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Estimation of Open Water Evaporation

A Review of Methods

R&D Technical Report W6-043/TR

Environment Agency

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This report reviews available methods for estimating evaporation from open water and recommends methods to be used by the Environment Agency.

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Professor Mike Depledge Head of Science

EXECUTIVE SUMMARY

Estimates of evaporation from open water are increasingly required for several Environment Agency functions, particularly Water Resources and Ecology. These estimates are used mainly for water balance studies to support appraisals of applications for abstraction licences, in wetlands and still waters management. Current methods of estimating open water evaporation vary between and, in some cases, within Regions; there is no generally adopted best method. In addition, there is often a mismatch between the accuracy of estimates produced by current methods (they are generally crude and subject to large uncertainties) and their significance in the calculations that are used as a basis for decision making.

There is therefore a need to review the Agency's requirements for estimating open water evaporation in relation to the methods and data available. The objectives of this project are:

- evaluate current methods of estimating open water evaporation;
- recommend the best available practicable methodologies for producing robust estimates;
- assess the associated uncertainty of these methodologies

The project consists of two Phases. Phase 1 essentially dealt with the first two objectives and made recommendations for more quantified assessments of uncertainty that were carried out under Phase 2.

A survey of Agency Regional staff in summer 1999 established that the Agency currently requires estimates of open water evaporation for three purposes: abstraction licensing, water balance studies and management of wetlands. Empirical factors applied to estimates of potential evaporation are used to produce these estimates in all Areas and Regions. However, the values of the factors and data sets of potential evaporation employed vary between Regions and may vary between Areas. Thus, although there is a consensus on the method, there is little on how it is operated. Significant errors, up to 30%, can occur when using the empirical factors method for estimating evaporation if inappropriate factors are used, e.g. if the empirical factors calibrated on values of potential evaporation calculated using one model are applied to values calculated using a different model.

The review of methods of estimating open water evaporation identified seven methods; pan evaporation, mass balance, energy budget models, bulk transfer models, combination models, the equilibrium temperature approach and empirical factors. A ranking of the seven methods of estimating open water evaporation against nine criteria (covering accuracy, robustness, ease of use etc.) established that the equilibrium temperature approach would best serve the Agency's purposes. The use of empirical factors and the combination models were ranked equal second.

The spatial variability of the meteorological variables that drive evaporation is strongly influenced by proximity to the coast. Inland, the spatial variability of wind speed and air temperature is low whilst incoming solar radiation and relative humidity show significantly more spatial variability. Topography has a strong effect on the driving variables, either directly, in terms of the lapse rates of air temperature and vapour pressure, or indirectly, through the formation of clouds affecting the amount of incoming solar radiation.

Correcting for the effect of altitude on the driving variables can be achieved except for the effect of cloudiness on incoming solar radiation. Empirical corrections to evaporation estimates for altitude have also been found, and although this is not as physically rigorous as correcting the driving variables, the analysis indicated that the latter performed better.

The size of the water body affects evaporation rates in two ways. Firstly, there is evidence that, for water surfaces greater than 10 m in diameter, the rate of evaporation over water is enhanced due to increased wind speed resulting from the smoother surface. However, no means of taking these effects into account has been found. Secondly, the size of the water body in England and Wales affects the development of thermal stratification. The maximum depth of the warmer surface layer is a function of the surface area of the water body. This can have a significant effect on open water evaporation.

The methods recommended for use by Agency staff (equilibrium temperature and empirical factors) have been tested against measurements of evaporation made at Kempton Park between 1956 and 1962. The combination models of Penman (1948) and Penman-Monteith were included for the purposes of comparison. The results show that the estimates of open water evaporation made by the equilibrium temperature model are in excellent agreement with the measured values. Using the empirical factors of Penman (1948) with MORECS grass PE, reasonable agreement was achieved. Other methods performed poorly.

The sensitivity of the estimates of open water evaporation made using the equilibrium temperature method to the parameters required has been carried out and values for these parameters have been recommended. Comparison between measured evaporation at Kempton Park reservoirs and the values predicted by the equilibrium temperature model suggest an accuracy of ± 2 % can be achieved for annual values and ± 30 % for monthly values.

New empirical factors, for use with MORECS grass PE, PENSE and PETCALC, have been calculated from the measured values at Kempton Park. The effect of different climate types in England and Wales on the accuracy of estimates of open water evaporation made using these factors has been investigated. The results show that estimates of annual evaporation based on MORECS grass PE and PENSE should be regarded as having an accuracy of ±15 %, whilst those based on PETCALC have an accuracy of ±25 %. The accuracy in estimates of monthly evaporation have a strong seasonality with an accuracy of ±50 % in the summer months (July to October incl.) and ±200 % during the winter months.

The accuracy of corrections for differences in altitude, between the meteorological station and the site for which an estimate of open water evaporation is to be made, have been tested. It has been shown that an accuracy of between ±5 and ±30 % in annual estimates of open water can be achieved using data from a station in the same climate region and correcting the driving variables. However, there is a strong seasonal difference in the accuracy on estimates of monthly open water evaporation. Correcting estimates of evaporation directly results in the error being decreased by between 30 % and 50 % from no altitudinal correction being applied.

It is recommended that the Agency should, in the long term, adopt the equilibrium temperature model as its method of estimating open water evaporation. In the meantime, it should make use of the empirical factors developed by this project to be applied to MORECS grass PE, PENSE and PETCALC. Practical guidance in the use of the empirical factors and equilibrium temperature methods by Agency staff has been provided in a handbook

"Estimating open water evaporation – guidance for Environment Agency practitioners" produced as part of the output from this R&D project.

Further research is recommended into determining values for the parameters required by the equilibrium temperature model, appropriate lapse rates for the driving variables and how thermal stratification of water bodies affects evaporation rates.

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1. INTRODUCTION

Estimates of evaporation from open water are increasingly required for several Environment Agency functions, particularly Water Resources and Ecology. These estimates are used mainly for water balance studies to support appraisals of applications for abstraction licences, in wetlands and still waters management, and will increasingly be used in modelling work in future. The demand for such estimates is likely to grow in the immediate future driven by the Abstraction Licence Review, the Habitats Directive and the Regional Still Waters Group (in the North West Region). The result will be that more robust estimates than at present will be required.

Current methods of estimating open water evaporation vary between, and in some cases within, Regions; there is no generally adopted best method. In addition, there is often a mismatch between the accuracy of estimates produced by current methods (they are generally crude and subject to large uncertainties) and their significance in the calculations that are used as a basis for decision making. It is important, and should be possible, to assess this uncertainty in numerical terms, allowing it to be taken into account in the decision making.

In addition to the uncertainty associated with the methodologies used for estimating open water evaporation, there is considerable uncertainty over the importance (to heat storage and hence evaporation rates) of the dimensions of the water body. The Agency is concerned with water bodies covering a very wide range of dimensions and there is a need to establish the nature and significance of this influence.

To achieve these aims the objectives of this project are:

- evaluate current methods of estimating open water evaporation;
- recommend the best available practicable methodologies for producing robust estimates;
- assess the associated uncertainty of these methodologies.

The project consists of two Phases. Phase 1 essentially deals with the first two objectives although it does review the literature relevant to the third. Recommendations made for more quantified assessments of uncertainty were carried out under Phase 2.

Chapters 2 to 9 are essentially concerned with Phase 1. Chapter 2 gives an introduction to the process of evaporation from open water and the factors that affect the rate of evaporation. Methods of estimating open water evaporation are reviewed in Chapter 3 which also discusses the source and size of errors involved with each method. The methods of estimating open water evaporation currently used by the Agency are given in Chapter 4, as is the availability of data. The methods of estimating open water evaporation are ranked in Chapter 5 and recommendations made as to which should be adopted by the Environment Agency. The spatial and temporal variability of evaporation are discussed in Chapter 6. Data sets potentially available for testing methodologies are described in Chapter 7. Chapter 8 presents the conclusions and recommendations of Phase 1.

Chapter 9 provides an introduction to Phase 2. The methods of estimating open water evaporation that have been tested under Phase 2 are described in Chapter 10. The absolute accuracy of estimates of open water made using these methods is discussed in Chapter 11. New empirical factors for use with PE data sets are described in Chapter 12. The sensitivity of estimates of open water evaporation made with the equilibrium temperature model to input parameter values is investigated in Chapter 13. In Chapter 14, the effect of differences in climate on the accuracy of estimates of open water evaporation made using the empirical factors method is discussed. The accuracy of corrections for differences in altitude between a meteorological site and the site where the estimate of open water evaporation is required is investigated in Chapter 15. (It should be noted that there is little information in the literature on the issues raised in these two chapters; indicating that there have been few previous studies.) Chapter 16 presents the conclusions and recommendations of the project.

PHASE 1

2. EVAPORATION FROM OPEN WATER

This chapter provides an introduction to the process of evaporation from open water and describes the factors that may affect the rate of evaporation. It also defines the terms and concepts that are used in the rest of the report.

Evaporation occurs when liquid water is converted into water vapour. The rate is controlled by the availability of energy at the water surface, and the ease with which water vapour can mix into the atmosphere.

The molecules comprising a given mass of water are in constant motion. A molecule must have a minimum energy if it is to leave the surface of the water and the number of such molecules is related to the surface temperature. Thus, adding heat to the water body raises the energy of the molecules (and hence the temperature) with the result that more molecules leave the surface. Sources of heat include solar (short-wave) radiation, thermal (long-wave) radiation and inflows of water into the body of water. The water vapour molecules in the lower portion of the overlying air are also in motion, and some of these will strike the water surface to either rebound or be captured. The capture rate of the water vapour molecules is proportional to their rate of collision with the surface and therefore to the vapour pressure adjacent to the water surface. Thus, evaporation is the difference between two rates, a vaporisation rate determined by the temperature of the water, and a condensation rate determined by vapour pressure.

If the air above the water body were still, then the movement of water molecules from the water surface into the air would lead to the saturation of the lowest portion of the air, with the result that net evaporation would cease. Normally, mixing processes, such as turbulence and convection, act to combine the air near the water surface with the drier air overlying it, allowing evaporation to continue. The stronger the wind, the more vigorous and effective will be the turbulent mixing. The greater the difference between the temperatures of the surface and overlying air, the greater the convective mixing.

2.1 Meteorological factors affecting evaporation

2.1.1 Net radiation

The amount of radiant energy captured by the water body (net radiation) is generally the dominant control on the annual evaporation rates. The radiation balance is illustrated in Figure 2.1

A portion, S_t, of the solar radiation incident at the top of the atmosphere, S_0, reaches the water body directly. Some reaches the water body in a diffuse form, S_d, after scattering by atmospheric particles and clouds. Part of the solar radiation reaching the water body, S_d+S_t, is reflected. The reflection coefficient, or short-wave albedo, α_s, depends on transient factors, such as the direction of the solar beam, but a value of 0.08 is generally used for water. This implies that 92% of the solar radiation striking the surface of the water body is absorbed. However, the radiation is absorbed throughout a thickness of the water column that is dependent on the wavelength of the radiation. Thus blue light may penetrate a column of clear water for tens of metres whilst near-infrared radiation is absorbed in less than a metre. The net short-wave radiation, S_n, is that portion of the incident solar radiation captured by the water body, taking into account losses due to reflection.

There is a significant exchange of radiant energy between the water body surface and the atmosphere in the form of long-wave (thermal) radiation, the incoming radiation, $L\downarrow$, and outgoing radiation, $L\uparrow$. The dominant source of the incoming radiation is the atmosphere and is mainly due to short wave radiation being absorbed and re-emitted at longer wavelengths. The outgoing radiation is that emitted from the surface of the water body.

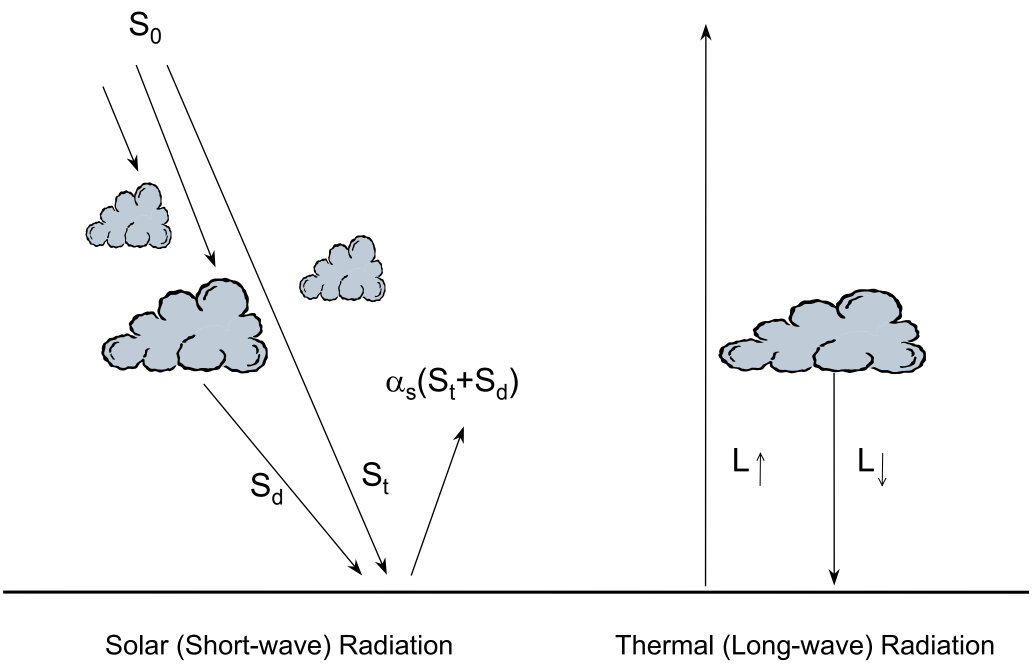

Figure 2.1 Radiation balance at the water surface.

The net radiation, R_n, is the net input of radiation at the surface, i.e. the difference between the incoming and reflected solar radiation, plus the difference between the incoming and outgoing long-wave radiation:

$$R_n = (1 - \alpha_s)(S_d + S_t) + (L_\uparrow - L_\downarrow) \tag{1}$$

2.1.2 Humidity

Absolute humidity varies only slightly throughout the day, except with a change of air mass. However, relative humidity is more variable and, as the relative humidity of the air over the water surface increases, so the net transfer of water molecules from the surface is reduced.

Relative humidity increases as the temperature of the air decreases with the result that the evaporation rate may decrease.

2.1.3 Diffusion processes

Diffusion of water molecules from an open water surface into completely still air will reduce as the air adjacent to the water approaches saturation. Air movement is necessary to remove the air in contact with the water surface and mix it with the drier air above. Hence, the rate of evaporation is generally influenced to some extent by turbulent air movement. The degree of turbulence is strongly related to the wind speed and surface roughness. This frictional air turbulence may be enhanced by convective turbulence if there is an appropriate gradient in mean air temperature away from the water surface. The positive relationship between wind speed and evaporation only holds up to a critical value. Thus, the energy available and the humidity determine the maximum evaporation rate.

2.2 Properties of the water body affecting evaporation

2.2.1 Water depth

The effect of the water depth on the seasonal distribution of evaporation can be considerable as a result of the heat storage capacity of the water body which is, to a large extent, determined by its depth. In higher latitudes (such as the UK), where there is a strong seasonal variation in the sun's elevation at noon, the increasing incoming solar radiation serves to warm the water body during the spring and early summer. During the autumn and early winter, as the incoming solar radiation decreases, the water body cools as the stored energy is released. The result is that the evaporation rate can be de-coupled in time from the net radiation. It is generally considered that the effect can be ignored for water bodies with a depth less than 0.5 m. and that the effect reaches a maximum (i.e. the seasonal evaporation ceases to change) once the depth increases beyond 4.5 m (because little of the incoming solar radiation penetrates below this depth) .

The effect of the water depth on the annual total evaporation is likely to be small since the seasonal variations tend to cancel out from one year to another, i.e. the annual energy input is approximately constant.

2.2.2 Thermal stratification

This is a function of the surface area of the water body. Stratification occurs in large, deep water bodies (at mid and high latitudes) and may accentuate the time lag between the net radiation and the evaporation rate. The temperature dependence of the density of water is a key factor (the maximum density occurs at a temperature of 4^{o} C).

During the early spring, most large, temperate water bodies exhibit a nearly uniform temperature distribution with depth (homothermal conditions). As the year progresses and the weather warms up, the water body receives heat at an increasingly rapid rate. Initially, the water body remains homothermal because the heat that is received at the surface layers is transported to deeper layers by wind-induced currents and turbulence. As the rate of heating continues to increase, it begins to exceed the rate of transfer to deeper layers with the result that the temperature of the surface layers increases faster than those of the deeper layers. As the heating continues, a point of inflection develops in the temperature depth profile and a well-mixed upper layer (the epilimnion), with relatively intense gradients at its bottom boundary is formed. The plane of maximum temperature gradient is known as the thermocline. During the remainder of the heating period, the thermocline slowly descends into the lake. Once a thermocline has formed, the deeper regions of the lake are relatively uninfluenced by changes in surface conditions. The maximum thickness of the epilimnion is dominantly a function of the surface area of the water body and the climate.

In the autumn, after the water body has attained its maximum heat content, the thermocline moves down rapidly into the deeper layers of the lake, often referred to as turnover. This is because the wind mixing is augmented by convective mixing due to surface cooling (resulting in an increase in density so that the water sinks). The thermocline continues to move down rapidly as the well-mixed upper layers cool further, until the whole water body again attains homothermal conditions.

A ‘reverse’ stratification can be created in winter, especially in continental climates, but the cool layer is much thinner than the epilimnion of summer. Sufficient cooling may permit the water body to freeze over, whilst retaining the temperature of the deeper water in the range of 2-4ºC.

If the minimum (winter) temperature of the water body is greater than 4ºC then there is only one turnover (in the autumn). Large water bodies that are rarely stratified are generally tropical with high temperatures.

The net result of the heat storage is that water temperatures are lower than air temperatures during the summer and vice versa during the winter. Thus, the evaporation rates from large deep water bodies may be higher in winter than in summer.

2.2.3 Size of surface

With constant wind speed, the evaporation rate is related to the size of the water body surface and to the relative humidity. As a parcel of air moves across a large water body, there will be a decrease in the rate of evaporation as the ‘vapour blanket’ increases in thickness. The larger the water body, the greater will be the total reduction in the depth of water evaporated. If the body of water is large enough (e.g. the oceans) the humidity of the air will be largely independent of the distance it has travelled from the edge and hence the evaporation will be closely related to the amount of energy available. On the other hand, small water bodies, such as evaporation pans, exert little influence on the temperature or humidity of the overlying air and so a continuous high rate of evaporation is maintained.

The magnitude of differences in evaporation rates from water bodies of different sizes will be considerably affected by the humidity of the incoming air. If this is initially high then it can only be modified slightly and thus the difference in evaporation rates will be small.

The wind speed tends to increase away from the edge of the water body in response to the smoother surface presented by the water surface. This results in an increase in the rate of evaporation which tends to counter the reduction in the rate of evaporation resulting from the decrease in relative humidity.

2.2.4 Rainfall

Rainfall directly on to the water body may change the heat content although this is often considered sufficiently minor to be ignored because the volume of rainfall is generally small compared to that of the water body, as is the temperature difference.

2.2.5 Inflow and outflow

Water flowing into and out of the water body may have a volume comparable to that of the water body and, if the temperature of the inflow is significantly different to that of the water body, there may be a significant contribution to the heat stored in the water body. An example of this is gravel pits with a significant throughflow of groundwater.

2.2.6 Vegetation

The presence of vegetation protruding from the water may affect the evaporation rate from the water body. Firstly, the vegetation canopy may shade the water and thus reduce the amount of heat gained by the water body. Secondly, it may change the aerodynamic roughness of the surface, either increasing it in the case of a sparse canopy or decreasing it in the case of a dense canopy.

Transpiration from vegetation may result in loss of water from the water body provided that the water supplying the roots can be considered part of the water body (e.g. in the case of floating vegetation). The energy used to drive the transpiration is then not available for evaporation from the body of water and so this may result in a reduction in the rate of water loss due to the higher albedo of the vegetation.

The vegetation on the land surrounding the water body may have an effect locally around the edge by changing the roughness. For example, the edge of a forest will result in increased turbulence of offshore winds compared to grass. This effect will be quite local (a few tens of metres) and so its impact on the total evaporation will be dependent on the size of the water body.

2.2.7 Turbidity and bottom reflectance

The presence of suspended particulate matter may increase the short wave albedo of the water body. A turbid water body will reflect more of the incoming solar radiation with the result that evaporation will be reduced. The effect of this may be large as hyper-turbid lakes can have albedos as high as 0.2, compared to 0.08 for clear water.

For shallow water bodies it is possible that the albedo of the bottom of the water body may affect the amount of heat in the water. If the bottom has a high albedo then more of the solar radiation will be reflected. A low albedo will result in more solar radiation being absorbed by the substrate which may then warm the water by conduction.

2.2.8 Salinity

The evaporation rate decreases by about 1 per cent for each 1 per cent increase in salinity because of the reduced vapour pressure of the saline water. This effect is normally small enough to be discounted when estimating evaporation rates from 'fresh' water bodies.

3. METHODS FOR ESTIMATING OPEN WATER EVAPORATION

A wide variety of methods for estimating open water evaporation have been reported in the literature. They can be categorised into seven types; pan evaporation, mass balance, energy budget models, bulk transfer models, combination models, the equilibrium temperature method and empirical factors. This chapter reviews the methods and discusses their relevance to the Environment Agency's objectives. It does not include all possible methods as it excludes methods that are clearly impractical (e.g. micrometeorological measurements, Bowen ratio or eddy correlation – see Brutsaert, 1982). A list of the symbols used is given in Annexe A.

3.1 Pan evaporation

The use of pans of water for measuring evaporation dates back to the 18th century. It is easy to understand their intuitive appeal as they measure open water evaporation in a visible way. However, despite numerous studies, it is very difficult to use data from pans except in specific circumstances. Hounam (1973) carried out a review of methods for estimating lake evaporation from measurements of pan evaporation and much of the following is drawn from this source.

3.1.1 Measurements

The pan adopted by the Meteorological Office (sometimes known as the Symons pan) is a galvanized iron tank 1.83 m square and 0.61 m deep. It is set in the ground with the rim 100 mm above the ground and the water level is kept near to the ground level. The interior is painted black. Measurements are made daily with a hook gauge attached to a vernier scale. Any rainfall recorded in the previous 24 hours must be allowed for.

A pan that has found wide use around the world is the US Class A pan. This is a circular galvanized iron tank with a diameter of 1.21 m and is 255 mm deep. It is mounted on an open wood frame so that air circulates round and under the pan. The water level is kept about 50 mm below the rim. The level is measured daily using a hook gauge. Again any rainfall must be allowed for. In the standard setup, a thermometer measures the water temperature and a three cup anemometer measures the wind speed 150 mm above the pan rim. The interior is generally painted black.

Another pan that has found world wide use is the USSR GGI-3000 pan. This is a cylindrical tank with a diameter of 0.618 m and is 0.6 m deep at the walls and 0.685 m deep at the centre. The pan is sunk in the ground with the rim approximately 75 mm above the surface. It is painted white.

3.1.2 Sources of measurement errors

The evaporation from a pan will be enhanced if it is surrounded by a dry surface. This is called the Oasis Effect. Energy from the surrounding surface will be transferred horizontally from the dry surface and provides extra energy for evaporation of water in the pan. Thus a pan will have a higher evaporation than a water body. In addition, specific pans differ due to their different constructions.

The US Class A pan suffers from the disadvantage that the sides are exposed to the sun with the result that it reaches a higher temperature (and consequently increased evaporation) than pans sunk in the ground. Conversely, sunken pans can sometimes overestimate evaporation due to heat transfer from the surrounding soil. Leakage is also much more difficult to detect in sunken pans.

Pans sunk in the ground are vulnerable to splash in and splash out. In times of hot weather, wildlife may use the pans as sources of drinking water. Attempts to overcome this by covering the pans with mesh have resulted in significant reductions in evaporation.

Despite their apparent simplicity, all pans need to be carefully maintained. The water level must be kept close to the prescribed level. Regular cleaning and periodic re-painting are necessary.

The siting of the pan can have a major impact on the measurements. For example, a pan sited on bare soil may record higher evaporation rates than one sited on grass because the air moving over the pan will tend to be drier (Allen et al. 1998). The difference in evaporation rates can be around 20%.

3.1.3 Data analysis

Measurements of pan evaporation can rarely be used directly as estimates of evaporation from a water body due to differences in size between the pan and the water body and, possibly, differences in the overlying air. Winter (1981) suggests that the use of data from pans located some distance away from the water body can result in considerable errors.

There are two approaches to estimating the evaporation of a water body from pan measurements; pan coefficients and pan conversions.

Pan coefficients are simply the ratio of the water body evaporation to pan evaporation. Numerous coefficients have been reported in the literature, although most apply to the US Class A pan. However, the coefficients are generally specific to the pan type, its location and the nature of the water body. In addition, they may vary with time. This variation with time takes account of the lag, due to heat storage, in large water bodies whereas the pans are too small for any lag effect.

Lapworth (1965) compared the evaporation from a 16 ha reservoir near London, calculated using the water balance, with that of a Symons pan and a US Class A pan over a seven year period. For annual totals, the pan coefficient for the Symons pan was 1.1 whilst that for the US Class A pan was 0.7. Crowe (1974) reported values of 0.94 and 0.78 respectively for a seven month study at Farmoor reservoir in Oxfordshire. Lapworth also found a strong monthly variation in the pan coefficients which varied between 0.47 and 1.18 for the US Class A pan. Winter (1981), in a hypothetical study, suggested errors of 10% for measurement errors, 50% for application of pan coefficients and 15% for areal averaging.

Pan conversions are achieved by taking the ratio of the bulk mass transfer equations of the lake and the pan:

$$\overline{E} = K \frac{(\overline{e}_s^* - \overline{e}\ )}{(\overline{e}_p - \overline{e})} \overline{E}_p \tag{2}$$

where $\overline{E}$ is the mean evaporation rate from the water body, $\overline{E}_p$ is the mean evaporation rate of the pan, K is an empirical constant, $\overline{e}_s^*$ is the mean saturated vapour pressure of the air at the water surface temperature, $\overline{e}_p$ mean saturation vapour pressure at pan surface temperature and $\overline{e}$ is the mean vapour pressure of the air at reference height.

However, this method is dependent on knowing the surface temperatures of the lake and the pan, which is rarely practical. In addition, an empirical coefficient is still required which has to be determined for the specific situation.

3.2 Mass balance

The mass balance method of measuring open water evaporation is simple in principle. Evaporation is calculated as the change in volume of water stored and the difference between inflow and outflow, i.e.:

$$E = P + \frac{(Q_{ri} + Q_{gi}) - (Q_{ro} + Q_{go}) - dV/dt}{A_s} \tag{3}$$

where E is the evaporation rate from the water body, P is the mean rate of precipitation over the sampling period, Q_{ri} is the surface inflow rate, Q_{ro} is the surface outflow rate, Q_{gi} is the groundwater and seepage inflow rate, Q_{go} is the groundwater and seepage outflow rate, V is the water stored and A_s is the surface area.

The relative importance of the terms depends on the hydrological and physiographical setting. The feasibility of determining evaporation depends primarily on the relative magnitudes of the terms. It is very difficult to obtain a reliable estimate whenever the evaporation is of the same order of magnitude as the errors inherent in the measurements. Thus, the method is unsuited to water bodies with large flows passing through.

An example of a very detailed analysis of the mass balance of a lake is provided by Harbeck *et al.* (1954) who describe a study of Lake Hefner (13.8 km^2 surface area near Oklahoma, USA) over a 16 month period. They estimated that the error in the monthly estimates of evaporation was less than five percent. This must be taken as the highest accuracy that is likely to be achieved using this method.

In the UK, Lapworth (1965) estimated the evaporation from a 16 ha reservoir near London over a period of seven years. The reservoir was man-made and an assumption was made that seepages were negligible. There were no inflows and outflows (except for a single lowering) during the period of the study and the rainfall inputs were measured with a raingauge at the site. An assessment of the errors suggested that the estimated evaporation was within

5 per cent of the true value. Given the exceptional nature of the conditions at this site due to lack of inflows and outflows, the results from this study must be accepted as having a high accuracy. Crowe (1974) reported the results of a seven month study during the commissioning of Farmoor reservoir in Oxfordshire but gave no assessment of the accuracy.

3.2.1 Measurements

Depending on the size of the water body, one or more raingauges are required to estimate precipitation. In most cases, precipitation is estimated from gauges on the surrounding land. In the case of large water bodies, the difference in the properties of the land and water surfaces, in particular through the partition of the incoming energy by the land surface between the latent heat flux and the sensible heat flux into the atmosphere, may result in the water body having a distinct micro-climate with the result that the precipitation could be appreciably different from that on the land.

The surface outflow of larger water bodies may be measured to a reasonable accuracy. However, surface inflow is generally known less accurately. Commonly, flows of only the major water courses are measured. Unmeasured flood water inflow may be large. If flow is seasonal, surface inflow during the summer may be small enough in comparison with the evaporation for the evaporation to be calculated with reasonable accuracy during this period. In order to achieve an acceptable error in the estimation of evaporation, the volumes of inflow and outflow should be comparable to the evaporation losses. For example Gangopaghaya *et al.* (1966) pointed out that, in the case of the Franklin D Roosevelt Lake on the Colorado River, the errors in measuring the outflow would result in an uncertainty that was ten times the amount of evaporation. In the case of the Lake Hefner study, the measured inflows and outflows were 10% greater than the evaporation over the 16 month period (Harbeck et al. 1954).

The volumes of groundwater and seepage inflow and outflow are usually unknown. In some situations it may be possible to assume that these are negligible. A further complication can arise if bank, or groundwater, storage occurs (water can be temporarily stored in the voids of materials forming the sides of the water body). Gangopaghaya *et al.* (1966) has pointed out that this can increase the total storage capacity by as much as 12% with the consequent error in the estimation of evaporation if this is not taken into account.

Water level recorders and a reliable depth-storage relationship are required. The use of more than one water level recorder should be used for large water bodies, in order to avoid errors due to seiches and wind set-up.

In view of the possible errors, the mass balance method is unlikely to be applicable over periods shorter than a month.

3.3 Energy budget

In this method, as its name implies, evaporation from a water body is estimated as the energy component required to close the energy budget when all the remaining components of the budget of the water body are known, i.e. it is the residual. The energy associated with evaporation comprises two components; the heat required to convert liquid water into water vapour (the latent heat of vaporisation) and the energy of the water vapour molecules carried from (advected) the water body. The latent heat of vaporisation ranges between 2.5 and 2.4 MJ kg^{-1} for liquid water between 0 and 40°C.

The energy budget of a water body is given by:

$$N = S(1-\alpha_s) + L_{\downarrow}(1-\alpha_L) - L_{\uparrow} - \lambda E - c(T_s - T_b)E - H + F_{in} - F_{out} + F_p - G \tag{4}$$

where N is the change in the energy storage in the water, S and $L\downarrow$ are the incident short and long-wave radiation respectively, and α_S and α_L are the albedos (reflectivities) for short and long-wave radiation, $L\uparrow$ is the long-wave radiative loss from the water, λE is the flux of latent heat (evaporation rate in energy flux units; λ is the latent heat of vaporisation and E is the evaporation rate in mass units), c is the specific heat of water, T_s and T_b are the temperature of the evaporated water and an arbitrary base temperature respectively, H is the flux of sensible heat (the energy used in warming the atmosphere in contact with the water which is then convected upwards), F_{in} and F_{out} are the heat fluxes associated with water flows in and out of the water body, F_P is the heat inflow associated with precipitation, and G is the heat conduction occurring between the water and its substrate. All the energy components are in units of energy per unit surface area of the water.

The three radiation terms together give the net radiation, R_n (see Eq. 1). so rewriting Eq. (4) gives

$$\lambda E + c(T_s - T_b)E = R_n - H + N + F_{in} - F_{out} + F_P - G \tag{5}$$

Usually the sensible heat term (the amount of energy directly warming the air) cannot be readily determined and it is eliminated from Eq. (5) through use of the Bowen ratio, β. This is defined as the ratio between the sensible and latent heat fluxes. It can be expressed:

$$\beta = \frac{H}{\lambda E} = \frac{c_p \phi (T_s - T_a)}{\varepsilon_m \lambda (e_s^* - e)} \tag{6}$$

where c_p is the specific heat of air at constant pressure, ϕ is the atmospheric pressure, T_s and T_a are the temperatures of the water surface and the air at a reference height, ε_m is the ratio of the molecular weight of water to that of dry air, and e_s^* and e are the saturated vapour pressure of the air at the water surface temperature and the vapour pressure of the air at the reference height. The ratio $c_p\phi / \varepsilon\lambda \equiv \gamma$ is also known as the psychometric coefficient. More detail on the Bowen ratio and many other aspects concerning evaporation physics are given in Brutsaert (1982).

From Equation 6, $H = \beta\lambda E$ which when substituted into Eq. (5) gives the evaporation rate,

$$E = \frac{R_n + N + F_{in} - F_{out} + F_P - G}{\lambda(1+\beta) + c(T_s - T_b)} \tag{7}$$

The second term in the denominator represents a correction term for the difference between the temperature of the evaporated water and an arbitrary base temperature.

By suitable selection of averaging period it is sometimes possible to neglect the F_{in}, F_{out} and G terms. Indeed, it is usually the case that the energy content of a water body is chiefly governed by the exchange of energy through the surface, rather than the inflows, including precipitation, and outflows and the water-substrate interface (Henderson-Sellers, 1986) . This would certainly be the case if the volumes of water flowing in and out of the water body are small compared to the volume, or the temperatures are close to the temperature of the water body. Therefore, the last four terms in the numerator of Eq. (7) can often be neglected if $T_b = T_s$ and the energy budget is then given by

$$E = \frac{R_n + N}{\lambda(1+\beta)} \tag{8}$$

This is sometimes referred to as the reduced energy budget equation ((Simon and Mero, 1985), (Assouline and Mahrer, 1993), and (dos Reis and Dias, 1998)).

The energy budget method consists of determining, by measurement or estimation, the different terms in either Eq. (7) or (8).

3.3.1 Measurements

After the direct measurement of evaporation, the energy budget is widely considered to be the most accurate method of estimating evaporation (Assouline and Mahrer, 1993 quoting Hoy and Stephens, 1977). As such it is often used as a reference method against which other methods are validated or calibrated. The accuracy depends upon the timescale and size of the water body. Because of the heat storage, the larger the water body, the longer the time interval required between measurements of the temperature profile to attain acceptable accuracy in the temperature differences. In the classic study at Lake Hefner (Anderson, 1954) an accuracy of 5% in the evaporation estimate was achieved for periods of a week or more but decreased for shorter periods. For a shallow (average depth 0.6 m) lake, Stewart and Rouse (1976) assumed that daily values were sufficiently accurate to use them as a standard against which an alternative method was validated.

The disadvantages of the energy balance method are the large number of measurements needed, the frequency of the measurements, and the difficulties inherent in making some of them. Consequently it is expensive and has not often been used in the more complete form of Eq. (7). Exceptions being the Lake Hefner study (Anderson, 1954), the work by Stauffer (1991) on Lake Mendota, Wisnonsin, and Sturrock *et al.* (1992) on Williams Lake, north central Minnesota, and more recently a comparative study of evaporation from two lakes in Florida (Sacks et al. 1994).

For accuracy, measurements of surface and profile water temperatures and the micrometeorology should be made at a representative point or points over the water body. This has often been achieved using an anchored instrumented raft (e.g. Anderson, 1954; Assouline and Mahrer, 1993; Sturrock *et al.*, 1992) but also for ease of maintenance and cost, measurements have been made over land and sometimes data used from distant weather stations. Work has been done to determine the effect on the accuracy of the evaporation estimates of using land-based and distant data sources (e.g. Keijman, 1974; Rosenberry *et al.*, 1993; Winter *et al.*, 1995).

The importance of the net radiation in the energy budget makes its accurate measurement or estimation paramount. Modern instrumentation allows the direct measurement of the net radiation to an accuracy of about 5%. However, where net radiometers are unavailable, R_n is calculated from either measurements or estimates of the radiation components (Eq. 1) and over the years there has been much work on improving the accuracy of these. A review of the many equations that have been developed to allow the short and long-wave radiation to be estimated from astronomical, meteorological and climatological data is given by Henderson-Sellers (1986), and Brutsaert (1982, pp 128-144). Major factors affecting the value of the incoming solar radiation, *S,* are atmospheric scattering, absorption and reflection, so that cloud amount and type are important, as well as season and latitude. The reflected component depends upon the albedo, which in turn varies. to a greater or lesser extent depending upon the degree of cloudiness. with solar angle. $L_\downarrow$ the long-wave radiation emitted by the atmosphere can be calculated from vertical profiles of temperature and humidity. However such data are not often available and it is usual to calculate it using the Stefan-Boltzmann relationship,

$$L_\downarrow = \varepsilon\sigma T_a^4 \qquad (9)$$

where T_a is the air temperature near the surface σ is the Stefan-Boltzmann constant and ε is the clear-sky atmospheric emissivity which can be calculated from air temperature and humidity near the surface. Like S, $L_\downarrow$ is also affected by cloudiness. The Stefan-Boltzmann equation, with appropriate surface values for the temperature and emissivity, is also used to calculate long-wave radiative loss, $L_\uparrow$, from the water. Stannard and Rosenberry (1991) found that measuring the incoming radiation and modelling the outgoing radiation resulted in overestimates of lake net radiation compared with directly measured values. One possible reason for this was differences in incoming radiation between the lake and the site where they measured it, 4.5 km away. Whether estimated or measured, the radiation values are integrated to produce period estimates consistent with the other measurements.

Estimation of the Bowen ratio, β, requires measurement of air temperature and specific humidity above the water and temperature and saturated specific humidity at the temperature of the water surface. This is usually achieved using wet and dry bulb thermometers at a reference height on a raft and a thermistor within the top few cm of the water.

The change in heat storage N per unit surface area is calculated from the following:

$$N = \rho c d \frac{\Delta T_w}{\Delta t} \qquad (10)$$

where ρ, c, d and ΔT_w are the density, specific heat, depth, and change in spatially averaged temperature of the water body in time step Δt. For pans and shallow lakes that are well mixed, T_w can be approximated by the surface temperature (Keijman, 1974). This however begs the question as to a suitable average value for the surface temperature; in calm conditions and high solar radiation, spatial variation in surface temperature can be large over short time scales. For deep lakes it is necessary to conduct thermal surveys consisting of temperature profiles with depth, measured ideally at a sufficient number of stations to produce a good average. For example, in the exceptional Lake Hefner study, surveys were made at weekly intervals at 16 stations and daily at one of two stations (Anderson, 1954), while at Williams Lake, surveys were made fortnightly at 16 stations (Sturrock et al. 1992). Selection of the appropriate time interval, which will depend upon the size of the water body, can result in the value of N being small enough to be neglected.

Estimation of the energy advected in and out of the lake requires that the inflow and outflow are gauged accurately and the water temperature measured. Inflow includes rivers and land surface run off, bank storage and seepage from groundwater. Outflow includes rivers, controlled withdrawals (reservoirs) and leakage to groundwater. Where inflow or outflow are large relative to the volume of the water body, and cannot be accurately gauged, the energy balance method may become unusable. However, in many lakes the relative inflow and outflow are small (e.g. Williams Lake, Sturrock *et al.*, 1992). Sturrock *et al.* (1992) calculated the groundwater volumes using the Darcy equation and used water temperature from wells for inflow, and surface temperature for leakage. The energy advected by rainfall is usually determined from the recorded rainfall and the wet-bulb temperature recorded during rainfall. Sacks *et al.* (1994), Stauffer (1991) and Sturrock *et al.* (1992) concluded that for the lakes that they studied, the advected energy was trivial compared to the other terms; e.g. 1% of the radiation terms. However, of the advected terms, Sacks *et al.* (1994) and Stauffer (1991) found that the largest was that due to precipitation. In non-natural or semi-natural water bodies, other advective components may be large but easy to measure, e.g. reservoirs and cooling ponds.

In some circumstances, the heat conduction term G can be significant, Sturrock *et al.* (1992) found that in the summer neglecting it made a 7% difference to the estimated evaporation from Williams Lake (average depth 5.2 m) in Minnesota. Stauffer (1991) states that ignoring sediment heat exchange can be a major source of error in estimation of evaporation and implies that the Lake Hefner results may be flawed through ignoring this component. He used annual sine-wave functions to model the sediment-water heat exchange (Likens and Johnson, 1969).

3.3.2 Errors

Comparisons have been made of the evaporation estimated using the energy balance with direct measurements using eddy correlation equipment mounted over lakes (Sene *et al.*, 1991, Stannard and Rosenberry, 1991 and Assouline and Mahrer, 1993). These show that, for deep lakes, the hourly or daily evaporation rates are determined primarily by the wind speed and atmospheric stability, with the energy being supplied from the heat storage in the lake. In consequence, estimates of evaporation on a short timescale determined from the energy balance method for deep lakes may not be accurate. Assouline and Mahrer (1993) found that, for a period of high wind speeds and sensible heat advection, the daily average evaporation rate estimated from the energy budget method was 2.8 mm day^{-1} compared to 4.1 mm day^{-1} measured using eddy correlation. However, they also found much closer agreement for a second period when wind speeds were lower and advection was less. Good agreement between the energy budget and eddy correlation estimates of evaporation can be obtained for longer time scales. Anderson (1954) gives an accuracy of evaporation estimates of 5% for periods of a week or more for Lake Hefner.

3.4 Bulk or mass transfer

A simple derivation of the bulk transfer equation is given by Sene *et al.*, (1991). It has the form

$$E = Cu(e_s^* - e) \tag{11}$$

where C is the mass transfer coefficient, u is the wind speed and e_s^* and e are the saturated vapour pressure of the air at the water surface temperature and the vapour pressure of the air at the reference height. The mass transfer coefficient can be thought of as the total drag coefficient; the combination of skin friction and a force resulting from the deceleration of the wind in the direction of flow. It can be shown that the mass transfer coefficient and the roughness lengths used in the Penman-Monteith equation (see Section 3.5.2) are linked. Over a uniform surface C can be calculated from theory which indicates that it is a function of the atmospheric stability and the roughness of the surface which itself is affected by the wind speed (Brutsaert, 1982; see p.117). The value of the coefficient has often been determined for sea surfaces although there is considerable scatter in the results (Brutsaert, 1982 see Table 5.3). For most inland water bodies the conditions of surface uniformity are not met and it is necessary to make more restrictive assumptions to obtain a theoretical solution to the evaporation and heat transfer equations (Brusaert, 1982; pp. 167-171). The value of C reflects the transfer characteristics of the particular water body which are determined by its geometry, plant cover, and the topography, land use and climate of the surrounding land. Moreover the value of the coefficient is specific for the characteristics of the site used to record the meteorological data; e.g. a value derived for wind speed measured at 2 m will not be correct for use with wind speeds measured at 10 m, even at the same site. Over the years, meteorological data have been inconsistently measured using a variety of different standards, resulting, according to Singh and Xu (1997), in over 100 such evaporation formulae. It is therefore not possible to find a value of C that is applicable to all water bodies. Rather, it is best to determine it empirically for a particular water body from the ratio of the mean evaporation rate, measured using a standard method e.g. eddy correlation or the energy budget, to the mean vapour pressure gradient. Nevertheless, attempts have been made to produce a generally applicable value. On the basis of an extensive measurement programme on reservoirs in the western USA, Harbeck (1962) suggested an expression for C that incorporated the area of the water body. In appropriate units (Shuttleworth, 1993) the transfer equation is

$$E = 2.909 A_s^{-0.05} u_2 (e_s^* - e) \tag{12}$$

where A_s is the area of the water surface in m^2, and u_2 is the wind speed at 2 m above the water surface. This is suitable for lakes in the range of 50 m < $A_s^{0.5}$ < 100 km that are in a relatively arid environment. A similar expression for pans in the range 0.5 m < $A_s^{0.5}$ < 5 m, is also given by Shuttleworth based on the work of Brutsaert and Yu (1968) viz.,

$$E = 3.623 A_s^{-0.066} u_2 (e_s^* - e) \tag{13}$$

The weak inverse dependence of the transfer coefficient upon the size of the water body reflects the effect of the reduced efficiency of turbulent transfer over the smooth water surface (Shuttleworth, 1993). However, some observations indicate that transfer is enhanced over large water bodies. For example, Venalainen *et al*. (1998) observed, from direct micrometeorological and eddy correlation measurements over two lakes in Sweden, that evaporation rates were greater from the larger of the two lakes. They attributed this to the effect of increased wind speed more than compensating for opposite effect of the increased humidity of the air associated with the larger distance travelled by the air over water. They also noted that evaporation from lakes with forest at the edge would be reduced through sheltering: apparently the reduction in turbulence associated with the reduced wind speeds more than compensates for the increased aerodynamic roughness of the forest.

An alternative form for the mass transfer equation dating from the 19th century has also been widely used. This takes the form

$$E = f(u)(e_s^* - e) \tag{14}$$

where *f*(*u*) is a function of the wind speed, $f(u) = a + bu$ with empirical constants *a* and *b*, that allows for free convection, i.e. evaporation when there is no wind. Sweers (1976) reviewed wind speed functions and concluded that, for a temperate climate, best results were obtained using the wind function of McMillan (1973) adjusted for the area of the water body in relation to the lake studied by McMillan. This function is,

$$f(u) = \left(\frac{5 \times 10^6}{A_s} \right)^{0.05} (3.6 + 2.5u_3) \tag{15}$$

where u_3 is the wind speed measured over the water at 3 m above the surface.

3.4.1 Measurements

Once a value for *C* has been determined, this method requires routine measurements of wind speed and vapour pressure at the same height as the measurements used in the determination of *C*. Unless the water body is less than a few metres across these measurements should be made over the water so that they are representative of conditions prevailing over most of the water surface. In addition, to determine e_s^* the average surface temperature of the water must also be measured.

When evaporation estimates are required on hourly or daily time scales then the effects of atmospheric stability must also be considered (e.g. Stauffer, 1991). However, for long-term estimates these effects can usually be neglected.

Other functional forms, some of which include air temperature, have been used. Singh and Xu (1997) tested 13 mass transfer equations, transformed into seven generalised forms using climatological data from north west Canada. They compared monthly evaporation estimates with pan data at four sites after calibration of each equation for each site. Agreement was generally good between the estimates and measurements for a particular site but the equations did not give good results at sites for which they were not calibrated. On a monthly time scale the humidity gradient, rather than the wind speed, primarily controlled the evaporation. This is at variance with the observations on two Swedish lakes (see above and Venalainen *et al.*, 1998) and demonstrates a limitation of pan estimates of evaporation.

3.4.2 Errors

Simon and Mero (1985) gave up trying to use the mass transfer method to estimate evaporation from Lake Kinneret in Israel because of inconsistent results and large scatter in estimates of the transfer coefficient. In contrast Sacks *et al.* (1994) found good agreement (generally within 8%) between the energy-budget evaporation and monthly mass-transfer evaporation for a shallow lake in Florida, but larger discrepancies (mean monthly difference of 24%) for a similar but deeper lake, also in Florida. Correcting the mass transfer coefficient for stability effects (Stauffer, 1991; Harbeck et al. 1958) did not improve matters. Sacks *et al.* suggested that the differences might be a smoothing effect caused by using long-term mean vapour pressure gradients; one of the main problems with this method is that it is sensitive to the errors in the vapour pressure gradient. They also found that using the Harbeck (1962) form for the mass transfer coefficient (Eq. 12) produced lower values that resulted in underestimates of the evaporation from the shallow lake by 14% and from the deep lake by 27%. This was in contrast to Sturrock *et al.* (1992) who found that the Harbeck prescription gave higher values than those based upon energy budget estimates. The reasons for these discrepancies remain unclear.

3.5 Combination equations

3.5.1 The Penman and Priestley-Taylor equations

In the last fifty years the most widely used formula to estimate evaporation from water, or vegetation, has been the Penman equation (Penman, 1948). Its success when applied in many different locations is attributable to its physical basis. Linacre (1993) presents a table comparing monthly or annual measured evaporation with Penman estimates for a wide range of water bodies from around the world. The median value of the ratio of measured to estimated evaporation is 0.99 with a standard deviation of 0.12.

Penman combined the mass transfer and energy budget approaches and eliminated the requirement for surface temperature to obtain his expression for the evaporation in mm per day from open water:

$$E = \frac{\Delta R_n'}{\Delta + \gamma} + \frac{\gamma f(u)(e_a^* - e)}{\Delta + \gamma} \qquad (16)$$

where R_n' is the net radiation in units of equivalent depth of water (mm day^{-1}), Δ is the slope of the saturated vapour pressure-temperature curve and γ is the psychometric coefficient (or c_p/λ). Penman subsequently modified this to a form commonly known as Penman E_T, the evaporation rate expected from short well-watered vegetation. The open water form (Eq. 16) does not allow for heat storage and was not intended for use in estimating evaporation from deep water bodies with or without components of advected energy. To incorporate advected energy, R_n is replaced in Eq. (16) with A, the available energy, which is the sum of net radiation and any energy advected into the water body less any that goes into storage.

When air travels a long distance over a wet surface it will tend to saturation so that the second term in Eq. (16) tends to zero. The first term represents the lower limit of evaporation and is referred to as the equilibrium rate. However, in practice equilibrium evaporation is rarely found; because the atmosphere near the surface is never truly homogeneous and, even over oceans, atmospheric humidity deficits develop. Priestley and Taylor (1972) analysed data collected over oceans and extensive saturated land surfaces and found that the evaporation values were fitted using

$$E = \alpha \frac{\Delta A}{\Delta + \gamma} \tag{17}$$

where A is the available energy and the constant α accounts for the evaporation arising from the humidity deficit in addition to the equilibrium term. The equation is now known as the Priestley-Taylor equation. Priestley and Taylor found that the average value of α was 1.26 from the data they examined and there has been subsequent corroboration of this value by other studies. de Bruin and Keijman (1979) used the Priestley-Taylor equation to estimate the evaporation from a large shallow lake (Lake Flevo; 460 km^2, mean depth 3 m) in the Netherlands and found very good agreement with daily evaporation measured by the energy budget and water budget methods during the summer and early autumn with $\alpha = 1.25$. However, they also found diurnal variation in the value of α which they attributed to the variation of the difference between air and water temperatures and suggested that the conditions producing such variation would be expected from many lakes. They also found evidence of seasonality in the value of α, of at least the same magnitude as the diurnal variation in evaporation. This variation is the result of some evaporation occurring when the available energy was zero. de Bruin and Keijman also found very good agreement between the evaporation estimated from the energy budget and that estimated using the following formula:

$$E = \frac{\Delta A}{\lambda(0.85\Delta + 0.63\gamma)} \tag{18}$$

derived from the Priestley-Taylor equation, the relationship between α and β, and an empirical relationship, $\beta = 0.63\gamma/\Delta - 0.15$, given by Hicks and Hess (1977).

Stewart and Rouse (1976) derived a variation of Eq. (17) by using a linear function of incoming solar radiation to replace the net radiation and heat storage. The parameters, a and b, of the function were obtained by regression and the values are necessarily specific to their lake. However the resulting equation is identical to the formula of Makkink (1957) who used it to estimate the evaporation from well-watered grass and is:

$$E = a\frac{\Delta}{\Delta + \gamma} S + b \tag{19}$$

As with other methods the disadvantage of the Priestley-Taylor equation is the requirement for measured $R_n^{'}$ and N values, especially the latter (the change in the heat stored in the water); it is often not possible or too expensive to make adequate measurements of N for a large water body. de Bruin (1978) overcame this difficulty by combining the Penman and Priestley-Taylor equations, thus eliminating the energy term to give the relationship

$$E = \left(\frac{\alpha}{\alpha - 1}\right)\left(\frac{\gamma}{\Delta + \gamma}\right) f(u)(e_a^* - e) \tag{20}$$

This formula requires only measurements of air temperature, humidity deficit and wind speed at 2 m. de Bruin tested the method by using a form of the wind function given by Sweers (1976) with time-averaged input data measured at the centre of Lake Flevo to calculate evaporation for varying time intervals. He found good agreement with estimates from the energy budget method for intervals of 10 days or more. He also found that the Priestley-Taylor coefficient was not constant for intervals of a day or less.

3.5.2 The Penman-Monteith equation

The Penman-Monteith equation (Monteith, 1965) is a more general form of combination equation. It was developed to describe the evaporation of water vapour from the sub-stomatal cavities of plants into the atmosphere. Essentially the evaporation rate is obtained from the simultaneous solution of diffusion equations for heat and water vapour, and the energy balance equation. When applied to open water it takes the form

$$E = \frac{1}{\lambda}\left[\frac{\Delta A + \rho c_p (e_a^* - e)/r_a}{\Delta + \gamma}\right] \tag{21}$$

where the aerodynamic resistance r_a is the resistance that the water molecules encounter in moving from the water surface to a reference height in the atmosphere and is inversely proportional to the wind speed. This equation has the same physical basis as the Penman equation but does not contain the empirical calibration factors inherent in the wind function used by Penman. Thus it represents the best description of the evaporation process and is to be preferred to other estimates provided the necessary input data are available (the same proviso as required by the Penman model). The heat storage and net energy advected into the water body need to be included in the available energy, A. Accurate estimates also require that the value of the aerodynamic resistance, r_a, accounts for the effects of surface roughness, size of the water body, and atmospheric stability.

3.5.3 Measurements

The combination equations proper require values of net radiation, air temperature, vapour pressure and wind speed. Fewer input data are required by the simpler, derived equations like the Priestley-Taylor equation. Unlike the energy balance, and mass balance methods they do not require values of surface temperature. However, for accurate estimates of evaporation it is necessary to estimate or measure the heat storage in the water, unless the time interval over which evaporation estimates are required is such that the heat storage can be neglected. Linacre (1993) derived a simplified version of the Penman equation requiring just air temperature, wind speed and dew point data. He suggested two different methods for estimating solar irradiance, one of which used rainfall as a surrogate indicator of cloudiness and the other which accounted for temperature variation with latitude, altitude and distance from the sea. Using this method with monthly or longer average input data, he obtained good agreement (within 5%) with measured evaporation rates for a range of different sites in Australia, USA and Copenhagen.

3.5.4 Errors

As with the other methods already considered the uncertainties in the evaporation estimates are larger for bodies of deep water because of the larger heat storage component. For large water bodies this component is determined primarily by the surface energy exchange which in turn is affected by the atmospheric stability, and must be allowed for when daily, or shorter, estimates are needed. When water bodies exceed a certain depth then stratification occurs and the heat storage has to be determined from measured temperature profiles or hydrodynamic models. However, for lakes in tropical climates the water temperature can be nearly constant all year round so that the change in heat storage can be neglected (Sene *et al.*, 1991).

On the basis of data collected from the literature Linacre (1993) states that the probable error associated with monthly or annual evaporation estimated using the Penman equation with monthly data is about 8%.

Stewart and Rouse (1976) tested the Priestley-Taylor equation (Eq. 17) using data from a small (10^5 m^2) shallow (mean depth 0.6 m) lake and found very good agreement with evaporation estimated by the energy budget method on a half-hourly and daily basis. They concluded that evaporation could be estimated within 5% using this method. They also tested the Makkink formula (Eq. 19) and found that it gave estimates of evaporation to within 10% over fortnightly to monthly intervals.

In addition to the uncertainty connected with the heat storage and the measurement errors of the driving data there can also be systematic uncertainty associated with the aerodynamic resistance in the Penman-Monteith equation. Near the edge of a body of water the aerodynamic resistance will be determined chiefly by the aerodynamic roughness of the surroundings in the direction of the prevailing wind. For example if there is forest in that direction then it will generate large turbulent eddies but it will also reduce the wind speed. The effect of the surroundings of the water body on the aerodynamic resistance will reduce with distance. Usually, because the water is smoother than most other surfaces, the wind speed will increase with distance over water resulting in a smaller value of resistance, unless the higher wind speed causes waves with associated increased roughness. Shuttleworth (1993) suggested that using the aerodynamic resistance implicit in the Penman equation (Eq. 16) in the Penman-Monteith equation (Eq. 21) might result in overestimates of evaporation from very large lakes of 10-15%. However, this is probably an upper limit because Shuttleworth does not appear to have taken into account the increase in wind speed that occurs (see Section 3.4).

The model of de Bruin removes the requirement to know the heat storage term but its effects will be reflected in variation in the value of the Priestley-Taylor 'constant' α. If the appropriate value is not known then errors may be quite large because of the sensitivity of the evaporation estimate to this parameter. The model is also sensitive to errors in the vapour pressure gradient.

3.6 The equilibrium temperature method

Useful models have been derived from a more detailed consideration of the heat transfer processes occurring at the surface of a water body. These require the same driving data as the combination equations except the water heat storage which is calculated within the models. Edinger *et al.,* (1968) introduced the concept of an equilibrium temperature and associated time constant, determined from meteorological data, towards which the water temperature is driven by the net heat exchange, i.e. when the water is at equilibrium temperature the net rate of heat exchange is zero. From this he was able to derive an expression for the temperature of a well-mixed body of water as a function of time and water depth. Once the water temperature is estimated then it can be used to estimate the evaporative and sensible heat fluxes, the heat storage and the long wave radiative loss from the water. A similar approach was taken by Keijman (1974) who then used the calculated heat storage in the Penman equation (see, Eq. 16) to estimate the evaporation from the shallow Lake Flevo. de Bruin (1982) used a slightly different approach to obtain an expression for the water temperature that also used an equilibrium temperature, but one that was constant and equal to the mean value of that used by Keijman. Using this model with ten-day means of standard land-based meteorological data, de Bruin achieved good agreement between measured and predicted water temperatures over several years for two reservoirs of different depths in Holland. This type of work was extended by Fraedrich *et al.,* (1977) by considering the effect of energy advected to a reservoir by inflow and outflow. Two characteristic temperatures and related time constants allow simulation of the energetics of the reservoir in response to surface-transfer and hydrological forcing mechanisms. They applied the model to a large shallow reservoir and got good agreement between simulated and predicted monthly averages of water temperatures and upward long-wave radiation. They also used the water temperature to calculate the energy storage term, which in turn they used in the Penman (Eq. 16) and Priestley-Taylor (Eq. 17) equations to estimate evaporation rates. The best estimate of evaporation was given using the Penman equation.

3.6.1 Measurements

Keijman (1974) used daily mean values of dry and wet bulb air temperature, and wind speed together with sunshine duration, measured around the perimeter of a lake, from which he estimated net radiation, to drive his model. He also compared the effect of using the data collected at the centre of Lake Flevo with data collected at two stations on the perimeter of the lake. Equally good results were achieved when using data from a perimeter station if it was downwind of the lake. Fraedrich *et al.*, (1977) used monthly mean weather data together with rates and temperatures of the inflow and outflow to drive their more sophisticated model.

3.6.2 Errors

There is little reported in the literature on the errors associated with this method. de Bruin (1978) found good agreement between estimated and measured lake surface temperatures. Good agreement of surface temperatures was also found by Keijman (1974) and reflected in estimates of daily lake evaporation estimated by the Penman equation that had a standard error of 0.6 mm. Fraedrich (1977) also found that the Penman equation gave better estimates of evaporation using this method than the Priestley-Taylor equation.

3.7 Empirical factors

In operational estimates of evaporation, empirical factors to convert evaporation rates measured or estimated for one type of land surface (the reference evaporation) to those of another have been used for some time. They are comparable to the use of pan coefficients to convert measurements of evaporation from evaporation pans to those of other water bodies or land surfaces (see Section 3.1), and consist of simply multiplying the reference evaporation by an empirical factor (sometimes referred to as the crop coefficient).

Although the source of the reference evaporation could be any method, in practice it has been combination equations. This is because these equations use readily available meteorological data and have proven to be robust at estimating evaporation.

Penman (1948) gives factors to convert evaporation rates from "turf with a plentiful water supply" to an open water surface exposed to the same weather conditions as:

Midwinter (November – February)	1.67
Spring and autumn (March – April, September – October)	1.43
Midsummer (May-August)	1.25

These values were derived from measurements of evaporation at one site (Rothamsted Experimental Station) in southern Britain using cylinders 0.76 m in diameter and 1.83 m deep. Therefore, the use of these factors outside these conditions should be treated with caution. Measurements of evaporation from water were used to calibrate Penman's model of evaporation and so these factors should be used with estimates of reference evaporation calculated using this model.

Doorenbos and Pruitt (1984) list empirical factors (crop coefficients) to allow evaporation to be estimated for a wide range of land surfaces from time series of evaporation calculated using a modified version of the Penman (1948) model for grass. The modification involved changing the wind function. The factors given for open water evaporation are:

Humid environment – light to moderate wind	1.1
Humid environment – strong wind	1.15
Dry environment – light to moderate wind	1.15
Dry environment –strong wind	1.2

No background or justification for the values of these factors is given. These coefficients can be used for calculating annual totals of evaporation for all water bodies and monthly totals for shallow water bodies (less than 5 m). They can be used to estimate the monthly totals of evaporation from deep water bodies in equatorial regions. However, Doorenbos and Pruitt warn that, when applied to deep water bodies (greater than 25 m) with a change in climate during the year, in spring and the early summer the correct coefficients may be 20-30% lower due to heat storage in the water body. Conversely, due to heat release, the correct coefficients may be 20-30% higher in later summer and autumn.

Recently, Allen *et al.* (1998) have given crop coefficients for use with Penman-Monteith estimates of evaporation, for a hypothetical crop with a bulk surface resistance of 70 s m^{-1} and a height of 0.12 m (which can be taken as corresponding to short grass freely supplied with water). The coefficient given for water bodies in subhumid climates or tropics and water bodies less than 2 m in depth is 1.05. Two coefficients are given for water bodies greater than 5 m depth, clear of turbidity, in temperate climates. A value of 1.25 is recommended for the autumn and winter when the water body is releasing thermal energy and 0.65 when the water body is gaining thermal energy (spring and summer). Allen *et al.* urge caution in using these coefficients and give no background or justification for them. It should be noted that the version of the Penman-Monteith model used in the Meteorological Office Rainfall Evaporation Calculation System (MORECS) varies the bulk canopy resistance of grass through the year, to simulate the annual cycle, and so does not correspond exactly to the version given by Allen *et al.* (1998).

3.7.1 Sources of error

The potential errors in using empirical factors are very similar to those incurred by using pan coefficients with data from evaporation pans.

There are inevitably measurement errors inherent in the meteorological data used to calculate the reference evaporation. The dominant driving variable is the net radiation which is generally derived from measurements of the sunshine hours or incoming solar radiation. In the case of the latter, modern instruments are generally accurate to around ±5%. Comparisons between solar radiation values derived from sunshine hour recorders and direct measurements suggest that accuracies of around 5% are consistently achieved except in conditions of very low incident radiation (generally very cloudy days in winter).

It is essential that the meteorological data used to calculate the reference evaporation are representative of the meteorological conditions over the water body. It is difficult to quantify the error that may arise from failing to do this but it could amount to around 10%. In general meteorological data should be used from a nearby site that reflects the general topography and land cover in the vicinity of the water body.

The main source of error is likely to be the use of an inappropriate coefficient for the water body in question. To achieve a high level of accuracy, the coefficient(s) should be determined for each particular water body and should vary throughout the year. In practice, it is not feasible to do this and thus the coefficients should ideally only be relied upon when used in the conditions that they were determined. In particular, differences in the depth of water, and possibly the surface area, may result in errors of up to 30% in evaporation totals for time periods less than a year. The use of a single set of empirical factors for every year will potentially result in errors due to variations in the weather from year to year. For monthly estimates, this is likely to be up to 15 or 20%. In addition, the empirical factors used should have been developed for the estimates of potential evaporation of a specific source, e.g. Penman-Monteith. For example the differences between estimates of PE by the Penman (1948) model and the MORECS implementation of the Penman-Monteith model are likely to result in differences in estimates of open water evaporation using the same set of empirical factors of around 30%.

Therefore, annual estimates of open water evaporation made using empirical factors are likely to have an accuracy of about 30% whilst, for monthly values, it is likely to be around 50% (but see Chapter 11).

3.8 Comparison of methods

There are very few comparisons of methods of estimating open water evaporation in the literature. Winter *et al.* (1995) evaluated 11 different equations for a small lake (about 600 m across and a maximum depth in excess of 9 m) in a humid climate of the USA. They compared the estimated monthly evaporation with that determined using the energy balance method, for 22 months over a period of five years. The equations included forms of the mass transfer and combination equation methods. The equations were evaluated using meteorological data from a raft on the lake and from a land-based station. Winter *et al.* concluded that the combination equations (Eq. 16 to 18) of Penman (1948), Priestley and Taylor (1972) and de Bruin and Keijman (1979) best described the evaporation in terms of close agreement with the energy budget values, small standard deviations from the energy budget values, lack of seasonal bias and similarity of results whether using raft-based or land-based data. However it should be noted that these equations used the available energy, i.e. they included a term for the change in heat storage of the water body which was based on repeated thermal surveys of the lake.

3.9 Discussion

Pan evaporation and empirical factors can be considered as similar methods as they rely on the use of factors (ideally time varying) to convert 'standard' estimates of evaporation to those of the water body. The difference between the methods is the source of the reference evaporation: measurements from an evaporation pan or estimates of evaporation calculated using meteorological data. The consensus in the UK is that measurements using evaporation pans are more subject to error than estimates based upon the measurements of meteorological variables. The development of physically based models, such as the Penman-Monteith combination equation, has resulted in reliable estimates of evaporation being readily derived from meteorological data to the extent that it is generally considered to be the norm in the UK. As a consequence, pan evaporation measurements have been discontinued at the majority of sites and, at the few remaining sites, data are generally held as paper records whilst the meteorological data is in computer form.

The difficulty and expense of measuring all the elements that are required for the mass balance means that this method has only been applied in a few, exceptional circumstances. These tended to be in the 1950s and 60s. Since then, developments in instrumentation have meant that the energy budget method has become a more practical proposition. However, both these methods rely on calculating a balance, so that the errors accumulate in the estimate of evaporation. The result is that, unless the evaporation losses are comparable in magnitude to the other changes in the budget, the errors are likely to be large. Nevertheless, the energy balance method is considered to give the most accurate estimates of evaporation. For both methods, the estimates of evaporation are specific to the site where the measurements are made and cannot be transferred to other water bodies. The advantage is that local factors, such as thermal stratification, are taken into account.

The bulk transfer method seems initially very attractive as it makes use of data that are easily measured: the meteorological variables and the water body's surface temperature. In practice, the sensitivity to vapour pressure measurements combined with the difficulty of defining the wind function reduce the accuracy of this method. Because of the need for measurements of the surface temperature of the water body, the estimates of evaporation are specific to the site. However, this ensures that local conditions, such as thermal stratification, are handled.

Combination equations are the most widely used method of estimating evaporation. They are particularly attractive because they make use of readily available meteorological data. The major limitation is that they do not take the heat storage of the water body into account if driven by net radiation data. This can be remedied by carrying out periodic thermal surveys of the water body and inputting the available energy, rather than the net radiation, into the equation. However, this makes the methods site specific.

The equilibrium temperature method is a relatively new method, which might explain why there are few references to it in the literature. It is an attractive method because it is physically based, uses readily available meteorological data and takes the heat storage of the water body into account. The only major limitation is that it assumes that the water body is uniformly mixed and thus it does not consider thermal stratification.

4. METHODS CURRENTLY USED BY THE ENVIRONMENT AGENCY

A telephone survey of the current methods for estimating open water evaporation used in the Regional and Area offices of the Agency has been carried out. This established:

- the purposes for which estimates are required,
- how the estimates are made,
- how much time is spent on the estimates,
- what data are available for estimating open water evaporation.

4.1 Purpose

There was a strong consensus that the estimates were required for three purposes: abstraction licensing, water balance studies and management of wetlands.

4.1.1 Abstraction licensing

Abstraction licenses are most often applications by farmers for winter storage, the water from which is used for irrigation during the summer. The size of these water bodies is variable, but an 'average' case has a diameter of about 100 m and a depth of up to 6 metres. Other types of water body that fall within this category are ornamental and amenity ponds/lakes which require abstractions to maintain their levels during the summer. The size of these vary widely from a diameter of about 10 m to substantial ornamental lakes.

The requirement is to assess the risk of failure of the supply. The approach adopted generally is to estimate the worst case scenario by calculating the evaporation during the year or summer of an exceptionally hot year, e.g. 1976. The accuracy required is commonly agreed to be 10%.

The frequency of requests per Area varies between a couple per month to very infrequent.

4.1.2 Water balance studies

The water balances of lakes and reservoirs are required for planning and modelling. The size of the water body involved is variable but includes strategic reservoirs.

A more thorough analysis is required for these water bodies as the data are often incorporated in a model. Generally daily values of evaporation are required as this is the time step of the catchment models. However, it is accepted that accurate daily values are not essential and that weekly values intelligently disaggregated to daily values are acceptable. The time period used is variable but is generally several decades and could extend from 1918 to the present. An accuracy of 10% is commonly seen as acceptable but several people expressed the view that 5% is preferable. However, there is consensus that the accuracy should be the same as other hydrological variables such as runoff and rainfall.

The need for estimates is infrequent, with a couple a year per Region being the highest mentioned.

4.1.3 Wetlands

These were identified particularly by Southern Region but were also commented on by some of the others. Estimates of open water evaporation are required for the wetlands in the context of both abstraction licensing and water balance studies. The abstraction licensing applications are made by farmers who have land within areas designated as Environmentally Sensitive Areas and who are applying for grants to manage the land. The requirement is for at least 20% of the area to be covered with water to a depth of at least 0.1 m during the winter months. The areas involved are typically 40 to 100 ha. About six applications for abstraction licenses are received per year. Estimates are required as monthly values, either for the fifth driest year in 20 or for the driest year in 20.

During the summer months, there may be a need to maintain target water levels in marsh drainage strips.

The requirement in water balance studies is to ensure that flows in streams, some fed by springs, have sufficient volumes to maintain wetlands. Therefore, estimates of losses from the wetlands, due to open water evaporation, are required.

4.1.4 Others

The Thames Region has an exceptional requirement to estimate open water evaporation from gravel pits. These are extensive, for example covering 10 km^2 in the area of the Cotswold Water Park. The estimation of evaporation is complicated by a throughflow of groundwater which may be very variable. In addition there may be leakage from the river into the gravel pits. The estimates of open water evaporation are required to quantify the increased losses to evaporation due to the land cover change from grass to open water. In the future, some of these gravel pits may be converted to wetlands to enhance their ecological value.

There was also some mention of open water evaporation from canals and rivers, the latter in the context of naturalisation and transfer schemes. There can be significant evaporation losses if there are ponded stretches.

4.2 Methods

The dominant method employed by the Agency's staff is the use of empirical factors. The majority apply these factors to the Meteorological Office Rainfall and Evaporation Calculation System (MORECS) grass potential evaporation but Penman grass potential evaporation is also used. Only one contact had made use of a model, which was that of Penman (1948).

4.2.1 Anglian

Estimates of open water evaporation are often obtained by the consultant employed on a particular project or by the applicant, in the case of abstraction licensing. MORECS weekly potential evaporation (PE) for grass have been used with empirical factors applied that were given by Penman (1948). The values are disaggregated to daily values by dividing by 7. The time period is from 1961 onwards. On some occasions, open water evaporation has been directly calculated, from the data provided by one of the Region's climate stations, using the model of Penman (1948).

4.2.2 Midlands

Few estimates of open water evaporation are carried out in this Region. Those which are, are dominantly for abstraction licensing. Estimates for reservoirs were done long ago and so are built into models. In the 1970s, an evaluation of data from evaporation pans that had been installed was carried out and it was found that there was considerably variation from site to site which could not be linked to factors such as altitude, wind speed etc. Therefore, a decision was made to use potential evaporation data adjusted by a factor. Following the advice of the Met. Office, a factor of 1.2 was used to take into account the difference in albedo (Grindley, 1969). This method is in use today whereby an empirical factor of 1.2 is applied to MORECS grass PE to obtain annual or monthly totals.

4.2.3 North East

Open water evaporation is calculated by multiplying MORECS grass PE by an empirical factor of 1.2 which was obtained from Grindley (1969). This is used to generate either monthly values or the long term average.

4.2.4 North West

Use is made of evapotranspiration data provided by PETCALC, a system for calculating potential evaporation for grass based on the model of Penman (1948). The sources of the driving variables are the Met. Office Monthly Weather Reports from 1918 onwards for individual stations. Empirical corrections are applied to each driving variable for elevation and latitude and longitude to extrapolate the time series of the nearest station to the location for which the estimate is required. Thus the output is estimates of grass potential evaporation for a given location, at monthly intervals for the specified time period. In order to calculate open water evaporation, the grass PE is multiplied by a series of empirical factors (1.25 May-Aug, 1.67 Nov-Feb, 1.43 Mar, Apr, Sep, Oct) given by Penman (1948)).

4.2.5 Southern

Monthly estimates are obtained from a table of open water evaporation values derived from average values of PE given by MAFF (1967). (The exact way in which the table was derived from MAFF (1967) is not known.) The MAFF document is a set of tables that give, for each county, the monthly average potential evaporations for grass, calculated using the model of

Penman (1948), for the period 1950 to 1964. Correction factors are included to allow the values to be adjusted for altitude. The average winter and summer potential evaporations for the coastal strip (5-10 miles) are also provided.

4.2.6 South West

Two rules of thumb are generally used for abstraction licenses; either 500 mm per year or 8 mm per day. For water balance studies, MORECS grass PE data are used, either in weekly or monthly time series, with the empirical factors of Penman (1948)) applied.

4.2.7 Thames

Estimates of open water evaporation are based on Penman grass PE. In the past, these were obtained from the Met. Office. However, this service has been withdrawn so the time series is being maintained by using MORECS grass PE data, adjusted to Penman grass PE by applying empirical correction factors obtained by regressing overlapping time series of the Penman and MORECS data. A time series from 1918 to the present is available. The empirical factors of Penman (1948) are applied to the estimates of grass PE to convert them to open water evaporation.

4.2.8 Wales

A rule of thumb of evaporation losses of 610 mm during a dry summer is often applied for abstraction licenses. Otherwise, use is made of tables of PE given by MAFF (1967) to check applications for an abstraction license. These are based on Penman (1948) grass potential evaporation with values given as monthly averages (1950-64), on a county basis, corrected for the average altitude of the county and with correction factors for altitude. No factor is applied to adjust from grass to open water.

4.3 Time and Value

Agency staff typically quoted half an hour as the amount of time that people were prepared to spend getting estimates for abstraction licensing. In the case of water balance studies, a couple of days was generally thought reasonable. Given the relatively low number of times per year that the estimates are required, this suggests that Agency staff spend comparatively little time making estimates of open water evaporation at present.

Decisions made by Agency staff must be based upon, and justified by, estimates made using methods that are robust and defensible, both internally and externally. The use of inaccurate estimates may result in resources having to be spent either justifying the basis for the decisions or rectifying the consequences.

Consequences specifically applying to estimates of open water evaporation include:

- A licensee having insufficient water for his purpose (e.g. irrigation from winter storage) if the evaporation estimate is too low.

- The licensee may be allocated more water than is actually required if the evaporation estimate is too high. This will reduce the amount available to other potential license holders, or even in licenses being refused. This has relevance to the Agency's national initiatives for Abstraction Management Strategies and the Abstraction Licensing Review.

- There may be detrimental impacts on the environment due to water being allocated to other uses as a result of either the abstraction licensing being set too high or of open water evaporation losses of wetlands being underestimated. The Agency has a duty to protect and, where possible, enhance the environment. This is particularly relevant to the EU Habitat Directive.

- Inappropriate amounts of water may be allocated for amenity uses, such as navigation on canals.

Therefore there are three benefits to the Agency of robust estimates of open water evaporation:

1. Agency staff can carry out their duties in a consistent and efficient manner.

2. All potential abstractors and amenity users have sufficient resources for their requirements.

3. The Agency can carry out it's core duty of protecting and enhancing the environment.

4.4 Data availability

4.4.1 Meteorological data

The number of stations maintained by the Regions is very variable. Table 4.1 shows the number of meteorological stations and how many of them have data that can be used with the combination equation or equilibrium temperature methods. Although there are 119 stations recording climate variables, only 39 of these record all the variables required to use either combination equation or equilibrium temperature methods. The spatial distribution of stations varies with Thames Region having no stations. The length of record is also very variable with comparatively few extending back prior to 1960.

Table 4.1 Number of stations at which climate variables are measured (based on National Rivers Authority, 1992, updated where changes are known)

Region	Number of sites	Combination equation compatible
Anglian	16	3
Midlands	19	2
North East	12	9
North West	2	2
Southern	15	8
South West	47	10
Thames	0	0
Welsh	8	5
TOTAL	119	39

4.4.2 Evaporation data sets

Most of the Regions have acquired data from MORECS of one type of another. MORECS is capable of providing weekly and monthly averages of evaporation and soil moisture deficit over 40 x 40 km squares (Hough and Jones, 1997). Great Britain is covered by 190 grid squares and MORECS uses daily data from about 100 synoptic weather stations as inputs. In the MORECS model, the evaporation is calculated by a modified form of the Penman-Monteith equation:

$$E = \frac{1}{\lambda}\left[\frac{\Delta(R_n - G_s) + \rho c_p (e_s - e)(1 + br_a/\rho c_p)/r_a}{\Delta + \gamma(1 + r_s/r_a)(1 + br_a/\rho c_p)}\right] \qquad (22)$$

where b is calculated as:

$$b = 4\varepsilon\sigma(273.1 + T_a)^3 \qquad (23)$$

r_s is the bulk surface resistance and is the resistance to water vapour diffusing out of the pores (stomata) of the land surface. For water, this value is set to zero. In the case of a land cover of permanent grass, the value used varies during the year (80 Nov – Feb, 70 Aug, Sep and Oct, 60 Mar, Jun and Jul, 50 Apr, 40 May, in s m^{-1})

The MORECS model is run with a daily time step. However, weekly and monthly grid square average values of potential evaporation, actual evaporation, soil moisture deficit and HEP (the hydrologically effective rainfall or potential ground water recharge) for 14 crops/surface covers on soils with median available water capacity can be output. The five basic meteorological parameters of sunshine hours, air temperature, vapour pressure, wind speed and rainfall, for each of the MORECS squares, are also potentially available for output as weekly or monthly values. The only output archived by the Met. Office are the weekly estimates of potential evaporation for grass for the grid data. MORECS grass PE data held by the Agency's Regions generally covers the period from 1961 to 1990 at weekly or monthly time steps, and often extends up to the present. The Northwest Region does not currently receive any MORECS grass potential evaporation data whilst the Northeast Region has it from 1988 to the present. Thames Region has data for a few squares from 1985.

The Met. Office view the grid data as a product required rapidly by the user to monitor the state of a variable, such as soil moisture. Therefore it is based on the network of primary meteorological stations and is run to produce weekly or monthly estimates. No provision is made to ensure that the calculations are performed on each occasion with data from the same meteorological stations. Hence, if data are missing from a particular station, no attempt is made to infill this gap, rather, data from another, more distant, station will be used. Thus there is not necessarily any consistency over time in the meteorological data used to drive the model. This potential source of inaccuracy is exacerbated by the fact that since MORECS was first set up (in the late 1970s) there has been a decline in the overall number of stations providing data.

Although meteorological data from about 100 stations are used, less than 70 of these stations have data from sunshine recorders (which are used to calculate the net radiation, the dominant driving variable of evaporation). In addition, the location of these stations does not represent a uniform coverage as stations tend to be located in the lowlands, at airfields. There is an additional bias in that a significant number of the sites are located on, or very near to, the coast with the result that they may not be representative of the inland areas that may dominate a particular square.

PETCALC is a system specifically developed for the North West Region. It produces monthly estimates of Penman (1948) potential evaporation for grass by using an albedo of 0.25 when calculating the net radiation. It also includes a procedure for quantifying the soil heat flux. The input data are provided by the Met. Office monthly weather reports for individual stations within the Region, from 1918 onwards. In order to infill missing data in some of the stations, regressions were derived to predict the variation in the driving variables (sunshine hours, air temperature, wind speed and wet bulb depression) as functions of the station's position in terms of altitude and latitude and longitude. The Region was divided into nine sub-areas which were selected to have a degree of meteorological homogeneity. The model was run to create time series of grass potential evaporation for each sub-area.

PENSE is a somewhat similar system developed for Southern Region. The system uses the Penman-Monteith model for a land cover of grass, following as closely as possible the procedures in MORECS. The input data are the Met. Office monthly weather reports, from 1918 on, for ten sites. Missing data was infilled using regressions based on the station's position in terms of altitude and distance from the coast. The model was then run to create time series of potential evaporation at each of the ten sites. The system has been further developed to allow the calculation of potential evaporation for any location by synthesising a time series of driving variables at that location from the data from the three nearest climate stations.

5. THE FUTURE

This chapter assesses the methods described in Chapter 3, in terms of the Agency's requirements, and makes recommendations as to which methods should be adopted. The implications of these recommendations for data are then discussed and suggestions are made as to the strategies that the Agency could adopt.

5.1 Assessment of methods

The methods have been ranked according to nine criteria:

1. Robustness – can the physical principles and assumptions be justified against the actual processes and variation from site to site,

2. QA – can the input data and the output estimates be checked rigorously,

3. Ease of use – is little knowledge and expertise in the method required from users and, potentially, little effort required to assemble the data and carry out the necessary calculations,

4. Data requirements – does the method require little data and are these data readily available to Agency staff,

5. Accuracy – is the method accurate enough for the purposes of the Agency,

6. Daily time step – is the method capable of giving estimates of evaporation with a daily time step,

7. Cost – is the cost of implementing and using the method low,

8. Breadth of applicability – is the method capable of being applied to all the sites for all the purposes of the Agency,

9. Climate change – is the method capable of being used to assess the impact of climate change.

Each method has been awarded a score on a scale of zero to three against each of the criteria:

0 - does not meet the criteria in any way
1 - performs poorly against the criteria
2 - performs well against the criteria but there are some significant deficiencies
3 - performs well against the criteria with only minor deficiencies

The scores were awarded on the basis of discussions with a range of people with relevant expertise. The average score has been calculated and the results are presented in Table 5.1.

Table 5.1 Ranking of methods for estimating open water evaporation

Criteria Method	1	2	3	4	5	6	7	8	9	Average
Pan evaporation	1	1	2	2	1	1	2	2	0	1.33
Mass balance	3	3	1	1	2	1	1	0	0	1.33
Energy balance	3	3	1	1	3	2	1	0	0	1.56
Mass transfer	3	3	2	1	3	3	2	1	0	2.00
Combination	2	3	2	2	2	3	2	2	3	2.33
Equilibrium temperature	3	3	2	2	3	3	2	3	3	2.67
Empirical factors	1	2	3	3	1	3	3	2	3	2.33

Pan evaporation scores poorly because of its lack of rigour and the absence of data being acquired now. The poor scores of the mass balance and energy balance methods are because they are essentially site specific. The mass transfer method scores slightly better because it is less demanding in terms of input data but these data are still site specific. The three methods that score best are the combination equations, equilibrium temperature and empirical factors. The latter scores well because of the simplicity and breadth of applicability, but these are offset by the lack of physical rigour. The same average score is achieved by combination equations but for the opposite reasons, these are more physically rigorous but would cost more and would not be as easy to use. The highest ranking is achieved by the equilibrium temperature method. This method has a good breadth of applicability and is capable of delivering both a daily time step series and dealing with climate change scenarios. It is physically rigorous whilst requiring only meteorological data as input. Thus it is ranked higher than the combination equations because it uses the same input driving variables but is capable of taking into account the heat storage of the water body. A limitation of both the equilibrium temperature method and the combination equations is that they cannot take into account thermal stratification of the water body.

5.2 Recommendations

From the discussion above, it is clear that the Environment Agency's purposes could be achieved best by using the equilibrium temperature method and therefore it is recommended that it should be adopted. This method will take time and resources to implement and so there may be a role for the continued use of empirical factors in the short term. Effectively, the Agency staff are consistent in the method they use (empirical factors). More consistency between Regions could be achieved if they were to use the same evaporation data set and the same empirical factors. This is capable of being implemented now at a small cost.

5.3 Data requirement and availability

The following sections consider the data requirements and present availability of that data, should the Agency decide to use either the empirical factors method or the equilibrium temperature method. The assumption is made that the use of empirical factors would be a short term measure to establish consistency across the Agency before the use of the equilibrium temperature method is adopted. We are assuming that the Agency would wish to make maximum use of existing data sets and would not want to invest in any new data sets for the empirical factors method.

5.3.1 Empirical factors

The use of the empirical factors method requires a data set of time series of potential evaporation. The most widely available data set within the majority of the Agency's Regions is the MORECS grass potential evaporation. The data sets generally covers the period from 1961 to 1990 at weekly or monthly time steps, and often extends up to the present. The Northwest Region does not currently receive any MORECS grass potential evaporation data whilst the Northeast Region has it from 1988 to the present. Thames Region have data for a few squares from 1985. It should be noted that the values of grass PE issued weekly by the Met. Office are not necessarily the same as those obtained if MORECS is run at a later date. Differences may result if data from different stations are available.

To make use of the MORECS data set for estimating open water evaporation it will be necessary to apply some empirical factors. Since the basis of the MORECS model is the Penman-Monteith equation the empirical factors would ideally be appropriate to this equation. The only empirical factors that meet this criterion are those given by Allen *et al.* (1998). A further advantage of these factors is that they were recommended to the FAO by an international committee of experts. Therefore at the end of Phase 1 of the R&D project it was recommended that, for water bodies less than 2 m in depth, the MORECS grass PE values should be multiplied by a coefficient of 1.05. For deeper water bodies the MORECS grass PE should be multiplied by 1.25, for November, December and January to April, and 0.65 for the rest of the year. These should be applied after the data has been corrected to the altitude of the site of interest using the method described in Section 6.2.2. However, work carried out under Phase 2 showed these factors to be unsuitable (see Chapter 11).

In the case of applications for license abstractions, it is recommended that the highest annual evaporation from the period 1961 to 1990 be used. This 30 year period is recommended by the WMO as a standard for analysing rainfall records and it would be consistent to accept this period for the purposes of licensing. Currently, it is unclear whether this standard will change to 1971 to 2000 as this is subject to debate.

The procedure is not as simple for water balance studies and assessing the evaporation losses from wetlands. Extending the time series back prior to 1961 can be achieved using a time series of potential evaporation, solar radiation or air temperature data (in that order of preference) from a station in a similar climate and physiographic setting to the water body. The data must include a significant part of the time period covered by the MORECS data and extend back to cover the period for which the estimates are required. The data should be aggregated, if necessary, to the same time step as the MORECS data and a linear regression performed between the two data sets, using the values from the common time period. The coefficients from this regression can then be used to estimate the MORECS grass PE from the other time series. The data should be corrected for altitude. If it is necessary to disaggregate the resulting time series to daily values then this can be achieved by linear interpolation between the values, see Section 6.4.

5.3.2 Equilibrium temperature model

The equilibrium temperature model should be implemented in the form given by de Bruin (1982). In addition to the depth of the water body, the data required are time series of meteorological variables: net radiation (which can be derived simply from records of sunshine hours or incoming solar radiation), mean air temperature, mean relative humidity, the wet bulb temperature (which is usually recorded to allow the relative humidity to be calculated) and mean wind speed. Although daily values would ensure that the daily variability in evaporation is properly represented, this is not as much an issue with open water as with other land cover types due to heat storage reducing the amount of variability. de Bruin (1982) got good agreement with measured evaporation using a ten day time step. Whether a monthly time step could be used as successfully is unproven.

The availability of meteorological data within the Environment Agency is very variable (see Section 4.4.1, Table 4.1) ranging from the Thames Region which has none, to the South West which has ten stations providing data. This means that, in order to use the equilibrium temperature method, the Agency will have to invest in acquiring time series of historic meteorological data and will need to continue to add data to these to ensure that they are current. The simplest source would be to purchase the data from the Met. Office. However, this is likely to prove expensive. A source of monthly meteorological data is the monthly weather summaries that were published by the Met. Office until recently. These are available as paper records and so would need to be entered manually into computer files. Since the storage of heat by water bodies tends to 'smooth' out the seasonal variability of evaporation, compared to other land covers monthly values are acceptable. The day-to-day variability will be dominantly determined by the wind speed and humidity.

The equilibrium temperature is essentially a simple model and therefore could be programmed relatively easily using a spreadsheet. However, the input, verification, storage and spatial interpolation of the time series of driving data is a considerably more time consuming task and so it would be more appropriate if the model and data handling were combined in one system. It would be necessary for the Agency to commission an appropriate software package in order to keep the amount of time spent by Agency staff down to the same or less than they spend currently on the task of estimating open water evaporation. An existing package, PENSE, would make a very good basis for this as it already performs most of the data handling activities required and so the cost of modifying the program to include the equilibrium temperature model, rather than using a spreadsheet, is likely to be low. The largest cost that will be incurred is in obtaining the monthly values of the driving variables. These will need to be manually entered and then verified. This has in effect already been achieved for the Southern and Northwest Regions when creating PETCALC and PENSE and so an approximate cost per Region of this activity could be obtained from these. The number of stations required will be the same as any other requirement for estimating evaporation, i.e. it should be sufficient to represent the spatial heterogeneity of the meteorological variables.

There will be two further costs. The first is in training Agency staff how to use the new system. This will be low as the system should be relatively easy to use and so a half day per staff member is the most that is likely to be required. The second is to add new meteorological data each year if a current data set is to be maintained. This will involve entering data from the Agency's own station but it will need to be supplemented by buying data from the Met. Office which is likely to have a significantly higher cost. However, the data required will be the same for any other estimates of evaporation made by the Agency using a combination equation approach and the Agency should therefore consider the data purchases in this context. It would probably be more efficient if the Agency was to consider implementing a system for estimating open water evaporation as part of a general system for estimating evaporation for a range of purposes, e.g. groundwater recharge, rainfall-runoff modelling etc.

6. SPATIAL AND TEMPORAL VARIABILITY OF EVAPORATION

This chapter considers the spatial and temporal variability of open water evaporation in terms of the driving variables of evaporation.

6.1 Spatial variability of evaporation

Evaporation shows a significant variability across England and Wales. This is illustrated by the mean annual potential evaporation from grass calculated by MORECS, Figure 6.1, which is likely to be comparable to the variability of open water evaporation. The values shows a range in mean annual totals from 450 mm year^{-1} in the north west, to 620 mm year^{-1} in coastal areas of the south west. Upland areas are notable for reduced evaporation, typically by about 50 mm year^{-1} compared with adjacent, lowland areas. This map almost certainly underestimates the real spatial variability as it is based on the MORECS grid cells which only deal with mean conditions over a 40 x 40 km grid cell.

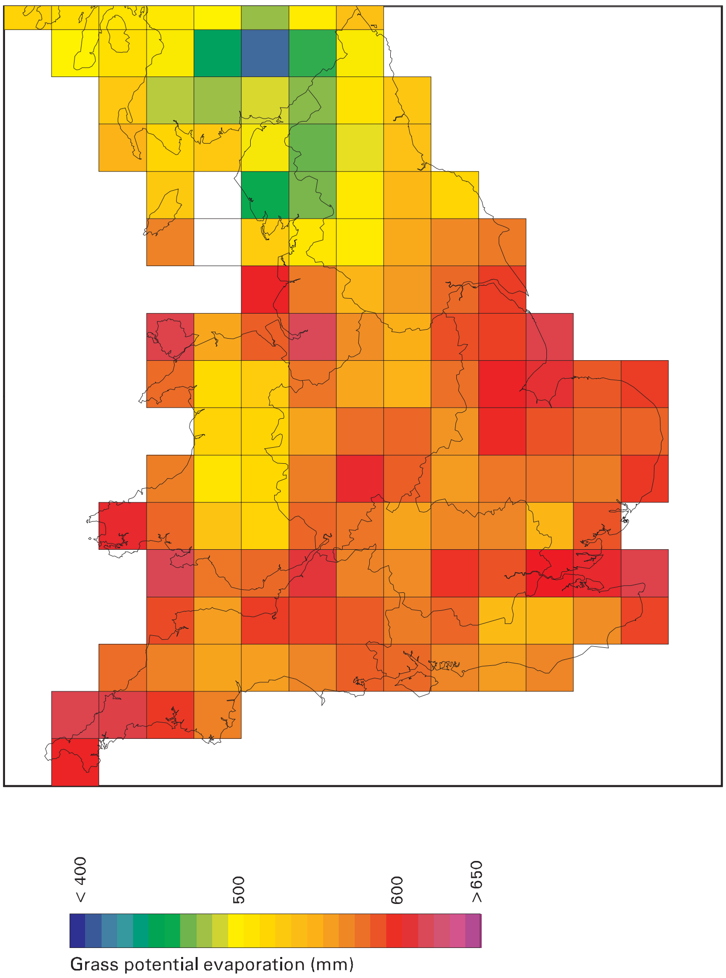

Figure 6.1 Mean annual MORECS grass potential evaporation (1961-1990).

For a given land surface, the spatial variability in evaporation is caused by variability in the driving variables, i.e. incoming radiation, wind speed, humidity and air temperature. The spatial variability of these will be considered in turn.

6.1.1 Incoming solar radiation

The dominant driving variable for evaporation is the incoming radiation which can be taken as being represented by the incoming solar radiation. The spatial variability of this has been analysed by Cowley (1978). There is a general decrease with increasing latitude, Figure 6.2, as a result of the reduction in solar elevation angles. The difference is less marked in summer than winter due to the longer period of daylight compensating for the reduction in fluxes with decreasing solar elevation angle. A consequence of this is that there is a limit to the latitudinal extent over which the data from a given station can be used when calculating evaporation. Allen *et al.* (1998) suggest a maximum of 50 km.

The remaining variability is essentially a function of the amount of cloud. Thus, incoming solar radiation is higher along a coastal strip, 5-10 km wide (MAFF, 1967) due to the relative absence of cloud and haze. There is a reduction in incoming solar radiation associated with upland areas, notably the Welsh uplands, the Pennines and the Lake District, but the effect is also detectable in the higher areas of the Southwest Peninsula. This is mainly as a result of the topography causing the air masses to rise and cool resulting in condensation causing increased probability of cloudiness in these upland areas.

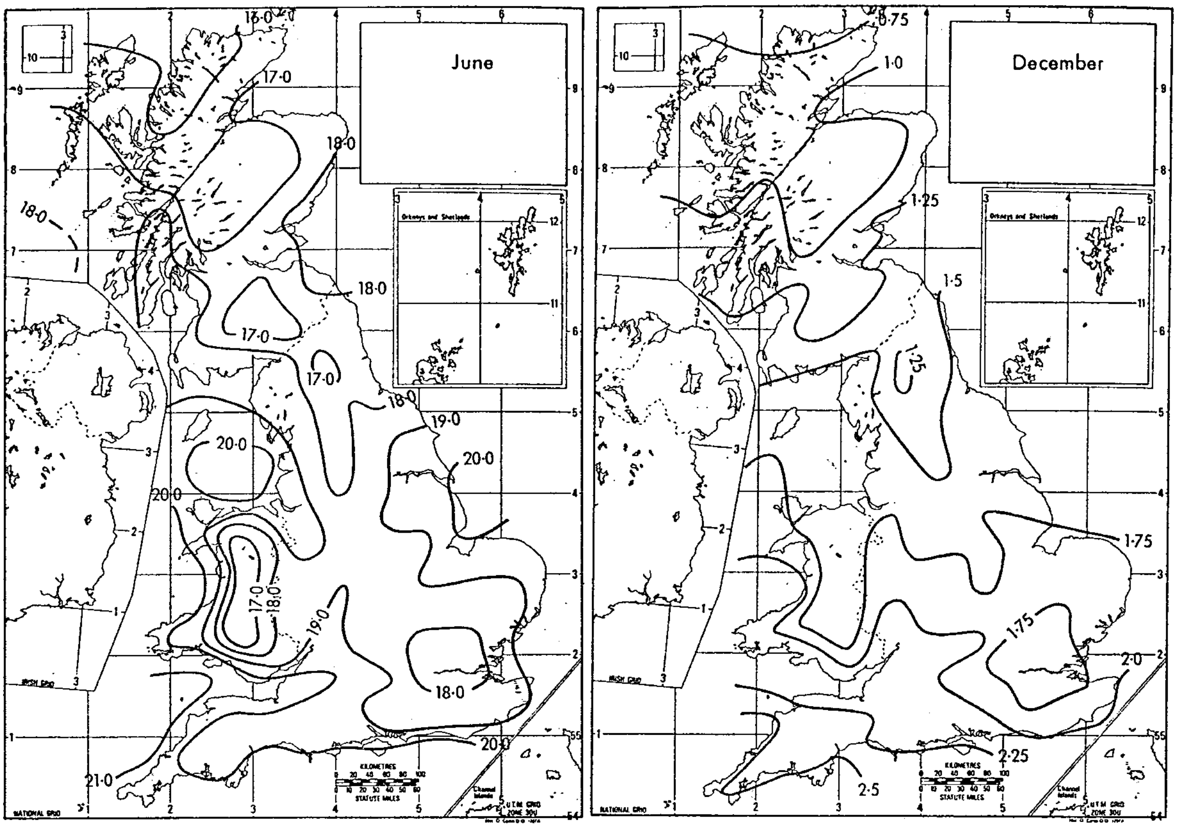

Figure 6.2 Distribution of mean daily incoming solar radiation (MJ m^{-2}) for June and December, 1966-75 (Cowley, 1978).

Shadowing due to topography may have an effect at a few sites in the uplands. This will occur if the water body is sited in an incised valley such that all, or a significant part of it, is in shade for a proportion of the day. The incoming solar radiation into the water body will therefore be reduced. It is more likely to occur in winter with lower solar elevation angles, but this is also the period when the incoming radiation is least which will serve to reduce the effect.

6.1.2 Air temperature

The mean daily air temperature for June shows a general reduction northwards, Figure 6.3. The major conurbations are marked by areas of higher temperature; about 1ºC above the surrounding areas. The upland areas of Wales, the Pennines and the Lake District tend to be slightly cooler than the surrounding lowlands even when adjustments are made for altitude. However, the changes are gradual which demonstrates that, provided the data are corrected for altitude, the spatial variability is not great during the summer.

Although there is warming inland in June, it is not particularly marked. In contrast, December is notable for pronounced temperature gradients towards the coast as a result of the warming effect of the sea. Inland, the effect of the uplands and conurbations is more marked but the gradients are still relatively gentle, suggesting that data can be extrapolated over several tens of kilometres

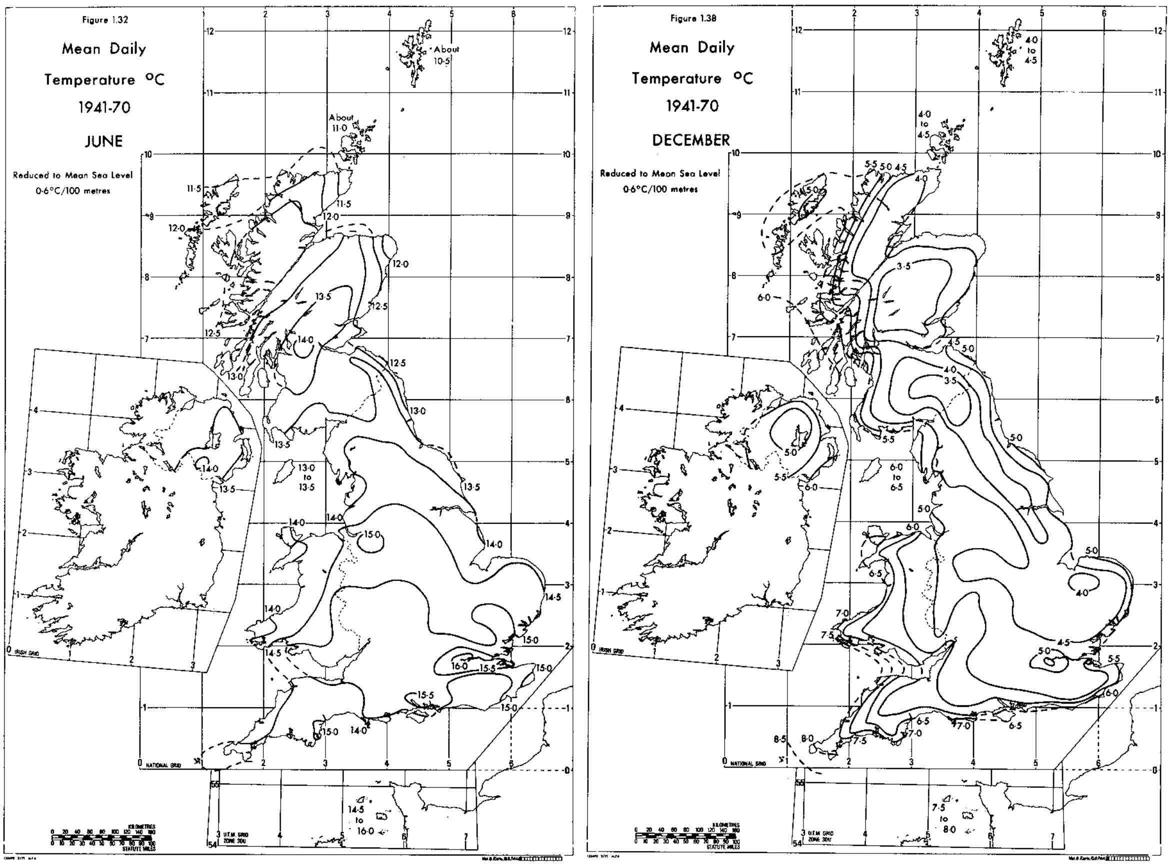

Figure 6.3 Distribution of mean daily air temperature (ºC) for June and December, 1941-70 (Met.Office, 1975a).

6.1.3 Wind speed

Mean hourly wind speeds are generally higher around the coast of England and Wales than inland, Figure 6.4. This effect only extends for 30 km at the most. Inland, there is remarkably little variability in the mean wind speeds. The uplands are marked by lower mean wind speeds but this should be treated with some caution. It may be an artefact due to the siting of the meteorological stations which tend to be in valley bottoms. Figure 6.4 is taken from a report (Met.Office, 1976) that emphasises the uncertainties in using the data for altitudes above 70 m. However, provided an adjustment is made for altitude, this suggests that, inland, the effect of the spatial variability on evaporation is low.

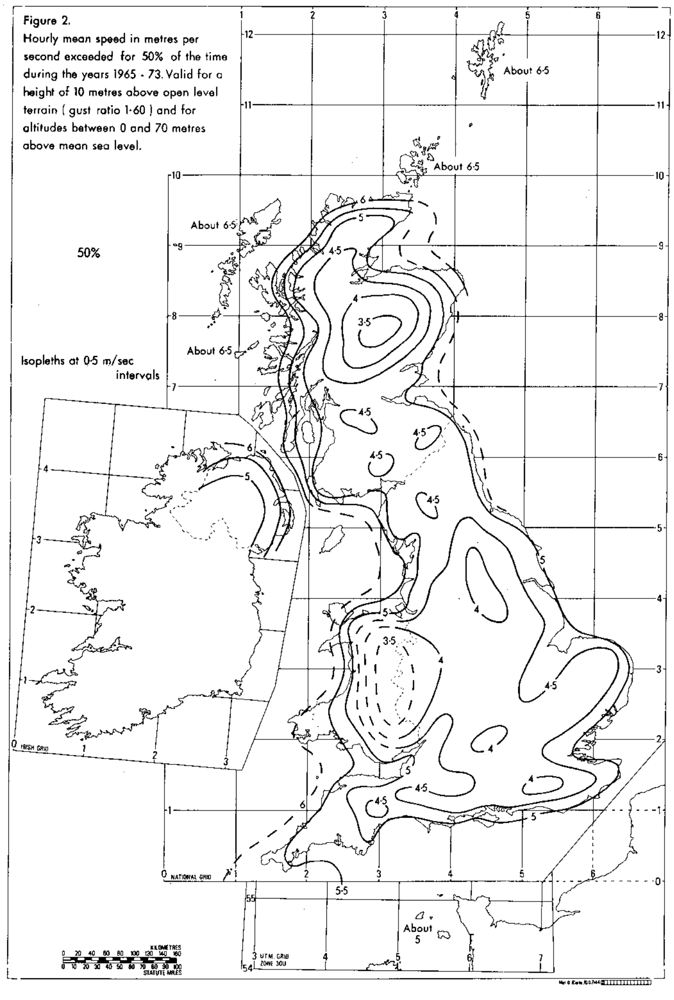

Figure 6.4 Distribution of hourly mean wind speed (m s^{-1}) exceeded for 50% of the time, 1965-73 (Met.Office, 1976).

6.1.4 Humidity

Relative humidity displays significant spatial heterogeneity across England and Wales, Figure 6.5. This map should be treated with some caution, particularly in the upland areas, as it is based on data from only 75 stations. There is a tendency for the relative humidity to be higher around the coast and it is possible to speculate that this might offset the effect of higher mean wind speed on evaporation in coastal districts. The east of England is generally an area of lower relative humidity whilst Wales and the west tend to have higher relative humidity. This is as a result of the UK generally being affected by air masses coming from the Atlantic Ocean and therefore having a high relative humidity in the west of the country. Passing over land tends to result in a warming of the air mass with an ensuing decrease in relative humidity eastwards. However, there is significant variability at a scale of a few tens of kilometres which suggests that extrapolating data over distances of this scale will result in increased errors in estimates of evaporation.

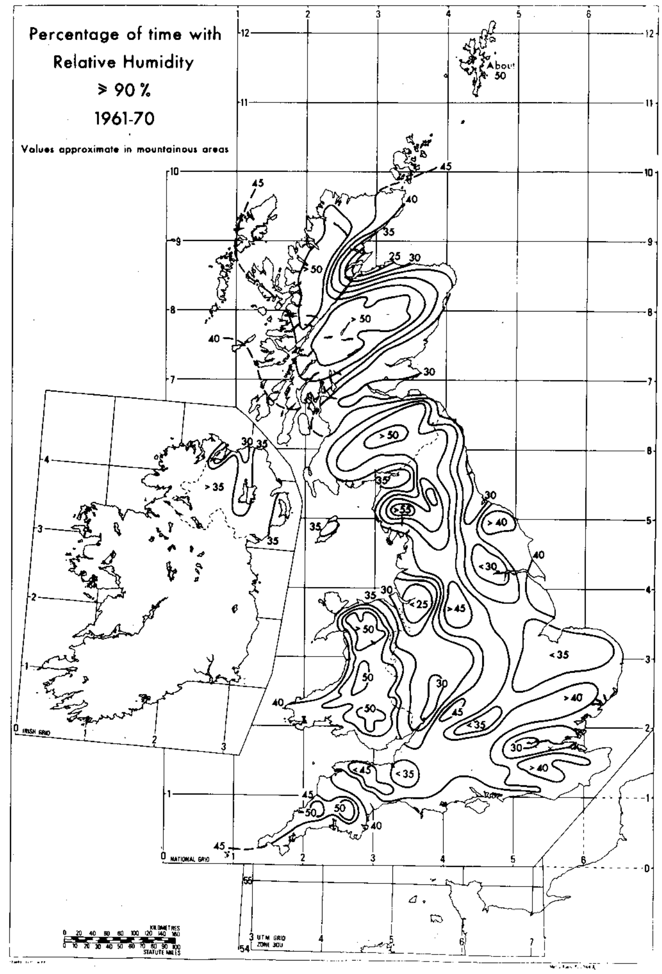

Figure 6.5 Distribution of percentage of time with relative humidity ≥90%, 1961-70 (Met.Office, 1975b).

6.1.5 Implications

The result of the spatial variability in the driving variables is that only small errors will be incurred extrapolating estimates of evaporation from the site of measurements in inland areas with subdued topography, e.g. East Anglia. However larger errors are likely to occur when extrapolating estimates of evaporation derived from sites within 5-10 km of the coast to inland locations, or vice versa, without appropriate corrections. Evaporation in this coastal strip is likely to be greater than adjacent inland areas because most of the driving variables have higher values.

Changes in the scale of the topography will also have a pronounced effect. Given that the prevailing winds in England and Wales are from the south-west, the effect is likely to be most pronounced on the western sides of upland areas due to the formation of clouds reducing the amount of incoming solar radiation. It is the spatial heterogeneity in the topography which often determines the spatial heterogeneity of evaporation. The consequence is that the errors are likely to be minimised if data for estimating evaporation are used from sites with a similar physiographic setting, i.e. altitude, aspect and slope.

For the practical purposes of estimating open water evaporation, this means that the effects of spatial heterogeneity can be minimised by using data from meteorological stations that are close to and in the same setting as the site for which the estimate is required. Data from stations greater than 50 km. away should never be used. The station(s) should be in the same climatic region, i.e. if the site for which an estimate is required is within 5-10 km of the coast then only data from meteorological stations within 5-10 km of the cost should be used. In hilly areas, if possible, the physiographic setting should be consistent, e.g. aspect, altitude band and exposure.

6.2 Adjustments for altitude

For correctness, adjustments for altitude should be applied to the driving variables before calculating evaporation. This is because the variations with altitude are different for each of the variables. In addition, the variables are often present in separate terms of a model. For example, the Penman-Monteith combination equation (see Section 3.5.1) can be viewed as having two terms, the first describing the input of energy from radiation, the second describing energy due to aerodynamics. Thus the net radiation occurs in the first part, whilst the wind speed and humidity occur in the second. Nevertheless, empirical coefficients to correct estimates of evaporation have been reported. Both approaches to corrections will be described in this section.

6.2.1 Adjustment of driving variables

The mean environmental lapse rates for air temperature and vapour pressure are based on physical principles and have been measured by the Met. Office for the UK. The average values given by Hough and Jones (1997) are –0.006ºC/m for temperature and –0.025 kPa/m for vapour pressure.

The variation of wind speed with altitude is not as well founded because a simple link through physical principles cannot be shown. Nevertheless, empirical relationships have been derived. Caton (1976) reported lapse rates of between 7 and 9 percent per 100 m for four sites with an altitudinal range of 220 m in the Midlands. Harding (Institute of Hydrology, pers. comm.) found a lapse rate of 0.006 m s^{-1}/m for 19 sites in northern England and southern Scotland with an altitudinal range of 830 m. The same analysis applied to four sites in mid-Wales resulted in a value of 0.005 m s^{-1}/m. Manley (1995) found a rate of between 0.5 and 0.008 m s^{-1}/m for 40 stations in the Agency's North West Region. The values reported by these studies are reasonably consistent and suggest that a value of 0.006 m s^{-1}/m would be appropriate generally. However, all the authors point out that there are both seasonal and regional variations. Whilst the net radiation is the dominant factor for the long term (monthly or greater) variation of evaporation, the wind speed is the dominant factor determining daily variability, except for shallow water bodies. The result is that errors in the wind speed will result in higher errors in daily estimates of evaporation than those for monthly estimates.

Reliable adjustments for variations of incoming solar radiation with altitude have not been reported. As discussed in Section 6.1.1, altitude does not have a direct effect on the incoming solar radiation. It is through the formation of clouds in response to topography that there is a causal link. The result is that other factors, such as the vapour pressure of the air mass, are also involved so any empirical relationship is only likely to be valid locally.

6.2.2 Adjustment of evaporation estimates

The driving variables used by MORECS can be thought of as applying to hypothetical weather stations, located at the centre of, and at the average elevation of, each square. Empirical monthly environmental lapse rates for MORECS grass potential evaporation have been derived by the Institute of Hydrology by analysing the within month variations in potential evaporation between MORECS grid cells as a function of elevation. (This was done on the project "Enhanced low flow estimation at the ungauged site", co-funded by NERC and the Environment Agency. The Agency project leader was Dr. R Grew, Regional Hydrologist, South Western Region.) An initial analysis showed that there was no firm evidence of the lapse rates being latitude dependent and so it was possible to derive relationships between altitude and potential evaporation for each month, Table 6.1. There is a strong seasonality to the lapse rates, hence the need to use monthly values. The mean elevation of each of the MORECS squares, calculated from the 50 m grid cell digital terrain model at IH, is given in Figure 6.6.

Table 6.1 Mean monthly lapse rates for MORECS grass PE

	Lapse rate (mm m^{-1})	Standard Error (mm m^{-1})
January	-0.0143	0.0012
February	-0.0140	0.0009
March	-0.0180	0.0015
April	-0.0237	0.0024
May	-0.0344	0.0038
June	-0.0314	0.0046
July	-0.0388	0.0061
August	-0.0411	0.0051
September	-0.0316	0.0028
October	-0.0225	0.0017
November	-0.0177	0.0015
December	-0.0136	0.0012
Annual	-0.3011	0.0276

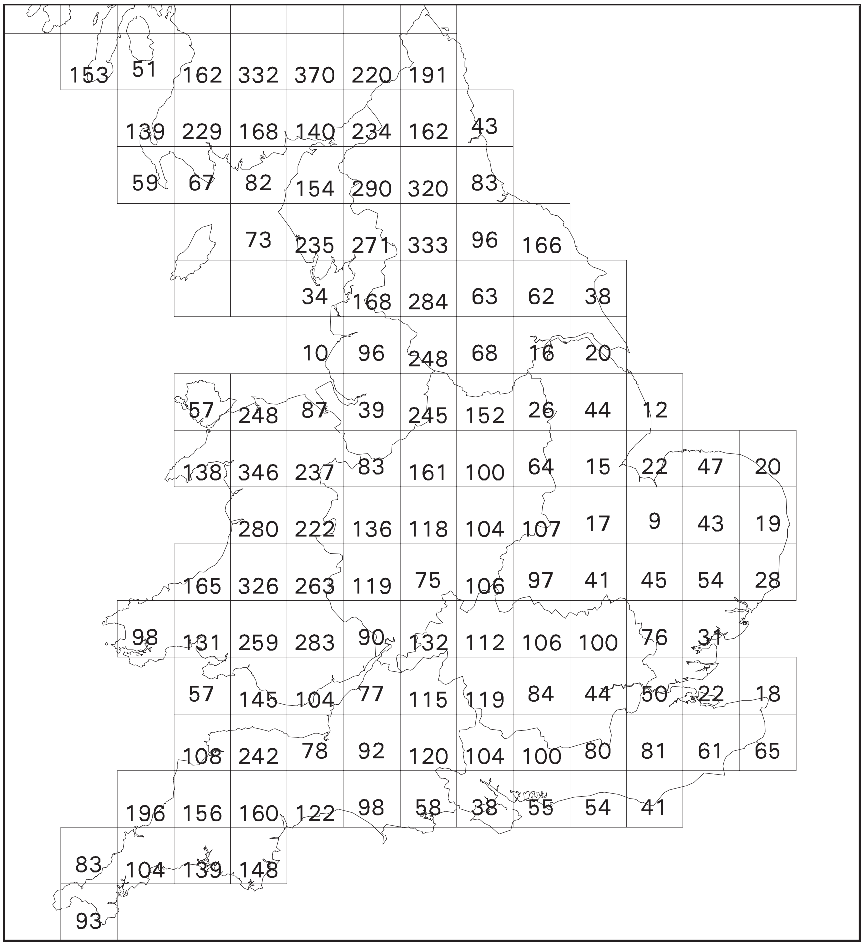

Figure 6.6 Mean elevation of the MORECS squares (m).

6.3 Effect of water body dimensions

There is little specific information in the literature on the effect of the water body dimensions on evaporation. Thus, although there is much speculation about the effect of size on meteorological variables such as relative humidity, air temperature and wind speed, there is little specific information and none that can be applied generally.

Some observations indicate that evaporation is enhanced over large water bodies. For example, Velalainen *et al*. (1998) observed, from direct micrometeorological and eddy correlation measurements over two lakes in Sweden, that evaporation rates were greater from the larger of the two lakes. They attributed this to the effect of increased wind speed more than compensating for the opposite effect of the increased humidity of the air associated with the larger distance travelled by the air over water.

Empirical and theoretical studies have been made of the evaporation from small areas (the oasis effect). As an airstream moves over an essentially homogeneous land cover, the energy exchange between the atmosphere and the land surface will tend to reach an equilibrium condition (principally by changes to the humidity of the air immediately above the surface). If the land cover over which the airstream is passing changes, then the energy exchange will also change to establish a new equilibrium. Shuttleworth (1993) points out that, although the effect of a change in surface cover propagates slowly up into the atmosphere, the adjustment of the surface evaporation rate into a moving airstream occurs quickly as the air passes on to a different type of evaporative surface (generally, within 5 to 10 m of the boundary between the two surfaces). This suggests that the surface characteristics of the surrounding land surface will dominate the evaporation rates for water bodies with a diameter less than 10 m. No definitive statement can be made as to what the effect will be as it will be dependent on the specific circumstances.

In practice, there is no means by which Agency staff can make a correction for the area of the water body when estimating open water evaporation.

The one factor affected by the water body size that is reasonably well documented is the thickness of the layer of warm water formed when thermal stratification takes places (i.e. the summer mixing depth). Straskraba (1980) gives a review of previous work and the results of some numerical modelling studies. Unfortunately, no information is given on the size of the water body, below which thermal stratification does not occur. Empirical relations are given for determining the summer mixing depth, z_{mix} (m), of lakes in various regions of the world. The relationship quoted for the Scottish Highlands is:

$$z_{mix} = 4.66 L_e^{4.66} \tag{24}$$

where L_e is the effective length of the water body, defined as half the product of the maximum length and maximum width, in kilometres. Straskraba found remarkable similarity of the relationships reported from lakes having similar climate conditions. However he points out that the mixing depth is greater for lakes at higher elevations which would imply that the relationship derived for the East German lowlands may be more appropriate for the lowlands of England and Wales:

$$z_{mix} = 4.72 L_e^{4.72} \tag{25}$$

These empirical relationships are shown in Figure 6.7. Both models predict, for water bodies with effective lengths less than 1500 m, comparatively little difference in the depth. However, for large water bodies, the data for the lowlands suggests that the summer mixing depth is less. Both empirical relationships become approximately linear for water bodies with effective lengths greater than 2500 m.

Smaller water bodies may develop thermal stratification to a greater depth than predicted due to the restricted fetch limiting the amount of turbulent mixing.

The above discussion assumes that advected energy is small in comparison with the radiation component. When this is not the case the depth of the thermocline (and the period during which stratification occurs) may be substantially different.

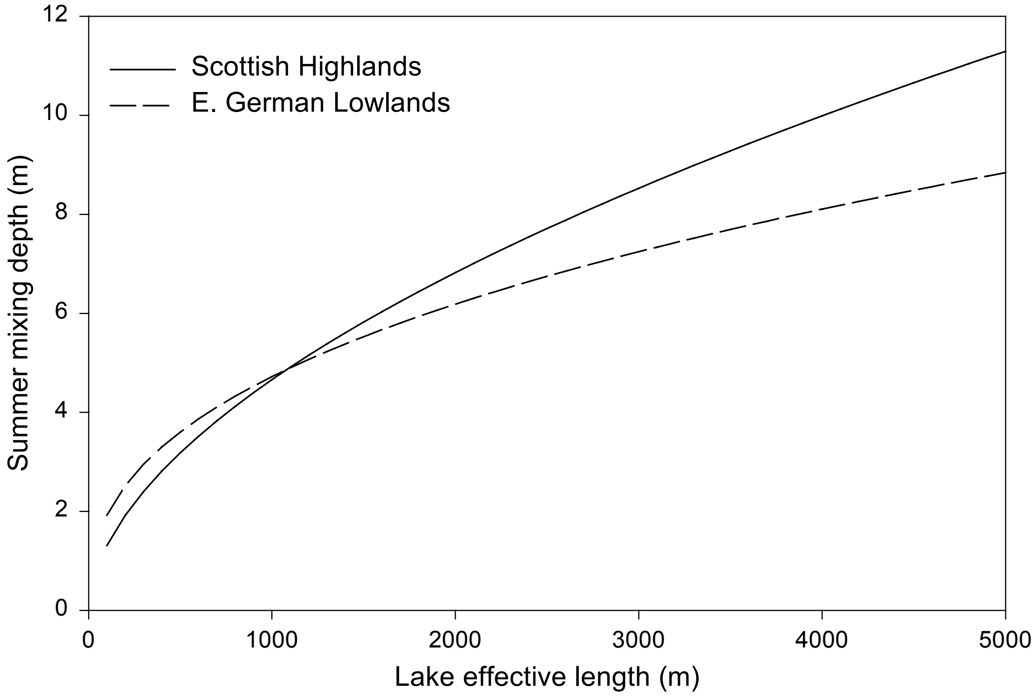

Figure 6.7 Relationship between size of water body and summer mixing depth.

The only means of predicting the development of thermal stratification are numerical hydrothermal models (Ahsan and Blumberg, 1999; Hostetler and Bartlein, 1990), however, these tend to be site specific.

The implication for models that assume a well mixed water body, e.g. the equilibrium temperature model, is that, during the period of thermal stratification, the model will underestimate the outgoing long wave radiation because it will be predicting a cooler surface temperature than would be observed. This will result in the model over estimating the evaporation due to the increased net radiation.

The conclusion is that there is no way, other than a spatially distributed numerical model, of taking the water body size into account when estimating evaporation. In a thermally unstratified water body, it is unlikely that the area of the water body has a significant effect, except when it is small (i.e. less than 10 m across), because it will be at the edges, where the air mass will interact with the land surface to change its characteristics, that the evaporation rates will vary. However, the depth of the water body will have an effect through changing the heat storage term, provided that the water body does not become thermally stratified then the equilibrium temperature model has the capability of handling this.

6.4 Temporal disaggregation

A variety of procedures are used to disaggregate a time series of evaporation data to shorter time steps, generally daily. Most procedures used are simple because the only information available is the time series of evaporation. However, if an appropriate additional time series is available, at the shorter time step, then it is possible to use a more sophisticated approach. The following discussion will assume that the required output is a time series with a daily time step.

6.4.1 Disaggregation with only one time series

In the absence of any additional time series, the simplest procedure is to divide each value by the number of days in the time step, e.g. divide by seven if the evaporation estimates are weekly. This method does not take into account any day-to-day variations in the driving variables on evaporation. It also assumes that there is no systematic variation, e.g. either increasing or decreasing, at the scale of the time step of the data. Hence it is not appropriate to input data with monthly or longer time steps.

A method more appropriate to data with longer time steps is to use linear interpolation between the centre points of adjacent time steps, e.g. mid-month dates. This allows seasonal trends to be taken into account. An improvement is to fit simple curves through a few adjacent values of the time series. Given that neither procedure reproduces the daily variability, the increase in complexity of using a curve rather than linear interpolation is rarely justified.

The ability of the two simple interpolation procedures to reproduce the daily variability can be demonstrated using a daily time series of meteorological data from Wallingford (51º36.1'N, 1º6.7'E, altitude 45 m A.O.D.) covering the period 1972 to 1998. The evaporation from open water has been calculated using the Penman-Monteith combination equation (see Section 3.5.2) assuming the net radiation is equal to the available energy, i.e. heat storage is negligible. The resulting time series was aggregated to monthly values. These were then used to generate daily time series using linear interpolation and the average daily evaporation which were compared with the original daily time series of evaporation using a series of statistical measures, Table 6.2.

Table 6.2 Comparison of disaggregated daily evaporation using simple methods

Time series	RMS error (mm)	Mean (mm)	Standard deviation (mm)	Minimum (mm)	Maximum (mm)
Daily average	0.902	1.34	0.679	1.57	3.54
Linear interpolation	0.896	1.34	0.676	1.57	3.54
Original		1.34	0.680	-2.23	12.74

The use of linear interpolation shows a small, but significant, reduction in the root mean square (RMS) error when compared with the original time series. Both methods result in time series whose mean is not significantly different to the original. However, although both methods result in time series whose standard deviations compare favourably with the original, they poorly reproduce the range of values. This is unlikely to be a significant problem over a long time period, similar to that used in this analysis, but may be over short time periods, e.g. one year. Care should be taken in interpreting the results of this analysis in the context of different climates or conditions where the heat storage in the water body is significant.

6.4.2 Disaggregation with additional daily time series

A more sophisticated approach is to use a daily time series of another meteorological variable, e.g. average daily air temperature or average daily wind speed, to control the disaggregation. However, this requires that there are daily time series, of open water evaporation and of the meteorological variable, that cover the same period for several years representative of the general climate conditions. The objective of the method is to use the meteorological variable to quantify the day-to-day variability of evaporation around the mean daily evaporation for a given time period, typically a month. Where the time period of the evaporation time series that is to be disaggregated is less than a year, it is recommended that the analysis to establish the relationship between the variability of the daily evaporation and the meteorological variable should be carried out on the basis of months to allow seasonal variations to be catered for. The procedure is:

1. Using the daily time series, calculate the mean daily evaporation and the mean daily value of the meteorological variable for each month.

2. Using the daily time series, calculate the departure from the mean daily value for that month of each daily value of evaporation and the meteorological variable.

3. Carry out a linear regression between the departures from the monthly mean of the evaporation (dependant variable) and the meteorological variable. This should be done for all the values available for each month of the year, i.e. use all the available values for January from all the years and repeat for each subsequent month. The result will be 12 values of slopes and intercepts, one pair for each month.

4. Using the aggregated evaporation time series and the daily time series of the meteorological variable for the same period, calculate the mean daily evaporation and the mean daily value of the meteorological variable for each month.

5. Calculate the time series of disaggregated daily values of evaporation. To do this, subtract the mean daily value of the meteorological variable of that month, $\overline{V}_m$, from the value of the meteorological variable, V_i, for each day. Calculate the departure from the mean daily evaporation using the appropriate values from the linear regression for that month for each day (the slope g_j and the intercept c_j for the j^{th} month of the year). Add these to the monthly mean daily evaporation, $\overline{E}_m$, i.e. to estimate the evaporation on day i, E_i:

$$E_i = \overline{E}_m + g_j (V_i - \overline{V}_m) + c_j \tag{26}$$

Table 6.3 Comparison of disaggregated daily evaporation using an ancillary daily time series

Time series	RMS error (mm)	Mean (mm)	Standard deviation (mm)	Minimum (mm)	Maximum (mm)
Air temperature	0.927	1.30	0.627	-1.05	4.12
Wind speed	0.775	1.30	0.623	-8.49	10.78
Original evaporation		1.34	0.680	-2.23	12.74

The likely accuracy that can be achieved using this method is illustrated by a similar analysis to that described in Section 6.4.1, using the same daily evaporation time series. The method was applied using two different meteorological variables, average daily air temperature and average daily wind speed. The results, Table 6.3, show that using the daily average air temperature results in a higher RMS error than simple linear interpolation. However, it is more successful at reproducing the range of values. The use of a time series of daily average wind speeds results in a significant reduction in the RMS error, suggesting that much of the day-to-day variability in this time series is due to wind speed.

7. DATA SETS FOR TESTING METHODOLOGIES

This Chapter describes data sets that have been identified as being potentially useful for testing the recommended methods of estimating open water evaporation. Full details of the data sets are given in Annexe D.

Ideally, a data set would consist of a time series of evaporation that has been (or could be) calculated for a period of at least one year using either direct measurements, the energy balance or the mass balance methods. Data from a climatological station at the site would also be available. Unfortunately there are very few of these so data from evaporation pans have been included as these may be of use in testing some aspects of the methods.

7.1 Measurements from water bodies

The most exhaustive study, in the UK, is that of Lapworth (1965) who measured evaporation from two man-made reservoirs at Kempton Park, near London. The study used the mass balance method and generated monthly estimates of open water evaporation from 1959 to 1962. Comparisons were also made with pan evaporation measurements and Penman evaporation.

Another data set based on mass balance measurements is given by Crowe (1974). These are for the Farmoor reservoir, in Oxfordshire. However, they are only for a seven month period and so are of less use as they do not represent an annual cycle.

Between 1966 and 1970, the Central Electricity Generating Board (CEGB) carried out an extensive study of cooling from open water surfaces based on measurements at Lake Trawsfynydd (McMillan, 1973). Estimates of evaporation are not given as the objective of the study was to estimate the combined heat loss (latent and sensible heat fluxes). The situation is complicated by significant advection from the power station at the site. However, the data listed in the report include weekly values of surface water temperature, net radiation, plus the difference between the values at the water surface and a height of 3 m for the air temperature and vapour pressure from 1 May 1969 to 26 December 1970. These would allow evaporative losses to be calculated using a bulk transfer model and thus could be used to test the values estimated by other methods. However, there must be considerable reservations about using these data as the large advective term in the energy balance makes the situation unique and thus not representative of the conditions generally of interest to the Agency.

The Institute of Hydrology carried out measurements of evaporation, using a Bowen Ratio system (see Brutsaert, 1982), from a site in Somerset where a reed bed was being re-created. The measurements cover the months of April to September inclusive in 1996 and April to June 1997. The reed bed was poorly established in 1996 so these data may represent open water conditions. The water is shallow so the lack of a full annual cycle may not be a limitation.

7.2 Evaporation pan measurements

Pan evaporation was recorded at a number of the Met. Office stations and therefore there are long term records available for a variety of sites across England and Wales. Holland (1967) lists 39 stations operating in the late 1960s. However, most measurements ceased during the 1980s as it was perceived that the errors involved in applying pan evaporation data to other land surfaces and locations were at least as great as those involved in estimating evaporation from meteorological data. Measurements have continued to be recorded at some sites and, although these data are no longer archived at the Met. Office, are often available from the measurement site.

Measurements of pan evaporation are available from sources such as research organisations (e.g. the Institute of Hydrology and the Institute of Arable Crop Research stations) and some university departments. However, these data are often not readily accessible. In addition, there has been a trend to discontinue these measurements so the data often does not extend into the 1990s. Nevertheless ten data sets have been documented, the details of which can be found in Annexe D.

8. PHASE 1 CONCLUSIONS AND RECOMMENDATIONS

The conclusions and recommendations presented here were relevant at the completion of Phase 1. Many have been superseded by the results of Phase 2, and these are clearly identified.

The objectives of this project are:

- evaluate current methods of estimating open water evaporation;
- recommend the best available practicable methodologies for producing robust estimates;
- assess the associated uncertainty of these methodologies.

Phase 1 is concerned with the first two.

8.1 Conclusions

Seven methods of estimating open water evaporation were identified; pan evaporation, mass balance, energy budget models, bulk transfer models, combination models, the equilibrium temperature method and empirical factors.

Pan evaporation and empirical factors are very similar in that they employ factors to convert ‘standard’ time series of evaporation into estimates of open water evaporation for specific water bodies. As such there is a lack of physical rigour and considerable uncertainty as to which values to use for the factors.

The mass balance method requires considerable investment in collecting data and is not appropriate if the evaporation losses are comparable in size to other changes in storage. The energy balance method is generally accepted as the most accurate method but requires considerable resources to accurately measure all the components of the balance. Nevertheless, the two balance methods are the only methods that consider the water body as a whole. In application, the other methods are one-dimensional and are assumed to be representative of the water body.

The mass transfer method requires measurements of surface temperature and is sensitive to errors in the vapour pressure data and the formulation of the wind function. The combination models have gained wide acceptance but, when applied to open water deeper than 0.5 m and for time intervals less than a year, it is necessary to use the available energy, rather than the net radiation, in order to take the heat storage into account. This requires measurements of surface temperature.

The equilibrium temperature method requires the same meteorological data as combination models but can take the heat storage into account, provided that the water body does not become thermally stratified. As such it combines physical rigour with a requirement for time series of meteorological data that are readily available.

The empirical factors method is used to produce estimates of open water evaporation in all Areas and Regions. However, the values of the factors and data sets employed vary between Regions and may vary between Areas. Thus, although there is a consensus on the method, there is little on how it is used.

A ranking of the seven methods of estimating open water evaporation against nine criteria established that the equilibrium temperature method would best serve the Agency's purposes. The use of empirical factors and the combination models were ranked equal second.

The spatial variability of the meteorological variables that drive evaporation is strongly influenced by proximity to the coast. However, inland, the spatial variability of wind speed and air temperature is low while incoming solar radiation and relative humidity show significantly more variability. Topography has a very strong effect on the driving variables, either directly, in terms of the lapse rates of air temperature and vapour pressure, or indirectly, through the formation of clouds affecting the amount of incoming solar radiation.

Correcting for the effect of altitude on the driving variables can be achieved except for the incoming solar radiation. Empirical corrections to evaporation estimates for altitude have also been found, although this is not as physically rigorous as correcting the driving variables.

The size of the water body has an effect on evaporation rates in several ways. There is evidence that the rate of evaporation over water is enhanced due to increased wind speed resulting from the smoother surface. The effect of the transfer of the airstream from the surrounding land surface to the water surface extends a maximum of ten metres from the edge. No means of taking these effects into account has been found. The main impact of the size of the water body in England and Wales is the development of thermal stratification. The maximum depth of the warmer surface layer is a function of the surface area of the water body. However, studies of thermal stratification published in the literature have tended to be site specific and so there is a lack of information on how thermal stratification affects the evaporation rates in general terms.

There is very little discussion in the literature on the impact of errors and uncertainties on estimates of open water evaporation. This makes it difficult for the Agency to assess the accuracy of the estimates of open water evaporation and then to analyse the effect this may have on other information derived from these data, e.g. catchment water balance models. Phase 2 of this project aims to address these issues by quantifying as far as possible the accuracy of the recommended methods (empirical factors and the equilibrium temperature method) and the accuracy of interpolating the driving variables.

8.2 Recommendations

- In the short term, the Agency should continue to use empirical factors to estimate open water evaporation but should use consistent data sets and factors. This could be achieved by using MORECS grass potential evaporation as the data set.

- At the end of Phase 1 it was recommended that, for water bodies less than 2 m in depth, the MORECS grass PE values should be multiplied by a coefficient of 1.05. For deeper water bodies the MORECS grass PE should be multiplied by 1.25, for November, December and January to April, and 0.65 for the rest of the year. However, Chapter 11 in Phase 2 shows that these factors are unsuitable.

- The MORECS data should be corrected for altitude using the empirical factors given in Section 6.2.2. (It should be noted that the values of grass PE issued weekly by the Met. Office are not necessarily the same as those obtained if MORECS is run at a later date. Differences may result if data from different stations are available.)

- In the longer term, the Agency should adopt the equilibrium temperature method for estimating open water evaporation. This will require an investment in meteorological data to be input into the model. This action would be more cost effective to the Agency if it was embedded in procedures for estimating evaporation for other purposes.

- There is a lack of information on the impact of errors and uncertainty on estimates of open water evaporation. Therefore Phase 2 should proceed and consist of:

  1. The estimates of open water evaporation produced by the empirical factors method and equilibrium temperature method should be tested against the measurements that were made at the Kempton Park reservoirs to establish their absolute accuracy.

  2. The altitudinal corrections should be tested by estimating the open water evaporation at a site using data from that site and data from another site that had been corrected for altitudinal difference. This would be repeated for a several sites across the England and Wales representing a variety of altitudes, latitudes and physiographic settings.

  3. The accuracy of the empirical factors method should be established by comparing it with estimates produced using the equilibrium temperature method for a number of stations representing a variety of climate conditions in England and Wales.

- Further research is required into how thermal stratification affects evaporation rates. This could best be achieved by a sensitivity study using a numerical hydrothermal model. This would allow the impact of the water bodies dimensions on evaporation rates to be determined for a range of climate conditions. The model would need to be tested against real data.

PHASE 2

9. INTRODUCTION TO PHASE 2

The following five chapters describe the work carried out under Phase 2 of the project, following the recommendations made at the end of Phase 1. It should be noted that the PENSE system used by the Southern Region is a simulation of MORECS and as such it need not be considered separately.

9.1 Absolute accuracy of estimates of open water evaporation

Estimates of open water evaporation, made using the equilibrium temperature method, combination model and empirical factors, have been tested against the seven year record of monthly evaporation from a reservoir at Kempton Park. The estimates of open water evaporation have been compared with the measured evaporation from the reservoir and the accuracy of the estimates determined. Two sets of empirical factors have been considered, the FAO values recommended for use by the Agency in the Phase 1 report and the original factors given by Penman (1948) (currently used by several Regions).

It is not clear, from the literature, what impact the time step used in the equilibrium temperature model has on the accuracy of the estimates. Therefore, the equilibrium temperature model has been run with time steps of 1, 5 and 10 days and one month. The resulting values have been aggregated to monthly values and compared with each other to determine whether any significant difference occurs in the resulting estimates. In addition, the results have been compared with the observed values from Kempton Park to determine the effect of the time step on the absolute accuracy. This is described in Chapter 11.

9.2 Optimum values for empirical factors

The opportunity has also been taken to calculate the optimum monthly values for the empirical factors (used to determine open water evaporation rates from evaporation rates for other surfaces). These have been calculated using the observed evaporation data at Kempton Park by comparing the values with simulated MORECS grass PE data. The exercise has been repeated for PE data from PETCALC. The results can be found in Chapter 12.

9.3 Sensitivity analysis

A sensitivity analysis has been carried out on the equilibrium temperature model, simulated PETCALC and simulated MORECS to investigate the impact of measurement errors in the input data on the accuracy of estimates of open water evaporation. This involved determining the typical errors in measuring the driving variables and then calculating the evaporation using time series of driving variables where each variable has the error added and repeating this with it subtracted. It has been assumed that the actual data represents accurate data and so the impact on daily estimates can be assessed. This is effectively investigating a worst case scenario as it is unlikely that a measured value will be consistently in error by the maximum.

In addition, the impact of uncertainties in the values for the three parameters (albedo, roughness length and water depth) required for the equilibrium temperature model have been investigated. These are described in Chapter 13.

9.4 Effect of differences in climate on estimates of open water evaporation using empirical factors

This has been investigated by using time series of driving variables from meteorological stations representing a range of conditions in England and Wales. The selection of stations has been made to reflect the broad climatic regions of England and Wales, i.e. Southeast England (high solar radiation and little maritime influence), the Southwest Peninsular (high solar radiation and strong maritime influence), upland Wales (windy, with extensive cloud cover), the Northwest (less solar radiation and prevailing airmasses from the Atlantic) and the Northeast (less solar radiation and prevailing airmasses having crossed upland areas). The choice of stations was also influenced by data availability and cost.

The equilibrium temperature model has been used to calculate time series of estimates of open water evaporation (the water depth at Kempton Park has been assumed to apply). These have been assumed to be accurate for a water body at the location of the meteorological station. The empirical factors method has been used to produce time series estimates of the open water evaporation. These have been compared with the ‘accurate’ time series and the differences analysed to identify any impact of climate, considering both annual totals and seasonal variations. The results can be found in Chapter14.

9.5 Accuracy of evaporation estimates after correcting for altitude

The accuracy has been assessed of evaporation estimates calculated using driving variables corrected for altitude, and of estimates of evaporation corrected directly for altitude as recommended in the Phase 1 report. This has been achieved by calculating time series of the open water evaporation using the meteorological data from one station, which have been assumed to be an accurate estimate of open water evaporation for that site. Time series of open water evaporation have then been calculated using the meteorological data from each of the other stations, correcting the driving variables and the estimated evaporation for altitude.

The potential impacts of differences in the solar radiation have been determined by using the solar radiation recorded at the site of the ‘accurate’ evaporation and that at the other stations.

The simulations have been carried out for each of the stations in turn and the results statistically analysed to quantify the likely errors involved in using meteorological data not recorded at the site where the estimate of evaporation is required. The results can be found in Chapter 15.

10. THE METHODS TESTED

Five methods will be discussed in the following chapters. They are defined here in detail.

10.1 Penman (1948)

Penman (1948b) made a classic study of evaporation using measurements from brick lined pits at Rothamsted in Hertfordshire. He succeeded in combining the thermodynamic and aerodynamic aspects of evaporation into a single equation. For daily evaporation this is:

$$\lambda E = \frac{\Delta(R_n - N) + \gamma 6.43(1.0 + 0.536 u_z)(e_a^* - e)}{\Delta + \gamma} \tag{27}$$

where λ is the latent heat of vaporization ≈ 2.45 (MJ kg^{-1}), E the evaporation rate (mm d^{-1}), Δ the slope of the temperature-saturation water vapour curve at air temperature (kPa ^{o}C^{-1}), R_n the net radiation (MJ m^{-2} d^{-1}), N the change in heat storage of the water (MJ m^{-2} d^{-1}), γ the psychrometric constant (kPa ^{o}C^{-1}), u_z the wind speed at height z_r (m s^{-1}), e_a^* the saturated vapour pressure at screen height (kPa) and e the vapour pressure at screen height (kPa). The net radiation is calculated as:

$$R_n = K^{\downarrow}(1 - \alpha_s) + L^{\downarrow} - p\sigma(T_a + 273.13)^4 \tag{28}$$

where $K^{\downarrow}$ is the incoming short-wave radiation (MJ m^{-2} d^{-1}), α_s the short-wave albedo of the water surface, $L^{\downarrow}$ the incoming long-wave radiation (MJ m^{-2} d^{-1}), p the cloudiness factor, σ the Stefan-Boltzman constant = 4.9 x 10^{-9} (MJ m^{-2} ºC^{-4} d^{-1}) and T_a the air temperature at the screen height (ºC). The third term in this equation is the outgoing thermal radiation and is estimated using the air temperature, because measurements of the surface temperature are not generally available. The implicit assumption is that there is little difference between the air and surface temperatures. This may be true of a land cover of vegetation well supplied with water but is less likely to be true for a body of water because of the inertia in the water temperature resulting from the larger heat storage component. It should be noted that it is important to calculate the incoming and outgoing long-wave radiations using the methods given by Penman (1948). This is discussed in Annexe G.

The aerodynamic term (the second in the numerator) of the Penman model contains several empirical constants. These should be considered strictly to apply to the conditions for which they were determined (i.e. the climate at Rothamsted and with poor exposure, as at the experimental site, and for very small bodies of water). Nevertheless, these constants have been used in numerous studies.

A variety of forms of the Penman model have subsequently appeared in the literature, generally with a modified aerodynamic term. The majority of these have been in response to using the model for land surfaces other than water, indeed, the Penman and Penman-Monteith models are most commonly employed for estimating the evaporation from short crops such as wheat or grass. As a result, confusion has arisen in the literature as to what precisely is meant by the Penman model. In this report, it will be consistently referred to as the Penman (1948) model in order to make the definition explicit.

10.2 Penman-Monteith

Monteith (1965) introduced greater physical rigour into the aerodynamic term of the Penman model by incorporating the concept of resistances, in particular, the aerodynamic resistance, r_a, to produce the Penman-Monteith model. In the form for water and daily evaporation the equation is:

$$\lambda E = \frac{\Delta(R_n - N) + 86400\rho_a c_p (e_a^* - e)d / r_a}{\Delta + \gamma d} \tag{29}$$

where ρ_a is the density of air = 1 (kg m^{-3}) and c_p is the specific heat of air = 0.001013 (MJ kg^{-1} ^{o}C^{-1}). The constant of 86400 (the number of seconds in a day) is required in order to keep the units of the different terms in the equation consistent. The aerodynamic resistance is defined as:

$$r_a = \frac{\log(z_r / z_0)^2}{k^2 u_z} \tag{30}$$

where k is von Karman's constant = 0.41, z_r the height of the measurements above the surface (m) and z_0 is the roughness length for momentum and water vapour (m). The roughness lengths are assumed to have the same value because water is a comparatively smooth surface.

The variable d was introduced by Monteith (1981) in order to correct for the use of the air temperature, rather than the surface temperature, in calculating the outgoing long-wave radiation component of the net radiation:

$$d = \frac{1 + 4\sigma(T_a + 273.13)^4 r_a}{86400\rho_a c_p} \tag{31}$$

10.3 Equilibrium temperature

In both the Penman and Penman-Monteith models given above, the variable that is generally not measured is the change in heat storage, N. When the water is shallow (less than 1 m), setting this to zero is unlikely to introduce any significant errors. However, as the water depth increases the errors will become significant at time scales less than a year. Measurements of the water temperature can be used to quantify this variable but these will rarely be available to Agency staff. However, in the absence of such measurements, Edinger *et al.* (1968) introduced a model based on the concept of an equilibrium temperature, T_e and an associated time constant, τ (days). The equilibrium temperature is the temperature towards which the water temperature is driven by the net heat exchange, i.e. when the water is at equilibrium temperature the net rate of heat exchange is zero. From this he was able to derive an expression for the temperature of a well-mixed body of water as a function of time and water depth. Thus the change in heat storage can be calculated. The time constant is defined as:

$$\tau = \frac{\rho c z}{4\sigma(T_n + 273.13)^3 + \lambda f(u)(\Delta_w + \gamma)} \tag{32}$$

and the equilibrium temperature as:

$$T_e = T_n + \frac{R_n^*}{4\sigma(T_n + 273.13)^3 + \lambda f(u)(\Delta_w + \gamma)} \tag{33}$$

where ρ is the density of water = 1000 (kg m^{-3}), c the specific heat of water = 0.0042 (MJ kg^{-1} ºC^{-1}), z the depth of the water, T_n the wet bulb temperature (ºC) and Δ_w the slope of the temperature-saturation water vapour curve at the wet bulb temperature (kPa ºC^{-1}). The wind function, $\lambda f(u)$, is that of Sweers (1976):

$$\lambda f(u) = 0.864(4.4 + 1.82u_z) \tag{34}$$

and the net radiation, if the surface were at the wet bulb temperature, is calculated using:

$$R_n^* = K^{\downarrow}(1 - \alpha_s) + L^{\downarrow} - p(\sigma(T_a + 273.13)^4 + 4\sigma(T_a + 273.13)^3(T_n - T_a)) \tag{35}$$

de Bruin (1982) showed that the water temperature on day j, $T_{w,j}$, could be calculated as:

$$T_{w,j} = T_e + (T_{w,j-1} - T_e)e^{t/\tau} \tag{36}$$

where t is the model time step in days. Thus the change in heat storage is given by:

$$N = \rho c z(T_{w,j} - T_{w,j-1}) \tag{37}$$

Estimates of the heat storage made in this manner can then be used in either the Penman (1948) or Penman-Monteith models. The latter is preferred due to its greater physical rigour. Furthermore, since the temperature of the water surface has been estimated, the net radiation can be calculated without assuming that the surface and air temperatures are the same or applying a correction to the Penman-Monteith model. The net radiation is then:

$$R_n = K^{\downarrow}(1 - \alpha_s) + L^{\downarrow} - p(\sigma(T_a + 273.13)^4 + 4\sigma(T_a + 273.13)^3(T_{w,j-1} - T_a)) \tag{38}$$

and the evaporation can be estimated using the Penman-Monteith model in the form:

$$\lambda E = \frac{\Delta(R_n - N) + 86400\rho_a c_p(e_a^* - e)/r_a}{\Delta + \gamma} \tag{39}$$

FORTRAN90 code for the equilibrium temperature model can be found in Annexe C.

10.4 Penman empirical factors

Penman (1948b) gives factors to convert evaporation rates from "turf with a plentiful water supply" to an open water surface exposed to the same weather conditions as:

Midwinter (November – February)	1.67
Spring and autumn (March – April, September – October)	1.43
Midsummer (May-August)	1.25

These values were derived from measurements of evaporation, for two years, at one site (Rothamsted Experimental Station) in southern England using cylinders 0.76 m in diameter and 1.83 m deep. (It will be appreciated that the size of the these cylinders is untypical of the water bodies of interest to the Agency.) The cylinders were filled either with water or soil on which mature grass, which was kept well watered, was growing. This enabled Penman to calculate the average difference between the evaporation from the two different surfaces. At the time, the site was criticised as having poor exposure. In addition, the empirical factor for the winter months is derived from measurements at Fleam Dyke, in Cambridgeshire.

10.5 FAO-56 empirical factors

Recently, Allen *et al.* (1998) have given crop coefficients for use with Penman-Monteith estimates of evaporation, for a hypothetical crop with a bulk surface resistance of 70 s m^{-1} and a height of 0.12 m (which can be taken as corresponding to short grass freely supplied with water). Two coefficients are given for water bodies greater than 5 m depth, clear of turbidity, in temperate climates. A value of 1.25 is recommended for the autumn and winter when the water body is releasing thermal energy and 0.65 when the water body is gaining thermal energy (spring and summer). No justification for these factors is given.

11. ACCURACY OF ESTIMATES OF OPEN WATER EVAPORATION

Lapworth (1965) describes an experiment to measure the evaporation from a reservoir at Kempton Park, near London (NGR 51211705), from 1956 to 1962 inclusive. The resulting data provide an excellent set of observations against which to test predictions of the different methods.

11.1 Evaporation measurements

The experiment made use of two reservoirs at the site. Most of the measurements were made on the East reservoir, which had an area of 17 ha and a maximum depth of 7.2 m, whilst some measurements were made on the smaller, West reservoir, for comparison. The water level was generally between 1.2 and 1.5 m below the top of the banks in the East reservoir, until a single lowering after which the levels were between 1.8 and 2.1 m. The mass balance method (see Section 3.2) was used to measure evaporation and so the water level was continuously recorded by a float-operated water–level recorder fixed over the outlet well of the reservoir. A pair of standard raingauges were installed at the site with, according to Lapworth, results being “in close accordance”. No other measurements were necessary to measure evaporation as, apart from the one lowering of the water levels, no inflows or outflows to the reservoir occurred. The evaporation could be calculated as the total rainfall (direct input plus runoff from the banks), plus or minus the change in water level. To quote Lapworth “This simple relationship, however, gives no indications of the difficulties which were experienced in obtaining a satisfactory measurement of evaporation”.

Lapworth suggests that the estimate of total evaporation over the 7 year period is within 5% of the true value, but does not give an assessment for the monthly values. Harbeck *et al.* (1954) report that they achieved an accuracy of better than 5% for 62% of their daily measurements at Lake Hefner. Given that the annual evaporation at Lake Hefner was twice that at Kempton Park but that the situation was complicated by inflows and outflows to the lake, it seems reasonable to assume that Lapworth achieved a comparable accuracy.

The water temperature was recorded, at approximately weekly intervals, near the centre of the East reservoir, at depth intervals of 1 m. Lapworth reports that, during the winter, the temperature was generally uniform with depth but, during the summer months, the temperature decreased with depth and the difference between top and bottom varied between 0.5 and 2.2ºC. Therefore, it is clear that the water column in the reservoir did become thermally stratified during the summer months and so the assumption of a well mixed body of water can not be justified for these periods.

Lapworth tabulates the results of his studies as monthly totals or, in the case of the water temperature, the reading on the first day of the month. The measurements gave a mean annual evaporation of 662 mm. The measurements of monthly evaporation from the East reservoir are shown in Figure 11.1 (a). These show a clear annual cycle with, on average, the minimum occurring in January and the maximum in July. There is a significant variation from year to year with the highest evaporation occurring in the summer of 1959 which was a particularly sunny period. The water temperature at a depth of 3 m, Figure 11.1 (b), shows a similar seasonal cycle, albeit, with less variability due to the smoothing effect of the heat storage.

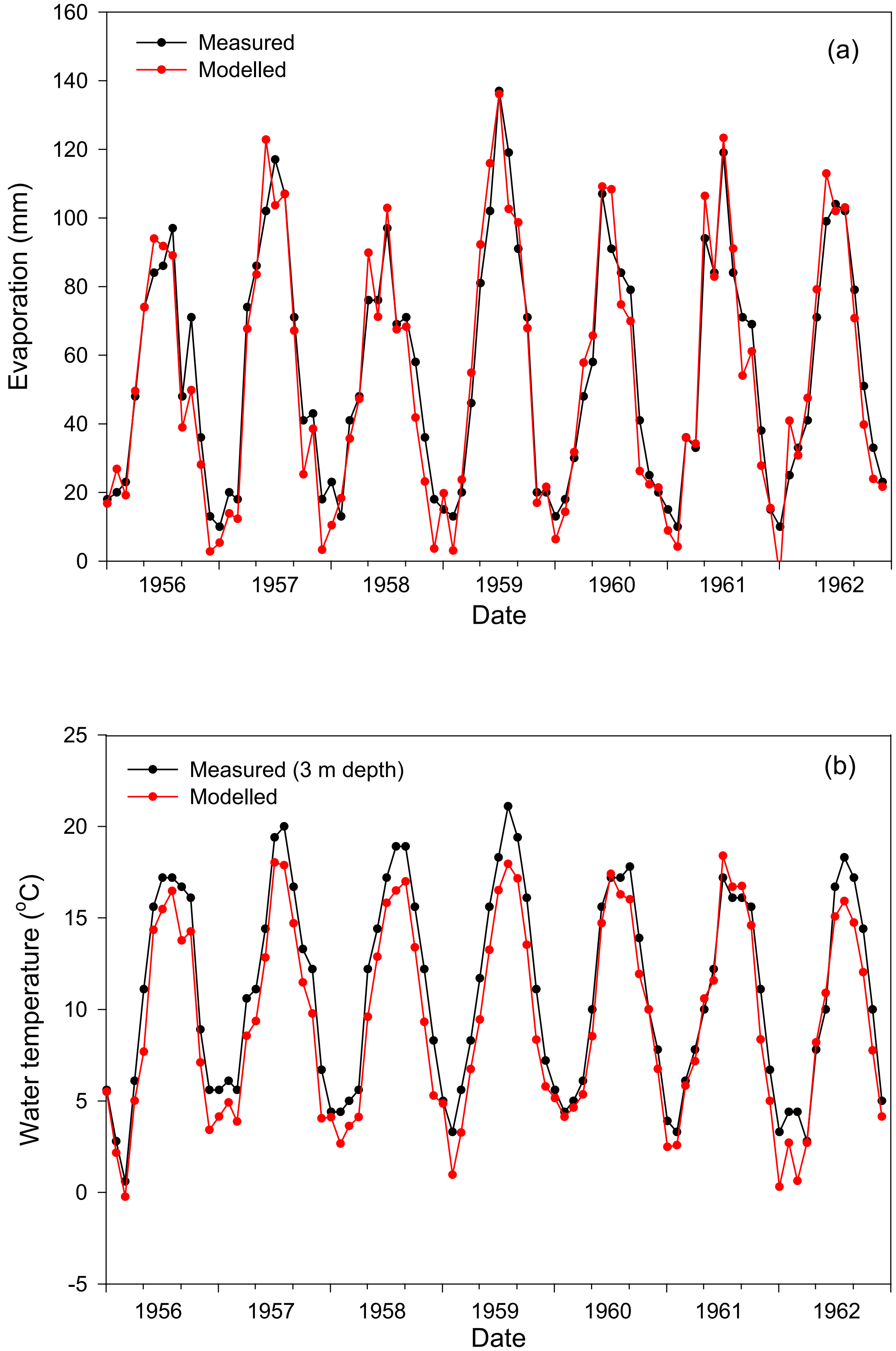

Figure 11.1 Measured monthly evaporation and water temperature at Kempton Park and that predicted by the equilibrium temperature model.

11.2 Meteorological data

Daily meteorological observations were obtained from The Met. Office for the period 1956 to 1962 inclusive. In the absence of a meteorological station at Kempton Park, the data were obtained for the station at Heathrow (NGR 50771767), approximately 7 km to the north-west. Sunshine hours were not recorded at Heathrow until 1957 and so the record was extended using data from Hampton (NGR 51311695), 2 km to the south-east of Kempton Park, for 1956. No overlapping data were available to check for consistency and so the data were used without any correction applied. However, sunshine hours normally show little variation over the short distance between the two stations, 9 km, and so it is reasonable to assume that the data from Hampton are representative of those from Heathrow.

11.3 The accuracy of estimates of evaporation

Five different methods of determining open water evaporation have been used to estimate the evaporation from the Kempton Park East reservoir:

- Equilibrium temperature model
- Penman-Monteith model with the change in heat storage set to zero
- Penman (1948) model with the change in heat storage set to zero
- Penman empirical factors applied to MORECS grass PE and PETCALC methodologies
- FAO-56 empirical factors applied to MORECS grass PE and PETCALC methodologies

The Penman-Monteith and Penman (1948) models with the change in heat storage set to zero should be considered as representative of shallow water, i.e. less than 0.5 m (see Section 2.2). The MORECS model is based on the Penman-Monteith model whilst PETCALC is based on the Penman model, see Section 4.4.2.

Table 11.1 Performance indicators of the five methods of estimating open water evaporation

Method	RMSE (mm)	MBE (mm)	Error in mean annual evaporation (mm)	RMSE (%)	MBE (%)	Error in mean annual evaporation (%)
Equilibrium temperature	9	2	-11	31.8	10.2	-1.6
Penman-Monteith	26	12	137	56.9	22.1	20.8
Penman (1948)	28	14	172	59.9	27.1	26.0
Penman factors × MORECS	18	6	72	48.8	19.0	10.9
Penman factors × PETCALC	30	18	220	74.3	38.2	33.6
FAO-56 factors × MORECS	34	-19	-227	48.4	-20.5	-34.3
FAO-56 factors × PETCALC	27	-12	-135	59.6	-7.5	-20.6

Estimates of evaporation and, in the case of the equilibrium temperature model, water temperature were generated for the 7 year period using a daily time step. An albedo of 0.065 was assumed (see Section 13.2.1) when calculating the net radiation and, for the Penman-Monteith model, a roughness length of 0.001 m (see Section 13.2.2). The resulting time series of evaporation were then aggregated to monthly totals for comparison with the results of Lapworth (1965).

The results are presented as graphs, Figures 11.1 to 11.3, for a subjective comparison and a series of performance indicators have been calculated, Table 11.1. The root mean square error (RMSE) measures systematic and non-systematic errors whilst the mean bias error (MBE) measures systematic errors. The values have been calculated both as the percentage error and the absolute error in order to reflect the seasonal cycle in the data. In addition, the absolute and percentage errors in the seven year mean annual evaporation have been computed.

11.3.1 Equilibrium temperature

The equilibrium temperature model performs extremely well with the best error measures of all the methods, except for the mean monthly bias percentage in which it is second best. The slightly higher value of this indicator is a result of a tendency for the model to underestimate the evaporation in the winter months and overestimate it during the summer. It is particularly impressive that there is little, if any, time lag between the measured and modelled time series, Figure 11.1 (a), demonstrating that the model is duplicating the heat storage term well. The error in the mean annual evaporation is less than the likely error in the measured values.

The model also duplicates the seasonal variations in the water temperature well, Figure 11.1 (b), although there is a consistent tendency to underestimate the temperature by about 1.5 ºC. However this may be an illusion as the modelled temperature is that of a well mixed water column whilst the measured values are point measurements at a depth of 3 m in a water column that is known to become thermally stratified during the summer months. In fact, the good agreement between the measured and modelled evaporations suggest that the effect of thermal stratification on the evaporation rates is negligible. This is likely to be true for other water bodies with similar dimensions.

11.3.2 Penman-Monteith

The Penman-Monteith model consistently overestimates the evaporation during the summer months, Figure 11.2. This is also demonstrated by the high RMSE and MBE percentages. The result is that the model overestimates the mean annual evaporation by 137 mm (20.8%). The error is a consequence of not estimating the outgoing long wave radiation accurately and suggests that the correction given by Monteith (1981) (for the assumption that the surface and air temperatures are approximately the same) has little impact in this case.

The Penman-Monteith model is the fourth most accurate in terms of the monthly percentage errors but third in both the monthly errors and the error in the mean annual evaporation. However, there is a pronounced time shift between the modelled and measured values as a result of the heat storage in the water body not being taken into account. The modelled values could be made to fit the observed data better by adjusting the values used for the albedo and roughness length, but these would then not conform to the values observed by other researchers.

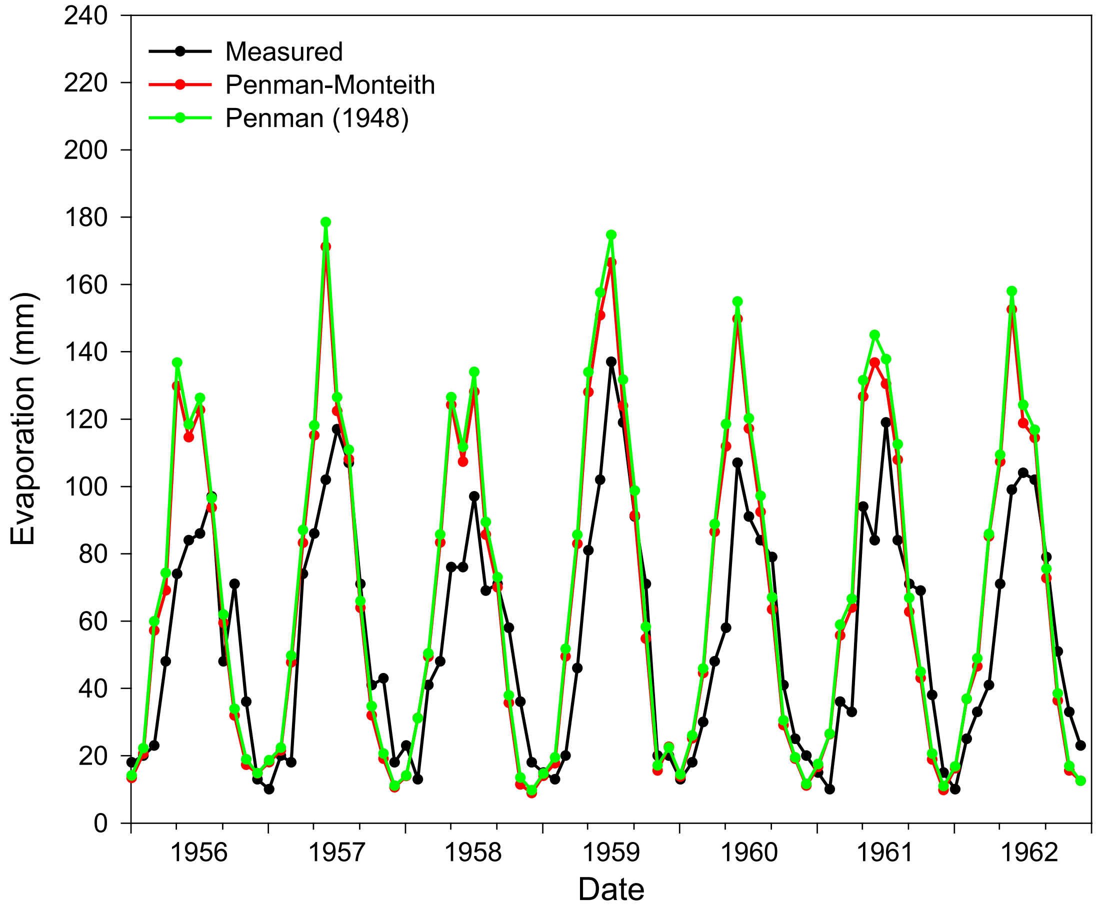

Figure 11.2 Comparison between observed monthly evaporation and that predicted by the Penman (1948) and Penman-Monteith models for open water.

11.3.3 Penman (1948)

The Penman (1948) model considerably overestimates the evaporation during the summer months but tends to get the evaporation during the winter months about right. This is a consequence of not estimating the outgoing long-wave radiation accurately due to using the air temperature in the calculation, rather than the surface temperature. This is demonstrated if the net radiation calculated by the equilibrium temperature model is input to the Penman model. The error is reduced to 59 mm (8.9%). This confirms that the assumption that the surface temperature of water is close to the air temperature is correct during winter but not during the summer. As a consequence, the Penman (1948) model has the third highest errors in terms of percentages and the second highest errors in terms of the absolute values.

There is a pronounced shift in the modelled time series compared with the measured values, Figure 11.2. The rise in evaporation rates in the spring and the fall in the autumn is predicted to occur earlier than is observed. This is a consequence of the model failing to take account of the heat storage term.

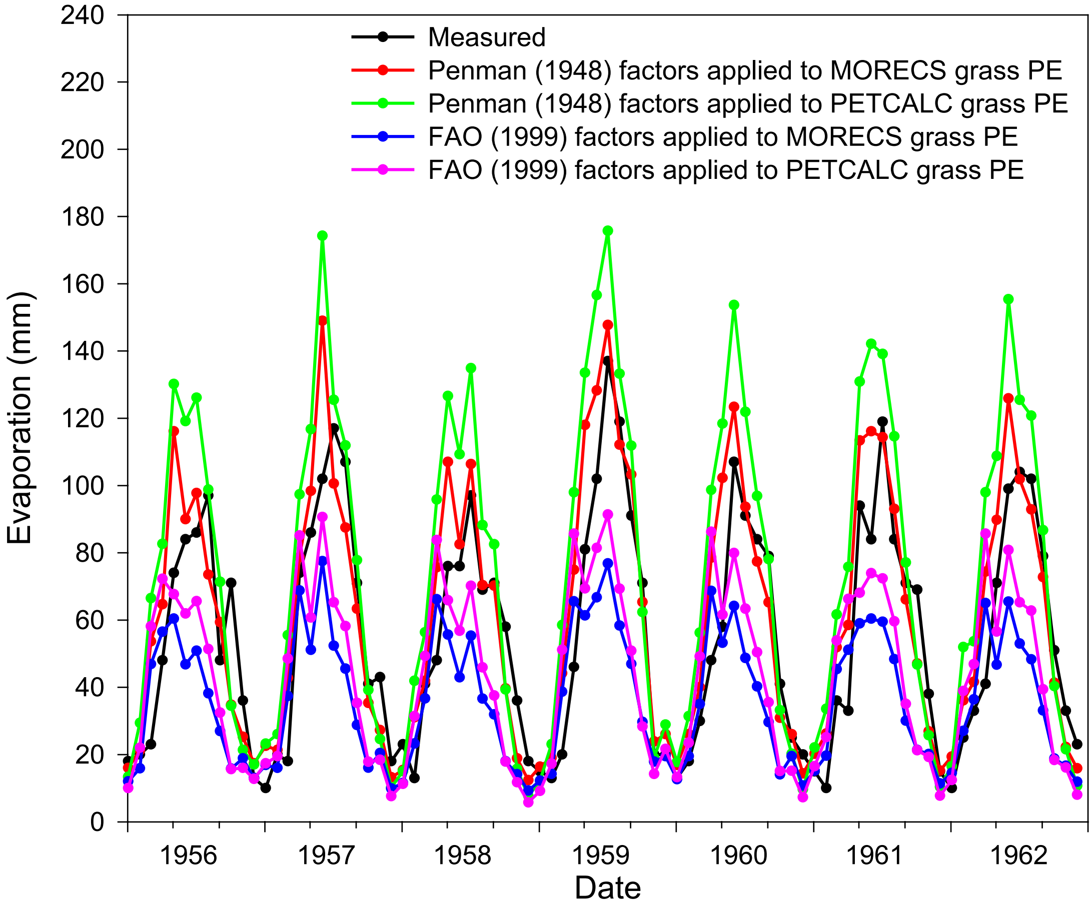

Figure 11.3 Comparison between observed monthly evaporation and that predicted by the empirical factors method.

The Penman model was originally calibrated against measurements of open water evaporation made from small brick lined pits. The conditions at the Kempton Park reservoirs are significantly different from those for which the calibration was performed (e.g. size of water body, setting of the site) which might be an additional factor in explaining the poor performance of this model.

11.3.3 Penman empirical factors

See Section 10.4 for the assumptions made in deriving these factors. When applied to simulated MORECS grass PE (it should be noted that the factors were not developed for use with the MORECS data) the method has the second best error measures and the predicted values compare very favourably with the measured values, Figure 11.3. The error in the mean annual evaporation is 10.9%, although this is greater than the estimated errors in the measured values. The RMSE and MBE are also the second lowest of the methods and about twice those of the equilibrium temperature model.

When applied to estimates of grass PE from PETCALC then the errors increase significantly. In particular, the result is an overestimate in the average annual evaporation of 33.6%. All the other error measures are the worst of any of the methods. The reason for this is that the estimates of grass PE from PETCALC are higher than those from the simulated MORECS so that, when multiplied by the empirical factors the estimates of open water evaporation are far too high. This is partly due to no correction being made for the assumption of air temperature being the same as the surface temperature (Note, both the MORECS and the Penman-Monteith model for open water used in this project attempt to make a correction) and because PETCALC uses the Penman (1948) formula, developed for water (although PETCALC calculates the net short-wave radiation using an albedo of 0.25, as is used in MORECS for grass). Thus the aerodynamic term of the model is calibrated for water rather than grass.

11.3.5 FAO-56 Empirical factors

The values generated using this method compare very poorly with the measured values when the PE data set is simulated MORECS grass PE. The method seriously underestimates the mean annual evaporation, by 227 mm (34.3%) although the RMSE and MBE are comparable to the other methods. The reason for the serious underestimate of the annual evaporation is that the factors applied to grass PE during the summer result in a large reduction in the evaporation from the values of grass PE, rather than the slight increase that is required. The factors for the winter months result in evaporation significantly greater than grass PE and thus are comparable to those for water.

However, when the factors are applied to grass PE calculated using PETCALC then the errors are lower. The better results obtained when these factors are applied to PETCALC are because the estimates of PE by PETCALC are greater than those from MORECS with the result that the evaporation from water is under estimated by less. A case of two errors almost cancelling each other out.

11.3.6 Discussion

The results clearly show that the equilibrium temperature model provides the most accurate estimates of evaporation of any of the methods tested. The estimates of annual evaporation are likely to be of the order of 5% and so well suited to the Agency's requirements. For monthly estimates the error is likely to be higher, probably around 10% but this is still conformable to the acceptable accuracy for the Agency (see Section 4.1) and it is clear that only this method is capable of simulating the lag observed in the measurements. The model's ability to handle the heat storage term of the water body and to directly estimate the outgoing long-wave radiation are important factors that contribute to its success.

Both the Penman (1948) and Penman-Monteith models suffer from the disadvantage that the outgoing long-wave radiation is poorly represented, with the result that both tend to overestimate evaporation by a significant amount. Therefore, neither of these models should be used by the Agency for estimating open water evaporation as the resulting errors will be significantly greater than the acceptable error of 10% (see Section 4.1).

Using the empirical factors of Penman (1948) with the simulated MORECS grass PE values results in a accuracy of around 10% in estimates of annual evaporation, suggesting that this method can be used with some confidence. However, these results are based on using MORECS single station data, i.e. where the meteorological data is for the specific site. The use of MORECS grid data is likely to result in greater errors due to the meteorological data used being averages of the grid cell. When applied to PETCALC grass PE, the result is large errors due to the significantly higher estimates of PE during the summer months. A further concern is that the empirical factors were calibrated for a site in southern England and their accuracy in a different climate, e.g. the Pennines, must be suspect.

The large errors produced by the use of the FAO-56 empirical factors means that they are totally unsuited for UK conditions and should not be used by the Agency.

Therefore the only method that fulfils the Agency's requirements on accuracy is the equilibrium temperature method.

11.4 Impact of different time steps on the equilibrium temperature model

The equilibrium temperature model, described in Section 10.3, is designed for use with a daily time step. However, daily values of meteorological data will not be available on all occasions. There is the potential for errors to arise because the calculated evaporation values are not independent of each other since the change in heat storage is, in part, a function of the water temperature at the end of the previous time step and therefore a function of the evaporation the previous period. There are two fundamentally different approaches that could be adopted:

- Run the model with the time step of the meteorological data
- Disaggregate the meteorological data to daily values

The first method is probably preferred when either monthly or annual values of evaporation are required, e.g. licensing, as it minimises the amount of calculation required. The second is probably more attractive when daily values are required, e.g. water resource modelling, as the disaggregation is then performed directly on the meteorological data and it is potentially possible to combine data with different time steps.

Running the model with a large time step may result in the equilibrium temperature time constant becoming less than the model time step, resulting in an instability. The use of meteorological data disaggregated to daily values may result in errors because there will not be as much variability in the daily values with the consequence that the predicted water temperature may become closer to the equilibrium temperature than would actually occur.

Both strategies have been tested for time steps of 5, 10 and 30 days. The daily meteorological data was aggregated to daily averages for the relevant time step. For testing the effect of disaggregating the meteorological data, these time series were then disaggregated to daily values using linear interpolation (see Section 6.4.1).

The percentage error in the mean annual evaporation and the RMSE and MBE as percentages of the monthly total evaporations have been calculated against modelled evaporation with a time step of one day and the measured values. It is encouraging that the error in the mean annual evaporation remains below the estimated error in the measured values for all time steps. The MBE of the estimates of evaporation compared with the model run with a one day time step remain less than 5% for both strategies for time steps of both 5 and 10 days. However, it increases significantly for a time step of 30 days. The trend of increasing error with length of time step is also shown by the RMSE. The modelled evaporation with a one day time step was less than that measured and so the errors when comparing the modelled with the measured values decrease because increasing the time step increases the modelled evaporation rates.

Table 11.2 Errors occurring when using meteorological data with a time step greater than one day

		Model time step (days)			Meteorological data aggregation time step (days)		
		5	10	30	5	10	30
Modelled 1 day time step as "truth"	% error in mean annual	-0.3	1.2	1.4	1.1	1.7	1.6
	RMSE %	22.9	35.9	72.3	52.5	65.0	80.8
	MBE %	-0.6	4.9	12.2	-2.4	1.1	6.8
Measured data as "truth"	% error in mean annual	-2.2	-1.7	-1.7	-3.0	-2.1	-2.0
	RMSE %	46.2	47.6	49.0	32.5	29.4	27.5
	MBE %	9.5	7.5	6.0	9.7	7.5	3.7

It is clear that, in terms of the errors, there is little to chose between the two different strategies for handling meteorological data with time intervals greater than one day. Both strategies result in increasing errors within the measured error for intervals of 5 and 10 days. The error does increase for a time step of 30 days but is still just acceptable. This demonstrates that acceptable estimates of open water evaporation can be achieved with meteorological data with a time period up to and including one month. The choice of strategy for handling the data can therefore be based on the application for which the estimates are required rather than on the required accuracy.

11.5 Summary of results

- The evaporation rates predicted by the equilibrium temperature model are in excellent agreement with the measured data with errors generally less than the errors in the observations.

- The Penman (1948) model performs poorly because the outgoing long wave radiation is calculated on the assumption that the water surface temperature is approximately the same as the air temperature.

- The Penman-Monteith model performs better than the Penman (1948) model but still has errors that are significantly greater than the errors in the observations, despite incorporating a correction for the surface temperature of the water.

- The empirical factors given by Penman (1948) and applied to simulated MORECS grass PE predicts evaporation rates that are in reasonable agreement with the measurements but when applied to PETCALC grass PE the errors become unacceptably large.

- The FAO-56 empirical factors applied to simulated MORECS grass PE result in very inaccurate estimates of evaporation and are therefore inappropriate to UK conditions.

- Driving the equilibrium temperature model using meteorological data with a time step of up to and including 30 days can be accomplished either by running the model with the time step of the data or by disaggregating the data to daily values with no significant difference in the resulting errors.

12. OPTIMUM VALUES FOR EMPIRICAL FACTORS

The empirical factors commonly used are those of Penman (1948b) and, in the original paper, were given to convert estimates of evaporation for open water, calculated using the model, to that for well watered turf. They are based on measurements of evaporation, for two years, at one site (Rothamsted Experimental Station) in southern England using cylinders 0.76 m in diameter and 1.83 m deep. At the time, the site was criticised as having poor exposure. It will be appreciated that the size of the these cylinders is untypical of the water bodies of interest to the Agency. In addition, the empirical factor for the winter months is derived from measurements at Fleam Dyke, in Cambridgeshire. The Kempton Park reservoirs are much more typical of the water bodies that the Agency is concerned with and the time span of measurements, seven years as against two, much longer. In addition, the time series of grass potential evaporation used as the basis for operational estimates of open water evaporation are different to the model used by Penman. Therefore, there is the opportunity to use the measured open water evaporation from Kempton Park to calculate empirical factors calibrated explicitly to the grass potential evaporation data sets used by the Agency.

It had originally been planned to produce two sets of optimum empirical factors. One set for the situation when the water is too shallow for the heat storage term to be significant and the other for deeper water when heat storage does have an effect. The evaporation for shallow water was to have been calculated using the Penman-Monteith combination model on the assumption that this was an accurate representation of the evaporation. However, the results of the comparison between evaporation estimated using this model and the measured values at Kempton Park (see Chapter 11) raised serious concerns about the parameterisation of this model. The resulting uncertainty meant that there was no basis upon which results from the Penman-Monteith model could be used with confidence.

The empirical factors for converting estimates of grass potential evaporation to open water evaporation have been calculated using the measured evaporation from Kempton Park. Two sources of grass potential evaporation are currently in use by the Agency. These are the PETCALC system used by the Northwest Region and MORECS (see Section 4.4.2 for descriptions of these systems). The PENSE system used by the Southern Region is a simulation of MORECS. As such, it need not be considered separately but the empirical factors calculated for use with MORECS grass PE can also be considered as applicable for use with PENSE. Daily time series of grass PE were generated, using the meteorological data from Heathrow as the driving variables, by programs that used essentially the same equations as in these systems (see Annexes F and G). The resulting values were aggregated up to monthly totals and the average conversion factor to the measured values at Kempton Park have been determined for each calendar month. These are shown graphically in Figure 12.1 and are given in Table 12.1. The empirical factors of Penman (1948b) are included in the table for the purpose of comparison. The effect of the heat storage in the water body is clearly present in the empirical factors, demonstrated by the fact that the smallest empirical factors occur in March whilst the largest are in November. The accuracy of estimates of open water evaporation using these factors at other locations in England and Wales is explored in Chapter 14.

Table 12.1 Empirical factors for converting grass PE values to open water evaporation

Month	MORECS grass PE	PETCALC	Penman (1948b)
Jan	1.43	1.57	1.67
Feb	1.14	0.88	1.67
Mar	0.92	0.71	1.43
Apr	0.95	0.75	1.43
May	0.91	0.78	1.25
Jun	1.02	0.81	1.25
Jul	1.24	0.99	1.25
Aug	1.37	1.08	1.25
Sep	1.47	1.25	1.43
Oct	1.99	1.98	1.43
Nov	2.29	2.63	1.67
Dec	1.95	2.68	1.67

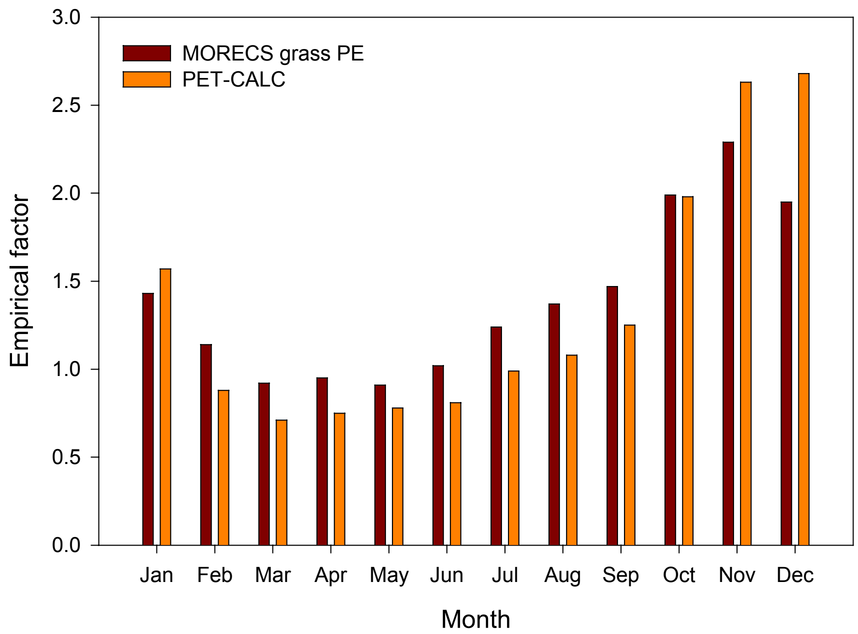

Figure 12.1 Empirical factors for converting grass PE to open water evaporation.

13. SENSITIVITY ANALYSIS

A sensitivity analysis has been carried out on the equilibrium temperature model, a simulation of MORECS grass PE and a simulation of PETCALC to investigate the impact of likely errors in the measurements of the driving variables on the evaporation calculated using these models. In addition a sensitivity analysis has also been carried out into uncertainty in the values for the model parameters of the equilibrium temperature model. The data from four meteorological stations, Table 13.1, have been used to investigate whether the magnitude of the driving variables has an effect on the uncertainty. These stations are located in the south-west, south-east and north of England and in mid-Wales and so can be anticipated to have different ranges of the driving variables. The time period used was 1989 to 1999 inclusive for the four stations.

Table 13.1 Location of meteorological stations used in sensitivity analysis

Name	Location	Easting	Northing	Elevation (m AOD)	Start date
Cefn Brwyn	Mid-Wales	282540	283870	359	1 Jan 1989
Coalburn	North England	369670	578320	292	1 Jan 1989
North Wyke	Southwest England	267050	96580	180	1 Jan 1989
Wallingford	South Midlands	461520	189630	47	1 Jan 1988

The following analysis considers the driving variables and model parameters one at a time, i.e. in isolation. Obviously, this is not a realistic scenario but is used in order to clearly identify the impact of a particular error or uncertainty. In practice, estimates of evaporation will be subject to errors in all the driving variables and parameters. In this case the errors will be the sum of the errors of the individual driving variables and model parameters since the models are linear.

In the case of the empirical factors method, there are two potential impacts of the location, one is through the PE model and the other is the possibility of bias in the empirical factors themselves. The result is that it is not possible to usefully comment on the sensitivity of the error to the climate or the magnitude of the driving variables.

13.1 Errors in the driving variables

The measurements made from which the driving variables of the models are calculated, are:

- incoming solar radiation (often measured as sunshine hours)
- air temperature
- wind speed
- vapour pressure deficit (generally expressed as relative humidity)

The accuracy to which these are generally measured are given in Table 13.2. These numbers were derived from a combination of WMO (1983), Met. Office (1982), the specifications for instruments from several manufacturers and by consulting staff who had experience in using the instruments.

The three models were run at a daily time step. The resulting estimates of evaporation were taken as having no errors. The models were run with each variable in turn having the measurement accuracy added and then subtracted. The average daily and annual percentage differences were calculated to give estimates of the impact of errors in the measurements on estimates of evaporation. These estimates are 'worst case scenarios' as they represent the situation of a consistent measurement error. They should be interpreted in the context of the accuracies required for the Agency's purposes, Section 4.1, which are 10% for abstraction licensing and, for water balance studies, an acceptable accuracy of 10% and a desirable accuracy of 5%.

Table 13.2 Measurement accuracy of driving variables

Measurement	Accuracy
Incoming solar radiation (W m^{-2})	± 5%
Air temperature (ºC)	± 0.5ºC
Wind speed (m s^{-1})	± 3%
Vapour pressure deficit (kPa)	± 5%

13.1.1 Equilibrium temperature model

The results of the sensitivity analysis, Table 13.3, show that the errors in the estimates of solar radiation have a significant impact on the estimates of evaporation produced by the equilibrium temperature model. The errors in daily estimates range from 8.1 to 9.9% with the result that the acceptable accuracy can be met but it is potentially difficult to meet the desirable accuracy in the worst case. This emphasises the need to ensure that good estimates of the incoming solar radiation are available. The differences in the sensitivity between the stations probably reflects the decreasing importance of the thermodynamic term resulting from a combination of higher latitude and greater cloud amounts at Cefn Brwyn and Coalburn.

The model is insensitive to errors in the measurements of wind speed but moderately sensitive to errors in air temperature and vapour pressure deficit. However, the errors likely to arise from measurement errors in the latter two variables are well within the desirable accuracy for estimates of evaporation.

Table 13.3 Percentage errors in evaporation estimated using the equilibrium temperature model due to errors in measurements

	North Wyke		Wallingford		Cefn Brwyn		Coalburn	
	Daily	Annual	Daily	Annual	Daily	Annual	Daily	Annual
Incoming solar radiation (W m^{-2})	±9.9	±8.1	±9.9	±8.8	±8.1	±6.6	±8.5	±7.4
Air temperature (ºC)	±1.7	±1.5	±1.9	±1.7	±1.2	±1.3	±1.5	±1.6
Wind speed (m s^{-1})	±0.3	±0.5	±0.5	±0.4	±0.2	±0.4	±0.2	±0.2
Vapour pressure deficit (kPa)	±1.8	±1.5	±0.7	±0.6	±1.5	±1.1	±1.0	±0.7

In general, the errors arising in annual estimates tend to be lower than those in the daily estimates but are of a comparable size. Although there are differences in the errors between the different stations, these are comparatively small and so are of little significance.

13.1.2 MORECS

Grass PE estimated using the simulation of MORECS is less sensitive to errors in the incoming solar radiation than estimates of open water evaporation using equilibrium temperature model, see Table 13.4, as a result of the higher value of the albedo, 0.25 compared to 0.065. Nevertheless, errors in the measurement of solar radiation result in higher inaccuracies than any of the other driving variables. The errors are within the acceptable criteria and comparable to the desirable accuracy but, when used to estimate open water evaporation, the errors will increase for most months due to the use of a single, average, empirical constant for each month.

The sensitivity to air temperature and vapour pressure deficit are higher than for the equilibrium temperature model, reflecting the different proportions between the thermodynamic and aerodynamic terms in the model as a result of the higher value for albedo. The two variables are comparable in their impact on the errors and are within the desirable accuracy for estimates of evaporation. There is little sensitivity to errors in wind speed.

It is not possible to comment on the effect of the magnitude of the driving variables because uncertainties resulting from errors in the measured driving variables can not be isolated from those due to using empirical factors derived from data from one climate region and being used in another, different region (see Chapter 14).

Table 13.4 Percentage errors in evaporation estimated using the simulated MORECS grass PE model due to errors in measurements

	North Wyke		Wallingford		Cefn Brwyn		Coalburn	
	Daily	Annual	Daily	Annual	Daily	Annual	Daily	Annual
Incoming solar radiation (W m^{-2})	±6.2	±4.2	±5.7	±4.2	±4.7	±4.0	±5.1	±4.7
Air temperature (ºC)	±2.2	±0.1	±2.3	±0.2	±0.1	±0.4	±0.1	±0.7
Wind speed (m s^{-1})	±0.7	±0.1	±0.6	±0.1	±0.4	±0.1	±0.4	±0.3
Vapour pressure deficit (kPa)	±2.7	±1.2	±2.2	±1.1	±2.3	±1.2	±2.2	±1.2

13.1.3 PETCALC

The results for PETCALC are significantly different to those for simulated MORECS grass PE, see Table 13.5. The sensitivity to errors in incoming solar radiation are lower, whilst those for the air temperature are much higher. This reflects the different methods used by the two models for calculating the incoming long-wave radiation (see Annexe G) and the absence of a correction for using air temperature rather than surface temperature in calculating the outgoing longwave radiation. The result is that the model is very sensitive to the errors in the air temperature. Surprisingly, the sensitivity to errors in the vapour pressure deficit is similar to that of the simulated MORECS grass PE, despite the difference in the formulation of the aerodynamic term in the two models. As with MORECS, the errors will increase for most months due to the use of a single empirical constant for each month, when used to estimate open water evaporation.

It is not possible to comment on the effect of the magnitude of the driving variables because uncertainties resulting from errors in the measured driving variables can not be isolated from those due to using empirical factors derived from data from one climate region and being used in another, different region (see Chapter 14).

Table 13.5 Percentage errors in evaporation estimated using the PETCALC model due to errors in measurements

	North Wyke		Wallingford		Cefn Brwyn		Coalburn	
	Daily	**Annual**	**Daily**	**Annual**	**Daily**	**Annual**	**Daily**	**Annual**
Incoming solar radiation (W m^{-2})	±3.8	±1.2	±3.3	±1.1	±1.8	±1.2	±1.9	±1.3
Air temperature (^oC)	±27.2	±11.2	±20.7	±8.9	±22.7	±12.4	±26.8	±14.3
Wind speed (m s^{-1})	±0.7	±0.4	±0.5	±0.3	±0.5	±0.4	±0.3	±0.2
Vapour pressure deficit (kPa)	±2.7	±1.5	±2.3	±1.6	±2.0	±1.5	±1.5	±1.0

13.2 Uncertainties in the equilibrium temperature model parameters

There are three model parameters that are required by the equilibrium temperature model:
- Albedo for short-wave radiation (α_s)
- Roughness length (z_0)
- Water depth (z)

A study of the sensitivity of the predicted evaporation (at the time scale of day, month and year) to each of these parameters has been carried out in order to determine the likely error that would occur due to uncertainties in these parameters. It is very unlikely that there will be measurements of the first two parameters but it is possible that the water depth will be known.

Each parameter was varied in turn over a range that represented the minimum and maximum values that it could be reasonably expected to occur, based on values reported in the literature. The percentage differences between the evaporation predicted for a particular value and that predicted with the value used with the Kempton Park data were calculated for time periods of a day, a month and a year.

13.2.1 Short-wave albedo

This is required for estimating the net radiation. There are a number of factors that can potentially affect this parameter and there is a considerable amount of information in the literature (mainly relating to oceanography). Detailed discussions can be found in Jerlov (1976), Kirk (1994) and Mobley (1994).

The albedo is potentially time variant as it is a function of the solar elevation angle. For an infinitely deep body of pure water with no suspended particles and a perfectly smooth surface, the albedo can be calculated from a combination of Snell's law and Fresnel's equation, Figure 13.1. This shows that the albedo maintains a very low value, about 0.02, at high elevation angles of the illumination source. However, as the elevation angle decreases the albedo steadily increases until, at an angle of 37 degrees, it has doubled. At lower elevation angles it rises rapidly. The consequence is that, for the solar direct beam, the albedo will be at a high value around dawn but will diminish as the day progresses to a minimum at solar noon before rising again towards dusk. The minimum albedo will be a function of both the latitude of the site and the time of year. As England and Wales lie at comparatively high latitudes the effect is significant in winter. However, the periods of high albedo coincide with those of low incoming solar radiation so the impact of the temporal variations in albedo on evaporation rates are diminished.

The situation is complicated by the fact that a proportion of the incoming solar radiation is in the form of diffuse radiation due to scattering within the atmosphere. This radiation is coming from the entire hemisphere and may be anisotropic, due mainly to the amount and type of clouds present. Thus the albedo will also be a function of the amount of diffuse solar radiation and its angular distribution. The effect will be to increase the albedo.

The surface of the water will rarely be perfectly flat, due to wind-induced waves. The size and orientation of these waves will be a function of the strength and direction of the wind and the fetch and depth of the water body. The effect of waves on the water surface will be to change the distribution of elevation angles of the incident radiation resulting in a tendency to increase the albedo.

Suspended particles in the water body will tend to increase the albedo, mainly by backscattering the radiation that has penetrated into the water body. This effect will be a function of the size of the suspended particles and the number present. These will depend on the type of water body as the phytoplankton abundance in the upper layers will have an effect as will the amount of suspended sediments.

Finally, there is the potential for reflectance from the bottom. This will be a function of the albedo of the bottom materials, the depth of the water, the turbidity of the water and the incident angle of the incoming radiation.

From the discussion above it will be appreciated that predicting the albedo of water is a non-trivial issue. Physically rigorous numerical models have been constructed which are capable of this, e.g. Mobley (1995), however, they are demanding both computationally and in terms of the input parameters and so are best regarded as research tools. Therefore, empirical models have tended to be used. Harbeck *et al*. (1954) measured albedo at Lake Hefner and produced a series of regression equations relating the albedo to the solar elevation angle and the cloud type, e.g. low overcast, high broken, etc. It would be difficult to use this model in practice as the information on the clouds is not readily available. Oceanographic modellers make extensive use of the work of Payne (1972) who made measurements of albedo from platforms at two sites (the Sargasso Sea and coastal waters near Massachusetts) to produce a look up table that related the albedo to the solar elevation and the transmissivity of the atmosphere (defined as the ratio of the incoming solar radiation at the earth's surface to that at the top of the atmosphere). It is uncertain to what extent Payne's model can be applied to inland waters. The optical characteristics of large water bodies, such as major water supply reservoirs, are probably not too different from oceanic waters. However, for smaller, shallower water bodies this is not necessarily true. This is due to the possibility of reflectance from the bottom, differences in the waves on the surface, and differences in the amount and type of suspended particles; all of which will tend to increase the albedo. Nevertheless, it is the only practical method available.

The average daily albedo was calculated for the Kempton Park site using the model of Payne (1972). The results showed that the albedo varied between a maximum of 0.40 and a minimum of 0.06 with a mean of 0.095. The daily evaporation was calculated using the equilibrium temperature model with the time series of daily albedo and then with a constant value of 0.065. The difference in the mean annual evaporation was 15 mm (2.3%) with a maximum daily difference of 0.26 mm and an average daily difference of 0.04 mm. This suggests that the accuracy gained by using an albedo that varies daily against a constant value is not warranted and that a value of 0.065 should be used for water bodies in the England and Wales.

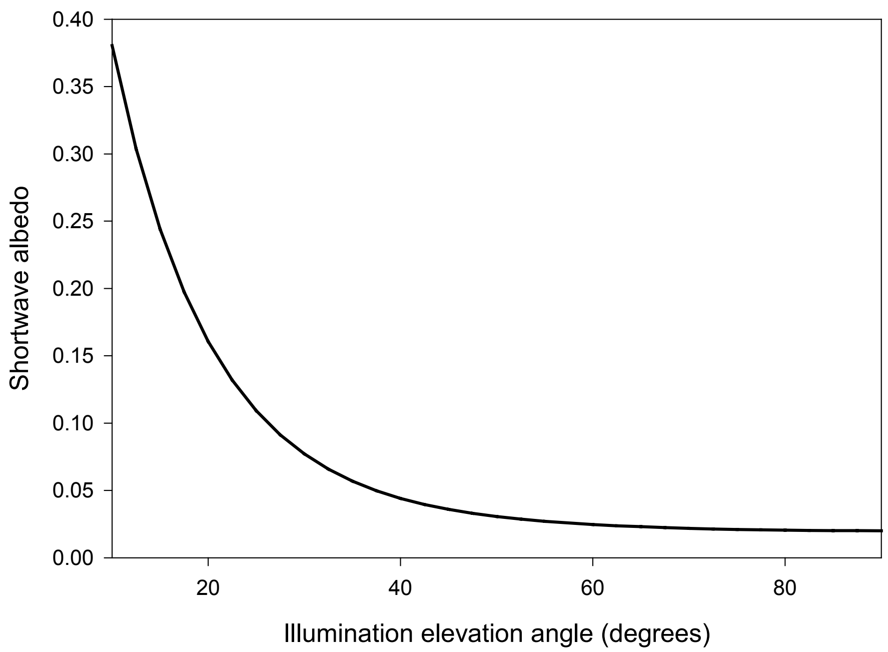

Figure 13.1 Variation in albedo of water with the elevation of the illumination source.

Unfortunately there are virtually no measurements of the short-wave albedo for inland water bodies reported in the literature to compare this value with. This is because many studies make use of micrometeorological instruments that measure the net radiation directly and so have no requirement for a value of albedo. Other studies have tended to use values of 0.06 to 0.08 which are cited as coming from major texts such as Brutsaert (1982). The basis for these values is not evident. Thus, a value of 0.065 is consistent with other studies..

The sensitivity analysis was carried out with the albedo varying between 0.02 and 0.12 and show that the estimated evaporation is relatively insensitive to the value of albedo. The results showed essentially the same sensitivity for values at the daily, monthly and annual time scale and so only the results at the annual scale are reproduced here, Figure 13.2. There is relatively little difference between the results from the four stations. The small differences that do exist are a function of the relative sizes of the thermodynamic (radiation) and aerodynamic (wind and vapour pressure) terms. Thus Cefn Brwyn has the lowest sensitivity as it has less sunshine and is windy, whilst Wallingford has the lowest wind speed and the second highest amount of sunshine. The albedo needs to be known to ±0.032 around the 'typical' value of 0.065 for the estimates of evaporation to be within 5%. Unless there is clear evidence to the contrary, e.g. measurements, a value of 0.065 should be used.

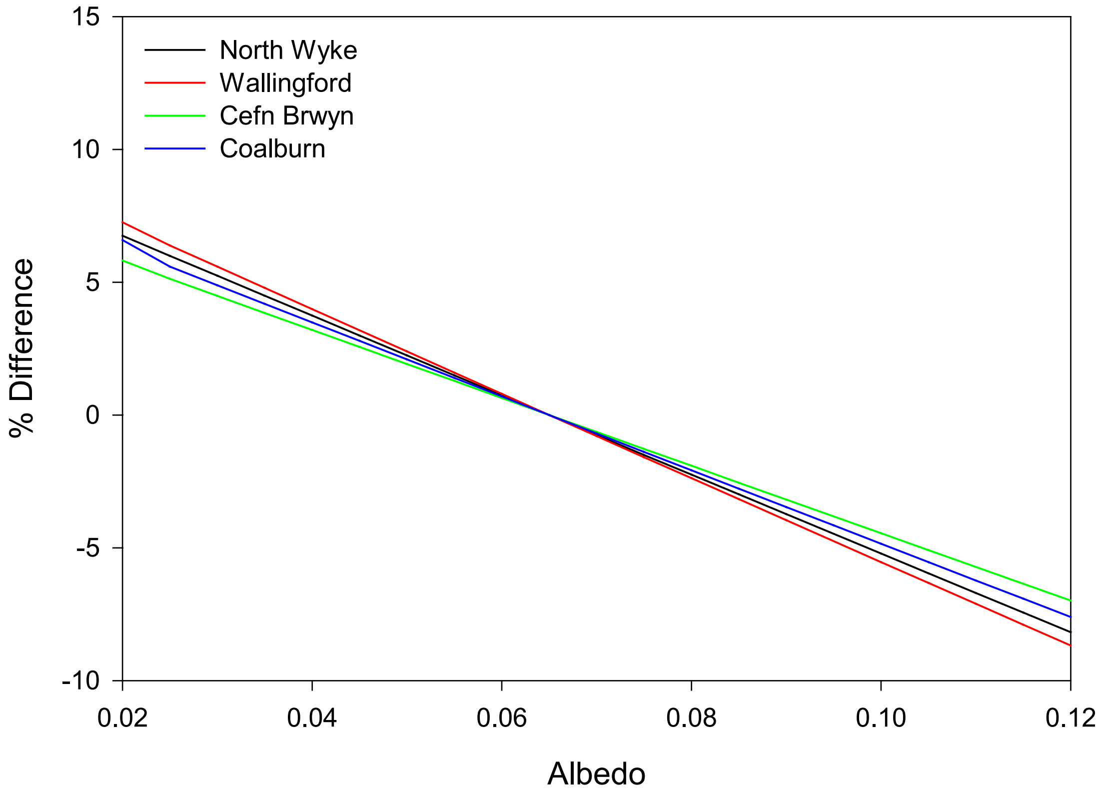

Figure 13.2 Percentage difference in estimated annual evaporation with variation in albedo.

13.2.2 Roughness length

The roughness length, z_o, is a parameter which represents the effect of skin friction and form drag as a force on the air flowing over the water body. (The skin friction and form drag are often combined into a form which can then be defined as the drag coefficient.) Increasing the roughness length results generally in increased evaporation due to increased turbulence in the air flow over the surface, see Section 2.1.3.

Roughness lengths have been rarely reported in the literature since most studies have made use of the mass transfer method (see Section 3.4) in which the transfer (drag) coefficient, C, is explicit. The roughness length can be calculated from the drag coefficient by:

$$z_o = \frac{z_r}{e^{\sqrt{k^2/C}}} \qquad (40)$$

Intuitively, it would seem inappropriate to use a constant value for the roughness length as it will depend on the wave field, which will in turn be a function of the wind speed, fetch and possibly water depth. A study by Garratt (1973) on Lough Neagh (Northern Ireland) found a linear relationship between wind speed, u (ms^{-1}), and drag coefficient of the form:

$$C = (0.36 + 0.1u)10^{-3} \tag{41}$$

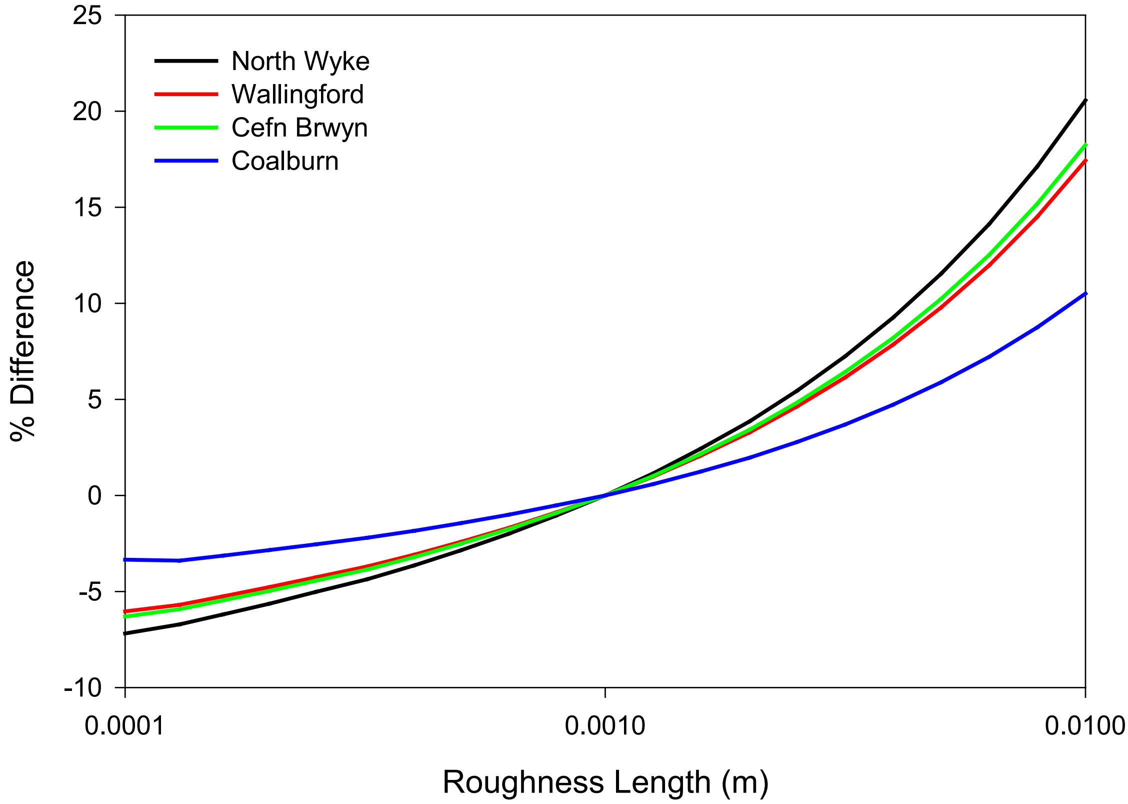

Figure 13.3 Percentage difference in annual evaporation with changes in the roughness length.

This relationship was tested against the data for Kempton Park. The average roughness length calculated from Eq. (40) and (41) was 0.000014 m, which agrees well with the values generally cited in the literature, e.g. Brutsaert (1982). However, using this value resulted in an underestimation of the evaporation by about 25%. Consequently, it was concluded that the relationship was not applicable to smaller fresh water bodies.

Therefore a pragmatic approach has been adopted. Penman(1948b) had calibrated his model against measurements of evaporation from water. Hence, the roughness length implied by the empirical values given in the aerodynamic term of the Penman (1948) model was calculated by equating the aerodynamic terms in equations (27) and (29), see Section 9.1. The resulting value is 0.001 m. This is significantly higher than values generally reported in the literature but it does give good results against the observations at Kempton Park although it is difficult to justify in physical terms.

The value of the roughness length was varied over a wide range in the sensitivity analysis, from 0.0001 to 0.01 m. The results for daily, monthly and annual time periods were essentially the same and so only the sensitivity of the annual total evaporation will be presented here. The results show essentially the same relationship for three of the stations, Figure 13.3, but for the data from Coalburn the sensitivity is significantly less. There is no obvious explanation for this except that the data from this station has the lowest average incoming solar radiation and vapour pressure deficits.

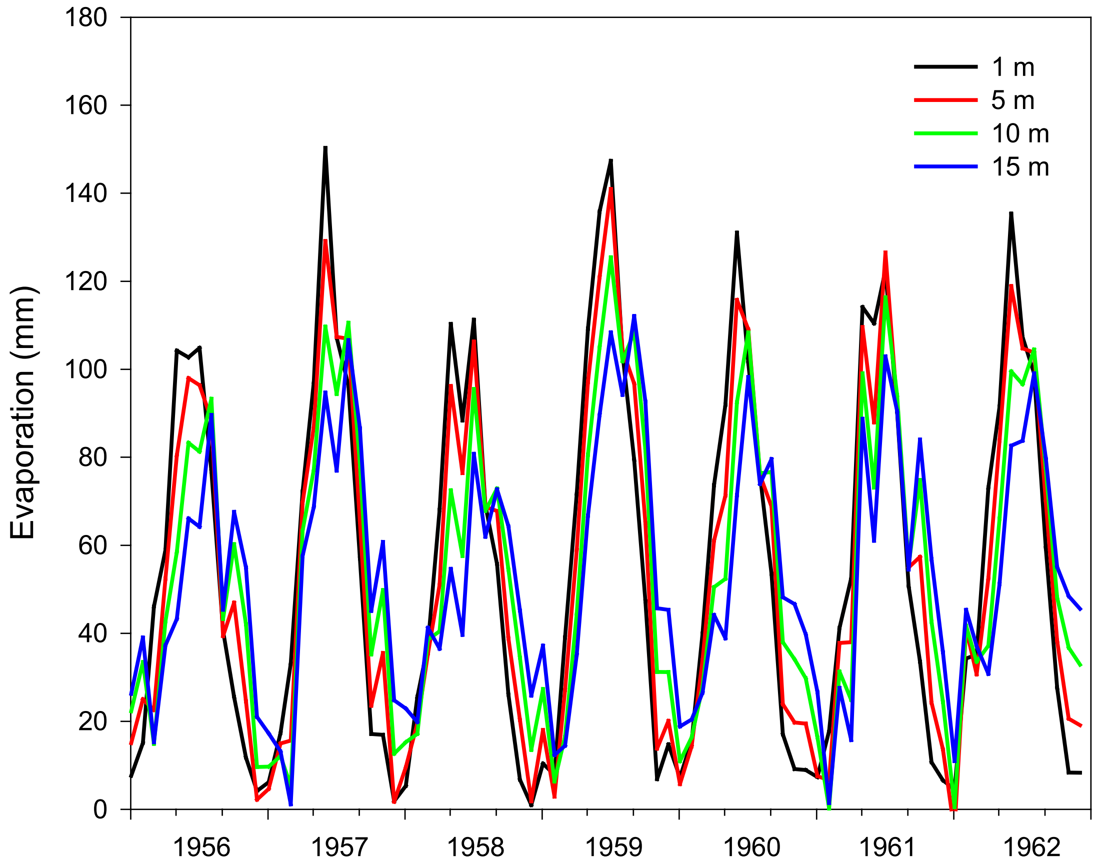

Figure 13.4 Impact of differences in water depth on the predicted evaporation at Kempton Park

The sensitivity to the value of the roughness length is such that a decrease in the roughness length by an order of magnitude results in a decrease in the estimated evaporation by about 7% whilst an increase by an order of magnitude results in a corresponding increase in evaporation by about 17%. This appears to mean that there is quite a tolerance to uncertainty in the roughness length but in fact, since there is little relevant information, there is no value that can be recommended with any real confidence. However, the value of 0.001 used for Kempton Park is based on observations and so, until research is carried out to throw more light on the problem, this value should be used.

13.2.3 Water depth

Potentially, the water depth can vary over a wide range of values. However, the equilibrium temperature model is based on the assumption that the water body is well mixed, i.e. it does not become stratified. Therefore, estimates of evaporation cannot be relied upon for water bodies deeper than 10 m and possibly as shallow as 5 m. Also, although the maximum depth of water will probably be known, the actual depth may be less well known as it will vary with time according to inflows and outflows.

The effect of the water depth on evaporation rates is through the heat storage in the water body. The deeper the water, the larger is this term. This is illustrated by using the meteorological data for Kempton Park to predict the evaporation from water bodies with depths of 1,5, 10 and 15 m, Figure 13.4. It can be seen that the effect of increasing water depth is to decrease evaporation rates during the summer months and increase it during the winter months. There is also an increasing phase shift with the peak evaporation rates occurring later in the year.

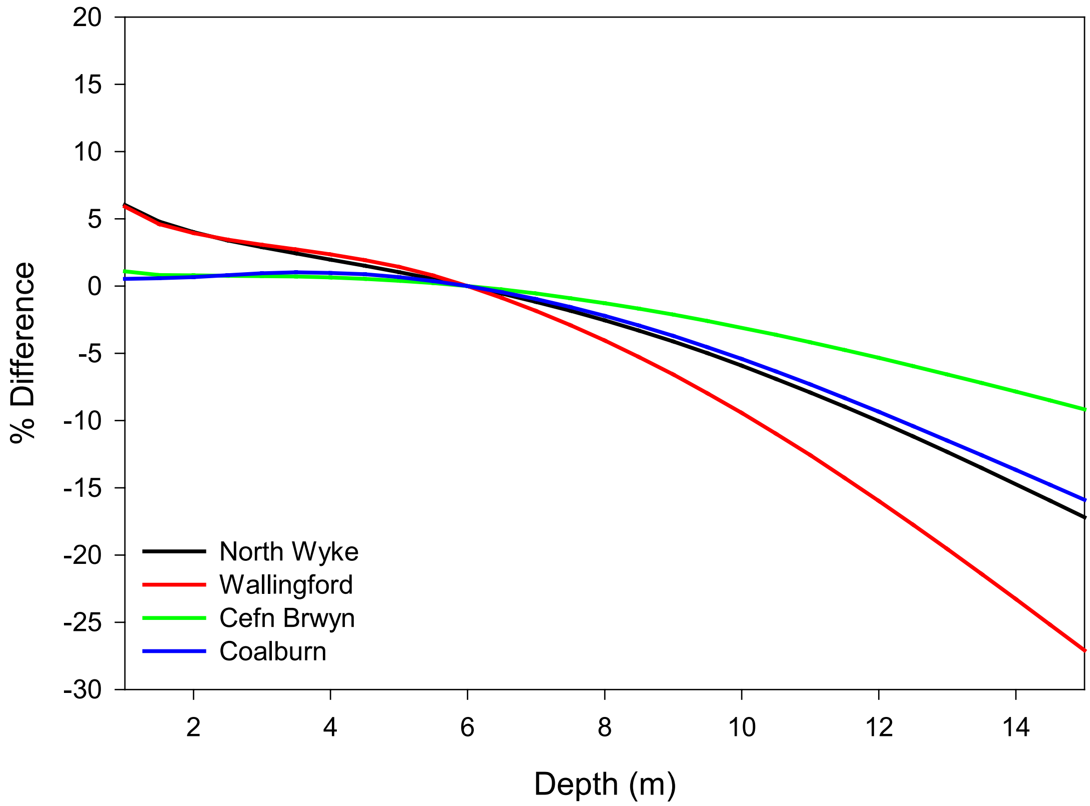

Figure 13.5 Percentage difference in predicted mean monthly evaporation with changes in the water depth.

The sensitivity analysis showed that the daily and annual evaporation values were very insensitive to variations in the water depth. This is because the daily variability is dominated by the day-to-day variations in the driving variables, such as wind speed. The annual variability tends to be low because there is little variability in the driving variables at this time scale. There is a greater sensitivity at the monthly time scale, Figure 13.5, reflecting the seasonal variations in the driving variables. There is a significant variation between the stations, presumably reflecting different seasonal trends in the driving variables. It might be expected that the strong seasonality in the incoming solar radiation would be reflected in the results. This seems to be partly borne out because there is a trend of decreasing sensitivity in the predicted evaporation from the station at Wallingford (which has high incoming solar radiation and low wind speeds) to the station at Cefn Brwyn (low solar radiation and high wind speed).

However, the predicted monthly evaporation values are not particularly sensitive to the water depth and a 5% difference greater or less than the values predicted for a water depth of 6 m is within the depth range of 1 to 8 m. Therefore, a rough estimate of the water depth is sufficient.

It should be noted that this analysis presumes that the water body does not become stratified. The evaporation rates predicted by the equilibrium temperature model and the measured values at the Kempton Park reservoir (which is known to have become thermally stratified) are in good agreement which suggests that a significant error is unlikely with a water depth of 6 m. Although it cannot be demonstrated, it is probably reasonable to assume that thermal stratification can be ignored up to depths of 10 m.

13.3 Summary of results

- Both the equilibrium temperature model and the use of empirical factors with MORECS grass PE are more sensitive to errors in the incoming solar radiation than the other driving variables, and least sensitive to the wind speed. However, when PETCALC PE is used with empirical factors then it is most sensitive to the air temperature.

- Small differences were observed in the sensitivity of the equilibrium temperature model due to differences in climate / magnitude of the driving variables.

- The albedo of the water body needs to be known to ±0.032 to attain an accuracy of ±5% in the estimates of evaporation with the equilibrium temperature model and, if there is no evidence to the contrary, a value of 0.065 should be used.

- There is little information on what the appropriate value for the roughness length is with the result that, although the model is not particularly sensitive to this parameter, there is considerable uncertainty over which value to use. A value of 0.001 m should be used because the studies with the data from Kempton Park using this value gave the best agreement with .

- The depth of water has little impact on the estimates of evaporation at the daily or annual time step but is potentially significant at the monthly time step due to the heat storage However an estimate ±25% is sufficient.

14. THE EFFECT OF CLIMATE ON ESTIMATES OF OPEN WATER EVAPORATION USING EMPIRICAL FACTORS

The empirical factors calculated from the data for Kempton Park can only be assumed to be accurate at that location. The effect of differences in climate on estimates of open water evaporation made using the empirical factors method has been investigated using data from the ten meteorological stations listed in Table 14.1 (Their locations are shown on the topographic map in Figure 14.1). In the absence of measured evaporation, this has been achieved by calculating times series of open water evaporation, using the equilibrium temperature model with a daily time step and assuming the same water depth as at Kempton Park (i.e. 6 m), from the meteorological data for each station. These have been assumed to be accurate values of the open water evaporation at the locations of the meteorological station, i.e. the sole source of variability in evaporation is the meteorological data because the water body and its surroundings are identical to Kempton Park. Thus, there is a reasonably high probability that the assumption is valid. Daily time series of potential evaporation for simulated MORECS grass PE and PETCALC were then calculated and the empirical factors determined from the Kempton Park data applied. For the purpose of comparison, the Penman (1948) factors were also applied to the MORECS grass PE. Two measures of the differences between the estimates of evaporation using the equilibrium temperature model and those from the empirical factors have been used. The root mean square error (RMSE) measures systematic and non-systematic errors whilst the mean bias error (MBE) measures systematic errors. It should be noted that ten stations must be considered as insufficient to draw firm conclusions throughout England and Wales. These results should only be seen as indicative.

Consideration was given to calculating empirical factors for each of the meteorological stations by assuming that the evaporation rates predicted by the equilibrium temperature model were accurate values for open water. This was rejected for two reasons. Firstly, it has only been demonstrated at one location that the equilibrium temperature model is able to simulate the observed evaporation reasonable accurately. Secondly, it is not at all clear how the results from the ten stations could be reliably extrapolated to cover England and Wales.

Table 14.1 Locations of meteorological stations used to test altitude corrections

Name	Easting	Northing	Elevation (m AOD)	Start date
Bicton	307300	86100	51	1 Jan 1988
Brooms Barn	575300	265600	75	1 Jan 1988
Cefn Brwyn	282540	283870	359	1 Jan 1989
Coalburn	369670	578320	292	1 Jan 1989
Eistaddfa Gurig	280320	285380	529	1 Jan 1989
High Mowthorpe	488800	468500	175	1 Jan 1994
North Wyke	267050	96580	180	1 Jan 1989
Rothamsted	513200	213400	128	1 Jan 1988
Sutton Bonington	450700	325900	48	1 Jan 1988
Wallingford	461520	189630	47	1 Jan 1988

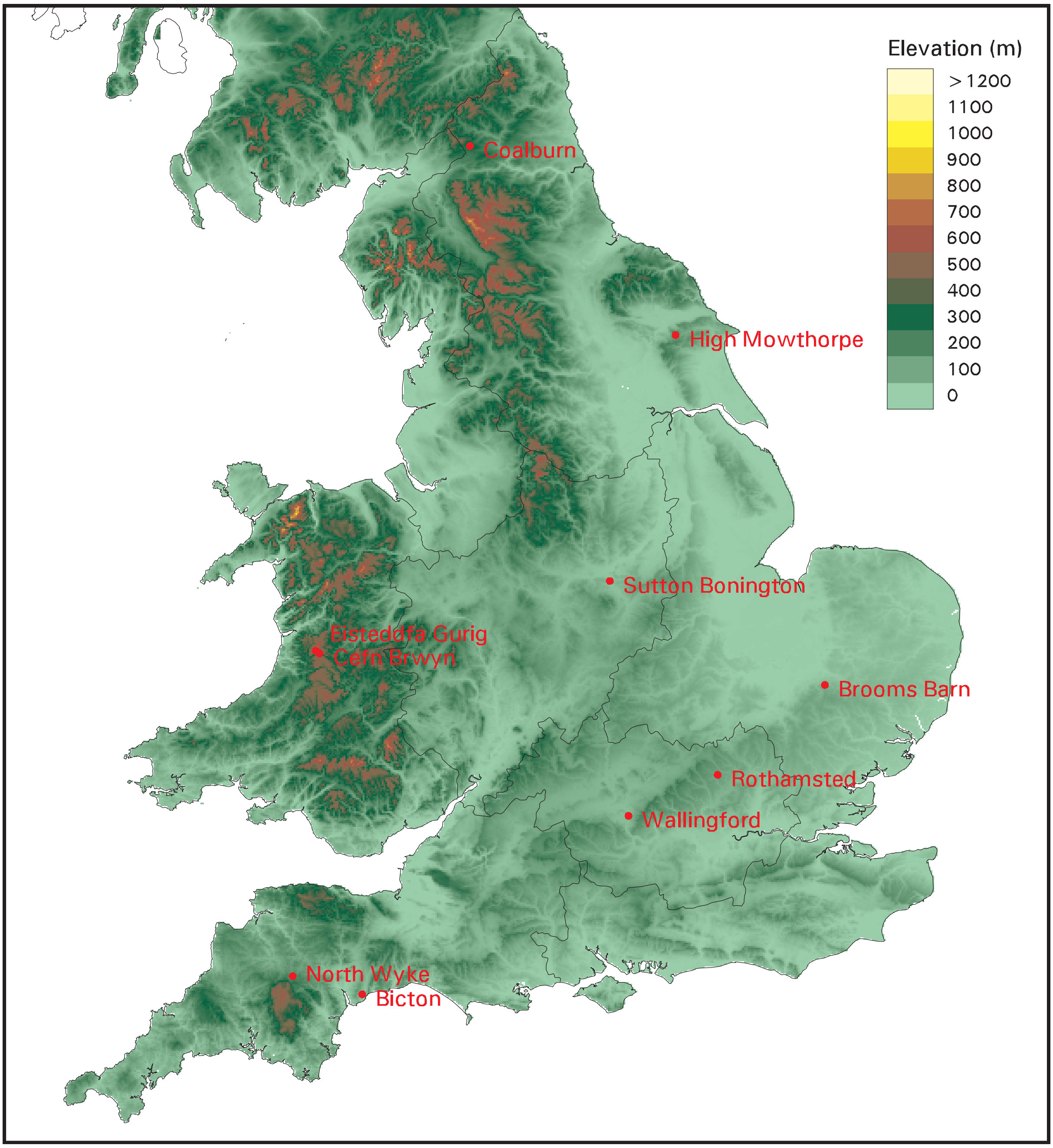

Figure 14.1 Locations of meteorological stations.

Daily values have not been included in this analysis because, when estimated daily values of evaporation are compared with measured values, considerable variability is found. This is because an "average" transmission is assumed for the clouds. However, if the values over periods of five to ten days are compared then good agreement is usually observed. Therefore, it was felt that no clear conclusions would result if daily values were considered.

14.1 Annual values

The results for annual total evaporation are given in Table 14.2. The values for Kempton Park are included as a basis for comparison. When the results from all the stations are considered, then the estimates based on the simulated MORECS grass PE do not show any significant bias whereas those based on PETCALC PE do (the latter generally overestimating the annual evaporation). The most likely explanations for this are the difference in the aerodynamic term and/or the method of calculating the net long-wave radiation in the Penman-Monteith and Penman models which are the basis for these two systems respectively.

The high errors in the estimates based on the Penman (1948) factors used with MORECS grass PE demonstrates that using a set of empirical factors with estimates of grass PE for which they were not developed is ill-advised.

The systematic errors in estimates based on the simulated MORECS grass PE with the optimum empirical factors show a wide range of values between the stations, with a maximum of –23.2% and there is a tendency for the non-systematic errors to be between 10 and 20%. There is some indication that the errors are a function of the location of the station as there are correlations between the MBE and the station altitude (r^2 = -0.83) and easting (r^2 = 0.71) and the RMSE with northing (r^2 = 0.70) but these correlations cannot be considered particularly significant, given the small sample size.

The errors in estimates based on PETCALC show similar characteristics to those from the simulated MORECS grass PE, although the values tend to be higher, as discussed above. However, there are no correlations between the errors and the locations of the stations.

Table 14.2 Percentage errors in estimates of annual open water evaporation using empirical factors (negative values occur when estimates based on empirical factors are less than those from the equilibrium temperature model)

Met. Station	MORECS grass PE × optimum factors		PETCALC PE × optimum factors		MORECS grass PE × Penman factors	
	% MBE	% RMSE	% MBE	% RMSE	% MBE	% RMSE
Kempton Park	3.0	4.4	4.5	5.5	43.2	43.3
Bicton	1.7	9.9	-22.3	22.5	58.5	59.0
Brooms Barn	17.3	17.8	21.6	23	73.7	74.0
Cefn Brwyn	-15.3	15.9	18.7	19.1	50.4	50.7
Coalburn	-22.5	26.0	22.4	26.9	53.8	54.4
Eistaddfa Gurig	-23.2	25.7	15.9	28.2	50.0	53.9
High Mowthorpe	1.9	23.3	38.2	40.8	97.8	101.5
North Wyke	0.9	4.2	21.8	22.3	69.7	69.9
Rothamsted	18.7	19.2	22.5	23.9	80.7	81.1
Sutton Bonington	8.0	8.6	17.7	18.6	61.1	61.3
Wallingford	16.5	17.0	-6.1	8.3	74.2	74.5
Average	0.6	15.6	15.0	23.4	64.8	65.8

The results suggest that annual estimates of open water evaporation using the Kempton Park empirical factors and MORECS grass PE should be regarded as having an accuracy of ±15% whilst with PETCALC it is ±25%.

14.2 Monthly values

As might be expected, the errors are significantly higher when the monthly values are considered. The details of the following discussion apply to estimates of evaporation based on the simulated MORECS grass PE but the general comments also apply to PETCALC PE. During the summer months, May to October, the average MBE is small, -3% (standard deviation 47%), although the average RMSE is larger, 63% (standard deviation 128%). However, during the winter months the errors are much larger; the average MBE is –77% (standard deviation 662%), and the average RMSE is larger, 855% (standard deviation 1920%). At first sight these figures may seem alarmingly high but they are due partly to a few exceptionally large values which tend to distort the average values. Therefore, the modal values are more useful than the averages. For the summer months, the modal MBE is 11% and the modal RMSE is 54% whilst, for the winter months, the modal MBE is 102% and the modal RMSE is 205%. The higher values in the winter months are less significant than it might first seem because this is the period when evaporation rates are low. The source of the higher errors is that the equilibrium model predicts negative evaporation rates (i.e. condensation) for some days during the winter which the use of empirical factors cannot duplicate. For this reason, estimates of evaporation based on the empirical factors method are particularly unreliable during March when the temperature of the water body is likely to be low whilst the temperature of the air is likely to be higher.

14.3 Summary of results

These are based on the assumption that the equilibrium temperature model accurately predicts the evaporation from open water.

- Annual estimates of open water evaporation using optimum empirical factors and MORECS grass PE should be regarded as having an accuracy of ±15%, due to differences in climate associated with a meteorological stations location.
- Annual estimates of open water evaporation using optimum empirical factors and PETCALC PE have an accuracy of ±25%, due to differences in climate associated with a meteorological stations location, and have a consistent trend of overestimating the values.
- Monthly estimates using either potential evaporation data set have an accuracy of ±50% during the summer months (July to October inclusive) and ±200% during the remainder of the year, due to differences in climate associated with a meteorological stations location.

15. ACCURACY OF ALTITUDE CORRECTIONS

Meteorological data will rarely be available specifically from the site where an estimate of open water evaporation is required. The norm will be to use data from the nearest meteorological station that is in a similar climatic region and apply corrections for differences in altitude. This Chapter investigates the likely impact of these procedures on estimates of open water evaporation. However, it should be noted that, as the analysis is based on data from only ten meteorological stations, the results should be considered as qualitative rather than quantitative.

Daily values have not been included in this analysis because, when estimated daily values of evaporation are compared with measured values, considerable variability is found. This is because an “average” transmission is assumed for the clouds. However, if the values over periods of five to ten days are compared then good agreement is usually observed. Therefore, it was felt that no clear conclusions would result if daily values were considered.

15.1 Correction of the driving variables

The accuracy of evaporation estimates calculated using driving variables corrected for altitude, using the procedures described in Section 6.2.1 has been investigated. This has been achieved by calculating time series of the open water evaporation, with the equilibrium temperature model, using the meteorological data from a station. It has been assumed that this is an accurate estimate of open water evaporation for that site. Time series of open water evaporation were then calculated using the meteorological data from each of the other stations, correcting the driving variables for altitude (but using the incoming radiation from the ‘reference’ station so as to only deal with variables for which an altitude correction was possible). The lapse rates used were -0.006ºC/m for air temperature, -0.025 kPa/m for vapour pressure and +0.006 m s^{-1}/m for wind speed. This procedure has been applied to the data from each of the meteorological stations listed in Table 14.1 in turn.

A similar procedure has been used to assess the potential impact of differences in the incoming radiation. Time series of open water evaporation have been calculated using the incoming radiation data from each of the other stations (but using the driving variables from the ‘reference’ station). This procedure has been applied to the data from each of the meteorological stations in turn.

The root mean square error (RMSE) has been used as the measure of the accuracy of the corrections.

As a basis for comparison, the RMSEs in estimating the annual evaporation using the data without any corrections have been calculated and are given in Table 15.1.

15.1.1 Driving variables excluding incoming radiation

The errors in estimating the annual evaporation totals are given in Tables 15.1 and 15.2. These show that, although the errors are generally ±25% (the average is ±28%), errors as large as ±72% can arise if data from an inappropriate meteorological station are used. However, when the results for stations that can be considered as climatically similar (e.g. Wallingford and Rothamsted, Cefn Brwyn and Eistaddfa Gurig) are examined then the RMSE is between 5% and 44%. This emphasises the importance of using data from meteorological stations representative of the site for which the estimate of evaporation is required.

Table 15.1 Percentage RMSE in annual evaporation when using uncorrected meteorological data

Transposed Meteorological Station Reference Meteorological Station	Bicton	Brooms Barn	Cefn Brwyn	Coalburn	Eistaddfa Gurig	High Mowthorpe	North Wyke	Rothamsted	Sutton Bonington	Wallingford
Bicton		36.7	27.4	70.6	39.8	82.9	21.6	49.2	17.8	42.1
Brooms Barn	26		8.5	28.5	10	42	10.1	10.8	14.1	6.5
Cefn Brwyn	20.8	9.7		37	16.3	49	6.9	18.7	9.6	14.1
Coalburn	40.5	21.2	26.4		20.3	15.2	29.2	14.6	31.4	18.1
Eistaddfa Gurig	27.8	9.2	13.3	27.9		38.5	14.9	12.7	18.4	10
High Mowthorpe	39.5	26	28.9	15	24.1		30.5	22.5	33.4	24.8
North Wyke	17.3	11.4	7.6	42.6	18.4	53.3		22.2	5.2	17.3
Rothamsted	32.3	9.6	15.1	18.4	10.8	32.3	17.9		21.7	5.1
Sutton Bonington	14.5	16.6	11.1	47.5	23.5	60.8	5.7	27.9		22.1
Wallingford	29	5.9	11.6	23.7	9.4	38	14.4	5.4	17.9	

This is confirmed if the values in Table 15.2 are compared with those resulting from no correction at all being applied, Table 15.1. Many of the errors increase if a correction is applied. In some cases this is not unreasonable as the implicit assumption being made in applying the corrections is that the location of the meteorological station and that for which an estimate of evaporation is required are subject to the same air mass. This is clearly not a valid assumption when, for example, using data from Bicton to estimate the evaporation at Coalburn. However, the errors also increase when the two sites can be assumed to be in the same air mass (e.g. Cefn Brwyn and Eistaddfa Gurig, Wallingford and Rothamsted and Bicton and North Wyke). This suggests that other, local factors (e.g. exposure, surrounding land cover etc.) can also result in significant variations in the driving variables.

The monthly values have generally the same features but with a distinct seasonal difference. During the summer months (May to October incl.) the errors are much lower than during the winter months. For stations that can be considered climatically similar, the average errors during the summer months are ±15%. During the winter months this rises to an average error of ±120% with the highest errors occurring in January and February. The errors are significantly higher when data from stations that are clearly not in the same climatic region are used. During the summer months the average error is ±35% but during the winter months this increases to ±250%.

No statistically significant relationships were found between the RMSEs and the relative locations of the stations represented by: difference in altitudes, distance apart, easting difference and northing difference.

Table 15.2 Percentage RMSE in annual evaporation when using altitude corrected driving variables (excluding incoming solar radiation)

Corrected Meteorological Station Reference Meteorological Station	Bicton	Brooms Barn	Cefn Brwyn	Coalburn	Eistaddfa Gurig	High Mowthorpe	North Wyke	Rothamsted	Sutton Bonington	Wallingford
Bicton		53	48.9	67.1	62.6	71.7	44.4	71.2	17.2	60.7
Brooms Barn	37.4		13.6	22.7	14.4	31.4	11.8	14.5	33.5	9.6
Cefn Brwyn	31.9	10.3		16.8	15	31.9	9.9	20	26.8	17.4
Coalburn	37.2	11.5	13.9		19.2	24.2	13.7	15.7	31	11.8
Eistaddfa Gurig	30.4	13.2	12.2	23.2		36.1	12	26.5	25	23.6
High Mowthorpe	27.5	26.3	19.6	24	25.1		14.7	37.9	30	36.7
North Wyke	32.3	13.8	13.1	21.7	16.9	24.6		25	29.1	22.6
Rothamsted	44	12.7	20.1	31.6	18.3	31.6	18.1		41.6	5.6
Sutton Bonington	15.4	40.6	32.8	43.5	43.3	65.7	31.8	58		50.8
Wallingford	41.5	9.2	18	27.2	18.1	35.7	16.6	6.3	38.9	

15.1.2 Incoming radiation

The errors in estimating the annual evaporation totals, using radiation values from another station, are given in Table 15.3. These show lower values, approximately half, than is the case with the other driving variables. The highest error is 32% but the average error is ±15% and the average error when considering climatically similar stations is ±4%. This suggests differences in altitude may not have much impact on that the incoming radiation, within the same climate region, and so using the data from a meteorological station to estimate annual evaporation at a site in the same climatic situation results in remarkably small errors.

The errors in estimating the monthly evaporation are also approximately half those for the other driving variables and show the same seasonal trend of lower errors in summer than winter. During the summer months (May to October incl.) the average error is ±50% when all stations are included but falls to ±7% when only climatically similar stations are considered. During the winter months the errors are significantly bigger, on average ±200% for all stations and ±150% for climatically similar stations.

No statistically significant relationships were found between the RMSE and the relative locations of the stations represented by: difference in altitudes, distance apart, easting difference and northing difference.

Table 15.3 Percentage RMSE in annual evaporation when using incoming radiation from another station

Transposed Meteorological Station / Reference Meteorological Station	Bicton	Brooms Barn	Cefn Brwyn	Coalburn	Eistaddfa Gurig	High Mowthorpe	North Wyke	Rothamsted	Sutton Bonington	Wallingford
Bicton		12.5	15.1	5.3	14.7	22.4	15	18.1	5.8	15.1
Brooms Barn	11		20.8	13.4	20.3	29.9	15.9	4.3	6	2.4
Cefn Brwyn	9	23.9		15.4	6.7	22.2	5.4	30.2	14.8	27
Coalburn	14.5	11.9	23.5		20.3	20.8	19.5	14.6	13.1	14.1
Eistaddfa Gurig	15.3	24.1	14.8	16.4		17.3	11.7	28.1	18.7	25.7
High Mowthorpe	13.6	15.1	12.8	17	18		11.6	16.5	9	14.8
North Wyke	10	21.6	6.1	10.7	5.5	21.3		27.2	14.2	24.1
Rothamsted	13.7	4.2	23.8	17.9	23.4	32.7	18.6		9.3	3
Sutton Bonington	6.1	6.5	16.4	6.4	16.1	25.7	13.7	11.4		8.7
Wallingford	12.2	2.6	22.1	14.6	21.8	30.4	17.5	2.8	7.7	

15.2 Correction of estimated evaporation

The accuracy of evaporation estimates corrected for altitude, using the procedures described in Section 6.2.2 has been investigated. This has been achieved by calculating time series of the open water evaporation, with the equilibrium temperature model, using the meteorological data from a station. It has been assumed that this is an accurate estimate of open water evaporation for that site. Time series of open water evaporation were then calculated using the meteorological data from each of the other stations and the daily values corrected for altitudinal differences using the lapse rates given in Table 15.4 (see Section 6.2.2. for the source of these). This procedure has been applied to the data from each of the meteorological stations listed in Table 13.1 in turn.

The root mean square error (RMSE) has been used as the measure of the accuracy of the corrections.

The results show that these corrections are effective in the majority of cases with a significant reduction in the RMSE, compared to Table 15.1. The average error is ±24%, varying between ±3% and ±73%. For stations that can be assumed to be climatologically similar the RMSE is reduced by between half and a quarter (compared with applying no correction), e.g. correcting estimates of open water evaporation from Cefn Brwyn to estimate it at Eistaddfa Gurig or Rothamsted to estimate it at Wallingford. However, when the evaporation data is from stations that are in distinctly different climate regions then the correction has very little effect or may increase the error.

The monthly values have generally the same features but with a distinct seasonal difference. During the summer months (May to October incl.) the errors are much lower than during the winter months. For stations that can be considered climatically similar, the average errors during the summer months are ±20%. During the winter months this rises to an average error of ±150% with the highest errors occurring in January and February. The errors are significantly higher when data from stations that are clearly not in the same climate region are used. During the summer months the average error is ±50% but during the winter months this increases to ±250%.

The analysis indicates that the errors arising when environmental lapse rates are applied to the driving variables are greater than the errors arising when the evaporation estimates themselves are adjusted for altitude.

Table 15.4 Mean monthly lapse rates for evaporation

	Lapse rate (mm m^{-1})	Standard Error (mm m^{-1})
January	-0.0143	0.0012
February	-0.0140	0.0009
March	-0.0180	0.0015
April	-0.0237	0.0024
May	-0.0344	0.0038
June	-0.0314	0.0046
July	-0.0388	0.0061
August	-0.0411	0.0051
September	-0.0316	0.0028
October	-0.0225	0.0017
November	-0.0177	0.0015
December	-0.0136	0.0012

Table 15.5 Percentage RMSE in annual evaporation when correcting the evaporation for altitude

Corrected Meteorological Station / Reference Meteorological Station	Bicton	Brooms Barn	Cefn Brwyn	Coalburn	Eistaddfa Gurig	High Mowthorpe	North Wyke	Rothamsted	Sutton Bonington	Wallingford
Bicton		35.0	11.5	49.4	11.7	72.7	14.4	43.2	18.0	42.4
Brooms Barn	24.7		25.2	13.0	30.2	34.8	16.5	7.1	12.5	8.2
Cefn Brwyn	9.2	27.8		43.0	11.5	65.0	9.7	36.2	12.4	36.3
Coalburn	27.8	9.9	30.8		36.4	19.6	22.1	9.6	16.9	10.3
Eistaddfa Gurig	8.8	30.4	10.9	47.9		66.8	13.0	39.9	16.0	39.5
High Mowthorpe	33.6	21.9	40.1	16.4	44.8		30.9	21.3	26.9	21.8
North Wyke	11.4	18.6	9.5	32.7	13.7	53.8		26.2	4.8	27.1
Rothamsted	28.3	6.2	30.1	10.8	35.9	29.6	21.3		16.7	2.7
Sutton Bonington	14.7	14.7	13.0	26.3	17.7	50.4	5.0	21.6		22.2
Wallingford	29.2	7.5	31.6	11.9	37.4	30.4	22.8	2.7	18.0	

15.3 Summary of results

- The errors in estimating the annual evaporation at a site, using meteorological data from another site but applying a correction to the air temperature, vapour pressure deficit and wind speed for altitude, are between ±5 and ±45% when using data from a station in the same climate region. These errors rise to an average of ±25% and a maximum of ±70% when the station is not in the same region.

- The errors in estimating the monthly evaporation at a site, using meteorological data from another site but applying a correction to the air temperature, vapour pressure deficit and wind speed for altitude show a strong seasonality with much lower errors in the summer months (May to October incl.) than in the winter. For stations in the same climate region, errors in the estimated monthly evaporation are of ±15% during the summer and ±120% during the winter months (this corresponds to about ± 17 mm and ± 13 mm respectively).

- Using the incoming radiation from a meteorological station to estimate the annual evaporation at a site in the same climate region results in relatively low errors, generally ±4%. When monthly values are estimated there is a strong seasonality to the errors with low values during the summer months, of ±7%, but much larger during the winter of ±150% (this corresponds to about ± 8 mm and ± 17 mm respectively).

- Significant variations in estimated open water evaporations result from factors local to the meteorological station and cannot be accounted for by simple altitudinal corrections.

- Correcting the daily evaporation values for differences in altitude is reasonably effective. The average error is ±25%, comparable with that achieved using lapse rates for the driving variables. When using data from a station in the same climate region the error varies between ±3% and ±14% but rises to ±73% if data from a station in a different climate region is used. The errors in monthly estimates of open water evaporation show a similar result but also have a strong seasonality with low values in the summer months and high values in the winter.

- The errors arising when environmental lapse rates are applied to the driving variables are greater than the errors arising when the evaporation estimates themselves are adjusted for altitude. This finding, taken together with the relative amount of work involved using the two alternative methods, indicates that it would be preferable to use the altitude corrections applied directly to evaporation estimates.

16. PROJECT CONCLUSIONS AND RECOMMENDATIONS

The objectives of this project are:

- evaluate current methods of estimating open water evaporation;
- recommend the best available practicable methodologies for producing robust estimates;
- assess the associated uncertainty of these methodologies.

Phase 2 is concerned with the last two.

This Chapter includes the conclusions and recommendations from both phases of the R&D project.

16.1 Conclusions

1. The review of methods of estimating open water evaporation identified seven methods; pan evaporation, mass balance, energy budget models, bulk transfer models, combination models, the equilibrium temperature method and empirical factors.

2. Pan evaporation and empirical factors are very similar in that they employ factors to convert 'standard' time series of evaporation into estimates of open water evaporation for specific water bodies. As such there is a lack of physical rigour and considerable uncertainty as to which values to use for the factors.

3. The mass balance method requires considerable investment in collecting data and is not appropriate if the evaporation losses are comparable in size to other changes in storage. The energy balance method is generally accepted as the most accurate method but requires considerable resources to accurately measure all the components of the balance. Nevertheless, the two balance methods are the only methods that consider the water body as a whole. In application, the other methods are one-dimensional and are assumed to be representative of the water body.

4. The mass transfer method requires measurements of surface temperature and is sensitive to errors in the vapour pressure data and the formulation of the wind function. The combination models have gained wide acceptance but, when applied to open water deeper than 0.5 m and for time intervals less than a year, it is necessary to use the available energy, rather than the net radiation, in order to take the heat storage into account. This requires measurements of surface temperature.

5. The equilibrium temperature method requires the same meteorological data as combination models but can take the heat storage into account, provided that the water body does not become thermally stratified. As such it combines physical rigour with a requirement for time series of meteorological data that are readily available.

6. The empirical factors method is used to produce estimates of open water evaporation in all Areas and Regions. However, the values of the factors and data sets employed vary between Regions and may vary between Areas. Thus, although there is a consensus on the method, there is little on how it is used.

7. A ranking of the seven methods of estimating open water evaporation against nine criteria established that the equilibrium temperature method would best serve the Agency's

purposes. The use of empirical factors and the combination models were ranked equal second.

8. The spatial variability of the meteorological variables that drive evaporation is strongly influenced by proximity to the coast. However, inland, the spatial variability of wind speed and air temperature is low while incoming solar radiation and relative humidity show significantly more variability. Topography has a very strong effect on the driving variables, either directly, in terms of the lapse rates of air temperature and vapour pressure, or indirectly, through the formation of clouds affecting the amount of incoming solar radiation.

9. The size of the water body has an effect on evaporation rates in several ways. There is evidence that the rate of evaporation over water is enhanced due to increased wind speed resulting from the smoother surface. The effect of the transfer of the airstream from the surrounding land surface to the water surface extends a maximum of ten metres from the edge. No means of taking these effects into account has been found. The main impact of the size of the water body in England and Wales is the development of thermal stratification. The maximum depth of the warmer surface layer is a function of the surface area of the water body. However, studies of thermal stratification published in the literature have tended to be site specific and so there is a lack of information on how thermal stratification affects the evaporation rates in general terms.

10. Four methods of estimating open water evaporation (the equilibrium temperature model, Penman (1948) and Penman-Monteith models and empirical factors using two different sets of factors) have been tested against the measurements of evaporation made at a reservoir, at Kempton Park, between 1956 and 1962 The results show that the values predicted by the equilibrium temperature model are significantly more accurate than those produced by any of the other methods. The Penman (1948) and Penman-Monteith models lack of accuracy is attributed to two causes: (i) not taking into account the heat storage in the water body and (ii) assuming that the air temperature can be used to estimate the outgoing long-wave radiation. The FAO-56 empirical factors are totally inappropriate to UK conditions. However, the empirical factors of Penman (1948b) applied to MORECS grass PE do provide reasonably accurate values at this location (but see Section 14.1).

11. The measurements of evaporation at Kempton Park have been used to calculate empirical factors for estimating open water evaporation using the MORECS grass PE, PENSE and PETCALC data available in the Agency.

12. A sensitivity analysis has shown that both the equilibrium temperature model and the use of empirical factors with MORECS grass PE are more sensitive to errors in the incoming solar radiation than the other driving variables, and least sensitive to the wind speed. However, when PETCALC PE is used with empirical factors then it is most sensitive to the air temperature.

13. Three parameters are required by the equilibrium temperature model: the albedo, roughness length and depth of the water body. In order to achieve an accuracy of ±5% in estimates of open water evaporation, the albedo needs to be known to ±0.032 and, if no information to the contrary is available, a value of 0.065 should be used. Estimates of open water evaporation are not particularly sensitive to the roughness length but it can vary over a couple of order of magnitude. A value of 0.001 should be used unless until

more research provides more information. The water depth is not a particularly critical parameter and values to ±25% are sufficient.

14. The empirical factors calculated from the Kempton Park data have to be assumed to apply across the whole of England and Wales. However, by comparing estimates of open water evaporation made using these factors with those calculated using the equilibrium temperature model, it has been established that annual estimates of evaporation should be considered to be accurate to ±15% when the PE data set used is MORECS grass PE or PENSE and ±25% when it is PETCALC. Monthly estimates of open water evaporation have an accuracy of ±50% during the summer months (May to October incl.) and ±200% during the winter months.

15. Corrections for the effect of altitude on the driving variables (air temperature, humidity and wind speed) are available but there are none for the incoming solar radiation. Empirical corrections to evaporation estimates for altitude have also been found, although this is not as physically rigorous as correcting the driving variables.

16. Correcting the driving variables for the difference in altitude when estimating the annual evaporation at one location using the meteorological data from another results in errors of between ±5 and ±45% when using data from a station in the same climate region. Much higher errors occur if data from another climate region are used. Using the incoming radiation values, from a different site but in the same climate region, results in errors of around ±4% in the estimated annual open water evaporation. This rises to around ±19% when values from a site in a different climate region are used. The errors when estimating monthly totals are higher and have a strong seasonality, ±7% during the summer months and ±150% during the winter.

17. Correcting the daily evaporation values for differences in altitude is reasonably effective. The average error is ±25%, comparable with that achieved using lapse rates for the driving variables. When using data from a station in the same climate region the error varies between ±3% and ±14% but rises to ±73% if data from a station in a different climate region is used. The errors in monthly estimates of open water evaporation show a similar result but also have a strong seasonality with low values in the summer months and high values in the winter.

18. The errors arising when environmental lapse rates are applied to the driving variables are greater than the errors arising when the evaporation estimates themselves are adjusted for altitude. This finding, taken together with the relative amount of work involved using the two alternative methods, indicates that it would be preferable to use the altitude corrections applied directly to evaporation estimates.

16.2 Recommendations

- The Agency should adopt the equilibrium temperature model for estimating open water evaporation in view of its significantly higher accuracy than other methods. This will involve an investment in meteorological data (incoming solar radiation or sunshine hours, air temperature, wet bulb depression, wind speed and vapour pressure deficit at monthly or, preferably, daily intervals) which could be minimised if the model was incorporated within procedures for estimating evaporative losses for other purposes, e.g. estimating groundwater recharge.

- Until the equilibrium temperature model is available, estimates of open water evaporation should be made using the empirical factors calculated in this project for estimating open water evaporation from MORECS grass PE, PENSE or PETCALC. Practical guidance in the use of the empirical factors and equilibrium temperature methods by Agency staff has been provided in a handbook "Estimating open water evaporation – guidance for Environment Agency practitioners" produced as part of the output from this R&D project.

16.3 Further Research

- There is uncertainty in the appropriateness of altitude lapse rates for correcting air temperature, vapour pressure deficit and wind. This is a general problem for estimating evaporation from any land surface. Research should be carried out into improving the accuracy of these lapse rates and to determine the conditions in which they are appropriate.

- There is significant uncertainty in the values for albedo and roughness length that should be used to parameterise the equilibrium temperature model. Research should be carried out to establish the appropriate values to be used for water bodies in England and Wales for which the Agency requires estimates of evaporation.

- Further research is required into how thermal stratification affects evaporation rates. This could best be achieved by a sensitivity study using a numerical hydrothermal model. This would allow the impact of the water bodies dimensions on evaporation rates to be determined for a range of climate conditions. The model would need to be tested against real data.

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ANNEXE A – LIST OF SYMBOLS

A	available energy
a	regression parameter (gradient)
A_s	area of the water surface
b	regression parameter (intercept)
C	mass transfer coefficient
c	specific heat of water
c_a	specific heat of air
c_j	intercept of the linear regression between departures of the daily evaporation from the mean daily evaporation and the daily meteorological variable and the mean daily meteorological variable for the j^{th} month of the year
c_p	specific heat of air at constant pressure
E	evaporation rate from a water body
E_i	estimated daily evaporation on day i
$\overline{E}_m$	monthly mean daily evaporation of month m
E_p	evaporation rate of an evaporation pan
e	vapour pressure of the air at the reference height.
e_s^*	saturated vapour pressure of the air at the water surface temperature
e_p	saturated vapour pressure of the air at pan surface temperature
e_a^*	saturated vapour pressure of the air at air temperature
$f(u)$	wind function of wind speed u
F_{in}	heat fluxes associated with inflows
F_{out}	heat fluxes associated with outflows
F_P	heat inflow associated with precipitation
G	heat conduction occurring between the water and its substrate
G_s	soil heat flux
g_j	slope of the linear regression between departures of the daily evaporation from the mean daily evaporation and the daily meteorological variable and the mean daily meteorological variable for the j^{th} month of the year
H	flux of sensible heat
K	empirical constant
$K\downarrow$	incoming short-wave (solar) radiation
k	von Karman's constant
$L\downarrow$	incoming long-wave (thermal) radiation
$L\uparrow$	outgoing long-wave (thermal) radiation
L_e	effective length of the water body (km)
m	ratio of observed sunshine hours to total possible hours of sunshine in a day
N	change in the energy storage in the water
P	atmospheric pressure
P	mean rate of precipitation over a sampling period
p	cloudiness factor
Q_{ri}	surface inflow rate
Q_{ro}	surface outflow rate
Q_{gi}	groundwater and seepage inflow rate
Q_{go}	groundwater and seepage outflow rate
R_n	net input of radiation at the surface of the water body
R_n'	net radiation in units of equivalent depth of water

R_n^*	net radiation when the water temperature is equal to the wet bulb temperature
r_a	aerodynamic resistance
r_s	bulk surface resistance
S	incident short-wave radiation
S_d	incoming diffuse solar radiation
S_n	net short-wave radiation at the surface of the water body
S_t	solar radiation incident at the top of the atmosphere
S_0	incoming direct solar radiation
T_a	air temperature at a reference height
T_b	arbitrary base temperature
T_e	equilibrium temperature
T_n	wet-bulb temperature
T_s	temperature of the water at the surface
$T_{w,i}$	water temperature at the end of the current day
$T_{w,i-1}$	water temperature at the end of the previous day
t	length of the model time step
u_z	wind speed at z m above the surface
V	water stored in water body
V_i	value of the meteorological variable on day i
$\overline{V}_m$	mean daily value of the meteorological variable of month m
z	water depth
z_{mix}	summer mixing depth of the water body (m)
z_o	roughness length
z_r	height of the meteorological observations above the surface
α_S	albedo for short wave radiation
α_L	albedo for long-wave radiation
α	Priestley-Taylor coefficient
β	Bowen ratio
ε	clear-sky atmospheric emissivity
ε_m	ratio of the molecular weight for water to that for dry air
Δ	slope of the saturated vapour pressure-temperature curve at air temperature
Δt	time step
ΔT_w	change in spatially averaged temperature of the water body
Δ_w	slope of the temperature-saturation water vapour curve at the wet bulb temperature (kPa $^oC^{-1}$)
σ	Stefan-Boltzmann constant
ϕ	atmospheric pressure
ρ	density of water
ρ_a	density of air
τ	time constant
γ	psychometric constant
λ	latent heat of vaporisation
λE	flux of latent heat (evaporation rate in energy flux units)

ANNEXE B - LIST OF ABBREVIATIONS

AWS	Automatic Weather Station
CEGB	Central Electricity Generating Board
FAO	Food and Agriculture Organization of the United Nations
IACR	Institute for Arable Crop Research
IH	Institute of Hydrology
MAFF	Ministry of Agriculture Food and Fisheries
MBE	Mean bias error
MORECS	Meteorological Office Rainfall Evaporation Calculation System
NERC	Natural Environment Research Council
PE	Potential Evaporation
RMSE	Root mean square error
STWA	Severn-Trent Water Authority
WMO	World Meteorological Organization

ANNEXE C – FORTRAN-90 SUBROUTINE CODE OF THE EQUILIBRIUM TEMPERATURE MODEL

```
!***********************************************************************
      subroutine equibtemp(albedo,depth,tstep,vpd,u,ta,tn,solrad,
     & longrad,fc,tw0,z0,rn,le,deltas,tw,evap,ierr)
!***********************************************************************
!    SUBROUTINE TO CALCULATE THE EVAPORATION, USING DAILY DATA, FROM A
!    WATER BODY USING THE EQUILIBRIUM TEMPERATURE MODEL OF de Bruin,
!    H.A.R., 1982, J.Hydrol, 59, 261-274
!    INPUT:
!       ALBEDO - ALBEDO OF THE WATER BODY
!       DEPTH - DEPTH OF THE WATER BODY (m)
!       TSTEP - THE TIME STEP FOR THE MODEL TO USE (days)
!       VPD - VAPOUR PRESSURE DEFICIT (mb)
!       U - WIND SPEED (m s-1)
!       TA - AIR TEMPERATURE (deg.C)
!       TN - WET BULB TEMPERATURE (deg.C)
!       SOLRAD - DOWNWELLING SOLAR RADIATION (W m-2 per day)
!       LONGRAD - DOWNWELLING LONG WAVE RADIATION (W m-2 per day)
!       FC - CLOUDINESS FACTOR
!       TW0 - TEMPERATURE OF THE WATER ON THE PREVIOUS TIME STEP (deg.C)
!    OUTPUT:
!       RN - NET RADIATION (W m-2 per day)
!       LE - LATENT HEAT FLUX (W m-2 per day)
!       DELTAS - CHANGE IN HEAT STORAGE (W m-2 per day)
!       TW - TEMPERATURE OF THE WATER AT THE END OF THE TIME PERIOD
!          (deg.C)
!       EVAP - EVAPORATION CALCULATED USING THE PENMAN-MONTEITH FORMULA
!            (mm per day)
!       IERR - ERROR FLAG
!            0 = OK
!            1 = ALBEDO =< 0 OR => 1
!            2 = DEPTH =< 0
!            3 = AIR TEMPERATURE < WET BULB TEMPERATURE
!            4 = DOWNWELLING SOLAR RADIATION =< 0
!            5 = WIND SPEED < 0.01 m/s
!            6 = VPD =< 0
!    CONSTANTS
!       LAMBDA - LATENT HEAT OF VAPORISATION (MJ kg-1)
!       GAMMA - PSCHROMETRIC CONSTANT (kPa deg.C-1)
!       RHOW - DENSITY OF WATER (kg m-3)
!       CW - SPECIFIC HEAT OF WATER (MJ kg-1 deg.C-1)
!       RHO - DENSITY OF AIR (kg m-3)
!       CP - SPECIFIC HEAT OF AIR (KJ kg-1 deg.C-1
!       SIGMA - STEFAN-BOLTZMANN CONSTANT (MJ m-2 deg.C-4 d-1)
!       K - VON KARMAN CONSTANT
!       DEGABS - DIFFERENCE BETWEEN DEGREES KELVIN AND DEGREES CELSIUS
!       ZR - HEIGHT OF MEASUREMENTS ABOVE WATER SURFACE (m) ASSUMED TO
!          BE SCREEN HEIGHT
!    OTHERS
!       DELTAW - SLOPE OF THE TEMPERATURE-SATUARTION WATER VAPOUR CURVE
!              AT WET BULB TEMPERATURE
!              (kPa deg C-1)
!       DELTAA - SLOPE OF THE TEMPERATURE-SATUARTION WATER VAPOUR CURVE
!              AT AIR TEMPERATURE
!              (kPa deg C-1)
!       TAU - TIME CONSTANT OF THE WATER BODY (days)
!       TE - EQUILIBRIUM TEMPERATURE (deg. C)
```

```
!         WINDF - SWEER'S WIND FUNCTION
!
      implicit none
           integer ierr
      real albedo,cp,cw,degabs,depth,deltaa,deltas,deltaw,evap,
     & evappm,fc,gamma,k,lambda,le,lepm,longrad,lradj,ra,rho,rhow,rn,
     & rns,sigma,solrad,sradj,ta,tau,te,tn,tstep,tw,tw0,u,ut,vpd,vpdp,
     & windf,z0,zr
      real alambdat,delcalc,psyconst
!
!     SETUP CONSTANTS
!
      lambda=alambdat(ta)
      gamma=psyconst(100.0,lambda)
      rhow=1000.0
      cw=0.0042
      rho=1.0
      cp=1.013
      sigma=4.9e-9
      k=0.41
           degabs=273.13
           zr=10.0
!
!     INITIALISE OUTPUT VARIABLES
!
      ierr=0
           deltas=0.0
           evap=0.0
           evappm=0.0
           le=0.0
           lepm=0.0
           rn=0.0
           tw=0.0
!
!     CHECK FOR SIMPLE ERRORS
!
      if (albedo.le.0.0.or.albedo.ge.1.0) then
             ierr=1
             return
           endif
           if (depth.le.0) then
             ierr=2
             return
           endif
           if (tn.gt.ta) ierr=3
           if (solrad.le.0.) ierr=4
           ut=u
           if (ut.le.0.01) then
             ierr=5
             ut=0.01
           endif
           if (vpd.le.0.0) then
             ierr=6
             vpd=0.0001
           endif
!
!     CONVERT FROM W m-2 TO MJ m-2 d-1
!
      sradj=solrad*0.0864
      lradj=longrad*0.0864
```

```
!
!    CONVERT FROM mbar TO kPa
!
    vpdp=vpd*0.1
!
!   CALCULATE THE SLOPE OF THE TEMPERATURE-SATURATION WATER VAPOUR CURVE
!   AT THE WET BULB TEMPERATURE (kPa deg C-1)
!
    deltaw=delcalc(tn)
!
!   CALCULATE THE SLOPE OF THE TEMPERATURE-SATURATION WATER VAPOUR CURVE
!   AT THE AIR TEMPERATURE (kPa deg C-1)
!
    deltaa=delcalc(ta)
!
!    CALCULATE THE NET RADIATION FOR THE WATER TEMPERATURE (MJ m-2 d-1)
!
    lradj=fc*sigma*(ta+degabs)**4*(0.53+0.067*sqrt(vpd))
         rn=sradj*(1.-albedo)+lradj-fc*(sigma*(ta+degabs)**4+
         &   4.*sigma*(ta+degabs)**3*(tw0-ta))
!
!    CALCULATE THE NET RADIATION WHEN THE WATER TEMPERATURE EQUALS THE
!         WET BULB TEMPERATURE. ASSUMES THE EMISSIVITY OF WATER IS 1
!    (MJ m-2 d-1)
!
         rns=sradj*(1.-albedo)+lradj-fc*(sigma*(ta+degabs)**4+
         &    4.*sigma*(ta+degabs)**3*(tn-ta))
!
!    CALCULATE THE WIND FUNCTION (MJ m-2 d-1 kPa-1) USING THE METHOD OF
!         Sweers, H.E., 1976, J.Hydrol., 30, 375-401, NOTE THIS IS FOR
!         MEASUREMENTS FROM A LAND BASED MET. STATION AT A HEIGHT OF 10 m
!    BUT WE CAN ASSUME THAT THE DIFFERENCE BETWEEN 2 m AND 10 m IS
!    NEGLIGIBLE
!
    windf=(4.4+1.82*ut)*0.864
!
!    CALCULATE THE TIME CONSTANT (d)
!
    tau=(rhow*cw*depth)/
         &    (4.0*sigma*(tn+degabs)**3+windf*(deltaw+gamma))
!
!    CALCULATE THE EQUILIBRIUM TEMPERATURE (deg. C)
!
    te=tn+rns/(4.0*sigma*(tn+degabs)**3+windf*(deltaw+gamma))
!
!    CALCULATE THE TEMPERATURE OF THE WATER (deg. C)
!
    tw=te+(tw0-te)*exp(-tstep/tau)
!
!    CALCULATE THE CHANGE IN HEAT STORAGE (MJ m-2 d-1)
!
    deltas=rhow*cw*depth*(tw-tw0)/tstep
!
!    z0 - ROUGHNESS LENGTH
!    DUE TO SMOOTHNESS OF THE SURFACE THE ROUGHNESS LENGTHS OF MOMENTUM
!    AND WATER VAPOUR CAN BE ASSUMED TO BE THE SAME
!
    z0=0.001
!
!    CALCULATE THE AERODYNAMIC RESISTANCE ra (s m-1)
```

```
!
    ra=alog(zr/z0)**2/(k*k*ut)
!
!     CALCULATE THE PENMAN-MONTEITH EVAPORATION
!
    le=((deltaa*(rn-deltas)+86.4*rho*cp*vpdp/ra)/(deltaa+
   &      gamma))
           evap=le/lambda
!
!     CONVERT THE FLUXES TO W m-2
!
          rn=rn/0.0864
          le=le/0.0864
          deltas=deltas/0.0864
    return
    end
!**********************************************************************
    function delcalc(ta)
!**********************************************************************
!     FUNCTION TO CALCULATE THE SLOPE OF THE VAPOUR PRESSURE CURVE
!     INPUT
!        TA - AIR TEMPERATURE (deg. C)
!     OUTPUT
!        DELCALC - SLOPE OF THE VAPOUR PRESSURE CURVE (kPa deg. C-1)
!
    implicit none
    real delcalc,ta,ea
    ea=0.611*exp(17.27*ta/(ta+237.3))
    delcalc=4099*ea/(ta+237.3)**2
    return
    end
!**********************************************************************
    function alambdat(t)
!**********************************************************************
!
!     FUNCTION TO CORRECT THE LATENT HEAT OF VAPORISATION FOR TEMPERATURE
!
!     INPUT:
!        T = TEMPERATURE (deg. C)
!     OUTPUT:
!        ALAMBDAT = LATENT HEAT OF VAPORISATION (MJ kg-1)
!
    implicit none
    real alambdat,t
    alambdat=2.501-t*2.2361e-3
    return
    end
!**********************************************************************
    function psyconst(p,alambda)
!**********************************************************************
!     FUNCTION TO CALCULATE THE PSYCHROMETRIC CONSTANT FROM ATMOSPHERIC
!     PRESSURE AND LATENT HEAT OF VAPORISATION
!     SEE ALLEN ET AL (1994) ICID BULL. 43(2) PP 35-92
!     INPUT:
!         P = ATMOSPHERIC PRESSURE (kPa)
!         ALAMBDA = LATENT HEAT OF VAPORISATION (MJ kg-1)
!     OUTPUT:
!         PSYCONST =  PSYCHROMETRIC CONSTANT (kPa deg. C-1)
!
    implicit none
```

```
real psyconst,p,alambda,cp,eta
cp=1.013
eta=0.622
psyconst=(cp*p)/(eta*alambda)*1.0e-3
return
end
```

ANNEXE D – EVAPORATION DATA SETS

Location of site(s)	Arley Brook Catchment
Location of data	Dept. of Geological Sciences University of Plymouth 37 Portland Square Plymouth
Local contact	Mr Ken Vines
Ownership	
Period of record	1974 to 1980
Frequency of observations	Daily
Completeness	Short gaps in climate data have been infilled but longer gaps remain. Gap in 1979 due to fuel crisis.
Data quality	Variable
Accessibility	All data that was originally on paper tape (mainly climate data) are on computer. Remainder is on paper
Cost	
Measurement details	Evaporation pan (probably Met. Office type) was at the principal climate station (Northwood Lane site).
Other issues	

Location of site(s)	Eskdalemuir
Location of data	The Met. Office Edinburgh Climate Office (0131 244 8366)
Local contact	
Ownership	The Met. Office
Period of record	1970-present for met. data 1970-86 evaporation pan data
Frequency of observations	Hourly met. data
Completeness	
Data quality	Data has been quality controlled
Accessibility	Evaporation data on paper record Met. data computerised.
Cost	
Measurement details	
Other issues	Some difficulties due to marked growth of the Eskdalemuir forest and a major change in the way the wind data was processed. IH has a copy of the met. data for the period 1970-1986 but this cannot be copied to anyone else.

Location of site(s)	Farmoor reservoir, Oxfordshire
Location of data	(Crowe, 1974) original data thought to be in the library of Binnie, Black and Veatch, Grosvenor House, 69 London Road, Redhill, Surrey, RH1 1LQ
Local contact	
Ownership	
Period of record	4 Apr – 30 Oct 1966
Frequency of observations	unknown
Completeness	unknown
Data quality	good
Accessibility	unknown
Cost	
Measurement details	Met. Office and US Class A pan evaporation Net radiometer Reservoir observations
Other issues	Meteorological data may also have been recorded. Paper records totals of pan evaporation, Penman evaporation and reservoir observations.

Location of site(s)	Ham Wall, Somerset
Location of data	Centre for Ecology and Hydrology Wallingford Oxon OX10 8BB
Local contact	Dr. Mike Acreman
Ownership	NERC
Period of record	Apr – Sept 1996 Apr – Jun 1997
Frequency of observations	hourly
Completeness	A few gaps
Data quality	good
Accessibility	All data are in computer format
Cost	Cost of reproduction
Measurement details	Bowen ratio measurements giving actual evaporation AWS deployed at same location
Other issues	The site is a reed bed that is being re-established so the early data may be representative of open water conditions. The work is described in Gilman *et al.* (1998)

Location of site(s)	Institute of Arable Crop Research (IACR) - Rothamsted
Location of data	IACR – Rothamsted Harpenden Herts. AL5 2JQ
Local contact	Colin Peters
Ownership	
Period of record	1960-present
Frequency of observations	
Completeness	
Data quality	
Accessibility	
Cost	
Measurement details	
Other issues	Information can be found on http://www.res.bbsrc.ac.uk/era/

Location of site(s)	Institute of Arable Crop Research (IACR) - Broom's Barn
Location of data	IACR - Broom's Barn Higham Bury St. Edmunds Suffolk IP28 6NP
Local contact	Alan Thornhill
Ownership	IACR
Period of record	1970-1996
Frequency of observations	daily
Completeness	
Data quality	good
Accessibility	Paper record 1970-1990 1990-1996 computer database
Cost	Cost of reproduction
Measurement details	Daily met. values suitable for calculating Penman evaporation Evaporation pan
Other issues	Information can be found on http://www.res.bbsrc.ac.uk/era/

Location of site(s)	Centre for Ecology and Hydrology (formerly Institute of Hydrology)
Location of data	Centre for Ecology and Hydrology Wallingford, Oxon OX10 8BB
Local contact	Mr M J Lees
Ownership	NERC
Period of record	Met. data 1962-present pan evaporation 1970 - present
Frequency of observations	Daily
Completeness	Some small gaps
Data quality	Good
Accessibility	All met. data on computer Evaporation pan data from 1995 on computer, paper record before that
Cost	Cost of reproduction
Measurement details	Standard manual met. data of sunshine hours, wind run, rainfall, average air temperature, max. min- temperature and dry bulb wet bulb temperatures Met. Office evaporation pan
Other issues	

Location of site(s)	Kempton Park
Location of data	(Lapworth, 1965) and Met. Office
Local contact	
Ownership	
Period of record	1956-62
Frequency of observations	Daily observations were made but the reported data is in monthly totals
Completeness	full
Data quality	good
Accessibility	Monthly values are reported in a journal paper. The original data has not been located.
Cost	None for the evaporation estimates but meteorological data would have to be purchased from the Met. Office
Measurement details	Mass balance of two reservoirs Evaporation pan data from US Class A and Met. Office pans Climate data is standard meteorological station held at the Met. Office
Other issues	

Location of site(s)	Lake Trawsfynydd
Location of data	(McMillan, 1973)
Local contact	
Ownership	
Period of record	1966-70
Frequency of observations	Hourly data recorded Weekly totals from 1 May 69 to 26 Dec 70 in report
Completeness	Problems with instruments left gaps in the early years but last two years are essentially complete
Data quality	Variable – radiation detectors used up to 1969 degraded
Accessibility	Paper records in report
Cost	none
Measurement details	Instruments were mounted on nine pontoons and a land based met. station. Measurements of wind speed, air temperature, vapour pressure at 0.6 and 3 m above water surface, plus net radiation and water surface temperature. On land same up to 10 m,
Other issues	Measurements are for a lake used for cooling a power station and so the atmosphere above the lake may well have been unstable. There was a strong advective component to the energy balance and the objective of the study was to estimate the combined heat loss (latent and sensible heat fluxes).

Location of site(s)	Otterbourne
Location of data	
Local contact	
Ownership	
Period of record	1920-68
Frequency of observations	monthly
Completeness	Complete until Dec 1967, except March 1963
Data quality	Some negative values
Accessibility	Held on floppy disk
Cost	
Measurement details	
Other issues	A copy of these data is held at IH

Location of site(s)	Reading University
Location of data	Dept. of Meteorology University of Reading Whiteknights, PO Box 27 Reading RG6 2AB
Local contact	
Ownership	University of Reading
Period of record	1971 onwards
Frequency of observations	Daily
Completeness	Full record
Data quality	Good
Accessibility	All on Access database
Cost	Negotiable
Measurement details	Standard manual met. data of sunshine hours, wind run, rainfall, average air temperature, max. min- temperature and dry bulb wet bulb temperatures Met. Office evaporation pan
Other issues	

Location of site(s)	Sutton Bonington
Location of data	The University of Nottingham Sutton Bonington Campus Loughborough, Leics, LE12 5RD
Local contact	Chris Deuchar
Ownership	The University of Nottingham
Period of record	1962-present
Frequency of observations	daily
Completeness	Full records
Data quality	Good
Accessibility	Evaporation pan data on paper records Met. data on computer format
Cost	Cost of reproduction
Measurement details	Rainfall 1916-present Sunshine hours 1929-present Wind run 1962-present Air temperature 1921-present Max-min temperature 1946-present Evaporation pan 1972-present (US class A and Met. Office)
Other issues	Evaporation pan originally installed by STWA Met. station is going automatic in 1999 so pan readings will probably stop.

Location of site(s)	South Wales – Penmaen, Bridgend, Llanelle, Dale Fort, Otielton, Trawscoed, Llynfan
Location of data	Environment Agency Wales, South West Area
Local contact	Jean Frost
Ownership	Environment Agency
Period of record	1967 to present for whole data set but only two are current. Most were operating in the period 1977 to 1983
Frequency of observations	
Completeness	
Data quality	
Accessibility	Paper records
Cost	
Measurement details	Class a Pan, Met. Office pan and USSR pan were compared at one site. The rest are Met. Office pans. Met. data is available at three of the sites.
Other issues	

Location of site(s)	Stocks Reservoir
Location of data	Department of Environmental Sciences University of Lancaster Lancaster LA1 4YQ
Local contact	Dr N Chappell
Ownership	
Period of record	1955 to 1963
Frequency of observations	daily
Completeness	
Data quality	
Accessibility	
Cost	
Measurement details	Met. Office evaporation tank Standard meteorological climatological station
Other issues	

ANNEXE E – BIBLIOGRAPHY

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ANNEXE F – SIMULATION OF MORECS GRASS PE

For the purposes of this project, it has been necessary to simulate MORRECS grass PE. This has been achieved with a FORTRAN subroutine that then calls a standard Penman-Monteith model. The inputs to the subroutine are the daily values of driving variables that are utilised by all the other models (including the equilibrium temperature model) and the day number and month of the day for which the estimate is required.

```
!**********************************************************************
      subroutine morecspe(statx,staty,jul,mth,vpd,wind,ta,srad,
           & lrad,fc,evap)
!**********************************************************************
!     SUBROUTINE TO SIMULATE MORECS GRASS PE AT A DAILY TIME STEP
!     INPUTS:
!        STATX – GRID REFERENCE EASTING OF THE SITE (m)
!        STATY – GRID REFERENCE NORTHING OF THE SITE (m)
!        JUL – DAY NUMBER
!        MTH – MONTH NUMBER
!        VPD – VAPOUR PRESSURE DEFICIT (mb)
!        WIND – WIND SPEED (m s-1)
!        TA – AVERAGE AIR TEMPERATURE (deg. C)
!       SRAD – INCOMING SOLAR RADIATION (W m-2)
!       LRAD – INCOMING LONG-WAVE RADIATION (W m-2)
!       FC – CLOUDINESS FACTOR
!     OUTPUT:
!        EVAP –EVAPORATION (mm/day)
!
      implicit none
      integer*4 jul,mth
      real*8 r8east,r8lat,r8long,r8north
      real*4 albedo,del,delta,evap,fc,g,gd,lrad,phi,pi,rn,srad,
     & t1,t2,ta,vpd,wind,statx,staty
      real*4 p(12),rs(12)
      data p/-137.,-75.,30.,167.,236.,252.,213.,69.,-85.,-206.,-256.,
    &-206./
      data rs/80.,80.,60.,50.,40.,60.,60.,70.,70.,70.,80.,80./
      real*4 delcalc
      pi=4.0*atan(1.0)
      r8east=statx
      r8north=staty
      albedo=0.25
!
!     GET THE LATITUDE OF THE SITE
!
      call bngtgeo(r8east,r8north,r8lat,r8long)
      phi=r8lat*pi/180.
!
!     CALCULATE TIME OF SUNRISE AND SUNSET
!
      del=0.41*cos(2.*pi*(float(jul)-172.)/365.)
      t1=(24.-acos(-tan(phi)*tan(del))*24.0/pi)/2.
      t2=24.-t1
!
!     ESTIMATE SOIL HEAT FLUX IN DAYLIGHT HOURS
!
      gd=0.2*((lrad-(fc*0.95*5.67e-8*(ta+273.13)**4))*(t2-t1)/24.+
     &        (1-albedo)*srad)
!
```

```
!     FOR NIGHT USE MEAN HEAT FLUX MODERATED BY LENGTH OF NIGHT
!
      g=gd+(gd-p(mth))/(2.*t1)
!
!     GET THE SLOPE OF THE TEMPERATURE/VAPOUR PRESSURE CURVE
!
      delta=delcalc(ta)*10.
!
!     CALCULATE THE NET RADIATION
!
      rn=(1-albedo)*srad+lrad-(fc*0.95*5.65E-8*(ta+273.13)**4)
!
!     INPUT TO THE STANDARD DAILY PENMAN-MONTEITH EQUATION
!
      call penmon(rs(mth),0.15,delta,vpd,wind,ta,rn,g,evap)
      return
      end
C************************************************************************
      subroutine penmon(rs,hc,delta,vpd,u,ta,rn,g,evap)
C************************************************************************
C     SUBROUTINE TO CALCULATE THE PENMAN_MONTEITH EVAPORATION
C     INPUT:
C         RS - BULK CANOPY RESISTANCE (s m-1)
C         HC - HEIGHT OF CANOPY (m)
C         DELTA - SLOPE OF THE SATURATION VAPOUR PRESSURE CURVE (mb deg.
C        C-1)
C         VPD - VAPOUR PRESSURE DEFICIT (mb)
C         U - WIND SPEED (m s-1)
C         TA - AIR TEMPERATURE (deg C) SET TO -999.99 IF UNKNOWN
C         RN - NET RADIATION (W m-2)
C         G - SOIL HEAT FLUX (W m-2)
C     OUTPUT:
C         EVAP - EVAPORATION (mm)
C     CONSTANTS
C         LAMB - LATENT HEAT OF VAPORISATION (MJ kg-1)
C         GAMMA - PSCHROMETRIC CONSTANT (kPa C-1)
C         CP - SPECIFIC HEAT OF MOIST AIR (KJ kg-1 C-1)
C         RHO - DENSITY OF AIR (kg m-3)
C         K - VON KARMAN'S CONSTANT
C         ZR - HEIGHT OF METEOROLOGICAL MEASUREMENTS (m) ASSUMED 1.2
C
      implicit none
      real*4 cp,d,delta,deltap,evap,g,gamma,gj,h,hc,k,lamb,ra,
     & rho,rn,rnj,rs,ta,tax,u,ut,vpd,vpdp,zoh,zom,zr,b,c,sigma
      real*4 alambdat,psyconst
C
C     SETUP CONSTANTS
C
      h=hc
      if (h.le.0.01) h=0.01
      tax=ta
      if (ta.eq.-999.99) tax=20.0
      lamb=alambdat(tax)
      gamma=psyconst(100.0,lamb)
      rho=1.0
      cp=1.013
      k=0.41
      zr=1.2
      evap=0.0
      ut=u
```

```
      sigma=4.9e-9
     if (ut.le.0.01) ut=0.01
C
C     CONVERT FROM W m-2 TO MJ m-2 d-1
C
      rnj=rn*0.0864
      gj=g*0.0864
C
C     CONVERT FROM MB TO kPa
C
      vpdp=vpd*0.1
      deltap=delta*0.1
C
C     CALCULATE THE AERODYNAMIC RESISTANCE RA (s m-1)
C
       if (h.le.eq.0.0) then
       ra=(alog(zr/zom)*alog(zr/zoh))/(k*k*ut)
            else
C
C     CALCULATE THE ZERO PLANE DISPLACEMENT, D,
C
        d=0.67*h
C
C     ZOM - ROUGHNESS PARAMETER FOR MOMENTUM
C     ZOH - ROUGHNESS PARAMETER FOR HEAT AND WATER
C
        zom=0.123*h
        zoh=0.1*zom
C
C     CALCULATE THE AERODYNAMIC RESISTANCE RA (s m-1)
C
        ra=(alog((zr-d)/zom)*alog((zr-d)/zoh))/(k*k*ut)
            endif
C
C     CALCULATE THE CORRECTION FOR THE SURFACE NOT BEING AT AIR
C    TEMPERATURE
C     THIS IS GIVEN IN Hough and Jones, 1997, Hydrology and Earth System Science,
C     1(2), 227-239 BUT DERIVED BY Monteith, 1981, Quart. J. Roy. Met. Soc., 107, 1-27
C
      b=4.0*0.95*sigma*(273.13+tax)**3
      c=1.+b*ra/(86.4*rho*cp)
C
C     CALCULATE THE PENMAN-MONTEITH EVAPORATION
C
      evap=((deltap*(rnj-gj)+86.4*rho*cp*vpdp*c/ra)/(deltap+
     &    gamma*c*(1.0+rs/ra))) /lamb
      return
      end
```

ANNEXE G – ESTIMATION OF LONG-WAVE RADIATION FOR THE PENMAN EQUATION

As part of this project, the PETCALC method of calculating PE was simulated in a FORTRAN program. Initially, a significant discrepancy was found between values calculated by PETCALC and those from the simulation. This discrepancy was found to occur because different methods were being used to calculate the net long-wave radiation.

The net long-wave (thermal) radiation is calculated as the difference between the outgoing, $K_\uparrow$, and incoming, $K_\downarrow$, long-wave radiations. PETCALC uses the method given by Penman (1948) which is based on the method given by Brunt (1939) to calculate values for these from the air temperature and the number of hours of sunshine in the day. Thus, the clear sky outgoing long-wave radiation is calculated as:

$$K_\uparrow = \sigma(T_a + 273.1)^4 \tag{42}$$

where σ is the Stefan-Boltzman constant and T_a is the air temperature (^{o}C). It should be noted that this makes two assumptions. Firstly that the emissivity of the land surface is unity and secondly that the temperature of the land surface is equal to the air temperature. Also, the units of $K_\downarrow$ are mm/day in order to simplify the subsequent calculation of the evaporation rate. Thus, both $K_\downarrow$ and σ have been divided by the latent heat of vaporisation, λ. The clear sky incoming long-wave radiation is then calculated as:

$$K_\downarrow = \sigma(T_a + 273.1)^4 (0.44 + 0.092\sqrt{e}) \tag{43}$$

where e is the vapour pressure (mm Hg) of the air at screen height. The two factors, 0.44 and 0.092, are the averages of values measured at six sites in Europe. The net long-wave radiation, K_n, (mm/day) is calculated by combining the two equations above and introducing a 'cloudiness factor':

$$K_n = \sigma(T_a + 273.1)^4 (0.092\sqrt{e} - 0.56)(1 - 0.09m) \tag{44}$$

The 'cloudiness factor' is the last term in this equation and allows for the amount of cloud present during the day. The variable *m* is the ratio of the number of sunshine hours during the day to the total number of daylight hours. It should be noted that the convention adopted is that the radiation incoming is considered positive and that outgoing is negative.

The method for calculating the net long-wave radiation adopted currently, e.g. Hough (1997), Allen *et al.* (1998), differs from this. The clear sky outgoing long-wave radiation is calculated as:

$$K_\uparrow = \varepsilon\sigma(T_a + 273.1)^4 \tag{45}$$

where ε is the emissivity of the land surface, commonly taken as 0.95. It should be noted that modern practice is to specify the long-wave radiation in units of energy, generally W m^{-2} or MJ m^{-2} d^{-1}. The clear sky incoming long-wave radiation is based on Brutsaert (1975) and is calculated as:

$$K_{\downarrow} = 1.28\sigma(T_a + 273.1)^4 \left(\frac{e}{T_a + 273.1} \right)^{1/7} \tag{46}$$

where the factor of 1.28 has been determined by the Met. Office as appropriate for UK conditions. The vapour pressure, *e*, is in units of millibars. The net long-wave radiation is then calculated as:

$$K_n = \varepsilon\sigma(T_a + 273.1)^4 (1.28 \left(\frac{e}{T_a} \right)^{1/7} - 1)(0.2 - 0.8m) \tag{47}$$

The method used by Penman and that commonly used today, for calculating the net long-wave radiation, produce values that are significantly different. The method used today gives values that are lower by between 5 and 14%. This has implications for the estimation of evaporation.

It would seem reasonable that the evaporation model should be independent of the methods used to produce the driving variables required by the model. Indeed, this is the current situation since modern instruments allow us to measure the driving variables to a reasonably high accuracy and the modern models are physically based and so do not require calibration. However, Penman was not in this position when he developed his model. Thus, he had to estimate the incoming and outgoing long-wave radiations from measurements of sunshine hours, using the best available methods of the time. He then calibrated the constants in the wind function of his model, using these estimates of radiation, against observations of the evaporation from water in brick-lined pits. Thus there is an implicit assumption that the estimates of radiations are compatible with those used by Penman. As discussed above, there are differences between the methods for estimating long-wave radiation used now and those used by Penman (1948). Therefore, when using the Penman (1948) model, the net long-wave radiation should be calculated in the way described by Penman (1948) in order to conform to the assumptions made when the wind function was calibrated.