A climate change report card for water

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5. A hydrological perspective on UK evaporation: historical trends and future projections

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Summary

- Evaporation is an important component of the hydrological cycle. This document presents a review of historical trends and future projections for evaporation in the UK, from a hydrological perspective.

- Potential evaporation (PE) is generally considered to represent the amount of water that would be lost to the atmosphere if there were no limits to supply. Actual evaporation (AE) can be estimated as a fraction of PE dependent on soil wetness.

- There are many formulae for estimating PE from meteorological data, of varying complexity, from simple empirical formulae to data-intensive physically-based formulae (like Penman-Monteith).

- PE is usually required, along with rainfall, as an input for hydrological modelling. MORECS PE, based on Penman-Monteith, is often used in the UK.

- PE accuracy is generally considered less important than rainfall accuracy for hydrological model performance, as PE is less spatially and temporally variable than rainfall and has a well-defined seasonal pattern.

- There is relatively little information on historical trends in AE or PE in the UK, or indeed globally. The few UK studies generally indicate increases in PE, and one recent study has shown an increase in national average AE.

- There is little consensus on the best formulae to derive future PE projections from climate model data. The choice is further complicated when considering possible changes in plant behaviour under higher CO_2 concentrations.

- Most studies presenting future PE projections for the UK indicate increases in annual PE totals, although some studies suggest small decreases in PE for one or two months of the year.

- The approach used to estimate PE could be particularly important in regions where precipitation and PE are in close balance with each other, but PE uncertainty could be less important than climate model uncertainty for hydrological impacts.

- Further investigation is needed into which PE formulae are likely to be more (or less) reliable when applied with climate model data, and into the feedbacks between climate change and plant transpiration, plant growth and land-cover.

Introduction

Evaporation is an important component of the hydrological cycle, as it transfers water from the land surface to the atmosphere. The term evaporation, or its more expansive name "evapotranspiration" can encompass the transfer via both evaporation and transpiration; the former is the loss of water lying on a surface (e.g. lake, soil, plant leaf), while the latter is the loss of water via plants, through the stomata in their leaves. While there is considerable and ongoing debate in the scientific literature as to the preferred terminology (summarised by Lhomme 1997 and Cain et al. 1998), the quantities potential evaporation (PE) or potential evapotranspiration (PET) are generally considered to represent the amount of water that would be lost to the atmosphere if there were no limits to supply (that is, when soil moisture is at or near field capacity; Federer et al. 1996). Actual evapotranspiration (AE or AET) can therefore be estimated as a fraction of PE dependent on soil wetness, as it can be less than PE if the soil is dry, but cannot usually be greater than PE. Evapotranspiration is difficult to measure directly, although several techniques attempt to measure AE. Some of the complexities of measuring and modelling evaporation are described by Shuttleworth (2007).

Four meteorological variables influence PE; radiation (or sunshine), temperature, humidity (or vapour pressure) and wind speed. Further variables influence the transpiration component; plant height, rooting depth, leaf area and vegetation roughness. Thus PE varies for different plant types, but cannot be accurately estimated for mixed plant communities as the relevant parameters are difficult to measure. To simplify matters, the concept of the 'reference crop' was introduced, with average crop parameters provided for the estimation of reference PE. The reference crop is often taken as short grass (Pereira et al. 1999).

Many formulae have been developed for estimating PE, some for particular reference crops and some where crop parameters can be specified. The simplest, empirical formulae involve a single meteorological variable (e.g. the temperature-based formulations of Thornthwaite (1948) and Oudin et al. (2005a)), while the most complex and physically-based PE formulae involve all four meteorological variables (e.g. Penman-Monteith; Monteith 1965), with a range in between (e.g. the Blaney-Criddle method involving temperature and sunshine (Blaney and Criddle 1950) or the Priestley-Taylor method involving temperature and radiation (Priestley and Taylor 1972)). Oudin et al. (2005a) provide a useful summary of 17 variations found in the literature. Penman-Monteith is the method recommended by the United Nations Food and Agricultural Organisation (FAO) for deriving grass reference PE (Pereira et al. 1999), and is used by the UK Climate Projections 09 (UKCP09) weather generator (Jones et al. 2009). The UK Met Office Rainfall and Evaporation Calculation System (MORECS; Thompson et al. 1981, Hough et al. 1996) is based on a modified version of the Penman-Monteith formulation of PE, as is the more recent Met Office Surface Exchange Scheme (MOSES; Cox et al. 1998).

PE estimates are of interest in themselves but, more importantly from the hydrological perspective, they are a required input for hydrological modelling of river flows, floods, droughts etc., alongside rainfall. Changes in PE, either historical or future, could contribute to changes in hydrological indices like mean monthly river flows, both on their own and in combination with changes in rainfall, so have to be taken into account. Although we focus here on the hydrological perspective, PE and AE are of interest in other areas (e.g. agriculture and ecology; Fisher et al. 2011), and many of the same issues will apply.

This paper begins by providing some background on estimates of AE and PE in the UK, the influence of PE in hydrological modelling, and the ways in which environmental change can affect PE (Section 2). A review of historical trends in evapotranspiration in the UK is then presented, in the context of global changes (Section 3). Future projections of PE are then reviewed, along with the possible consequences of uncertainty in PE changes for modelling the hydrological impacts of climate change (Section 4). An assessment of the confidence in the direction of historical trends and future projections for evaporation in the UK is then presented (Section 5). Finally, Section 6 presents a discussion and conclusions.

Background

Measuring and modelling AE

There are several techniques that measure AE. Comprehensive reviews of evaporation measurement methods are provided by Verhoef and Campbell (2005), Shutov et al. (2006), Burt et al. (2005) and Shuttleworth (2007), covering Bowen ratio techniques, Eddy Covariance (EC), Lysimeters and the new remote method of Large Aperture Scintillometers (LAS). A less direct method to estimate AE is by calculating the difference between catchment rainfall and runoff, however this estimate of ‘losses’ includes AE together with other catchment losses like abstractions, as well as possible catchment imports like groundwater inflows. Where imports and exports of water from the catchment are minimal and the surface and groundwater catchment areas are sufficiently similar, the average catchment losses provide a reasonable assessment of the mean AE across the catchment. The UK National River Flow Archive provides estimates of Mean Annual Loss for most of its catchments, calculated as mean annual rainfall minus mean annual runoff. More complex catchment water balance methods can include changes in water storage in the catchment (Senay et al. 2011).

Recently there have been attempts to measure evaporation using Earth Observation (EO) products (e.g. Miralles et al. 2011). This method uses a simple sub-model fed by a variety of satellite data to calculate interception, soil surface evaporation and transpiration. The Priestley-Taylor formula is used to calculate PE and then uses a soil-moisture accounting model to limit the soil-surface evaporation, with factors based on the aridity and the land-type. It calculates the interception loss independently using the Gash (1979) model of interception.

There are essentially two types of models that aspire to model AE; hydrological models and land-surface models. Hydrological models need to estimate the proportion of rainfall lost from the land-surface via evaporation in order to estimate runoff and river flow. Land-surface models additionally require the estimate of evaporation in order to provide an essential boundary condition for a meteorological model. Both types of model run on a sub-daily (typically 1 hour or less) time-step, although some hydrological models run at longer (e.g. daily) time-steps. Typically, hydrological models determine PE by using meteorological data as input to one of several available estimation schemes, and AE is a fraction of PE dependent upon soil moisture, modelled continuously using a water-accounting scheme. A leading example of this is the PDM (Moore 2007). For a land-surface model, there is a need to obtain a full energy balance (radiation and heat) and carbon balance at the same time as the water balance, which means there can be less emphasis on modelling particular processes (such as PE) in isolation. Such models tend to be more complex as a result, and explicitly involve vegetation processes such as photosynthesis responses to atmospheric carbon dioxide. The wide range of physical processes included in the scheme requires estimation of a considerable number of model parameters, many of whose values can only be approximated at certain scales of application. A leading example of a land-surface model is JULES: the Joint UK Land Environment Simulator (Best et al. 2011), which has been benchmarked against EC data from around the world (Blyth et al. 2010).

A comparison of the two types of model and their performance at the global scale was made in WaterMIP (Water Model Inter-comparison Project; Haddeland et al. 2011). No such comparison has been made for the UK, however both types of model are used in the UK and a comparison would be possible.

Penman-Monteith PE, MORECS and MOSES

The Penman-Monteith formulation for short grass PE (Monteith 1965) is given by:

$$PE_{PM} = \frac{1}{\lambda \rho_w} \frac{\Delta(R_n - G) + \rho_a c_p (e_a - e_d)/r_a}{\Delta + \gamma(1 + r_c / r_a)}$$

where

λ is the latent heat flux (J/kg),

ρ_w is the density of water (kg/m^3),

ρ_a is the density of air (kg/m^3),

c_p is the specific heat of air (J/kg/^{o}C)

γ is the psychrometric constant (kPa/^{o}C),

R_n is the net solar radiation (J/m^2/s),

G is the soil heat flux (J/m^2/s),

$e_a=e(T_a)=0.611 \exp(17.27\ T_a / (T_a+237.3))$ is the saturation vapour pressure (kPa) with T_a air temperature (^oC),

$\Delta=de_a/dT_a=17.27 x 237.3\ e_a / (T_a+237.3)^2$ is the slope of the vapour pressure curve (kPa/^oC),

$e_d=e(T_d)$ is the actual vapour pressure (kPa) with T_d dew-point temperature (^oC),

$r_a=208/W_2$ is the aerodynamic resistance (s/m) with W_2 wind speed (m/s) at a height of 2 m, and

r_c is the canopy surface resistance (of short grass),

giving PE_{PM} in units of m/s. The FAO recommend a grass crop height of 0.12 m and surface resistance of 70 s/m (Allen et al. 1994). Penman PE (Penman 1948) has $r_c=0$ and $G=0$, and essentially represents loss from open water (or a freely evaporating water surface). The Penman-Monteith equation was developed for use with weather data at a daily time-step, but tests have shown that using monthly mean weather data gives monthly PE_{PM} very similar to the average of daily PE_{PM} computed from daily weather data (Allen et al. 1998; Oudin et al. 2010).

Penman-Monteith forms part of the MORECS system (Thomson et al. 1981, Hough et al. 1996), which converts daily synoptic station data into estimates of weekly and monthly PE, AE and soil moisture deficit (SMD) for a short grass land cover (and a selection of other land covers, such as deciduous trees) on a 40x40 km grid over Great Britain, for a range of soils defined by their available water capacity (AWC; Hough and Jones 1997). It implements a slightly modified version of Penman-Monteith to that recommended by the FAO. In particular, MORECS includes a correction for the assumption that surface temperature equals the measured (air) temperature, and employs monthly-varying leaf area index and canopy surface resistance values for short grass (height 0.15 m), resulting in r_c ranging from 44.5 s/m (late spring) to 88.7 s/m (winter) (with an average of about 73 s/m).

The more recent scheme, MOSES, was originally developed to calculate surface-to-atmospheric fluxes as part of a General Circulation Model. MOSES calculates water and energy fluxes between the land-surface and atmosphere alongside estimates of CO_2 fluxes and their effects on vegetation physiology (further detail is provided by Cox et al. 1999). For UK applications, MOSES can now run operationally as part of the UK Post Processing system at a 2km resolution and hourly time-step, and can provide a wide variety of outputs at that resolution including soil moisture, soil temperature, AE and PE (Smith et al. 2006). These quantities are typically produced for a range of vegetation-types (crops, shrubs, C3 and C4 grass, broadleaf and needle-leaf trees) and four non-vegetated surface types (urban, inland open water, bare soil and ice), with maps of UK land-cover used to determine the proportion of each surface type within each MOSES grid-box. Like MORECS, MOSES calculates evaporation according to the Penman-Monteith equation but the surface resistance is variable and assumed to be dependent on photosynthesis, which in turn depends on air temperature, incident radiation, humidity deficit and vegetation-type. MOSES also encompasses a wider range of land-atmospheric fluxes including canopy evaporation and sublimation from a snow surface. MOSES now forms the basis of the Joint UK Land Environment Simulator (JULES; Best et al. 2011) which is a research tool for the scientific community, for use in carbon cycle, climate change and impacts studies at both local and global scales.

PE estimates arising from MOSES, and its successor JULES, can be expected to change in line with contemporary understanding of earth-system processes, while MORECS PE estimates reflect a standardised, but arguably older, understanding of land-atmospheric physics.

PE and hydrological modelling

PE estimates (or the variables used to make them) are required inputs for most hydrological models. This includes models that provide continuous simulation of river flow at a catchment outlet, like PDM (Probability Distributed Model; Moore 2007), CATCHMOD (Wilby et al. 1994) or CLASSIC (Climate and LAnd-use Scenario Simulation In Catchments; Crooks and Naden 2007), and fully-distributed area-wide models, like Grid-to-Grid (G2G; Bell et al. 2009) or WASIM (WAter flow and balance SIMulation; Kleinn et al. 2005). It also applies to models such as the Palmer Drought Severity Index (e.g. van der Schrier et al. 2011).

PE is a much more conservative quantity than rainfall (it has much less spatial and temporal variability) and is highly seasonally predictable (Calder et al. 1983), so relatively simple data are considered sufficient to obtain a closed water balance in hydrological modelling. Thus accuracy of rainfall inputs is generally considered more important than accuracy of PE inputs for hydrological model performance, as has been demonstrated in a number of sensitivity studies (e.g. Nandakumar and Mein 1997; Boughton 2006; Paturel et al. 1995; Manning et al. 2009). However, Parmele (1972) shows varying sensitivity to PE biases of +10% and -20% in nine US catchments, indicating that the relative importance of PE accuracy is likely to vary according to location, as well as how model performance is assessed (whether it focuses more on high or low flows for example). Thus care should be taken not to over-generalise from relatively limited studies. It should also be noted that some of the perceived insensitivity of hydrological models to PE accuracy could, in some cases, be due to model recalibration, which can allow model parameters to compensate for differing PE amounts (Andréassian et al. 2004).

Oudin et al. (2005a) tested 27 PE formulations (including 10 based on linear relationships with one or more weather variables) with four rainfall-runoff models for 308 catchments in France, Australia and the US. They found that PE based on temperature or radiation often provided more accurate streamflow simulations than more complex formulae. However, it should be noted that PE from all formulae was scaled by mean annual Penman PE, so only the effects of fluctuations within the period were tested, and that the models were recalibrated for each of the alternative inputs. Vorosmarty et al. (1998) compare 11 PE formulae over the US and state that ‘there is negative feedback, in that the drier the climate, and the larger the PE, the less important the PE estimate becomes in determining AE and thus runoff’. Manning et al. (2009) suggest that, for their modelling of water resources in the Thames, ‘possible underestimation of PE in very hot summers is offset in the subsequent hydrological modelling because AE is limited by moisture supply rather than determined by PE'.

The lower year-to-year variation in PE than rainfall even means that climatological estimates of PE (i.e. a seasonally varying pattern duplicated for each year) can be used, rather than PE time-series. Calder et al. (1983) showed that the use of sophisticated PE equations led to no clear improvement in SMD estimation compared to climatological PE, for six grassland sites in the UK. Further, Fowler (2002) showed that using climatological monthly mean PE distributed equally over each day of the month led to no significant degradation in SMD estimation when compared to either more complex climatological mean methods or PE time-series, for a site in Auckland, New Zealand. This was the case even for years that were much wetter or dryer than average. Similarly, Oudin et al. (2005a,b) found no significant improvements in river flow simulation when using PE time-series, rather than climatological mean PE, as input to four rainfall-runoff models for 308 catchments in France, Australia and the US. However, Oudin et al. (2005a) noted that arid and/or small catchments (PE ≥ rainfall; area < 150km^2) generally gained more benefit from use of PE time-series than wet and/or large catchments (PE < rainfall; area > 150km^2).

PE and hydrology in the UK

Monthly MORECS short grass PE is often used to provide hydrological model inputs in the UK. Figure 1 shows MORECS annual mean rainfall, PE and AE (for short grass and median AWC soils), and the ratios PE/rainfall and AE/PE, for the period 1961-1990. There is a south/east to north/west rainfall gradient, with an approximately reverse gradient for PE, which is around 20% higher in the south/east. However, the latter does not carry straight through to AE, which decreases again in the south and east of England. The ratio PE/rainfall demonstrates the dryness of eastern England (PE/rainfall ~> 1) and the wetness of the north/west of Britain (PE/rainfall << 1). The ratio AE/PE demonstrates that AE in the north/west of Britain is generally energy-limited (AE ~ PE) whereas AE in the east of England is generally water-limited (AE << PE). The dry/wet 'boundary' (where PE ~ rainfall) and the water-limited/energy-limited 'boundary' (where AE approaches PE) are related to each other, creating a transition region where the supply of water to the land surface (as rainfall) and the supply of energy (to evaporate water from the land surface) are in closer balance with each other (hereby termed water-energy balanced), as discussed below.

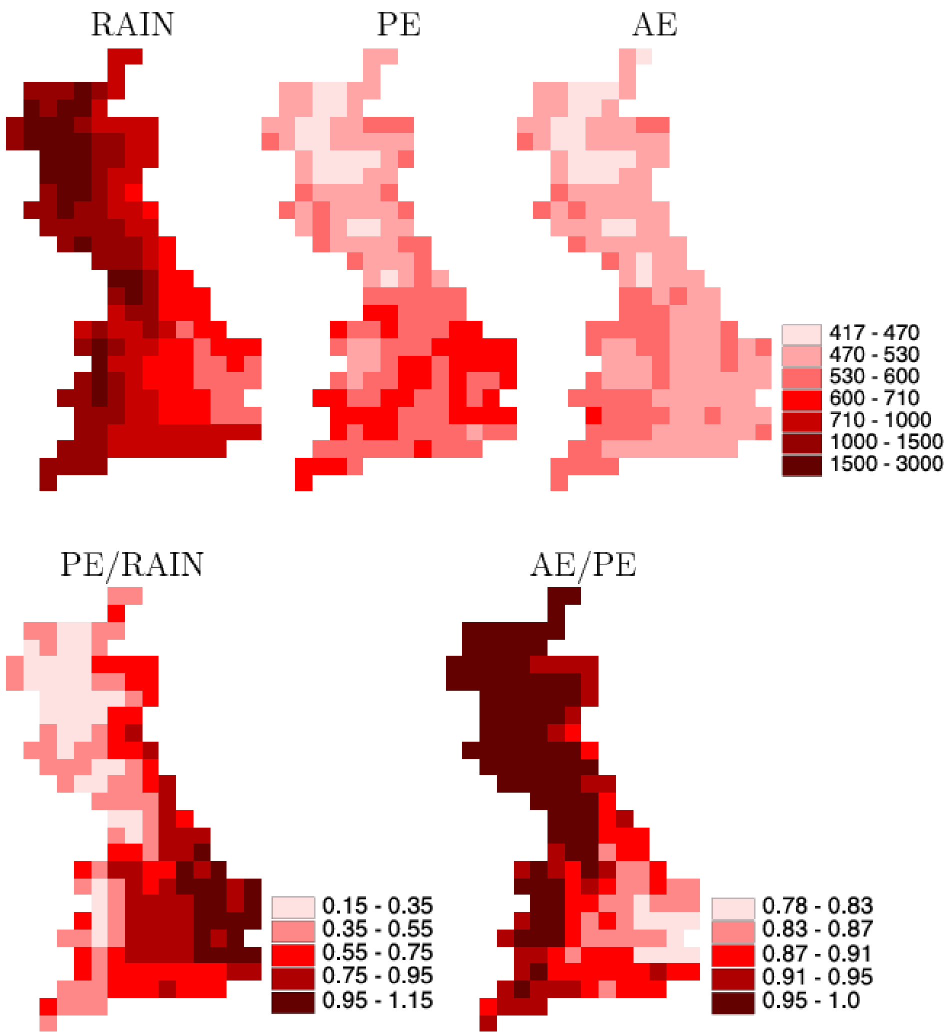

Figure 1 Maps of MORECS annual mean Rainfall, PE and AE (mm) for 1961-1990, and the ratios PE/rainfall and AE/PE.

It should be noted that the plots in Figure 1, using annual mean 40x40 km MORECS data (for short grass and median AWC soils), are only indicative; the patterns will vary to some extent from year to year and month to month. For instance, in a hot, dry summer, AE in the north-west could also be water-limited. Generally, hydrological models account for this by using rainfall at a finer spatial and temporal resolution, and producing their own estimates of AE using their own soil moisture accounting schemes. Model parameters are often calibrated to the provide the best at-site performance (by comparing simulated and observed river flows for instance).

By way of example, Figure 2 shows 1961-1990 monthly mean rainfall, PE, AE and runoff for example catchments located in three contrasting regions: wet/energy-limited, energy-water balanced and dry/water-limited. A given change to PE for a wet/energy-limited catchment is likely to result in a similar change to AE, whereas the same change to PE for a dry/water-limited catchment is likely to result in little change to AE (assuming little/no change to seasonal rainfall amounts). Thus dry/water-limited catchments in the south-east of England are likely to be less sensitive to PE accuracy. But wet/energy-limited catchments in the north-west are also likely to be less sensitive to PE accuracy, since they get more rainfall and lose a much smaller proportion of rainfall to evaporation than catchments in the south/east.

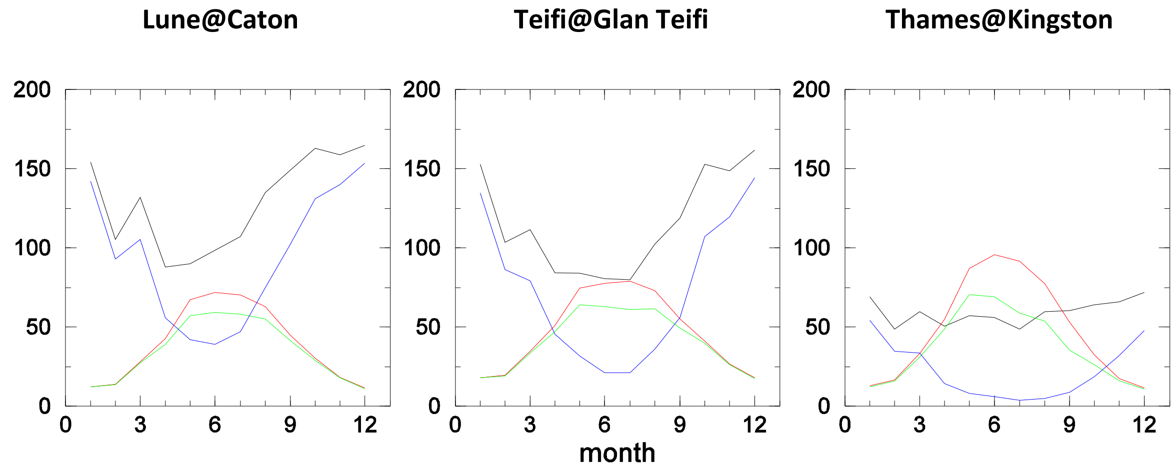

Figure 2 Monthly mean rainfall (black), PE (red), AE (green) and runoff (blue) (in mm) for three contrasting catchments when modelled with the CLASSIC rainfall-runoff model. The left-hand plot shows the energy-limited Lune@Caton, in north-west England; the right-hand plot shows the water-limited Thames@Kingston, in south-east England; the middle plot shows the more energy-water balanced Teifi@Glan Teifi, in south-west Wales.

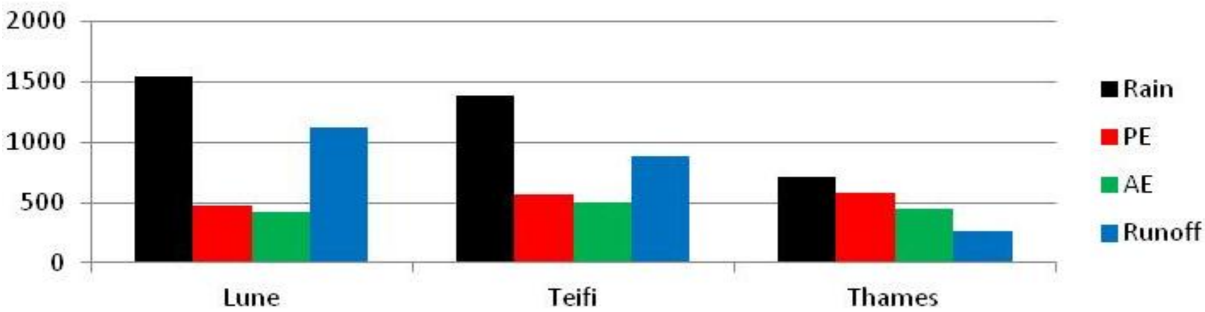

Figure 3 Mean annual totals of rainfall, PE, AE and runoff (in mm) for the three catchments in Figure 2.

Catchments most sensitive to PE accuracy are those in the more energy-water balanced transition region, close to the boundary between wet/energy-limited and dry/water-limited regions, since PE forms a larger proportion of their rainfall and so PE changes can more easily push them into a different regime. The histograms of mean annual rainfall, PE, AE and runoff in Figure 3 for the three example catchments show that the energy-water balanced catchment has the highest AE. The summer rainfall in this catchment is able to sustain a higher rate of AE than in the water-limited catchment. While there is available rainfall, a higher rate of PE results in higher AE, but where rainfall is more limited then AE will be reduced; the impact on runoff depends on the PE/rainfall balance. Thus such catchments are more sensitive to PE accuracy, particularly in terms of runoff through the summer and early autumn.

PE and environmental change

The situation becomes more complicated under climate change, as the amounts and seasonal distribution of rainfall are likely to change, as well as the atmospheric demand for water (PE). This means that dry/wet and energy-limited/water-limited boundaries are likely to shift, in a way which will be highly dependent on the specifics of the seasonal changes in rainfall.

Changes in PE under climate change can occur through changes in the meteorological controls (radiation, temperature, humidity and wind speed), but also through changes in the transpiration controls (surface resistance and leaf area). The relationship between climate change, meteorology and plant physiology is highly complex (Wullschleger et al. 2002). For example, higher concentrations of carbon dioxide in the atmosphere can lead to plant stomata opening less widely, resulting in a higher value of surface resistance and less transpiration (Betts et al. 2007). But higher carbon dioxide concentrations can also enhance plant growth, leading to a greater leaf area and more stomata, possibly counter-acting the effect of stomatal closure (Betts et al. 2007). Lower light levels, due to sunshine changes or to greater shading of the canopy, can also lead to stomatal closure (Wullschleger et al. 2002).

Frequently, climate change impact studies consider only the meteorological controls on PE, with values for the transpiration controls kept constant. Recently however, Gedney et al. (2006) showed that increasing trends in continental runoff through the twentieth century were most consistent with the reduction in transpiration due to CO_2-induced stomatal closure, suggesting the possible importance of this mechanism of change in PE for hydrological modelling under climate change.

A further complication is land-use change, either on its own or in combination with climate change. Different land-covers have different PE rates, so large-scale changes in land-cover can influence the total amount of AE, and so affect runoff and river flows. Some hydrological models allow for different land covers. For example, CLASSIC uses derived relationships between alternative land cover PE and short grass PE (Crooks and Naden 2007). Crooks and Davies (2001) investigated the sensitivity of the Thames catchment to land-use change, using CLASSIC. Their results showed that, while there was little difference in flood frequency for the land-use of 1960 compared to that of 1990, covering the catchment with 100% grass resulted in lower flood frequency than with the 1990 land-use, and covering the catchment with 100% trees resulted in even lower flood frequency. Similarly, Dunn and Mackay (1995) investigated the sensitivity of two sub-catchments of the river Tyne, in north-east England, to land-use change, using a distributed hydrological model. Their results showed that the same land-use change had a negligible effect on the hydrology of the highland catchment but much more effect on the lowland catchment, because evaporation is a greater proportion of total rainfall in the lowland catchment.

Ideally, climate change and land-use change would be considered together, as there are possible two-way feedbacks between them. At the local scale, eco-hydrological indicators such as those of Hill et al. (2000) can be used to assess which vegetation types are most likely to be found for particular combinations of soil type and hydrological conditions. These types of indices can be used to assess which species could appear or disappear following environmental change at a site, and feedbacks on evaporation and humidity could follow, particularly if the environmental changes affect large regions. For example, Teuling et al. (2010) show how forests and grasslands differentially affect overlying air-temperature under drought conditions.

Historical trends

PE and AE trends in the UK

There are very few published analyses of historical evapotranspiration trends in the UK, and most look at specific sites; none have national coverage.

The longest series is analysed by Burt and Shahgedanova (1998), who use daily observations from the Radcliffe Meteorological Station (Oxford) for 1815-1996. As only temperature and rainfall data were available, the authors used the Thornthwaite PE formulation to obtain an initial monthly PE time-series, then developed and applied seasonal regression equations to correct this towards what would have been obtained from the Penman formulation. They then estimated monthly AE, using rainfall and PE to estimate SMDs. The resulting plots of PE and AE time-series suggest an increase in PE and decrease in AE, but with considerable annual and decadal variation. No quantification of trends or their significance was provided.

Crane and Hudson (1997) used daily data from a site in the Welsh Uplands to calculate Penman PE for 1969-1995. They found no clear trend in PE, although there were trends in some of the components contributing to PE. The authors concluded that local factors (e.g. tree-felling) have had a greater impact on climate variables and PE at this site than have any larger-scale climatic changes.

A report for the Scottish Executive reviewing flood protection levels (Price and McKenna 2003) includes a plot of annual PE averaged over the 72 MORECS grid squares covering Scotland, for 1961-1993. This shows an increasing trend, but no quantification is given. Similarly, Yang et al. (2005) plot MORECS PE for 1961-1993, separately for each month, for a single MORECS grid square in Surrey. They indicate that increasing trends are apparent in almost all months, but provide no quantification.

Unfortunately, PE is not included in the UKCP09 trends report (Jenkins et al. 2007), or a Met Office report on spatial trends in UK climate (Perry 2006), based on 5x5km grids of numerous climate variables. This is possibly due to a shorter period of available wind data (from 1969) compared to other variables. It should be noted that any apparent trends in relatively short records need to be treated with caution, as they may be due to natural climatic variations rather than climate change (Robson 2002). This problem is likely to be exacerbated if there happen to be particularly wet or dry periods towards the start or end of the record under investigation.

There are no long term observations of AE in the UK. There are several sites where evaporation measurements are taken (for example, Auchenforth – a moorland site in Scotland; Easter Bush – a grassland site in Edinburgh; Grimsby Wood – a deciduous forest in the South of England; Wytham Wood – a deciduous forest in the South of England; Sheep Drove – a grass and crop site in the South of England; Tadham Moor – a watered grassland site in the South of England; Griffin Forest – a conifer site in Scotland). With the exception of Griffin Forest, these sites have been in operation for less than 10 years. This is far from long enough to establish a trend, although a review of these data sets is needed.

A long term analysis of AE (Blyth pers. comm.) modelled by JULES has shown a slow upward trend over 37 years (Figure 4). This result is similar to the trend given by a satellite Earth-Observation product (Miralles et al. 2011). Both the satellite product and the JULES model calculate an AE that is very close to PE. A comparison of the relationship between AE and PE as modelled and that from Eddy-Covariance observations suggests this is a reasonable model assumption. Thus, if PE increases, it is reasonable to expect that AE will follow (with the exception of very dry years such as 2003 and regions such as the south-east of England).

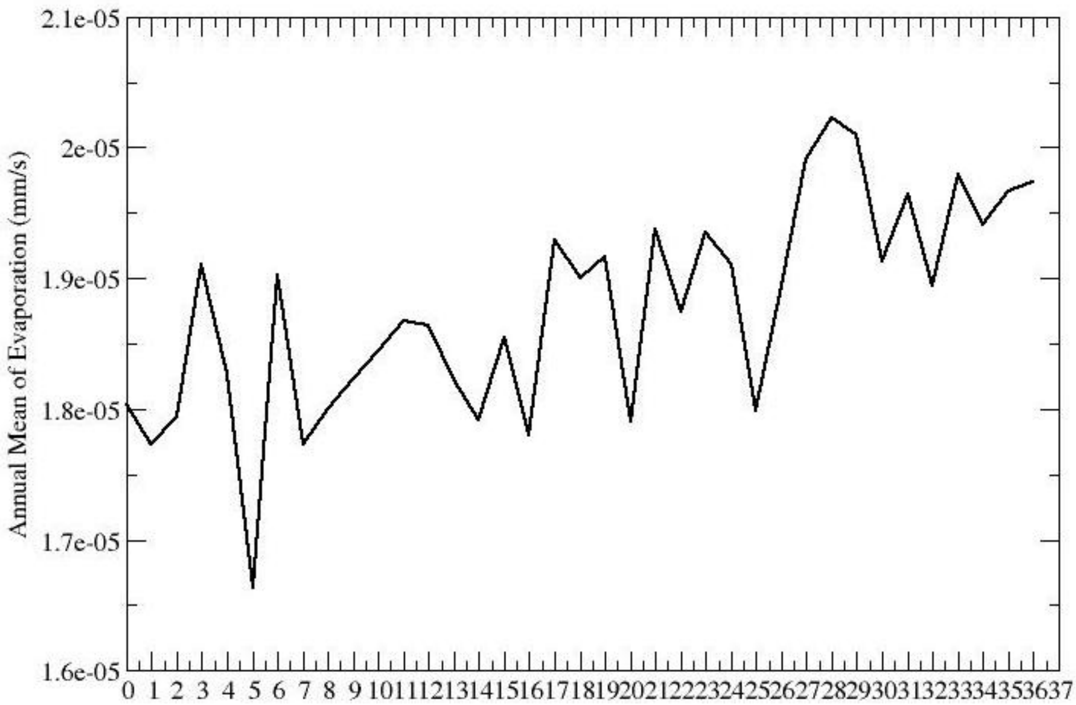

Figure 4 AE modelled by JULES for 1971-2007.

Global context

The shortage of information on evapotranspiration trends is not limited to the UK; in their global summary of observed and projected changes in water-related climate, the IPCC conclude that 'there is little literature on observed trends in evapotranspiration, whether actual or potential' (Bates et al. 2008).

There is evidence that pan evaporation (a proxy for PE) has been decreasing in a number of regions of the globe over the last 5 decades (Fu et al. 2009). Since this has been occurring despite increases in temperature, it has become known as the 'pan evaporation paradox'. The global review of trends in wind speed by McVicar et al. (2012) suggests that declining rates of evaporative demand are primarily due to declining wind speeds. This implies that it could be important to include wind in the calculation of PE, so that any wind speed trends under climate change can be transferred.

However, declining rates of PE are not universal. The global study of Weedon et al (2011) for 1958-2001 finds increasing trends in some basins (e.g. Niger), decreasing trends in some basins (e.g. Amazon and Murray-Darling), and no significant trends in some basins (e.g. Congo and Lena). Similarly, Thomas (2000) shows moderate increases in PE in north-east and south-west China but larger decreases are seen in the north-west and south-east, while, for the Canadian Prairies, trends in open water PE were typically found to be increasing in the north but decreasing in the south (Burn and Hesch 2007). Analyses in the latter case suggested that decreasing wind speed trends were the main influence on the decreasing PE trends, while vapour pressure deficit trends were the main influence on the increasing PE trends. Irmak et al. (2012) showed a significant decreasing trend in grass reference PE in the Platte river basin in Nebraska, US, but with a non-significant increase in wind speeds. They suggest, instead, that increasing rainfall and so decreasing short-wave radiation are responsible for the PE decrease.

Jung et al. (2010) examine data-driven estimates of global land AE, which show a (highly statistically significant) positive trend for 1982-1997 but a (less significant) negative trend for 1998-2008. Their correlation analyses suggest that the recent decline is due to increasing soil moisture limitation.

Future projections

Issues with PE estimation under climate change

Projections of PE from vegetated surfaces under climate change are generally not produced directly by Global or Regional Climate Models (GCMs or RCMs), and so have to be made offline using other climate model variables in PE formulae. Since there are so many formulae for estimating PE, the issue arises of which is likely to perform best when using climate model data (as opposed to observed weather data). Part of this issue is whether the climate model PE estimates will be used directly to drive the hydrological model, in which case the absolute values of the estimates in both current and future periods need to be reasonable, or whether the climate model PE estimates are simply used to estimate changes in PE between baseline and possible future climates, which are then applied to baseline observed PE data using the change factor method (Kay and Jones 2012). Note that a slightly different way of obtaining PE estimates under a future climate is to use a statistical downscaling model to fit a regression relationship between large-scale weather data and catchment-scale PE data, then to apply that relationship using large-scale GCM data to estimate future PE for the catchment (e.g. Wilby and Harris 2006). Similarly, Chun et al. (2012) develop a generalised linear model (GLM) for PE, to use as a substitute for the Penman-Monteith formula in the UK. These methods are likely to have many of the same issues as estimating PE using climate model variables in more standard PE formulae.

Using more complex formulae with climate model data is not necessarily straightforward, since some variables may be less reliable than others (Kingston et al. 2009, Vorosmarty et al. 1998), or simply not available from all climate models. For example, Arnell (2011) calculated changes to Penman-Monteith PE by applying changes to temperature, vapour pressure and net radiation but keeping wind speed fixed, as wind speed changes were not available under his future scenarios. The probabilistic projections in UKCP09 (Murphy et al. 2009) did not initially include wind speed, as wind projections were not available from all of the GCMs used in the statistical methodology (Sexton and Murphy 2010). Fisher et al. (2011) discuss issues with input data uncertainty and sensitivity in PE formulae. A further issue is that climate model data may have different inter-variable correlations to observed data (Chun et al. 2012).

As simpler, empirical PE formulae rely on fixed relationships between atmospheric variables and PE, these may not hold under extrapolation to very different future climates (Shaw and Riha 2011, Chun et al. 2012), although some level of extrapolation may be acceptable for formulae that show good performance under the current climate for a wide range of locations and climatic conditions (Sperna Weiland et al. 2011). Irmak et al. (2012) suggest that, as many climate variables affecting PE have been changing and are expected to change in the future, single-variable PE equations should be avoided when estimating PE (historical or future) trends 'due to the inherent nature of the trend passed to PE from the variable'. This view is echoed by Donohue et al. (2010), who state that 'the greater the number of the four key variables that are incorporated in a formulation, the more realistic the trends from that formulation become', although this clearly depends on both the relative sensitivity of a formula to each variable and on the relative strengths and directions of any trends in each variable (which are likely to have some level of inter-dependency; Fisher et al. 2011). It is currently not clear which meteorological variables are most important in the UK, although an analysis by Reynard and Arnell (1993) suggested that Penman-Monteith PE was particularly sensitive to increases in humidity compared to the other meteorological variables, because of the relatively high humidity levels in the UK.

The sensitivity of a PE formula cannot necessarily be predicted directly from the meteorological variables it includes; Bormann (2011) compared the PE changes given by 18 PE formulae, for six locations in Germany, and found as much variability between projections from formulae of the same type as between formulae of different types. However, Shaw and Riha (2011) illustrate the sensitivity of five PE formulae to increasing temperatures (i.e. purely temperature-sensitivity, neglecting changes in other variables), and show that the temperature-only Hamon and Thornthwaite PE formulae are much more temperature-sensitive than the more complex Penman-Monteith and Priestley-Taylor PE formulae. This result should be interpreted carefully though, since other variables (like radiation) are likely to vary with temperature (i.e. not independently; Fisher et al. 2011, Chun et al 2012), and inclusion of these could increase the response of the more complex formulae. Interestingly, Shaw and Riha (2011) show that the temperature-sensitivity of the Oudin PE formula is mid-way between that of the other two pairs of formulae. Oudin PE is usually described as temperature-based but also involves extraterrestrial radiation (dependent only on Julian day and latitude, not cloud cover etc.), so has some dependence on radiation even though its only true variable is temperature.

Shaw and Riha (2011) also show that radiation, rather than temperature, is the main driver of PE in broadleaf forests in the US. They suggest that it is the strong relationship between radiation and temperature for longer averaging-periods (weekly and above) which allows temperature-only formulae to work well in the current climate, but that this relationship will alter under climate change, leading to temperature-only formulae perhaps overestimating PE changes. They thus recommend the Priestley-Taylor formula, if Penman-Monteith is not used, but suggest that other formulae which include radiation could also work; even those that only include extraterrestrial radiation (e.g. the Oudin and Hargreaves formulae), as long as they are applied over longer averaging-periods rather than daily. Hargreaves PE (Hargreaves and Samani 1985) is an efficient empirical formula that includes the diurnal temperature range (as a proxy for humidity) as well as temperature and extraterrestrial radiation.

Sperna Weiland et al. (2011) compared Penman-Monteith, Priestly-Taylor, Hargreaves and Blaney-Criddle PE formulae for estimating global reference PE using daily reanalysis data. They found that Blaney-Criddle PE using reanalysis data provided the best match to monthly Penman-Monteith PE using Climate Research Unit (CRU) observation data. However, they concluded that the need for cell-specific calibration of the coefficients in the Blaney-Criddle formula, with large spatial variation in the calibrated coefficients, meant that it was unsuitable for climate change applications. They also concluded that, since Penman-Monteith did not outperform other methods, has a high data demand and is sensitive to input data accuracy, it too was not ideal for climate change applications. Their preferred formula was a globally-recalibrated version of Hargreaves PE, which performs well across a range of climate zones so is likely to do the same under climate change.

Prudhomme et al. (2012) applied 12 PE formulae (five temperature-based, including Oudin and Blaney-Criddle, five radiation-based, including Priestley-Taylor, and two versions of Penman-Monteith) for estimating reference PE in Great Britain, using MORECS' meteorological data. They compared the results to MORECS PE data for 1961-1990, in terms of monthly mean PE for one month in each of the four seasons (winter - January, spring - April, summer - July and autumn - October). They found that Penman-Monteith gave the best match (unsurprisingly since MORECS PE is based on Penman-Monteith), and that the temperature-based methods (apart from Blaney-Criddle) tended to give higher PE (particularly in April and July). Similar results were found when using HadRM3 RCM data for 1961-1990.

Kay and Davies (2008) compared Penman-Monteith PE and Oudin temperature-based PE over Britain when calculated using weather data directly from five GCMs and eight RCMs. The results showed, perhaps surprisingly (and contrary to Prudhomme et al. 2012), that Oudin PE from climate model temperature data matched MORECS PE better than did Penman-Monteith PE from climate model data, for the baseline period 1961-1990. Penman-Monteith PE was lower than MORECS PE for most months of the year and most climate models, leading to clear underestimation of mean annual PE. The authors postulated that this was due to lower reliability of some of the extra variables required for the calculation of Penman-Monteith PE (i.e. radiation, wind speed, and humidity or dew point temperature). Ekstrom et al. (2007) similarly found that Penman-Monteith PE derived from climate model variables for a baseline period was lower, in every month, than PE derived from weather observations. Prudhomme et al. (2012) suggested that RCM (short and long wave) radiation data should not be used directly but instead derived from other RCM variables.

Sometimes, bias-correction schemes are employed to allow for perceived biases in climate model data (Piani et al. 2010). However, it should be borne in mind that variables produced by climate models are expected to be internally consistent; applying bias correction separately to each variable required for PE estimation, or only to some variables, could introduce inconsistency and lead to incorrect PE estimates. A possible alternative is direct bias-correction of PE. This was attempted by Ekstrom et al. (2007), using simple monthly correction factors, but they still found the resulting baseline PE time-series to be unrealistic in terms of the range of daily values. Furthermore, bias correction assumes that any differences between observations and climate model estimates are due solely to bias in the climate model, rather than to multi-decadal natural variability for example, and that precisely the same bias applies for any future period. Thus any bias correction should be undertaken with extreme care.

PE projections for the UK

Arnell (2011) uses data from 21 climate models, for six catchments in Britain, and constructs a climate scenario representing a 2°C rise in global mean temperature. Increases in annual PE (calculated using Penman-Monteith but with fixed wind speeds) are shown for all but one climate model, with significant variation between climate models and catchments (ranging from approximately -4% to +40%). Furthermore, annual PE changes are shown to be broadly related to annual temperature changes, although changes to relative humidity and (to a lesser extent) radiation also have an effect.

Fowler et al. (2008) use data from 13 RCMs, for the 2080s time-horizon under the A2 emissions scenario, with a weather generator to develop probabilistic estimates of future changes in temperature, precipitation and Penman-Monteith PE for the river Eden, Cumbria. They show increases in mean PE in all seasons, with the largest percentage increases in autumn (approximately +30% to +80%) and winter (+30% to +60%) and smallest percentage increases in spring (0% to +50%) and summer (+20% to +40%).

Kay and Davies (2008) use data from five GCMs and eight RCMs, again for the 2080s time-horizon under the A2 emissions scenario, to calculate changes in Penman-Monteith PE and Oudin temperature-based PE for Britain. They show that percentage changes in Oudin PE are positive throughout the year, and larger in winter than summer, whereas changes in Penman-Monteith PE can be negative for some months and some climate models, and show greater monthly variability. The annual PE increases are generally lower for Penman-Monteith PE than for Oudin PE in the north, but generally higher in the south. Percentage changes in annual Oudin PE range from about +12% to +34%, whereas those in Penman-Monteith PE range from about +6% to +56%.

Ekstrom et al. (2007) use data from the HadRM3H RCM over Europe, again for the 2080s time-horizon under the A2 emissions scenario, and show seasonal absolute differences between future and baseline Penman-Monteith PE. Differences are largest in the summer (4-8 mm/day over large parts of Europe; less over Britain) and smallest in the winter (less than 0.5 mm/day for most of Europe, including Britain). However, for hydrological modelling of a catchment in north-west England they chose to use Blaney-Criddle PE instead of Penman-Monteith PE, as the latter was considered to give a spread of daily values that was too large compared to observations, and to lead to percentage increases in PE that were too high (up to 80%) in summer; Blaney-Criddle gave a summer increase of less than 20%. Based on the latter research, Cameron (2006) decided to apply the Thornthwaite formula, rather than Penman-Monteith, to estimate changes in PE for modelling flood changes in the Lossie catchment in Scotland. They obtained changes in PE, for the 2080s time-horizon under the A2 emissions scenario, in the range 6.2% (June) to 20.8% (January).

Wilby and Harris (2006) use data from four GCMs with statistical downscaling, for the 2080s time-horizon under the A2 and B2 emissions scenarios, to model changes in low flows in the Thames. They obtain PE increases of 5-43% in winter and 11-22% in summer, but for two GCMs (ECHAM4 and HadCM3) the summer changes are larger than the winter ones, whereas for the other two GCMs (CGCM2 and CSIRO) the winter changes are larger than the summer ones. Wilby et al. (2006) use a similar method for modelling water resources and quality in the Kennet, using three of the same GCMs, but find much smaller increases in PE (3-9% in winter and 5-16% in summer) presumably due to the use of a different regression model for PE. Diaz-Nieto and Wilby (2005) showed that projected PE increases for the Thames derived via statistical downscaling (using temperature and specific and relative humidity) were roughly half those derived using the Penman-Monteith formula.

Chun et al. (2012) use data from the HadCM3 GCM for 1950-2099 under the A2 emissions scenario, to estimate changes in PE at 25 sites across Britain using both the Penman-Monteith formula and their GLM, which incorporates four meteorological variables (radiation, temperature, wind speed and relative humidity). They show annual PE increases almost everywhere, with greater percentage increases in the south-east than the north-west, but smaller percentage increases using the GLM than using Penman-Monteith. However, the site-specific parameter calibration required by the GLM may make its use under climate change questionable, as for Blaney-Criddle (Sperna Weiland et al. 2011).

Christierson et al. (2012) use UKCP09 probabilistic temperature change data, for the 2020s time-horizon under the A1B emissions scenario, and the Oudin temperature-based PE formula to calculate corresponding PE changes for each river-basin region in the UK, for their hydrological modelling of water resources. They present maps of the central estimate of PE changes which show an increase in PE across the country, with the largest changes occurring in the winter months. Plots of the range of monthly PE changes for two example catchments almost always show increases, usually in the range 0% to +30%, with a greater variation in winter than summer (but only present a 20-member subset of the full 10,000 UKCP09 set, selected via latin hypercube sampling). These were shown to be generally consistent with PE changes from an older set of scenarios, and with those calculated using RCM temperature data within the Oudin formula.

Similarly, Kay and Jones (2012) use UKCP09 probabilistic temperature change data, this time for the 2080s time-horizon under the A1B emissions scenario, and the Oudin PE formula to calculate corresponding PE changes for nine catchments in Britain. As shown by Christierson et al. (2012), the PE percentage changes are almost always positive, and are generally larger, with greater variation, in winter than summer. Here though, the changes are usually in the range 5% to 60%, which is larger than suggested by Christierson et al. (2012) because of the different time-horizons (2080s versus 2020s).

Prudhomme et al. (2012) use data from one of the 11 UKCP09 RCM ensemble members, for the 2050s time-horizon under the A1B emissions scenario, and compare the RCM-derived changes in PE across Great Britain for 12 PE formulae (five temperature-based, five radiation-based and two versions of Penman-Monteith). PE almost always shows an increase, although the magnitude varies between methods as well as by season and location. The largest variation between methods occurs during summer, and the temperature-based methods often (though not always) give larger increases than the other methods, with the radiation-based method often giving the smallest increases (or decreases).

Kingston et al. (2009) applied six different PE formulae, using data from five GCMs for a scenario representing a 2°C rise in global mean temperature, and compared the resulting latitudinally-averaged annual PE changes (for 60°S to 60°N). PE increases were found for all latitudes, GCMs and PE formulae, and broadly followed corresponding temperature changes. However, significant differences were found between different PE formulae; Hamon (Hamon 1961) and Jensen-Haise (Jenson and Haise 1963) gave the largest PE changes at most latitudes, then Hargreaves, Penman-Monteith and Priestly-Taylor, although the differences are less at the UK's latitude than at lower latitudes. Blaney-Criddle gave PE changes that varied much less with latitude than the other methods.

PE projections with changes in transpiration controls

Bell et al. (2011) developed a new method to estimate Penman-Monteith PE, which uses time-varying values of surface resistance produced by an RCM with an embedded land-surface scheme. This allows the surface resistance (r_c) to vary with the level of CO_2 in the atmosphere, which could be an important mechanism of change in PE. Comparisons over Britain, using data from the HadRM3 RCM for 1961-1990, showed that using r_c time-series from the RCM gave PE comparable to MORECS, but generally lower than when using MORECS' 12 fixed monthly values of r_c, particularly in spring. Similarly, for 2070-2099, using r_c time-series from the RCM gave lower PE than when using fixed monthly values of r_c. In terms of percentage increases in PE between the two periods, using fixed r_c gives increases in the range 15%-34%, whereas using r_c time-series gives increases in the range 3%-7% — significantly lower. However, it should be noted that not all possible feedbacks are included in this RCM run (for example, no changes to leaf area index or land cover are applied); PE changes may be somewhere between the two shown if other mechanisms were included.

Kay and Jones (2012) used the PE method developed by Bell et al. (2011), when applying RCM data alongside UKCP09 probabilistic data. They showed that percentage changes in PE from an 11-member RCM ensemble (with corresponding r_c time-series) were more variable than those derived using probabilistic temperature changes to calculate changes in Oudin PE. The biggest difference between the two sets of percentage changes in PE occurred in winter and spring, where the RCM PE changes were often lower than the probabilistic PE changes. However, even large percentage changes in PE are not really significant in winter in the UK, since the baseline PE is very small. Differences in spring could be more important.

Moratiel et al. (2011) investigate possible changes in grass reference PE in a part of Spain, including the effect of a change in r_c due to CO_2 concentrations rising from 372ppm to 550ppm by 2050. They do this by quantifying trends in observed weather variables for 1980-2009, extending these over the next 50 years, and using them to calculated Penman-Monteith PE both with the current r_c (70 s/m) and an approximated future r_c assuming stomatal closure (87 s/m). They find PE increases of about 10% with current r_c but only about 5% when r_c is also changed, thus reducing the PE increase by about half. This is similar to the findings of Bell et al. (2011), although there the differences are larger, perhaps because of the later time-horizon.

Effect of PE uncertainty on hydrological projections

From a hydrological perspective, differences in PE projections are only crucial if they are likely to result in different hydrological projections. The apparently frequent insensitivity of hydrological modelling to PE accuracy (Section 0) suggests that accurate PE projections will not always be important. Sperna Weiland et al. (2011), using reanalysis data for 1979-2002, show a general reduction in the variability of results between six different PE methods when moving down their global hydrological modelling chain, from PE to AE to runoff to river flows. They thus suggest that 'the selection of a PET method may be of minor influence on the resulting river flow modelled with a hydrological model', except for relatively limited regions where the variability remains high (which do not include the UK).

Kay and Davies (2008) look at the extra uncertainty introduced into hydrological impacts for three catchments in Britain, using PE changes derived from Penman-Monteith and Oudin formulae with data from 13 climate models (5 GCMs and 8 RCMs). The results suggest that the extra uncertainty is greatest for changes in low to median flows, although possibly still important for changes in high flows in some catchments, but that climate model uncertainty dominates.

Arnell (1999) tested the effect of using Priestly-Taylor PE in place of Penman-Monteith PE, in his modelling of changes in runoff in Europe for the 2050s under four different scenarios of climate change. As the PE increases using Priestly-Taylor PE were generally smaller than those for Penman-Monteith PE, the pattern of runoff changes followed that of rainfall changes more closely. However, the PE uncertainty was generally less than climate change scenario uncertainty.

Kingston et al. (2009) assessed the effect of six different PE formulae on the global extent of arid (rainfall/PE < 1) and humid (rainfall/PE ≥ 1) regions, and on regional water surpluses (annual rainfall minus PE, for months where rainfall > PE), with 5 GCMs for a 2°C rise in global mean temperature. They showed that, although almost all GCMs and PE formulae agreed on an increase in the extent of arid areas (and corresponding decrease in the extent of humid areas), they disagreed on the amount of change. Since the east of England is verging on being arid (Figure 1), this uncertainty could affect parts of Britain. Furthermore, the regional water surplus analysis of Kingston et al. (2009) indicates that PE uncertainty is of comparable magnitude to GCM uncertainty, although this analysis was only done for Mediterranean, East African and Southeast Asian regions, so did not cover Britain.

Betts et al. (2007) modelled the change in continental runoff under a doubling of CO_2, and showed that including the effect of stomatal closure under higher CO_2 concentrations led to larger increases in runoff (17±5%) than including only radiative forcing (11±6%). The difference was slightly less when changes in land cover and leaf area index were also included. Similar effects were seen for changes in flood peaks in the Thames Basin, using data from an 11-member perturbed-physics RCM ensemble to drive the Grid-to-Grid hydrological model (Bell et al. 2012). There, the percentage changes in flood peaks were higher when averaged over the 6-member sub-ensemble which included the effect of stomatal closure, and lower when averaged over the 5-member sub-ensemble which did not included the effect of stomatal closure. However, the simultaneous variation of other parameters in the climate model perturbed-physics ensemble means that this result should be interpreted with caution.

Confidence assessment

Figure 5 presents an assessment of the confidence in the direction of historical trends and future projections for evaporation in the UK, using the “evidence” and “agreement” classifications of the IPCC. (www.ipcc.ch/pdf/supporting-material/uncertainty-guidance-note.pdf). The “evidence” classification draws on the amount and quality of evidence of spatially and temporally consistent changes, while the “agreement” classification considers the level of agreement between different studies.

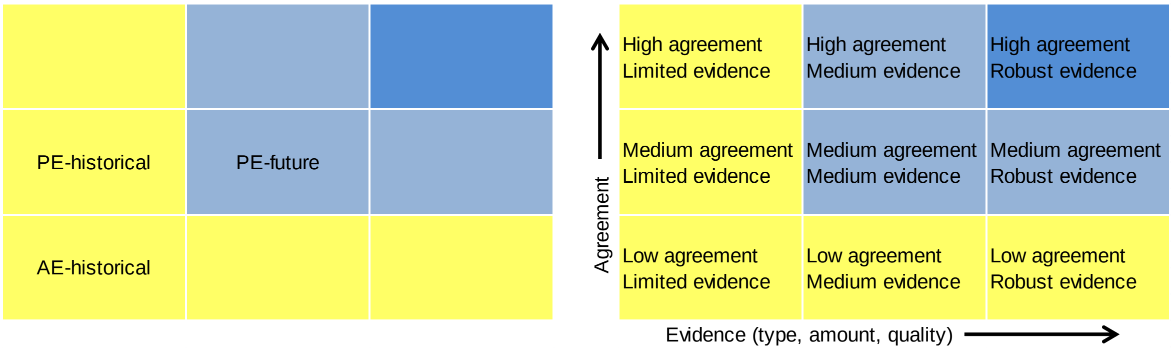

Figure 5 Confidence assessment for the direction of historical trends in PE and AE in the UK (PE-historical and AE-historical) and the direction of future projections of PE in the UK (PE-future) (left), with the assessment key (right).

The assessment of the direction of historical trends in PE and AE is based on the studies discussed in Section 0, while the assessment of the direction of change indicated by future PE projections is based on the studies discussed in Sections 0 and 0. Judgment of each class is clearly subjective. In this case there is judged to be limited evidence of historical trends in either PE or AE, but with more agreement about increasing trends in PE (Medium agreement) than about any trends in AE (Low agreement). There are more studies discussing future PE projections (Medium evidence) and most agree on an increase in PE in the UK (Medium agreement).

Note that this confidence assessment considers only the direction, not the magnitude, of any evaporation trends or projections. This is due to the general lack of quantification of historical trends and the difficulties in comparing projections between disparate studies (see discussion in Section 6).

Discussion and conclusions

The aim of this paper was to present a review of historical trends and future projections for evaporation in the UK, and to provide a hydrological perspective on these.

Evaporation is frequently estimated from meteorological data, using either empirical formulae (of varying complexity) or formulae based on representation of physical processes. The Met Office's MORECS and MOSES systems apply the physically-based Penman-Monteith PE formulation to derive PE estimates from meteorological observations over Britain. These PE data are often used, along with rainfall, as inputs for hydrological modelling, although there is ongoing debate as to which to use: MORECS or MOSES. While hydrological models may be relatively insensitive to the absolute accuracy of PE data, inaccurate data can cause difficulties with calibration such as failure to close the annual water balance or instability in model parameters. Although PE is less spatially and temporally variable than rainfall, and has a well-defined seasonal pattern, AE accounts for the loss of a significant proportion of incoming rainfall across much of southern and eastern England. These factors are important considerations when hydrological models are applied, especially when simulating the impacts of environmental change (including climate change).

There is relatively little information on historical trends in evaporation in the UK, or indeed globally, whether actual or potential evaporation (AE or PE). The few studies there have been in the UK generally indicate increases in PE, and one study has shown an increase in AE. Globally, both increases and decreases in PE have been detected in different areas, with different causes, although decreasing wind speeds tend to be the primary cause suggested for any PE decreases. A more recent global decline in data-driven estimates of AE has been attributed to decreasing soil moisture.

Direct measurement of evaporation is progressing from localised observations to remotely sensed estimates from large areas. Combined analyses for the UK of AE measurements, satellite data, observed changes in the meteorological variables which affect evaporation, and observed changes in river flow data should, together with further modelling, enable a better understanding of the direction, magnitude and causes of any changes in AE and PE (e.g. Donohue et al. 2010).

There is little consensus within the scientific community on the best approach for deriving future projections of PE; some authors believe that PE projections must use formulae which include all meteorological variables that influence PE, while others believe that the sensitivity of such formulae to data quality makes this inadvisable, or that the choice will not make much difference in subsequent hydrological modelling. The dilemma is summarised by Kingston et al. (2009), who considers 'whether more reliable estimation of changes in PET can be obtained from physically-based methods (e.g. Penman-Monteith) with uncertain data quality, or more empirical methods (e.g. Hargreaves) with more reliable input data'. Vorosmarty et al. (1998) argue that 'Although these [physically-based PE] methods are attractive on theoretical grounds, the degree to which the necessary input data sets can be successfully assembled…remains an open question. Use of more physically realistic evaporation functions must be weighed against potential inaccuracies in, and inconsistencies among, the several climatic forcing fields used by these methods'. The choice is further complicated when considering possible changes in plant behaviour under higher carbon dioxide concentrations (stomatal closure, increased plant growth etc).

Most studies presenting PE projections for the UK indicate increases in annual PE totals, although some studies suggest (small) decreases in PE for one or two months of the year. However, there is considerable variation in the magnitude of the projections, caused not just by the PE formula applied but by the climate model, emissions scenario and time-horizon, and according to the location investigated. These factors make it difficult to compare PE projections between studies, as do the varied ways of describing the changes (i.e. percentage or absolute changes, and monthly, seasonal or annual changes).

It has been suggested that the choice of PE scheme could be particularly important in regions where precipitation and PE are in close balance with each other (e.g. Kingston et al. 2009). The maps in Section 0, based on MORECS data for 1961-1990, show that this is likely to include parts of the UK, although changes in rainfall as well as PE make it difficult to predict precisely which areas are likely to be most affected. Bormann (2011) suggests calculating the change in the annual climatic water budget (change in annual precipitation minus annual PE, from baseline to future climate), to test the sensitivity of a given location to choice of PE scheme. Where the changes are relatively consistent between different PE schemes, the choice is probably not crucial, but the choice needs greater consideration where there is more variation between schemes. However, likely changes in the seasonality of UK rainfall under climate change (Murphy et al. 2009) mean that such annual water balance tests may mask important seasonal changes. Plots of monthly mean rainfall and PE under climate change may be more informative.

In some circumstances, it may be necessary to extend the sensitivity study through to the hydrological modelling, as the importance of PE is likely to vary according to the hydrological aspect under investigation as well as the catchment location (e.g. Kay and Davies 2008). Nevertheless, a number of studies have shown that impacts uncertainty due to climate model structure is greater than uncertainty due to PE formulation, which suggests that, where the capacity for hydrological model runs is limited, the priority should be to cover climate model uncertainty more comprehensively.

The question remains: which PE schemes are likely to be more (or less) reliable when applied with climate model data? The answer requires investigation of 1) which meteorological variables are likely to be most important in terms of possible PE changes in the UK (i.e. what complexity of model is really necessary here?), and 2) the reliability of each of the meteorological variables when taken from climate models for baseline and future climates in the UK. Together, these could enable the derivation of improved projections of PE for the UK from climate model data, provided the reliability of the appropriate climate model variables can be demonstrated. Further investigation is also needed into the effect of climate change on plant transpiration, and the feedbacks between climate change, plant growth and land-cover change.

Clarification of the best ways to both measure and model evapotranspiration (AE or PE) would be particularly useful, as the global hydrological and climate modelling of Ziegler et al. (2003) suggests that the detection time for trends in evaporation should be less than that for either precipitation or runoff, because of the much greater natural variability of the latter two. For example, for Europe for 2000-2099 (under the A2 emissions scenario) their analysis gives an AE trend of 0.45 mm/yr, detectable in 48 years, whereas the precipitation trend is larger (0.62 mm/yr) but only detectable in 58 years, and the runoff trend is smaller (0.14 mm/yr) and only detectable in 132 years.

References

Allen, R.G., Pereira, L.S., Raes, D. and Smith, M. (1998). Crop evapotranspiration (guidelines for computing crop water requirements). FAO Irrigation and Drainage Paper No. 56.

Allen, R.G., Smith, M., Pereira, L.S. and Perrier, A. (1994). An update for the calculation of reference evapotranspiration. ICID Bulletin, **43**(2), 35-92.

Andréassian, V., Perrin, C. and Michel, C. (2004). Impact of imperfect potential evapotranspiration knowledge in the efficiency and parameters of watershed models. Journal of Hydrology, **286**, 19-35.

Arnell, N.W. (1999). The effect of climate change on hydrological regimes in Europe: a continental perspective. Global Environmental Change, **9**, 5-23.

Arnell, N.W. (2011). Uncertainty in the relationship between climate forcing and hydrological response in UK catchments. Hydrology and Earth System Sciences, **15**, 897-912.

Bates, B.C., Kundzewicz, Z.W., Wu, S. and Palutikof, J.P. (Eds.), (2008). Climate Change and Water. Technical Paper of the Intergovernmental Panel on Climate Change, IPCC Secretariat, Geneva, 210pp.

Bell, V.A, Gedney, N., Kay, A.L., Smith, R., Jones, R.G. and Moore, R.J. (2011). Estimating potential evaporation from vegetated surfaces for water management impact assessments using climate model output. Journal of Hydrometeorology, **12**, 1127-1136, doi: 10.1175/2011JHM1379.1.

Bell, V.A., Kay, A.L., Cole, S.J., Jones, R.G., Moore, R.J. and Reynard, N.S (2012). How might climate change affect river flows across the Thames Basin? An area-wide analysis using the UKCP09 Regional Climate Model ensemble. Journal of Hydrology, **442-443**, 89-104, doi: 10.1016/j.jhydrol.2012.04.001.

Bell, V.A., Kay, A.L., Jones, R.G., Moore, R.J. and Reynard, N.S. (2009). Use of soil data in a grid-based hydrological model to estimate spatial variation in changing flood risk across the UK. Journal of Hydrology, **377**, 335-350.

Best, M.J., Pryor, M., Clark, D.B. et al. (2011). The Joint UK Land Environment Simulator (JULES), model description – Part 1: Energy and water fluxes, Geosci. Model Dev., **4**, 677-699.

Betts, R.A., Boucher, O., Collins, M., Cox, P.M., Falloon, P.D., Gedney, N., Hemming, D.L., Huntingford, C., Jones, C.D., Sexton, D.M.H. and Webb, M.J. (2007). Projected increase in continental runoff due to plant responses to increasing carbon dioxide. Nature, **448**, 1037-1041.

Blaney, H.F. and Criddle, W.D. (1950). Determining water requirements in irrigated areas from climatological and irrigation data. US Department of Agriculture (Soil Conservation Service) Technical Paper 96.

Blyth, E.M., Gash, J.H.C., Lloyd, A., Pryor, M., Weeden, G.P. and Shuttleworth, W.J. (2010). Evaluating the JULES land surface model energy fluxes using FLUXNET data. Journal of Hydrometeorology, **11**, 509-519.

Bormann (2011). Sensitivity analysis of 18 different potential evapotranspiration models to observed climatic change at German climate stations. Climatic Change, **104**, 729-753.

Boughton, W. (2006). Calibrations of a daily rainfall-runoff model with poor quality data. Environmental Modelling and Software, **21**, 1114-1128.

Burn, D.H. and Hesch, N.M. (2007). Trends in evaporation for the Canadian Prairies. Journal of Hydrology, **336**, 61-73.

Burt, C.M., Mutziger, A.J., Allen, R.G. and Howell, T.A. (2005). Evaporation research: Review and interpretation. Journal of Irrigation and Drainage Engineering, ASCE **131**, 37-58.

Burt, T.P. and Shahgedanova (1998). An historical record of evaporation losses since 1815 calculated using long-term observations from the Radcliffe Meteorological Station, Oxford, England. Journal of Hydrology, **205**, 101-111.

Cain, J.D., Batchelor, C.H., Gash, J.H.C. and Harding, R.J. (1998). Comment on the paper ‘Towards a rational definition of potential evaporation’ by J.P. Lhomme. Hydrology and Earth System Sciences, **2**, 257-264.

Calder, I.R., Harding, R.J. and Rosier, P.T.W. (1983). An objective assessment of soil-moisture deficit models. Journal of Hydrology, **60**, 329-355.

Cameron, D. (2006). An application of the UKCIP02 climate change scenarios to flood estimation by continuous simulation for a gauged catchment in the northeast of Scotland, UK (with uncertainty). Journal of Hydrology, **328**, 212-226.

Chun, K.P., Wheater, H.S. and Onof, C. (2012). Projecting and hindcasting potential evaporation for the UK between 1950 and 2099. Climatic Change, **113**, 639-661.

Christierson. B.V., Vidal, J. and Wade, S.D. (2012). Using UKCP09 probabilistic climate information for UK water resource planning. Journal of Hydrology, **424-425**, 48-67.

Cox, P.M., Huntingford, C. and Harding, R.J. (1998). A canopy conductance and photosynthesis model for use in a GCM land surface scheme. Journal of Hydrology, **213**, 79-94.

Cox, P.M., Betts, R.A., Bunton, C.B., Essery, R.L.H., Rowntree, P.R. and Smith, J. (1999). The impact of new land surface physics on the GCM simulation of climate and climate sensitivity. Climate Dynamics, **15**, 183–203.

Crane, S.B. and Hudson, J.A. (1997). The impact of site factors and climate variability on the calculation of potential evaporation at Moel Cynnedd, Plynlimon. Hydrology and Earth System Sciences, **1**, 429-445.

Crooks, S. and Davies, H. (2001). Assessment of land use change in the Thames catchment and its effect on the flood regime of the river. Physics and Chemistry of the Earth B, **26**, 583-591.

Crooks, S.M. and Naden, P.S. (2007). CLASSIC: a semi-distributed modelling system. Hydrology and Earth System Sciences, **11**, 516-531.

Diaz-Nieto, J, and Wilby, R.L. (2005). A comparison of statistical downscaling and climate change factor methods: impacts on low flows in the River Thames, United Kingdom. Climatic Change, **69**, 245-268.

Donohue, R.J., McVicar, T.R. and Roderick, M.L. (2010). Assessing the ability of potential evaporation formulations to capture the dynamics in evaporative demand within a changing climate. Journal of Hydrology, **386**, 186-197.

Dunn, S.M. and Mackay, R. (1995). Spatial variation in evapotranspiration and the influence of land use on catchment hydrology. Journal of Hydrology, **171**, 49-73.

Ekstrom, M., Jones, P.D., Fowler, H.J., Lenderink, G., Buishand, T.A. and Conway, D. (2007). Regional climate model data used within the SWURVE project 1: projected change in seasonal patterns and estimation of PET. Hydrology and Earth System Sciences, **11**, 1069-1083.

Federer, C.A., Vorosmarty, C. and Fekete, B. (1996). Intercomparison of methods for calculating potential evaporation in regional and global water balance models. Water Resources Research, **32**, 2315-2321.

Fisher. J.B., Whittaker, R.J. and Malhi, Y. (2011). ET come home: potential evapotranspiration in geographical ecology. Global Ecology and Biogeography, **20**, 1-18.

Fowler, A. (2002). Assessment of the validity of using mean potential evaporation in computations of the long-term soil water balance. Journal of Hydrology, **256**, 248-263.

Fowler, H.J., Tebaldi, C. and Blenkinsop, S. (2008). Probabilistic estimates of climate change impacts on flows in the river Eden, Cumbria. Proceedings of the British Hydrological Society 10th National Hydrology Symposium, Exeter, 15-17th September 2008.

Fu, G., Charles, S.P. and Yu, J. (2009). A critical overview of pan evaporation trends over the last 50 years. Climatic Change, **97**, 193-214.

Gash, J.H.C. (1979). An analytical model of rainfall interception by forests. Q. J. R. Meteorol. Soc., **105**. 43-55.

Gedney, N., Cox, P.M., Betts, R.A., Boucher, O., Huntingford, C. and Stott, P.A. (2006). Detection of a direct carbon dioxide effect in continental river runoff records. Nature, **439**, 835-838.

Haddeland, I., Clark, D., Franssen, W. et al. (2011). Multi-model estimate of the global water balance: setup and first results. Journal of Hydrometeorology. doi:10.1175/2011JHM1324.1.

Hamon, W.R. (1961). Estimating potential evaporation. In: Division, J.o.H. (Ed.), Proceedings of the American Society of Civil Engineers, 107-120.

Hargreaves, G.H. and Samani, Z.A. (1985). Reference crop evapotranspiration from temperature. Appl. Eng. Agric., **1**, 95-99.

Hill, M.O., Roy, D.B., Mountford, J.O. and Bunce, R.G.H. (2000). Extending Ellenberg's indicator values to a new area: an algorithmic approach. Journal of Applied Ecology, **37**, 3-15.

Hough, M. and Jones, R.J.A. (1997). The United Kingdom Meteorological Office rainfall and evaporation calculation system: MORECS version 2.0 — an overview. Hydrology and Earth System Sciences, **1**, 227-239.

Hough, M., Palmer, S., Weir, A., Lee, M., and Barrie, I. (1996). The Meteorological Office Rainfall and Evaporation Calculation System: MORECS version 2.0 (1995). An update to Hydrological Memorandum 45, The Met. Office, Bracknell.

Irmak, S., Kabenge, I., Skaggs, K.E. and Mutiibwa, D. (2012). Trend and magnitude of changes in climate variables and reference evapotranspiration over 116-yr period in the Platte River Basin, central Nebraska-USA. Journal of Hydrology, **420-421**, 228-244.

Jenkins, G.J., Perry, M.C. and Prior, M.J.O. (2007). The climate of the United Kingdom and recent trends. UKCP09 scientific report, Met Office Hadley Centre, Exeter, UK.

Jenson, M.E. and Haise, H.R. (1963). Estimating evapotranspiration from solar radiation. Journal of Irrigation and Drainage Division, ASCE **89**, 15-41.

Jones, P.D., Kilsby, C.G., Harpham, C., Glenis, V., Burton, A. (2009). UK Climate Projections science report: Projections of future daily climate for the UK from the Weather Generator. University of Newcastle, UK.

Jung, M., Reichstein, M., Ciais, P., Seneviratne, S.I. et al (2010). Recent decline in the global land evapotranspiration trend due t o limited moisture supply. Nature, **467**, 951-954.

Kay, A.L. and Davies, H.N. (2008). Calculating potential evaporation from climate model data: a source of uncertainty for hydrological climate change impacts. Journal of Hydrology, **358**, 221– 239.

Kay, A.L. and Jones, R.G. (2012). Comparison of the use of alternative UKCP09 products for modelling the impacts of climate change on flood frequency. Climatic Change, **114**(2), 211-230, doi:10.1007/s10584-011-0395-z.

Kingston, D.G., Todd, M.C., Taylor, R.G., Thompson, J.R. and Arnell, N.W. (2009). Uncertainty in the estimation of potential evaporation under climate change. Geophysical Research Letters, **36**, doi:10.1029/2009GL040267.

Kleinn, J., Frei, C., Gurtz, J., Luthi, D., Vidale, P.L. and Schar, C. (2005). Hydrological simulations in the Rhine Basin driven by regional climate models. Journal of Geophysical Research, **110**, doi:10.1029/2004JD005143.

Lhomme, J.-P. (1997). Towards a rational definition of potential evaporation. Hydrology and Earth system Sciences, **1**, 257-264.

Manning, L.J., Hall, J.W., Fowler, H.J., Kilsby, C.G. and Tebaldi, C. (2009). Using probabilistic climate change information from a multimodel ensemble for water resources assessment. Water Resources Research, **45**, doi:10.1029/2007WR006674.

McVicar, T.R., Roderick, M.L., Donohue, R.J., Tao Li, L., Van Niel, T.G., Thomas, A. Grieser, J., Jhajharia, D., Himri, Y., Mahowald, N.M., Mescherskaya, A.V., Kruger, A.C., Rehman, S. and Dinpashoh, Y. (2012). Global review and synthesis of trends in observed terrestrial near-surface wind speeds: Implications for evaporation. Journal of Hydrology, **416-417**, 182-205.

Miralles, D.G., De Jeu, R.A.M., Gash, J. H., Holmes, T.R.H. and Dolman, A. J. (2011). Magnitude and variability of land evaporation and its components at the global scale. Hydrology and Earth System Sciences, **15**, 967-981.

Monteith, J.L. (1965). Evaporation and environment. Symposia of the Society for Experimental Biology, **19**, 205-234.

Moore, R.J. (2007). The PDM rainfall-runoff model. Hydrology and Earth System Sciences, **11**, 483-499.

Moratiel, R., Snyder, R.L., Duran, J.M. and Tarquis, A.M. (2011). Trends in climatic variables and future reference evapotranspiration in Duero Valley (Spain). Natural Hazards and Earth System Sciences, **11**, 1795-1805.

Murphy, J.M., Sexton, D.M.H., Jenkins, G.J., et al. (2009) UK Climate Projections Science Report: Climate change projections. Met Office Hadley Centre, Exeter, UK.

Nandakumar, N. and Mein, R.G. (1997). Uncertainty in rainfall-runoff model simulations and the implications for predicting the hydrologic effects of land-use change. Journal of Hydrology, **192**, 211–232

Oudin, L., Hervieu, F., Michel, C., Perrin, C., Andreassian, V., Anctil, F. and Loumagne, C. (2005a). Which potential evapotranspiration input for a lumped rainfall-runoff model? Part 2 — Towards a simple and efficient potential evapotranspiration model for rainfall-runoff modelling. Journal of Hydrology, **303**, 290-306.

Oudin, L., Michel, C. and Anctil, F. (2005b). Which potential evapotranspiration input for a lumped rainfall-runoff model? Part 1 — Can rainfall-runoff models effectively handle detailed potential evapotranspiration inputs? Journal of Hydrology, **303**, 275-289.

Oudin, L., Moulin, L., Bendjoudi, H. and Ribstein, P. (2010). Estimating potential evapotranspiration without continuous daily data: possible errors and impact on water balance equations. Hydrological Sciences Journal, **55**, 209-222.

Parmele, L.H. (1972). Errors in output of hydrologic models due to errors in input potential evapotranspiration. Water Resources Research, **8**, 348-359.

Paturel, J.E., Servat, E. and Vassiliadis, A. (1995). Sensitivity of conceptual rainfall-runoff algorithms to errors in input data — case of the GR2M model. Journal of Hydrology, **168**, 111–125

Penman, H.L. (1948). Natural evaporation from open water, bare soil and grass. Proceedings of the Royal Society of London, **193**, 120-145.

Pereira, L.S., Perrier, A., Allen, R.G. and Alves, I. (1999). Evapotranspiration: Concepts and future trends. Journal of Irrigation and Drainage Engineering, **125**, 45-51.

Perry, M. (2006). A spatial analysis of trends in the UK climate since 1914 using gridded datasets. Climate Memorandum No 21, Met Office, UK.

Piani, C., Weedon, G.P., Best, M., Gomes, S.M., Viterbo, P., Hagemann, S. and Haerter J.O. (2010). Statistical bias correction of global simulated daily precipitation and temperature for the application of hydrological models. Journal of Hydrology, **395**, 199-215.

Price, D.J. and McKenna, J.E. (2003). Climate change: review of levels of protection offered by flood prevention schemes: UKCIP02 update (2003). Report to Scottish Executive, Babtie Group.

Priestley, C.H.B. and Taylor, R.J. (1972). On the assessment of surface heat fluxes and evaporation using large-scale parameters. Monthly Weather Review, **100**, 81-92.

Prudhomme, C., Crooks, S., Boelee, L., Williamson, J. and Davies, H. (2012). Derivation of RCM-driven potential evapotranspiration for Great Britain. Report to Environment Agency, Project Note SC090016/PN2, CEH Wallingford.

Reynard, N.S. and Arnell, N.W. (1993). Sensitivity of potential evaporation to climate change. Paper to conference on Rainfall and Evaporation, Bratislava, Slovak Republic, 20-24 September 1993. pp106-113.

Robson, A.J. (2002). Evidence for trends in UK flooding. Phil. Trans. R. Soc. Lond. A, **360**, 1327-1343.

Senay, G.B., Leake, S., Nagler, P.L., Artan, G., Dickinson, J., Cordova, J.T. and Glenn, E.P. (2011). Estimating basin scale evapotranspiration (ET) by water balance and remote sensing methods. Hydrological Processes, **25**, 4037-4049.

Sexton, D.M.H. and Murphy J. (2010). UKCP09: Probabilistic projections of wind speed. Met Office Hadley Centre, Exeter, UK.

Shaw, S.B. and Riha, S.J. (2011). Assessing temperature-based PET equations under a changing climate in temperate, deciduous forests. Hydrological Processes, **25**, 1466-1478.

Shutov, V., Gieck, R.E., Hinzman, L.D. and Kane, D.L. (2006). Evaporation from land surface in high latitude areas: A review of methods and study results. Nordic Hydrology, **37**, 393-411.

Shuttleworth, W.J. (2007). Putting the 'vap' into evaporation. Hydrology and Earth System Sciences, **11**, 210-244.

Smith, R.N.B., Blyth, E.M., Finch, J.W., Goodchild, S., Hall, R.L. and Madry, S. (2006). Soil state and surface hydrology diagnosis based on MOSES in the Met Office Nimrod nowcasting system. Meteorological Applications, **13**, 89–109.

Sperna Weiland, F.C., Tisseuil, C., Durr, H.H., Vrac, M. and van Beek, L.P.H. (2011). Selecting the optimal method to calculate daily global reference potential evaporation from CFSR reanalysis data. Hydrology and Earth System Sciences Discussion, **8**, 7355-7398.

Teuling, A.J., Seneviratne, S.I., Stöckli, R. et al. (2010). Contrasting response of European forest and grassland energy exchange to heatwaves. Nature Geoscience, **3**, 722–727 doi:10.1038/ngeo950.

Thomas, A. (2000). Spatial and temporal characteristics of potential evapotranspiration trends over China. International Journal of Climatology, **20**, 381-396.

Thompson, N., Barrie, I.A. and Ayles, M. (1981). The Meteorological Office Rainfall and Evaporation Calculation System: MORECS (July 1981). Hydrological Memorandum No. 45, Met Office, Bracknell.

Thornthwaite, C.W. (1948). An approach towards a rational classification of climate. Geographical Review, **38**, 55-94.

Van der Schrier, G., Jones, P.D. and Briffa, K.R. (2011). The sensitivity of the PDSI to the Thormthwaite and Penman-Monteith parameterizations for potential evaporation. Journal of Geophysical Research, **116**, doi:10.1029/20120JD015001.

Verhoef, A. and Campbell, C.L. (2005). Evaporation measurement. In: M.G. Anderson (Ed.), Encyclopedia of hydrological sciences. J. Wiley, Chichester, West Sussex, England; Hoboken, NJ, pp589-601.

Vorosmarty, C.J., Federer, C.A. and Schloss, A.L. (1998). Potential evaporation functions compared on US watersheds: Possible implications for global-scale water balance and terrestrial ecosystem modeling'. Journal of Hydrology, **207**, 147-169.

Weedon, G.P., Gomes, S., Viterbo, P. et al. (2011). Creation of the WATCH Forcing Data and Its use to assess global and regional reference crop evaporation over land during the twentieth century. Journal of Hydrometeorology, **12**, 823-848.

Wilby, R.L., Greenfield, B. and Glenny, C. (1994). A coupled synoptic-hydrological model for climate change impact assessment. Journal of Hydrology, **153**, 265-290.

Wilby, R.L. and Harris, I. (2006). A framework for assessing uncertainties in climate change impacts: Low-flow scenarios for the River Thames, UK. Water Resources Research, **42**, doi:10.1029/2005WR004065.

Wilby, R.L., Whitehead, P.G., Wade, A.J., Butterfield, D., Davis, R.J. and Watts, G. (2006). Integrated modelling of climate change impacts on water resources and quality in a lowland catchment: River Kennet, UK. Journal of Hydrology, **330**, 204-220.

Wullschleger, S.D., Gunderson, C.A., Hanson, P.J., Wilson, K.B. and Norby, R.J. (2002). Sensitivity of stomatal and canopy conductance to elevated CO_2 concentration - interacting variables and perspectives of scale. New Phytologist, **153**, 485-496.

Yang, C., Chandler, R.E., Isham, V.S., Annoni, C. and Wheater, H.S. (2005). Simulation and downscaling models for potential evaporation. Journal of Hydrology, **302**, 239-254.

Ziegler, A.D., Sheffield, J., Maurer, E.P., Nijssen, B., Wood, E.F. and Lettenmaier, D.P. (2003). Detection of intensification in global- and continental-scale hydrological cycles: Temporal scale of evaluation. Journal of Climate, **16**, 535-547.