

A hydrological perspective on evaporation: historical trends and future projections in Britain

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ABSTRACT

Evaporation is an important component of the hydrological cycle. Potential evaporation (PE) from a vegetated surface is the amount of water that would be lost to the atmosphere were the supply unlimited; actual evaporation (AE) is a fraction of PE dependent on soil wetness. Many formulae exist for estimating PE from meteorological data. PE is usually a required input, with rainfall, for hydrological modelling, but PE accuracy is generally considered less important than rainfall accuracy for model performance. Few studies investigate historical evaporation trends in Britain, but generally indicate increases. Most studies presenting future PE projections for Britain indicate increased annual PE, but some suggest small decreases in some months. Limited consensus on the best formulae to derive PE projections from climate model data is further complicated by possible changes in plant behaviour (transpiration and growth) under higher carbon dioxide concentrations. Appropriate PE estimation could be particularly important in regions where precipitation and PE are in close balance, but PE uncertainty could be less important than climate model uncertainty for hydrological impacts. Further research is needed into which PE formulae are likely to be most reliable when applied with climate model data, and into climate change and plant feedbacks.

Key words | actual evaporation, climate change impacts, evapotranspiration, hydrology, potential evaporation

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INTRODUCTION

Evaporation transfers water from the land-surface to the atmosphere, so is an important part of the hydrological cycle. While there is ongoing debate on terminology (Lhomme 1997; Cain *et al.* 1998), here the terms evaporation and evapotranspiration each encompass transfer via both evaporation (loss of water lying on a surface, e.g., lake, soil, leaf) and transpiration (loss via plant stomata). Potential evaporation (PE or PET) is generally considered as the amount of water that would be lost to the atmosphere if there were no limits to soil-moisture supply (Federer *et al.* 1996). So actual evaporation (AE or AET) can be estimated as a fraction of PE dependent on soil wetness; it can be less than PE if soils are dry, but generally cannot be greater than PE. PE 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. To simplify matters, the 'reference crop' concept was introduced, with average crop parameters provided for estimation of reference PE. The reference crop is often 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), Hamon (1961) and Oudin *et al.* (2005a)), while the most complex and physically based formulae involve all four meteorological variables (e.g., Penman–Monteith; Monteith 1965), with a range in between (e.g., Blaney–Criddle, involving temperature and sunshine (Blaney & Criddle 1950), and Priestley–Taylor, involving temperature and radiation (Priestley & Taylor 1972)). Oudin *et al.* (2005a) provide a useful summary of 17 variations found in the literature. Penman–Monteith is 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) and the Met Office Surface Exchange Scheme (MOSES; Cox *et al.* 1998) use modified versions of Penman–Monteith.

PE estimates are a required input for hydrological modelling, alongside rainfall. PE changes, on their own or in combination with rainfall changes, can contribute to changes in hydrological indices like mean monthly river flows. This paper provides some background on AE and PE in Britain, influence of PE in hydrological modelling, and the ways in which environmental change can affect PE. A review of historical evaporation trends in Britain is presented, with a global context. Future PE projections in Britain are reviewed, with estimation difficulties and effects of PE uncertainty on hydrological climate change impacts discussed. Finally, a discussion and conclusions are presented. Although the focus here is the hydrological perspective, evaporation is of interest in other areas (e.g., agriculture and ecology; Fisher *et al.* 2011), and many of the same issues will apply.

BACKGROUND

Penman–Monteith PE, MORECS and MOSES

Penman–Monteith PE for short grass (m/s) is given by:

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

with λ latent heat flux (J/kg), ρ_w water density (kg/m³), ρ_a air density (kg/m³), c_p specific heat of air (J/kg/°C), γ the psychrometric constant (kPa/°C), $e_a = e(T_a) = 0.611 \exp(17.27T_a/(T_a + 237.3))$ saturation vapour pressure (kPa) with T_a air temperature (°C), $e_d = e(T_d)$ actual vapour pressure (kPa) with T_d dew-point temperature (°C), $\Delta = de_a/dT_a = 17.27 \times 237.3e_a/(T_a + 237.3)^2$ the slope of the vapour pressure curve (kPa/°C), R_n net solar radiation (J/m²/s), G soil heat flux (J/m²/s), $r_a = 208/W_2$ aerodynamic resistance (s/m) with W_2 wind speed (m/s) at a 2 m height, and r_c canopy surface resistance of short grass (s/m). The FAO recommend $r_c = 70$ s/m for short grass (height 0.12 m) (Allen *et al.* 1994). The equation was developed for daily weather data but its applicability has been demonstrated for monthly mean data (Allen *et al.* 1998; Oudin *et al.* 2010). Penman PE (Penman 1948) has $r_c = 0$ and $G = 0$, essentially representing loss from open water.

Penman–Monteith forms part of MORECS, which converts daily synoptic weather station data into estimates of weekly and monthly PE, AE and soil moisture deficit (SMD) for short grass (and several other land covers) on a 40 × 40 km grid over Britain, for a range of soils defined by their available water capacity (AWC; Hough & 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 uses monthly-varying r_c for short grass (height 0.15 m), ranging from 44.5 s/m (late spring) to 88.7 s/m (winter) (average ~73 s/m).

MOSES calculates water and energy fluxes between the land-surface and atmosphere alongside estimates of carbon dioxide (CO₂) fluxes and their effects on vegetation physiology (Cox *et al.* 1999). It can run operationally at a 2 km resolution and hourly time-step (Smith *et al.* 2006), providing outputs including soil moisture, AE and PE for a range of vegetated and non-vegetated surface types, with land-cover maps used to determine the proportion of each type within each grid-box. Like MORECS, MOSES calculates PE using Penman–Monteith but with variable r_c assumed to depend on photosynthesis, which depends on air temperature, radiation, humidity and vegetation type. MOSES also includes processes such as canopy evaporation and sublimation from snow. MOSES now forms the basis of the

Joint UK Land Environment Simulator (JULES; [Best *et al.* 2011](#)), which is a land-surface model used for research into the carbon cycle and climate change at local and global scales. JULES has been benchmarked against eddy-covariance AE data from around the world ([Blyth *et al.* 2010](#)). MOSES and JULES PE can be expected to change in line with contemporary understanding of earth-system processes, while MORECS PE reflects a standardised, but arguably older, physical understanding.

PE, AE and hydrological modelling

PE estimates (or the variables 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 CATCHMOD (the Thames Catchment Model; [Wilby *et al.* 1994](#)) or CLASSIC (Climate and LAnd use Scenario Simulation In Catchments; [Crooks & Naden 2007](#)), and fully distributed area-wide models, like G2G (Grid-to-Grid; [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 ([Sheffield *et al.* 2012](#)). Typically, hydrological models estimate AE as a fraction of PE dependent upon soil moisture, modelled continuously using a water-accounting scheme.

PE is much less spatially and temporally variable than rainfall, and highly seasonally predictable ([Calder *et al.* 1983](#)), so relatively simple data are considered sufficient to obtain a closed water balance in hydrological modelling. Thus rainfall accuracy is generally more important than PE accuracy for hydrological model performance ([Paturel *et al.* 1995](#); [Nandakumar & Mein 1997](#); [Boughton 2006](#); [Manning *et al.* 2009](#)). However, [Parmele \(1972\)](#) shows varying sensitivity to PE biases of +10% and -20% in nine US catchments, indicating that relative importance is likely to vary by location, as well as how performance is assessed (e.g., focussing on high or low flows). Also, some of the perceived insensitivity of models to PE accuracy could be due to recalibration, which can allow model parameters to compensate for differing PE ([Andréassian *et al.* 2004](#)).

[Oudin *et al.* \(2005a\)](#) tested 27 PE formulations with four rainfall-runoff models for 308 catchments in France, Australia and the USA. They found that PE based on temperature or radiation often provided more accurate streamflow

simulations than more complex formulae, but PE from all formulae was scaled by mean annual Penman PE, so only the effects of variability were tested, and the models were recalibrated for each set of inputs. [Vorosmarty *et al.* \(1998\)](#) compare 11 PE formulae over the USA and state ‘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, for their modelling of Thames water resources, ‘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 PE (i.e., a seasonal pattern duplicated each year) can be used, rather than PE time-series. [Calder *et al.* \(1983\)](#) showed that using PE time-series led to no clear improvement in SMD estimation compared with climatological PE, for six grassland sites in Britain. 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 with either more complex climatological means 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 PE, as input to their four rainfall-runoff models for 308 catchments, although they noted that arid and/or small catchments (PE \geq rainfall; area < 150 km²) generally gained more benefit from time-series than wet and/or large catchments.

PE, AE and hydrology in Britain

Monthly MORECS short grass PE is often used to provide hydrological model inputs in Britain. [Figure 1](#) shows maps of MORECS annual mean rainfall, PE and AE (for short grass and median AWC soils), and the ratios PE/rainfall and AE/PE, for 1961–1990. There is a south/east to north/west rainfall gradient, with an approximately reverse PE gradient. However, the latter does not carry through to AE, which decreases again in the south/east. The ratio PE/rainfall shows the dryness of eastern Britain and wetness (PE/rainfall $\ll 1$) of north/west Britain. The ratio AE/PE

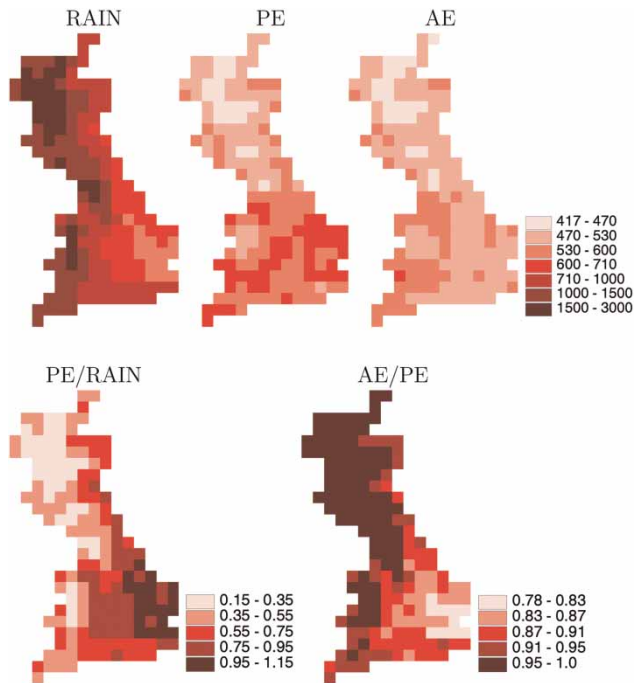


Figure 1 | Maps of MORECS annual mean rainfall, PE and AE (mm) for 1961–1990 (for short grass and median AWC soils), and the ratios PE/rainfall and AE/PE.

shows that AE in the north/west is generally energy-limited ($AE \sim PE$), with AE in the east generally water-limited ($AE < PE$). The dry/wet ‘boundary’ (where PE approaches rainfall) and water-limited/energy-limited ‘boundary’ (where

AE approaches PE) are related (e.g., Arora 2002), creating a transition region where the supply of water and energy to the land surface are in closer balance (hereby termed energy-water balanced).

The maps in Figure 1, using annual mean data, are only indicative; PE/rainfall and AE/PE patterns will vary month-to-month and year-to-year (e.g., AE in the north-west could also be water-limited in a hot, dry summer like that of 1976; see Rodda & Marsh (2011)). To illustrate the effect of seasonal variation of PE and rainfall on AE and runoff, Figure 2(a) shows 1961–1990 monthly means of these variables for three contrasting catchments (modelled with CLASSIC): wet/energy-limited, energy-water balanced and dry/water-limited. In summer, the wet/energy-limited catchment shows $PE < rainfall$ with near-constant AE close to PE, while the energy-water balanced catchment has PE close to rainfall and a slightly larger difference between PE and AE. In the dry/water-limited catchment $PE > rainfall$ for the middle part of the year, leading to higher SMDs and $AE < PE$, with reduced AE as summer progresses (peak AE precedes peak PE). A given PE change for a wet/energy-limited catchment is likely to result in a similar AE change, whereas the same PE change for a dry/water-limited catchment is likely to result in little AE change (assuming little/no rainfall change). Thus

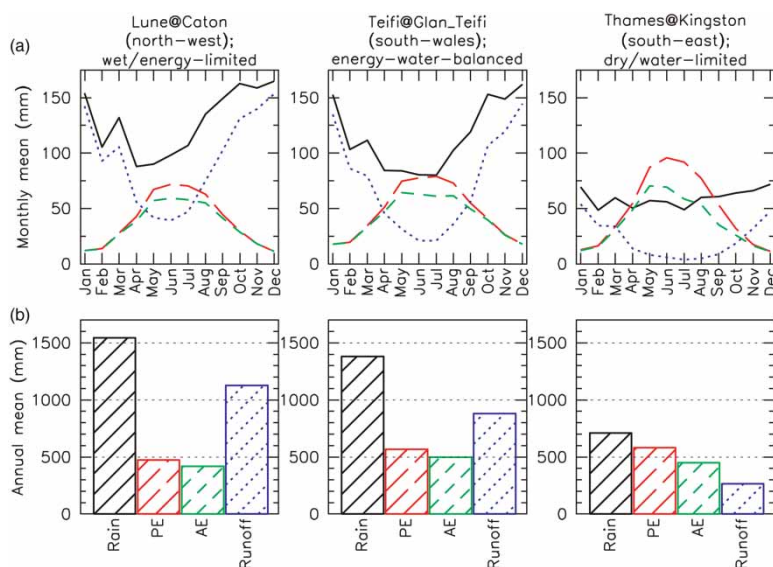


Figure 2 | Rainfall (solid line, black), PE (long-dashed line, red), AE (short-dashed line, green) and runoff (dotted line, blue), plotted as (a) monthly means; and (b) mean annual totals for 1961–1990 (mm), for three contrasting catchments modelled with CLASSIC. For references to colour see the online version of the paper: <http://www.iwaponline.com/jwcc/toc.htm>.

dry/water-limited catchments in the south/east 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. Catchments most sensitive to PE accuracy are those in the more energy-water balanced transition region, since PE forms a larger proportion of their rainfall and PE changes can more easily push them into a different regime, via their affect on soil moisture storage. Histograms of mean annual rainfall, PE, AE and runoff for the three example catchments (Figure 2(b)) show that the energy-water balanced catchment has the highest AE; summer rainfall in this catchment is able to sustain a higher AE rate than in the water-limited catchment. Similar conclusions on the sensitivity of energy-water balanced catchments to PE changes are reached by Wang & Alimohammadi (2012) from annual water balance analyses in 277 US catchments.

PE and environmental change

The situation becomes more complicated under climate change, as amounts and seasonal distribution of rainfall, as well as atmospheric demand for water (PE), are likely to change. PE changes can occur through changes in the meteorological controls, but also through changes in the transpiration controls. The relationship between climate change, meteorology and plant physiology is highly complex (Wullschleger *et al.* 2002). For example, higher concentrations of CO₂ in the atmosphere can lead to plant stomata opening less widely, giving greater surface resistance and less transpiration, but higher CO₂ concentrations can also enhance plant growth, leading to greater leaf area and more stomata, possibly counter-acting the effect of stomatal closure (Betts *et al.* 2007). Lower light levels, due to sunshine reductions or greater canopy shading, can also lead to stomatal closure (Wullschleger *et al.* 2002). Climate change impact studies generally consider only meteorological controls on PE, with transpiration controls kept constant. 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 CO₂-induced stomatal closure, suggesting the

importance of this mechanism of PE change for hydrological modelling under climate change.

A further complication is land-use change. Different land-covers have different PE rates, so large-scale changes in land-cover can influence AE, affecting runoff and river flows. Some hydrological models, e.g., CLASSIC, allow for different land-covers. Crooks & Davies (2001) used CLASSIC to investigate the sensitivity of flooding in the Thames to land-use change, showing that covering the catchment with 100% grass resulted in lower flood frequency than for 1990 land-use, with 100% trees resulting in even lower frequency. Similarly, Dunn & Mackay (1995) investigated the sensitivity of two sub-catchments of the River Tyne (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 hydrology in the highland catchment, but much more effect in the lowland catchment because its evaporation is a greater proportion of total rainfall.

Ideally, climate and land-use change would be considered together, as there are possible two-way feedbacks between them (e.g., Dirmeyer *et al.* 2010). At the local scale, eco-hydrological indicators, such as those of Hill *et al.* (2000), can be used to assess which species could appear/disappear following environmental change at a site. Feedbacks on evaporation, humidity, etc. could follow (e.g., Teuling *et al.* (2010) show how forests and grasslands differentially affect overlying air-temperature during droughts).

HISTORICAL TRENDS

PE and AE trends in Britain

There are few published analyses of historical evaporation trends in Britain, and most look at specific sites; none has national coverage. The longest used daily temperature and rainfall observations from Oxford for 1815–1996 (Burt & Shahgedanova 1998). Thornthwaite monthly PE was first estimated from temperature data, then seasonal regression equations were applied to correct this towards Penman PE. Monthly AE was then estimated, via SMDs. The results suggest PE increases and AE decreases, but with considerable annual and decadal variation (no trend quantification was provided). Crane & Hudson (1997) used daily data

from a site in Wales to calculate Penman PE for 1969–1995. They found no clear PE trend, although there were trends in some components, and concluded that local factors (e.g., tree-felling) have had a greater impact on climate variables and PE at this site than any larger-scale climatic changes.

Plots of trends in MORECS data include those of Rodda & Marsh (2011), who show water-year total PE and AE for 1961–2009 averaged over England and Wales, and Price & McKenna (2003), who show annual total PE for 1961–1993 averaged over Scotland. Each shows apparent increases, but with no quantification. Similarly, Yang *et al.* (2005) plot MORECS PE for 1961–1993, separately for each month, for a single MORECS grid-box in Surrey; increasing trends are apparent in most months (with no quantification). Figure 3 shows MORECS PE and AE for 1961–2012 (for short grass and median AWC soils) averaged over grid-boxes covering England and Wales, and Scotland, again showing increasing trends in each case. Fitted linear trend lines indicate very similar trends in PE and AE over Scotland (~0.6 mm/year), whereas over England and Wales trends in AE are slightly higher (~0.7 mm/year) and those in PE are higher still (~1.0 mm/year); all trends are significant at the 1% level ($p < 0.01$), based on permutation tests. An analysis of UK AE modelled by JULES also shows a slow upward trend over 1971–2007 (Blyth pers. comm.). Much more analysis is required, however, to investigate spatial and seasonal variation and consistency between modelling systems.

Unfortunately, neither PE nor AE is included in the UK climate trend report of Jenkins *et al.* (2007), based on 5 ×

5 km grids of numerous climate variables. Although there are several methods for measuring AE, including eddy-covariance, lysimeters and large aperture scintillometers (e.g., Shuttleworth 2007), there are no long-term AE observations in Britain. Any apparent trends in relatively short records should be treated with caution, as they may be due to natural climatic variations rather than climate change (Robson 2002); a problem potentially exacerbated by particularly wet/dry periods towards the start/end of the record (Hannaford & Buys 2012).

Global context

The shortage of information on evaporation trends is not limited to Britain; the Intergovernmental Panel on Climate Change concludes ‘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 several regions of the globe over the last five decades (Fu *et al.* 2009); known as the ‘pan evaporation paradox’ since it has occurred despite temperature increases (Shuttleworth *et al.* 2009). The global review of McVicar *et al.* (2012) suggests that declining evaporation is primarily due to declining wind speeds, implying the importance of including wind when estimating PE.

A global study finds large basins with positive (e.g., Niger), negative (e.g., Amazon) and non-significant (e.g., Congo) PE trends over 1958–2001 (Weedon *et al.* 2011), and both increases and decreases have been found in

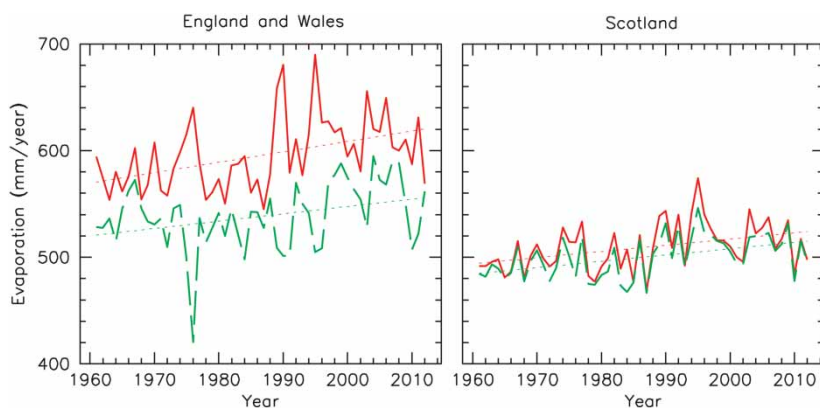


Figure 3 | MORECS PE (solid line, red) and AE (dashed line, green) for 1961–2012 (for short grass and median AWC soils), averaged over England and Wales (left) and Scotland (right). In each case a fitted linear trend line is shown (dotted). For references to colour see the online version of the paper: <http://www.iwaponline.com/jwc/toc.htm>.

China (Thomas 2000). Analysis of open-water PE for the Canadian Prairies suggests that decreasing wind speeds are the main factor decreasing PE in the south, while vapour pressure deficits are the main factor increasing PE in the north (Burn & Hesch 2007). Irmak *et al.* (2012) show a significant decrease in grass reference PE in the Platte basin, USA, but with no significant change in wind speeds, and suggest that increasing rainfall and related decreasing short-wave radiation are instead responsible.

Jung *et al.* (2010) examine data-driven estimates of global land AE, which show a (highly significant) increase for 1982–1997 but a (less significant) decrease for 1998–2008. Their correlation analyses suggest the recent decline is due to increasing soil moisture limitation, but the time periods studied are too short to imply trends; differences could be due to decadal variability.

FUTURE PROJECTIONS

Issues with PE estimation under climate change

Projections of PE from vegetated surfaces are generally not produced directly by Global or Regional Climate Models (GCMs or RCMs), so have to be made offline. But which PE formulae are likely to perform best when using climate model data, as opposed to observed weather data? One consideration is whether the climate model PE estimates will be used directly, so the absolute values need to be reasonable, or whether they are simply used to estimate PE changes applied to baseline observed PE using the change factor method (Kay & Jones 2012). Note that an alternative way of estimating future PE is via statistical downscaling; fitting a regression relationship between large-scale weather data and catchment PE data, then applying it using GCM data to estimate future catchment PE (e.g., Wilby & Harris 2006). Similarly, Chun *et al.* (2012) develop a generalised linear model (GLM) for PE. These methods are likely to have many of the same issues as estimating PE using climate model variables in more standard PE formulae.

The use of more complex formulae with climate model data is not necessarily straightforward, since some variables may be less reliable (Vorosmarty *et al.* 1998; Kingston *et al.* 2009), or simply not available from all climate models. For

example, the probabilistic UKCP09 projections (Murphy *et al.* 2009) did not initially include wind speed, as wind data were not available from all the GCMs used in the statistical methodology (Sexton & Murphy 2010). Fisher *et al.* (2011) discuss issues with input data uncertainty and sensitivity in PE formulae. A related issue is that climate model data may have different inter-variable correlations than observed data (Chun *et al.* 2012).

As simpler, empirical formulae rely on fixed relationships between atmospheric variables and PE, these may not hold under extrapolation to very different future climates (Shaw & Riha 2011), although some level of extrapolation may be acceptable where formulae perform well for a wide range of locations and climatic conditions (Sperna Weiland *et al.* 2012). Irmak *et al.* (2012) suggest that, as many climate variables affecting PE have been changing and are expected to change in future, single-variable PE formulae should be avoided when estimating PE (historical or future) trends ‘due to the inherent nature of the trend passed to PE from the variable’. A view echoed by Donohue *et al.* (2010) as ‘the greater the number of the four key variables ... in a formulation, the more realistic the trends from that formulation become’, although this clearly depends on the sensitivity of a formula to each variable and the relative strengths and directions of any trends in those variables (likely to have some level of inter-dependency; Fisher *et al.* 2011).

The sensitivity of PE formulae cannot necessarily be predicted directly from the meteorological variables included; Bormann (2011) compared PE changes given by 18 formulae, for six locations in Germany, and found as much variability for formulae of the same type as for different types. However, Shaw & Riha (2011) show that the temperature-only Hamon and Thornthwaite formulae are much more temperature-sensitive than the more complex Penman–Monteith and Priestley–Taylor formulae. This should be interpreted carefully, since changes in other variables were neglected but are likely to co-vary with temperature (Fisher *et al.* 2011; Chun *et al.* 2012), and their inclusion could increase the response of more complex formulae. Interestingly, Shaw & Riha (2011) show that the temperature-sensitivity of the Oudin formula is mid-way between that of the other two pairs of formulae. Oudin PE is temperature-based but also includes extraterrestrial radiation (dependent only on Julian

day and latitude), so has some dependence on radiation even though its only true variable is temperature.

Shaw & Riha (2011) also show that radiation, rather than temperature, is the main driver of PE in broadleaf forests in the USA. They suggest that the strong relationship between radiation and temperature for longer averaging-periods (weekly and above) 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 including radiation could also work; even those that only include extraterrestrial radiation (e.g., Oudin or Hargreaves), as long as they are applied over longer averaging-periods. Hargreaves PE (Hargreaves & Samani 1985) is an empirical formula using temperature, extraterrestrial radiation and diurnal temperature range (a humidity proxy).

Sperna Weiland *et al.* (2012) compared Penman–Monteith, Priestley–Taylor, Hargreaves and Blaney–Criddle formulae for estimating global reference PE using daily reanalysis data. Blaney–Criddle provided the best match to monthly observation-based Penman–Monteith PE, but they concluded that its need for cell-specific calibration, with large spatial variation in calibrated coefficients, meant it was unsuitable for climate change applications. They also concluded that Penman–Monteith was not ideal, since it did not outperform other methods, has a high data demand and is sensitive to data accuracy. 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.

Kay & Davies (2008) calculated Penman–Monteith and Oudin PE over Britain, using monthly data from 13 climate models for 1961–1990, and compared them with MORECS PE. Perhaps surprisingly, Oudin PE matched better than Penman–Monteith PE; the latter was lower than MORECS for most months 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 Penman–Monteith. 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.

Sometimes, bias-correction is applied to allow for perceived biases in climate model data (Piani *et al.* 2010). However, it should be remembered that climate model variables are expected to be internally consistent; applying bias-correction separately to each variable, or only to some variables, could introduce inconsistency (e.g., Fischer & Knutti 2013) and potentially cause incorrect PE. A possible alternative is direct bias-correction of PE. This was attempted by Ekstrom *et al.* (2007), but the resulting PE time-series was still considered 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 climate model bias, rather than multi-decadal natural variability for example (e.g., Deser *et al.* 2012), and that the same bias applies for future periods.

PE projections for Britain

Table 1 summarises the PE changes for Britain presented in several references, described in more detail below.

Arnell (2011) uses data from 21 GCMs, for six catchments in Britain, for a scenario representing a 2 °C rise in global mean temperature. Annual PE (calculated using Penman–Monteith but with fixed wind speeds) increases for all but one climate model, with significant variation between climate models and catchments (range –4–40%). 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 with A2 emissions, with a weather generator for the River Eden (Cumbria). They show increases in mean Penman–Monteith PE in all seasons, with the largest percentage increases in autumn (30–80%) and winter (30–60%) and smallest in spring (0–50%) and summer (20–40%). Similarly, Kay & Davies (2008) use data from 13 climate models (2080s A2) to calculate changes in Penman–Monteith and Oudin 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. Increases in annual Oudin PE are

Table 1 | Summary of PE projections for Britain

Reference	Spatial coverage	Climate data; scenario (e.g., time-horizon; emissions)	PE formula(e)	Projected PE changes
Arnell (2011)	6 British catchments	21 GCMs; 2 °C global mean <i>T</i> rise	Penman–Monteith (wind speed fixed)	Annual increase for all but 1 climate model
Fowler <i>et al.</i> (2008)	River Eden (Cumbria)	13 RCMs; 2080s; A2	Penman–Monteith	Increase in all seasons
Kay & Davies (2008)	Great Britain: northern and southern averages	5 GCMs, 8 RCMs; 2080s; A2	Oudin; Penman–Monteith	Increase in all months; Increase in most months, but decrease in some months for some climate models
Ekstrom <i>et al.</i> (2007)	Europe	1 RCM; 2080s; A2	Penman–Monteith	Increase in all seasons
Cameron (2006)	River Lossie (Scotland)	1 RCM; 2080s; A1F1, A2, B2, B1	Thornthwaite	Increase in all months
Wilby & Harris (2006)	Thames (SE England)	4 GCMs; 2020s, 50s, 80s; A2, B2	Statistical downscaling	Increase in winter and summer
Wilby <i>et al.</i> (2006)	Kennet (SE England)	3 GCMs; 2020s, 50s, 80s; A2, B2	Statistical downscaling	Increase in winter and summer
Diaz-Nieto & Wilby (2005)	Thames (SE England)	1 GCM/RCM; 2020s, 50s, 80s; A2, B2	Statistical downscaling; Penman–Monteith	Increase in all months except Dec 2020s; Increase, often greater than above
Chun <i>et al.</i> (2012)	25 sites in Great Britain	1 GCM; 2080s; A2	GLM; Penman–Monteith	Annual increase almost everywhere; Annual increase almost everywhere, greater than above
Christierson <i>et al.</i> (2012)	UK (plus 2 example catchments)	UKCP09; 2020s; A1B	Oudin	Increase in ensemble median all months; for example catchments, decrease in some months for some ensemble members
Kay & Jones (2012)	9 British catchments	UKCP09; 2080s; A1B	Oudin	As above
Prudhomme & Williamson (2013)	Great Britain	1 RCM; 2050s; A1B	Blaney–Criddle, Hamon, Oudin, Thornthwaite, McGuinness–Bordne, Makkink, Jensen–Haise, 2 Priestley–Taylor, Turc, 2 Penman–Monteith	Increase in month representing each of 4 seasons, everywhere, for all formulae except Priestley–Taylor
Kingston <i>et al.</i> (2009)	Globe, latitudinally-averaged	5 GCMs; 2 °C global mean <i>T</i> rise	Blaney–Criddle, Hamon, Hargreaves, Jensen–Haise, Priestley–Taylor, Penman–Monteith	Annual increase in all cases

~12–34%, whereas those in Penman–Monteith PE are ~6–56%; increases are generally lower for Penman–Monteith than Oudin in the north, but generally higher in the south.

Ekstrom *et al.* (2007) use data from the HadRM3H RCM over Europe (2080s A2) and show seasonal absolute differences between future and baseline Penman–Monteith PE. Differences are largest in summer (4–8 mm/day over large

parts of Europe; less over Britain) and smallest in winter (<0.5 mm/day for most of Europe). However, for hydrological modelling of a catchment in north-west England they used Blaney–Criddle instead of Penman–Monteith PE, as the latter gave a spread of daily values too large compared with observations, and percentage increases that were too high (up to 80%) in summer (Blaney–Criddle gave a

summer increase <20%). Based on this, Cameron (2006) decided to apply Thornthwaite, rather than Penman–Monteith, to estimate PE changes in the Lossie catchment (Scotland), obtaining changes ranging from 6.2% (June) to 20.8% (January) for the 2080s with A2 emissions.

Wilby & Harris (2006) use data from four GCMs (2080s A2 and B2) with statistical downscaling for the Thames. They obtain PE increases of 5–43% in winter and 11–22% in summer, but for two GCMs, the summer changes are larger than winter, with the opposite for the other two GCMs. Wilby *et al.* (2006) use a similar method for the Kennet, using three of the same GCMs, but find much smaller PE increases (winter 3–9%; summer 5–16%) presumably due to a different regression model. Diaz-Nieto & 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 Penman–Monteith. Chun *et al.* (2012) use data from the HadCM3 GCM (1950–2099 A2) to estimate PE at 25 sites across Britain using Penman–Monteith and a GLM with four variables (radiation, temperature, wind and humidity). They show annual percentage increases almost everywhere by the 2080s, with greater increases in the south/east than north/west, but smaller increases using the GLM than 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.* 2012).

Christierson *et al.* (2012) use UKCP09 probabilistic temperature data (2020s A1B) to calculate Oudin PE percentage changes for each river-basin region in the UK. They present maps of the central estimate of PE changes, which show an increase across the country, with the largest changes in winter. The range of monthly PE changes, for two example catchments, almost always show increases, usually 0–30%, with a greater variation in winter than summer (but only a 20-member subset of the full 10,000 UKCP09 set is shown, selected via latin-hypercube sampling). These PE changes are shown to be generally consistent with an older set of scenarios, and with those calculated using temperature data from the 11-member UKCP09 RCM ensemble in the Oudin formula. Similarly, Kay & Jones (2012) use UKCP09 probabilistic temperature data (2080s A1B) to calculate Oudin PE changes for nine catchments in Britain. Again,

the PE changes are almost always positive, and generally larger (with greater variation) in winter than summer, but the range of 5–60% is larger than that of Christierson *et al.* (2012), because of the later time-horizon. Prudhomme & Williamson (2013) use data from one member of UKCP09 RCM ensemble (2050s A1B) and compare PE changes across Britain for 12 formulae for a month representative of each of the four seasons. PE almost always shows an increase, although the magnitude varies by season, location and formula.

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

PE projections with changes in transpiration controls

Bell *et al.* (2011) developed a method to estimate Penman–Monteith PE using surface resistance (r_c) values produced by an RCM with an embedded land-surface scheme, so allowing r_c to vary with the level of CO₂ in the atmosphere. Comparisons over Britain, using data from the HadRM3 RCM for 1961–1990, showed that using RCM r_c gave PE comparable with MORECS, but generally lower than when using MORECS' 12 fixed monthly r_c values, particularly in spring. Similarly, for 2070–2099, using RCM r_c gave lower PE than when using fixed r_c . Looking at percentage PE changes between the two periods, using fixed r_c gave increases of 15–34%, whereas using RCM r_c gave significantly lower increases (3–7%). Thus r_c changes could be an important mechanism for limiting PE change. However, not all possible feedbacks are included in this RCM run (e.g., changes to leaf area or land cover); PE changes may be somewhere in between if other mechanisms are included.

Kay & Jones (2012) used the method above with data from the UKCP09 11-member RCM ensemble (with

corresponding r_c time-series) and showed that the corresponding percentage PE changes were more variable than those derived using UKCP09 probabilistic temperature data to calculate Oudin PE changes. The biggest difference occurred in winter and in spring, where varying r_c PE changes were often lower than Oudin PE changes. However, even large percentage PE changes are not really significant in winter in Britain, since baseline PE is very small. Differences in spring could be more important.

Moratiel *et al.* (2011) investigate possible changes in grass reference PE in part of Spain, including the effect of r_c changes due to CO₂ concentrations rising from 372 to 550 ppm 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 calculate Penman–Monteith PE both with current r_c (70 s/m) and an approximate future r_c assuming stomatal closure (87 s/m). They find PE increases of about 10% with current r_c but only about 5% with future r_c . This is similar to Bell *et al.* (2011), although there the differences are larger, perhaps because of the later time-horizon. Islam *et al.* (2012) also find much smaller PE increases (or decreases) for the US Great Plains when the effect of elevated CO₂ on stomatal resistance is included.

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 suggests that accurate PE projections will not always be important. Sperna Weiland *et al.* (2012), using reanalysis data for 1979–2002, show a general reduction in variability of results between six different PE formulae when moving down their global hydrological modelling chain, from PE to AE to runoff to river flows. They thus suggest ‘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 (not Britain).

Kay & 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. The results suggest that the extra uncertainty is greatest for changes in low to median flows, although possibly still important for high flows in some catchments, but that climate model uncertainty dominates. Arnell (1999) tested the effect of using Priestly–Taylor in place of Penman–Monteith PE, when modelling changes in European runoff under four scenarios for the 2050s. As Priestly–Taylor PE increases were generally smaller than those for Penman–Monteith, the pattern of runoff changes followed that of rainfall changes more closely. However, PE uncertainty was generally less than scenario uncertainty.

Kingston *et al.* (2009) assessed the effect of six PE formulae on the global extent of arid (rainfall/PE < 1) and humid (rainfall/PE ≥ 1) regions, with five GCMs for a 2 °C rise in global mean temperature. They showed that, although almost all GCMs and PE formulae agreed on an increased arid area (and corresponding decreased humid area), they disagreed on the amount of change; eastern England verges on aridity (Figure 1) so this uncertainty could apply. Furthermore, regional analysis of water surpluses (annual rainfall minus PE, for months where rainfall > PE) suggests that PE uncertainty is of comparable magnitude to GCM uncertainty, but the analysis only covers the Mediterranean, East Africa and Southeast Asia (not Britain). Sheffield *et al.* (2012) showed that the global drought severity trend is exaggerated when modelled using Thornthwaite PE rather than Penman–Monteith PE, while Milly & Dunne (2011) showed that using Jensen–Haise PE for 10 catchments in the USA gave smaller runoff changes than suggested by the driving climate models directly.

Betts *et al.* (2007) modelled change in continental runoff under a doubling of CO₂, and showed that including the effect of stomatal closure led to larger runoff increases (17 ± 5%) than including only radiative forcing (11 ± 6%). The difference was slightly less when changes in land cover and leaf area were also included. Similar effects were seen for percentage changes in flood peaks in the Thames, using data from the UKCP09 11-member RCM perturbed-physics ensemble to drive the G2G hydrological model (Bell *et al.* 2012); changes were higher when averaged over the six members which included stomatal closure, and lower when averaged over the five members which did not. However, simultaneous variation of other parameters in the

perturbed-physics ensemble means that this should be interpreted with caution.

DISCUSSION AND CONCLUSIONS

The aim of this paper was to present a review of historical trends and future projections for evaporation in Britain, from a hydrological perspective. Evaporation is generally estimated from meteorological data, using formulae of varying complexity. The UK Met Office's MORECS and MOSES systems apply the physically based Penman–Monteith formula to estimate PE from meteorological observations over Britain. These PE are often used, with rainfall, as inputs for hydrological modelling. While hydrological models may appear to 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 parameter instability. Although PE is less variable than rainfall, with a well defined seasonal pattern, AE accounts for loss of a significant proportion of rainfall across much of south/east Britain. These factors are important considerations when hydrological models are applied, especially when simulating impacts of environmental change.

There is relatively little information on historical evaporation trends in Britain, or indeed globally, whether actual or potential. The few British studies generally indicate PE increases, and some also show AE increases. Globally, both increases and decreases in PE have been detected in different areas, with different causes, although decreasing wind speeds tend to be implicated in PE decreases. A more recent global decline in data-driven AE estimates has been attributed to decreasing soil moisture. Direct AE measurement is progressing from localised observations to remotely sensed estimates from large areas. Combined analyses of measurements, satellite data, observed changes in meteorological variables affecting evaporation, and observed changes in river flows should, together with further modelling, enable a better understanding of the direction, magnitude and causes of AE or PE change (e.g., Donohue *et al.* 2010).

There is little consensus on the best approach for deriving future PE projections; some authors believe that formulae including all meteorological variables influencing PE must be applied, while others believe that sensitivity of such

formulae to data quality makes this inadvisable, or that the choice will make little 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 CO₂ concentrations (stomatal closure, increased plant growth, etc.).

Most studies presenting PE projections for Britain indicate increases in annual PE, although some studies suggest (usually small) PE decreases in some months (Table 1). However, there is considerable variation in the magnitude of projections, caused not just by the PE formula applied but by the climate model, emissions scenario and time-horizon, and by location. These factors make it difficult to compare PE projections between studies, as do the varied ways of describing changes (i.e., percentage or absolute; monthly, seasonal or annual).

It has been suggested that the choice of PE formula could be particularly important in regions where precipitation and PE are in close balance (e.g., Kingston *et al.* 2009). The maps in Figure 1, based on MORECS data, show that this is likely to include parts of Britain, 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 annual climatic water budget (annual precipitation minus annual PE) between baseline and future climates, to test the sensitivity of a given location to choice of PE formula; the choice is probably not crucial if changes are relatively consistent between formulae, but needs greater consideration if there is more variation. However, likely changes in the seasonality of 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 to the hydrological modelling, as the importance of PE is likely to vary according to the hydrological aspect under investigation as well as catchment location (e.g., Kay & Davies 2008). Nevertheless, several studies have shown that uncertainty due to climate model structure is greater than that due to PE formulation, implying that, where there is limited capacity for hydrological model runs, the priority should be to cover climate model uncertainty more comprehensively.

The question remains: which PE formulae are likely to be more (or less) reliable when applied with climate model data? For each region of the globe, the answer requires investigation of: (1) which meteorological variables are most important for PE changes (i.e., the necessary level of complexity), and (2) the reliability of each of these meteorological variables when taken from different climate models. Together, these could enable the derivation of improved PE projections, and so improved hydrological projections. Further investigation is also needed into feedbacks between climate change and plant transpiration and growth. Clarification of the best ways to measure/model evaporation would also be useful for detection and attribution, as evaporation trends are easier to detect and attribute than precipitation or runoff trends (Ziegler *et al.* 2003; Douville *et al.* 2012).

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