on the soil surface. *Atmospheric Research*, 64 (1-4): 325-336.

2.1. Introduction

Dew in arid and semi-arid ecosystems is considered to be of great importance (Atzema et al., 1990) and is a water source for the bacteria of biological crusts (Lange et al., 1992, Lange et al., 1998) and for plants (Pitacco et al., 1992, Mabbayad and Watson, 1995). It is therefore of interest to quantitatively describe its role in the short (daily) and long (seasonal) term water balance of arid environments.

During the night, the latent heat flux towards the soil surface is very small, and therefore the amounts of dew deposition are very small as well. This fact poses some very special technical measurement difficulties. Various methods for measuring dew are described in the literature, most of them using artificial condensing plates with varying physical properties (Duvdevani, 1947, Lomas, 1965, Noffsinger, 1965, Bunnenberg and Kuhn, 1980, Zangvil and Druian, 1980, Severini et al., 1984, Janssen et al., 1991, Jacobs et al., 1994, Zangvil, 1996, Kidron, 1998, Liu and Foken, 2001). One of these methods is the Hiltner dew balance (Lambrecht) which was used for several years in the Negev desert (Zangvil and Druian, 1980, Zangvil, 1996) to continuously record dew deposition. The Hiltner dew balance is based on the continuous weighing of an artificial condensation plate that hangs from a beam 2 cm above the soil surface. This device is very convenient and simple to use, yet its adequacy has to be tested since the energy balance of its condensation plate is very different from that of the soil surface above which it is installed due to the fact that: (1) It hangs above the soil surface, the air gap thus effectively isolating it from the soil; (2) The properties of the material of which the condensation plate is made (a thin plastic plate) are very different from those of the soil; and (3) The dew condensing on a plate accumulates on it, while dew formed on the soil surface may infiltrate into the soil.

The Hiltner dew balance could, therefore, be considered as a “potential dew” gauge, the results of which are probably mainly correlated to atmospheric conditions. These limitations apply as well to the other methods mentioned above. A method that detects the actual dew deposition on the soil surface under natural conditions is clearly needed. Examination of the adequacy of a micro-lysimeter (ML) for this purpose is proposed.

MLs have been widely used to measure evaporation from the soil surface of irrigated crops (Shawcroft and Gardner, 1983, Lascano and van Bavel, 1986, Plauborg, 1995), and their use for dew measurements has been suggested (Sudmayer et al., 1994, Jacobs et al., 2000a). Typically, an undisturbed soil sample (a representative vertical section of the soil profile) is inserted into a small cylinder open at the top. The ML is inserted back into the soil with its upper edge level with the soil surface and weighed continuously. For bare soil, any change in weight reflects a flux. The recommended materials and dimensions for ML's have been determined for cases in which the evaporation flux after irrigation was of interest. In these cases the soil is saturated or close to saturation, and the latent heat flux is relatively large (Boast and Robertson, 1982).

Theoretically the ML will provide the absolute reference for latent heat fluxes, as long as the soil and the heat balance of the ML are similar to those of the surrounding area. The heat balance of the soil sample can be considerably affected by insufficient surface area, insufficient depth and wall material:

(1) <u>Small surface area</u>. A small diameter may result in edge problems. Boast and Robertson (1982) and Walker (1983) used ML's with a diameter of 7.6 cm. and wall thickness of 3 mm; Evett et al. (1995) used an ML with a 8.5 cm diameter and wall thickness of 3.5 mm. There are no reports on the effect a change in diameter has on the representativity of the ML.

(2) <u>Insufficient depth.</u> In a shallow ML distorted water and temperature profiles may occur resulting in heat and water fluxes different from those in the surrounding soil. Boast and Robertson (1982) tested the effect of ML depth on evaporation. They concluded that for periods of 1-2 days, the 7 cm depth ML was accurate

enough to describe evaporation from the soil shortly after irrigation (heat flux was not measured).

(3) Wall material. The performance of ML in the field is strongly affected by the thermal conductivity of the wall (λw). λw should be equal or smaller than the thermal conductivity of the surrounding soil (λs) to eliminate vertical heat conduction through the ML cylinder and therefore minimizing horizontal heat flux in the deeper layers of the sample. Evett et al. (1995) found that PVC is the best material among those they tested.

The soil heat flux (in both the ML sample and the surrounding soil) has to match the latent and sensible heat fluxes from the boundary layer at the soil surface. The energy-balance of bare soil at the soil-atmosphere interface is:

$$NR + G + H + E = 0 \tag{2.1}$$

in which all fluxes are positive when directed towards the soil surface (the interface) and measured in (W m^{-2}): NR – net radiation; E – latent heat flux (negative for evaporation and positive for condensation); H – sensible heat flux; G – soil heat flux.

Rewriting the energy balance equation, the latent heat flux (E) can be derived:

$$E = -(NR + G + H) \tag{2.2}$$

The sensible heat flux (H) can be computed by the stability corrected aerodynamic equations (Brutsaert, 1982):

$$H = \frac{\rho C_p u_z k \Delta T}{\left[\left(\ln\left(Z_u / Z_0\right) - \psi_m(Z_u)\right)\left(\ln\left(Z_1 / Z_2\right) - \left(\psi_h(Z_1) - \psi_h(Z_2)\right)\right)\right]} \tag{2.3}$$

in which: ρ - dry air density (kg m^{-3}); Cp – heat capacity of the air (J K^{-1} kg^{-1}); uz – mean horizontal wind speed (m s^{-1}) as measured at height Z (m); k – von Karman constant (=0.41); ΔT – air temperature difference between Z_1 and Z_2 (K) ; Z_o – the roughness length (m) ; ψm and ψh are the stability corrected functions at the pointed height for momentum and latent heat flux accordingly.

The soil heat flux (G) can be calculated by:

$$
\begin{aligned}
g_{j+\frac{1}{2}} &= Cv \frac{\left(T_{j+1}^{i} + T_{j+1}^{i+1}\right) - \left(T_{j}^{i} + T_{j}^{i+1}\right)}{2} \frac{dZ}{dt} \\
G &= \sum_{j=1}^{n} g_{j+\frac{1}{2}}
\end{aligned}
\qquad (2.4)
$$

in which $g_{j+\frac{1}{2}}$ is the mean heat gain/loss for a soil layer of thickness dZ (m) between depths j and j+1 for time interval dt (s) (between i and i+1); Cv – volumetric heat capacity of the layer (J K^{-1} m^{-3}); T – soil temperature (K); n – number of soil layers.

The daily mean value of G is often one or more orders of magnitude smaller than the remaining terms in the energy-balance (equation 2.1) (Brutseart, 1982). This is not the case during shorter periods of time, during which it may be one of the dominating fluxes. The soil heat flux (G) can play a very important role in the energy balance at the soil surface during nighttime.

The soil surface temperature is influenced by the atmospheric and the soil conditions and is one of the main factors that determine whether dew will deposit. Small changes in soil surface temperature of the ML sample, when compared to the surrounding soil, may result in preferential dew deposition if its surface is slightly cooler than the surrounding. It is very important, therefore, to ensure similar temperature profiles (and consequently similar soil heat fluxes) inside the ML and in the surrounding soil. Provided the soil sample is undisturbed and representative of the area, similar temperature profiles will yield equal surface temperatures and hence guarantee that the latent heat fluxes measured with the ML represent the surrounding soil.

The objectives of this work were to test the adequacy of ML's to estimate dew deposition, and to compare the dew deposition measured with it to that measured by a Hiltner dew balance in order to validate the long term data collected by the latter in the research area.

2.2. Materials and Methods

The research was carried out at the Wadi Mashash Experimental Farm in the Northern Negev, Israel (31^o 08’ N, 34^o 53’ E; 400 a.m.s.l., see Map 2.1, page 31). Mean annual rainfall at the farm is 115 mm, most of which occurs between October and April. Long-term maximum and minimum temperatures for January are 14.7^{o}C and 4.8^{o}C; for July they are 32.4^{o}C and 18.6^{o}C. Class A pan evaporation is 2500-3000 mm.year^{-1}. The soil is a sandy loam Aridisol (Loess) with 10% clay, 54% silt and 36% sand.

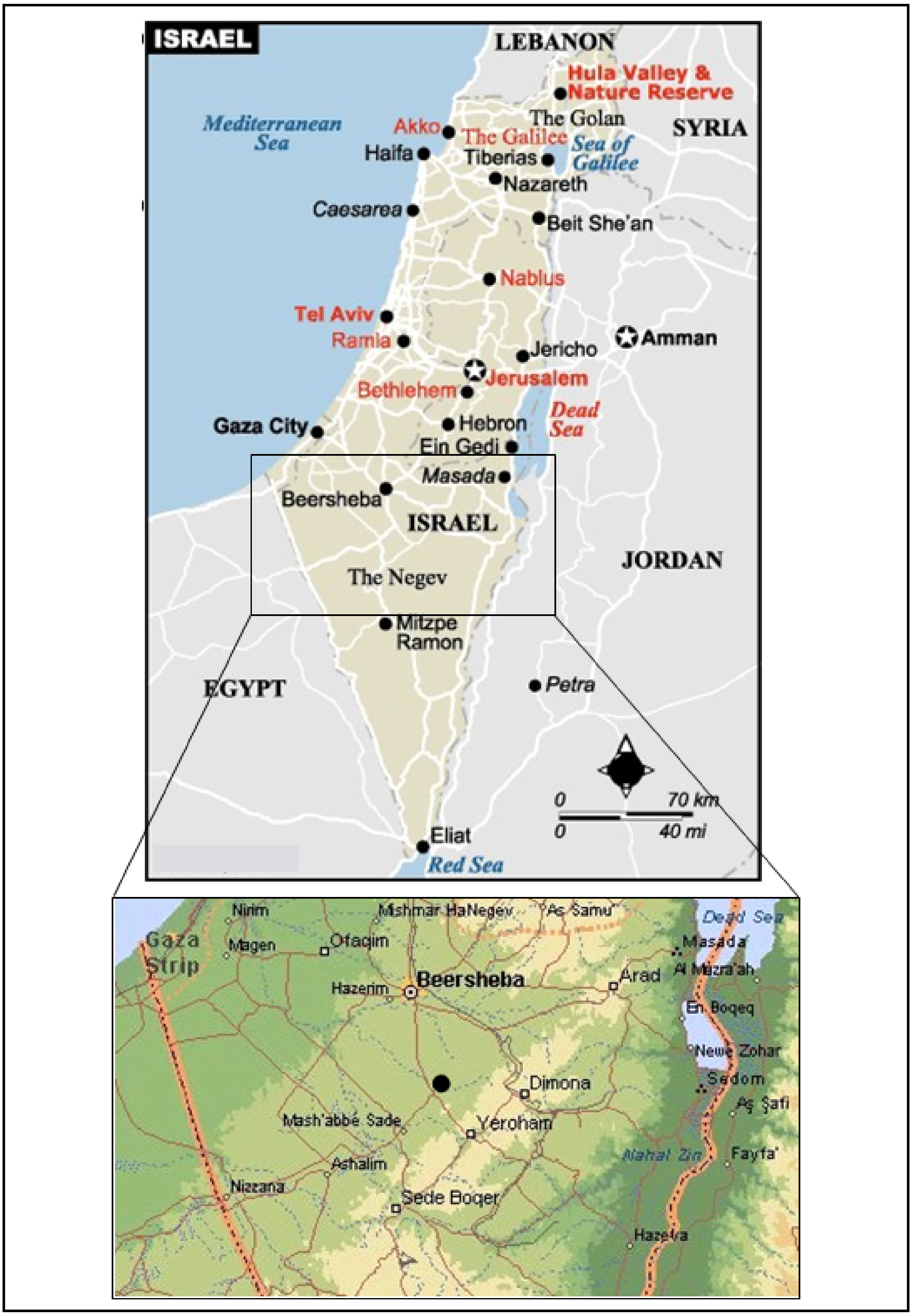
Measurements were carried out during two periods. The first period was from day of the year (DOY) 87 to 126 (March 29 to May 7), 2000 and the second from DOY 163 to 235 (June 12 to August 23), 2001. During both measuring periods the following were measured: incoming and reflected short-wave radiation using two pyranometers (LI2003S, Campbell Scientific Inc.); two net-radiometers (Q-7, Campbell Scientific Inc.) installed 1.5 m above soil surface; wind speed at four heights (2, 1, 0.5, 0.25 m), using cup-anemometers (014A Met-One); soil heat flux at three different locations in the field with heat flux plates (HFT-3, Campbell Scientific Inc.) at depth of 5 cm and measuring temperature above them at 1 cm intervals. The mechanical weighing system of the Hiltner balance was replaced by connecting the weighing arm to a load-cell, thereby reducing the time lag of the original setup. The modified Hiltner dew-balance was placed in close proximity to the ML and its condensing plate was protected from wind using the original windshield designed for this device.

During the first measuring period, the sensible heat flux was measured using six interchangeable self-designed aspirated psychrometers measuring dry-bulb temperatures. A ML was built out of a PVC cylinder (thermal conductivity of ~0.1

W m^{-1} K^{-1}) with 7 mm thickness, 25 cm diameter and 15 cm length. An *undisturbed* soil sample was taken by digging a trench around the desired sample area, to reduce the pressure on the soil crust. The ML cylinder was then forced into the soil by applying pressure with a hydraulic jack and the ML with the soil sample was dug out and positioned above a scale (EK-12Kg A&D) so that the top end of the sample was level with the soil surface. The output of the scale was registered automatically every half hour by a palm computer (48GX, Hewlett Packard). The resolution of the scale was 1g and the ML surface area was 490 cm^2, yielding a resolution of 0.02 mm (in equivalent depth of water) or 27.78 W m^{-2} (in energy terms). One set of 7 differentially connected thermocouples was inserted radially at a depth of 2 cm in the ML soil sample to measure lateral temperature gradient; another set of 6 thermocouples was inserted at depths of 15, 10, 5, 3.75, 2.5 and 1.25 cm inside the ML, to compare its vertical temperature gradients to those measured in the surrounding soil.

During the second period, the aspirated psychrometers were replaced by a sonic anemometer (CA27, Campbell Scientific Inc.) and a deeper ML used. The dimensions of the new PVC ML had a diameter of 18.6 cm and 55 cm of effective depth with an additional 5 cm of polypropylene insulation. A scale with a maximum weighting capacity of 30 Kg and resolution of 0.1g (HP 30K, A&D) was used. The resolution of the deeper ML was 0.004 mm (in equivalent depth of water) or 5.11 W m^{-2} (in energy terms). Temperatures in the ML and in the surrounding soil were measured from 50 cm to 5 cm depth in 5 cm intervals and every 1 cm from 5 cm to soil surface.

Temperatures in the soil and in the ML, as well as the weight of the Hiltner condensation plate were measured for the last two minutes of every half hour. All data excluding the above were measured every 10 seconds and averaged half hourly. Data were measured and collected by a data-logger (23X, Campbell Scientific Inc.).



Map 2.1: *Location of the Wadi Mashash Experimental Farm in the Northern Negev, Israel (31º 08' N, 34º 53' E; 400 a.m.s.l.) is marked by the black circle.*

2.3. Results and Discussion

Latent heat flux measured with the Hiltner dew balance (Hiltner) and the 15cm depth ML (ML_{15}) together with the latent heat flux calculated using the Energy-Balance equation (EB) for one representative day of the first measurement period (day of year (DOY) 105-106, 16-17 April 2000) are presented in Figure 2.1. Positive values represent condensation and negative values - evaporation.

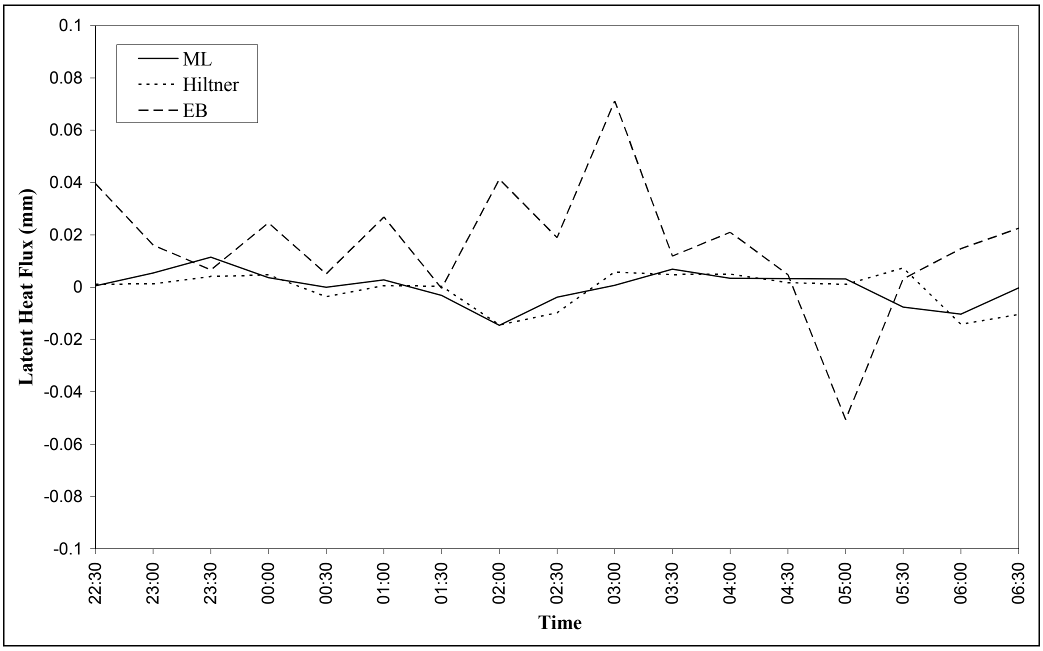


***Figure 2.1**: Latent heat flux as measured by the Hiltner dew balance (Hiltner) and the micro-lysimeter (ML), and as calculated from the energy-balance equation (EB), for DOY 105-106, 2000, a representative day of the first measurement period.*

The latent heat fluxes measured with the Hiltner and the ML15 are similar, while the EB computations yield much larger amounts of condensation. A good agreement between the Hiltner and the ML15 was observed as well for the total condensation per night during the first measurement period, while EB yielded condensation amounts that were much higher than those of the Hiltner and the ML15 (Figure 2.2).

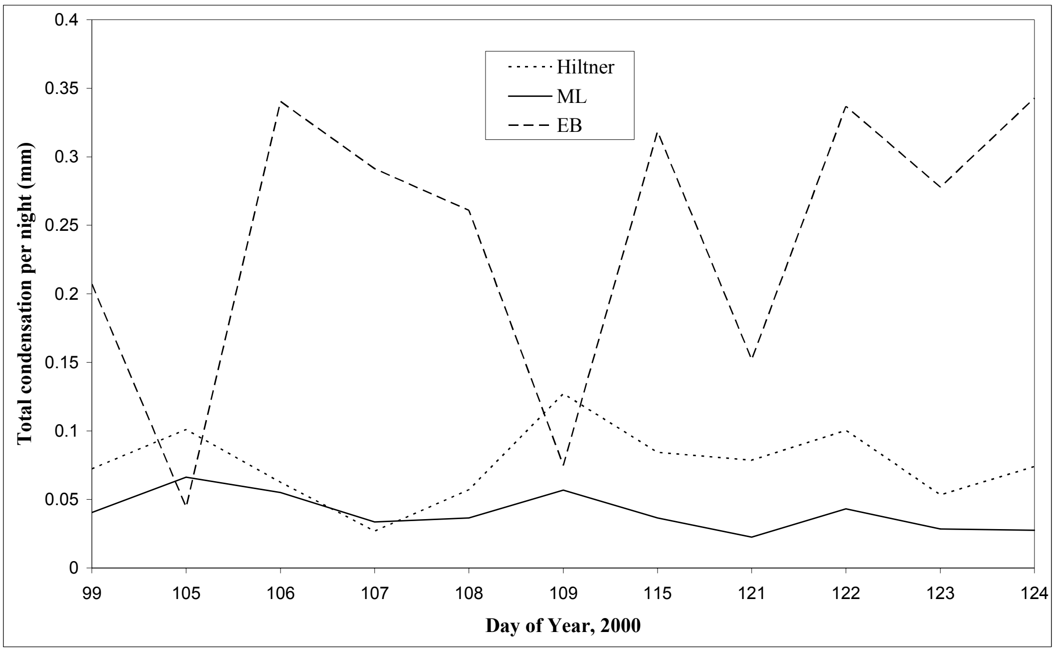


***Figure 2.2**: Total condensation per night measured by the Hiltner dew balance (Hiltner) and the micro-lysimeter (ML) and calculated with the energy-balance equation (EB), for the first measurement period.*

As mentioned previously, it is unlikely to assume that the Hiltner dew balance can accurately describe the actual condensation on the soil surface. The ability of the ML15 to correctly detect the latent heat fluxes at the surrounding soil surface can, therefore, be questioned. One of the critical conditions to ensure the representativity of the ML15 is that a similarity in the temperature profiles close to the soil surface exists. Reliable measurements of surface temperature are difficult to obtain even when an infra-red thermometer is used due to the sensitivity of the former to even extremely small changes of wind speed. We compared therefore the temperatures of the ML15 and the surroundings at 1 cm depth (Figure 2.3). During the day, the ML15 is warmer than the surrounding soil and cooler during the night. These differences are probably amplified at the soil surface.

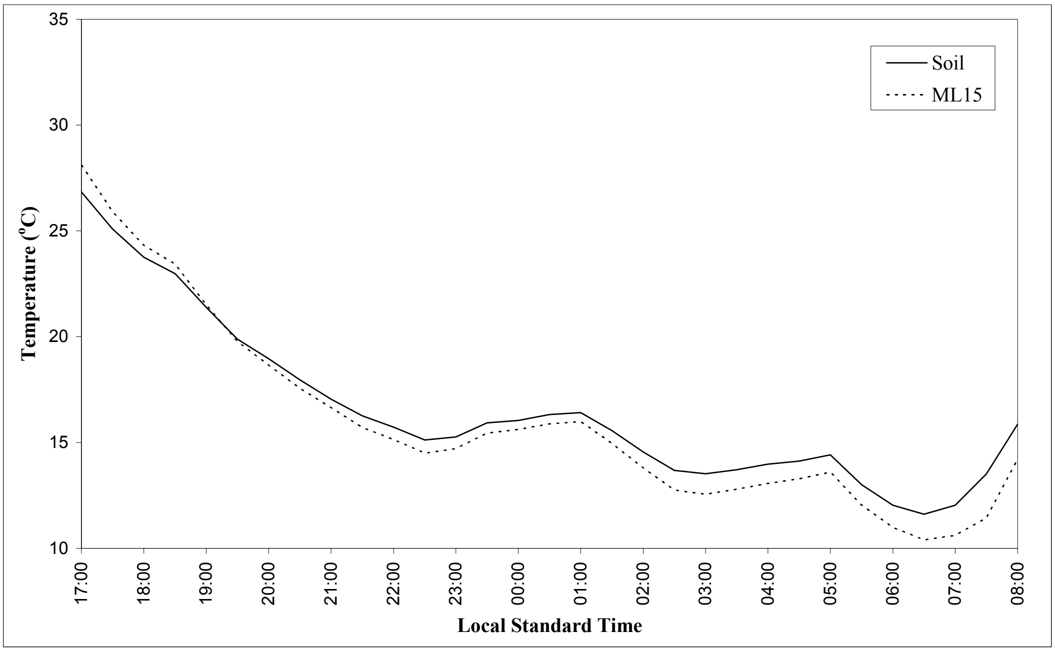


***Figure 2.3**: Soil temperature at 1 cm depth inside (ML) and outside (Soil) the micro-lysimeter, for the 15 cm depth micro-lysimeter, measured at DOY 105-106, 2000 (first measurement period).*

Two possible mechanisms may explain these temperature differences: (1) lateral heat flux exchange between the ML15 and the surroundings, and (2) insufficient depth of the ML, thus preventing the conduction of heat from or to the deeper soil layers. In Figure 2.4, the lateral and vertical temperature gradients inside the ML15 are compared. The lateral gradient, measured at 2 cm depth in the ML is very small when compared to the vertical gradient at that depth, especially during the night (order of $0.0001^{o}C\ m^{-1}$).

Outward radial flux of heat, therefore, cannot explain the preferential cooling of the soil sample inside the ML15 during the night. This cooling can, therefore, only be explained by the lack of deeper layers that could contribute to the upward heat flux and therefore prevent the temperature drop. It appears that heat storage below 15 cm plays an important role, and in order to determine the required depth of MLs in dew-related studies the temperature profile from the soil surface to a depth of 50 cm was measured during nights during which dew was recorded. The rationale of this examination was that the depth of the ML should exceed the shallowest depth at which the daily temperature is constant. In Figure 2.5 several soil temperature profiles as measured at the beginning of the second measuring period (DOY 183, 2001), are presented. It is very clear that a relatively large heat flux can be expected at 15 cm depth.

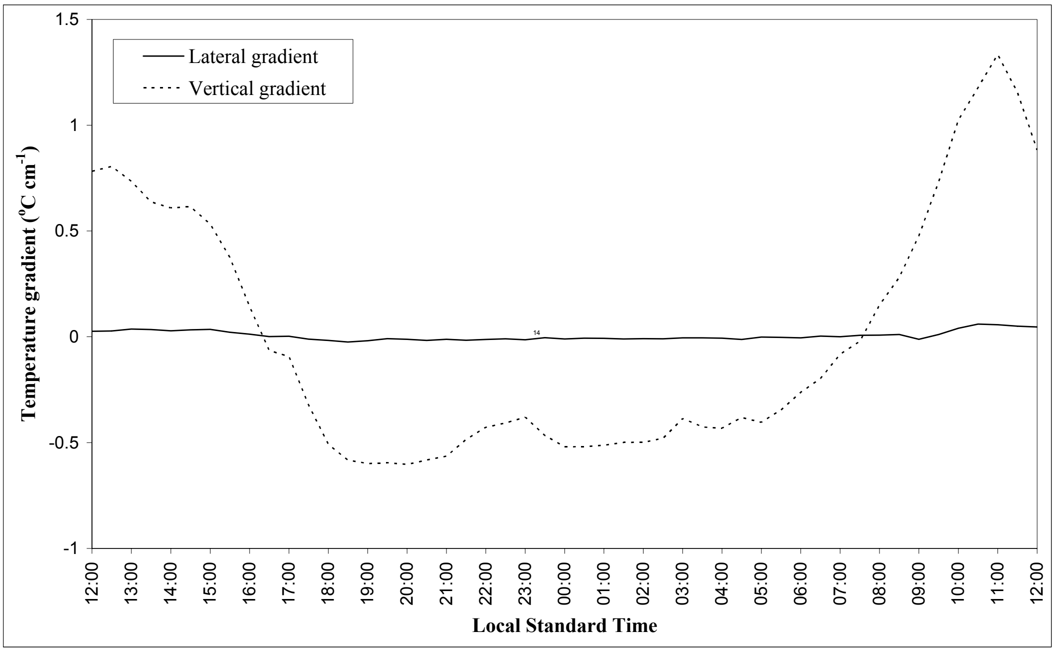


Figure 2.4*: Lateral and vertical temperature gradients inside the micro-lysimeter, measured at Day of Year 105-106, 2000 (first measurement period).*

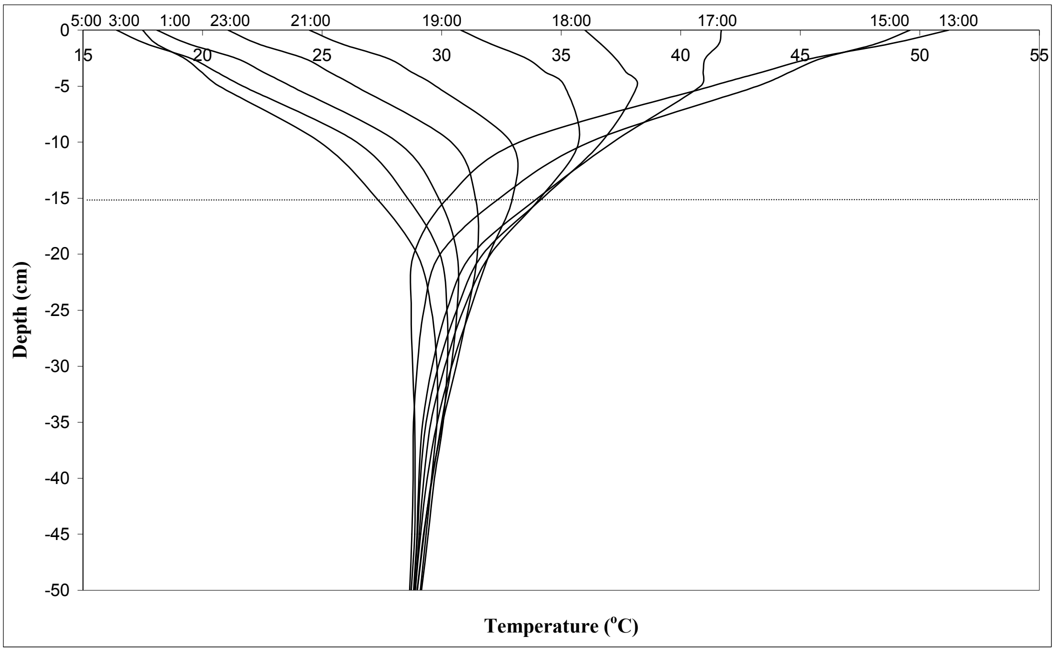


Figure 2.5*: Temperature profiles in the soil for different hours of the day, as measured on DOY 183, 2001 (beginning of the second measuring period).*

In Figure 2.6 the partial contribution of the different soil layers is presented.

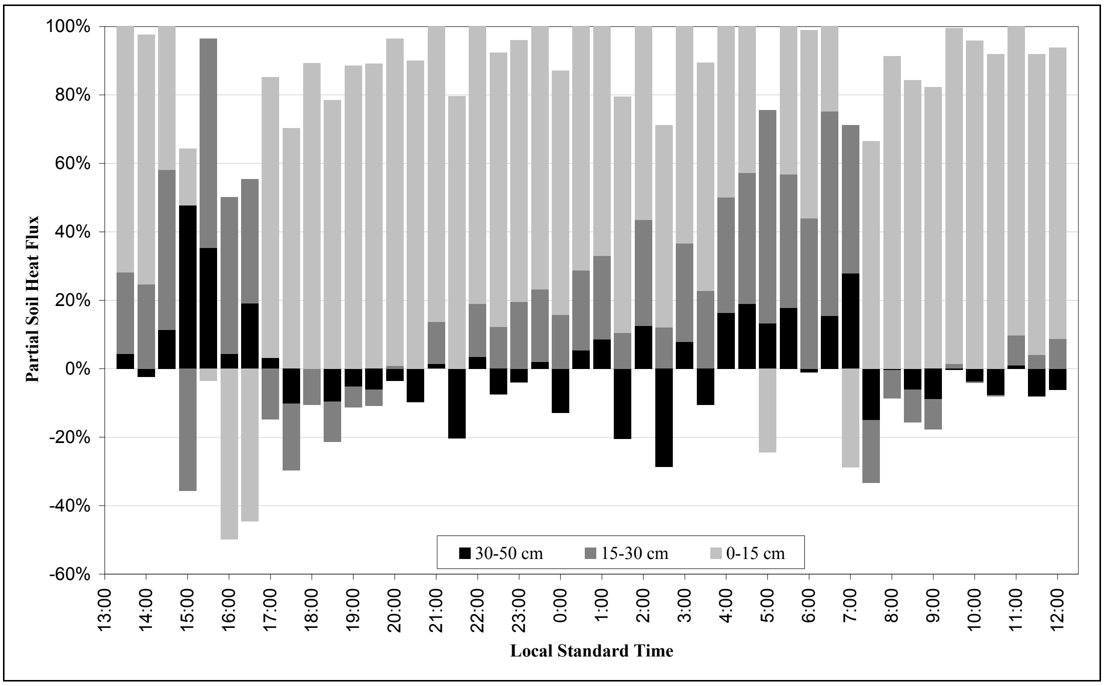


Figure 2.6*: Partial contribution of the different soil layers (0-15, 15-30, 30-50 cm) to the total soil heat flux, for Day of Year 183, 2001 (second measurement period).*

The partial contribution of each layer was calculated as the change in its heat storage $\left(g_{j+\frac{1}{2}}\right)$ divided by the total soil heat flux (G) during the corresponding time interval (equation 2.4). The increasingly important role of the 15-30 cm layer as of 21:00 and until sunrise is noteworthy as well as the fact that during some periods during the night the layers below 15 cm contribute even more than 60% of the total flux. The soil was very dry during this measurement period (average volumetric water content of the upper 5 cm of the soil was 2.1%), and that the conduction of heat in the soil would be higher for higher volumetric water contents. This implies that if the whole profile was wetter, the depth at which the daily temperature remains constant may be below 50 cm. As a result of these findings we built a 55 cm deep ML (ML55). To keep the lateral gradient as small as possible, the PVC tube was thermally insulated by wrapping it in glass wool. The daily course of the soil temperature at 1 cm depth inside and outside the ML55 is presented for a characteristic night in Figure 2.7.

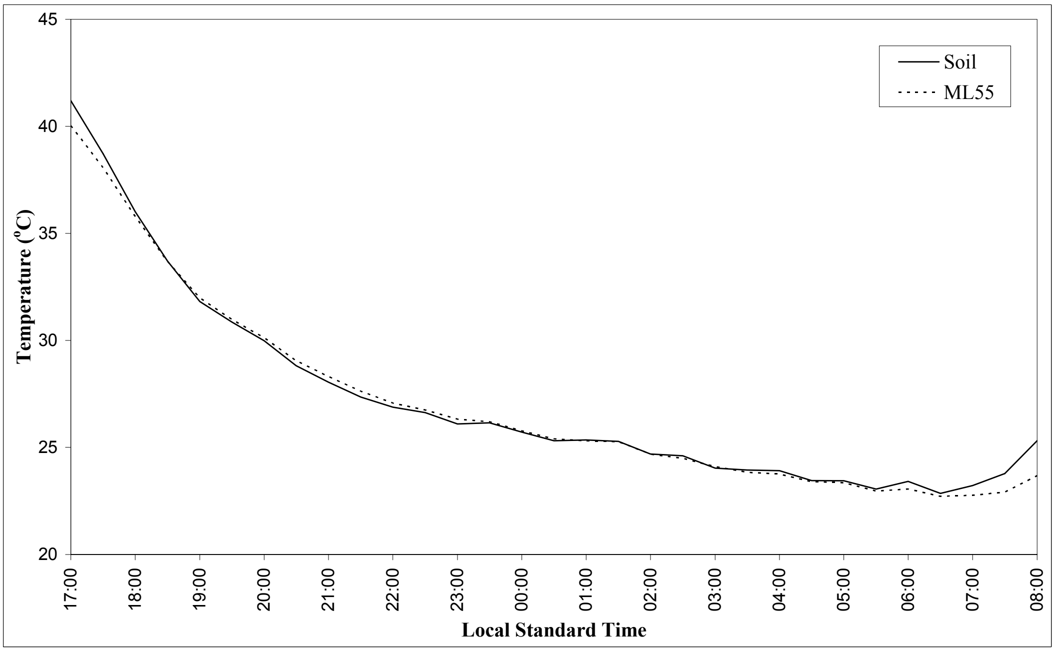


Figure 2.7: *Soil temperature at 1cm depth inside (ML) and outside (Soil) the micro-lysimeter, for the ML_{55}, measured at Day of Year 245-246, 2001 (second measurement period).*

The temperature differences are now much smaller than those recorded for the ML_{15}. The better correspondence between measurements inside and outside the ML for the ML_{55} can be gleaned from the results presented in Figure 2.8 and in Table 2.1. In figure 2.8, temperatures at 1 cm depth inside both ML_{15} and ML_{55} are plotted versus the corresponding surrounding soil temperature at the same depth for all measured nights (from 17:00 to 8:00, at half hourly intervals). Table 2.1 shows the results of the linear regression analysis between the temperatures inside and outside the ML's (both ML_{15} and ML_{55}), for both 1 and 5 cm depths. This pattern repeats for the whole profile. These results leave no doubt as to the fact that the ML_{55} represents the surrounding soil better than the ML_{15}.

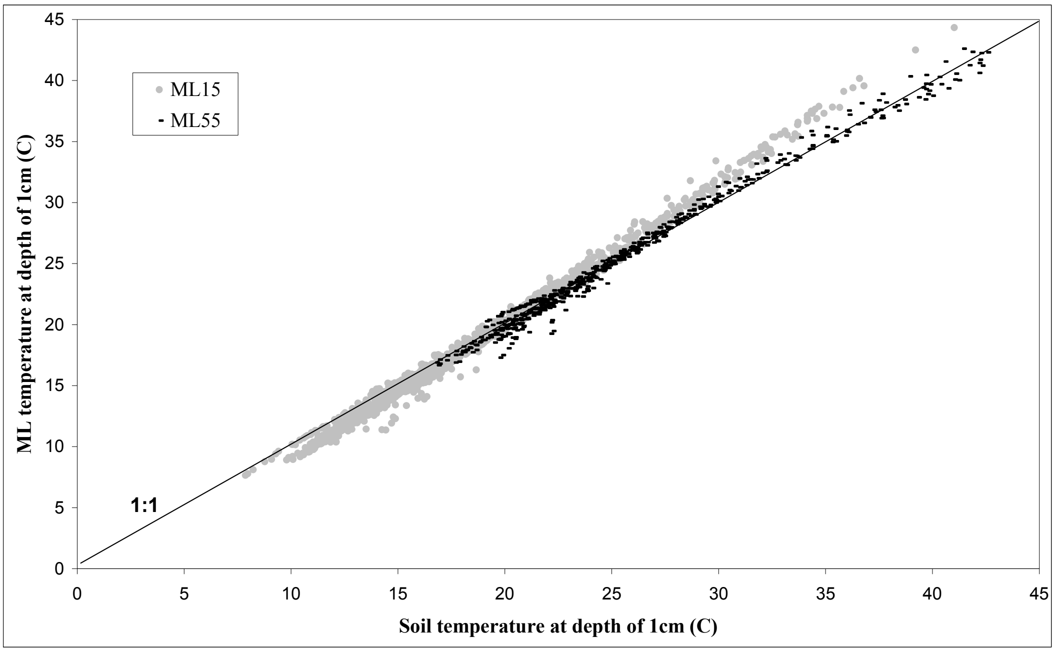


***Figure 2.8**: Temperatures at 1 cm depth inside both ML_{15} and ML_{55} plotted versus the surrounding soil temperature at the same depth for all measured nights (from 17:00 to 8:00, every half hour).*

Table 2.1: Summary of results from linear regression analysis and comparison of slopes between the temperatures inside the MLs (both 15 and 55) and the surrounding soil temperatures, for depths of 1 and 5 cm.

	Depth	N	Slope	Intercept	R^2	P Slopes Comparison	P Coefficients Comparison	P Slopes Comparison to 1
ML_{15}	1cm	1349	1.133	-2.475	0.994	0.0000*	0.0000*	0.0000*
ML_{55}	1cm	565	1.019	-0.439	0.990			0.394
ML_{15}	5cm	1349	1.230	-4.667	0.982	0.0000*	0.0000*	0.0000*
ML_{55}	5cm	565	1.037	-1.668	0.986			0.133

* Slopes/coefficients are significantly different.

Comparing the total condensation measured by the Hiltner, the ML_{55} and the computed latent heat flux (EB), a completely different pattern than that presented in Figure 2.2 emerges (Figure 2.9). The ML_{55} and the EB agree rather well. It seems, therefore, that a deeper ML resolves the apparent lack of agreement between atmospheric and soil measured fluxes as appeared to be the case from the results obtained with ML_{15}.

We suggested that the values registered by the Hiltner balance could be referred as "potential dew deposition", thus implying that larger dew amounts than those that actually condense on the soil surface should be detected by the balance. From our results it appears that the opposite is the case: the Hiltner balance underestimated total condensation. No surface wetting was evident during visual inspect of the soil surface on several occasions during which dew was registered by the Hiltner balance. This would indicate that water vapor is adsorbed within the soil profile, even though the soil surface temperature does not drop to dew point. Hints for the existence of this mechanism can be found in a theoretical analysis of Philip (1957) and in articles of Kosmas et al. (1998, 2001).

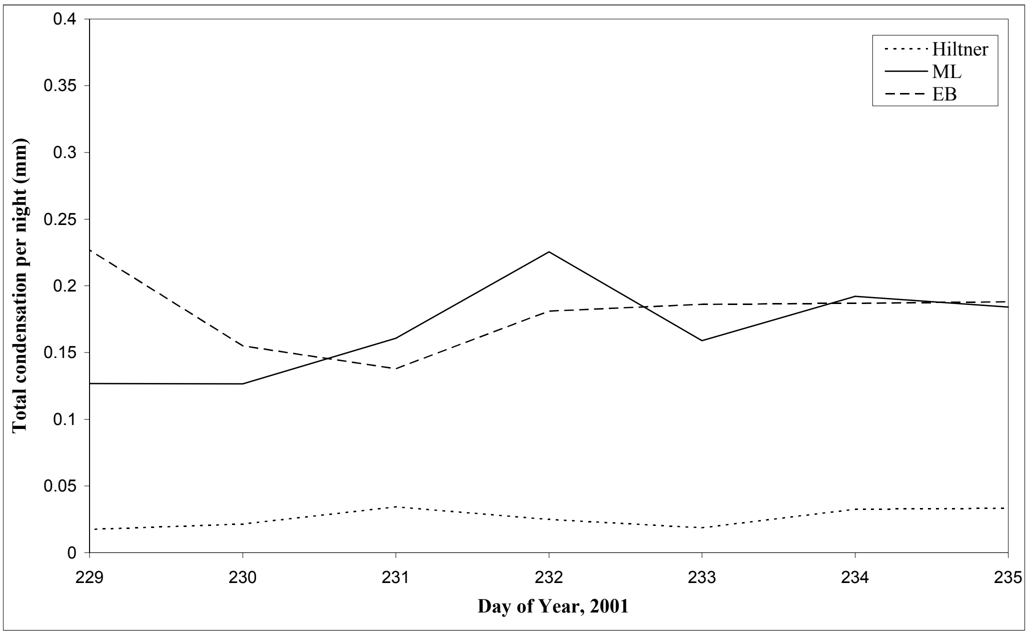


Figure 2.9: *Total condensation per night measured by the Hiltner dew balance (Hiltner) and the micro-lysimeter (ML) and calculated from the energy-balance equation (EB), for the second measurement period.*

2.4. Conclusions

The use of a ML appears to be the most accurate method to measure dew deposition on the soil surface. Nevertheless, it appears that the ML specifications for measuring evaporation published to date are not sufficient in the case of measuring condensation. We have shown that for measuring dew, the minimum depth of a ML should be the depth at which the diurnal temperature is constant in order to ensure similar temperature profiles inside and outside the ML.

For a dry loess soil in the Negev Desert, a minimum depth of 50 cm is required. Furthermore, our results indicate that during the night water vapor is adsorbed within the soil profile, even though the soil surface temperature does not drop to dew point and no surface wetting is evident. This process, which appears to be important at least for the area in which we carried out our trials, cannot be quantified with the Hiltner dew balance and casts serious doubts about the usefulness of the latter for the estimation of total absorption of water from the atmosphere by the soil profile.

Chapter 3

Results

Diurnal Water Content Cycle

As published in Journal of Hydrometeorology:

Agam (Ninari) N. and Berliner P.R. (2004). Diurnal water content changes in the bare soil of a coastal desert. Journal of Hydrometeorology, 5 (5), 922-933.

3.1. Introduction

The deposition of dew is a common meteorological phenomenon. The magnitude of the fluxes involved is very small, which would indicate a-priori, that the potential contribution of dew to the water balance of a given region would be very minor. Nevertheless, its contribution to the water budget of plants has been studied, and particular attention has been paid to the microclimate that develops within the plant canopy (Pedro and Gillespie, 1982a, Zuberer and Kenerley, 1993, Wilson et al., 1999, Jacobs et al., 2000a). It has been reported that in arid areas, dew is an important ecosystem input due to the lack of precipitation (Evenari, 1985). It is a major source of water for the growth and development of biological soil crusts (Lange et al., 1992, Lange et al., 1998, Jacobs et al., 1999); it plays a key role in the germination of annual seeds (Gutterman and Shem-Tov, 1997); it is an important source of water for small insects (Moffett, 1985); and finally, it plays a role in the water balance of some vascular plants (Willis, 1985, Zentay et al., 1985, Jacobs et al., 2000a).

Dew is not only important in a biological context. Its presence, for instance, may affect remotely sensed parameters. It has been shown that altimeter backscatter measurements are affected by the presence of dew on the surface (Ridley et al., 1996) and that surface reflectance values were affected by the daily cycle of dew deposition and subsequent evaporation (Menenti et al., 1989). For these and similar applications it is enough to know the time span during which dew is present on the ground. Dew may, in addition, play an important role in deserts by changing the water and energy surface balances. It is a commonly accepted fact that the magnitude of latent heat flux above desert areas is linked to the occurrence of precipitation events, and is negligible during the dry season during which the radiant energy reaching the surface of the desert is partitioned between convective and conductive sensible heat only (Cleugh and Roberts, 1994, Unland et al., 1996). This pattern could be altered by the presence of a wet soil surface, which could happen if enough dew would condensate.

In order to assess the degree to which dew may affect the above-mentioned balances, it is necessary to quantify the actual amounts of water deposited and their distribution within the soil matrix. Due to the fact that plants cover only a very small fraction of the surface of deserts, the study of dew deposition on bare soil surfaces is essential. This task is, however, challenging and introduces many difficulties, since the fluxes and the total amounts of deposited water are extremely small. To overcome these difficulties the usual approach has been to assess the condensation on detached surfaces (on which measurements are easier to perform) and assume that they are representative of the actual deposition on the substrate above which they are installed. A number of devices to determine the amounts of dew deposition have been proposed and are in use (Duvdevani, 1947, Lomas, 1965, Noffsinger, 1965, Bunnenberg and Kuhn, 1980, Zangvil and Druian, 1980, Severini et al., 1984, Janssen et al., 1991, Jacobs et al., 1994, Zangvil, 1996, Kidron, 1998, Liu and Foken, 2001), their common denominator being that the physical properties of the artificial condensing plates are very different from those of the bare soil.

Among these instruments, the Hiltner dew balance showed great promise, due to its simplicity and robustness. It was used for several years in the Negev desert in Israel in order to monitor dew deposition (Zangvil and Druian, 1980, Zangvil, 1996). This instrument consists of an artificial condensation plate (hanging from a beam 20 mm above the soil surface) that is continuously weighed. The energy balance of the condensation plate is very different from that of the soil above which it hangs due to the following: (1) the plate is isolated from the soil surface by an air gap; (2) the properties of the material of which the condensation plate is made (a thin plastic plate) are very different from those of the soil; and (3) the dew condensing on the plate accumulates on it and may evaporate, while dew formed on the soil surface may be absorbed by the porous matrix. In view of these limitations (which also hold true for the other dew-recording methods or devices mentioned above) the Hiltner dew balance could be considered as a "potential dew" gauge, whose results are mainly correlated to atmospheric conditions (Ninari and Berliner, 2002).

The amounts of dew deposition recorded in the Negev Highlands using the Hiltner balance, were in the range of 0.06-0.1 mm per night on average and 200 dewy nights per year were detected (Zangvil, 1996). Close by, in a very similar area (soil type, topographical elevation, etc.), Kidron et al. (2000) estimated dew using a different type of gauge (Kidron, 1998). This gauge was found to be well correlated with the Duvdevani dew gauge, but suffers essentially from the same set of drawbacks as described previously for the Hiltner balance. Kidron et al. (2000) made the interesting comment that even though dew deposition was recorded by their dew gauges, no visible moistening of the soil surface was evident. This observation raises the interesting possibility that the detection of dew by gauges located on or close to the soil surface may not be indicative of actual dew deposition on the soil surface. This, however, does not preclude the possibility that water vapor from the atmosphere may be directly absorbed by the soil matrix as a result of capillary condensation and/or physical adsorption. The former is the predominant mechanism when the relative humidity in the pores is high, while the latter predominates at low values of relative humidity (Philip and de Vries, 1957). Water adsorption has been put forward as being an important link in the water cycle of arid and semi-arid regions (Danalatos et al., 1995, Kosmas et al., 1998, Kosmas et al., 2001).

Kosmas et al. (1988) proposed an empirical model to predict the water vapor adsorption as a function of the minimum daily relative humidity, the daily amplitude of relative humidity, and the soil water tension of the 50mm topsoil. The usefulness of such a model is, however, limited, as water tension of the uppermost soil layer is rarely measured, and most definitely not on a regular basis.

The previously mentioned studies (Zangvil, 1996, Kosmas et al., 1998, Kidron, 2000) indicate that there are a large number of nights during the dry season during which, irrespective of the mechanism (dew deposition or direct vapor absorption), water is added to the bare soil surface in arid environments. The added water will evaporate during the following day, thereby changing the pattern of radiant energy dissipation.

The purpose of this study was to describe the daily pattern of changes in water content in the upper soil layers and to identify the mechanism by which water is added to the soil (deposition or direct absorption). Moreover, the gains in soil water content during the night are compared to the dew amounts recorded by the Hiltner balance and the losses and gains of water in terms of easily measurable environmental parameters are parameterized.

3.2. Materials and Methods

The measurements were carried out at the Wadi Mashash Experimental Farm in the Northern Negev, Israel (31o 08’ N, 34o 53’ E; 400m A.M.S.L., 60Km from the Mediterranean Sea). Rainfall events occur between October and April, and the mean long-term annual rainfall at the farm is 115 mm. Long-term maximum and minimum temperatures are 14.7^{o}C and 4.8^{o}C for January; and 32.4^{o}C and 18.6^{o}C for July, respectively. Class A pan evaporation is 2500-3000 mm per year. The soil is a sandy loam Aridisol (Loess) with 13% clay, 15% silt and 72% sand and porosity of 0.45.

Data was collected during nine 24-h field campaigns that took place during the dry season of 2002. A total of 124 mm of rain were recorded during the rainy season of 2001-2002 (previous to the above mentioned measuring period). The first campaign took place 10 weeks after the last rainfall of the 2001-2002 season (March 29, 13.5

mm), and the last one ended about two hours before the first rainfall of the next season (October 30). The dates of the remaining 7 campaigns were randomly spread in between.

Daytime monthly averages of net-radiation and nighttime monthly averages of wind speed and air temperature are presented in Table 3.1. The corresponding values, as measured during the campaigns held in those months, are noted as well, showing that the randomly chosen dates for the field campaigns are representative of the season.

***Table 3.1**: Monthly averages of some meteorological conditions directly affecting dew deposition, and the same values, as measured during the campaigns held in those months, showing that the randomly-chosen dates for the field campaigns are representative of the season.*

Month		Daytime Net-Radiation (W m^{-2})	Nighttime Wind Speed (m s^{-1})	Nighttime Air Temperature (C)
June	Average	252.94 ± 9.07	2.09 ± 0.52	20.48 ± 1.53
	Campaigns	264.73	2.65	21.30
July	Average	247.27 ± 9.33	1.92 ± 0.57	23.00 ± 1.21
	Campaigns	267.33	2.29	24.89
August	Average	241.77 ± 14.46	1.83 ± 0.45	23.94 ± 1.43
	Campaigns	240.68	1.91	24.30
September	Average	195.29 ± 15.49	1.65 ± 0.34	20.50 ± 1.71
	Campaigns	207.10	1.48	19.44
October	Average	141.83 ± 24.83	1.55 ± 0.46	18.64 ± 2.47
	Campaigns	137.08	1.71	17.47

During each campaign, the 100mm topsoil was sampled hourly, and gravimetric water content of the samples determined at 10mm increments. A micrometeorological station was installed near-by for continuous measurement of incoming and reflected short-wave radiation with two pyranometers (CM5, Kipp & Zonen); net-radiation (Q-7, Campbell Scientific Inc.); wind speed at four levels (2, 1, 0.5, 0.25 m) with cup-anemometers (014A Met-One); dry- and wet-bulb temperatures at 1m height using a self-designed aspirated psychrometer; soil heat flux at three different locations in the field with heat flux plates (HFT-3, Campbell

Scientific Inc.) installed at a depth of 50 mm, and temperature measurements above them at 10 mm intervals, using differentially wired thermocouples. Data were measured and collected every 10s and averaged every 30min by a data-logger (23X, Campbell Scientific Inc.). In addition, the changes in mass of an improved micro-lysimeter (Ninari and Berliner, 2002) (186 mm diameter and 550 mm of effective depth with an additional 50 mm of polyurethane insulation) were recorded by placing the micro-lysimeter in a pit and weighing it every half hour. The scale (AND, maximum weighting capacity 30 kg) had a resolution of 0.1g, which resulted in a resolution of 0.004 mm (equivalent depth of water) or 5.11 W m^{-2} (in energy terms). The output of the scale was registered automatically every half hour by a palm computer (48GX, Hewlett Packard). Dew deposition amounts were measured using a conventional Hiltner dew-balance (Lambrecht ltd.) using the original windshield.

Table 3.2 summarizes the environmental conditions of eight out of the nine campaigns (a data-logger failure occurred during the August 27-28 campaign) for the daytime (from sunrise to sunset) and the nighttime (from sunset to sunrise) separately.

Maximum incoming short-wave radiation varied from more than 950 W m^{-2} at the beginning of the measuring period to ~700 W m^{-2} towards the end. Throughout the season, a decrease in both the daytime and nighttime net-radiation fluxes is as well apparent. The same trend is evident in the differences between the maximum and the minimum temperatures of the soil surface and the air. The daytime minimum relative humidity changed significantly during the season, even though daytime averages did not vary much. Maximum relative humidity reached ~95% on all nights. The variance of the average nighttime relative humidity was higher than the corresponding humidity during daytime. The average wind speed at all nights was rather high.

Table 3.2: *Specifications of the field campaigns conditions (note the variation between the different days along the dry season).*

Date			Jun. 17-18	Jun. 26-27	Jul. 18-19	Aug. 05-06	Sep. 05-06	Sep. 25-26	Oct. 15-16	Oct. 29-30
Incoming Short-Wave Radiation (W m-2)		light hours	14	13.5	14	14	12.5	12.5	11.5	11
		Max	991.94	958.15	952.57	944.68	985.16	829.35	744.65	670.49
Net-Radiation ($W\ m^{-2}$)	Day	Max	432.74	436.06	426.00	410.68	457.35	367.40	339.24	319.57
		Average	210.56	198.36	209.25	195.44	195.98	185.96	170.89	126.06
	Night	Min	-82.13		-70.54	-74.07	-67.69	-63.82	-55.68	-49.66
		Average	-53.97		-47.29	-53.07	-55.80	-40.08	-37.91	-39.33
Air Temperature (C)	Day	Max	35.01	35.91	38.93	37.06	32.72	32.98	33.06	27.56
		Average	27.07	28.63	32.77	31.03	28.66	27.98	28.84	24.18
	Night	Min	15.99		20.42	19.77	16.88	14.95	17.34	12.87
		Average	20.85		24.98	24.61	20.22	19.22	20.62	16.27
	Range		19.03		18.52	17.29	15.84	18.04	15.72	14.68
Soil Surface Temperature (C)	Day	Max	54.40	55.43	55.91	55.78	53.80	47.30	46.09	39.07
		Average	39.50	39.76	42.32	41.89	40.32	36.07	35.66	29.79
	Night	Min	17.05		19.43	20.22	16.61	14.99	16.12	12.66
		Average	21.55		25.60	25.33	20.48	19.21	20.56	15.99
	Range		37.35		36.48	35.56	37.19	32.31	29.98	26.42
Relative Humidity (%)	Day	Min	35.92		44.51	45.51	55.07	57.77	53.47	58.38
		Average	63.76		61.23	66.60	68.10	65.23	61.23	71.00
	Night	Max	96.78		98.79	95.03	93.13	99.32	97.41	96.61
		Average	86.98		85.44	88.20	85.16	93.71	88.41	85.67
Wind Speed ($m\ s^{-1}$)	Day	Max	7.43		7.03	6.47	6.07	5.91	6.95	5.27
		Average	3.39		4.08	2.85	2.94	3.73	3.47	2.83
	Night	Max	4.623		4.54	4.86	4.86	4.14	5.58	4.22
		Average	1.95		1.98	1.80	1.74	1.60	2.10	1.49

3.3. Computational Procedure

The soil surface temperature was measured at three locations as described in the previous section. On July 4, 2001 these average values were compared to soil surface temperature averages obtained using a hand-held infrared thermometer (AG42, Telatemp Corp.) pointed at the areas in which the thermocouples were installed. The analysis of variance of the regression (r^2=0.99, p=0.0000) indicates that the intercept was not significant and the slope (=1.07) not significantly different from 1. In view of the good agreement between both methods and the fact that it is easier to continuously record thermocouple output, we used the latter to compute average soil surface temperature.

Dew point temperature at the soil surface (DPTs) determines if dew deposits or not. Measurements have not been carried out very close to the soil surface. However, as latent heat fluxes were small and wind speeds relatively high, the vapor pressure gradient was very small, and the dew point temperature computed at a height of 1m height is probably a good rough estimator of the dew point temperature close to the soil surface.

The total potential evaporative energy for each day was estimated by summing the half-hourly potential evaporation computed using the Penman-Monteith equation (3.1) (Monteith, 1965), from the time when evaporation began (as derived from micro-lysimeter measurements, usually at sunrise) to the time when the minimum gravimetric water content was recorded (usually during the early afternoon).

$$PET = \frac{\Delta\left(NR + G\right) + \rho C_p \frac{\left(e_a^{sat} - e_a\right)}{r_a}}{\Delta + \gamma} \tag{3.1}$$

where Δ - slope of saturated water vapor pressure vs. temperature curve (mb $^oC^{-1}$); *NR* – net-radiation flux density (W m^{-2}); *G* – soil heat flux density (W m^{-2}); ρ - dry air density (kg m^{-3}); C_p – specific heat of dry air at constant pressure (J K kg^{-1}); e_a^{sat} - saturated water vapor pressure at air temperature (mb); r_a – aerodynamic resistance (sec m^{-1}) defined as:

$$r_a = \frac{\ln\left(Z_u/Z_0 - \psi_m \right) \ln\left(Z_e/Z_0 - \psi_v \right)}{k^2 U_z} \tag{3.2}$$

with Z_u – height of wind speed measurement (m); Z_e – height of water vapor pressure measurement (m); *k* – von Karman constant (≈0.41); U_z – wind speed at height Z_u (m s^{-1}); ψm and ψ_v – integrated diabatic influence functions (Paulson, 1970); Z_0 – the roughness length (m).

The roughness length was computed from wind speed profile measurements obtained under neutral conditions ($-0.01 \le Ri \le 0.01$) and from an additional set of 3D wind speed measurements, and was found to be 0.042±0.074 mm (n=1276).

3.4. Results and Discussion

In order for dew to deposit on a surface, the temperature of the surface should be equal to, or less then, the dew-point temperature of the air mass with which it is in contact. In Figure 3.1, the nighttime soil surface temperature and the dew point temperature at 1m height during eight out of the nine campaigns is presented.

It can be readily observed that the soil surface temperature usually did not drop below the estimated dew-point temperature, with the exception of September 25-26, and a short time interval during the early morning of October 30. From these observations it was to be expected that no dew would deposit on the soil surface. Indeed, dew deposition on the soil surface was not visually observed in any of the campaigns. However, a clear daily cycle in the gravimetric water content of the uppermost 10-mm soil layer was observed, indicating that moisture was absorbed by the soil during the late afternoon and night, and evaporated thereafter (henceforth the "absorption period" and "evaporation period", respectively). A third degree polynomial was fitted by least squares to the diurnal variation in the measured moisture content of the upper soil layer for each of the nine campaigns. The data and the fitted polynomials are presented in Figure 3.2. The polynomial regressions and the coefficients were significant for all campaigns, clearly indicating that the diurnal cycle is not an exceptional occurrence. The minimum water content was recorded close to 15:00 and the maximum around sunrise.

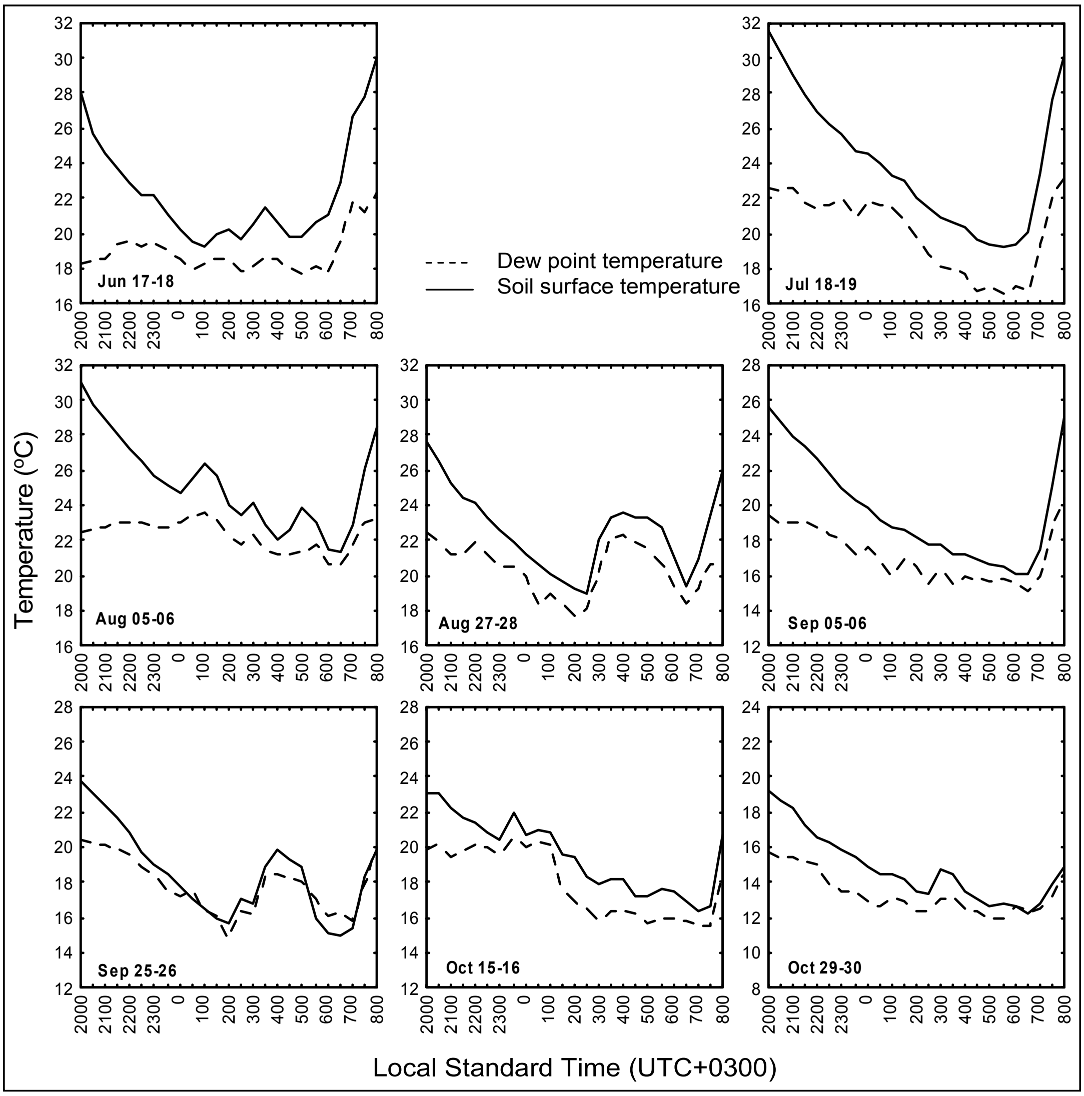


***Figure 3.1**: Night-time soil surface and dew point temperatures during the eight field campaigns. The soil surface temperature did not drop to the dew-point temperature, with the exception of September 25-26, and a brief period during the early morning of October 30.*

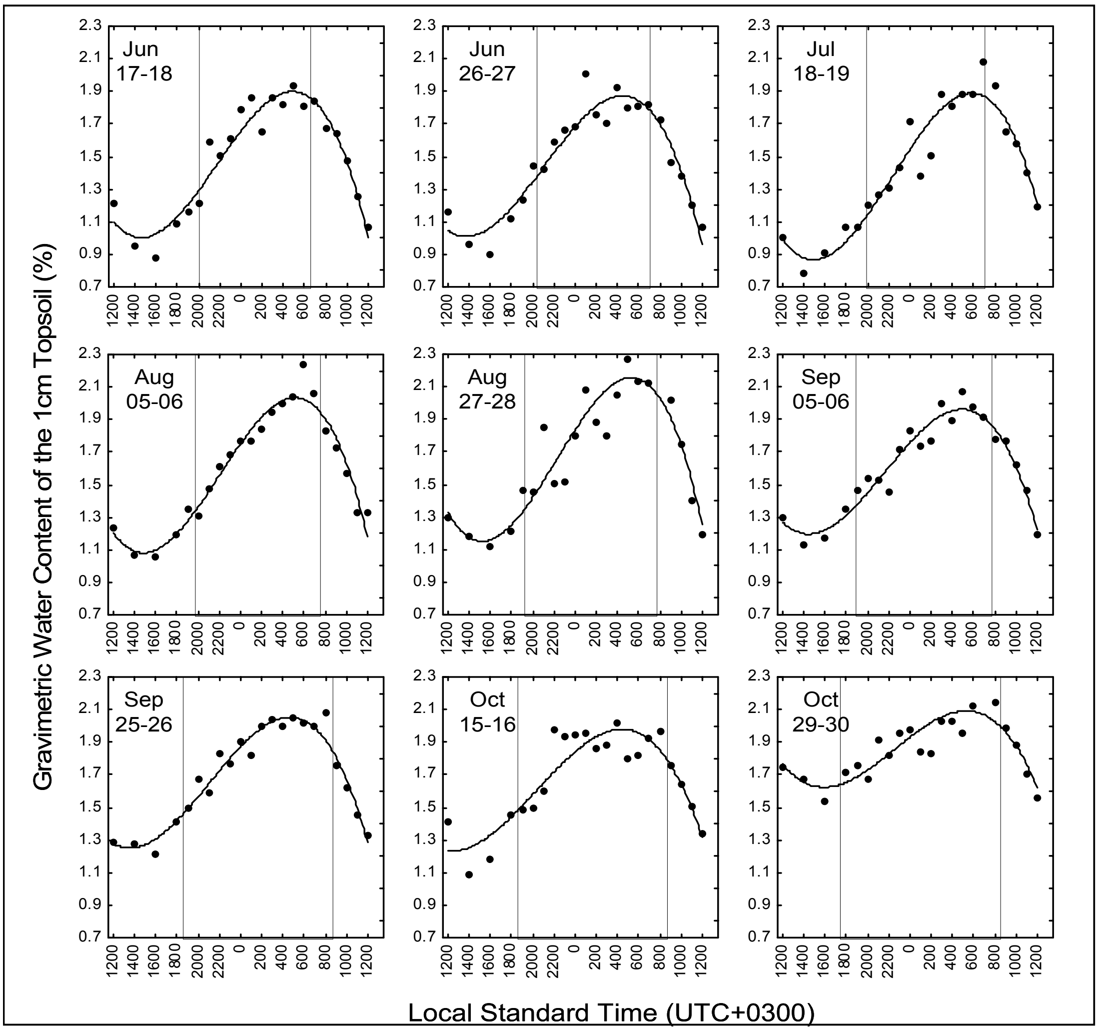


Figure 3.2*: Diurnal patterns of the gravimetric water content of the uppermost 1-cm soil layer*. *The vertical lines indicate the time of sunset and sunrise.*

The magnitude of the changes in water content was very small and the average water content throughout the entire measurement period very low: a maximum of approximately 2% and a minimum which ranged from 1 to 1.5 %. These gravimetric water contents correspond to approximately $4 \cdot 10^2$ and $5 \cdot 10^6$ bar respectively (computed using the Van Genuchten formulation (Van Genuchten 1980)). The relative humidity in the soil pores, corresponding to the above mentioned potentials (computed using the Kelvin equation (Hillel, 1971)) reached a maximum of 75% before sunrise and dropped close to 0 at noon. These are conditions for which the dominant mechanism is adsorption, rather than capillary condensation (Philip and de Vries, 1957).

Third degree polynomials were also fitted to each of the remaining 10-mm layers for all campaigns. In Figure 3.3 for example, the best-fit lines, drawn using the coefficients of third degree polynomials, for the uppermost six ten-mm soil layers for August 27-28 are presented. The fact that the amplitude in the water content decreased with depth is evident. This pattern repeated itself for all campaigns, although the depth at which no change was detected varied throughout the season. An objective criterion was needed to determine the depth to which the daily change in water content penetrated. In Table 3.3 the levels of significance for the coefficients of the polynomials are presented, with the exception of those that correspond to the intercept, as they were always significant but are not relevant to the particular issue under consideration. Layers in which a daily change in water content could be observed were defined as such, if at least two of the three coefficients (intercept excluded) were significant. This approach may be illustrated by the best fit lines presented in Figure 3.3 and the corresponding data in Table 3.3, from which we can conclude, for example, that on August 27-28 the daily change occurred in the five uppermost layers of the soil, and no change in water content was detected below that depth.

Table 3.3: *Significance levels of the slope parameters of third degree polynomial fit of the gravimetric water content change (GWC) with time (LST) for each of the 10 one-cm layers at the different dates ($GWC=b_0+b_1LST+b_2LST^2+b_3LST^3$). Layers in which changes in water content occurred were defined as those for which at least two of the coefficients were significant (α=0.05). The grayed areas mark the defined layers.*

Date		0-1 cm	1-2 cm	2-3 cm	3-4 cm	4-5 cm	5-6 cm	6-7 cm	7-8 cm	8-9 cm	9-10 cm
June 17-18	r^2	0.93	0.92	0.79	0.57	0.62	0.51	0.51	0.35	0.28	0.39
	b_1	0.0018	0.0000	0.0094	0.0480	0.0353	0.9191	0.2493	0.9975	0.8248	0.6476
	b_2	0.0000	0.0000	0.0010	0.0485	0.0291	0.8269	0.5605	0.7222	0.8519	0.7231
	b_3	0.0000	0.0000	0.0009	0.0694	0.0550	0.8789	0.6736	0.6931	0.7600	0.4231
June 26-27	r^2	0.92	0.87	0.93	0.62	0.50	0.27	0.48	0.53	0.60	0.18
	b_1	0.0166	0.0051	0.0472	0.9519	0.9589	0.1242	0.2981	0.0276	0.1241	0.1218
	b_2	0.0000	0.0001	0.0038	0.4421	0.5298	0.2112	0.3865	0.0433	0.3397	0.2928
	b_3	0.0000	0.0000	0.0021	0.4186	0.4530	0.2463	0.4080	0.0498	0.5681	0.3832
July 18-19	r^2	0.92	0.84	0.70	0.20	0.47	0.23	0.49	0.21	0.15	0.30
	b_1	0.0024	0.0018	0.0522	0.2847	0.3535	0.4322	0.1384	0.4944	0.7419	0.1386
	b_2	0.0000	0.0001	0.0060	0.2808	0.1426	0.3828	0.0864	0.2934	0.5418	0.1531
	b_3	0.0000	0.0000	0.0033	0.3454	0.1154	0.4396	0.1084	0.2435	0.4937	0.2209
August 05-06	r^2	0.95	0.79	0.77	0.82	0.37	0.37	0.28	0.35	0.32	0.47
	b_1	0.0001	0.0999	0.0305	0.0216	0.3588	0.4199	0.9845	0.7223	0.8631	0.2101
	b_2	0.0000	0.0032	0.0029	0.0020	0.1754	0.2853	0.6533	0.9206	0.6996	0.6642
	b_3	0.0000	0.0008	0.0018	0.0016	0.1492	0.3055	0.5427	0.8229	0.5445	0.9915
August 27-28	r^2	0.85	0.84	0.62	0.37	0.25	0.07	0.11	0.05	0.01	0.24
	b_1	0.0065	0.0025	0.0234	0.0389	0.0259	0.9992	0.8612	0.4341	0.8450	0.5224
	b_2	0.0001	0.0001	0.0089	0.0324	0.0299	0.9806	0.9818	0.4328	0.7920	0.3375
	b_3	0.0000	0.0000	0.0107	0.0446	0.0398	0.9128	0.9485	0.4302	0.7633	0.3089
September 05-06	r^2	0.93	0.71	0.69	0.58	0.46	0.55	0.50	0.32	0.27	0.33
	b_1	0.0073	0.0454	0.0069	0.0042	0.1861	0.1242	0.5427	0.7264	0.3725	0.3571
	b_2	0.0000	0.0082	0.0025	0.0082	0.1463	0.2112	0.6630	0.7328	0.3912	0.3577
	b_3	0.0000	0.0066	0.0034	0.0241	0.1980	0.2463	0.5711	0.6172	0.3394	0.4691
September 25-26	r^2	0.95	0.81	0.55	0.51	0.41	0.52	0.46	0.24	0.37	0.47
	b_1	0.0157	0.2166	0.5095	0.0006	0.7977	0.1386	0.3238	0.1131	0.1130	0.0181
	b_2	0.0000	0.0042	0.6826	0.0004	0.2805	0.6296	0.8588	0.2245	0.3308	0.0641
	b_3	0.0000	0.0005	0.3445	0.0004	0.1254	0.8494	0.6625	0.3720	0.6278	0.1779
October 15-16	r^2	0.83	0.50	0.45	0.24	0.25	0.09	0.13	0.22	0.22	0.05
	b_1	0.5346	0.2997	0.2452	0.6161	0.2691	0.8890	0.8956	0.6818	0.7741	0.3659
	b_2	0.0099	0.0457	0.0604	0.3223	0.1441	0.6705	0.6944	0.3847	0.4500	0.3312
	b_3	0.0007	0.0306	0.0302	0.2301	0.1196	0.5771	0.6410	0.2872	0.2858	0.3280
October 29-30	r^2	0.74	0.42	0.39	0.31	0.12	0.14	0.08	0.07	0.18	0.48
	b_1	0.0098	0.0693	0.0160	0.0209	0.4093	0.9714	0.5832	0.7153	0.4973	0.0054
	b_2	0.0004	0.0253	0.0074	0.0130	0.2740	0.9308	0.4901	0.5823	0.3220	0.0233
	b_3	0.0001	0.0203	0.0065	0.0124	0.2328	0.9916	0.4834	0.5496	0.2829	0.0555

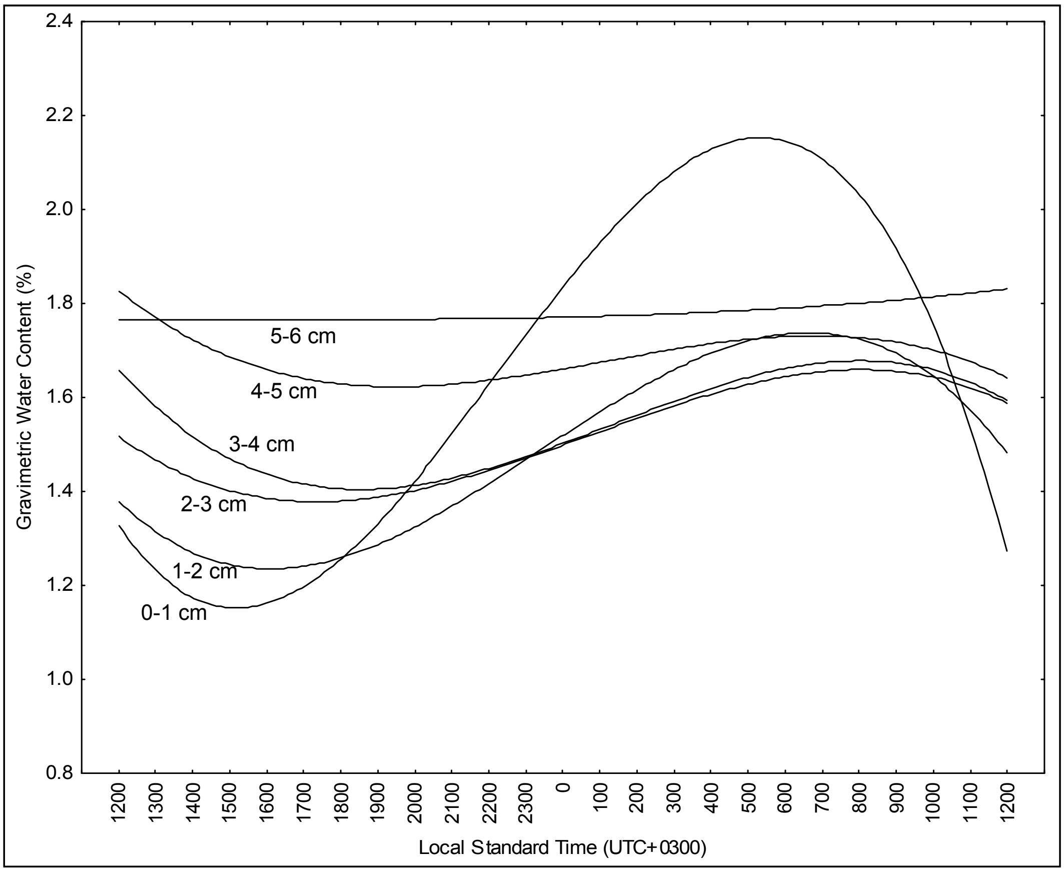


***Figure 3.3**: Best fit lines obtained from a third degree polynomial regression analysis for the uppermost six one cm soil layers for August 27-28.*

Diurnal changes in the water content of the uppermost soil layer have been predicted by various models (e.g. Parlange et al., 1998) that are based on the theories that describe the coupled flow of energy and mass in the soil (e.g., Philip, 1957, De Vries, 1958, Milly, 1982, Milly, 1984). The role of water vapor transport was recognized and incorporated in these models. Under extremely dry conditions, as was the case during the present study, water movement in the liquid phase becomes negligible and the change of water content at any given depth will, thus, be the result of water vapor movement and physical adsorption or desorption (Scanlon and Milly, 1994). Qin et al. (2002) presented a detailed model that linked energy fluxes in the atmospheric boundary layer to the coupled transport of mass and energy in a sand dune in the Negev desert. Their model does not distinguish between capillary flow due to condensation of dew and vapor adsorption, but predicts diurnal changes in the water content of the uppermost sand layers that correspond qualitatively with our data. No data set describin diurnal moisture changes in the upper soil layers for such extremely dry conditions has been previously presented.

The total change in water content (expressed as equivalent depth of water) was computed by summing the changes in volumetric water content for the depths for which a change was defined (as detailed above). Figure 3.4 presents the adsorption amounts measured with the micro-lysimeter and those measured with the soil samples. The maximum values on July 17-18 and August 27-28 are slightly higher for the soil samples, while on July 26-27 and September 25-26 the micro-lysimeter showed slightly higher amounts. No systematic under- or over- estimation is evident. The trends are similar and the differences are rather small.

The detection of the changes in the micro-lysimeter's mass, together with the abovementioned findings that there are no changes in the water content of the deeper layers of the soil, clearly indicate that the addition of water to the soil's uppermost layer is due to absorption of atmospheric water vapor, rather than redistribution of water within the soil.

In Figure 3.5 the total amounts of added water during the absorption period, as measured using the micro-lysimeter and the Hiltner dew balance, were plotted against those derived from the total changes in water content of the soil samples. The agreement between the amounts measured with the soil samples and the micro-lysimeter on one hand and the clear underestimation of the Hiltner balance on the other hand are noteworthy. The average water gain per absorption period was 0.26 and 0.25 mm for the soil samples and the micro-lysimeter respectively, with corresponding standard deviations (SD) of 0.05 and 0.04. In contrast, the average gain detected by the Hiltner balance (for the same dates) was much lower: 0.04 mm (SD=0.02). The fact that during the night no signs of dew deposition could be observed on the soil surface, indicate that the process by which water was added to the soil was adsorption rather than dew deposition.

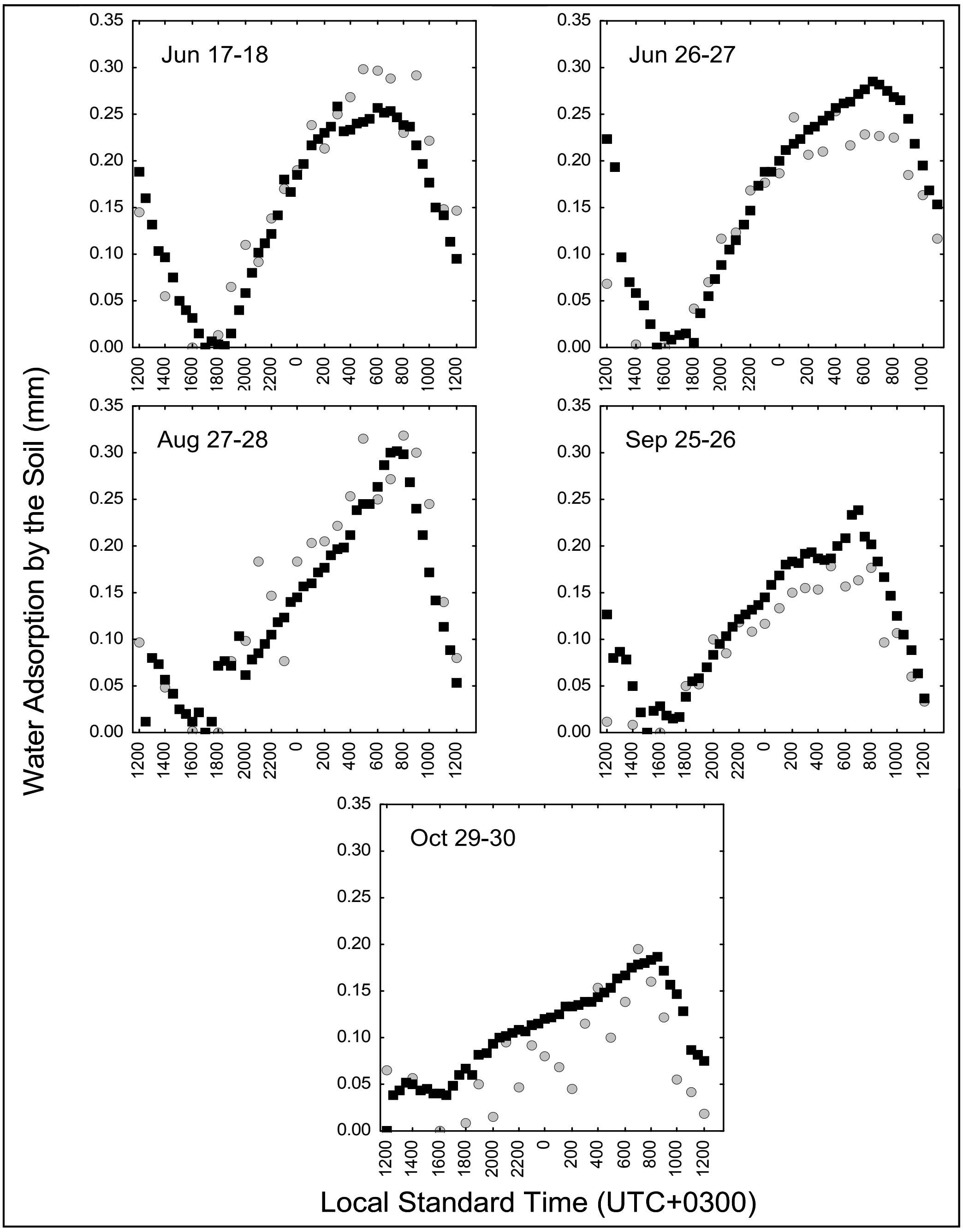


***Figure 3.4**: Diurnal patterns of total water content in the soil (expressed as equivalent water depth) as determined with the micro-lysimeter and with soil samples for five representative dates. The maximum depth to which computations were carried out for each date is described in the text.*

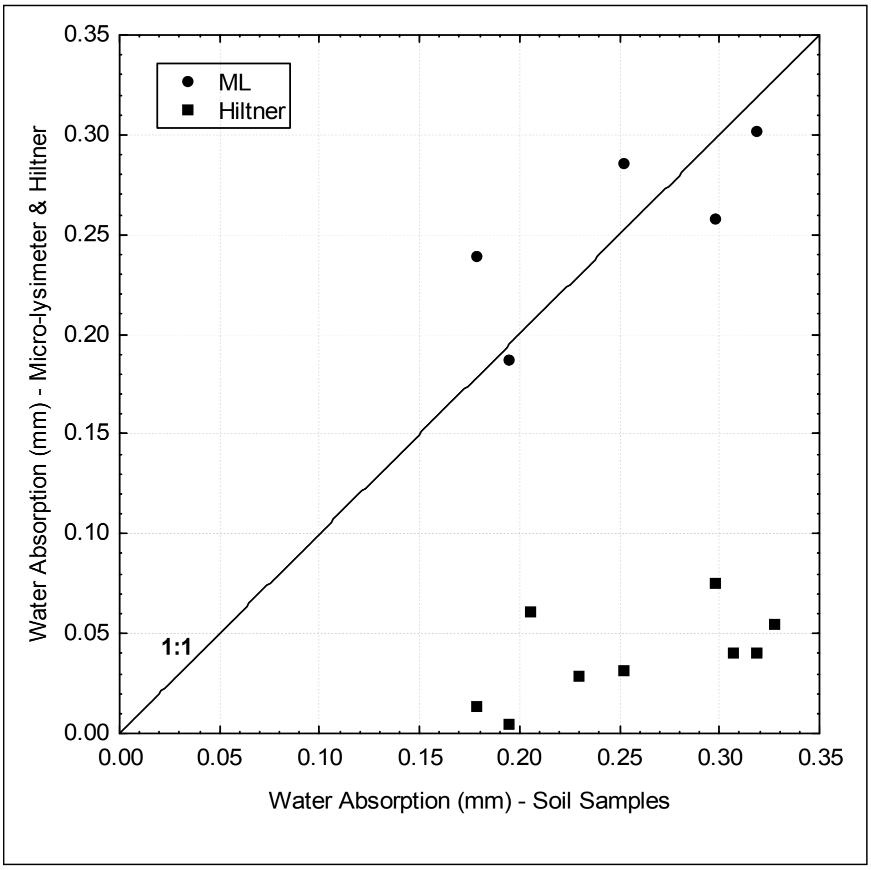


Figure 3.5: *Total amount of water added to the soil profile (expressed as equivalent water depth) during the night as measured using the micro-lysimeter and the Hiltner dew balance, as compared to the gains computed from water content changes in the soil samples.*

A salient feature of Figure 3.2 is that the maximum gravimetric water content of the first centimeter of the soil, which reaches for all campaigns a similar value (1.97± 0.08 %) and is apparently unrelated to the prevailing environmental conditions. In contrast, the minimum soil water content varies throughout the season and will therefore be the factor that determines the total amount of adsorbed water. The seasonal variation of the minimum soil water content suggests that it may be controlled by the atmospheric conditions and, therefore, should be linked to an index that integrates the various relevant factors. The "dryness of the atmosphere" is usually defined as being equal to the maximum rate at which water can be evaporated from a given surface under the prevailing atmospheric conditions (i.e. potential evaporation) and is computed using the Penman equation (Monteith, 1965). It was, therefore, hypothesized that the greater the "dryness of the atmosphere", the lower will the minimum water content be. The period during which the cumulative daily potential evaporation was computed began with the onset of evaporation (as determined from the micro-lysimeter measurements) and ended at the time at which the minimum water content in the uppermost soil layer was reached. In Figure 3.6 the plot of daily accumulated potential evaporation as a function of minimum measured daily water content in the uppermost soil layer is presented. A strong negative linear correlation was found (r=-0.93, p=0.0007).

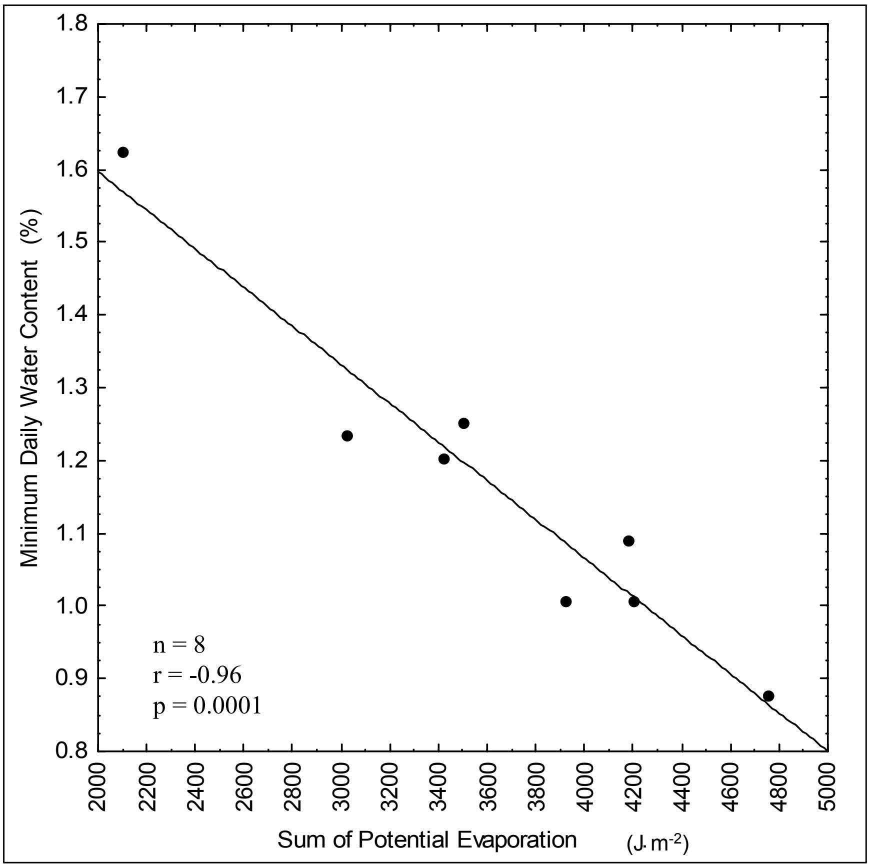


Figure 3.6: *Minimum daily water content as a function of the sum of potential evaporation from the beginning of the evaporation process until the minimum water content was reached.*

The strong dependence of the minimum water content on the total antecedent potential evaporation together with the previously mentioned fact that a relatively constant maximum water content was observed, implies that a correlation between the total antecedent potential evaporation and the amount of water adsorbed by the soil during the following “absorption period” must exist. In Figure 3.7 the total water gain is presented as a function of the antecedent potential evaporation. The correlation is, indeed, very strong (r=0.94, p=0.0005).

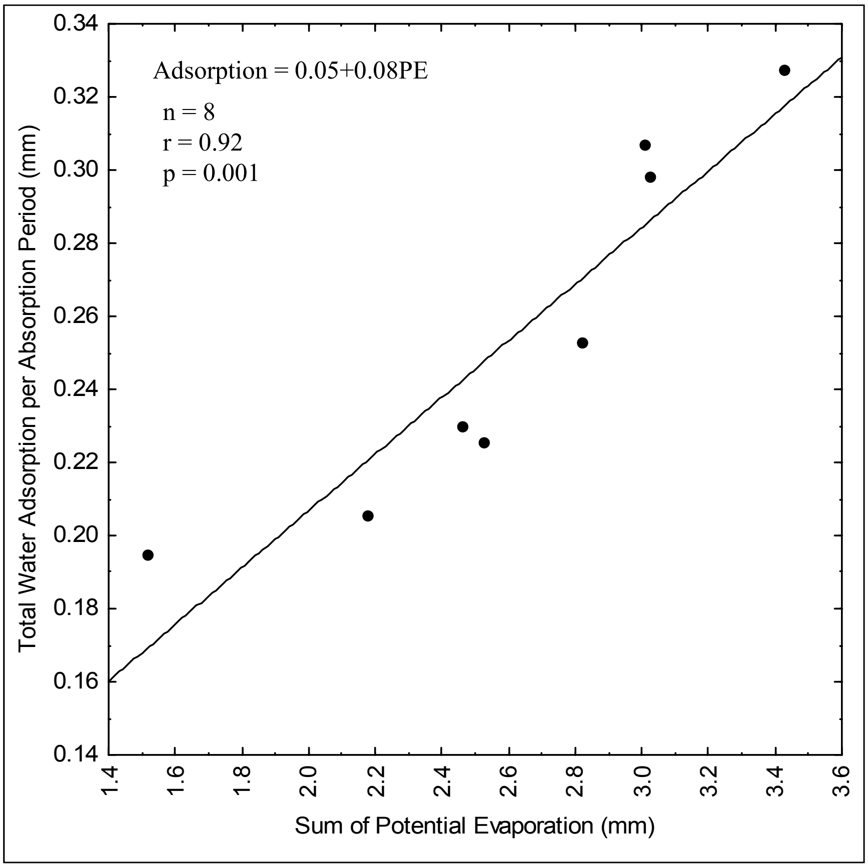


Figure 3.7*: The dependence of total water adsorption per night (expressed as equivalent water depth) on the sum of potential evaporation (PE) from the time when evaporation began to the time at which the minimum water content was reached on the day previous to the period during which adsorption takes place.*

For predictive purposes, however, this approach has one serious drawback, namely that the time of the day at which water absorption starts should be known. This information is usually not available. The link between the total water adsorption and the total daylight potential evaporation of the immediately preceding day was therefore tested. The results are presented in Figure 3.8 and the correlation is very strong and significant (r=0.95, p=0.0004). For a loess soil in the Negev desert, the total water adsorption during the dry period may, therefore, be determined using the following equation:

$$ApN = 0.09 + 0.04PE \tag{3.3}$$

in which ApN (mm) is the total daily water gain during the "absorption period" and PE (mm) is the total potential evaporation for the preceding period (sunrise to sunset) computed using the Penman equation.

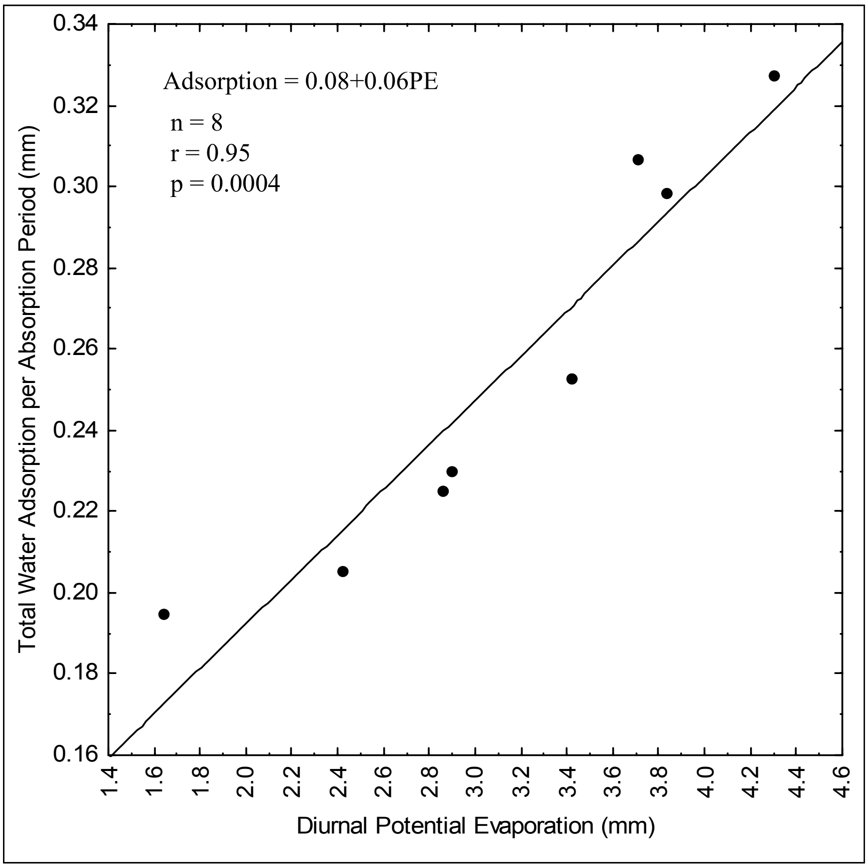


***Figure 3.8**: The dependence of total water adsorption (expressed as equivalent water depth) on the diurnal potential evaporation (PE) (total PE from sunrise to sunset on the day previous to the night during which adsorption was monitored).*

Kosmas et al. (1988) concluded that the diurnal fluctuations of soil moisture content due to water vapor adsorption increased with increasing clay content, a result that coincides with the findings of Li (2002). From the foregoing it may be concluded that equation (3.3) is probably site specific. Moreover, the influx of moist air during the late afternoon, which is a characteristic of the coastal desert in which this trial was carried out, may also contribute to the site-specificity of (3.3). These facts indicate that equations will have to be derived for different soil-climate combinations, but the approach should be valid for coastal deserts in the Southern Mediterranean. Moreover, the specificity is compensated because the proposed model is much simpler that the one proposed by Kosmas et al. (1988).

3.5. Summary and Conclusions

The results of this study indicate that in the area in which this study was carried out actual dew deposition on a bare soil surface is a rare occurrence. There is, however, a clear discernible daily cycle of water content in the upper soil layers. The lack of any evidence of soil surface wetting led to the conclusion that the main process responsible for the observed diurnal change in water content is the direct adsorption of water vapor by the soil. The maximum water content attained during each of the nine campaigns was relatively constant and independent of the prevailing meteorological conditions. Therefore, the minimum water content that was reached during the preceding day determines the total amount of water vapor adsorbed. A strong and significant correlation between the total adsorption of water vapor by the soil during the "absorption period" and the total potential evaporation between sunrise and sunset of the previous day was found. Based on this finding, an empirical model, based on commonly available data, is proposed to predict the total amount of water adsorbed by the soil during the absorption period. The proposed model is site specific but very simple and easy to implement.

The fact that there is no evidence of the deposition of liquid water on the soil surface has two additional consequences. The first is that even though the magnitude of the latent heat fluxes are larger than those previously reported, the long-held belief that dew is of prime importance in this type of ecosystem, needs to be re-assessed. The second consequence is the possible effects on remote sensing techniques. The reported changes in some of the remotely sensed data should be related to the small changes in the soil surface water content, as it appears that water drops are not formed on the surface of the soil.

An additional and interesting consequence of the finding of this study is that in coastal deserts there is a daily cycle of water vapor exchange between the soil and the atmosphere during the dry season. The latent heat flux density during the early morning may reach 20% of the net-radiation flux density and decreases thereafter. The magnitude of this flux is of the same order of magnitude as the expected soil heat flux density (Stull, 1988) and should, therefore, be considered when energy and water balances at the soil surface are computed for these regions.

Lastly, it can be stated that, in the area in which this study was carried out, artificial condensing plates of any kind cannot be used to evaluate dew deposition quantities on a bare soil surface.

Chapter 4

Potential Implementations

The Role of Water Content Changes in the Energy Partitioning

As published in the *Journal of Geophysical Research - Atmosphere*:

Agam (Ninari) N., Berliner P.R., Zangvil A. & Ben-Dor E. (2004). Soil water evaporation during the dry season in an arid zone. Journal of Geophysical Research – Atmosphere, 109, D16103, doi:10.1029/2004JD004802.

4.1. Introduction

Land surface processes in general, and the energy partitioning at the soil surface in particular, play an important role in global and meso-scale studies. These processes are usually integrated as sub-sets (or sub-models) of global and meso-scale models. During the last decade a number of land surface models have been developed (Yang et al., 1998). One of their main tasks is to describe the patterns of radiant energy dissipation at the land-atmosphere interface, which above bare soil are determined by the moisture content of the soil surface. It is, therefore, reasonable that the quality of land surface models should be judged by the accuracy with which they compute the aforementioned soil water content (Irannejad and Shao, 1998). The moisture level of the soil is the result of the interaction of atmospheric variables (radiation, temperature, wind speed, etc.) with the transport of mass and energy in the soil. The transport of mass and energy in the soil (and consequently evaporation of water from the soil surface) has been intensively studied, and various formulations and numerical solutions based on the general theory of Philip and de Vries (Philip and de Vries, 1957) have been presented (e.g., Milly, 1982, Milly, 1984, Kondo et al., 1990, Scanlon and Milly, 1994, Parlange et al., 1998, Qin et al., 2002). The use of such detailed descriptions is not appropriate for large scale modeling due to the vast amount of field data required (Yang et al., 1998) and the intense computational efforts involved. This problem is usually circumvented by parameterizing the evaporation flux as a function of the potential evaporation and some easily computed index of surface moisture (Shao and Henderson-Sellers, 1996).

One approach to the parameterization method is the supply and demand formulation, according to which the actual evaporation rate is computed from

$$E = \min\left\{E_p, E_c\right\} \tag{1}$$

in which E, E_p and E_c (W m^{-2}) are the actual evaporation, the potential evaporation and the maximum rate at which the soil profile can transport water to the soil surface, respectively. Even though this approach is simpler than simultaneously solving the coupled equations for mass and energy in the atmospheric boundary layer and in the soil, the drawbacks mentioned previously apply to the computation of E_c. The computation of the actual evaporation using formulae of the type:

$$E = \beta E_p \tag{2}$$

in which β depends explicitly on the soil water content is, therefore, favored and widely used (Shao and Henderson-Sellers, 1996).

Carlson et al. (1984) defined β as:

$$\beta \equiv \frac{R_b}{R_b + R_s} \approx \frac{\theta}{\theta_{sat}} \tag{3}$$

where R_b (m^{-1} sec) and R_s (m^{-1} sec) are the atmospheric and soil resistances respectively and θ and θ_{sat} the actual and field saturation values of volumetric soil water content. β varies from 0 (absolutely dry surface) to 1 (saturated surface), and its value is required for each grid point in the model. Carlson et al. (1984) suggested to derive β from the remotely sensed surface temperature and thermal inertia. The simplicity of the approach and the possibility to obtain maps of β using remotely sensed data are two attractive features of this approach. The linearity of the dependence appears, on the other hand, to be an oversimplification. As a result, an improved formulation, which accounts for the field capacity and the wilting point moisture contents of the soil, was incorporated in the land-surface model that is used as part of the fifth generation Meso-scale Model (MM5), jointly developed by Pennsylvania State University and the National Center for Atmospheric Research. In this model, β is computed using equation (4) (Chen and Dudhia, 2001):

$$\beta = \frac{\theta - \theta_w}{\theta_{ref} - \theta_w} \tag{4}$$

where θ is volumetric water content, and θ_{ref} and θ_w are field capacity and wilting point.

In this case as well, β varies linearly from 0 (no evaporation) to 1 (evaporation at a rate that equals the potential evaporation). This formulation is based on the implicit assumptions that the soil moisture content does not drop below the wilting point (as β cannot be negative), and therefore latent heat flux vanishes when the soil moisture content reaches the wilting point. Meso-scale models of this type were developed mostly for temperate climate zones (Bougeault, 1991), for which these assumptions are probably valid. Their applicability to other climatic regions should be considered with great care. A point in case is desert areas, for which it has been shown that the soil dries well below the wilting point (Boulet et al., 1997).

The detailed description of land-surface processes is not limited to meso-scale models, but is also a key feature of global circulation models. It was found that the exchanges of momentum, heat and moisture between the atmosphere and the earth's surface have a fundamental influence on the dynamics and thermodynamics of the atmosphere (Chen and Dudhia, 2001). Evidence from atmospheric general circulation model experiments suggests that climate is sensitive to variations in evaporation from the land surface (Schmugge and Andre, 1991). As the accuracy of the response of general circulation models is dominated by sub-grid scale parameterizations of inputs and parameters of land surface processes (Avissar and Pielke, 1989, Hu and Islam, 1997, Robock et al., 1998, Sridhar et al., 2003), the accuracy and sensitivity of these inputs are of great importance.

The parameterization method described above (i.e., parameterizing the actual evaporation according to the soil water content) is, however, very crude (Blondin, 1991). It was found that model simulation results are very sensitive to the soil parameters chosen for assessing β (Shao and Henderson-Sellers, 1996, Lohmann et al., 1998) especially during nighttime (Irannejad and Shao, 1998) and when the soil is relatively dry (Yang and Dickinson, 1996, Yang et al., 1998). Notwithstanding the uncertainties of this type of parameterization, it is still widely used (e.g., European Center for Medium-range Weather Forecast, 2002).

In general, land surface models differ only slightly from each other, and are based on different degrees of simplification of the detailed processes (Yang et al., 1998, Gusev and Nasonova, 2003). The urgent need is, therefore, not to develop "new" land surface models, but to evaluate and test the available models with field data. Moreover, high quality field data are crucial for improving land surface models (Yang et al., 1998).

Very few data sets that address the problematic environmental conditions stated above are available. Usually, in arid zones measurements of evaporation are carried during the period during which the soil remains wet in the wake of rainfall events (Mitsuta et al., 1995), the rationale being that during the dry period evaporation is negligible. Surprisingly, however, diurnal changes in the water content of the uppermost soil layer (with a corresponding latent heat flux density) were monitored in the desert even during the dry season (Agam (Ninari) and Berliner, in press). The commonly used approach to the parameterization of evaporation discussed previously (eq. 2), may in these circumstances, lead to significant errors in the assessment of the radiation dissipation patterns in arid zones. The objective of this study was to assess the relative magnitude of latent heat flux density over a bare loess soil in the Negev desert throughout the dry season, during which the atmospheric models usually assume the lack of latent heat flux.

4.2. Materials and Methods

The measurements were carried out at the Wadi Mashash Experimental Farm in the Northern Negev, Israel (31^o 08’ N, 34^o 53’ E; 400m A.M.S.L., 60Km from the Mediterranean Sea). Rainfall events occur between October and April, and the mean annual rainfall at the farm is 115 mm. Long-term maximum and minimum temperatures are 14.7^oC and 4.8^oC for January; and 32.4^oC and 18.6^oC for July, respectively. Class A pan evaporation is 2500-3000 mm per year. The soil is a sandy loam Aridisol (Loess) with 13% clay, 15% silt and 72% sand. Data was collected during nine 24-h field campaigns that took place during the dry season of 2002. A total of 124 mm of rain were recorded during the rainy season of 2001-2002 (prior to the above mentioned measuring period). The first campaign took place 10 weeks after the last rainfall of the 2001-2002 season (March 29, 13.5 mm), and the last one ended about two hours before the first rainfall of the next season (October 30). The remaining 7 campaigns were randomly spread within these two dates.

During each campaign, a micrometeorological station was installed near-by, in a spot in which the fetch in the prevailing wind direction (N-NW) is 200 m and devoid of vegetation, for continuous measurement of incoming and reflected short-wave radiation with two pyranometers (CM5, Kipp & Zonen); net-radiation (Q-7, Campbell Scientific Inc.); wind speed at four levels (2, 1, 0.5, 0.25 m) with cup-anemometers (014A Met-One); dry- and wet-bulb temperatures at 1m height using a self-designed aspirated psychrometer; soil heat flux at three different locations in the field with heat flux plates (HFT-3, Campbell Scientific Inc.) installed at a depth of 50 mm, and temperature measurements above them at 10 mm intervals, using differentially wired thermocouples. Sensible heat flux was measured with a sonic anemometer (CA27, Campbell Scientific Inc.). Data was measured and collected every 10 seconds and averaged every 30 minutes by a data-logger (23X, Campbell Scientific Inc.). Additionally, the changes in mass of an improved micro-lysimeter (Ninari and Berliner, 2002) (186 mm diameter and 550 mm of effective depth with an additional 50 mm of polyurethane insulation) were recorded by placing the micro-lysimeter in a pit and weighing it every half hour. The PVC tube containing the undisturbed soil sample was isolated by polypropylene to avoid lateral heat flux. The scale (AND, maximum weighting capacity of 30 kg) had a resolution of 0.1 gram, which resulted in a resolution of 0.004 mm (equivalent depth of water) or 5.11 W m^{-2} (in energy terms). The output of the scale was registered automatically every half hour by a palm computer (48GX, Hewlett Packard).

Thermal images of the micro-lysimeter surface together with its surroundings were acquired hourly during several campaigns throughout the season, using a thermal video radiometer (TVR) (INFRAMETRICS 760, 1994). The TVR is highly sensitive in terms of both radiometric temperature and spatial resolutions of the surfaces (±0.05° C and about 0.035 m from height of 20 m, respectively). For the current study, the sensor configuration was optimized to work across the spectral region of 3–14 μm, and the optic configuration enabled a field of view (FOV) of 20° along an instantaneous field of view (IFOV) of 1.8 mRad. Using an onboard internal calibration procedure, the calibration information for each scene of the raw data was recorded on an 8 mm NTSC magnetic tape. Selected images were saved as TIFF files and further processed with image processing software. Analysis and results of five out of the nine campaigns, for which complete sets of data are available, are presented.

4.3. Data Analysis and Evaluation

As the dates of the campaigns were selected randomly, it was important to substantiate that they are representative. Data of prime meteorological parameters (i.e. incoming short wave radiation, air temperature, relative humidity and wind speed) were collected from a meteorological station located not far from the research site (and influenced by the same meso-scale conditions). Data measured during the campaigns were compared to the corresponding monthly averages. Campaign dates in which the values measured were within the range of ±1 standard deviation from the monthly average were considered representative. The diurnal changes in the relative humidity measured for each of the campaigns, together with the monthly averages for the same time intervals, are presented as an example in Figure 4.1. The relative humidity during the campaigns was within 1 standard deviation from the monthly average, except for a few hours at noon of August 4. Similar patterns were found for the other parameters. It was concluded, therefore, that the meteorological conditions during these campaigns (of which the dates were randomly selected) are representative of the conditions for the season they stand for.

Various techniques have been used for measuring latent heat flux, most of them relying on micro-meteorological methods (Brutseart, 1982). However, in arid environments, the magnitude of latent heat flux during the dry season is very small, which poses special technical measurement difficulties (Ninari and Berliner, 2002). In such conditions, even small errors in the parameters used in those methods may result in errors which are of the order of magnitude of the flux itself. Direct methods for measuring latent heat flux are, therefore, advantageous. Theoretically, the use of a micro-lysimeter provides an absolute reference for latent heat fluxes, as long as the soil and the heat balance of the micro-lysimeter are similar to those of the surrounding area. Provided the soil sample is undisturbed and representative of the area, similar temperature profiles will yield equal surface temperatures and hence guarantee that the latent heat fluxes measured with the micro-lysimeter represent the surrounding soil (Ninari and Berliner, 2002).

The representativity of the micro-lysimeter was tested during several campaigns by comparing its surface temperature to that of the surrounding area. Thermal images of the micro-lysimeter surface together with its surroundings were acquired hourly. Figure 4.2 is an example for the acquired images.

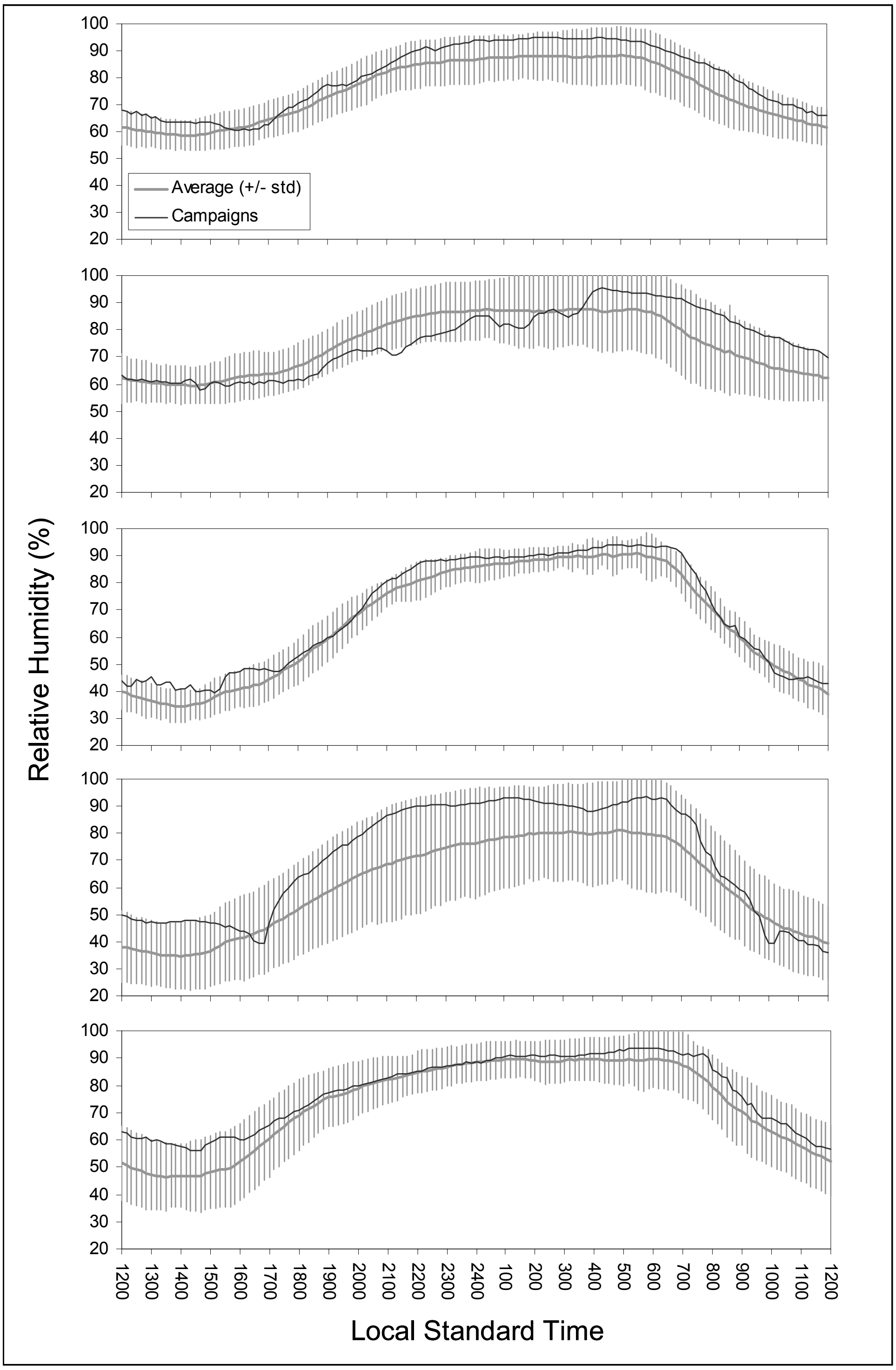


***Figure 4.1**: Diurnal changes in the relative humidity measured for each of the campaigns, together with the monthly averages (±1 standard deviation) for the same time intervals are presented. Data was collected from a meteorological station located not far from the research site (and influenced by the same meso-scale conditions).*

Figure 4.2: *A sample of a thermal image of the micro-lysimeter and its surrounding, acquired at mid-night June 26-27. The gray level indicates the surface temperature (the brighter the color, the warmer the surface). The two rectangles mark the subsets from which the pixels data were extracted.*

The representativity of micro-lysimeters containing undisturbed soil cores usually decreases with time, and in studies with wetting cycles (rain, irrigation) it is common to replace the sample. The disadvantage of this procedure is that it is time consuming and probably affected by the spatial variability of the soil characteristics. In this study, one core was used throughout the entire dry period, and its representativity was assessed shortly after insertion (June 26-27) and towards the end of the measuring season (September 25-26). For the selected campaigns, two subsets of the images (indicated in Figure 4.2 by the marked rectangles) were produced, representing the surface of the micro-lysimeter's soil sample and the surrounding soil. The rectangle representing the micro-lysimeter contained 1071 pixels and the one representing an undisturbed area in the surrounding soil 33472 pixels. For each date, a set of 20 pairs of subsets were analyzed to compare the surface temperature of the micro-lysimeter sample to the surrounding soil.

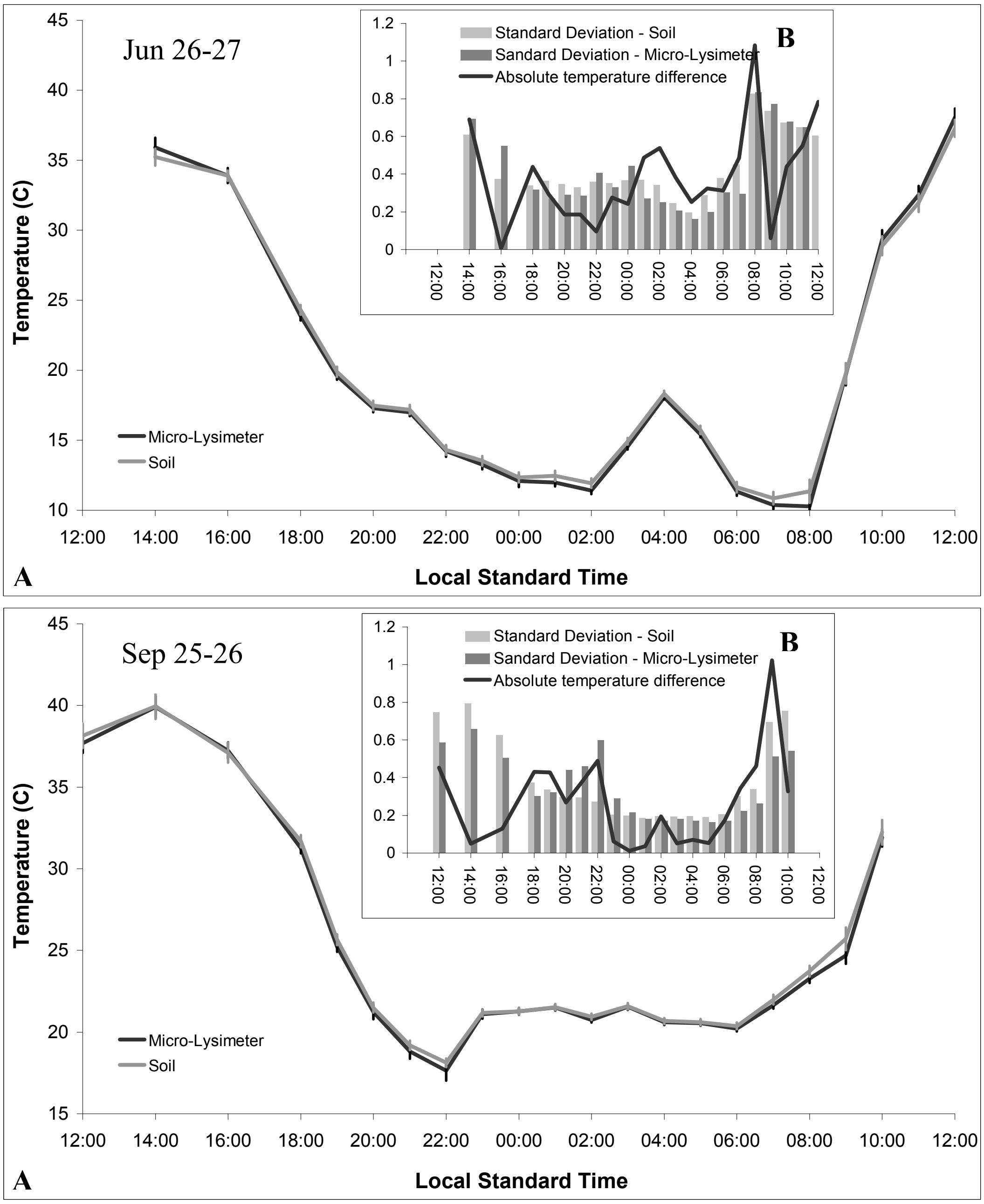


Figure 4.3*: The daily course of the average temperature (± standard deviation) of the soil and the micro-lysimeter (a) together with the absolute differences between the average temperatures of the soil and the micro-lysimeter compared to the standard deviation from their averages (b), for June 26-27 and September 25-26.*

The daily course of the average temperature (± standard deviation) of the soil and the micro-lysimeter are presented in Figure 4.3a. Figure 4.3b shows the absolute differences between the average temperature of soil and the micro-lysimeter compared to the standard deviation from their averages. Performance of a goodness-of-fit test showed that the daily course of temperatures of both the micro-lysimeter surface and the soil surface was similar (For June 26-27, χ^2=0.483, and for September 25-26 χ^2=0.271). During most of the day the temperature difference between the sample and the soil are less than 0.5 ^{o}C. A peak difference of 1 ^{o}C in the morning (9:00 in June and 8:00 in September) is due to a shadow partially falling on the micro-lysimeter sample for a very short time (because of technical constraints). Moreover, during most of the day, the standard deviations (for both the sample and the soil) are of the same order of magnitude as the temperature differences between them. It can, therefore, be concluded that the soil sample in the micro-lysimeter is representative of the surrounding soil, and that the changes in soil water content measured by the micro-lysimeter are reliable and representative of the changes in the soil water content (and thus the latent heat flux) of the surrounding soil.

In order to assess the relative contribution of the latent heat flux to the energy balance at the soil surface it is, however, important to carefully evaluate the quality of the entire energy balance data. The energy balance at the soil surface is described by equation (5)

$$NR + G + H + E = 0 \tag{5}$$

in which *NR* is net-radiation and *G*, *H*, *E* are soil, sensible and latent heat fluxes, respectively. All components of the energy balance equation are in (W m^{-2}).

An additional component of the energy balance that should be carefully treated is the soil heat flux. Although it's daily mean value is often one or more orders of magnitude smaller than the remaining terms in the energy-balance equation (5), this is not the case during shorter periods of time, during which it may be one of the dominating fluxes (Ninari and Berliner, 2002). Soil heat flux plates were inserted in three different places within the measurement site at a depth of 50 mm. The soil heat flux was computed as the sum of the soil heat flux measured at 50 mm depth (G_p) and the heat stored in the uppermost 50 mm of the soil (G_s):

$$G = G_p + G_s \tag{6}$$

The heat storage was computed using equation (7):

$$\begin{aligned} G_s &= \sum_{j=1}^{5} g_{j+\frac{1}{2}} \\ g_{j+\frac{1}{2}} &= Cv \frac{\left(T_{j+1}^{i} + T_{j+1}^{i+1}\right) - \left(T_{j}^{i} + T_{j}^{i+1}\right)}{2} \frac{\Delta Z}{\Delta t} \end{aligned} \tag{7}$$

in which $g_{j+\frac{1}{2}}$ is the mean heat gain/loss for a soil layer of thickness ΔZ (=0.01 m) between depths *j* and *j+1* for time interval Δt (=3600 sec) (between *i* and *i+1)*; *Cv* – volumetric heat capacity of the layer (J K^{-1} m^{-3}); T – soil temperature (an average of the three locations) (K).

Finally, an assessment of the quality of the complete energy balance is important. The degree of closure of the energy balance is a common approach for validating data. Complete closure $\left(H + E = -\left(NR + G\right)\right)$ is rarely, if ever, attained (Turnipseed et al., 2002). A scatter plot of (H+E) versus (NR+G) together with the corresponding regression analysis is presented in Figure 4.4. A slope of 0.9 and a correlation coefficient of 0.93 were computed for the present data set. The lack of closure is most likely the result of measurement errors, probably due to the fact that each of the four fluxes was measured using a different instrument. However, when this result is compared to the degree of closure reported in the literature by others (partial list in Table 4.1) the degree of closure achieved in the present study can be deemed as satisfactory.

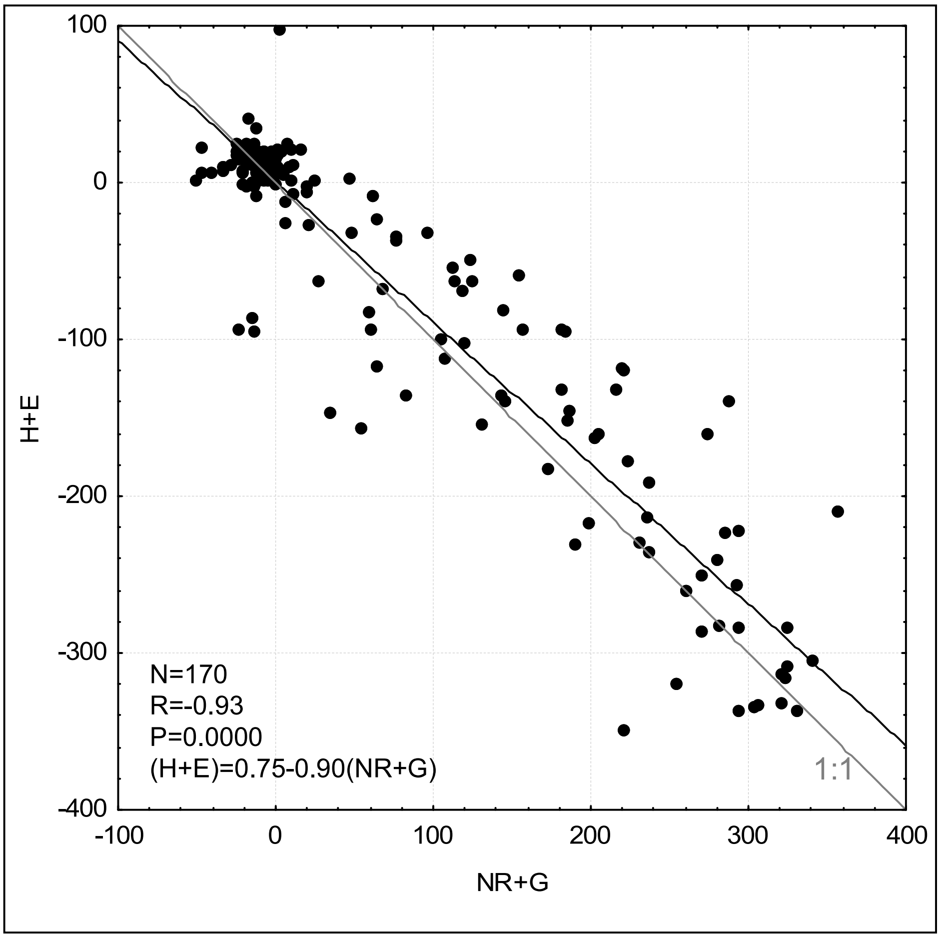


Figure 4.4: *The sum of sensible (H) and latent (E) heat flux densities versus the sum of net-radiation (NR) and soil heat flux density (G) together with the corresponding regression analysis. A slope of 0.9 and a correlation coefficient of 0.93 indicate satisfactory closure.*

Table 4.1*: Reported closure achievement*

Source	Degree of Closure
(Ma et al., 2003)	70%
(Turnipseed et al., 2002)	75-95% with average of 85%
(Tanaka et al., 2001)	70%
(Anthoni et al., 2000)	70-80%
(Unland et al., 1996)	Mention a regression coefficient of 0.96 without any information about the slope.

4.4. Results and Discussion

The matric potential corresponding to the maximum daily moisture content of the uppermost soil layer is less than -400 bar (computed using the Van Genuchten formulation (van Genuchten, 1980) using the coefficients suitable to the studied soil). However, even though the matric potential of the uppermost soil layer is extremely low, a clear daily cycle of water content change in the soil can be observed. In Figure 4.5, the differences between total soil water content and the maximum daily total water content (mm water) are presented. The largest diurnal change in the soil water content was measured in July, and the smallest in October. The soil matric potential at wilting point, which, as has been mentioned, is used in many models as the critic value for “latent heat flux shut down”, is commonly set at -15 bar. For the conditions prevalent at the research site, throughout the entire dry season, the soil is drier than the wilting point and most models would predict no latent heat flux.

The most important micro-meteorological parameters, which are expected to affect the diurnal change in soil water content, are presented in Figure 4.6 for the corresponding five campaigns. The high level of short wave radiation (Figure 4.6a) is a characteristic of the Negev desert in which the measurements were carried out (Berliner and Droppelmann, 2003). For all campaigns, except October 29, the short wave radiation reached a maximum at ~13:00. A very similar radiation regime can be observed for June-August, for which the highest radiation flux densities were measured, with a notable decrease in September. The abrupt course of the incoming radiation flux density during October 29 indicates that it was a cloudy day. The high levels of incoming radiation during the three first campaigns and the decrease from August to October agree well with the seasonal pattern found for the 24-h minimum water content. In general, the higher the radiation level was, the lower was the minimum water content.

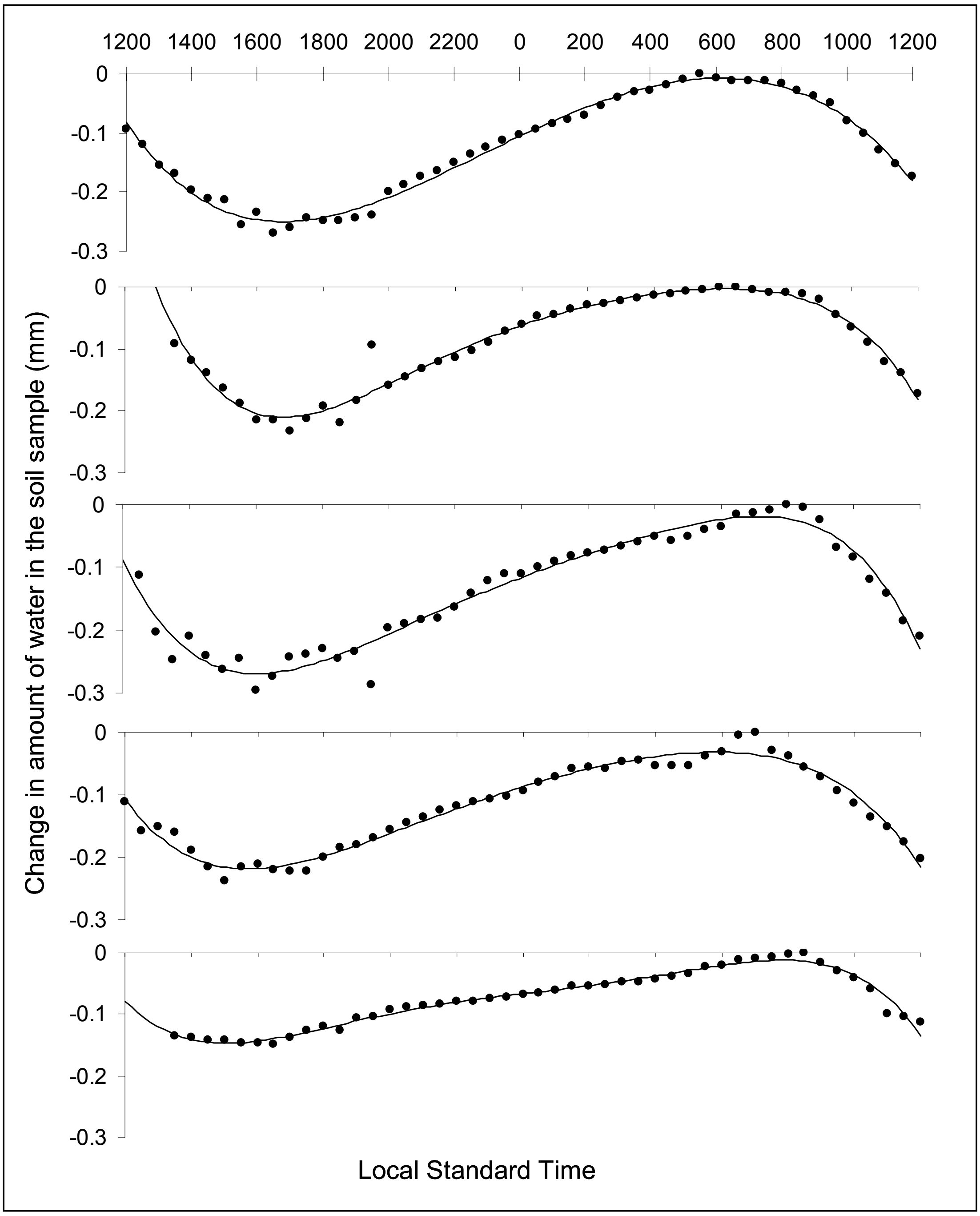


Figure 4.5: *The differences between total soil water content and the maximum daily total water content (mm water), as measured in June 16-17, July 17-18, August 4-5, September 25-26 and October 29-30.*

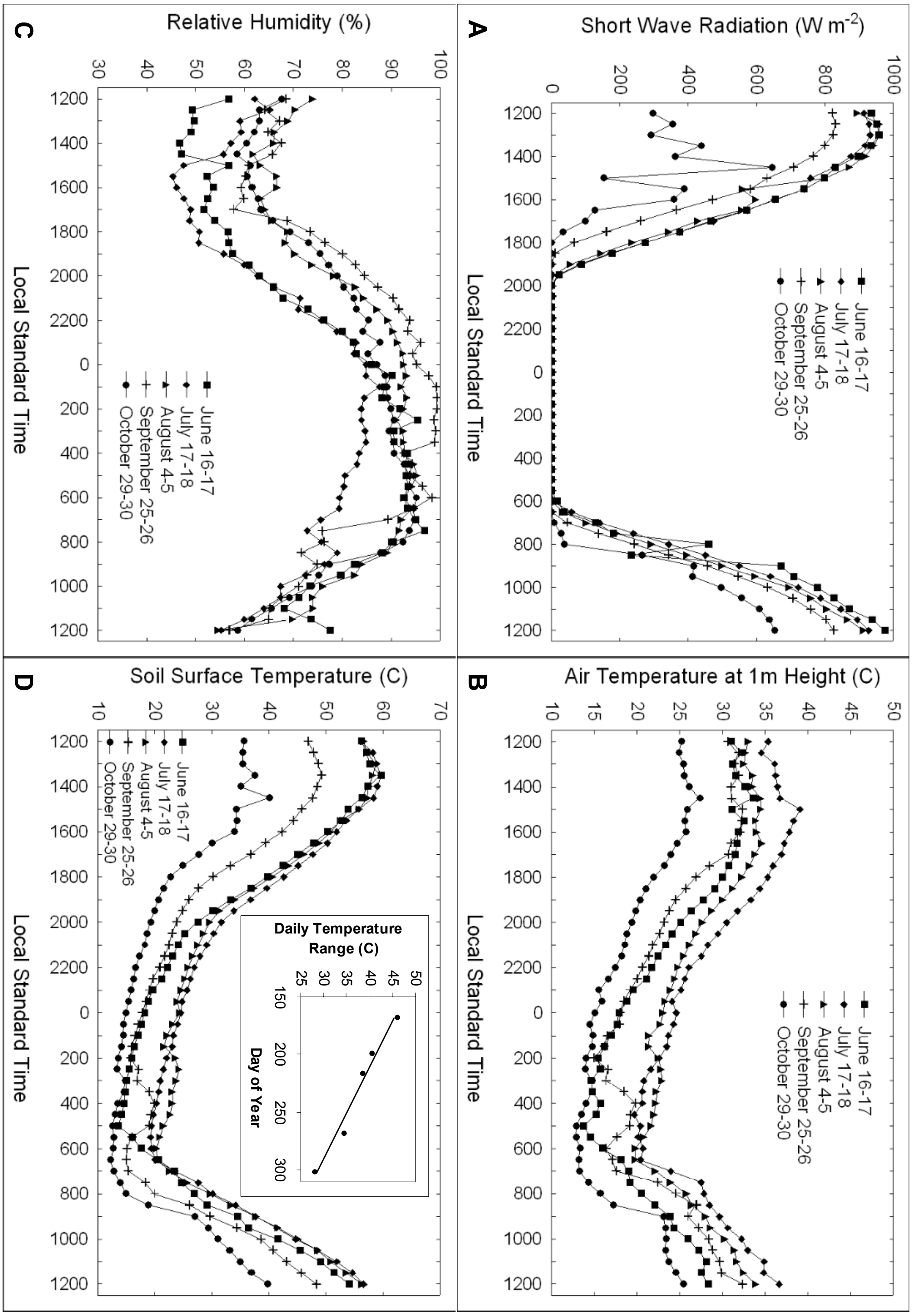


***Figure 4.6**: The daily course of important micro-meteorological parameters: (A) incoming sort-wave radiation, (B) air temperature, (C) relative humidity, and (D) soil-surface temperature, as measured in June 16-17, July 17-18, August 4-5, September 25-26 and October 29-30.*

The highest air temperatures were measured during the June campaign, and the lowest temperatures during the October campaign (Figure 4.6b). Maximum air temperature was measured at ~16:00, 3-hr later than the radiation peak. The wide peak during October 29 is the result of the presence of clouds. Minimum air temperature was monitored close to sunrise. The diurnal temperature range was large $\left(17.18 \pm 2.50\ ^{o}\mathrm{C}\right)$, as can be expected in a desert environment. No significant changes in the daily temperature range were observed for the measurement days.

The daily course of relative humidity (*RH*) (Figure 4.6c) has no clear seasonal trend. For all five campaigns, minimum *RH* was reached at the approximate time of maximum air temperature (14:00-16:00). A steep increase occurred thereafter, until midnight. During none of the nights, relative humidity reached 100%. It was very close to saturation for a few hours during the night of September 25-26. The drier night was July 17-18, but even during this night the *RH* was ~90% for a short period. The fact that saturation was not reached indicates that no fog event occurred during the campaigns. There seems to be no clear correlation between the daily course of *RH* and the change in the topsoil water content.

The soil surface temperature (Figure 4.6d) reached a daily maximum slightly later (~30 minutes) than the maximum incoming radiation, and slightly earlier (~30-60 minutes) than the minimum water content. The highest soil surface temperatures were measured during June-August. A notable decrease occurred from August to October. The ranges between maximum and minimum diurnal temperatures for the five dates are as well plotted in Figure 4.6d. Maximum range (47^{o}C) was observed on June 16-17 (from 59.5^{o}C at noon to 13.5^{o}C just before sunrise). A significant decrease in the diurnal range can be seen, with a minimum of ~28^{o}C on October 29-30.

The components of the energy budget at the soil surface are presented in Figures 4.7a-e. Fluxes directed towards the soil surface are defined as positive. Due to instrumentation breakdown, no data of the sensible heat flux density is available for June 16-17. The net-radiation level, during both daytime and nighttime, changed throughout the season in a pattern similar to the one exhibited by the incoming short wave radiation, decreasing (in absolute values) from August to October. A corresponding decrease is observed for the sensible heat flux density. To a lesser extent, the same pattern occurs for the soil- and the latent- heat flux densities.

During daytime, a large fraction of the net-radiation was dissipated as sensible heat. However, the soil and the latent heat flux densities are not negligible. During nighttime, the soil heat flux density is the most dominant component of the energy balance, the sensible heat flux density is very small, and in most campaigns the latent heat flux density is significantly larger than the sensible heat flux.

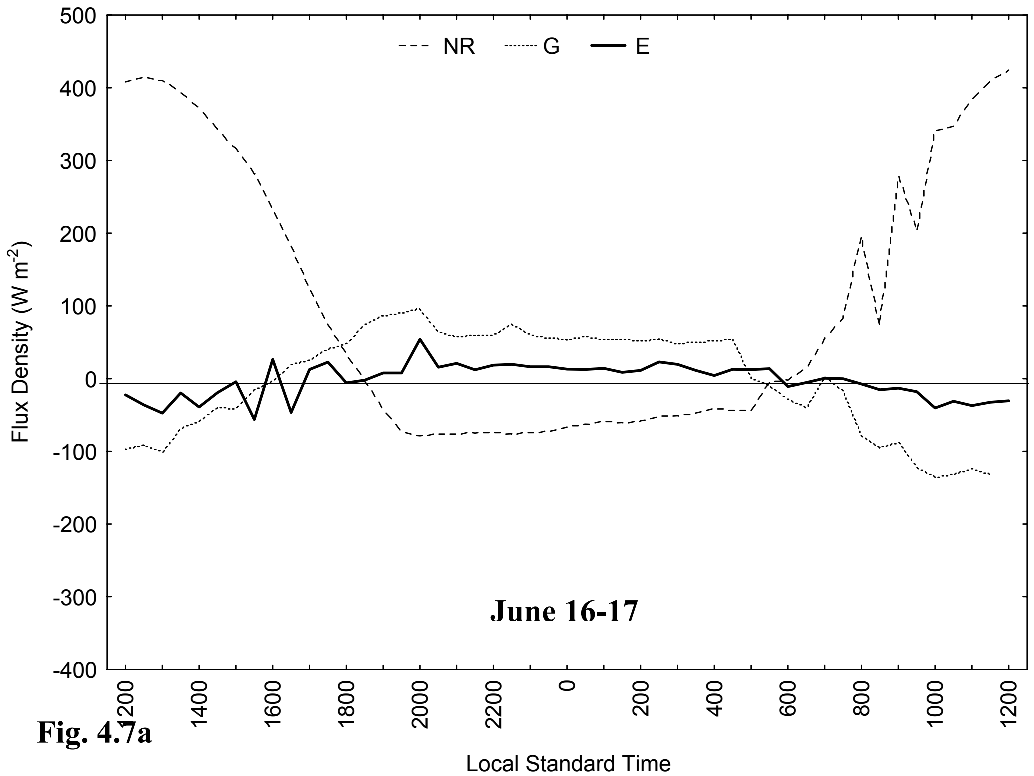


Fig. 4.7a

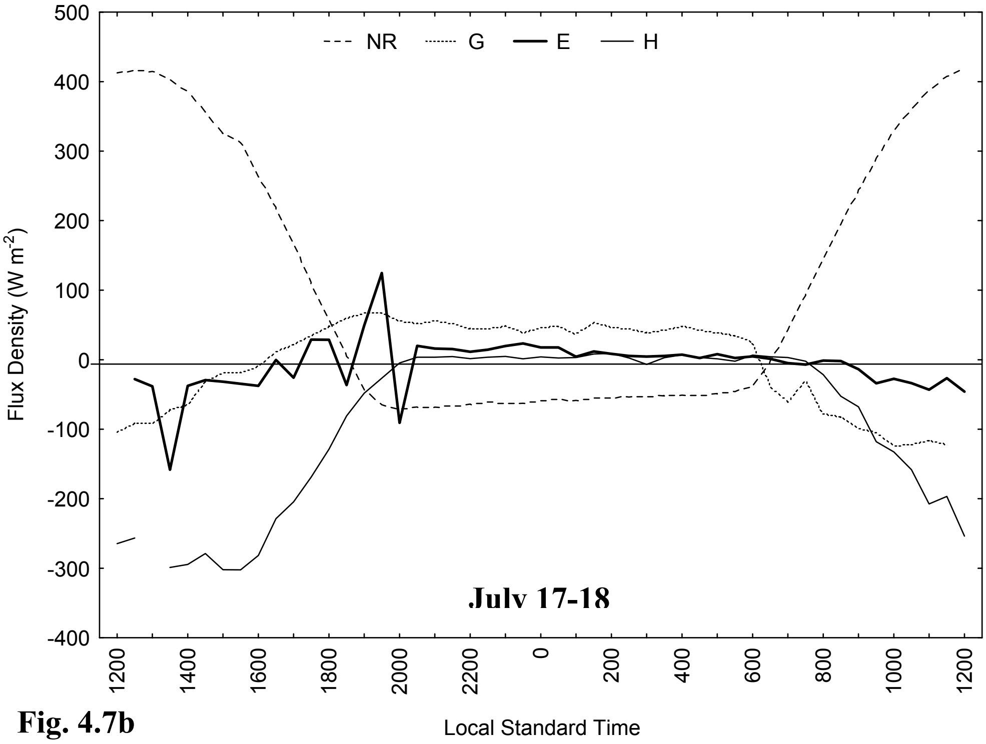


Fig. 4.7b

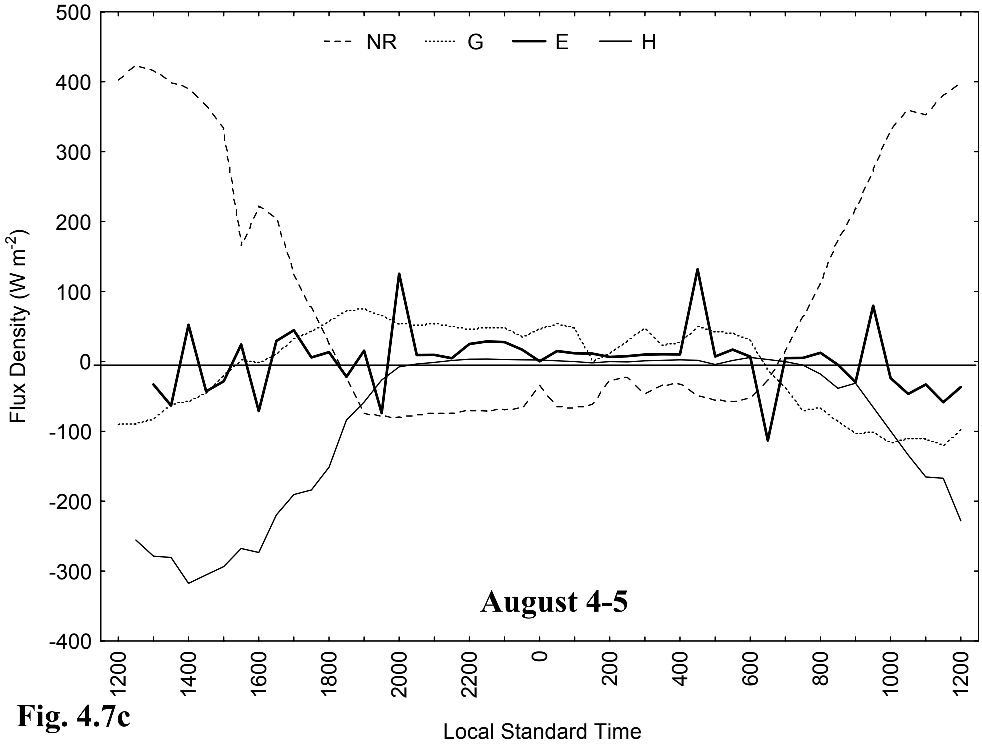


Fig. 4.7c

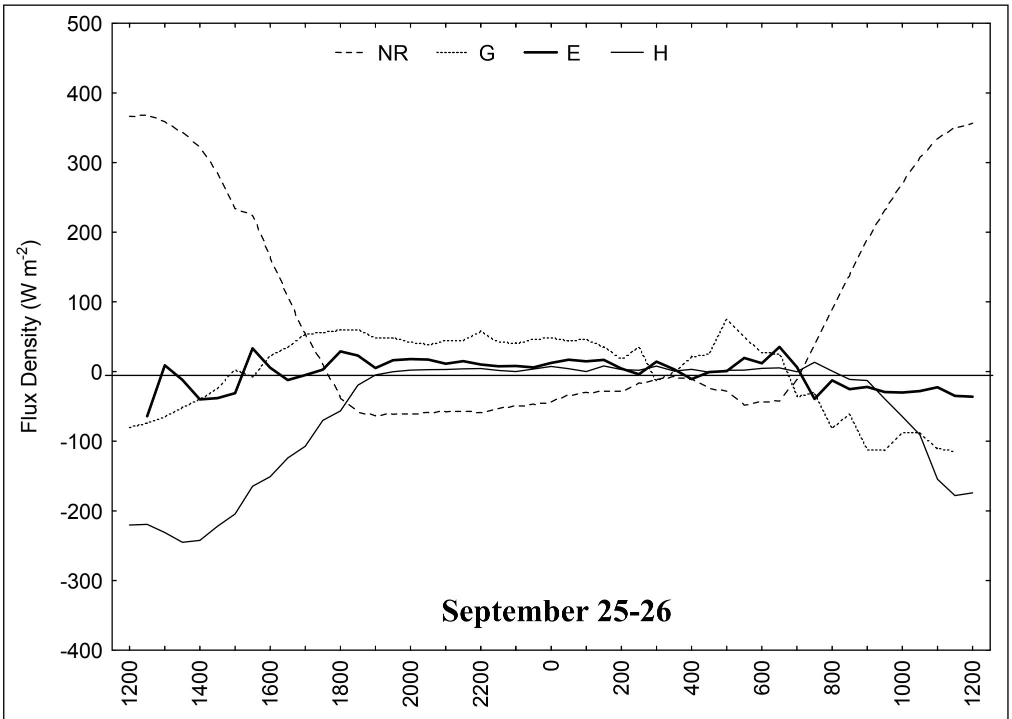


Fig. 4.7d

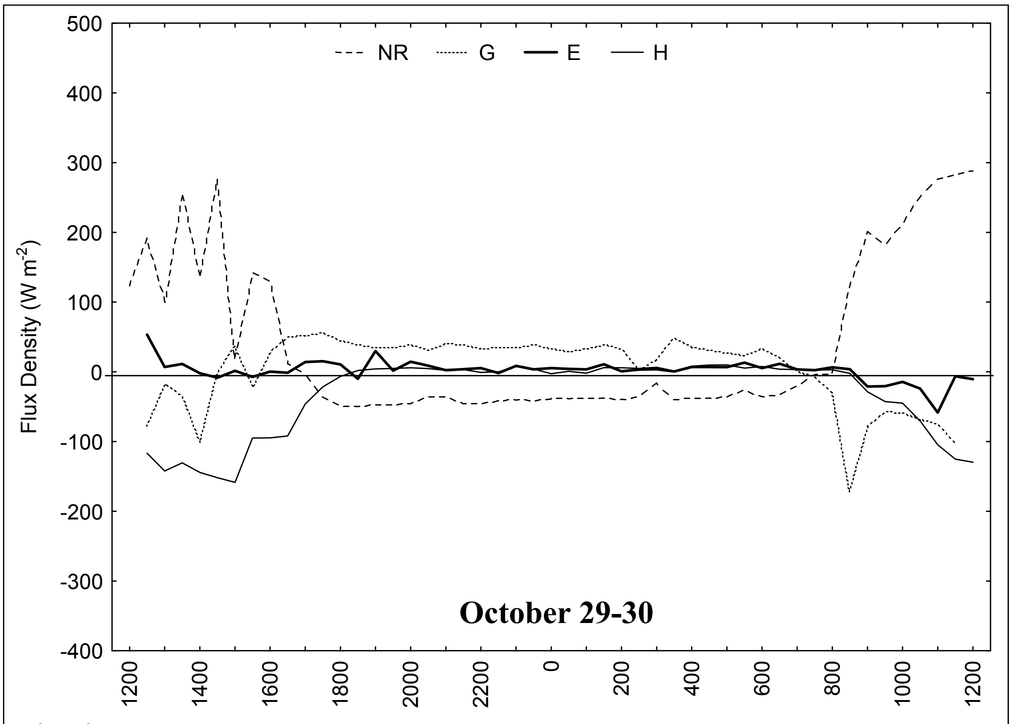


Fig. 4.7e

Figure 4.7*: The daily course of the energy balance components, as measured during the five 24-h campaigns, throughout the dry season of 2002.*

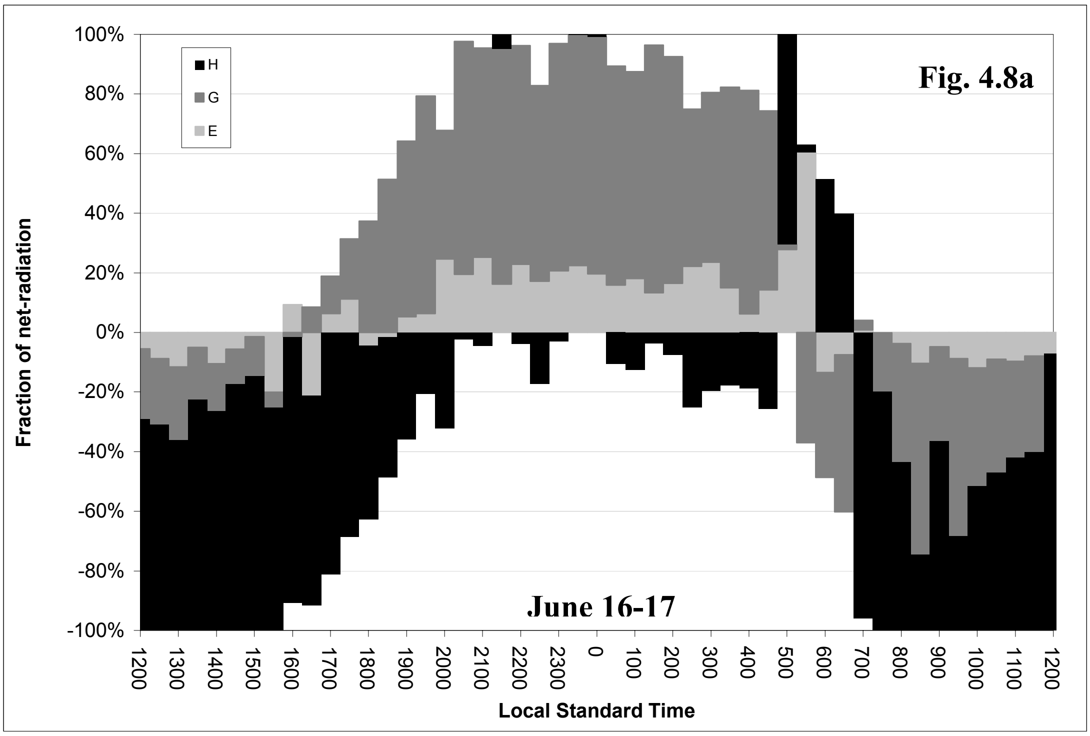
The average values of the Bowen ratio, computed from the above-mentioned fluxes, are presented in Table 4.2. These values were computed for several representative hours during the day and the night for each of the campaigns. These values confirm what can be visually interpreted from Figure 4.7. The Bowen ratio is consistently smaller than 1 during the night, a fact that indicates that the magnitude of the latent heat flux density during these hours is, indeed, significantly larger than the sensible heat flux density.

***Table 4.2**: Average values of the Bowen ratio – nighttime, daytime and total averages computed from several representative hours for each of the dates.*

Date	Day	Night	24-h
Jul 17-18	6.89	0.33	3.08
Aug 05-06	7.08	0.16	3.39
Sep 25-26	8.83	0.49	4.12
Oct 29-30	6.75	0.91	3.28
Average	**7.39**	**0.47**	**3.47**

In order to assess the role of the latent heat flux density in the energy balance, it is necessary to compare its magnitude to the magnitude of the sensible and soil heat flux densities. Figures 4.8a-e present the sensible-, latent- and soil- heat flux densities as fractions of *NR*, (i.e. *H*+*G*+*E*=100%). Since no direct measurements of sensible heat flux density were available for June 16-17, it was derived from the energy balance equation.

In the afternoon (~12:00-17:00) the sensible heat flux density is the dominant flux, being approximately 70% of the net radiation. The interesting fact is that the remaining 30% are evenly split between the latent and the soil heat flux densities (i.e., ~15% each). During the late afternoon (~17:00-20:00) the sensible heat flux direction remains the same, from the soil surface towards the atmosphere. The soil heat flux direction, however, has already changed and is now directed towards the soil surface. While the sensible heat flux density decreases, the soil heat flux density increases. The latent heat flux density during these hours does not show a clear trend, but its magnitude remains in the range of 10-15% of the net-radiation. During the night (~20:00-06:00) the soil heat flux dominates the scene, being approximately 70% of the net-radiation. The sensible heat flux density during these hours was only about 10% of the net-radiation, and the latent heat flux density was about 20%. In the morning (~06:00-12:00) the soil heat flux changes direction and decreases gradually, while the sensible heat flux changes direction about one hour later and increases gradually. The fraction of the latent heat flux remains approximately constant, about 10-15% of the net-radiation.



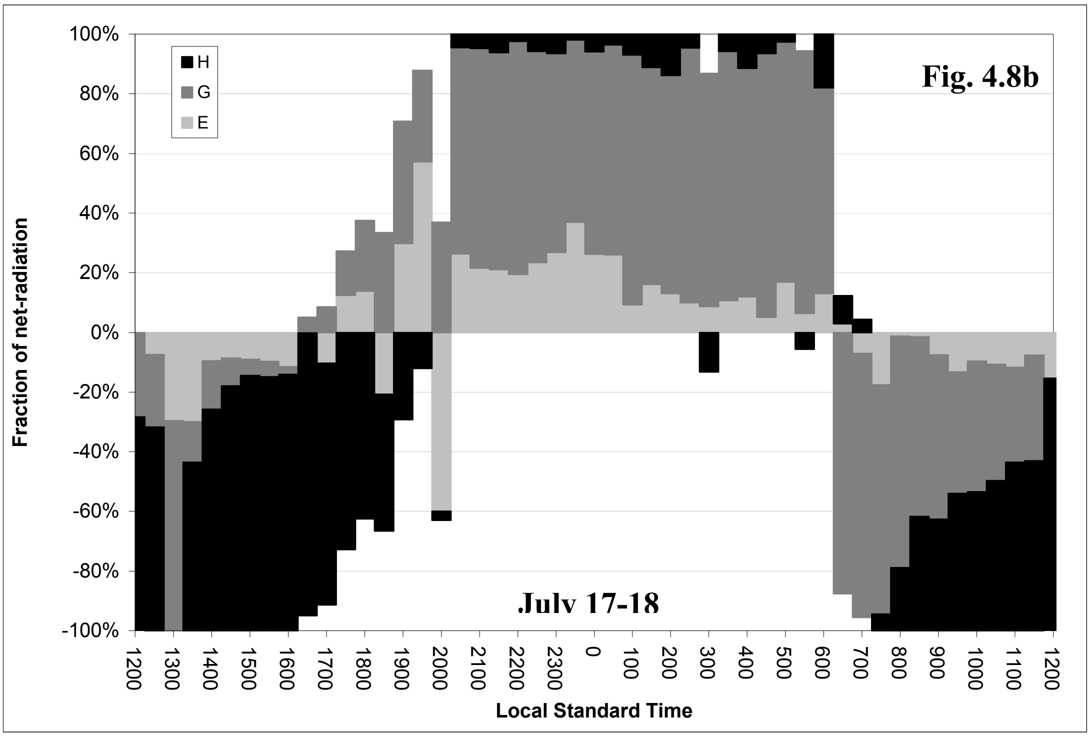


Fig. 4.8b
July 17-18
H
G
E
Fraction of net-radiation
Local Standard Time

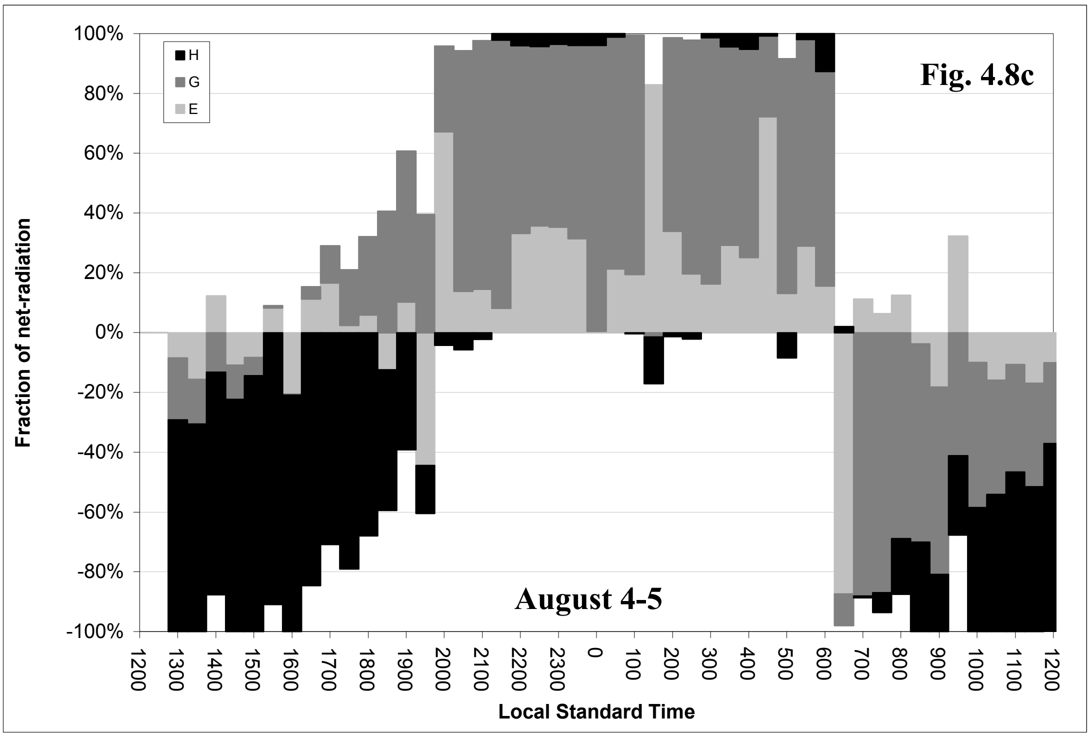
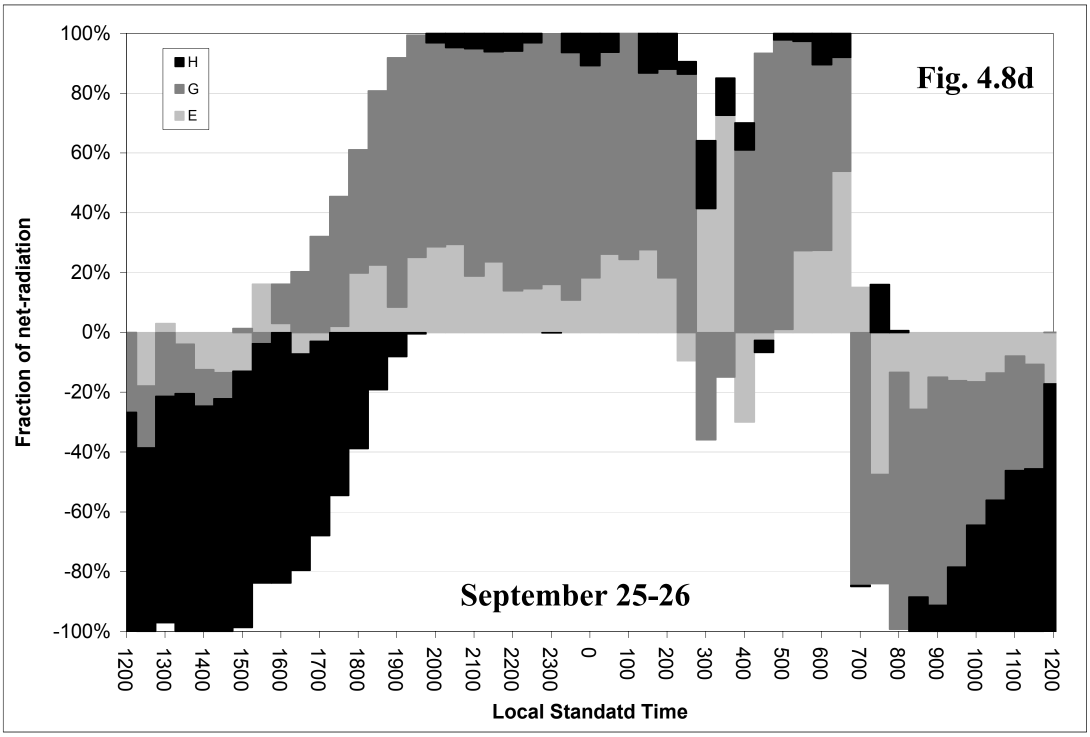


Fig. 4.8c
August 4-5
H
G
E
Fraction of net-radiation
Local Standard Time



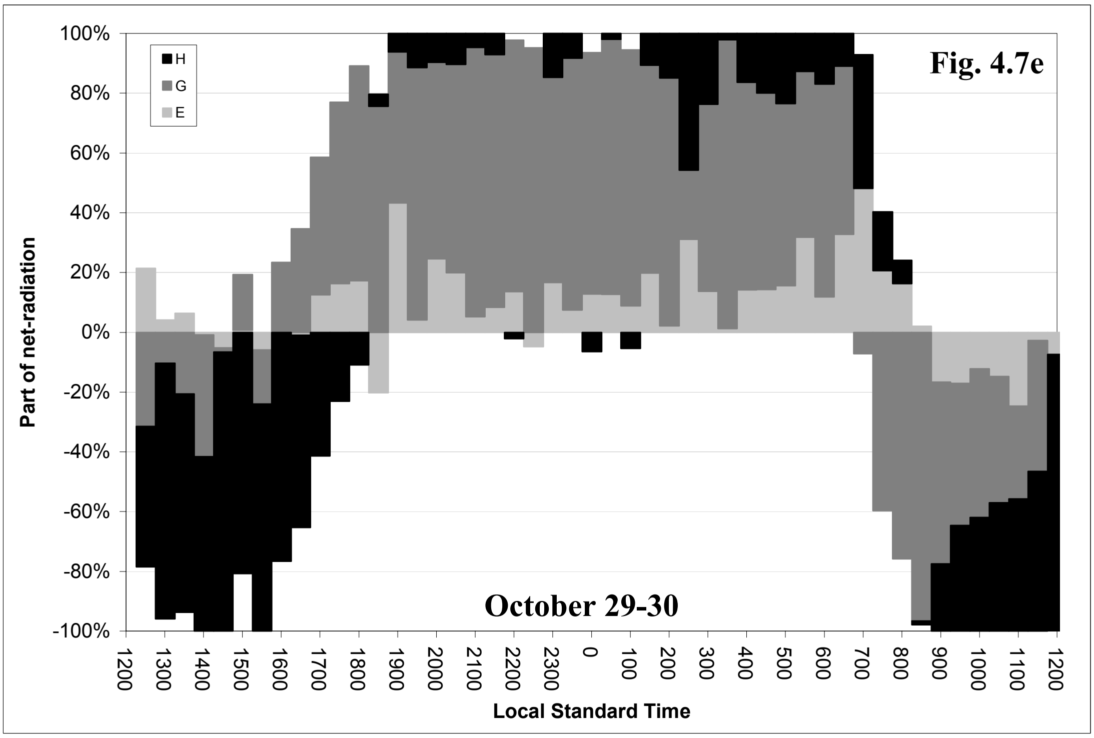


Figure 4.8*: The soil- (G), sensible- (H) and latent- (E) heat flux densities as fractions of the net-radiation (NR) (i.e. H+G+E=100%), as measured during the five 24-h campaigns. Note that the sensible heat flux density during June 16-17 was derived from the energy balance equation.*

4.5. Conclusions

Measurements carried out above a loess soil in the Negev desert, during the dry season indicate that the water content of the uppermost soil layer reach values which are considerably and systematically lower than the wilting point. Most of the commonly used meteorological models would, therefore, assume no latent heat flux. Nevertheless, latent heat flux densities were monitored throughout the dry season.

Latent heat flux density reaches ~20% of the net-radiation during the night, and thus plays a major role in the dissipation of the net radiation. However it should be kept in mind, that the overall flux densities at night are rather small. The 10-15% of the net-radiation during the day is even more significant as the magnitude of the flux densities is much larger. It is clear, thus, that meteorological models assuming no latent heat flux over deserts during the dry season may lead to erroneous results.

Chapter 5

Summary and Conclusions

In the frame of this research, the daily pattern of changes in the water content of the uppermost soil layers during the dry season was described, and the mechanisms by which water is added to the soil (dew deposition or direct adsorption) and their frequency were determined. The corresponding magnitude of latent heat flux and its effect on the energy partitioning were assessed in view of their importance for meso- and global-scale meteorological models. As these fluxes are very small, the development of special equipment was required, and the use of a micro-lysimeter was proposed.

The first step was to determine the dimensions of a micro-lysimeter that will confidently enable its use to detect dew. Measurements of the lateral temperature gradient inside the soil sample showed that a diameter ~20 cm is sufficient as the lateral heat flux was found to be negligible. A micro-lysimeter with a depth of 15 cm was initially constructed. The measured latent heat flux was compared to the amount of dew detected by a Hiltner dew balance and to the flux computed according to the energy balance approach. A good agreement between the micro-lysimeter and the Hiltner balance was found, while larger fluxes were computed via the energy balance approach. This result was surprising and, thus, the surface temperature of the soil sample was compared to the surrounding soil surface temperature as a means to evaluate its representativity. The micro-lysimeter was found to be cooler during the night with temperature differences (~2^{o}C) that were found to be large enough to affect the amounts of dew deposition. These differences can only be explained by the lack of deeper layers that contribute to the upward heat flux. In order to ascertain the depth required for a micro-lysimeter, the temperature profile in the undisturbed soil was monitored. The depth at which the changes in temperature are negligible on a diurnal cycle is 50 cm, and an improved micro-lysimeter, with a depth of 55 cm, was built. Its surface temperatures were compared to those of the surrounding soil. The temperature differences were significantly smaller than those measured with the previous micro-lysimeter. The measured latent heat flux was compared to the amount of dew detected by a Hiltner dew balance and to the flux computed according to the energy balance approach. The flux computed via the energy-balance approach was found to agree much better with the micro-lysimeter. The deeper micro-lysimeter clearly resolves the apparent lack of agreement between atmospheric and soil measured fluxes, as appeared to be the case from the results obtained with the 15 cm depth micro-lysimeter.

It was therefore concluded that **for measuring dew, the minimum depth of a micro-lysimeter should be the depth at which the diurnal temperature is constant in order to ensure similar temperature profiles inside and outside the micro-lysimeter**. This depth is probably site specific and depends on the thermal properties of the soil. **For a dry loess soil in the Negev Desert, a minimum depth of 50 cm is required**.

Having an accurate and reliable method for measuring very small latent heat flux at the soil-atmosphere interface, a study of the diurnal soil water content changes was carried out in order to assess the magnitude of these changes and to determine the mechanism responsible for the water gain during the night. Further, the relative magnitude of latent heat flux density over a bare loess soil in the Negev desert throughout the dry season was assessed.

Data were collected during nine 24-h field campaigns that took place during the dry season of 2002. Maximum incoming short-wave radiation varied from more than 950 W m^{-2} at the beginning of the measuring period to ~700 W m^{-2} towards the end. Throughout the season, a decrease in both the daytime and nighttime net-radiation fluxes is apparent. The same trend is evident in the difference between the maximum and the minimum temperatures of both the soil surface and the air. The daytime minimum relative humidity changed significantly during the season, even though daytime averages did not vary much. Maximum relative humidity reached ~95% in all nights. The variance of the average nighttime relative humidity was higher than the corresponding variance for daytime. It is worthwhile noticing that the average wind speed at all nights was rather high.

In order for dew to deposit on a surface, the temperature of the surface should be equal to, or less than, the dew-point temperature of the air mass with which it is in contact. The soil surface temperature usually did not drop below the estimated dew-point temperature. From these observations it was to be expected that no dew would deposit on the soil surface. Indeed, dew deposition on the soil surface was not visually observed in any of the campaigns. It was, therefore, concluded that in the area in which this study was carried out **actual dew deposition on a bare soil surface is a rare occurrence**. This does, however, not preclude that water vapor from the atmosphere may be directly adsorbed by the soil matrix. Water adsorption has been put forward as being an important link in the water cycle of arid and semi-arid regions. Indeed, a clear daily cycle of water content of the uppermost soil was observed. Throughout the season, the diurnal minimum water content was recorded at about 15:00 and the diurnal maximum to sunrise. The magnitude of the diurnal changes was very small and the average water contents throughout the entire measurement period were very low: a maximum of approximately 2% and a minimum which ranged from 1 to 1.5%.

The agreement between the amounts measured with the soil samples and the micro-lysimeter on the one hand and the clear underestimation of the Hiltner balance on the other hand are noteworthy. The average diurnal water gain was 0.26 and 0.25 mm for the soil samples and the micro-lysimeter respectively, with corresponding standard deviations (SD) of 0.05 and 0.04. The average gain detected by the Hiltner balance was much lower: 0.08 mm (SD=0.01). The fact that during the nights no signs of dew deposition could be observed on the soil surface, indicates that **the main process responsible for the observed diurnal change in water content is the direct adsorption of water vapor by the soil** rather than dew deposition.

The maximum daily water content of the uppermost 1 cm of the soil was found to be relatively constant and independent of the prevailing meteorological conditions. It was, therefore, concluded that the minimum water content that was reached during the preceding day determines the total amount of water vapor adsorbed. **A strong and significant correlation between the total adsorption of water vapor by the soil during the "absorption period" and the total potential evaporation between sunrise and sunset of the previous day was found**. Based on this finding an empirical model, based on commonly available data, is proposed in order to predict the total amount of water adsorbed by the soil during the absorption period. The proposed model is site specific, but very simple and easy to implement.

As no dew deposition was detected on the soil surface, but was detected by the Hiltner dew balance in the area in which this study was carried out, **artificial condensing plates should not be used to evaluate dew deposition quantities on a bare soil surface**. Moreover, even though the magnitude of the latent heat flux is larger than previously reported, the long-held belief that dew (as liquid water) is of prime importance in this type of ecosystems, needs to be re-assessed.

Lastly, it has been stated that land surface processes in general, and the energy partitioning at the soil surface in particular, play an important role in regional and meso-scale studies. These processes are usually integrated as sub-models of global and meso-scale models. One of their main tasks is to describe the patterns of radiant energy dissipation at the land-atmosphere interface, which above bare soil are determined by the moisture content of the soil surface. Many of the land surface models assume that when the soil moisture drops below the wilting point, there is no latent heat flux.

The results of this research indicate that the water content of the uppermost soil layer reaches values which are significantly and systematically lower than the wilting point. Most of the commonly used meteorological models would, therefore, assume no latent heat flux. During daytime a large fraction of the net-radiation was dissipated as sensible heat. However, the soil and **the latent heat flux densities are not negligible**. During nighttime, the soil heat flux density is the most dominant component of the energy balance, the sensible heat flux density is very small, and **in most campaigns the latent heat flux density is significantly larger than the sensible heat flux**. Being smaller than 1 during the night, the Bowen ratio supports these findings. **Latent heat flux density reached ~20% of the net-radiation during the night, and thus plays a major role in the dissipation of the net radiation.** However, overall flux densities at night are rather small. **The 10-15% of the net-radiation during the day is even more significant as the magnitude of the flux densities is much larger**. It is clear, thus, that models assuming that there is no latent heat flux over deserts during the dry season may lead to erroneous results.

The results of this study bring about new and important knowledge regarding the diurnal changes in soil water content and the energy partitioning over loess soil, under extremely dry conditions. This knowledge can improve the performance of meso- and global-scale meteorological models.

Chapter 6

References

Agam (Ninari) N. and Berliner P.R. (2004). Diurnal water content changes in the bare soil of a coastal desert. *Journal of Hydrometeorology*, 5 (5), 922-933.

Alnaser W.E. and Barakat A. (2000). Use of condensed water vapour from the atmosphere for irrigation in Bahrain. *Applied Energy*, 65, 3-18.

Anthoni P.M., Law B.E., Unsworth M.H. and Vong R.J. (2000). Variation of net radiation over heterogeneous surfaces: measurements and simulation in a juniper-sagebrush ecosystem. *Agricultural and Forest Meteorology*, 102, 275-286.

Atzema A.J., Jacobs A.F.G. and Wartena L. (1990). Moisture distribution within a maize crop due to dew. *Netherlands Journal of Agricultural Science*, 38, 117-129.

Auld B.A., McRae C.F. and Say M.M. (1988). Possible control of Xanthium spinosum by a fungus. *Agriculture Ecosystems and Environment*, 21(3-4), 219-223.

Avissar R. and Pielke R.A. (1989). A parameterization of heterogeneous land surfaces for atmospheric numerical models and its impact on regional meteorology. *Monthly Weather Review*, 117, 2113–2136.

Awanou C.N. and Hazoume R.P. (1997). Study of natural condensation of atmospheric humidity. *Renewable Energy*, 10, 19-34.

Berliner P.R. and Droppelmann K.J. (2003). Validation in an aid area of an algorithm for the estimation of daily solar radiation. *Journal of Hydrometeorology*, 4, 297-303.

Beysens D. (1995). The formation of dew. *Atmospheric Research*. 39(1-3), 215-237.

Beysens D., Muselli M., Ferrari J.P. and Junca A. (2001). Water production in an ancient sarcophagus at Arles-sur-Tech (France). *Atmospheric Research*, 57, 201-212.

Blondin C. (1991). Parameterization of land-surface processes in numerical weather prediction. In: Schmugge T.J. and Andre J.C. (eds.). *Land Surface Evaporation - Measurement and Parameterization*. Springer-Verlag, New-York, pp. 31-54.

Boast C.W. and Robertson T.M. (1982). A "micro-lysimeter" method for determining evaporation from bare-soil: description and laboratory evaluation. *Soil Science Society of America Journal*, 46, 689-696.

Bougeault P. (1991). In: Schmugge T.J. and Andre J.C. (eds.). *Land Surface Evaporation - Measurement and Parameterization*. Springer-Verlag, New-York, pp. 55-92.

Boulet G., Braud I. and Vauclin M. (1997). Study of the mechanisms of evaporation under arid conditions using a detailed model of the soil-atmosphere continuum. Application to the EFEDA I experiment. *Journal of Hydrology*, 193, 114-141.

Bourque C.P.A. and Arp P.A. (1994). Dawn-to-dusk evolution of air turbulence, temperature and sensible and latent heat fluxes above a forest canopy: concepts, model and field comparisons. *Atmosphere Ocean*, 32(2), 299-334.

Bristow K.L., Campbell G.S., Papendick R.I. and Elliott L.F. (1986). Simulation of heat and moisture transfer through a surface residue - soil system. *Agricultural and Forest Meteorology*, 36, 193-214.

Brutseart W. (1982). *Evaporation into the Atmosphere*. D. Reidel Publishing Company, Dordrecht.

Bunnenberg C. and Kuhn W. (1980). An electrical conductance method for determining condensation and evaporation processes in arid soils with high spatial resolution. *Soil Science*, 129, 58-66.

Butler D.R. (1980). Dew and thermal lag: measurements and an estimate of wetness duration on cocoa pods. *Quarterly Journal of the Royal Meteorological Society*, 106, 539-550.

Carlson T.N., Rose F.G. and Perry E.M. (1984). Regional-scale estimates of surface moisture availability from GEOS infrared satellite measurements. *Agronomy Journal*, 76, 972-979.

Castellani V. and Dragoni W. (1987). Some considerations regarding karstic evolution of desert limestone plateaus. In: *International geomorphology. Proceedings of 1st conference. Vol. 2*. Wiley, 1199-1206.

Chen F. and Dudhia J. (2001). Coupling an Advanced Land Surface–Hydrology Model with the Penn State–NCAR MM5 Modeling System. Part I: Model Implementation and Sensitivity. *Monthly Weather Review*, 129, 569-585.

Chiu-On N.G. (1999). Macroscopic equations for vapor transport in a multi-layered unsaturated zone. *Advances in Water Resources*, 22, 611-622.

Cleugh H.A. and Roberts T. (1994). Local-scale energy balances and microclimate in the desert ranges of central Australia. *Australian Meteorological Magazine*, 43, 219-228.

Danalatos N.G., Kosmas C., Moustakas N. and Yassoglou N. (1995). Rock fragments: II. Their effect on soil properties and biomass production. *Soil Use and Management Journal*, 11, 121-126.

De Vries D.A. (1958). Simultaneous transfer of heat and moisture in porous media. *Transactions, American Geophysical Union*, 39, 909-916.

Degen A.A., Leeper A. and Shachak M. (1992). The effect of slope direction and population-density on water influx in a desert snale, trochoidea-seetzenii. *Functional Ecology*, 6, 160-166.

Duvdevani S. (1947). An optical method of dew estimation. *Quarterly Journal of the Royal Meteorological Society*, 73, 282-296.

European Center for Medium-range Weather Forecast (2002). ECMWF.

Evenari M. (1985). Adaptations of plants and animals to the desert environment. In: *Hot deserts and arid shrublands, A*. Ecosystems of the world, 12A. Elsevier, pp. 79-92.

Evenari M. (1985). The desert environment. In: *Hot deserts and arid shrublands, A*. Ecosystems of the world, 12A. Elsevier. pp. 1-22.

Evenari M., Shanan L. and Tadmor N. (1971). *The Negev - The Challenge of a Desert*. Harward University Press, Cambridge.

Evett S.R., Warrick A.W. and Matthias A.D. (1995). Wall material and capping effects on microlysimeter temperatures and evaporation. *Soil Science Society of America Journal*, 59, 329-336.

Fraser A.B. (1994). The sylvanshine: retroreflection from dew-covered trees. *Applied Optics*, 33(21), 4539-4547.

Garratt J.R. and Segal M. (1988). On the contribution of atmospheric moisture to dew formation. *Boundary-Layer-Meteorology*, 45(3), 209-236.

Gilead M. and Rosenan N. (1954). Ten years of dew observation in Israel. *Israel Exploration Journal*, 4.

Goossens D. and Offer Z.Y. (1995). Comparisons of day-time and night-time dust accumulation in a desert region. *Journal of Arid Environments*, 31(3), 253-281.

Gusev Y.M. and Nasonova O.N. (2003). The simulation of heat and water exchange in the boreal spruce forest by the land-surface model SWAP. *Journal of Hydrology*, 280, 162-191.

Gutterman Y. and Shem-Tov S. (1997). Mucilaginous seed coat structure of Carrichtera annua and Anastatica hierochuntica from the Negev Desert highlands of Israel, and its adhesion to the soil crust. *Journal of Arid Environments*, 35(4), 695-705.

Hillel D. (1971) *Soil and Water*, Academic Press, New York.

Hillel D. (1998) *Environmental Soil Physics*, Academic Press, San Diego.

Hollermann P. and Zapp H. (1991). On the estimation and significance of condensation of dew in the eastern Canary Islands. *Erdwissenschaftliche-Forschung*. 25, 175-196.

Hu Z. and Islam S. (1997). Effects of spatial variabilities on the scaling of land surface parameterizations. *Boundary Layer Meteorology*, 83, 441–461.

Irannejad, P. and Shao, Y. (1998). Description and validation of the atmosphere-land-surface interaction scheme (ALSIS) with HAPEX and Cabauw data. *Global and Planetary Change*, 19, 87-114.

Jackson T.J. and Moy L. (1999). Dew Effects on Passive Microwave Observations of Land Surfaces. *Remote Sensing of Environment*, 70, 129-137.

Jacobs A.F.G., Heusinkveld B.G. and Berkowicz S.M. (1999). Dew deposition and drying in a desert system: a simple simulation model. *Journal of Arid Environments*, 42, 211-222.

Jacobs A.F.G., Heusinkveld B.G. and Berkowicz S.M. (2000a). Dew measurements along a longitudinal sand dune transect, Negev desert, Israel. *International Journal of Biometeorology*, 43, 184-190.

Jacobs A.F.G., Heusinkveld B.G. and Berkowicz S.M. (2000b). Force-restore technique for ground surface temperature and moisture content in a dry desert system. *Water Resources Research*, 36, 1261-1268.

Jacobs A.F.G., Van-Boxel J.H. and Nieveen J. (1996). Nighttime exchange processes near the soil surface of a maize canopy. *Agricultural and Forest Meteorology*, 82, 155-169.

Jacobs A.F.G., Van-Pul A. and El-Kilani R.M.M. (1994). Dew formation and the drying process within a maize canopy. *Boundary Layer Meteorology*, 69, 367-378.

Janssen L.H.J.M., Romer F.G. and Kema N.V. (1991). The freqeuncy and duration of dew occurrence over a year: model results compared with measurements. *Tellus, Series B*, 43B, 408-419.

Kidron G.J. (1998). A simple weighing method for dew and fog measurements. *Weather*, 53, 428-433.

Kidron G.J. (1999). Altitude dependent dew and fog in the Negev Desert, Israel. *Agricultural and Forest Meteorology*, 96, 1-8.

Kidron G.J. (2000). Analysis of dew precipitation in three habitats within a small arid drainage basin, Negev Highlands, Israel. *Atmospheric Research*, 55, 257-270.

Kidron G.J., Barzilay E. and Sachs E. (2000). Microclimate control upon sand microbiotic crusts, western Negev Desert, Israel. *Geomorphology*, 36, 1-18.

Kondo J., Saigusa N. and Sato T. (1990). A parameterization of evaporation from bare soil surfaces. *Journal of Applied Meteorology*, 29, 385-389.

Kosmas C., Danalatos N.G., Poesen J. and van Wesemael B. (1998). The effect of water vapour adsorption on soil moisture content under Mediterranean climatic conditions. *Agricultural Water Management*, 36, 157-168.

Kosmas C., Marathianou M., Gerontidis S., Detsis V., Tsara M. and Poesen J. (2001). Parameters affecting water vapor adsorption by the soil under semi-arid climatic conditions. *Agricultural Water Management*, 48, 61-78.

Lange O.L., Belnap J. and Reichenberger H. (1998). Photosynthesis of the cyanobacterial soil-crust lichen Collema tenax from arid lands in southern Utah, USA: Role of water content on light and temperature responses of CO_2 exchange. *Functional Ecology*, 12, 195-202.

Lange O.L., Kidron G.J., Budel B., Meyer A., Kilian E. and Abeliovich A. (1992). Taxonomic composition and photosynthetic characteristics of the 'biological soil crusts' covering sand dunes in the western Negev Desert. *Functional Ecology*, 6, 519-527.

Lascano R.J. and van Bavel C.H.M. (1986). Simulation and measurement of evaporation from a bare soil. *Soil Science Society of America Journal*, 50, 1127-1133.

Li X.Y. (2002). Effects of gravel and sand mulches on dew deposition in the semiarid region of China. *Journal of Hydrology*, 260, 151-160.

Liu H. and Foken T. (2001). A modified Bowen ratio method to determine sensible and latent heat fluxes. *Meteorologische Zeitschrift*, 10, 71-80.

Lohmann, D., Lettenmaier, D. P., Liang, X., Wood, E. F., Boone, A., Chang, S., Chen, F., Dai, Y., Desborough, C. and Dickinson, R. E. (1998). The Project for

Intercomparison of Land-surface Parameterization Schemes (PILPS) phase 2(c) Red-Arkansas River basin experiment: 3. Spatial and temporal analysis of water fluxes. *Global and Planetary Change*, 19, 161-179.

Lomas J. (1965). Note on dew duration recorders under semi-arid conditions. *Agricultural Meteorology*, 2, 351-359.

Ma Y., Su Z., Koike T., Yao T., Ishikawa H., Ueno K.I. and Menenti M. (2003). On measuring and remote sensing surface energy partitioning over the Tibetan Plateau--from GAME/Tibet to CAMP/Tibet. *Physics and Chemistry of the Earth*, Parts A/B/C, 28, 63-74.

Mabbayad M.O. and Watson A.K. (1995). Biological control of gooseweed (Sphenoclea zeylanica Gaertn.) with an Alternaria sp. *Crop Protection*, 14(5), 429-433.

Madeira A.C., Kim K.S., Taylor S.E. and Gleason M.L. (2002). A simple cloud-based energy balance model to estimate dew. *Agricultural and Forest Meteorology*, 111, 55-63.

Malek E. and Bingham G.E. (1997). Partitioning of radiation and energy balance components in an inhomogeneous desert valley. *Journal of Arid Environments*, 37, 193-207.

Marshall T.J., Holmes J.W. and Rose C.W. (1996). *Soil Physics*. Cambridge University Press, New York.

Menenti M., Bastiaanssen W.G.M. and Van-Eick D. (1989). Determination of surface hemispherical reflectance with Thematic Mapper data. *Remote Sensing of Environment*, 28, 327-337.

Milly P.C.D. (1982). Moisture and heat transport in hysteretic, inhomogeneous porous media: a matric-based formulation and a numerical model. *Water Resources Research*, 18, 489-498.

Milly P.C.D. (1984). A simulation analysis of thermal effects on evaporation from soil. *Water Resources Research*, 20, 1087-1098.

Minnis P., Mayor S., Smith W.L. and Young D.F. (1997). Asymmetry in the diurnal variation of surface albedo. *IEEE Transaction on Geoscience and Remote Sensing*, 35, 879-891.

Mitsuta, Y., Tamagawa, I., Sahashi, K. and Wang, J. (1995). Estimation of annual evaporation from the Linze desert during HEIFE. *Journal of the Metrological Society of Japan*, 73, 967-974.

Moffett M.W. (1985). An Indian ant's novel method for obtaining water. *National Geographic Research*, 1, 146-149.

Monteith J. L., Szeicz G. and Waggoner P.E. (1965). The measurement and control of stomatal resistance in the field. *Journal of Applied Ecology*, 2, 345-355.

Monteith J.L. (1957). Dew. *Quarterly Journal of the Royal Meteorological Society*, 83, 322-341.

Monteith J.L. (1965). Evaporation and environment. In: Fogg, G. E. (ed.). *The State and Movement of Water in Living Organisms*. Academic Press, New-York, pp. 205-234.

Monteith J.L. and Butler D.R. (1979). Dew and thermal lag - model for Cocoa pods. *Quarterly Journal of the Royal Meteorological Society*, 105, 207-215.

Moore W.J. (1963). *Physical Chemistry*. Longmans, London.

Morin L., Auld B.A. and Brown J.F. (1993). Host range of Puccinia xanthii and postpenetration development on Xanthium occidentale. *Canadian Journal of Botany*, 71(7), 959-965.

Neuma J. (1956). Estimating the amount of dewfall. *Arch. Meteorology*, 9, 197-203.

Nikolayev V.S., Beysens D., Gioda A., Milimouk I., Katiushin E. and Morel J.P. (1996). Water recovery from dew. *Journal of Hydrology*, 182, 19-35.

Ninari N. and Berliner P.R. (2002). The role of dew in the water and heat balance of bare loess soil in the Negev Desert: quantifying the actual dew deposition on the soil surface. *Atmospheric Research*, 64, 323-334.

Noffsinger T.L. (1965). Survey of techniques for measuring dew. In: Waxler (ed.). *Humidity and Moisture*, Vol. II. Reinhold, New-York, pp. 523-531.

Parlange M.B., Cahill A.T., Nielsen D.R., Hopmans J.W. and Wendroth O. (1998). Review of heat and water movement in field soils. *Soil and Tillage Research*, 47, 5-10.

Passerat S.A., Bruckler L., Thony J.L. and Vauclin M. (1989). Numerical modeling of coupled heat and water flows during drying in a stratified bare soil - comparison with field observations. *Journal of Hydrology*, 105, 109-138.

Paulson C.A. (1970). The mathematical representation of wind speed and temperature profiles in the unstable atmospheric surface layer. *Journal of Applied Meteorology*, 9, 857-861.

Pedro M.J. and Gillespie T.J. (1982a). Estimating dew duration. I. Utilizing micrometeorological data. *Agricultural Meteorology*, 25, 283-296.

Pedro M.J. and Gillespie T.J. (1982b). Estimating dew duration. II. Utilizing standard weather station data. Agricultural Meteorology, 25, 297-310.

Philip J.R. (1957). Evaporation, and moisture and heat fields in the soil. *Journal of Meteorology*, 14, 354-366.

Philip J.R. and de Vries D.A. (1957). Moisture movement in porous materials under temperature gradients. *Transactions, American Geophysical Union*, 38(2), 222-232.

Pinter-Jr P.J. (1986). Effect of dew on canopy reflectance and temperature. *Remote Sensing of Environment*, 19, 187-205.

Pitacco A., Gallinaro N. and Giulivo C. (1992). Evaluation of actual evapotranspiration of a Quercus ilex L. stand by the Bowen ratio-energy budget method. *Vegetatio*, 99-100, 163-168.

Plauborg F. (1995). Evaporation from bare soil in a temperate humid climate - measurement using micro-lysimeters and time domain reflectrometry. *Agricultural and Forest Meteorology*, 76, 1-17.

Qin Z., Berliner P.R. and Karnieli A. (2002). Numerical solution of a complete surface energy balance model for simulation of heat fluxes and surface temperature under bare soil environment. *Applied Mathematics and Computation*, 130, 171-200.

Ridley J., Strawbridge F., Card R. and Phillips H. (1996). Radar backscatter characteristics of a desert surface. *Remote Sensing of Environment*, 57, 63-78.

Robock A., Schlosser C.A., Vinnikov K.Y., Speranskaya N.A. and Entin J.K. (1998). Evaluation of AMIP soil moisture simulations. *Global Planet Change*, 19, 181–208.

Rosenberg N.J. (1969). Evaporation and condensation on bare soil under irrigation in the east central great planes. *Agronomy Journal*, 61, 557-561.

Scanlon B.R. and Milly P.C.D. (1994). Water and heat fluxes in desert soils. 2. Numerical simulations. *Water Resources Research*, 30, 721-733.

Scherm H. and Van Bruggen A.H.C. (1993). Sensitivity of simulated dew duration to meteorological variations in different climatic regions of California. *Agricultural and Forest Meteorology*, 66, 229-245.

Schmugge T.J. and Andre J.C. (eds.) (1991). *Land Surface Evaporation - Measurement and Parameterization*. Springer-Verlag, New-York.

Severini M., Moriconi M.L., Tonna G. and Olivieri B. (1984). Dewfall and evapotranspiration determination during day- and nighttime on an irrigated lawn. *Journal of Climate and Applied Meteorology*, 23, 1241-1246.

Shachak M., Leeper A. and Degen A.A. (2002). Effect of population density on water influx and distribution in the desert snail Trochoidea seetzenii. *Ecoscience*, 9, 287-292.

Shao, Y. and Henderson-Sellers, A. (1996). Validation of soil moisture simulation in landsurface parameterisation schemes with HAPEX data. *Global and Planetary Change*, 13, 11-46.

Shawcroft R.W. and Gardner H.R. (1983). Direct evaporation from soil under a row crop canopy. *Agricultural Meteorology*, 28, 229-238.

Slatyer R.O. (1967). *Plant - Water Relationships*. Academic Press, London.

Sridhar V., Elliott R.L. and Chen F. (2003). Scaling effects on modeled surface energy-balance components using the NOAH-OSU land surface model. *Journal of Hydrology*, 280, 105-123.

Stull R.B. (1988). *An Introduction to Boundary Layer Meteorology*. Kluwer Academic, New-York.

Subramanian A.R. and Kesava-Rao A.V.R. (1983). Dew fall in sand dune areas of India. *International Journal of Biometeorology*, 27, 271-280.

Sudmayer R.A., Nulsen R.A. and Scott W.D. (1994). Measured dewfall and potential condensation on grazed pasture in the Collie river basin, southwestern Australia. *Journal of Hydrology*, 154, 255-269.

Tanaka K., Ishikawa H., Hayashi T., Tamagawa I. and Ma Y.M. (2001). Surface energy budget at Amdo on the Tibetan Plateau using GAME/Tibet IOP98 data. *Journal of the Meteorological Society of Japan*, 79, 505-517.

Thomas M.D. (1928). Aqueous vapor pressure of soils: 4. influence of replaceable bases. *Soil Science*, 25, 485-493.

Thompson N. (1981). The duration of leaf wetness. *Meteorological Magazine*, 110, 1-12.

Turnipseed A.A., Blanken P.D., Anderson D.E. and Monson R.K. (2002). Energy budget above a high-elevation subalpine forest in complex topography. *Agricultural and Forest Meteorology*, 110, 177-201.

Unland H.E., Houser P.R., Shuttleworth W.J. and Zong-L. Y. (1996). Surface flux measurement and modeling at a semi-arid Sonoran Desert site. *Agricultural and Forest Meteorology*, 82, 119-153.

Van de Griend A.A. and Owe M. (1994). Bare soil surface resistance to evaporation by vapor diffusion under semiarid conditions. *Water Resources Research*, 30, 181-188.

Van Genuchten M.T. (1980). A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. *Soil Science Society of America Journal*, 44, 892-898.

Waggoner J.E., Begg J.E. and Turner N.C. (1969). Evaporation of dew. *Agricultural Meteorology*, 6, 227-230.

Walker G.K. (1983). Measurement of evaporation from soil beneath crop canopies. *Canadian Journal of Soil Science*, 63, 137-141.

Wallin J.R. (1967). Agrometeorological aspects of dew. *Agricultural Meteorology*, 4, 85-102.

Went F.W. (1955). Fog, mist dew and other sources of water. In: *Yearbook of Agriculture*. US Department of Agriculture, Washington, D.C. pp 103-109.

Wigneron J.P., Calvet J.C. and Kerr Y. (1996). Monitoring water interception by crop fields from passive microwave observations. *Agricultural and Forest Meteorology*, 80, 177-194.

Willis A.J. (1985). Dune water and nutrient regimes - their ecological relevance. In: Sand dunes and their management. Report of a meeting held in Swansea, 1984: *Nature Conservancy Council; Focus on Nature Conservation*, pp 159-174.

Wilske B., Burgheimer J., Karnieli A., Zaady E., Yakir D. and Kesselmeier J. (Submitted). The CO_2 exchange of Biological Soil Crusts in a semiarid shrub-grassland at the northern transition zone of the Negev desert, Israel. *Soil Biology and Biochemistry*.

Wilson T.B., Bland W.L. and Norman J.M. (1999). Measurement and simulation of dew accumulation and drying in a potato canopy. *Agricultural and Forest Meteorology*, 93, 111-119.

Wood D., McNairn H., Brown R.J. and Dixon R. (2002). The effect of dew on the use of RADARSAT-1 for crop monitoring; Choosing between ascending and descending orbits. *Remote Sensing of Environment*, 80, 241-247.

Yakirevich A., Berliner P.R. and Sorek S. (1997). A model for numerical simulating of evaporation from bare saline soil. *Water Resources Research*, 33, 1021-1033.

Yang, Z.-L. and Dickinson, R. E. (1996). Description of the Biosphere-Atmosphere Transfer Scheme (BATS) for the Soil Moisture Workshop and evaluation of its performance. *Global and Planetary Change*, 13, 117-134.

Yang, Z.-L., Dickinson, R. E., Shuttleworth, W. J. and Shaikh, M. (1998). Treatment of soil, vegetation and snow in land surface models: a test of the Biosphere-Atmosphere Transfer Scheme with the HAPEX-MOBILHY, ABRACOS and Russian data. *Journal of Hydrology*, 212-213, 109-127.

Zangvil A. (1996). Six years of dew observations in the Negev Desert, Israel. *Journal of Arid Environments*, 32, 361-371.

Zangvil A. and Druian P. (1980). Measurements of dew at a desert site in southern Israel. *Geographical Research Forum*, 2, 26-34.

Zentay T., Gerei L. and Balogh J. (1985). An investigation of some water budget properties of sand soils on the Danube-Tisza interfluve. *Foldrajzi Ertesito*, 34, 123-132.

Zhang W. and Watson A.K. (1997). Host range of Exserohilum monoceras, a potential bioherbicide for the control of Echinochloa species. *Canadian Journal of Botany*, 75(5), 685-692.

Zuberer D.A. and Kenerley C.M. (1993). Seasonal dynamics of bacterial colonization of cotton fiber and effects of moisture on growth of bacteria within the cotton boll. *Applied and Environmental Microbiology*, 59, 974-980.