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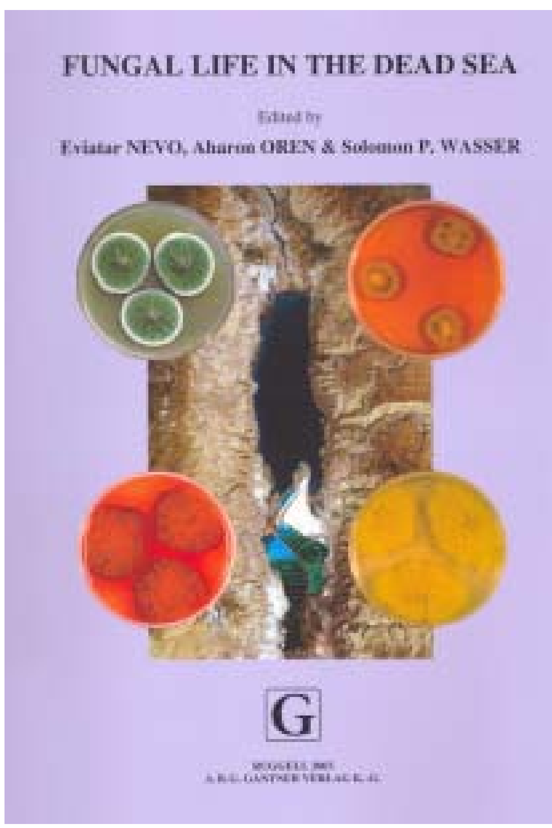

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ABSTRACT

Ten years of quasi-continuous meteorological data and subsurface temperatures, acquired from a buoy, anchored on the Dead Sea, are presented and analyzed. The analysis reveals and quantifies diurnal and annual periodic variations as well as interannual changes.

Insolation variations are not significantly different from that in other parts of the world for our latitude.

Air pressures are about 50 mb larger than those at the Mediterranean sea level. The amplitudes of the air pressure diurnal variation, the atmospheric tides, are significantly larger than those expected for this latitude, and also significantly larger than those at the Mediterranean sea level. They change seasonally and are related to the changes in the ambient air temperatures.

Air temperatures and humidities are affected by the proximity of the water, and, therefore, the temperatures are slightly lower and the humidities slightly higher than those observed at nearby land stations (e.g., Sdom). The air temperature reaches a peak in the afternoon, then it starts to decline but, shortly afterward, it starts to increase again and reaches a second peak, higher than the first one, late in the evening. The humidity reaches a minimum in the afternoon, starts to increase, and then recedes again, reaching a lesser second minimum late in the evening. The behavior of the temperature and the humidity are directly related to the arrival of the dry and hot Mediterranean sea breeze.

Sea surface temperatures have a typical bimodal distribution determined by insolation and by the rate of mixing in the water column. The interaction between the sea surface temperature and air temperature is explained.

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The wind field is also strongly affected by the Mediterranean sea breeze. Diurnally the strongest winds occur at midnight and the weakest in the early afternoon. Seasonally, the weakest winds were observed during the winter. Strong winds, i.e., larger than 10 m s^{-1}, consist of less than 1% of the observations. They are far more prevalent in the winter than in the summer and can induce very high and destructive waves.

The Dead Sea is “drying up”. Its level is declining and its salinity is increasing. The microclimate above it reacts to these changes, and during the ten years we collected the data, we observed significant changes in some of the measured parameters. Thus the atmospheric pressure increased in direct proportion to the decrease in the Dead Sea level. The insolation diminished, apparently due to an increase in aerosol content in the air. Some of these aerosols stem from the sea itself and the increase in their number are due to the higher salinity of the surface layers of the sea. Air temperatures and relative humidities did not change significantly, but the sea surface temperature increased.

1. Introduction

The Dead Sea is a unique environment endowed with very unusual properties, described in some detail in other papers in the present volume. Its meteorology and hydrography are also unusual, and, in essence, the following is a description of the Dead Sea meteorology, intended as a background to the other investigations presented in this volume. This description is based on a large database acquired by the authors from 1992 to 2002. The present investigation differs from previous investigations in that the entire meteorological database was acquired at sea and accompanied by simultaneous subsurface measurements of temperatures.

One must stress from the very beginning that these data were acquired at one single point on the sea, and not in the middle of the sea (Figure 1). Thus one cannot claim that the climatology computed from those measurements is representative of the entire sea. Nevertheless, this is the best data we have and one could argue that it is a more representative climatology than that derived from shore stations.

Since 1977 the descending surface water level of the Dead Sea has led to the separation between the southern and the northern basins of the Dead Sea (Steinhorn, 1981a). In the present paper, the Dead Sea will refer solely to what is today the northern basin. The paper is organized according to air and water properties (e.g., radiation, winds, air temperatures, water temperatures, etc.). In order to avoid unnecessary repetitions, the description of previous investigations in the following paragraph is, as far as possible, succinct and not exhaustive. Additional details on previous investigations will be presented and discussed in relevant sections of each topic.

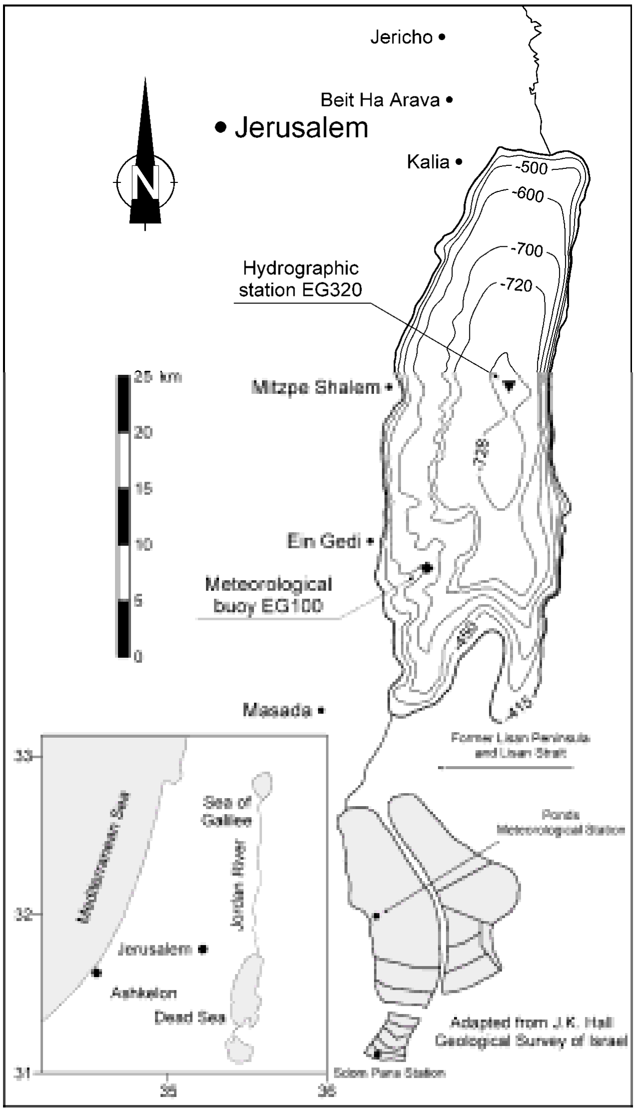

Figure 1. Location map and station positions.

Previous Investigations

Systematic meteorological investigation of the Dead Sea region was initiated in 1929 by the pioneering efforts of Ashbel which he continued until 1974 (Ashbel, 1975). Ashbel's data were acquired at land-based meteorological stations situated in the vicinity of the Dead Sea. During the entire period of Ashbel's investigations, the Dead Sea level was high enough to maintain the connection between the two basins. Still, Ashbel (1954, 1966, 1975) finds a significant difference between the northern and the southern parts of the sea. For instance, he showed that the wind regime in the southern basin was significantly different. There was less rain there, and the waters there were warmer than those in the northern basin (Ashbel, 1939). Ashbel ascribes the differences to orography and to the shallowness of the southern basin.

Bitan (1974, 1977, 1984) described the wind regime around the Dead Sea. His data were acquired later than Ashbel's data but, apparently, most of them also preceded the separation between the northern and the southern basins. Bitan (1977, 1984) also observed significant differences between the northern and the southern parts of the region. As far as we could understand, Bitan's database did not include any measurements at sea. Stanhill (1987, 1994) provides a description of the radiation climate over and around the Dead Sea computed from land-based measurements, adjusted to the sea environment via sporadic measurements made during occasional cruises on the Dead Sea. Gat and Karni (1995), in their description of the climate of the Jordan Valley, include a description of the northern shores of the Dead Sea. Their investigation is based on three land-based stations. Goldreich (1998), in a comprehensive investigation of the climate of Israel, also provides a fairly detailed description of the Dead Sea region. His data stem from Ashbel's work as well as the database of the Israel Meteorological Service, thus again land-based measurements.

In the marine environment Ashbel was, once again, one of the pioneers of field measurements. Ashbel (1975) provides the data of long series systematic measurements of the sea level, sea surface and subsurface temperatures, specific gravity of the water, and evaporation rates. Neev and Emery (1967) followed, and although hydrography was not their main objective, they acquired and presented a large amount of new hydrographic information. However, their investigation was limited by national boundaries to only a part of the Dead Sea. Steinhorn (1981a) carried out a very comprehensive investigation of the hydrography of the Dead Sea via a large number of data acquisition cruises. Many later papers on the hydrography of the Dead Sea are based on the data collected by and the innovations of Steinhorn (e.g., Steinhorn and Assaf, 1980; Steinhorn, 1981b; Steinhorn, 1983). Another series of extensive hydrographic cruises were carried out by Anati and resulted in a crop of important papers (e.g., Anati and Shasha, 1989; Anati, 1993; Anati, 1997; Anati, 1999). This series is still being continued under the supervision of Gertman (see Gertman and Hecht, 2002). Stanhill (1990) investigated the heat balance changes in the Dead Sea upper layers from temperature measurements obtained at sea during a series of cruises spanning four years.

The data obtained during cruises are very important and the effort required to sustain a series of cruises on the Dead Sea is not negligible. This data contributed enormously to understanding the environment of the Dead Sea. Nevertheless cruises suffer from an implicit deficiency and bias: they have to be carried out in relatively good weather, thus there are no representative data on "bad weather" conditions. This also pertains to the meteorological data acquired during the cruises, and in particular the data intended to "calibrate" land-based meteorological stations. Lack of data during bad weather conditions biases our investigation of the physics of the Dead Sea and dynamic processes there. For instance, it is difficult to ascertain to what degree the "calibrated" data of land-based meteorological stations are really representative of the conditions at sea. Understanding meteorological processes above the Dead Sea, and in particular understanding the interaction between the sea and the atmosphere above it, requires intensive meteorological measurements at sea, detailed comparisons with land-based measurements, as well as simultaneous hydrological measurements. Such measurements have to be carried out from buoys. On the other hand, instrument breakdowns interrupt the continuity of the buoy measurements and even if the data from the buoy is closely monitored and the breakdown is discovered almost immediately, it takes time to go out and repair or replace the instrument. Therefore data gaps are almost inevitable. Furthermore, in the long run, buoys are far more expensive than cruises, and financial considerations prevent us from deploying a sufficient number of buoys to obtain appropriate space continuity. Thus buoy measurements do not replace cruises. Rather, the two methods have to be used to complement each other.

A first attempt to obtain quasi-continuous meteorological data at sea was carried out by the IOLR between 1984 and 1988. The data acquired by this buoy were analyzed by a group of scientists from the Hebrew University (Cohen et al., 1983, 1986; Mahrer et al., 1984; Weiss et al., 1987). Unfortunately, as described by Weiss (1988), the measurements were beset by difficulties. The buoy often parted from its anchor and drifted about, and some of the buoys disappeared without a trace. Moreover, some of the sensors failed for extended periods. Therefore, the database lacked uniformity and continuity. Thus, although the reports presented and discussed some short period processes (e.g., diurnal processes), there is no attempt to present a climatological summary based on the buoy data. In particular, Weiss (1988) discussed the data acquired on the buoy during one year only, 1987 - 1988, presumably the best of the series, and even for that year there are extended gaps in the database. Therefore, Weiss (1988) presents more of a description of his findings rather than a statistical analysis of the data.

2. Data Acquisition

In 1990 we renewed our attempts to acquire continuous meteorological and hydrographic data on the Dead Sea. The first few attempts were not successful. The buoys lost their moorings and drifted away and some were lost. These experiences led us to the planning and construction of a new buoy, which was anchored in 100 m of water about 4 km east of Ein Gedi (Figure 1). The new buoy (Figure 2) was large enough to support heavier moorings as well as significant salt precipitation on the cable and on the buoy body. However, the buoy's capability to float the extra weight was not limitless, and in order to ensure its survival we had to replace the anchoring tackle and clean it thoroughly at relatively short intervals (e.g., two months). Even so, sometimes, we arrived at the buoy just in the nick of time (Figure 3). On the 19th of February 2002, the transmissions from the buoy ceased. We found out that it had turned over from the weight of the salt deposited on it and on its mooring cable. Thus, this buoy had survived at sea and acquired data for more than 9 years (17 June 1992 to 19 February 2002) and had provided the largest in situ quasi-continuous database on the meteorology and hydrography of the Dead Sea. We intend to renew the operation of the buoy in the near future.

The new buoy supported a standard meteorological screen in which we had transducers for the measurement of air temperature, air pressure, and relative humidity at 3 m above the sea surface. Wind speed and direction as well as total incoming radiation were measured on a mast at 3.7 m above the sea surface. Current speed and direction were measured at 1 m below the sea surface, and water temperatures were measured at 1, 2, 3, 4, 6, 8, 10, 15, 20, 30 and 40 m below the sea surface. The buoy orientation was also recor-

Figure 2. Meteorological buoy on the Dead Sea.

Figure 3. Meteorological buoy covered with salt and on the verge of sinking.

ded in order to compute the accurate absolute wind direction. The exact position of the buoy was monitored by GPS. Measurements were obtained at 20-minute intervals - hence quasi-continuous. All the measurements, with the exception of the wind speeds, were instantaneous. The measured wind speeds were arithmetic averages over the measuring interval (i.e., 20 minutes). In order to maintain time continuity, we did not change from summer- to winter-time and all times referred to in our investigation are Israel Summer Time, i.e., GMT-3. The data were stored on board the buoy on storage units, but they were also transmitted to a computer on shore in real time. The data stored in the computer were accessible at any time via cellular telephone. Aanderaa - a Norwegian manufacturer, produced all measuring instruments. The technical staff of the IOLR produced the hardware and the software necessary for the shoreward data transmission, communication with the computer, and communication with the Haifa laboratory.

Potentially we could have collected 254471 measurements per parameter, which including the data necessary for monitoring the operation of the buoy would have produced a database of some 6 million measurements. However, we had breakdowns as well as initiated interruptions, the later due to the necessity to clean and repair the buoy and its instruments. The availability of the data in quasi-real time enabled us to find out immediately when an instrument ceased to work and sometimes we could advance the next cruise so as to minimize the "down time" of the instrument, or at least come on the cruise well prepared to change the faulty instrument, thus the interruptions in the series were minimized. Nevertheless, we have gaps in the data. Figure 4 depicts the available database and the gaps in it. The marked gaps do not necessarily indicate that all the instruments failed. Sometimes there were partial failures only, e.g., a failure in one or two or three of the temperature measuring thermistors. Faults and problems with particular instruments will be indicated and discussed in the paragraph relating to the parameter.

More information on the buoy and its operation can be found on the internet at www.ocean.org.il. As long as the buoy was operational, a representative sample of the data was presented in quasi real time on the internet at the same site.

From April 1995 to December 1996, we also used Aanderaa instruments to collect meteorological data at Ashkelon, situated close to the Mediterranean seashore (Figure 1). This station was at a height of about 10 m above sea level.

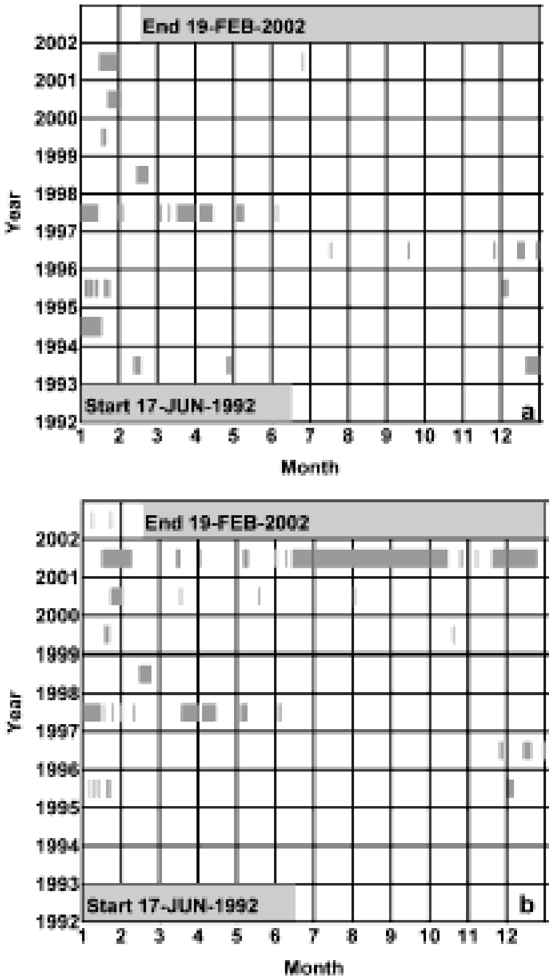

Figure 4. a- Distribution of meteorological data gaps,
b- Distribution of thermistor chain data gaps.

The data analysis was carried out on the original data. However, for certain purposes the data were averaged. The reasons and methods of averaging will be defined and explained below so as not to have to repeat the explanation in every one of the following sections.

Hourly averages: Since all the measurements were instantaneous (with the exception of wind speed), spurious jumps could mar the data: therefore, for some of our investigations the data were averaged over a period of one hour.

Diurnal oscillation detection: In order to identify time periodicity in the data, data for a given time of the day were averaged, irrespective of the date.

Diurnal averages: In order to smooth over any diurnal oscillations the data were averaged over a 24-hour period. However, if some data were missing during the 24 hours in question, that particular day was considered biased, and that particular data was not used in our computations. Radiation measurements were not averaged but summed up, however, the same criteria applied to them, i.e., if during a particular day any data were missing, that day was considered biased and that particular data was not used in our computations.

3. Solar Radiation

Incident solar radiation on the top of the atmosphere is depleted by absorption and scattering on its path to the surface of the Earth. Some of the scattered radiation also reaches the surface of the Earth. Global radiation, or insolation, consists of the direct incoming solar radiation together with the scattered radiation reaching the surface of the Earth. The insolation depends upon the length of the path that the solar radiation has to pass on its way to the Earth as well as upon the composition of the atmosphere, i.e., gas content (such as CO_2 and SO_2), water vapor, clouds, solid particles, etc. Hence, the insolation is a function of the angle of the sun above the horizon, and, therefore, the insolation varies periodically, diurnally as well as annually. These regular cycles are deformed by random variations introduced by changes in the composition of the atmosphere as well as by the amount of cloud coverage above the measuring site. The insolation reaching the Dead Sea is more depleted than that reaching the mean sea level due to the additional 400 m layer of air above it. Ashbel (1966) discusses this effect in detail but does not quantify it.

Solar radiation was measured by pyranometer, in the range of 0.3 to 2.5 micron, with an accuracy of ±20 W m^{-2}. The dome of the instrument was cleaned periodically. Nevertheless, some dust and dirt accumulated on the dome of the instrument between cleanings and probably increased the error. Moreover, sometimes, the instruments failed completely so that we acquired 242964 valid data, from which we computed 3335 diurnal insolation values, thus a failure rate of about 5.7%.

Diurnal variations

Following the variation in the sun's angle above the horizon and the changes in the length of the day, the diurnal insolation over the Dead Sea changes as well. Figure 5 depicts the average diurnal variation in the insolation as it changes from month to month. The instantaneous peak radiation is always at noon (about 12:30), but the amplitude varies from month to month and it is the smallest in December (471 W m^{-2}) and the largest in June (964 W m^{-2}). From June to January the insolation is diminishing but the traces are very similar. In order not to clutter the figure, only the six months of the year, from the minimum to the maximum are presented. The shortest day (in Figure 5 the distance between the two points where the graph reaches 0) also occurs in December (about 10 hours) and the longest in June (about 12 hours).

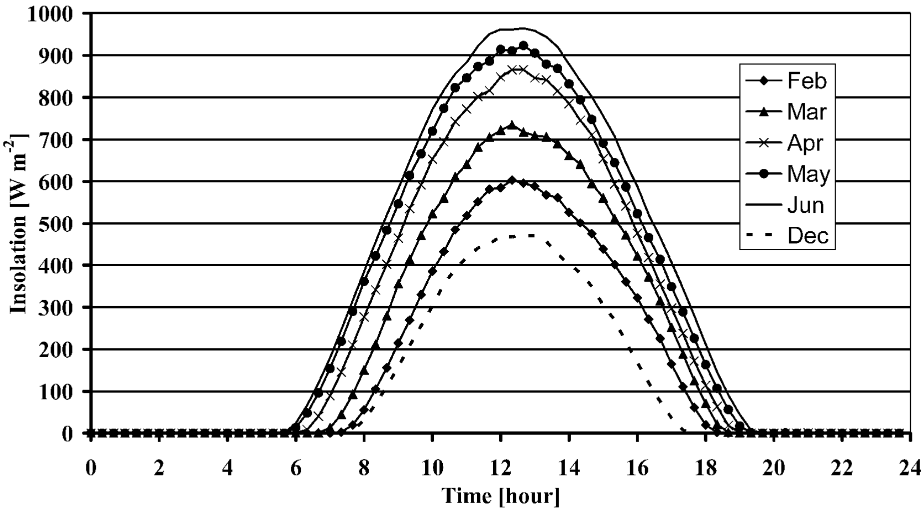

Figure 5. Seasonal changes in the diurnal variation of insolation over the Dead Sea.

Annual variations

The annual variation of the insolation, plotted from the total diurnal insolation (Figure 6), shows random fluctuations which are far more intense during the "winter" part of the year - November to June and less intense during the "summer" part of the year - June to November. These random fluctuations are, of course, induced by passing pressure systems and the accompanying cloud cover and/or aerosols in the air. The integrated smalllest average diurnal insolation occurs in December (about 10 MJ m^{-2} per month), and the larger average in June (about 27.5 MJ m^{-2} per month). The magnitude of the average insolation as measured by us appears to compare well with the values published in literature (e.g., Stanhill, 1987; Goldreich, 1998).

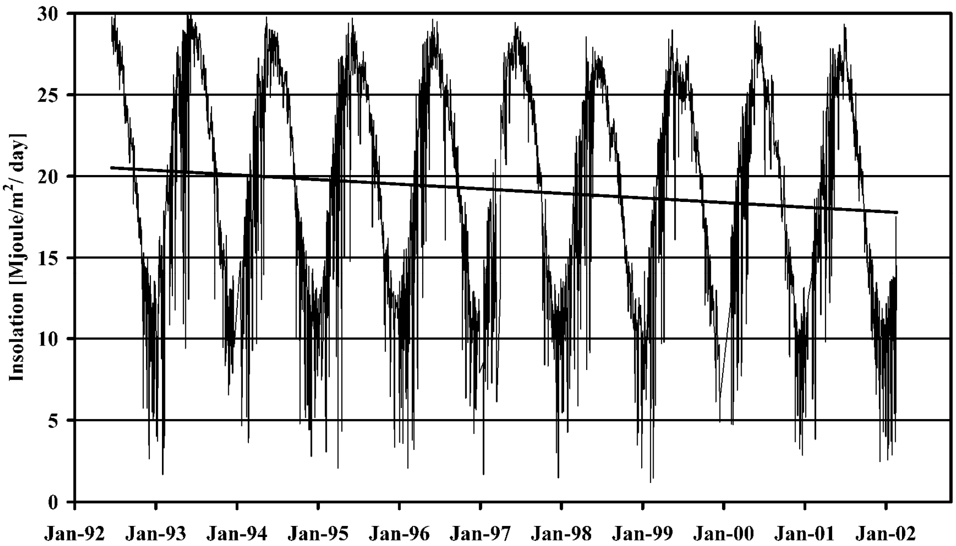

Figure 6. Interannual changes in the insolation over the Dead Sea, 1992-2002.

Interannual variations.

The data presented in Figure 6 seems to indicate a declining trend. The linear slope of the trend is - 0.28 MJ m^{-2} day^{-1} per year and a *t* test indicated that it is significant to better than 99.9%. Due to this trend, by the end of the period of our measurements, the insolation was reduced by about 2.7 MJ m^{-2} day^{-1}. Cohen et al. (1983) and Mahrer et al. (1984) also noticed that the insolation over the Dead Sea has diminished since Ashbel's (1964) report. They relate this to the Dead Sea level decrease and the subsequent increase in local sea aerosols, but they qualify this suggestion since they did not have sufficient data for a statistical analysis. Cohen and Stanhill (1996), who also found a large decrease in the global irradiance along the Jordan valley, suggest that this is due to the increase in anthropogenic aerosols in the atmosphere over the entire region. A three-year investigation of the dust particles over the Dead Sea indicated a significant increase in the dust deposition over the sea (Singer et al., 2003), related perhaps to the aridification of the region (e.g., Alpert et al., 1997). Throughout the period of our measurements, the Dead Sea level has descended by about 7 m and the salinity of the surface waters has increased (e.g., Gertman and Hecht, 2002). Therefore, one could probably assume that the local aerosol density (salt particles resulting from spray) as well as dust particles increased and that this is the main reason for the diminished insolation.

Comparison of Measurements at Mediterranean Sea Level.

We attempted to quantify the additional radiation depletion above the Dead Sea due to it's being 400 m below the Mediterranean Sea level. Thus we compared the measurements on the Dead Sea with those measured by us at Ashkelon. The results (Figure 7) show that there is a high degree of visual correlation between the measurements at the two sites. Thus one can see that significantly reduced insolation occurs simultaneously at both sites on a number of occasions. These ought to indicate cloud cover over the entire region, while a significant reduction of insolation at Ashkelon accompanied by a lesser reduction over the Dead Sea would indicate a more extensive cloud cover at Ashkelon than over the Dead Sea. However one does not see a significant difference between the two sites. In fact, an average of the differences in the insolation at the two sites indicates that the insolation at

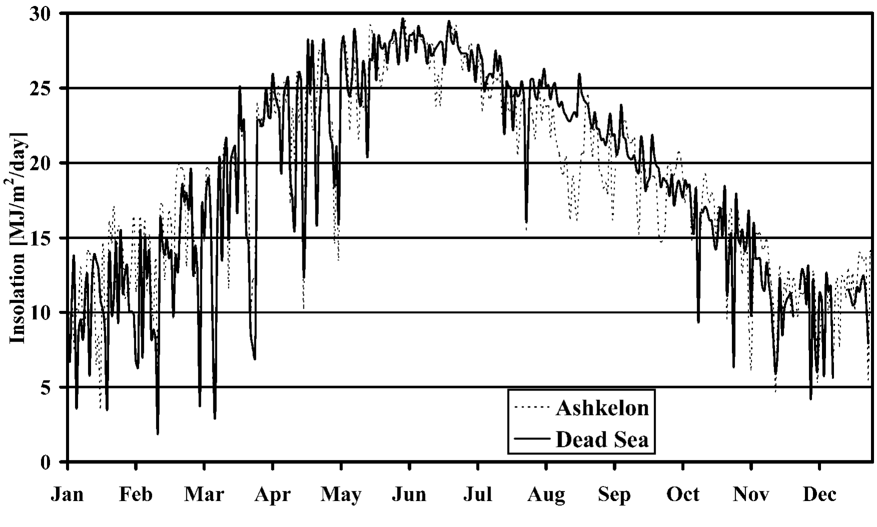

Figure 7. Insolation at Ashkelon versus that at the Dead Sea.

the Dead Sea is slightly larger than that at Ashkelon. This puzzling result can be ascribed to the differences in cloudiness and atmospheric composition between the two sites. Obviously a single year of measurements is insufficient to quantify the effect.

4. Air Pressure

The air pressures over the Dead Sea follow an annual cycle with higher pressures during the winter and lower ones during the summer (Ashbel 1954, 1966). From the end of September to the end of June, pressure systems (cyclones and anticyclones) crossing the region induce large local random fluctuations which could be of the order of 13 mb, or even as high as 20 mb. Pressure fluctuations diminish during the summer period, and then the dominant factor affecting the pressure changes is the semidiurnal variation. Note that due to the additional 400 m air layer above the Dead Sea, the air pressures there are about 50 mb larger than those at Mediterranean mean sea level.

We acquired 245700 valid measurements, thus a failure rate of about 3.5%. The failures were not entirely random. They occurred in blocks of data, often during the winter at the beginning and at the end of the year (see Figure 4). By and large, our measurements confirm Ashbel's description, however, we will attempt to quantify his statements statistically.

<u>Diurnal variations -Atmospheric tides.</u>

Semidiurnal pressure variations, known as atmospheric tides result from the combination between the gravitational tidal forces, the diurnal temperature variation, and the resonance of the atmosphere (e.g., Chapman, 1951; Giles, 1987; Glickman, 2000). In contrast to the sea tide, whose extrema shift by about 50 minutes every 24 hours, the peaks of the atmospheric tide always occur close to 10:00 (the highest peak) and 22:00, while the troughs occur close to 04:00 (the deepest trough) and 16:00 local time (Giles, 1987). The magnitude of the amplitude of the atmospheric tide changes from about 3.5 mb at the equator and diminish toward the poles where it is about 0.3 mb (Giles, 1987).

The computed pressure diurnal oscillations show that the atmospheric tides over the Dead Sea (Figure 8) follow the pattern described above. On the average the highest peak (1061.4 mb) occurs at 09:40; the lowest trough (1057.4 mb) occurs at 17:20; the next peak (1059.8 mb) occurs at midnight, and the next trough (1059.7 mb) occurs at 03:00. The time of the peaks and the troughs differ from the classical theory, however, what is particularly conspicuous is that the maximal amplitude, 4 mb, is significantly larger than the one indicated by Giles (1987). An inspection of our database indicated that neither the times of the occurrence of the extrema nor the size of the amplitudes vary significantly from year to year.

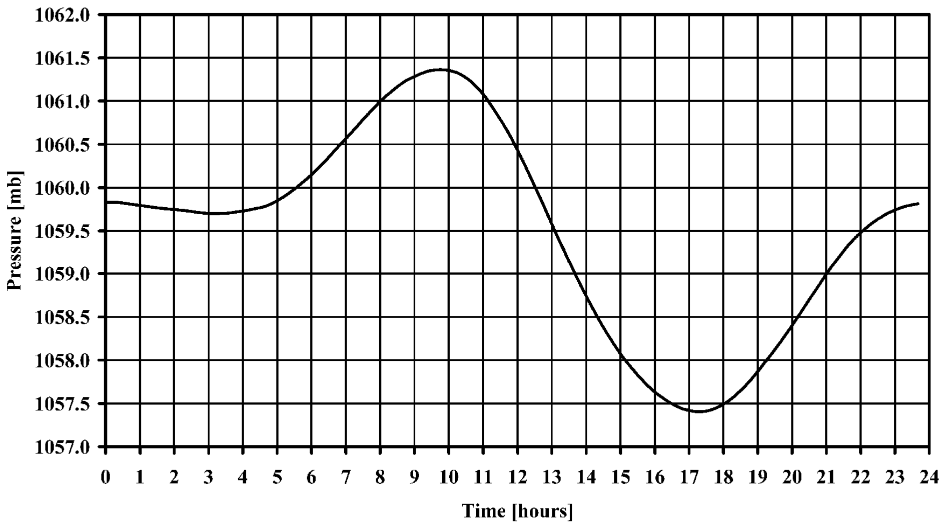

Figure 8. Annual average atmospheric tides over the Dead Sea, 1992-2002.

It is interesting to compare the atmospheric tide over the Dead Sea to that at Mediterranean mean sea level. Data at the Mediterranean sea level were available for 1996 at two sites: Ashkelon and Hadera (Figure 9). The peaks of the atmospheric tides at the three sites occur at exactly the same time (i.e., 10:20), the troughs occur within 40 minutes of each other (at about 17:20). However, the maximal amplitudes for the Dead Sea, 3.9 mb, versus those at Ashkelon and Hadera, 1.52 mb and 1.35 mb, respectively, differ significantly. Furthermore, the maximal amplitude of the atmospheric tides is different from that indicated by Giles (1987), who, for our latitude, gives a value of 2.5 mb. On the other hand, Reiter (1975) is less specific about magnitudes but he points out that the amplitude is larger inland than over the sea, that it is related to the temperature, and that there are seasonal changes in the amplitude of the atmospheric tides. The atmospheric tides over the Dead Sea are indeed changing with the seasons (Figure 10). The entire monthly trace shifts from January to July (from July to December the traces recede monotonously and were not presented here in order not to clutter the figure). The amplitude of the atmospheric tides as well as the times of the occurrence of the extrema also change seasonally (Table 1).

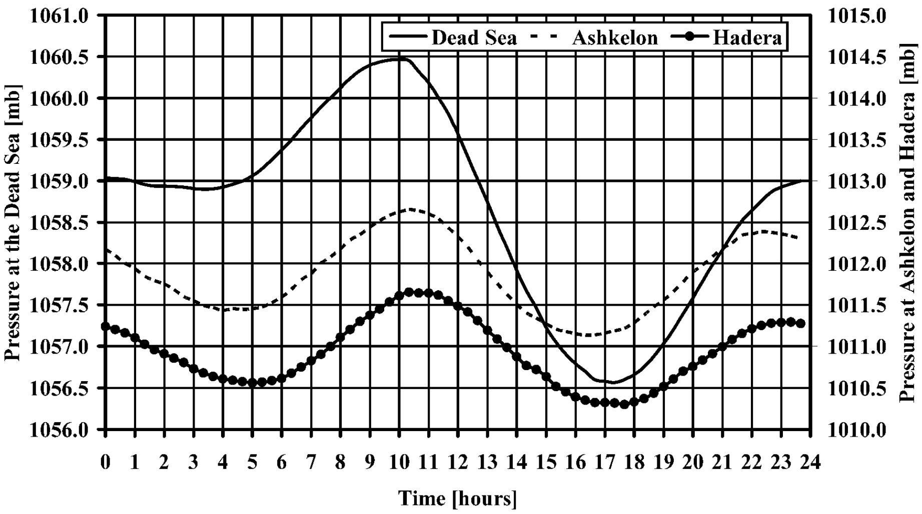

Figure 9. Atmospheric tides at the Dead Sea, Ashkelon and Hadera in 1996.

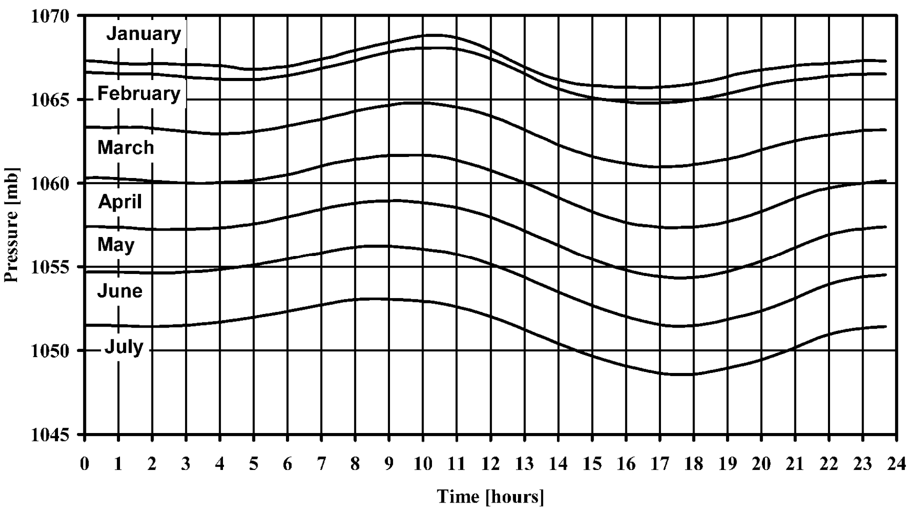

Figure 10. Seasonal changes in average atmospheric tides.

Table 1. Seasonal Changes in the Extrema of Atmospheric Tides (monthly averages)

	Pressure mb.						
	Avg.	STD	Time Max	Max	Time Min	Min	Amplitude
Jan	1067.1	4.7	10:20	1068.8	16:40	1065.7	3.1
Feb	1066.3	4.8	10:40	1068.1	16:40	1064.8	3.3
Mar	1063.0	5.1	09:40	1064.7	17:00	1061.0	3.7
Apr	1059.8	4.9	10:00	1061.7	17:20	1057.4	4.3
May	1056.9	3.7	09:20	1058.9	17:40	1054.4	4.6
Jun	1054.2	3.2	08:40	1056.2	17:40	1051.5	4.7
Jul	1051.1	2.7	08:40	1053.1	17:40	1048.6	4.5
Aug	1052.4	2.6	9:40	1054.4	17:40	1049.9	4.5
Sep	1056.0	2.9	9:40	1057.9	17:20	1053.6	4.3
Oct	1060.2	3.0	9:40	1062.1	17:00	1058.1	4.0
Nov	1063.7	3.7	10:00	1065.5	16:00	1062.1	3.4
Dec	1067.1	4.0	10:00	1068.8	15:40	1065.7	3.1

Thus, the smallest amplitudes (3.1 mb) were observed in January and December, the largest amplitude (4.7 mb) was observed in June. From June onward the amplitude diminishes. One can't help but relate this to the decrease in the average pressure from January to July and then the increase from August to December. Furthermore, the times of the peaks shift toward an earlier hour from winter to summer, from 10:00 in November and December to 8:40 in June and July. The time of the troughs also shift from 15:40 in December to 17:40 in June and July. The seasonal changes in the size of the amplitude of the monthly averaged temperature diurnal cycle (Figure 11b) coincide with the seasonal changes in the monthly averaged atmospheric diurnal tidal cycle (Figure11a). Furthermore, the seasonal changes are in phase, i.e., the time of the temperature maxima coincide with the times of the atmospheric tide minima and vice versa, the times of the temperature minima coincide with the times of the atmospheric tide maxima (Figure 11b vs. Figure 11a), both occurring later in the winter than in the summer. Thus, there is an apparent relation between the atmospheric tide and the air temperature as well as the length of the day (both related to the angle of incident radiation).

Annual variations.

As indicated above (Table 1), pressures are the largest during the winter (1067.1 mb, in December and January) and the smallest during the summer (1051.1 mb in July). Furthermore, as an indicator of the random pressure fluctuation induced by crossing pressure systems, we computed the standard deviations for each month (Table 1). On the average, we found that the largest fluctuations occur in March while the smallest fluctuations occur in August. Although the largest average standard deviation is of the order of 5 mb, however, single much larger pressure changes were encountered.

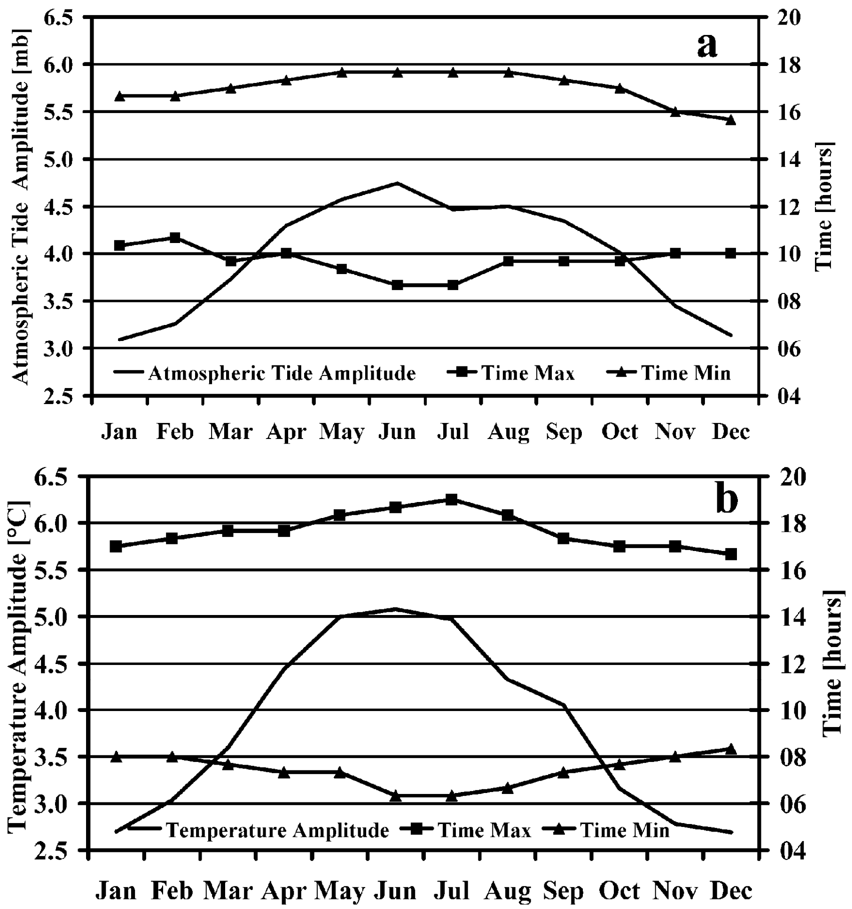

Figure 11. Seasonal changes in the amplitudes and times of occurrence of the minima and maxima of the diurnal cycles of atmospheric tides and air temperatures.

Interannual variations.

Diurnal averages for the entire period of our measurements (Figure 12) show a slight trend toward higher pressures. A *t* test indicates that the slope coefficient (0.11 mb per year) is significant at 99%. Thus, over the period of our measurements, the pressures have increased by 1.1 mb. Over this period, the Dead Sea level had dropped by 7 meters, and our air pressure gauge mounted on the buoy also dropped. Due to the change in the sea level, the air pressure would have been expected to increase by 0.9 mb, which seemed to explain the observed trend.

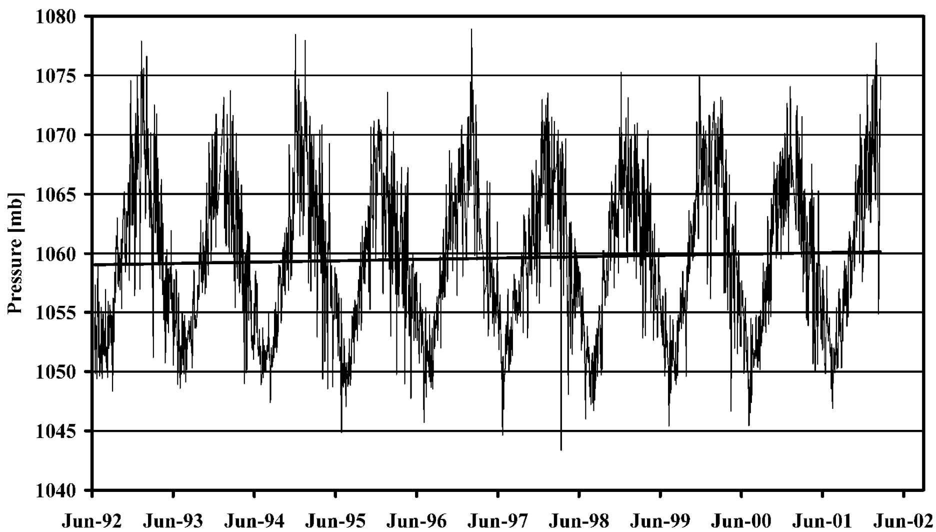

Figure 12. Interannual air pressure changes, 1992-2002.

5. Air Temperature

Air temperatures are directly related to the intensity of the insolation: hence, they change periodically throughout the day and throughout the year. Average annual air temperature in the Dead Sea region (based on measurements at Sdom) is the highest in Israel and exceeds the average temperature expected for the Dead Sea latitude (Goldreich, 1998). On our buoy, a platinum resistance thermometer measured the air temperature, and we acquired 245702 valid measurements, thus a 3.5% failure rate.

Diurnal variations.

In general, during a cloudless day, minimal air temperature will be observed early in the morning, before sunrise, and it will reach a maximum about an hour or two after the peak of the insolation at noon (e.g., Goldreich, 1998). The Dead Sea presents us with a different picture. Diurnal variation of air temperature averaged for the entire period of our measurements (Figure 13a) shows that the minimum (24.1°C) occurs as late as 07:40, while the maximum (27.8 °C) occurs at 17:20. However the diurnal oscillation extrema change seasonally as well (Figure 13b). Thus the minima can occur as late as 08:00 in the winter (November, December, January, and February) or as early as 6:20 in the summer (June, July, August), while the maxima could occur as early as 16:20 in the winter (December and January) or as late as 19:00 in the summer (July). The amplitude of the diurnal oscillations also changes with the seasons (Table 2). It is about 2.7 °C in the winter (e.g., January) and reaches a maximum of 5.1°C in the summer (e.g., June).

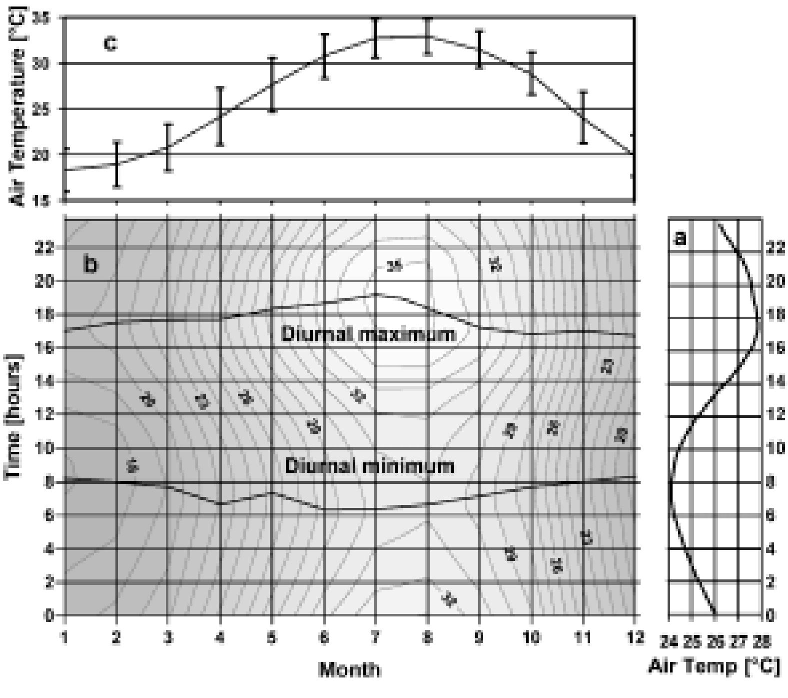

Figure 13. Air temperature variations over the Dead Sea. **a.** Annual averaged diurnal oscillations. **b.** Seasonal changes in the diurnal oscillations. **c.** Annual averaged changes in the air temperature.

The reason for this is that the amplitude is related to the intensity of the insolation and to the length of the day.

Table 2. Seasonal Changes in the Extrema of the Temperature Diurnal Oscillations.

	Minimum		Maximum		Amplitude
	time	°C	time	°C	
JAN	08:00	16.9	17:00	19.6	2.7
FEB	08:00	17.4	17:20	20.4	3.1
MAR	07:40	19.1	17:40	22.8	3.7
APR	06:40	22.1	17:40	26.6	4.5
MAY	07:20	25.4	18:20	30.4	5.0
JUN	06:20	28.4	18:40	33.6	5.1
JUL	06:20	30.4	19:00	35.4	5.0
AUG	06:40	30.9	18:20	35.2	4.4
SEP	07:20	29.7	17:20	33.7	4.1
OCT	07:40	27.3	17:00	30.6	3.2
NOV	08:00	22.5	17:00	25.4	2.8
DEC	08:20	18.4	16:40	21.2	2.8
average	07:40	24.1	17:20	27.8	3.7

A deeper insight into the air temperature patterns over the Dead Sea could be gained by comparing the average annual diurnal cycle over the Dead Sea with that on the Mediterranean seashore, at Ashkelon, both measured in 1996 (Figure 14). It is obvious from this figure that the minimum at Ashkelon preceded the minimum over the Dead Sea by one hour, and the maximum at Ashkelon preceded the maximum over the Dead Sea by almost two hours. At both sites, the times of the maxima and the minima change with the seasons, however the extrema at Ashkelon always precede the equivalent ones at the Dead Sea, albeit the time difference between the corresponding extrema is somewhat shorter in the winter and somewhat longer in the summer.

To understand the delay in the minima over the Dead Sea one must bear in mind that the air temperatures at Ashkelon were measured over ground, while the air temperatures over the Dead Sea were measured over water. Ground tends to warm up as well as lose heat faster than water, and, as the air is warmed up by the ground or the water and not directly by insolation, there is an obvious delay. Furthermore, the additional blanket of 400 m of air above the Dead Sea depletes the insolation (although we did not succeed in showing it in the section on insolation) as well as slows down the cooling. Ashbel (1954) as well as Goldreich (1998) discuss these effects.

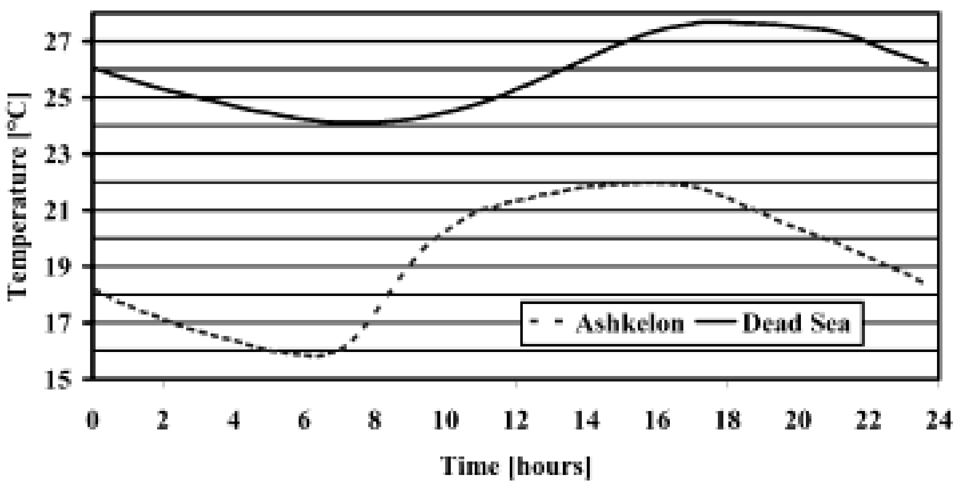

Figure 14. Air temperature diurnal oscillations Dead Sea versus Ashkelon 1996.

The air temperature maximum over the Dead Sea is also “late” relative to the maximum at Ashkelon (Figure 14). However, this is not just a shift in the phase of the oscillation, since it is delayed by two hours versus the delay of one hour for the minimum.

There certainly must be an air temperature afternoon maximum related to the peak radiation at noon and perhaps delayed by factors described above for the early morning delay in the minimum. However, there seems to be an additional factor that contributes to the rise in the air temperature above and beyond the peak due to the insolation. This additional factor is the Mediterranean sea breeze (e.g., Ashbel, 1966; Bitan, 1977; Goldreich, 1998). According to the Doron and Neumann (1977) model, the Mediterranean sea breeze reaches speeds of about 5 m s^{-1} at about 14:00 on the coast, and, when reinforced by the anabatic effect (“valley breeze”) it can reach a maximum speed of about 7 m s^{-1}. Indeed, in June 1996, at Ashkelon, we observed speeds of about 5 m s^{-1} at 14:00 (according to Goldreich, 1998, during June one encounters the largest sea to ground air temperature differences, thus the maximum sea breeze). For a speed of 7 m s^{-1}, the breeze reaches the Dead Sea, 100 km from Ashkelon, about 4 hours later, i.e., at about 18:00. During its travel the air climbs up the Judean Hills 800 m, cooling down adiabatically by about 8 °C (since it is summer the process is assumed to be adiabatically unsaturated), then descends down to the Dead Sea, 1200 m, warming up adiabatically dry by about 12 °C. Thus the air reaches the Dead Sea about 4 °C warmer then it started. During June 1996 we observed, at Ashkelon, maximal average air temperatures of about 25 °C, and, therefore, we would expect the air to reach the Dead Sea with a temperature of about 30 °C. The heated air reaches the Dead Sea in the afternoon, just in time to continue the heating due to the insolation and push the air temperature to a later and higher peak. More often than not, the two processes overlap and one cannot distinguish between the two, certainly not when averaging the data. If one is aware of the process, one can discern a first air temperature maximum, followed by a slight decline in the temperature, followed by a second, later and more intense maximum (e.g., Figure 15). The second maximum does indeed occur around 18:00. During the winter (e.g., Figure 15 March 98) it is about 24.5 °C, about 1.5 °C higher then the local solar maximum. During the summer (e.g., Figure 15, June 92) it is as high as 40 °C, about 3 °C higher than the local solar maximum. Bitan (1977) maintains that the Mediterranean sea breeze is active throughout the year, albeit weaker in the winter than in the summer. This is also indicated by the delay in the afternoon air temperature maximum, i.e., the afternoon maximum in the winter is not as late as in the summer, but still later than expected for the effects of the insolation alone, as well as in the lesser difference between the two maxima (e.g., Figure 15, 23 March 1998 versus 10 June 1992).

Goldreich (1998, p. 119) describes the diurnal air temperature variation near the Dead Sea as having a main maximum in the late afternoon and a minor maximum in the morning. The first is ascribed to the arrival of the Mediterranean sea breeze, while the second is ascribed to the local Dead Sea breeze. Associated with them are two minima, the main one early in the morning, the minor one at noon. He stresses that the minor extrema can be observed only on continuous recordings. We found such minor extrema in our records of the air temperature on the shore at Ein Gedi (although the temperatures there were recorded every 20 minutes), but we do not find them in our records at sea, since the local sea breeze cannot be observed at sea (minor maxima observed very early in the morning in Figure 14 are far too small to be considered as more than local fluctuations).

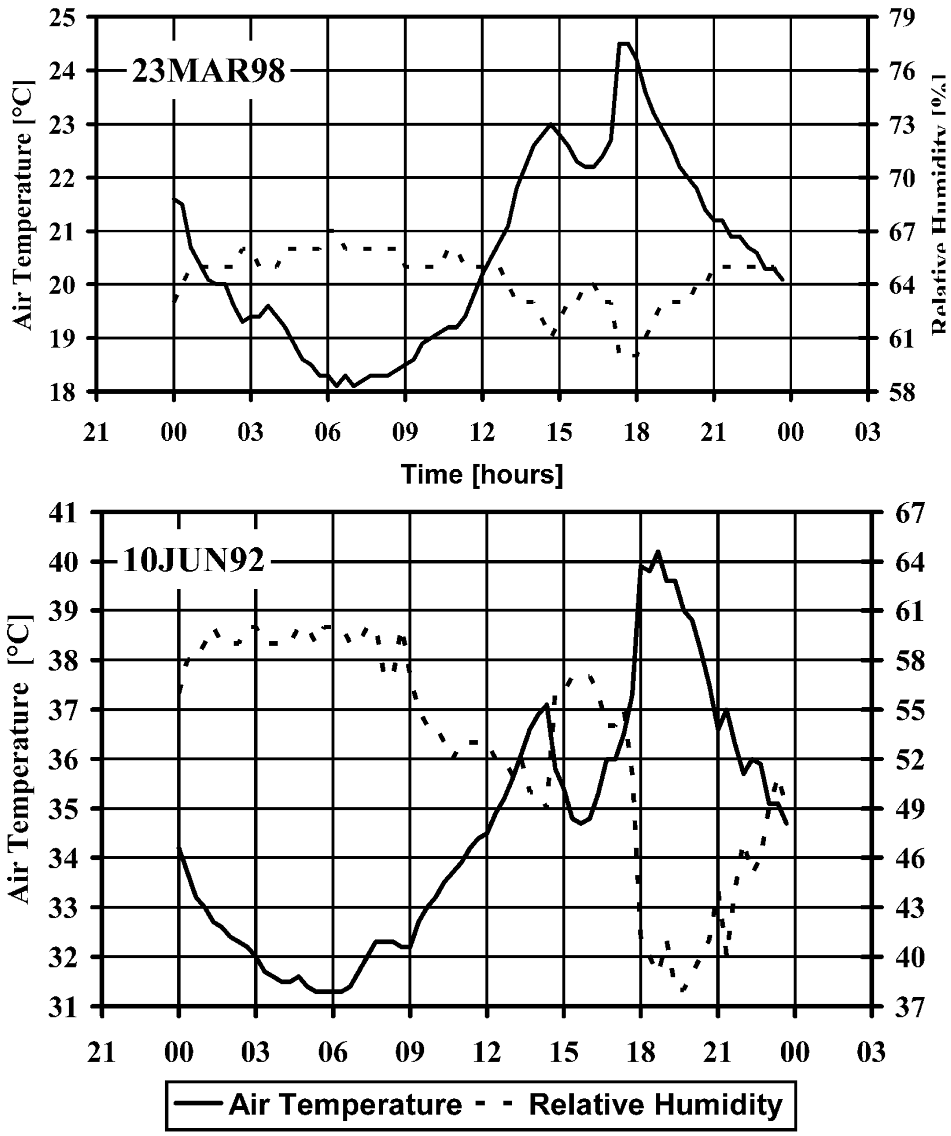

Figure 15. Examples of diurnal measurements of temperature and humidity.

Annual and interannual variations.

Average annual changes in air temperature (Figure 13c) are as expected, with low air temperatures during the winter and high temperatures during the summer. It was mentioned at the beginning of the present section that

Table 3. Monthly Averages of Air Temperature, Relative Humidity, and Sea Surface Temperature.

	AIR TEMPERATURE					RELATIVE HUMIDITY					SEA SURF. TEMP.			
Month	**Sdom***	**Aver**	**STD**	**Max**	**Min**	**Sdom***	**Aver**	**STD**	**Max**	**Min**	**Aver**	**STD**	**Max**	**Min**
JAN	15.7	18.3	2.4	24.9	7.5	51	65.9	3.6	83	44	21.6	2.3	24.4	14.0
FEB	17.5	18.9	2.4	26.6	9.2	48	65.3	3.8	81	44	21.5	2.2	24.6	14.5
MAR	20.7	20.8	2.6	31.0	14.4	44	64.4	3.2	73	42	22.3	1.3	28.8	18.3
APR	24.6	24.2	3.1	39.8	15.8	40	62.5	3.6	72	40	24.5	1.3	29.2	21.6
MAY	28.7	27.7	2.9	39.8	20.7	37	61.4	3.5	79	37	27.4	1.4	33.1	24.0
JUN	32	30.8	2.4	42.4	24.6	37	60.2	3.7	68	39	30.9	1.2	34.9	27.9
JUL	33.5	32.7	2.1	43.4	26.6	37	59.8	3.5	68	34	33.1	0.8	35.9	30.6
AUG	33.4	32.9	1.9	41.8	28.0	38	60.6	3.0	68	46	33.7	0.7	35.9	31.1
SEP	31.5	31.5	2.0	39.8	24.6	41	61.0	2.7	67	42	32.5	1.1	35.8	28.8
OCT	27.8	28.9	2.3	37.5	21.3	43	61.7	2.9	72	46	30.0	1.2	33.8	26.6
NOV	22.4	24.0	2.8	33.9	14.2	47	62.4	3.7	79	45	25.8	1.8	30.8	20.2
DEC	17.3	19.9	2.3	26.4	11.3	50	64.9	2.9	73	52	22.6	1.7	26.3	14.8
year	25.4	25.9				43	62.5				27.2			

*Average monthly air temperature and relative humidity at Sdom from Goldreich (1998)

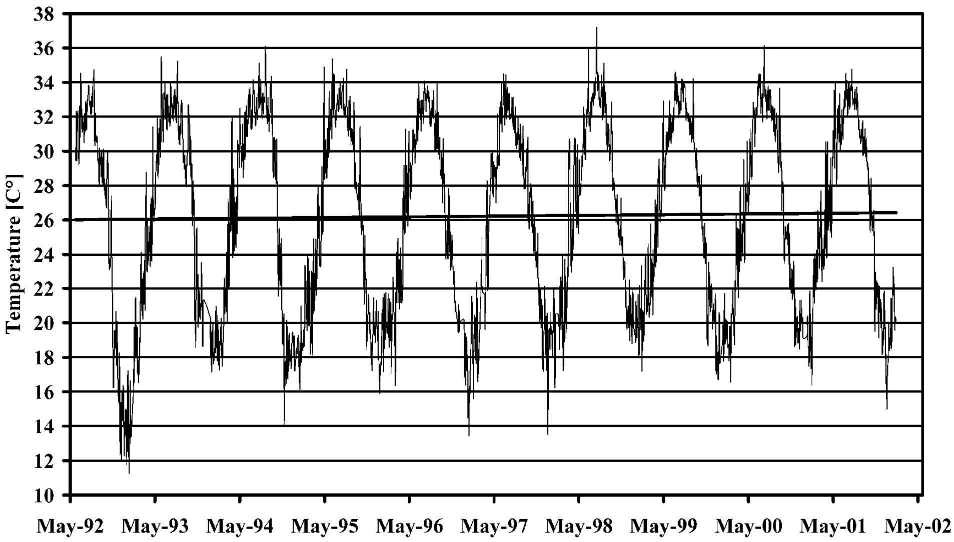

Figure 16. Interannual air temperature changes over the Dead Sea, 1992-2002.

according to Goldreich (1998) the average annual air temperature at Sdom are the highest in Israel (25.4 °C). It appears that the average annual temperature over the Dead Sea is even higher, 25.9 °C (Table 3). However, if one compares the average monthly temperatures over the Dead Sea with those at Sdom (Table 3), one finds that during the winter months (January to March, and October to December) the temperatures over the Sea are higher than those at Sdom, while during the summer months (April to August) the temperatures over the Sea are lower than those at Sdom. This effect is typical of the moderating influence of the Dead Sea.

Interannual changes, plotted from diurnal averages (Figure 16) appear to indicate a slight upward trend, however, a *t* test indicates that the slope of this trend is not significant.

6. Relative Humidity

The relative humidity in the air is strongly and inversely related to the air temperature: as the temperature decreases, the relative humidity increases; and as the temperature increases, the relative humidity decreases. Thus, the relative humidity goes through both a diurnal cycle as well as an annual cycle. On the Dead Sea these cycles have some peculiar properties that will be described below. On our buoy the humidity was measured instantaneously by a hygroscopic element. We acquired 245700 measurements, thus a failure rate of about 3.5%.

<u>Diurnal variations.</u>

On the Dead Sea shores, and in particular during the summer, the relative humidity reaches a maximum at noon and descends to a minimum in the late afternoon close to sunset (Ashbel, 1966; Goldreich, 1998). The maximum is ascribed to the local sea breeze, while the minimum is ascribed to the hot dry air descending from the Judean Mountains (the Mediterranean sea breeze, e.g., Ashbel, 1966; Bitan, 1977; Goldreich, 1998).

The annual average relative humidity diurnal oscillation observed by us (Figure 17a) indicates a maximum (64.2%) at about 08:00. There are two minima: one of them (60.9%) at about 17:00, the other one (60.8%) at about 20:40. However, the extrema change seasonally (Figure 17b) and for part of the year we have observed two maxima (in June to August) and two minima (in April to November). For these periods one finds an early maximum at about 06:00 and a late maximum at about 08:30. Furthermore, one finds an early minimum at about 16:00 and a late minimum at about 20:30.

As we can see, the pattern observed on the buoy differs from that described by Ashbel (1966) or by Goldreich (1998). We associate the morning maximum to the early morning low air temperatures. However during part of the year the terrestrial breeze, blowing southward, brings in drier air from the land (e.g., Gat and Karni, 1995) and the relative humidity decreases for a short while. The first afternoon diurnal minimum is associated with the peak in local insolation, while the second afternoon diurnal minimum is associated with the dry air brought in by the Mediterranean breeze, both processes having been described in the previous paragraph. The relationship between air temperature variations and relative humidity at various hours of the day is well demonstrated in Figure 15, where we can clearly see the decrease in humidity related to an increase in temperature and vice versa.

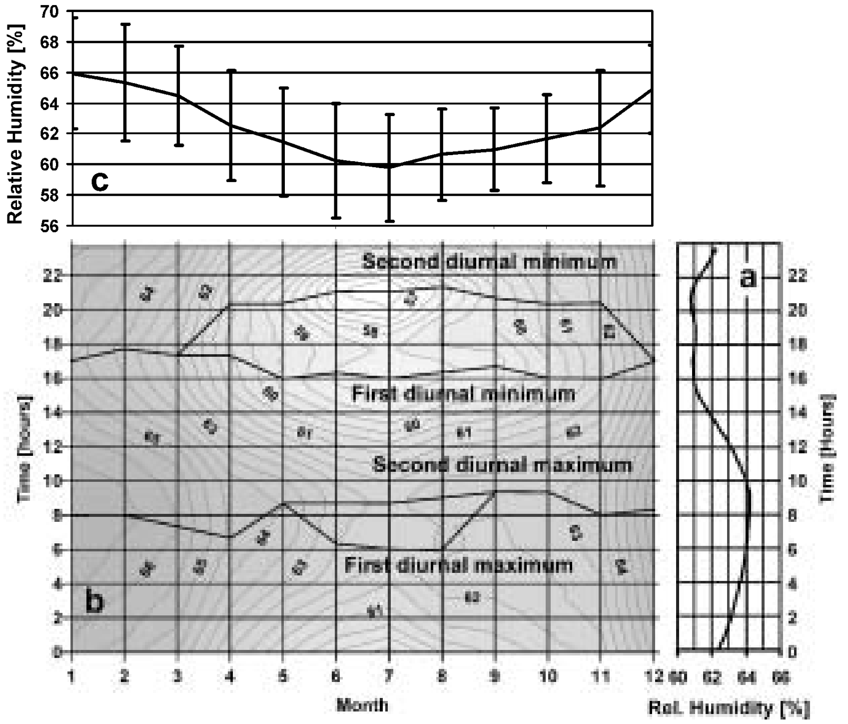

Figure 17. Relative humidity variations over the Dead Sea. **a.** Annual averaged diurnal oscillations. **b.** Seasonal changes in the diurnal oscillations. **c.** Annual averaged changes in the air temperature.

Annual and interannual changes.

As expected, winter months are more humid than summer months (Figure 17c). Monthly averages (Table 3) show that December has the highest relative humidity (64.9%), July has the lowest relative humidity (59.8%) and the average annual humidity is 62.5%. Compared with the monthly data at Sdom (in Table 3, data from Goldreich, 1998) or the relative humidity at Kalia and Mitzpe Shalem (Gat and Karni, 1995), the relative humidity over the sea is significantly larger. This is obviously expected since relative humidities at a land-based station are expected to be lower than those at marine based stations. Our measurements show that the relative humidity spreads between 35 to 85% (Figure 18). However, these extremes are very infrequent, and 98% of the measured relative humidity data are between 50 and 70%.

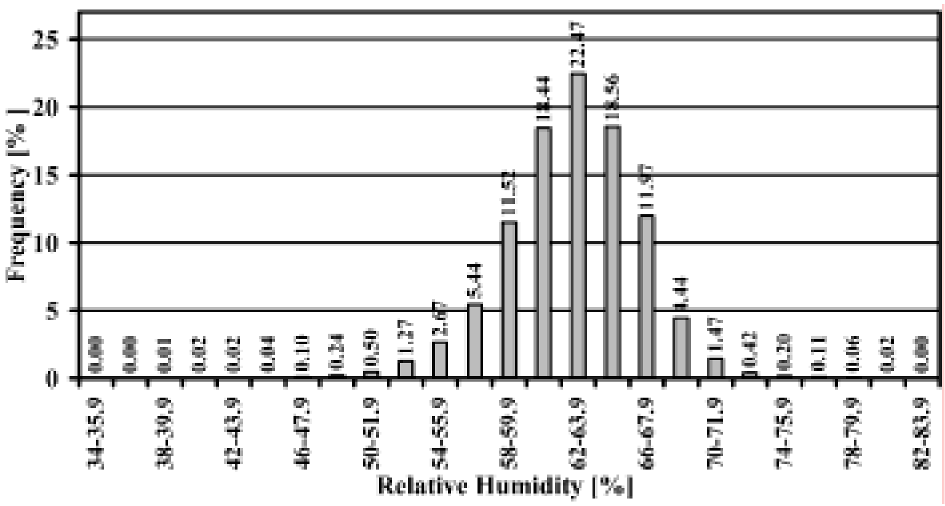

Figure 18. Histogram of relative humidity over the Dead Sea, 1992-2002.

Interannual changes, plotted from diurnally averaged data, in order to smooth diurnal fluctuations and eliminate biased data, (Figure 19) do not show any significant features, and although there seems to be a slight upward trend it is insignificant.

Particularly conspicuous on Figure 19 are the very large relative humidities observed during the winter of 1992. This winter was characterized by heavy rains and a significant addition of fresh water to the Dead Sea (e.g., Anati et al., 1995; also Anati, 1997; his figure 8-3a). These fresh waters form a thin, stable, lower salinity upper layer of about 3 to 5 m (Gertman and Hecht, 2002). Since diminishing salinity enhances the evaporation from the surface of the Dead Sea, the relative humidity over the sea is larger. The following years, as the runoff into the sea receded, the annual relative humidity pattern returned to that of previous years.

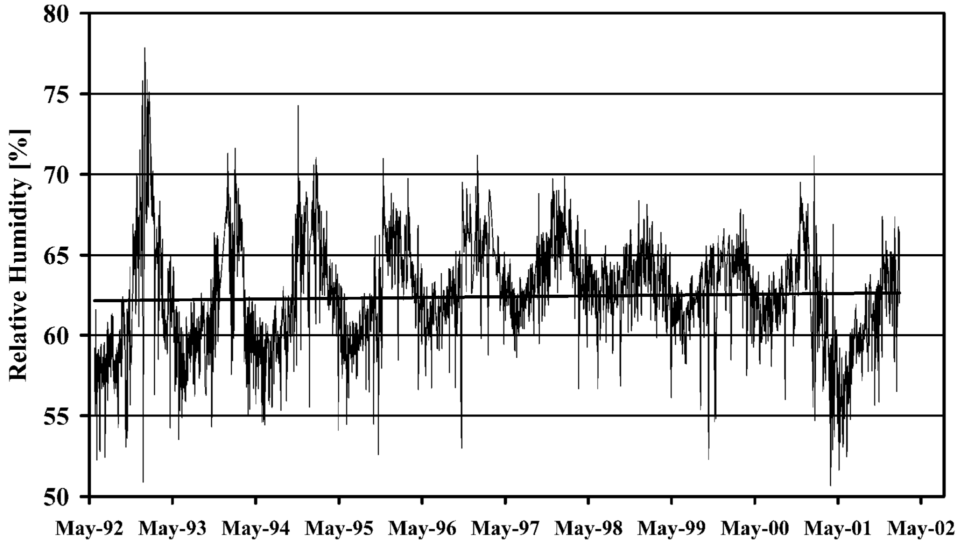

Figure 19. Interannual relative humidity changes over the Dead Sea, 1992-2002.

7. Sea Surface and Subsurface Temperature

The thermal structure of the Dead Sea waters and its seasonal changes were described in a number of investigations (e.g., Anati et al., 1987; Anati, 1997) and most recently by Oren in this volume. All these descriptions are based on data from extensive cruises (except Stanhill, 1990, who also included some satellite data). Buoy sea surface temperatures were measured at 1 m below surface by a thermistor. We obtained 235675 measurements, thus a failure rate of about 7.4%. The sea surface temperature is determined by insolation, long wave radiation, conduction, and evaporation. Thus we can expect the sea surface temperature to vary diurnally as well as seasonally.

Diurnal oscillations.

Diurnal oscillations of sea surface temperatures (SST) in the Dead Sea were not previously investigated. Annual average diurnal oscillations of the SST (Figure 20a) show that there is a minimum temperature in the morning (26.8 °C) at about 07:40 and a maximum temperature in the afternoon (27.6 °C) at about 17:00. As the other parameters, the extrema change seasonally (Figure 20b). The minimum was observed as early as 06:00 and the maximum as late as 16:00. The amplitude of the oscillation changes seasonally as well. It is almost insignificant during the winter (0.2 °C in January and December), reaching a maximum of 1.4 °C in June. The times of the extrema and the seasonal changes in the amplitude of the oscillation indicate a direct relationship with the insolation.

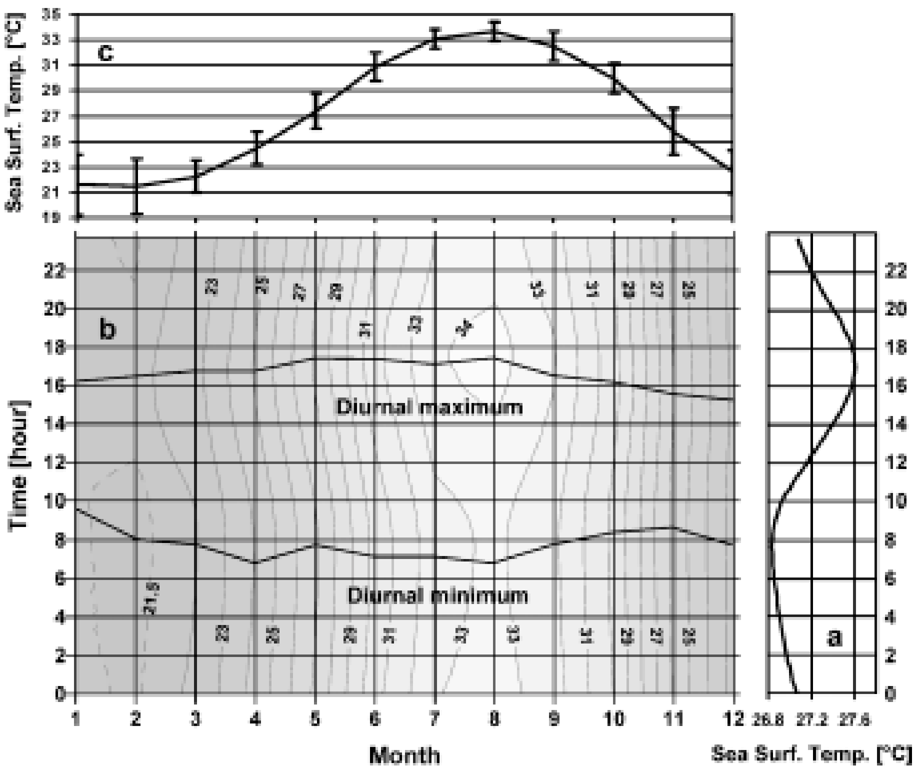

Figure 20. Sea surface temperature variations. **a.** Annual averaged diurnal oscillations. **b.** Seasonal changes in the diurnal oscillations. **c.** Annual averaged changes in the air temperature.

Annual and interannual changes.

The annual average SST (Figure 20c) presents us with an annual cycle with the lowest temperature in February (21.5 °C) and the highest (33.7 °C) in August, the largest fluctuations (indicated by the standard deviations) occur in the winter, and the smallest fluctuations occur in the summer (see Table 3).

The interannual changes of the SST (Figure 21) indicate a slight positive upward trend. A *t* test indicates that the slope is significant at 95%. Thus from 1992 to 2002 the water temperature has increased by about 0.06 °C per year.

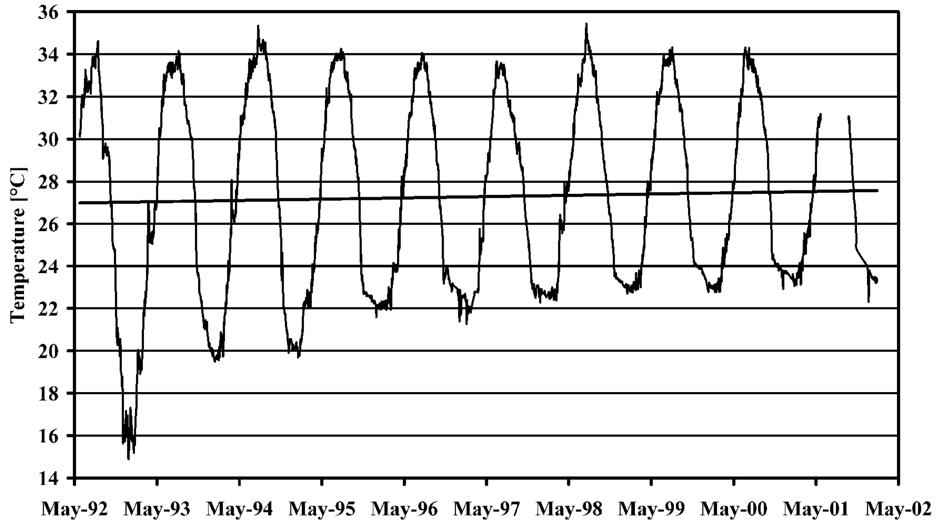

Figure 21. Interannual sea surface temperature changes, 1992-2002.

The most conspicuous feature in the interannual plot of the SST (Figure 21, plotted from the diurnally averaged SST) is the very large 1992 annual cycle. The winter of 1992-1993, a particularly rainy winter (e.g., Anati et al., 1995), which was already mentioned in the previous paragraph, added some very cold as well as fresh waters to the sea. This produced a relatively thin (3 to 5 m) stable layer that reacted rapidly to the factors determining its temperature and did not mix with the layers below. Therefore very low temperatures were observed during the winter of 1992-1993. During the following years, we noticed a gradual increase in the winter temperatures and, eventually, by the winter of 1995-1996, convection mixed the entire water column. From 1995 to the end of our observations, convection mixed the entire water column every winter and thus, throughout the winter, the water temperatures remained relatively stable.

Subsurface temperatures.

Subsurface temperature variations are a complex process (e.g., Weinstein et al., 2000) and can't be discussed in detail in the present investigation. A detailed description and discussion of the mixing as reflected in the subsurface temperatures can be found in Gertman and Hecht (2002). In general, our measurements indicate that the average temperature of the upper layer is fairly uniform (27.1-27.2 °C) down to 6 m (Table 4), from where it declines to an average of 22.5 °C at 40 m. The standard deviations (Table 4) indicate that seasonal variations penetrate down to 40 m. The vertical changes in the standard deviations indicate that intensity of the seasonal penetration is about the same down to 15 m, and it diminishes below that.

Table 4. Sea Temperatures 1992-2002

Depth meters	Average	Min	Max	StDev	Number of measurements
1	27.2	14.0	35.9	4.7	235675
2	27.2	14.6	35.9	4.7	235674
3	27.1	14.6	35.9	4.7	235675
4	27.1	14.7	35.7	4.6	235673
6	27.2	16.0	35.3	4.4	228068
8	27.0	17.1	35.2	4.4	219441
10	26.7	18.0	35.2	4.5	208319
15	26.2	18.9	34.8	4.3	221061
20	25.1	19.6	34.6	3.7	223574
30	22.9	19.7	32.5	1.2	227522
40	22.5	20.0	31.3	0.6	160436

8. Winds

The Dead Sea is a large elongated lake bounded on the east and the west by high and steep hills and relatively opened at the north and the south. Furthermore, it is situated within about 100 km from the Mediterranean coast on the west and a large desert on the east (Figure 1). These topographic conditions result in a very complex wind regime. Local winds, the local breeze (i.e., the Dead Sea breeze), the Mediterranean sea breeze, and the local katabatic winds have a considerable influence on the general wind regime of the region (e.g., Ashbel, 1954, 1975). As indicated in the introduction, there are a number of investigations of the wind regime in the region of the Dead Sea; however, practically all were carried out on data acquired on shore (e.g., Ashbel, 1954, 1975; Bitan, 1974, 1977, 1984; Gat and Karni, 1995). Perhaps the most conspicuous fact revealed upon studying those investigations are the large differences between the wind patterns observed at the land-based meteorological stations close to the Dead Sea (e.g., Bitan 1984 his Figures. 3, 6, 8, and 9). Thus, comparing our measurements at sea with any of those stations is problematic and should be attempted only with caution. This particular problem was discussed by the Hebrew University Group (Cohen et al., 1983, 1986; Mahrer et al., 1984; Weiss et al., 1987; Weiss, 1988), which was the only group to have data acquired at sea. Since their database was deemed to be insufficient for testing their meso-scale meteorological model, they attempted to enhance it with data acquired on shore. For this purpose they compared sea acquired data with what they considered the best of the shore acquired data (data obtained at Qidron, see Stanhill, 1987), as well as a comparison between the sea acquired data and data obtained at Sdom (Weiss et al., 1987). By and large, their conclusions were that the differences are significant indeed. Thus, in the description and discussion of our results we will stress the comparison between our data and the data previously obtained at sea, although we will also avail ourselves of the information obtained on shore. Unfortunately, as it was already mentioned before, the data previously obtained at sea are hampered by many gaps and failures (Weiss, 1988).

The general consensus of the investigations mentioned above appears to be that the wind regime follows a diurnal cycle as well as an annual cycle. In this respect we think that our measurements conform to the same pattern, and we will describe them accordingly.

<u>Measurements.</u>

Wind speed measurements on the buoy are the average of 20 minutes, while the wind direction is an instantaneous value at the beginning of those 20 minutes. In a turbulent environment instantaneous measurements could be erroneous. In order to minimize the potential for error, the winds were averaged over the entire hour, i.e., the wind vector components were averaged and the vector recomputed. If any of the components was missing, the particular data were ignored. We acquired 77151 valid measurements, thus a 9% failure rate. In judging the discussion of our results, one must be aware that most of our failure was during the winter months. In this respect, January and June were relatively “bad” months when we managed to acquire only 74% and 81% of the potential measurements, respectively.

<u>Annual Wind Variations.</u>

Hourly averaged wind speeds larger than 10 m/s are usually considered "strong winds" (e.g., Goldreich, 1998, p. 159). Our results (Table 5A) show that more than 99% of the wind speed events were less than 10 m/s. Strong winds have occurred throughout the year (Table 5A, 5D) and will be discussed below. Usually, we expect strong winds to occur mainly during the winter season. However, contrary to expectations, monthly averaged winds during the winter season are not necessarily significantly stronger than monthly averaged winds during the summer season (Table 5C). Therefore, the strength of the monthly averaged winds can’t be used as a criterion to identify the season (Goldreich, 1998). On the other hand, there is a significant difference in the number of strong wind events from month to month (Table 5D). Thus, there is a drastic reduction in the number of strong wind events between April and May (79 vs. 33) and a drastic increase in the strong wind events between October and November (25 vs. 74). We use this factor to define winter as November to April, and summer as May to October (in Table 5 summer months were underlined). Indeed, according to this definition far more strong wind events have occurred in the winter than in the summer (413 events vs. 119 events, Table 5D).

The wind direction and speed distribution for the entire year (Table 5B, 5C) shows that a significant part of the winds (62%) blow from the northern quarter with an average speed of 4.5 m/s, followed by winds (17%) from the southern quarter with an average speed of 3.4 m/s, 12% of the winds from the western quarter with an average speed of 3.8 m/s, and 9% of the winds from the eastern quarter with an average speed of 1.8 m/s. Thus the most prevalent northern winds are also the strongest. During the winter, we observed fewer northern winds (55%) with just about the same average speeds (4.6 m/s), but at the same time a larger percentage of southern winds (26%) with average speeds of 4.0 m/s. Westerly winds are driven by the Mediterranean sea breeze and the katabatic effects. They blow throughout the year and the proportion of westerly winds does not differ significantly from winter to summer (13% vs. 12%). The average speeds are slightly higher in the summer than in the winter (4.0 m/s vs. 3.5 m/s) indicating that the driving forces, the Mediterranean sea breeze, in particular, are somewhat stronger in the summer than in the winter. Easterly winds seem to be more prevalent and slightly weaker in the summer (11% and 1.6 m/s) than in the winter (6% and 2.0 m/s).

Table 5. Annual Wind Distribution (1992-2002, hourly averages)

A. Number of events for each wind speed class.

Speed	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	total	%	winter	%	summer	%
0-1.9	1377	1385	1350	1467	1545	1464	1737	2180	1987	2103	1615	1572	19782	25.64	8766	23.33	11016	27.84
2.0-3.9	1480	1766	1759	1702	1662	1706	2170	2143	1871	2048	1913	1873	22093	28.64	10493	27.92	11600	29.31
4.0-5.9	1207	1487	1553	1436	1393	1224	1437	1546	1437	1673	1571	1362	17326	22.46	8616	22.93	8710	22.01
6.0-7.9	930	1081	1026	973	1063	833	975	1228	999	1131	1270	1154	12663	16.41	6434	17.12	6229	15.74
8.0-9.9	454	437	385	357	418	242	287	335	342	275	650	573	4755	6.16	2856	7.60	1899	4.80
10.0-11.9	45	48	59	74	26	12	18	8	19	20	59	60	448	0.58	345	0.92	103	0.26
12.0-13.9	7	17	12	5	5	1	2		1	5	14	2	71	0.09	57	0.15	14	0.04
14.0-15.9		4	6		2						1		13	0.02	11	0.03	2	0.01

B. Number of events for each direction class.

Direction	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	total	%	winter	%	summer	%
N (315.0-44.9)	2623	3206	3503	4008	4306	3894	4517	5328	4446	4702	4282	3133	47948	62	20755	55	27193	69
E (45.0-134.9)	342	332	372	479	670	604	715	930	773	801	518	395	6931	9	2438	6	4493	11
S (135.0-224.9)	1903	1843	1163	709	575	340	590	325	507	914	1609	2431	12909	17	9658	26	3251	8
W (225.0-314.9)	632	844	1112	618	563	644	804	857	930	838	684	637	9363	12	4727	13	4636	12

C. Monthly average wind speed for each direction class.

	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	Avg	winter	summer
N (315.0-44.9)	4.6	4.8	4.5	4.6	4.7	4.1	4.1	4.2	4.3	4.4	4.9	4.8	4.5	4.6	4.3
E (45.0-134.9)	2.1	1.8	1.8	1.7	1.6	1.5	1.6	1.6	1.5	1.8	2.5	2.1	1.8	2.0	1.6
S (135.0-224.9)	4.3	4.3	4.1	3.2	2.5	2.8	3.4	2.5	2.4	2.6	3.9	4.1	3.4	4.0	2.7
W (225.0-314.9)	3.0	3.8	4.3	3.9	4.2	4.4	4.5	4.0	3.8	3.3	3.1	3.1	3.8	3.5	4.0

D. Strong winds - number of events for each direction class.

	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	total	%	winter	%	summer	%
N (315.0-44.9)	7	15	25	65	22	4	8	6	12	13	36	16	229	43	164	40	65	55
E (45.0-134.9)	3	0	1								2		6	1	6	1	0	0
S (135.0-224.9)	34	34	10			1		1		4	27	35	146	27	140	34	6	5
W (225.0-314.9)	8	20	41	14	11	8	12	1	8	8	9	11	151	28	103	25	48	40
total	52	69	77	79	33	13	20	8	20	25	74	62	532		413		119	
%	10	13	14	15	6	2	4	2	4	5	14	12		100		100		100

Weiss (1988) defines winter months as November, December, January, February, characterized by weak winds, of the order of 2 m/s, and a few strong wind events. The strongest winds observed by him are of the order of 14 m/s, which he describes as unusually strong. In his data, January is poorly represented and February comes out as the most "wintery" month with average winds of 5 - 6 m/s as well as some 12 m/s events. March is described as a transition period, most of the time quiescent but still with some 12 m/s events. There are no data for April and May. June, July, and August belong to the summer regime with daily quiescent periods as well as winds reaching up to 8 to 9 m/s. Weiss' (1988) description parallels ours in parts and the differences are probably due to his limited database, as well as to his different definition of the seasons.

Most of the investigators quoted above provide their quantitative statistics of the wind speed and direction distribution via wind roses, which are difficult to compare with our results.

For the northern basin, Ashbel (1954, 1975) shows that during the summer the most prevalent winds are northerly winds, followed by southerly winds and then, in the third place, by northwesterly winds. For the winter, Ashbel found that the northerly winds are still the most prevalent, but the northwesterly winds are in the second place and the southerly winds follow them. Ashbel does not define the seasons and, once more, some of the differences between our results and Ashbel's may stem from different definitions of the seasons.

<u>Diurnal oscillations.</u>

The most prominent feature of the wind regime over the Dead Sea is its diurnal variation as was discussed by Ashbel (1954, 1975); Bitan (1974, 1977, 1984); and Gat and Karni (1995). The diurnal wind changes are the result of the combination of the local sea breeze, the arrival of the Mediterranean sea breeze in the late afternoon, and the katabatic effects.

Figure 22, compiled from the vector averaged diurnal oscillations, depicts those oscillations and their seasonal changes. One of the features that stand out immediately in Figure 22 is the total lack of southerly winds. The southerly winds are related to two factors. One of them is passing storms, which by their nature are random and usually of short duration (see next section) and, therefore, are obliterated by averaging. The other one is the local sea breeze (e.g., Bitan, 1984, at various shore stations around the Dead Sea, in particular in the winter; Gat and Karni, 1995 at Kalia). The local sea breeze leaves practically no mark on our wind measurements at sea.

The most conspicuous feature of Figure 22 is that the strongest winds appear in the summer and, in particular, during the night. From our measurements, we can describe the winds over the Dead Sea as being very weak at about 13:00 to 14:00, intensifying towards the evening, and reaching their maximum (an average of as much as 6 m/s) around midnight. Our results for August appear to correspond to Bitan's (1984) description of the diurnal oscillation at Ein Gedi (his Figure 16), except that we observed the wind speed maximum at midnight rather than at 19:00, and we do not observe an additional wind speed maximum early in the morning (at about 05:00 in Bitan's Figure 16). Bitan relates the major wind speed peak to the Mediter-ranean sea breeze, the minor wind speed peak to the local land breeze, and as indicated above, we do not observe the local sea breeze. Weiss (1988) also emphasized the diurnal process and found that the wind speed peaks occur even earlier, at about 17:00 to 19:00.

During the winter, when both the Mediterranean sea breeze and the katabatic effects are weaker, wind speeds are lower but the pattern of the diurnal oscillation is still discernable (Figure 22).

<u>Strong winds - storms.</u>

As shown, strong winds account for less than 1% of the winds; however, they are important in forcing the currents and mixing the waters of the sea. Of the 532-recorded strong wind events (Table 5D), most (143) did not last for more than one hour and were in the class of 10 to 12 m/s. There were only 4 storms that lasted for 8, 10, 11, and 12 hours each. One must stress again that these are hourly averages, and, as one might expect, looking at the 20-minute averages much stronger winds can be found. Thus, the strongest wind observed on the Dead Sea was of 18.4 m/s and winds stronger than 16 m/s were observed on a number of occasions. The only previous report of strong winds on the Dead Sea can be found in Weiss (1988) who, as we have seen above, has observed a few 14 m/s and 12 m/s events in February and March. To the best of our knowledge none of the other investigations report winds stronger than 8 m/s, and, therefore, the reputation of the Dead Sea as a flat calm lake.

The traditional Beaufort scale defines winds in the range of 8 to 10.7 m/s as a fresh breeze; 10.8 - 13.8 as a strong breeze; 13.9 - 17.1 m/s as a moderate gale, and 17.2 - 20.7 m/s as a fresh gale. The same scale describes the respective sea states as moderate, rough, and very rough (the last referring to the 13.9 - 20.7 m/s range). In open seas, given sufficient fetch and duration, such winds can develop 1.2 - 2.4 m, 2.4 - 4 m, and 4 - 6 m waves, respectively. Thus, 3 m waves on the Dead Sea (Hecht et al., 1997) are not a freak occurrence and perhaps even higher waves do occur.

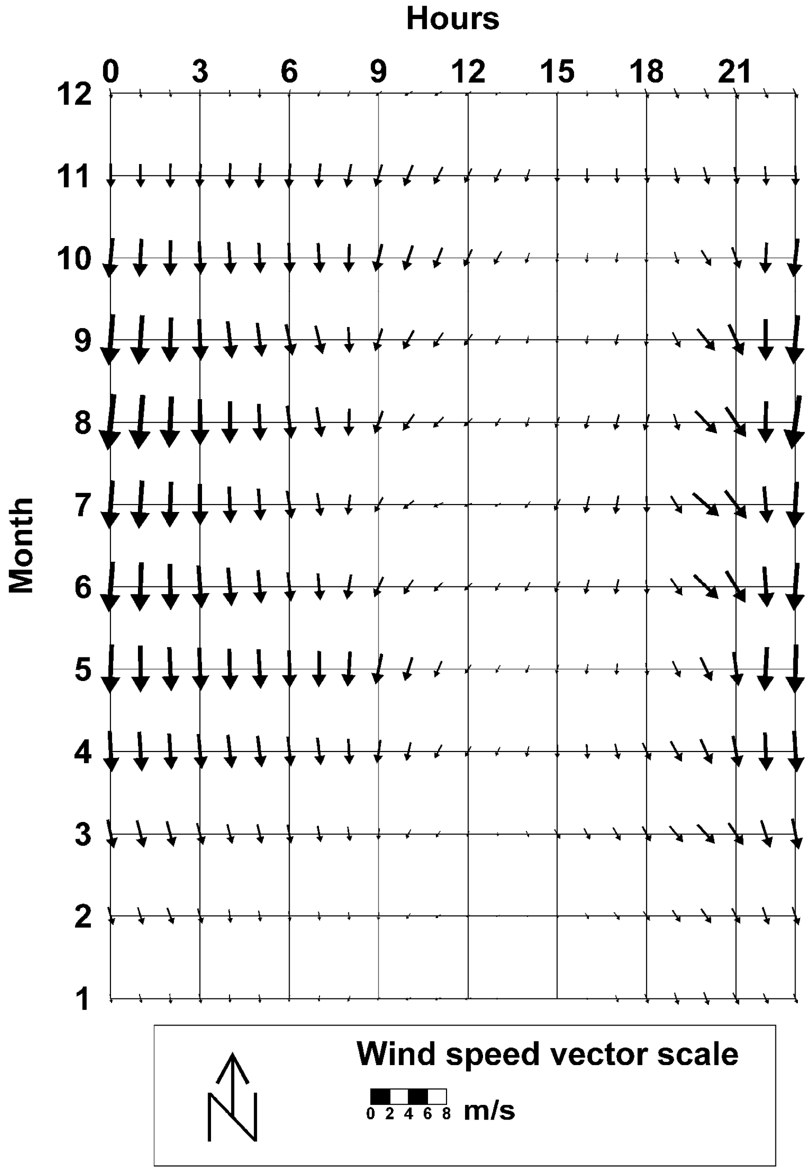

Figure 22. Wind speed diurnal oscillations over the Dead Sea.

As pointed out before, strong winds occur about twice as often in the winter than in the summer (Table 5D). Their direction is predominantly northerly. However, during the summer an almost equal number of strong winds are westerly, indicating that they are related to the Mediterranean sea breeze enhanced by katabatic effects. During the winter, there is an almost equal number of westerly and southerly strong winds. The last are probably related to the depressions crossing the region.

9. Summary

The results presented in this paper are a summary of 10 years of data acquired at sea, at 20-minute intervals from a buoy deployed in 100 m of water about 4 km southeast of Ein Gedi.

The Dead Sea environment, apart from being unique, is also dynamic. Thanks to its declining level ("drying up"), the changing concentration of the chemicals dissolved in its waters, and the diminishing freshwater resources in its vicinity, the microclimate above its surface as well as the thermohaline structure of its waters have changed significantly within a short time span, e.g., the period covered by the present investigation 1992-2002.

The meteorological parameters described in the present investigation follow a diurnal cycle as well as an annual cycle. Some of the oscillations observed at the Dead Sea are enhanced by the extreme environment, e.g., 415 m below the Mediterranean sea level, steep topography, higher temperatures, higher pressures, etc.

Solar radiation, the main driving force of all other meteorological phenomena, changes diurnally as well as seasonally. Throughout the year, the peak insolation occurs at noon (12:30 summer local time). The largest monthly averaged diurnal insolation (964 W m^{-2}) occurs in June when we also have the longest days of the year (about 12 hours). The least monthly averaged diurnal insolation occurs in December (471 W m^{-2}) when we have the shortest days of the year (about 10 hours). Over the period of our measurements, the insolation has diminished by 0.28 MJ m^{-2} day^{-1} per year and by the end of this period it was reduced by about 2.7 MJ m^{-2} day^{-1}. This reduction is ascribed to an increase in the dust and aerosols above the surface of the Dead Sea, possibly related to the desertification of the region as well as to an increase in the salt particles that were emitted from the sea surface due to its enhanced salinity.

Air pressures at the surface of the Dead Sea are larger than those at the Mediterranean sea level by about 50 mb. Random air pressure fluctuations, due to crossing pressure systems, can be as large as 20 mb. However, there are regular pressure changes as well such as semidiurnal air pressure variations known as atmospheric tides, which are driven by gravitational forces as well as by diurnal air temperature variation. Globally it is not a very intense process but it is important in prediction models (e.g., Reiter, 1975). The atmospheric tide varies from a maximum of 3.5 mb at the equator to 0.3 mb at the poles. On the Mediterranean Sea coast (at Ashkelon or Hadera), the yearly averaged maximal amplitude for 1996 was about 1.5 mb, when at the same time, the maximal amplitude over the Dead Sea was 3.9 mb. The atmospheric tide over the Dead Sea is a particularly large one. On the average its highest peak (1061.4 mb) occurs at 09:40, followed by the lowest trough (1057.4 mb) at 17:20, the next peak (1059.8 mb) at midnight, and the next trough (1059.7 mb) at 03:00. Air pressures change seasonally and are lower in the summer (1051 mb) than in the winter (1067 mb). The amplitude of the atmospheric tide also varies seasonally. The largest amplitudes (4.7 mb) were observed in June and the least amplitudes (3.1 mb) were observed in January and December. The times of the extrema also shifted. The peaks occurred earlier in the summer than in the winter, and the troughs occurred later in the summer than in the winter. Over the period of our measurements average air pressures have increased by 1.1 mb. However, over the same period the sea surface level has declined by 7 m, the equivalent of an increase of 0.9 mb in pressure. Since the buoy and the pressure gauge on it floated on the sea surface, the positive trend in the air pressure can be ascribed to sea level changes.

Winds over the Dead Sea are driven by crossing pressure systems, by the local breeze, by the Mediterranean breeze, and by the anabatic and katabatic processes. The morphological complexity of the region and the combination of the driving forces make modeling and prediction particularly difficult (e.g., Segal et al., 1983). Wind patterns and intensity change seasonally, however, indiscriminate wind speed averages provide a misleading index for the definition of the seasons. Instead, we assumed that the number of occurrences of strong wind events is a better indicator for the season.

Strong winds, i.e., wind speeds larger than 10 m/s, occurred in less than 1% of the hourly averaged wind speed measurements. Strong wind events occurred throughout the year but the number of strong wind events diminished significantly from April (15 events) to May (6 events) and increased significantly from October (5 events) to November (14 events). Accordingly, we defined "summer" as May to October (with 119 events) and "winter" as November to April (with 413 events).

In general, during the summer, winds are mostly northerly (69%). During the winter, northerly winds are still prevalent (55%) but there is an increase in the southerly winds (26% in the winter vs. 8% in the summer). Thus the winter is characterized not only by a larger number of strong wind events but also by an increase in the percentage of southerly winds. Westerly winds, associated with the Mediterranean breeze, occur at about the same rate throughout the year (12%) and, on the average, at about the same speeds (3.5-4.0 m/s).

Waves, driven by strong winds, rise up very fast but they also dissipate quickly (Hecht et al., 1997). Due to the much higher density of the water, breakers are far more destructive than their equivalent in the open ocean. Such events are not frequent but when they do occur they can lead to catastrophic results as, for instance, during the summer of 1983, when high waves destroyed the experimental Dead Sea floating solar pond in its early stages of construction.

The diurnal cycle of the winds is one of the most conspicuous characteristics of the wind regime over the Dead Sea and has elicited most of the interest of the researchers. Diurnally, the strongest winds occur at midnight. As the day progresses, wind strength diminishes with the weakest winds occurring between 13:00 - 14:00. However, the diurnal cycle is also affected by the seasons, and although the timing of the minima and the maxima of the wind intensity do not change significantly with the seasons, the intensity of the winds does change from winter to summer. Thus, we found the weakest winds in the winter and the strongest winds during the summer, e.g., during the summer the winds at midnight can reach an average speed of 6 m/s.

Air temperatures vary diurnally and annually. Annually we find August to be the hottest month with an average air temperature of 32.9 °C, but slightly lower than the average air temperature over land (33.4 °C at Sdom). The coldest month is January with an average temperature of 18.3 °C, but slightly warmer than the average air temperature over land (15.7 °C at Sdom). Normally, the peak diurnal air temperature occurs about two hours after the peak insolation. Over the Dead Sea, the air temperature changes are also closely associated with the arrival of the Mediterranean breeze as well as with the onset of the local breeze. Therefore, the times of the local extrema are entirely different than those along the Mediterranean coast. In particular, the local afternoon maximum is much higher as well as much later than one would expect solely from the insolation. The peak afternoon air temperature changes with the seasons from an average maximum of 35.4 °C in July, at 19:00, to an average minimum of 19.6 °C in January at 17:00. The highest air temperature observed over the Dead Sea, 43.4 °C, occurred on the 5^{th} of July 1998.

Relative humidity is one of the very important factors in the determination of evaporation from the Dead Sea (e.g., Steinhorn, 1997). In particular, the range of the relative humidity is crucial for the determination of the selective precipitation of minerals from its waters (e.g., Krumgalz et al., 1997, 2000). In general, the relative humidity at sea is significantly larger than that measured on shore. The relative humidity over the Dead Sea varies between 35 to 85%; however, these extreme values are infrequent, and 98% of the relative humidity is between 50 and 70%. Seasonally the changes in average relative humidity are small, with the lowest monthly average humidity in July (60%) and the highest monthly average humidity in December (65%). Diurnally, the relative humidity is high in the morning and low in the late afternoon. The times of the minima and the maxima are related to the local breeze and the Mediterranean breeze, respectively.

Sea surface temperature is directly related to the insolation, heat conduction, heat convection, and back radiation whereas the air above it receives most of its heat from the sea surface and a minor amount from radiation absorption by moisture. As such, it is instructive to observe the behavior of the sea surface temperatures versus that of the air above it (Figure 23). The two histograms depict a characteristic bimodal distribution and it is important to understand the meaning of this distribution.

The sea surface temperatures (SST) are distributed over a relatively narrow range: from 14 to 36 °C. There are two prominent peaks, one at about 23.5 °C and the other at about 33.5 °C as well as a smaller peak at about 16.5 °C. From Figure 21 we can learn that the rainy winter of 1992-1993 resulted in very low SSTs, due to the thin layer of relatively fresh water, which did not mix with the layers below it and therefore cooled rapidly. This was a short-lived meromictic period of the sea. During the following years, we did not encounter such massive fresh water inflow, and the sea became holomictic. The waters of the upper layers mixed with the layers beneath them and the temperature of the mixture was higher. The process, as can be seen in Figure 21, consists of the SST diminishing rapidly from the summer peak and, as the surface waters mix with the lower layers, the temperature stabilizes at around 22.3 °C. As we approach summer, the temperature increases rapidly (Figure 21) and a thermocline forms, which eventually settles at about 30 to 40 m. From this point on, the insolation absorbed in the upper layer has to heat up a thicker layer of water, and, therefore, the temperature remains stable at about 33.5 °C (Gertman and Hecht, 2002).

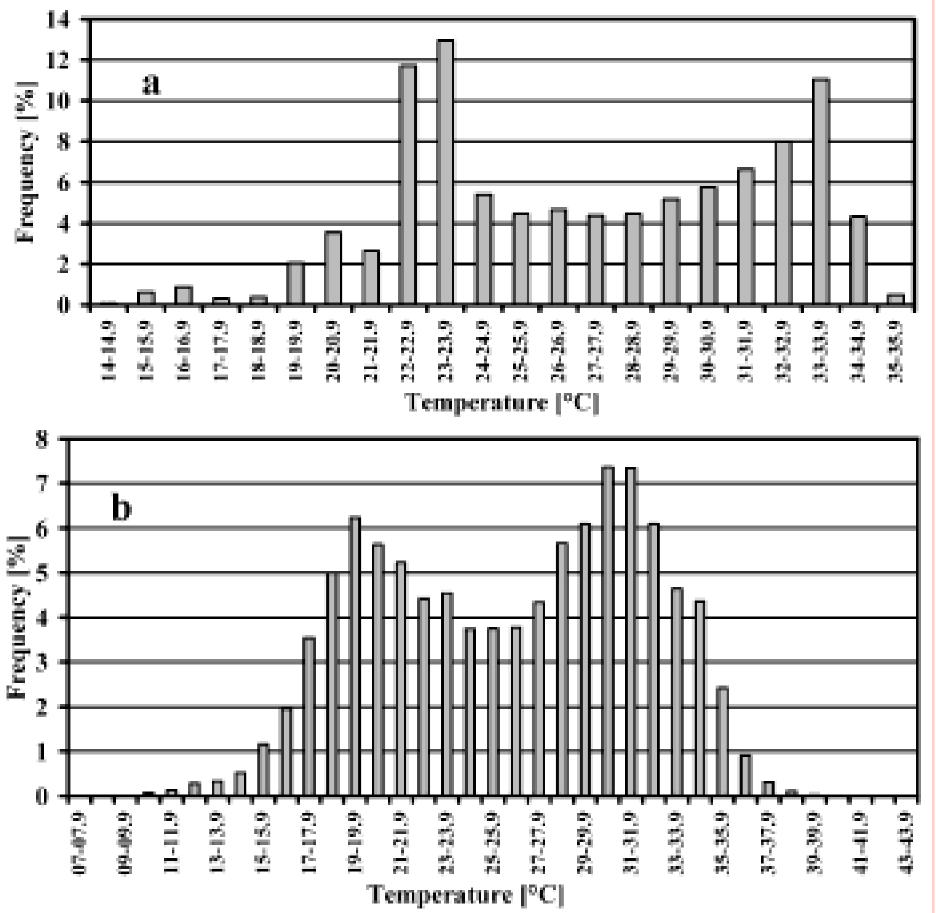

Figure 23. Sea surface (a) temperatures and air (b) temperature histograms, 1992-2002.

The air temperature above the sea follows suit and its histogram also depict two modes: one is the winter mode at about 19.5 °C and the other one is the summer mode at about 31 °C.

The air temperature however, is also influenced by the Mediterranean sea breeze and by the local breeze, and, therefore, we can expect a diurnal and an annual oscillation in the difference between the SST and the air temperature above it. Figure 24a depicts the average of this diurnal oscillation and shows that we can expect a maximum difference in the morning (of the order of 3 °C) and a slightly negative difference, about -0.5 °C (i.e., the air is warmer than the sea), in the evening. The last is the effect of the warm air brought in by the Mediterranean sea breeze. While the diurnal maximum is almost steady in time and in value throughout the year (Figure 24b), the diurnal minimum changes seasonally and is more intense and later during the summer months. Thus we find the largest difference in January and a negative difference in May (Figure 24c).

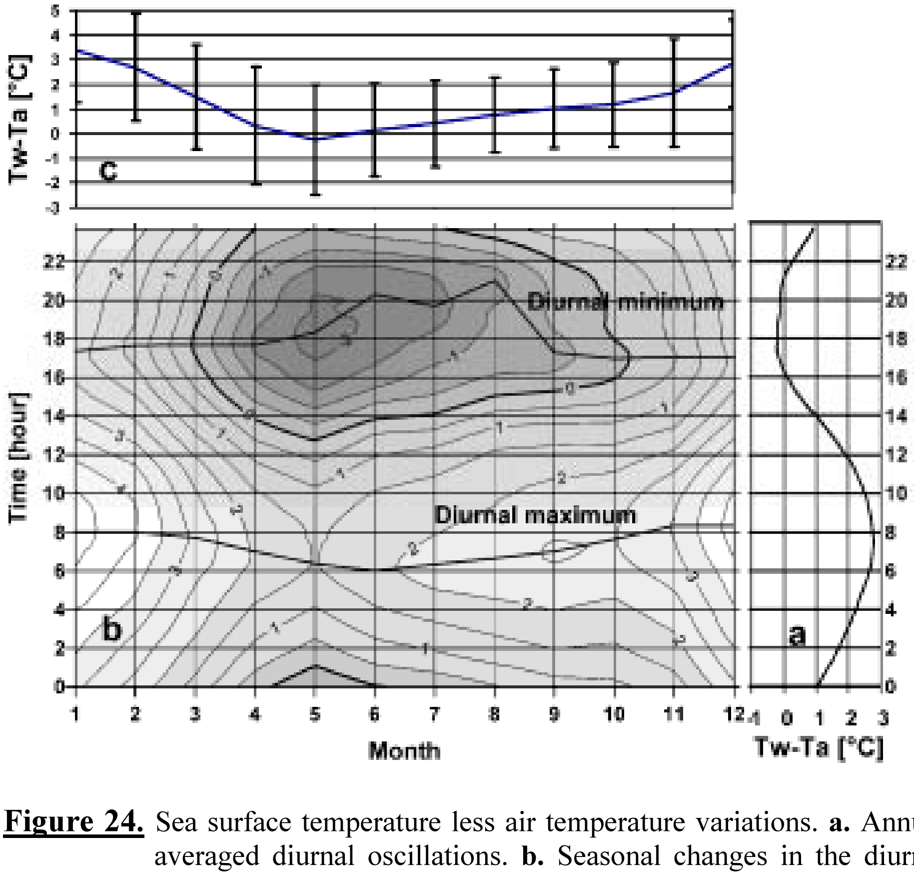

Figure 24. Sea surface temperature less air temperature variations. **a.** Annual averaged diurnal oscillations. **b.** Seasonal changes in the diurnal oscillations. **c.** Annual averaged changes in the air temperature.

Over the period of our measurements, our data indicates positive trends in the SSTs as well as in the air temperatures. The slope of the SST trend, 0.06 °C per year, proved to be significantly different from zero with a probability of 95%. The slope of the trend in the air temperature was not significantly different from zero.

In conclusion, our investigation indicates that data obtained from shore stations provide a poor approximation for the microclimate over the sea. Furthermore, almost ten years of meteorological data acquired at sea provided a large database and statistical information indispensable for any proper modeling effort in this region. Ten years of data acquisition have proven that long-term trends do occur. Therefore, it is imperative to continue the meteorological and oceanographic measurements both in order to be able to continue to monitor those trends as well as to test the validity of the models.

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