Response of potential evaporation to climate variability and change: what GCMs simulate

Wenju Cai and Roger Jones,
CSIRO Atmospheric Research
PB No 1, Aspendale, Victoria 3195, Australia
Email: Wenju.cai@csiro.au ; roger.jones@csiro.au

Potential evaporation (E*p*) as simulated by CSIRO climate model is presented in two forms. The first form of E*p* is calculated at the soil surface, produced as a diagnostic variable within the models and not used to calculate any other factors. The second form of E*p* has been calculated by Morton's (1983) Complementary Relationship Areal Evaporation (CRAE) method using other climate variables produced by the model. This type of output has been used to construct projections of regional change in E*p* and has also been extensively used in hydrological modelling.

Directly calculated E*p*

We present simulated E*p* from a control experiment in which no climate change forcing is included, and compare it with E*p* in a global change experiment in which forcing of a increasing CO_2, ozone, and aerosols are included following the Special Report on Emission Scenarios (SRES) A2. It follows the observed evolution of the atmosphere from 1870 to 2000 and projected CO_2-e levels from 2001-2100. By 2100, the equivalent CO_2 reaches a level that is more than three times the level of 1870 (concentration ppm). Thereafter, both the CO_2 and aerosol levels are held constant and the model integrated for another 150 years. The effect of aerosols is treated in a most rudimentary manner (see Rotstayn's Abstract), and the results may be modified somewhat if the more sophisticated treatment of aerosols is implemented.

In the CSIRO climate model, E*p* is calculated from:

$$\lambda E_p = \rho\lambda[q^*(T_o) - q]/r_{aV}$$

where λ is latent heat of vaporization of water, ρ is air density, $q^*(T_0)$ is the saturation specific humidity at surface temperature (T_0), q is the specific humidity of the lowest model level, and r_{aV} is the aerodynamic resistance. T_0 changes with each time step. This first-principle flux-gradient parameterization is energy balanced and is popularly used in most climate models. It can be shown to be similar to Penman-Monteith formulation, except that it uses actual surface temperature rather than a wet equilibrium temperature.

Results from the control experiment show that the seasonal climatology of E*p* is well simulated. Further, the model relationship between potential evaporation anomalies and anomalies of various meteorological parameters is realistic, and is consistent with what is expected from the Penman formulation, including:

- E*p* increases with increasing solar radiation,
- E*p* increases when the actual evaporation decreases (the complementary relationship),

- E*p* decreases when rainfall and cloud increase, i.e., the model physics are potentially capable of simulating the effect of global dimming, if dimming per se is simulated.

In response to global warming due to increasing atmospheric CO_2:

- Global mean E*p* increases,
- Global mean cloud cover decreases,
- Global mean downward solar radiation decreases despite the decrease in cloud cover (because of increased cloud water content, leading to increased cloud albedo),
- Global mean rainfall increases despite the decrease in cloud cover. Global mean actual evaporation increases, consistent with an increase in global mean rainfall,
- Global mean diurnal temperature range (DTR) decreases,
- Global mean surface energy fluxes increase (hence E*p* increases).

On regional scales, however, the trends of DTR can be very different; for example, outputs from a model grid near Victoria show that the model potential evaporation increase is accompanied by an increase in DTR.

Morton's CRAE E*p*

Morton's CRAE model utilises mean daily temperature, mean daily moisture content of the air and total downward shortwave radiation or sunshine ratio averaged over a period of greater than 5 days. It is essentially an energy-balance method following Bouchet's (1963) complementary relationship.

The relationship states that over areas of a regional scale and away from any sharp environmental discontinuities, there exists a complementary feedback mechanism between actual and potential evaporation. Energy at the surface that, because of limited water availability, is not taken up in the process of actual evaporation (E*a*) increases the temperature and humidity gradients of the overpassing air and leads to an increase in E*p.* This relationship is described by:

$Ea + Ep = 2Ew$.

Under conditions where E*a* equals E*p*, this rate is referred to as the wet-environment areal evaporation (E*w*). E*p* is sometimes called point potential evaporation to distinguish it from areal evaporation. Morton et al. (1986) consider E*p* to be comparable to Penman's (1948) equation and E*w* to Priestly and Taylor's (1972) equation.

For the past seven years, we have been using monthly values of air temperature at screen height, humidity (or dewpoint temperature) and downward shortwave radiation as simulated by climate models to produce potential evaporation for use in hydrological and crop impact modelling. E*p* was at first used because this was used to scale A Class pan data in hydrologic impact models but now we produce both E*p* and E*w*. E*w* is considered preferable for catchment scale hydrologic modelling (BoM, 2001).

The method applied is the same as Morton's formulation as used in the Bureau of Meteorology Atlas for Potential Evaporation (BoM, 2001), allowing the comparison of 1961–90 potential evaporation climatologies over Australia. In general, for Australia, most climate models (e.g. CSIRO, UK Hadley Centre, German Max Planck/DKRZ, the Canadian Climate Centre and US NCAR and GFDL models), produce a relationship between rainfall and potential evaporation that reproduces the historical relationship but shows less variability,

Walsh et al. (2000) describe a detailed error analysis of potential evaporation for the DARLAM 125 model over Australia, nested in the Mark2 model. This analysis was carried out using a 0.5°C climatology of E*p* derived from a global land climatology of temperature, humidity and global solar radiation described by New et al. (2000). This data set was used in preference to the Bureau's E*p* because it was generated in a grid of similar scale; both the model and the New et al. (2000) climatology are unable to reproduce the highly detailed spatial distribution of E*p* in the mountainous regions of eastern Australia.

The analysis showed that the spatial correlation between monthly (Jan–Dec) DARLAM and the BoM E*p* climatology ranges from 0.82 and 0.97 for E*p* and 0.57 and 0.97 for E*w*. Cool season correlations are very high (>0.9), correlations being lowest in summer. The conclusions of the error analysis are as follows:

Although all of the inputs from DARLAM are overestimated over much of Australia, these overestimates are partially compensated by underestimated vapour pressure deficit and overestimated radiation. Although both temperature and vapour pressure in DARLAM are too high, vapour pressure deficit, the difference between saturated and actual vapour pressure, is less in DARLAM compared to the vapour pressure from the CRU data set. As shown in the previous section, both Ep and Ew are overestimated over most of Australia, except over northwestern and southeastern Australia in the summer and winter respectively. However, these differences are usually under 25% over most areas in most months. This is a good result, as operational estimates of pan evaporation often attain this level of error (Walsh et al., 2000; McMahon, pers. comm.).

E*p* and E*w* output from a range of models has been used to produce patterns of climate change for calculating regional projections and for use in constructing climate scenarios. These patterns are calculated by regressing the change in both E*p* and E*w* for each month against mean global warming to create a standardised pattern of change in percent. A positive change denotes an increase, and a negative change, a decrease. The rationale behind calculating patterns of climate change is that it represents the global warming signal for each variable and location, limiting the model-dependent errors and uncertainties to those representing the simulation of climate change.

Over Australia, an increasing trend in annual E*p* and E*w* is produced in all climate models examined (Figure 1), although decreases are simulated in some other parts of the world. This is the source of the conclusion articulated in CSIRO (2001) that potential evaporation is likely to increase by 0 to 8% per degree of global warming. Increases in E*p* also contributed to projected changes in atmospheric moisture demand as expressed by E*p* – P (precipitation) (CSIRO, 2001).

Figure 1 shows change vectors for E*p* and P for a grid square centred over the upper Macquarie River catchment in NSW from six climate models, showing that Ep increases even with increases in rainfall. Changes in P and E*p* are codependent (r^2 of about 0.4), whereas changes in P and E*w* are only very weakly codependent.

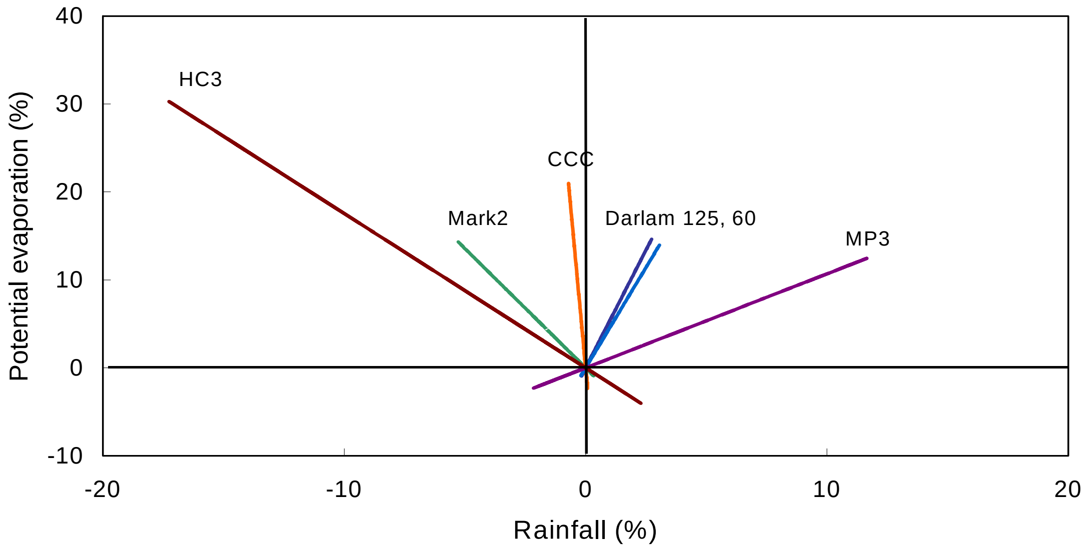

Figure 1: Trend of Ep in several climate models including the UK (HC3) , Germany (MP3), CSIRO Mark 2, Canadian (CCC), and CSIRO regional models (Darlam 125, 60) for the upper Macquarie catchment, NSW.

This codependence can also be investigated in terms of the transition between a stationary (proxy historical) and non-stationary climate. Figure 2 is in two parts. Figure 2a shows the mean change for both P and E*p* from the MP3 model (Max Planck ECHAM4/OPYC3 greenhouse gas only IS92a simulation; IPCC-DDC) shown with a 29-year running mean. Figure 2b shows the same data as an annual time series with a 29-year running mean, showing that periods of potential evaporation decrease are not necessarily incompatible with an overall trend towards long-term increase.

These changes are consistent with how the complementary relationship may be expected to behave under climate change. Figure 3 (solid lines) plots the relationship in the context of rainfall variations under a stationary climate. When rainfall increases, E*a* increases but E*p* decreases, consistent with what the model results described above. Under a warmed climate, because of the availability of more energy at surface (as simulated in the model), E*w* increases (dashed lines, Figure 3). The above complementary relationship still holds, but if rainfall increases from R to R', E*a* increases (as simulated in climate models), E*p* may increase or decrease. For E*p* to decrease, a threshold value of rainfall (R_{th}) increase has to be passed. Given that all models discussed above show an increase in E*p*, it suggests that this threshold value has not been reached in these models. But if rainfall decreases, E*p* will increase.

On the other hand, if the climate change leads to a decrease in surface net flux, as has been simulated for the climate of Western Europe over the past 50 years when both direct and indirect effects of aerosols are properly incorporated into a climate model (Liepert et al., 2004), then E*a*, E*w* and E*p* will decrease regardless of rainfall response. Rather than seeing the whole complementary relationship lift, as in Figure 2, it will lower under such conditions.

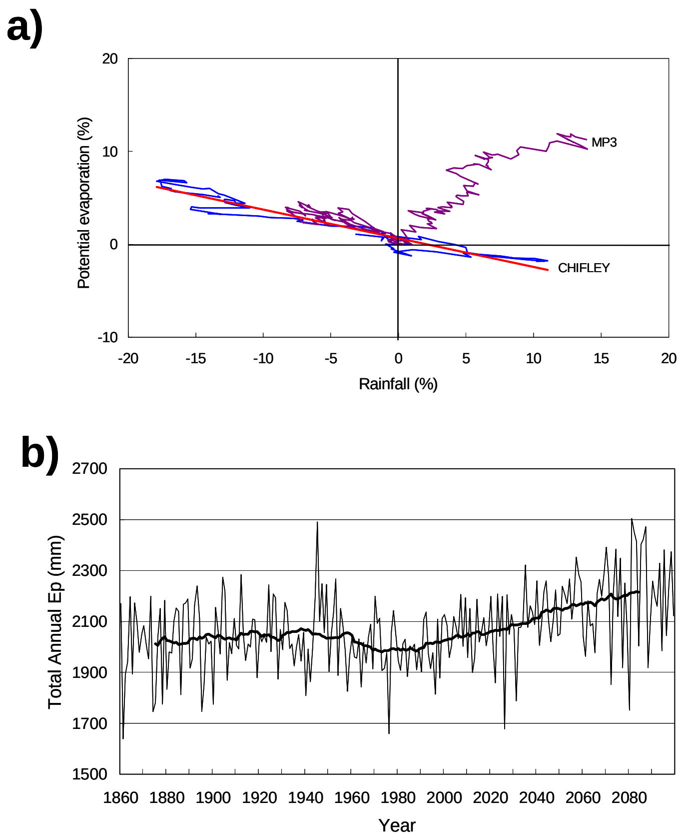

Figure 2a: Trend of P and Ep in the German MP3 model expressed as a 29-year running mean for the upper Macquarie catchment in NSW compared to historical data, showing the departure from stationary climate; 2b: Ep from 2a shown as a time series with a 29-year running mean showing a decrease in the second half of the 20th century followed by a subsequent long-term increase.

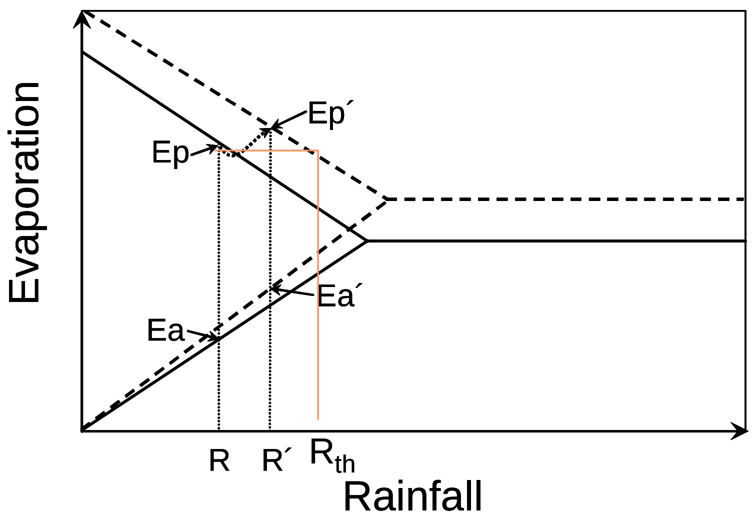

Figure 3: Schematic showing the complementary relationship between Ea, Ep and rainfall under a control control climate. When Ea increases, Ep decreases and the sum of Ea and Ep is approximately a constant (2Ew). Under a warmed climate if the heat flux at surface increases Ew increases, leading to an increase in Ep if R' is less than a threshold value (R_{th}).

Reconciling climate model output and historical trends

All the substantial work conducted by CSIRO Atmospheric Research on model-derived E*p* to date has compared model climatologies with observed climatologies using models of potential evaporation (e.g. BoM, 2001). Direct comparison between the climate model and the instrumental record is difficult, because the instrumental record of potential evaporation is restricted to the A-Class pan. Before the publication of Roderick and Farquar's (2004) on trends of Australian pan evaporation, there were several long-term records available: a long-term record of lake evaporation and E*p* derived from Morton's model for south-western Victoria from 1859–1990 reconstructed from a composite record free of observational inhomogeneities after 1863 (Jones et al., 2001); a combined record of Melbourne sunken tank and Wurdee Boluc reservoir pan evaporation from 1920–1990 (Nathan et al., 1984; Bruce, 1990) and a combined record of lysimeter, pan and model-derived evaporation from Griffith (Khan, pers. comm.). All these records show increases over the 20th century.

Three areas of reconciliation between different methods of deriving potential evaporation and the different types of potential evaporation need to be carried out:

1. Link both methods of estimating E*p* detailed in this abstract with other methods and reconcile their outcomes in climate model simulations,
2. Reconcile the differences between the different types of evaporation derived from instrumental records (e.g. A-Class pan, Morton, Penman), and
3. Combine these two efforts to understand the relationship between the observed and simulated changes over time.

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