4 Evapotranspiration losses from the Nylsvlei floodplain

4.1 Introduction

Evapotranspiration on the inundated Nylsvlei floodplain is a potentially significant loss to floodwaters that would influence the hydraulic behaviour of the floodplain, and therefore needs to be included in a hydraulic model of the floodplain. Evapotranspiration varies with the seasons and so it was necessary to include these variations on at least a seasonal basis. These variations were included on a monthly basis as an average daily loss for each month of the year, and this chapter is about how losses to evapotranspiration were measured and how a monthly averaged daily loss was derived.

All atmospheric moisture originates from the Earth's surface where water in its liquid and solid phases is transformed into water vapour at an evaporating surface and is transported into the atmospheric boundary layer (Schulze, 1997). This water originates from two processes, namely evaporation and transpiration (with sublimation also occurring from ice in cold climates). Evaporation occurs from soil surfaces and from open water surfaces. Transpiration is physically the same process as evaporation but represents water loss from soil through the stomata on leaves (mainly) of plants (Schulze, 1997). Usually evaporation and transpiration are lumped together in the term evapotranspiration, as it is difficult to measure them separately, but easier to measure them together.

Evaporation is controlled by three atmospheric conditions, namely (Schulze, 1997):

- The capacity of air to take up more water vapour (which is influenced by temperature and relative humidity (RH) of the air)
- The amount of energy available for the latent heat used in the process of evaporation, with the energy provided mainly by solar radiation

- The degree of turbulence (related to wind) in the lower atmosphere, where the turbulence is necessary to replace the saturated layers above the evaporating surface with unsaturated, drier air from higher levels or from different air masses

These three factors create an atmospheric demand and when this demand is met fully, when the soil is wet and vegetation covers the ground completely and is growing actively, potential evapotranspiration takes place. Potential evapotranspiration is defined as evaporation from an extended surface of a green crop, which fully shades the ground, exerts negligible resistance to the flow of water and is always well supplied with water (Rosenberg *et al*, 1983).

Since there are many methods available to determine evapotranspiration and it can become confusing trying to decide which one to use, the United Nations Food and Agricultural Organisation (FAO) attempted to standardise the methods used to find evapotranspiration and to define evapotranspiration itself (Allen *et al*, 1998). Once the standard FAO evapotranspiration is known for an area, crop factors can be applied to find the evapotranspiration for a specific crop. The FAO evapotranspiration is based on a hypothetical grass as no one grass species can grow in all places around the world. The FAO defines reference evapotranspiration as: "the rate of evapotranspiration from a hypothetical crop height of 0.12m, a fixed canopy resistance of 70s/m and albedo of 0.23, which would closely resemble evapotranspiration from an extensive surface of green grass cover of uniform height, actively growing, completely shading the ground and not short of water" (Schulze, 1997). Albedo is defined to be the fraction of incident light or radiation that is reflected by an object or surface (Encarta Dictionary, 2003). There are various equations that have been calibrated specifically to find the reference evapotranspiration.

The actual evapotranspiration rate can be either equal to or less than the potential rate, depending on the soil moisture state and the plant moisture stress. On the Nylsvlei floodplain under flooded conditions in summer, it is reasonable to expect that the actual evapotranspiration would be either equal to or very close to the potential evapotranspiration rate.

The usual method of determining the actual evapotranspiration rate for a site is to find its potential or reference evapotranspiration and then to adjust this figure for conditions where the soil moisture content is limiting, and for the effect of different vegetation types (Blight, 1992). The hydraulic model was concerned with the evapotranspiration that occurs from inundated areas and an estimate of this rate was required to account for this loss.

There are also methods that estimate the actual evapotranspiration without first estimating the potential evapotranspiration, but these are complex and difficult to use (Blight, 1992). However, Blight (2002a) carried out some measurements using one of these methods in the Nylsvley Reserve, based on the energy balance. As the process of evapotranspiration depends on the energy balance, and on the principle of conservation of mass, many methods of assessing evapotranspiration are based on determining these balances.

4.2 Methods to determine potential evapotranspiration

There are three main method classes for predicting potential evapotranspiration (Blight, 1992):

- climatological models
- micrometeorological methods
- direct measurements

4.2.1 Climatological models

Some of the more popular climatological models are given below, but there are many models available. These models are popular due to their ease of application and the fact that they do not require vast amounts of data. They are empirically derived, although some of them do have physically based elements (Blight, 1992). Due to this, care should be taken in applying them only to similar conditions to which they were developed (Blight, 1992). These models are based on either air temperature, or radiation or a combination of these. The temperature models are useful because they have minimal data requirements and are accurate in the regions in which they were developed and for periods greater than a week (Blight, 1992).

Thornthwaite (1948)

This temperature-based model requires day length but is unsuitable in the tropics and for periods of less than a month (Scholes & Savage, 1989). Potential evapotranspiration (E) in cm/month is given by:

$$E = 1.62\left(\frac{10T}{\sum i_j}\right)^a \tag{4.1}$$

where:

- T is the mean daily temperature (˚C)
- i is $(T_{mean}/5)^{1.514}$

and

$$a = 6.75\times10^{-9}\left(\sum i_j\right)^3 - 7.71\times10^{-7}\left(\sum i_j\right)^2 + 0.01792\left(\sum i_j\right) + 0.49239 \tag{4.2}$$

Blaney and Criddle (1950)

This temperature-based model needs day length and an empirical crop factor (Scholes & Savage, 1989). Potential evapotranspiration (E) in inches/month is given by:

$$E = K\sum\left(\frac{tp}{100}\right) \tag{4.3}$$

where:

- K is a crop constant
- t is temperature (˚F)

- p is monthly % of daytime hours in the year

If monthly values for the consumptive use coefficient K are available, monthly consumptive use (evapotranspiration) can be found by using (Viessman & Lewis, 1996):

$$u = \frac{Ktp}{100} \tag{4.4}$$

Where u is the monthly consumptive use (evapotranspiration) in inches.

Linacre (1977)

This temperature-based model for finding potential evapotranspiration (E) has two forms, one for the case of a lake surface with an albedo of 0.05 and one for the case of well-watered vegetation with an albedo of 0.25. These equations need latitude, altitude and either relative humidity or daily temperature range to provide an estimate for evaporation on a lake surface or evapotranspiration from well-watered vegetation accurate to (Linacre, 1977):

- 1.7mm daily, 0.5mm for a monthly mean, 0.9mm/day for a week and 0.3mm/day on a yearly basis for a lake
- 0.7mm/day for a monthly mean and 1.0mm/day for a weekly mean for well-watered vegetation

For the surface of a lake with an albedo of 0.05:

$$E = \frac{\left(\frac{700T_m}{100-A}\right) + 15(T - T_d)}{80 - T} \tag{4.5}$$

For well-watered vegetation with an albedo of 0.25:

$$E = \frac{\left(\frac{500T_m}{100-A}\right) + 15\left(T - T_d\right)}{80 - T} \tag{4.6}$$

where:

- T is the mean temperature in degrees Celsius
- $T_m = T + 0.006h$ (sea level equivalent temperature)
- h is the elevation in metres
- T_d is the mean dew point temperature in degrees Celsius
- $\left(T - T_d\right) = 0.0023h + 0.37T + 0.53R + 0.35R_{ann} - 10.9$ (˚C), provided precipitation is at least 5 mm/month and $(T-T_d)$ is at least 4˚C (Linacre, 1977)
- R is the mean daily temperature range in degrees Celsius
- R_{ann} is the difference between the mean temperatures of the hottest and coldest months in degrees Celsius
- A is the latitude in degrees

Jensen and Haise (1963)

This model is based on solar radiation and temperature and gives a monthly mean daily estimate of potential evapotranspiration (E) in mm, but underestimates in advective conditions (Scholes & Savage, 1989). The FAO has noted that radiation methods can give good results in humid climates but underestimate in arid climates (Allen *et al*, 1998).

$$E = \left(0.14T - 0.37\right)R_s \tag{4.7}$$

where:

- T is the monthly mean of the mean daily temperature (˚C)
- R_s is the monthly mean solar radiation at the top of the atmosphere in MJ/m^2/day

Hargreaves (1994)

Hargreaves's equation is based on temperature and radiation and has been found to give results that are closest to the classic Penman Equation (see later), for finding the reference evapotranspiration, by The Centre Commune de Recherche of the European Economic Community in 1992 (Hargreaves, 1994). The FAO found this equation to be the best of the empirical temperature and radiation methods for finding reference evapotranspiration, having global validity (Allen *et al*, 1998).

$$ET_0 = 0.0023 \times R_a \times (T_m + 17.8) \times (T_{\max} - T_{\min})^{0.5} \tag{4.8}$$

where:

- R_a is extraterrestrial radiation, which depends on the latitude and is measured in equivalent units of water evaporation, mm
- T_m is mean daily temperature (˚C)
- T_{max} is maximum daily temperature (˚C)
- T_{min} is minimum daily temperature (˚C)

4.2.2 Micrometeorological methods

There are a number of these models, which have been derived from physical bases, but they are complex and require numerous data that are seldom easily obtainable (Blight, 1992). These methods include those of Dalton (1802) and Penman (1948), a reduced form of the Dalton equation where the data are more easily measured. Many modified forms of the Penman model have been proposed over the years. A modified form of the Penman-Monteith method, itself a modified form of the Penman model, was adopted by the FAO (Allen *et al*, 1998) as a standard method to find the reference evapotranspiration.

Dalton (1802)

The Dalton Equation determines the potential evaporation rate only by the vapour pressure deficit of the air and the ability of the air to carry water vapour away from the surface given as a function of the wind speed (Evans and Jakeman, 1998):

$$E = f_D\left(\bar{u}\right)\left(e_s^* - e_a\right) \qquad (4.9)$$

where:

- $f_D\left(\bar{u}\right)$ is a function of mean wind speed $\bar{u}$
- e_s^* is the saturation vapour pressure at the temperature of the surface
- e_a is the vapour pressure of the air at the surface

The term e_s^* is difficult to assess, since the temperature of the soil surface is required (Blight, 1992).

Penman (1948)

Penman used an energy balance to reduce Dalton's equation to a form for which data may be more easily measured (Blight, 1992). Despite this, the data required are still often not readily available at most weather stations, such as the weather station at the Nylsvley Reserve. According to the FAO, the Penman method may require local calibration of the wind function to give satisfactory results (Allen *et al*, 1998).

$$LE = \left[\left(R_s\left(1-r\right)\left(0.18+0.55\frac{n}{N}\right)\right)-\left(\sigma T_a^4\left(0.56-0.09e_a^{0.5}\right)\left(0.1+0.9\frac{n}{N}\right)\right)\right]$$
$$+\frac{\gamma}{\Delta}\left(0.35\left(1+\frac{u_2}{100}\right)\left(e_s-e_a\right)\right)+\gamma \qquad (4.10)$$

where:

- E is potential evapotranspiration
- L is latent heat of vaporisation of water
- Δ is the slope of the saturated vapour pressure vs. temperature curve at air temperature
- R_s is solar radiation at the top of the atmosphere
- r is the albedo
- n/N is the ratio of bright sunshine to sunshine

- σ is the Stefan-Boltzman constant = 2.01 x 10^{-9}mm Hg/d
- T_a is the absolute mean daily air temperature (K)
- u_2 is the mean wind velocity at 2 metres above ground level
- e_d is the saturated vapour pressure at dew point
- e_a is the actual vapour pressure at air temperature
- e_s is mean saturation vapour pressure at air temperature
- γ is the psychometric constant (0.49mm Hg/ºC)

There are many modified forms of the Penman equation. These include the Penman-Monteith equation (Monteith, 1965), which takes stomatal resistance into account (Blight, 1992).

FAO Penman-Monteith Equation (Allen et al, 1998)

This is a modified form of the Penman-Monteith equation (Allen *et al*, 1998), adopted by the FAO as the recommended equation to find the reference evapotranspiration (as defined by the FAO) anywhere in the world, after an exhaustive study investigating many of the evapotranspiration methods available. This equation, like the Penman equation, unfortunately requires data that is not always readily available at weather stations, such as the one at the Nylsvley Reserve.

$$ET_0 = \frac{0.408\Delta(R_n - G) + \gamma \frac{900}{T+273} u_2 (e_s - e_a)}{\Delta + \gamma(1 + 0.34u_2)} \tag{4.11}$$

where:

- ET_o is the reference evapotranspiration (mm day^{-1})
- R_n is the net radiation at the crop surface (MJ m^{-2} day^{-1})
- G is the soil heat flux density (MJ m^{-2} day^{-1})
- T is the mean daily air temperature at 2 m height (°C)
- u_2 is the wind speed at 2 m height (m s^{-1})
- e_s is the saturation vapour pressure (kPa)

- e_a is the actual vapour pressure (kPa)
- e_s - e_a is the saturation vapour pressure deficit (kPa)
- Δ is the slope of the vapour pressure curve (kPa °C^{-1})
- γ is the psychrometric constant (kPa °C^{-1})

4.2.3 Direct measurements

Some of the direct evapotranspiration methods have been used at Nylsvlei in the past and are described below.

Lysimeters

A soil profile is reconstructed in a large container, usually in the field. The container is mounted on a pressure cell so that mass changes can be measured. Provision is usually made for the collection of water percolating through the bottom of the container. This is an expensive method and is not accurate since water movement is disrupted by the base of the pan, which is only about a metre deep (Scholes and Savage, 1989). Plants can also be planted in these lysimeters to simulate evapotranspiration for certain habitats and this was done for the savanna habitats at Nylsvley (Moore, 1980), but not for the wetland habitat.

Pan evaporation

Evaporation pans are often installed at weather stations and so this type of data is generally readily available. The standard pan usually used is the American ‘class A pan’. Other types include the Symons pan (S pan) and the ‘mini’ or M pan, a smaller portable evaporation pan. The different pan types evaporate at slightly different rates due to their size differences and the effect that this has on water temperature and other factors that affect evaporation. Conversion factors and equations are available to convert evaporation rates between the pan types.

The M pan measures 500mm in diameter and 150mm deep and has been found by calibration to evaporate 9% more than a standard A pan and 11% more than a standard Symons or S pan (Blight *et al*, 2001).

Bosman (1990) proposed the following two equations (mm/month) to convert A pan evaporation data to S pan evaporation data, or S pan evaporation data to A pan evaporation data for a representative environment (i.e. grass) in South Africa:

$$S\,pan = -16.2354 + 0.8793(A\,pan) \tag{4.12}$$

$$A\,pan = 26.3622 + 1.0786(S\,pan) \tag{4.13}$$

For the A pan, potential evaporation is taken to be equal to 0.7 times the pan evaporation (Hojem, 1988).

An S pan has been in operation at Nylsvley intermittently since the 1970s. There are also A pan and S pan records for weather stations near Nylsvley such as at Du Toit's Kraal (about 10km downstream of Nylsvley and still within the Nylsvlei floodplain), Bela-Bela (formerly Warmbad) and Mokopane (formerly Potgietersrus).

Evaporation pans can be sensitive to inaccuracies if not used correctly. Large differences in evaporation rates can be found over short distances, sometimes from a change in microclimate due to relief. Care must be taken to ensure that the microclimate of the evaporation pan matches the microclimate of the site to be simulated. Inaccuracies in pan measurements can also arise from reading errors.

Schulze and Maharaj (1991) mapped the potential A pan evaporation across South Africa using equations based on temperature devised specifically for South Africa. The country was divided into evaporation regions and an equation was developed for each region. These equations described the A pan evaporation (which represents the potential evaporation) across South Africa. Good agreement between A pan data and the equations was found. These data were subsequently used by Midgley *et al* (1994) to divide South Africa into evaporation zones. Nylsvley Nature Reserve falls into evaporation zone 1C, quaternary sub-catchment A61C, and this quaternary sub-catchment has according to the report, an MAE (mean annual evaporation) of 1741mm. The areas around the reserve, which includes the greater Nylsvlei floodplain area also fall into evaporation zone 1C and consist of quaternary sub-catchments A61A, A61B, A61D and A61E. These quaternary sub-catchments have very similar MAEs. Using these MAEs, the monthly evaporation for each quaternary sub-catchment could be found using a monthly percentage of the MAE for the evaporation zone (1C) as specified in the report.

4.3 Methods to determine actual evapotranspiration

Actual evapotranspiration is defined as the evapotranspiration that is 'actually' occurring in the environment. The actual evapotranspiration rate is often less than the potential evapotranspiration rate as the rate at which water can be 'drawn out' of the soil by plants and through overcoming suction in the soil is less than the rate at which water can be evaporated from the surface, which is determined by the radiation received. This is certainly the case for a relatively arid country like South Africa (Blight, 1997a).

There are various ways of determining actual evapotranspiration. These include methods where transpiration and evaporation are measured separately, and where it is measured as the lumped term evapotranspiration. To find transpiration, the uptake of water by vegetation depends on the amount of water stored in the soil, atmospheric conditions and the physical characteristics of the plant, such as leaf surface area and root distribution. The plant offers resistance to flow and this resistance can change, especially in the leaves where the extent to which the stomata open depends on the availability of water. There are various methods of measuring actual transpiration rates including techniques of (Scholes and Savage, 1989):

- gas analysis (where the plant is enclosed in a chamber and the humidity due to transpiration is measured)
- lysimetry (discussed previously)
- cut shoot methods (where a plant shoot is cut off and the mass of the excised shoot is accurately monitored over a short period of time, the

change in mass over this time is assumed to represent the pre-excision transpiration rate)

- micrometeorological methods

4.3.1 The surface energy balance method

This method was devised by Bowen (1926) and until recently has been difficult to use due to a lack of suitable instruments to measure solar radiation (Blight, 2003). The method is described in detail by Blight (2002b). The surface energy balance method was carried out on certain days between 1999 and 2001 by Blight (Blight, 2001; Blight *et al*, 2001; Blight, 2002a; Blight, 2003) on the floodplain at the Nylsvley Reserve.

Quoting Blight (1997a): "The evaporative process is primarily one of energy consumption. Energy must be supplied to provide the latent heat of vaporisation necessary for water to evaporate at the soil (or water surface), and the resultant water vapour must be swept away by air movement or dispersed by diffusion to maintain the evaporation gradient and so keep the evaporative process going. If the amount of energy consumed by evaporation can be computed, the corresponding mass of water evaporated can be deduced".

The method works as follows (Blight *et al*, 2001): incoming and reflected solar energy are measured, together with temperature changes of the near surface soil, and temperature and relative humidity gradients in the air above the soil surface. Wind speeds are also measured. The surface energy balance equation, which describes how R_n (the net incoming radiation) is converted at the ground surface, is (Bowen, 1926):

$$R_n = G + H + L_e \qquad (4.14)$$

where:

- G is the soil heat flux, the energy consumed in heating the near-surface soil

- H is the sensible heat flux, the energy consumed in heating the air above the surface (this is "sensible" because this is the heat that can be sensed or felt by the observer)
- L_e is the latent heat flux of evaporation, the heat consumed in evaporating water from the surface
- R_n is the net radiation flux for the surface (incoming solar and diffuse sky radiation, less reflected radiation and outgoing long-wave terrestrial radiation), see Figure 4.1 (a).

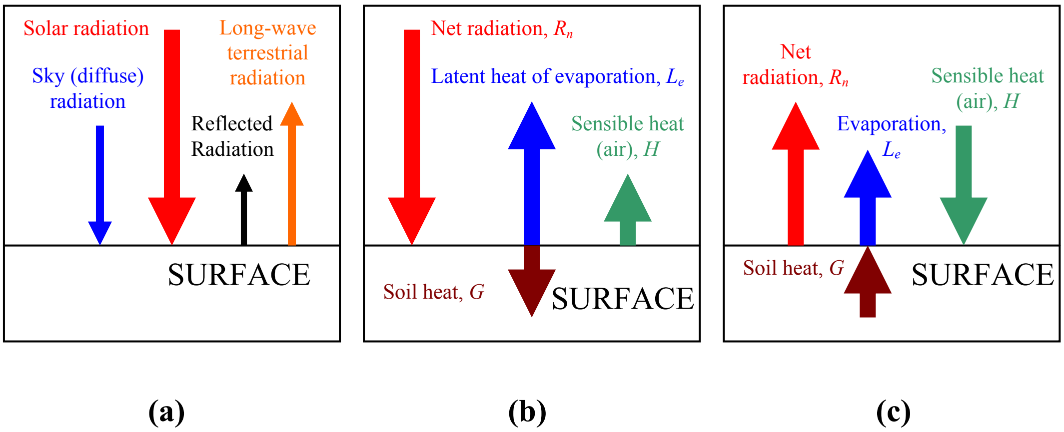

Figure 4.1: Components of the radiation balance at a soil surface: (a) incoming and outgoing radiation; (b) radiation interchange during the day; (c) radiation interchange at night (taken from Blight, 1997a)

Figure 4.1 (b) shows these energy components during the day, while there is a net inflow of energy, and Figure 4.1 (c) shows the balance at night, when the net energy flow is outwards (Blight, 1997a). Although the figure does not show this, the amounts of energy involved during the day are much larger than during the night.

The net radiation flux (R_n) can be measured or estimated directly by the difference between incoming and outgoing radiation. The soil heat flux (G) can be estimated from changes in temperature of the near-surface soil and the specific heat capacity of the soil at its current water content, and H can be calculated in a similar way, or from temperature and relative humidity gradients in the near-surface air. The sensible heat flux (*H*) is generally ignored though as it is a very small value due to the low density of air compared to that of the soil, and thus the far lower ability of the air to absorb energy. The energy fluxes are measured in Watts (Joules per second). When integrated over a period of time, e.g. 12 hours, the terms in the equations are expressed in Joules.

The soil heat flux *(G*) is calculated from the depth of the soil heated diurnally by the incoming radiation:

$$G = z_G(\Delta T)C_G \rho_G \tag{4.15}$$

where:

- z_G is the depth of soil heated (m) (z_G is usually 0.2 to 0.25m)
- (ΔT) is the average measured rise in temperature over depth z_G
- C_G is the specific heat of the soil (kJ/kg/˚C)
- ρ_G is the bulk density of the soil in kg.m^{-2}

The specific heat of the soil is conventionally taken as:

$$C_G = C_{Gd} + wC_w \tag{4.16}$$

where:

- C_{Gd} is the specific heat of the dry soil particles and has a value of about 0.85 kJ/kg/˚C
- C_w is the specific heat of the water and has a value of 4.19 kJ/kg/˚C
- w is the gravimetric water content of the soil (mass of water/mass of solids)

The large difference in the specific heat of water and soil means that even at very low moisture contents in the soil, water still has a significant effect on the overall specific heat of the body of soil.

With measurements of G and R_n, L_e can be calculated. The sensible heat flux (H) is ignored, due to its small value. Over the course of a day, the area under the curve of L_e can be integrated, and divided by the latent heat of vaporisation of water in J/kg (2.470MJ/kg) giving the potential evaporation loss in kg/m^2 of water per day, which is equivalent to mm/day. This gives the evaporation or evapotranspiration for the day or time period concerned.

For South African conditions, Blight (1997a) maintains that the major proportion of the net radiation is transformed into latent heat of evaporation, with components G and H being relatively minor.

Table 4.1: Example of a day's measurements (18 December 2000) for the energy balance method and method of calculation to find the evapotranspiration for a grassy area at the Nylsvley Reserve, after Blight (pers. comm.)

Grassy Area							
Time	**Rn W/m^2**	**Rn kJ/m^2**	**ΣRn kJ/m^2**	**Soil ΔT°C**	**ΣG kJ/m^2**	**ΣLe kJ/m^2**	**ΣET (mm)**
6:00	20	36	36	0	0	0	0.0
7:00	132	475	511	1	357	154.2	0.1
8:00	145	522	1033	2.3	821	212.2	0.1
9:00	407	1465	2498	3.7	1321	1177.4	0.5
10:00	440	1584	4082	2.7	964	3118.4	1.3
11:00	450	1620	5702	5	1785	3917.4	1.6
12:00	260	936	6638	4.3	1535	5103.4	2.1
13:30	270	1458	8096	5.3	1892	6204.4	2.5
14:30	200	720	8816	4.7	1678	7138.4	2.9
15:30	245	882	9698	4	1428	8270.4	3.3
16:30	230	828	10526	5.3	1892	8634.4	3.5
17:30	36	130	10656	5	1785	8871	3.6

Table 4.1 shows an example of a day's worth of measurements for a grassy area on 18 December 2000 observed at the Nylsvley Reserve (Blight, pers. comm.). Two sites were observed, a grassy area and a degrassed area where all the green grass was cut off, with a layer of dead surface litter left on the surface. The observations were used to calculate the evaporation for the day, by summing G and R_n cumulatively over the course of the day and then calculating the cumulative L_e for the day. This value of L_e was then divided by the latent heat of vaporisation of water to give the cumulative evaporation in mm for the day. G was found by measuring the temperature change in the soil and multiplying it by 357 $kJm^{-2}°C^{-1}$, a simpler method than using Equation 4.15. Blight (pers. comm.) indicated that this number may need some adjustment for different times of the day for a realistic answer. The method gave an almost identical evaporation of 3.5mm for 18 December 2000 in the degrassed area. Figure 4.2 shows the cumulative growth in R_n, G and L_e for both sites.

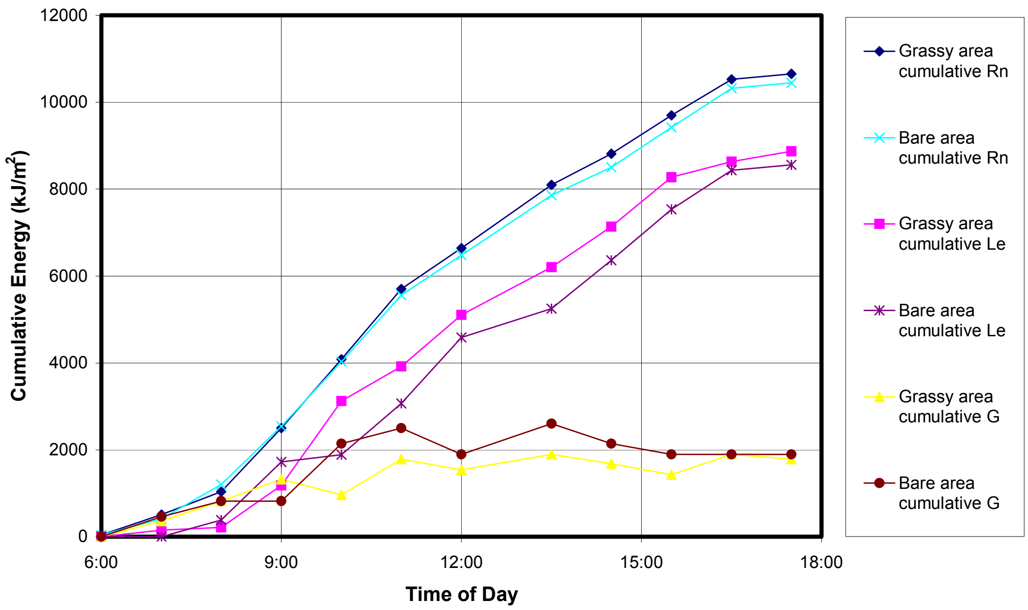

Figure 4.2: Graph showing cumulative growth in R_n, G and L_e over the course of the day 18 December 2000, after Blight (pers. comm.) in the grassy and the degrassed areas

Blight applied the surface energy balance method at three different places: Nylsvley, Clarens and at a landfill in Cape Town (Blight, 2002a). He verified his results at Clarens during periods of little or no rain when the change in soil water storage could be calculated. Agreement was reasonable in two of the cases but in the third, it badly underestimated evapotranspiration as calculated using the soil storage. He attributed this to the days when the evapotranspiration measured may not necessarily have been representative of average conditions for that time of year.

The evapotranspiration data derived from the energy balance given in Table 4.2 were collected at the Nylsvley Reserve over the course of individual, isolated days, therefore giving the evapotranspiration for those particular days only. Evaporation and evapotranspiration can vary quite significantly from one day to the next due to changes in cloud cover and temperature. Therefore, the evapotranspiration rates measured for those few days may not be representative of the average monthly rate, for the months in which these measurements were taken. Monthly mean daily evapotranspiration rates were needed to define evapotranspiration losses for the study area. Accurate monthly mean daily evapotranspiration rates are ideally obtained from long-term continuous measurements.

Evapotranspiration was measured in the wetland zone in a year when that zone was not inundated. The actual evapotranspiration in the inundated floodplain would differ from when it is not inundated (although the ground surface was still damp in this case) due to the differing albedos of soil and water. Blight (pers. comm.) asserts that the difference in evapotranspiration between the inundated floodplain and the floodplain that is not inundated but wet should be small. Blight (2002b) conducted an experiment with plant pots in a green house (which provided windless conditions). One pot was filled with soil and a growing grass cover, one was filled with only soil and one was filled with water only. The water surface evaporated at a constant 3.2mm/day, the bare soil surface at a rate of 4 mm/day initially and the grassy surface initially at a rate of 4.6 mm/day. The soil and grass surface evaporations dropped over the following days due to the depletion of water in the near surface soil. This shows that more evaporation can occur from a soil surface than from a water surface; and that transpiration by vegetation adds to evaporation, but the sum of the two (evapotranspiration) is not greatly in excess of evaporation. Thus on the Nylsvlei floodplain, evaporation from the open water surfaces may be less than that measured by Blight in the grassy and bare areas, but the presence of plants on the water surface may increase evaporation slightly.

4.3.2 Leaf area indices

Leaf area index (LAI) methods separate soil evaporation and plant evaporation components. The leaf area index is defined as the ratio of the one-sided leaf surface area to projected ground area beneath the canopy, giving a dimensionless number (Green *et al*, 2000). Relationships can be found between LAI and a range of ecological processes including photosynthesis, transpiration and evapotranspiration. These relationships are by empirical correlation. Thus, evapotranspiration can be estimated from the LAI of an area. LAI is difficult and destructive to measure physically as it requires the stripping of all leaves of the canopy in an area to determine the LAI or in some cases the collection of leaf litter from underneath the canopy (Green *et al*, 2000). However, developments in remote sensing have allowed LAI to be measured by satellites or aerial surveys which measure the fraction of photosynthetically active radiation (FPAR) absorbed by vegetation. For high values of LAI actual evapotranspiration may exceed potential evapotranspiration (Blight, 1992).

4.3.3 Study by Sanchez-Carrillo *et al* (2001) of evapotranspiration from a wetland

Actual data were obtained from measurements from plants in a study by Sanchez-Carrillo *et al* (2001) to determine transpiration losses from a wetland in Spain. A model was then devised that could predict transpiration from different plants at different times of the year. Transpiration was found as a function of the plant species and leaf area.

Total water transpired from the plant in a given time could be found and in turn, with information of plant density and leaf density for a unit area, transpiration could be predicted as a depth for a given area. It is not clear whether plants increase or decrease evapotranspiration in a wetland (Sanchez-Carrillo *et al*, 2001), whether transpiration exceeds the decreased evaporation due to leaf cover shading the water and sheltering of the water surface from the wind. This study showed that solar radiation, humidity and temperature control plant physiology and hence transpiration rates. Among the findings, it was determined that the reed species present in the wetland transpired at a maximum rate during the middle of the day. By contrast, the sedge and cattail species were found to transpire at a maximum during the early morning and late afternoon when it was cooler and they limited their transpiration in the middle of the day. It was also found that transpiration in summer was significantly higher than in autumn when they conducted tests again.

The model was able to estimate transpiration rates using mean daily values of solar radiation, humidity and temperature. Only having to know these three parameters was useful as it reduced the measurement and calculation complexity. They determined that transpiration had an important role in the water balance especially during dry periods. Transpiration was at times found to be up to three to four times that of evaporation for the same area. The evaporation data was taken from evaporation pans.

Unfortunately, the data and model are site specific and cannot be used for the Nylsvlei floodplain, as the vegetation and climate are different. However, this study showed that transpiration played a significant role in the evapotranspiration of the area and that transpiration varied with the time of year and the time of day.

4.4 Previous studies of evapotranspiration at Nylsvlei

According to Scholes and Walker (1993) annual evaporation is high on the Nylsvlei floodplain; on average evaporation from a standard A pan exceeds rainfall in every month of the year. This is due to the high annual solar radiation received: on average, the Nylsvley Reserve receives 75% of its potential annual total of 4371 sunshine hours. Evaporation was approximately 2400mm per annum using an unshielded class A pan, between 1975 and 1982 at the Nylsvley weather station, equivalent to 1870mm per annum for a Symons pan. The evaporation recorded at the Du Toits Kraal S pan (A6E005), the longest evaporation pan record available close to the reserve, is 1708mm/a for the period 1970/1971 to 2000/2001. The Symons pan evaporation for Nylsvley was 1770mm in the year 2000/2001, the year when the energy balance measurements were conducted (Blight, 2002a). This is slightly less than the average recorded by Scholes and Walker between 1975 and 1982 and slightly more than the long-term average for the Du Toits Kraal S pan. Scholes and Walker (1993) calculated that the evaporation from a hypothetical extensive water body at Nylsvley, as estimated by the Penman Equation, averaged 1897mm per annum.

4.4.1 Study of evapotranspiration on the floodplain at the Nylsvley Reserve using an energy balance

Blight (2001; 2002a; 2003) and Blight *et al* (2001) conducted evapotranspiration measurements using the energy balance method described previously at the Nylsvley Reserve. Equipment included thermocouples buried in the soil, radiometers to measure incoming and reflected radiation, a psychrometer to measure air temperature and humidity and an anemometer to measure wind speed. Measurements were started at sunrise and continued until sunset.

These measurements were conducted on certain days over the course of 1999, 2000 and 2001. The first three measurements were conducted on 24 August 1999 (not 24 August 2000 as stated by Blight (2002a) - (Blight, pers. comm.)), 20 October 2000 and 18 December 2000, when evapotranspiration was measured for both a grassy site (wild rice) and a bare soil site.

In August 1999, one sub-site was covered by vegetation, which was dormant at that time of year and the other had a bare soil surface and was adjacent to the water's edge, about 30m from the main stream channel. In October 2000, one sub-site was covered by vegetation starting to grow vigorously and the other was the bare soil surface of a warthog wallow. In December 2000, one sub-site was covered by vigorously growing vegetation and the other was similar but the vegetation was cut off just above the ground level to leave a surface covered by dead leaf litter, but devoid of any leaves or stems that could have transpired. See Figure 4.2 and Table 4.1 for the observations and calculation of evapotranspiration for that day. The grassy and bare sub-sites were selected to try to determine what proportion of evapotranspiration arises from evaporation from the soil and what proportion from transpiration from plants.

In August 1999, the water table was at a depth of about 50mm below the bare soil surface, and about 150mm beneath the vegetated surface. In October 2000, it was at a depth of approximately 50mm below the surface of both sites. In December 2000 the groundwater level was less than 150mm below the surface. The first three test days were calm so wind evaporation was negligible. In June 2001, the groundwater level was at a depth of 340mm. Blight (2002a) concluded that the groundwater level is likely to have been shallower than 700mm throughout the study period. For all of these months the soil evaporation was higher than the precipitation. Positive recharge to the soil (infiltration) only occurred between October and December 2000, for the remainder of the observation period recharge was negative with upward flow.

Blight (2002a) conducted more measurements on 6 February 2001, 27 March 2001, 6 July 2001 and 26 September 2001. The daily evapotranspiration values measured for each visit are tabulated in Table 4.2 and plotted in Figure 4.3. The daily time series of pan evaporations recorded at the Donkerpoort Dam S pan are shown for the same period in Figure 4.3 to give an indication of the seasonal trend and variability for the time of year of evaporation. The pan evaporations recorded at the Donkerpoort Dam S pan for the specific days when the measurements were conducted by Blight (also given in Table 4.2) are used later to determine average monthly values from periodic daily measurements at the Nylsvley Reserve.

Evapotranspiration measured at the grassy (wild rice) and bare sites peaked in mid-summer (February) and reached a minimum in mid-winter (July) as would be expected. The evapotranspiration from both the grassy and bare areas was lower than the corresponding Symons pan and Mini pan evaporations for all the measurement days. The evapotranspiration from the grassy and bare areas also plotted near the lower envelope of the daily S pan evaporations recorded at the Donkerpoort Dam for the same period (Figure 4.3). This suggests that shading of the ground by vegetation, the differing albedos of vegetation and soil compared to water and availability of water in the soil may limit the evapotranspiration rate on the floodplain.

Table 4.2: Evapotranspiration rates found using the energy balance method and Mini pan at the Nylsvley Reserve (Blight, 2001; Blight *et al,* 2001; Blight, 2002a) and the corresponding Symons pan evaporation for the same days at the Donkerpoort Dam S pan (A6E006)

Date	ET for grassy areas (mm/day)	ET for bare areas (mm/day)	M pan evaporation (equivalent S pan in brackets) (mm/day)	Donkerpoort Dam S pan (A6E006) (mm/day)	Weather conditions
24-Aug-99	0.5	0.7		5.0	Min temp: 8 deg C, max temp: 18 deg C, cloudy
20-Oct-00	2.6	2.0	3.5 (3.1)	7.0	Min temp: 13 deg C, max temp: 27deg C, afternoon cloudy with shower
18-Dec-00	3.6	3.5	6.0 (5.3)	6.0	Min temp: 14 deg C, max temp: 33 deg C, clear but partly cloudy in afternoon
6-Feb-01	4.6	3.7	6.0 (5.3)	6.0	Min temp: 16 deg C, max temp: 30 deg C, partly cloudy
27-Mar-01	3.5	3.0	4.0 (3.6)	7.0	Min temp: 16 deg C, max temp: 32 deg C, clear day
6-Jul-01	1.9	1.0	3.5 (3.1)	4.0	Max temp: 13 deg C
26-Sep-01	2.5	No Data		6.0	Max temp: 30 deg C

ET = evapotranspiration

The evapotranspiration from the grassy areas was greater than the evaporation from the bare areas except in August. This suggests that the evapotranspiration rate on the floodplain may be limited by shading of the ground by vegetation, vegetation albedo that varies with the time of year (highest in winter when the wild rice is dormant) and compared to the soil albedo, availability of water in the soil, and differing rates of transpiration in summer (when the wild rice is growing) and winter (when the wild rice is dormant).

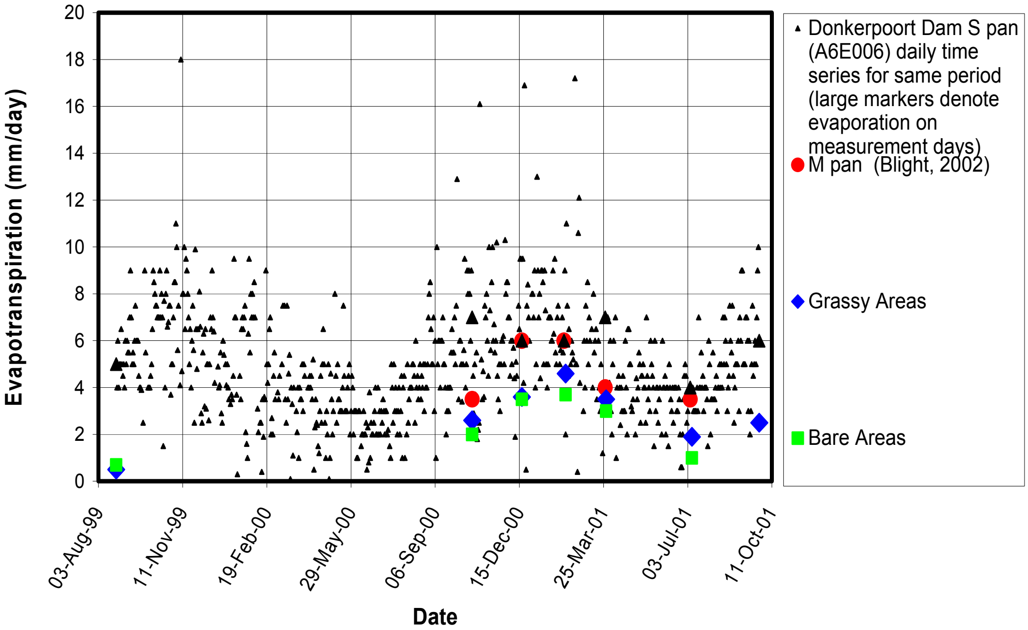

Figure 4.3: Evapotranspiration measured at the Nylsvley Reserve (Blight, 2002a) and from the S pan at the Donkerpoort Dam

Moore (1980), using lysimeters, reported for 4 to 10 December 1980 that areas with dead grass cover at the Nylsvley Reserve evapotranspired less water (3.8mm/day) than bare areas (4.1mm/day) and growing grass (4.9mm/day).

The annual evapotranspiration estimated by Blight (2002a), obtained by multiplying the ratio of the latent heat of evaporation (ΣL_e) to the net radiation (ΣR_n) for each measurement day by the cumulative pan evaporation for the periods between the measurement days taking into account the albedos of water and grass (Blight *et al*, 2001), was 920mm - 52% of the Symons pan evaporation of 1770mm for the study year. This agrees with the water balance produced by Morgan (1996). This water balance used the mean monthly-corrected evaporation of the A pans at Bela-Bela (formerly Warmbad) and Mokopane (formerly Potgietersrus) and it vastly overestimated the evaporation losses, the evaporation losses being larger than the floodplain storage. He attributed this over-prediction to the exaggerated inundation surface area calculated in his model and the artificial evaporation rate used. Morgan finally used an evaporation rate of 25% of the mean monthly calculated evaporation (based on the A pan evaporations at Bela-Bela and Mokopane).

4.4.2 Moore's study of evapotranspiration in the savanna at the Nylsvley Reserve

Moore (1980) calculated the total evaporation rate from 5 x 5 metre plots of savanna vegetation at Nylsvley (not on the floodplain) sealed from the surrounding soil by plastic barriers to a depth of 1.5m. By measuring the decline in soil water content over time using a calibrated "neutron probe" water meter at five depths and two locations in each plot, he found evapotranspiration to range from 1.6 to 5.3 mm/day (584mm/annum to 1934mm/annum). The evapotranspiration rate depended on the climatic conditions and to a lesser extent on the dominant species in the plot. The average over four habitat types (dominant species), three replicates and seven measuring periods was 3.6mm/day. This translates to 1314mm/annum. The evapotranspiration figure of 920mm/annum measured using the energy balance compares well with Moore's results, despite the difference in vegetation type and habitat.

The evapotranspiration results for the Burkea Savanna were compared with the predicted potential evapotranspiration for Nylsvley using various models. The Thornthwaite method gave a correlation coefficient of 0.9629, the Blaney-Criddle method a correlation coefficient of 0.9679 (using a crop factor derived for the savanna) and the class A pan using a vegetation factor, a correlation coefficient of 0.9932.

4.5 Average monthly evapotranspiration rates for the Nylsvlei floodplain

The evapotranspiration measurements collected using the energy balance method in 1999, 2000 and 2001 were measured on seven isolated days. These measurements are unlikely to be representative of the monthly mean daily evapotranspiration rates as evapotranspiration and evaporation can vary significantly from day to day due to changes in climatic conditions; such as air temperature, humidity, cloud cover and wind. The periodic evapotranspiration rates measured using the energy balance method were therefore used to find average values of daily evapotranspiration for each month of the year. This was achieved by comparing the evapotranspiration data for the grassy areas to the evaporation data from an evaporation pan nearby for the same days, and obtaining a pan factor to describe this relationship. The pan factor was then applied to the long-term monthly average of the daily pan evaporation to obtain an evapotranspiration rate for the floodplain for each month of the year. The grassy area (rather than the bare soil) evapotranspiration rate was chosen, as it is a better indicator of evapotranspiration from the inundated floodplain areas. Evapotranspiration rates were obtained by linear interpolation for the months when there were no energy balance measurements available. Various evaporation pans in the Nylsvley area were considered for this exercise and are tabulated in Table 4.3, with their record lengths and geographical positions.

Table 4.3: Evaporation pans considered for determining pan factors using the Nylsvley energy balance evapotranspiration data

Station	Pan type	Pan name	Latitude (DMS)	Longitude (DMS)	Data period
A2E006	Symons	Roodepoort at Warmbad Dam	24 51 30	28 14 45	1945/46 – 2001/02
A6E005	Symons	Du Toit's Kraal at Nylsvlei	24 34 00	28 46 00	1970/71 – 2000/01
A6E006	Symons	Donkerpoort at Donkerpoort Dam	24 40 15	28 19 15	1977/78 – 2001/02
B5E002	American	Amsterdam at Crecy	24 40 00	28 49 00	1956/57 – 1965/66
B5E003	Symons	Klavervalley at Klavervalley Dam	24 37 00	29 01 00	1969/70 – 1970/71
N/A	Symons	Nylsvley Reserve entrance gate	24 38 50	28 40 20	2000/01
N/A	Mini	Portable	24 38 50	28 41 30	2000/01

Station A2E006 was too far from the study site; stations A6E005, B5E002 and B5E003 had missing data for some or all of the days required in Table 4.2. The Nylsvley Reserve S pan had unreliable and missing data on certain days and were only available for 2000/2001 consequently having no long-term monthly average available; the M pan data were only available on the measurement days with no long-term monthly average available. The Symons pan data from the Donkerpoort Dam station (A6E006) were used, as it had data for all the energy balance measurement days (Table 4.2) and it had a long-term monthly average of the daily evaporation (23 years). Unfortunately, the Donkerpoort Dam S pan is some 35km from the study site.

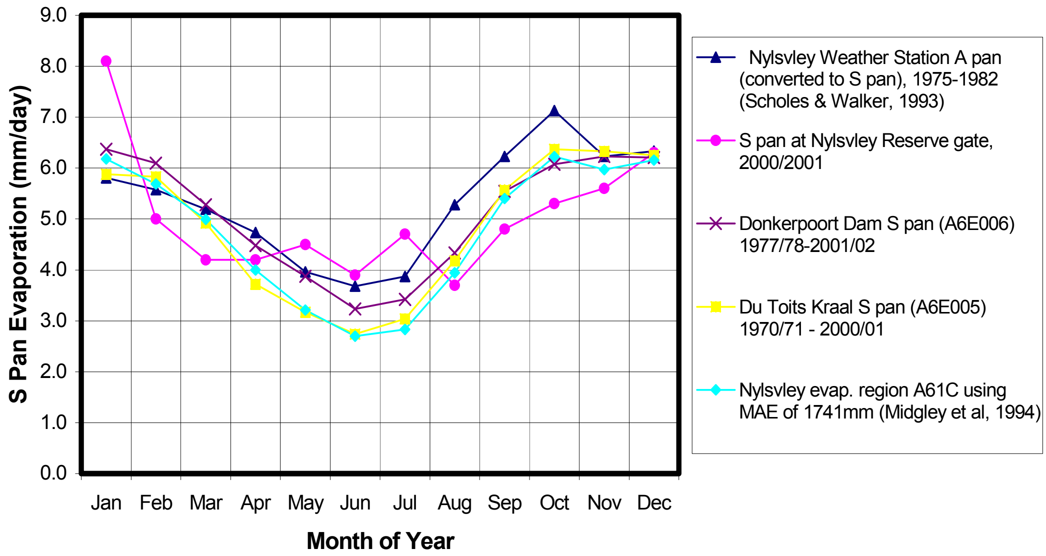

Figure 4.4: Comparison between pan data from Donkerpoort Dam and stations located near the study area of the Nylsvlei floodplain

The average monthly pan evaporation data expressed as S pan evaporation data for some of the stations in the study area of the Nylsvlei floodplain, together with the Donkerpoort Dam station, are shown in Figure 4.4. A pan data from Scholes and Walker (1993) were converted to S pan data using Equation 4.12. Data for evaporation region A61C (Midgley *et al*, 1994) are also plotted. Agreement between the data is reasonable except for the S pan located at the Nylsvley Reserve gate - due to the short record period of only one year. The data of Scholes and Walker (1993) are higher for June to October possibly due to the shorter record period (June 1975 to June 1982). S pan data given by Midgley *et al* (1994) for quaternary subcatchment A61C (calculated using the percentage of MAE given for each month for evaporation zone 1C) are based on pan data from: Doorndraai Dam S pan (A6E001), Potgietersrus A pan, Sterkriver A pan (A6E003), Du Toits Kraal S pan (A6E005) and Donkerpoortdam S pan (A6E006). The data given by Midgley *et al* (1994) agree well with the Du Toit's Kraal S pan data. Figure 4.4 shows that the evaporation data from the Donkerpoort Dam station compare well with data from stations near the study area of the Nylsvlei floodplain (despite its distance from the study area and the other stations) providing a reasonable estimate of average monthly pan evaporation from the floodplain.

Table 4.4 gives estimates of evapotranspiration from the Nylsvlei floodplain using pan evaporation recorded at Donkerpoort Dam to convert the periodic daily energy balance measurements to average daily rates for each month.

Table 4.4: Derivation of the average monthly evapotranspiration rates from pan data for the Donkerpoort Dam station and the periodic energy balance measurements in the grassy area at the Nylsvley Reserve

Month	Total daily evaporation or evapotranspiration (mm)		S Pan factor (A/B)	Average monthly evaporation or evapotranspiration (mm)	
	Nylsvley energy balance measurements (A)	S Pan at Donkerpoort (B)		S Pan at Donkerpoort Dam (C)	Nylsvley (factored S Pan) (A/B x C)
Oct	2.6	7.0	0.37	6.1	2.3
Nov			0.49*	6.2	3.0
Dec	3.6	6.0	0.60	6.2	3.7
Jan			0.68*	6.4	4.4
Feb	4.6	6.0	0.77	6.1	4.7
Mar	3.5	7.0	0.50	5.3	2.6
Apr			0.49*	4.5	2.2
May			0.49*	3.9	1.9
Jun			0.49*	3.2	1.6
Jul	1.9	4.0	0.48	3.4	1.6
Aug	0.5	5.0	0.10	4.3	0.4
Sep	2.5	6.0	0.42	5.5	2.3

* denotes interpolated pan factors

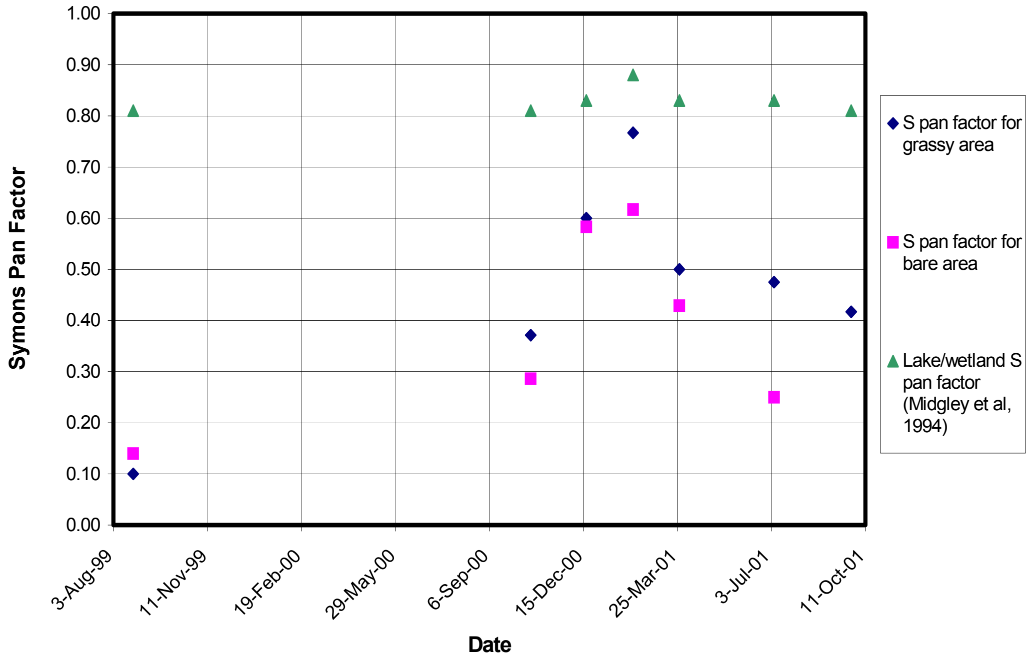

Figure 4.5: Symons pan factors derived from energy balance evapotranspiration data for grassy and bare areas and the Symons pan evaporation data for the same days at the Donkerpoort Dam Symons pan (A6E006), and the recommended Symons pan factor for lakes and wetlands in South Africa from Midgley *et al* (1994)

Figure 4.5 shows Symons pan factors derived using the energy balance measurements for grassy and bare areas (Table 4.2) at the Nylsvley Reserve and the evaporation recorded at the Donkerpoort S pan (A6E006) for the same day (as in Table 4.4), and the recommended Symons pan factors for lakes and wetlands in South Africa from Midgley *et al* (1994). Figure 4.5 shows that evapotranspiration and evaporation from the grassy and bare areas respectively is always lower than the Symons pan evaporation, especially so in winter. The S pan factors range from 0.1 in winter to 0.77 in summer. This indicates the effects of growth response and the limiting influence of water availability on the floodplain, discussed later. The recommended S pan factors for lakes and wetlands in South Africa (Midgley *et al*, 1994) are more constant and higher (>0.8) throughout the year than the S pan factors for the Nylsvlei floodplain. The generalised lake and wetland factors are for the whole of South Africa (not specific to the Nylsvlei floodplain) and are for open water surfaces, whereas the energy balance measurements were conducted on a damp vegetated floodplain.

4.6 Comparison of evapotranspiration rates derived from energy balance measurements at the Nylsvley Reserve with some empirical and pan methods

4.6.1 Empirical methods

The evapotranspiration rates derived from the energy balance measurements at the Nylsvley Reserve (Table 4.4) were compared with the predictions of some of the empirical evapotranspiration models discussed in 4.2.1. These are plotted in Figure 4.6 together with class A pan evaporation data and “equilibrium evapotranspiration” (based on class A pan evaporation data representing evapotranspiration over a large area of savanna), both for the Nylsvley Reserve, from Scholes and Walker (1993). The empirical models were chosen for their relative simplicity and modest data requirements (temperature, radiation at the edge of the Earth’s atmosphere and day length). The micrometeorological methods such as the Penman method and the FAO Penman-Monteith method were not used, as they required far more data that were not readily available. The empirical models used were: Thornthwaite, 1948 (Temperature Method); Blaney and Criddle, 1950 (Temperature Method); Jensen and Haise, 1963 (Radiation and Temperature Method); Linacre, 1977 (Temperature Method) and Hargreaves, 1994 (Radiation and Temperature Method).

All the empirical prediction models display smooth trends over the seasons, with maximum evapotranspiration rates predicted for December/January when the radiation received at the edge of the Earth’s atmosphere and the average temperature are at a maximum. The Thornthwaite (1948) and Jensen and Haise (1963) models compared best with the factored evapotranspiration rates derived from the energy balance measurements. The greater variability of the energy balance estimates - particularly in February and August - may be attributed to the life-cycle (and transpiration) responses of the wild rice (the dominant grass species in the area where the measurements were conducted, which evapotranspires at a maximum in summer and is dormant at the end of winter (Marneweck, 2003)), limiting water availability on the floodplain at certain times of year and the small number of periodic evapotranspiration measurements (seven points - Table 4.2).

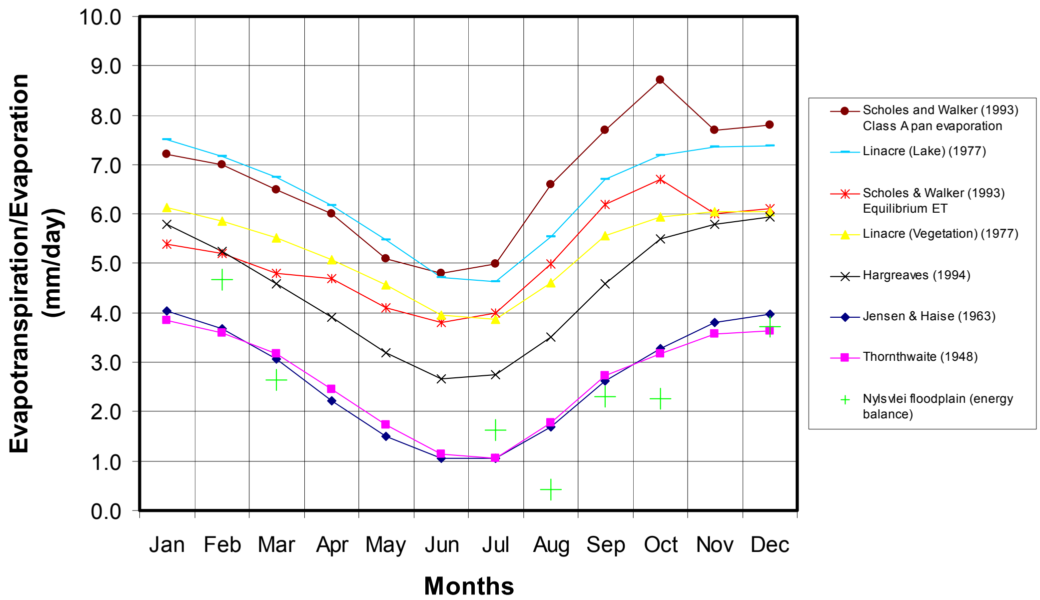

Figure 4.6: Comparison of the evapotranspiration derived from the energy balance measurements with various empirical model predictions and data based on class A pan evaporation from Scholes and Walker (1993)

4.6.2 Pan methods

The evapotranspiration rates derived from the energy balance measurements at the Nylsvley Reserve were also compared with the predictions of some vegetation pan factors (discussed in 4.2.3) applied to the average monthly pan evaporation for the Donkerpoort Dam gauge (A6E006) and are shown in Figure 4.7. This gauge was chosen as it was used to derive the factors and the monthly average daily evapotranspiration rates for the Nylsvlei floodplain from the energy balance measurements. The pan factors used are recommended by Midgley *et al* (1994) for South Africa and included the S pan factor for a lake or wetland, the A pan factor for true grassveld, the A pan factor for false grassveld, the A pan factor for tropical bush and savanna (bushveld) and the A pan factor for false bushveld. These pan factors vary for every month of the year and reach a maximum in summer and a minimum in winter. The Donkerpoort Dam S pan evaporation data were converted to A pan evaporation data using equation 4.13.

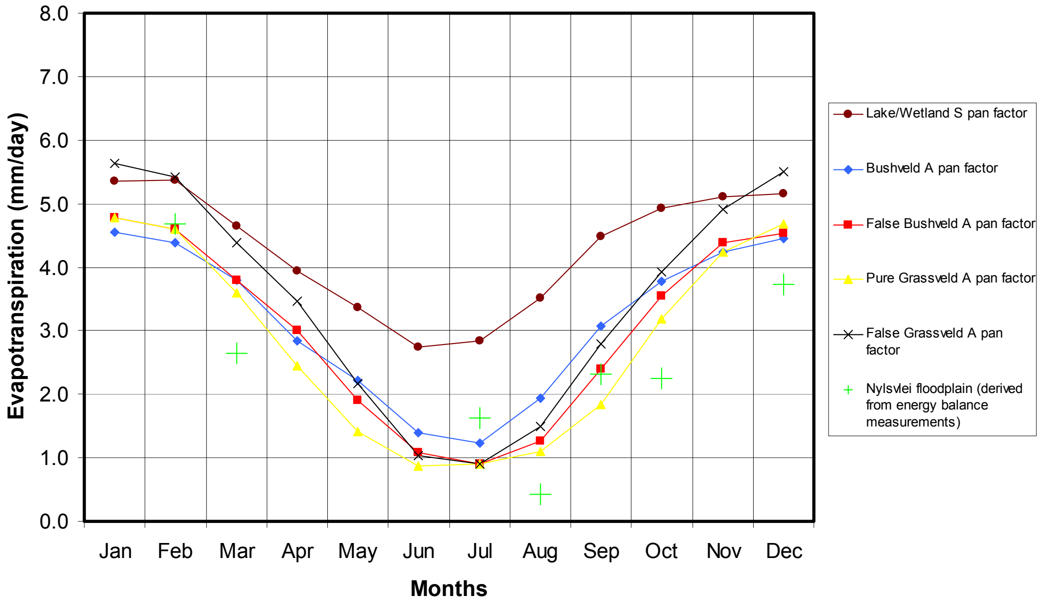

Figure 4.7: Evapotranspiration rates predicted using various pan factors and the rates derived from the energy balance measurements applied to the Donkerpoort Dam station pan data

The lake/wetland factors, despite their name, generally predict higher evapotranspiration rates throughout the year (especially in winter) than the other pan factor methods and the data derived from the energy balance measurements. The four vegetation pan factors (which are similar) agree quite well with the data derived from the energy balance measurements except from October to December, March and August where differences of up to 2mm/day occur. Overall, the Pure Grassveld A pan factor appears to agree the best with the data derived from the energy balance measurements.

4.7 Recommended evapotranspiration rates for the Nylsvlei floodplain

The evapotranspiration rates derived from the energy balance for the grassy areas were used in the hydraulic model to represent evapotranspiration losses from inundated areas of the Nylsvlei floodplain. These rates given in Table 4.5 are the best estimates of evapotranspiration rates for the Nylsvlei floodplain as they are based on measured data.

Table 4.5: Average monthly evapotranspiration rates used in the hydraulic model for inundated areas of the Nylsvlei floodplain

Evapotranspiration (mm/day)											
Oct	**Nov**	**Dec**	**Jan**	**Feb**	**Mar**	**Apr**	**May**	**Jun**	**Jul**	**Aug**	**Sep**
2.3	3.0	3.7	4.4	4.7	2.6	2.2	1.9	1.6	1.6	0.4	2.3

Reasonable agreement between these rates (Table 4.5) and the predictions from in particular the empirical methods of Jensen and Haise (1963) and Thornthwaite (1948) (discussed earlier) and the Pure Grassveld A pan factor can be noted, which supports the use of these particular methods in this wetland environment when no measured data are available.

4.8 Summary

Evapotranspiration measurements (using the energy balance method) on the Nylsvlei floodplain by Blight (2002a) and taken on individual days over the course of 1999, 2000 and 2001, allowed a monthly average daily evapotranspiration rate to be derived for each month of the year for the floodplain. This was achieved by deriving a pan factor using the measured evapotranspiration rates (from the energy balance) and comparing these to evaporation pan data from the Donkerpoort Dam (relatively nearby) on the same days. These pan factors were applied to the long-term monthly average of the daily pan evaporation data for the Donkerpoort Dam station, to find these desired evapotranspiration rates. Reasonable agreement was found between these data derived from the energy balance measurements and evapotranspiration rates predicted by the Jensen and Haise (1963) and Thornthwaite (1948) empirical methods and by the Pure Grassveld A pan factor, supporting the use of these particular methods in this wetland environment when no measured data are available. The other potentially significant loss to floodwaters which would affect the hydraulic behaviour of the floodplain and hence needs to be included in the model is infiltration. The investigation of this loss is described in Chapter 5.