Hydro-climatological variability in the Murray-Darling Basin

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Abstract Investigations into the recent drought in the Murray-Darling Basin have brought to light confusion surrounding the cause and effect of temperatures, potential evaporation and actual evaporation. In this study, a simple coupled land surface–planetary boundary layer model is used to illustrate the role of soil moisture in controlling evaporation and temperature, and to explore the interaction between potential and actual evaporation and temperatures under drought conditions. We demonstrate that increased temperatures during drought conditions are a result of the reduced soil moisture and actual evaporation. It is also shown that potential evaporation is increased under drought conditions as a result of increased atmospheric moisture demand, which is itself due to the decreased actual evaporation.

Key words evaporation; temperature; drought; soil moisture; Murray-Darling Basin

INTRODUCTION

Investigations into the recent drought in the Murray-Darling Basin (MDB) have brought to light confusion surrounding the cause and effect of temperatures and evaporation. These studies have noted that during the recent drought low rainfall totals have been accompanied by anomalously high air temperatures. In particular, Karoly *et al*. (2003) noted that whilst monthly rainfall totals were at extreme lows during the 2002 drought, the monthly average maximum temperatures were much higher than in previous droughts. This led the authors to state that "*…the higher temperatures caused a marked increase in evaporation rates, which sped up the loss of soil moisture and the drying of vegetation and watercourses. This is the first drought in Australia where the impact of human-induced global warming can be clearly observed…*".

Similarly, Nicholls (2004) investigated the anomalously high air temperatures that occurred during the 2002 cool season (May–October) in the MDB. This was achieved through a comparison to an identified negative correlation between average monthly temperature and average monthly rainfall, between 1952 and 2002. Nicholls (2004) then examined the residual time series of the correlation; it demonstrates a statistically significant monotonic increase toward higher air temperatures over the period of the regression data. It was then speculated that this was due to the increasing trend in atmospheric carbon dioxide and other greenhouse gases, and that "*the warming has meant that the severity and impacts of the most recent drought have been exacerbated by enhanced evaporation and evapotranspiration*".

In a more recent study Cai & Cowan (2008) suggest that increased temperatures are the cause of reduced inflows into the MDB since 1950. They show that a rise of 1°C leads to an approximate 15% reduction in annual inflows. Similarly Cai *et al*. (2009) speculate that increased temperatures have led to decreased soil moisture, and that annually a rise of 1°C leads to a 9% reduction in soil moisture over the southern MDB.

The actual relationship between temperatures and evaporation is driven by interactions between the land surface and the lowest part of the atmosphere, known as the planetary boundary layer (PBL). The land surface and PBL are a tightly coupled system (Santanello *et al*., 2005). The characteristics of the landscape (predominantly soil moisture) influence the atmosphere by controlling the division of net radiation into latent and sensible heat fluxes (Stensrud, 2007, p.12). Conversely the atmosphere forces the land surface through precipitation, momentum and radiative fluxes, due to the moving atmospheric fluid (Trier *et al*., 2008).

In this paper we employ a simple coupled Land Surface–PBL model, which simulates the evolution of the PBL and evaporation over the course of a day, to explore the role of soil moisture in the evolution of daytime air temperatures. The first part of the paper describes the model used, while the second part explores the role of elevated temperature on evapotranspiration for the

May–October 2002 MDB drought. The final part explores how the extreme range of soil moisture conditions influences the maximum air temperatures observed over a summer day.

COUPLED LAND SURFACE – PBL MODEL DESCRIPTION

Planetary boundary layer model

The PBL can be conceptualized to comprise three layers as seen in Fig. 1 (see Stensrud, 2007, for further details). These are the surface layer, where potential temperature increases towards the warmer ground surface, the uniformly mixed layer, which has constant profiles with height of potential temperature, θ (K), and specific humidity, q (kg kg^{-1}), and the inversion layer, above the mixed layer, where potential temperature increases with height. The inversion layer separates the turbulent boundary layer from the free atmosphere and is where entrainment occurs.

This investigation uses a simple convective model of the PBL to assess the evolution of daytime temperature under different soil moisture scenarios. The model predicts both the daily evolution of the PBL height and temperature allowing the evolution of daytime temperature under different conditions to be determined.

In this model the PBL is represented by a slab of air of height h (m) with uniform potential temperature and uniform specific humidity. A step inversion caps the slab which is in turn overlain by drier stably stratified air, as shown in Fig. 1. The inversion is idealised as a step discontinuity in temperature and humidity. The governing equations below are as given in Quinn *et al.* (1995).

The temperature of the slab (θ), is described using the differential equation:

$$\frac{d\theta}{dt} = \frac{H_S + H_i}{h\rho c_P} \tag{1}$$

where ρ is the density of air (kg m^{-3}), c_P is the specific heat capacity of the air at constant pressure and t is time (s).

The water vapour budget is similar and controlled by the fluxes of water vapour at the surface, E_S (kg m^{-2} s^{-1}), and at the inversion, E_i (kg m^{-2} s^{-1}). For the specific humidity, q:

$$\frac{dq}{dt} = \frac{E_S + E_i}{h\rho} \tag{2}$$

The sensible heat flux at the inversion, H_i (W m^{-2}), is given by:

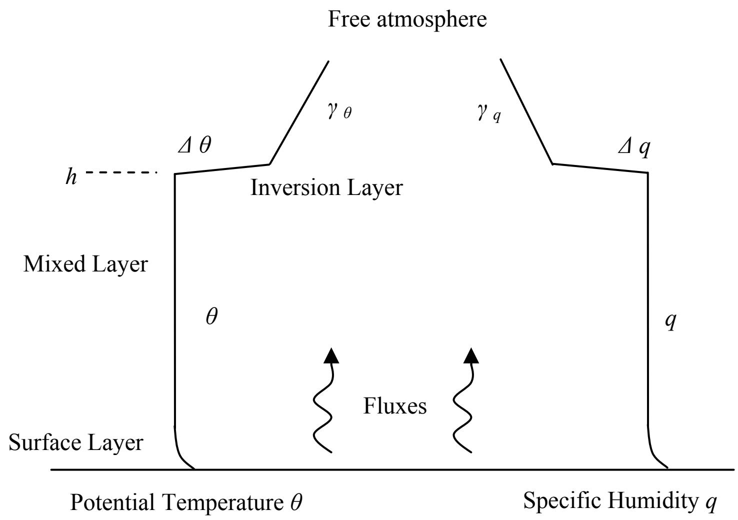

Fig. 1 Boundary layer profiles of potential temperature and specific humidity. Figure based on the schematics presented by Quinn *et al*. (1995) and Margulis & Entekhabi (2001).

$$H_i = \rho c_P \Delta\theta \frac{dh}{dt} \tag{3}$$

and the water mass flux at the inversion, E_i, is given by:

$$E_i = \rho \Delta q \frac{dh}{dt} \tag{4}$$

The sensible heat flux at the inversion is proportional to the sensible heat flux at the surface, H_S (W m^{-2}), where the constant of proportionality (the entrainment coefficient *c*) takes values in the range 0–1. This yields the rate of change of height, *h*, of the slab as follows:

$$\frac{dh}{dt} = \frac{cH_S}{\rho c_P \Delta\theta} \tag{5}$$

The inversion strength ($\Delta\theta$) tends to decrease as the boundary layer warms. Additionally the inversion strength tends to increase as entrainment into the stable air above the inversion base increases (Tennekes, 1973). Per unit time the entrainment increases the inversion strength by an amount $\gamma dh/dt$ giving a net rate of change of Δ as:

$$\frac{d\Delta\theta}{dt} = \gamma_\theta \frac{dh}{dt} - \frac{d\theta}{dt} \tag{6}$$

for temperature, and:

$$\frac{d\Delta q}{dt} = \gamma_q \frac{dh}{dt} - \frac{dq_m}{dt} \tag{7}$$

for humidity. In equations (6) and (7), γ_θ and γ_q are the gradients (lapse rates) of potential temperature (K m^{-1}) and specific humidity (kg kg^{-1} m^{-1}) in the overlying air.

To avoid harmful numerical instabilities and solution errors that can obscure the behaviour of the simple PBL model, the system of equations (1)–(7) was discretized in time using the implicit Euler scheme and solved using Newton-Raphson iteration (e.g. Kavetski & Clark, 2011).

Penman-Monteith model

The Penman-Monteith model is used to estimate the latent heat flux, *LE* (Monteith, 1965, 1981) as:

$$LE = \frac{\Delta(Rn - G) + \rho c_p \left(\frac{e_s - e_a}{r_a}\right)}{\Delta + \gamma\left(1 + \frac{r_s}{r_a}\right)} \tag{8}$$

where *Rn* is the net radiation (W m^{-2}), *G* is the soil heat flux (W m^{-2}), Δ is the slope of the saturated specific humidity temperature curve, $(e_s - e_a)$ is the specific humidity deficit, ρ is the density of air (kg m^{-2}), c_p the specific heat of air at constant pressure (J kg^{-1} $°C^{-1}$), γ is the psychrometric constant, r_s is the bulk stomatal resistance (s m^{-1}) and r_a is the aerodynamic resistance (s m^{-1}).

The aerodynamic resistance was calculated using:

$$r_a = \frac{\left[\log\left(\frac{z - d}{z_0}\right)\right]^2}{k^2 u} \tag{9}$$

where *u* is the mean wind speed (m s^{-1}), *z* is the reference height of the anemometer (m), *d* is the zero plane displacement (m), z_0 the roughness length (m) and *k* is von Karman's constant (= 0.41).

The stomatal resistance parameter is a function of soil moisture availability, solar radiation, temperature and CO_2. This parameter incorporates the physiological resistance that crops impose on water transfer from within their internal organs to their outer surfaces.

SIMULATION OF THE INFLUENCE OF AIR TEMPERATURE ON EVAPORATION DURING THE 2002 DROUGHT

Nicholls (2004) observed that the drought in May–October 2002 was about 2°C warmer than the 1952–2002 average, and suggested that these increased temperatures led to enhanced evaporation and evapotranspiration. Here, the coupled Land Surface–PBL model is used to test the claims that increased temperatures led to increased evaporation during the 2002 drought. We also explore how drought conditions influence potential and actual evaporation.

The PBL model was used to simulate wet and dry land surface conditions, achieved by changing the surface resistance parameter in the Penman-Monteith equation from 50 to 500 s m^{-1} for wet and dry scenarios, respectively. For each wet and dry moisture condition, the actual evapotranspiration at the land surface and the potential evaporation from an open body of water were calculated at each time step. The potential evapotranspiration is calculated to represent the expected evaporative losses from exposed surface water under both wet and dry landscape conditions.

To assess the potential impact of elevated air temperatures, three initial temperatures were specified based on the mean 1952–2002 May–October minimum temperature which was 6°C. The 2002 drought was approximately 2°C warmer than the 1952–2002 average. To represent the impact of elevated temperatures, model simulations were initialised with both wet and dry bulb temperatures set at 4°C, 6°C and 8°C. The wet and dry bulb temperatures are set equal so as to ensure that no initial vapour deficit exists, and that any consequent divergence is due to the initial temperatures, and the wet/dry scenarios.

Figure 2(a) and (b) shows the resultant simulations of actual evapotranspiration under wet and dry conditions, respectively. As expected the evaporative heat fluxes are higher in the wet scenario due to the ready availability of moisture. Importantly, both graphs demonstrate a very minor increase in evaporative fluxes associated with increased temperature. It was calculated that under the wet conditions an increase in air temperature of 2°C only gives an additional 0.076 mm of evapotranspiration over the entire day.

Figure 2(c) and (d) shows the simulations of potential evaporation as would occur from surface water under both wet and dry land surface conditions, respectively. Again, only minor differences occur as a function of air temperature. Importantly, the potential evapotranspiration is actually enhanced in the dry land surface scenario. This is due to the relatively high atmospheric moisture demand (vapour pressure deficit, shown in Fig. 2(f)), which itself is due to the lack of actual evapotranspiration in the dry scenario.

Figure 2(e) shows the simulated air temperature for both wet and dry scenarios. The simulations clearly show higher air temperatures occur under the dry land surface scenario, due to enhanced heating rates resulting from the lack of available moisture for evapotranspiration.

These results all demonstrate that potential evaporation under dry conditions is elevated not as a result of the air temperature, but as a result of the lack of actual evaporation which has the effect of increasing surface heating and hence air temperatures. This is an entirely natural consequence of the dynamics of drought. Importantly, the PBL model shows that antecedent temperature increases do not lead to significant increases in evapotranspiration.

SIMULATION OF THE INFLUENCE OF WET AND DRY SOIL MOISTURE ON TEMPERATURES

The PBL model was next used to determine the influence of extreme soil moisture conditions on the evolution of summer day time temperatures.

For this analysis the data required by the model were obtained from observations made during the First ISLSCP (International Satellite Land Surface Climatology Project) Field Experiment (FIFE). This was a land-surface–atmosphere experiment, conducted from May 1987 to late 1989, centred on a 15 × 15 km grassland site near Manhattan, Kansas, USA. During the intensive field campaigns, the fluxes of heat, moisture, carbon dioxide and radiation were measured with surface

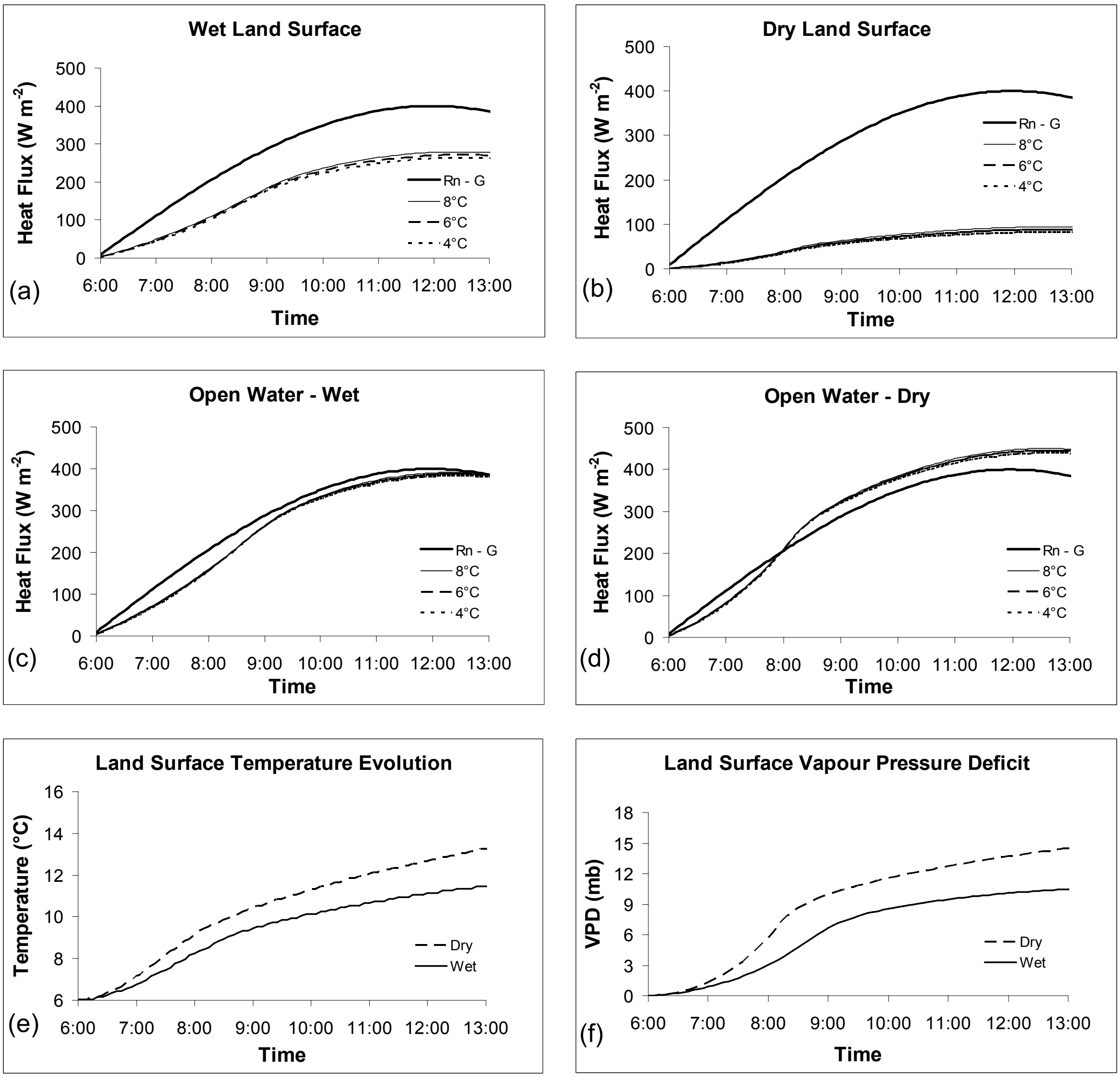

Fig. 2 Evolution of actual and potential evapotranspiration during a sunny day for initial temperatures of 4, 6 and 8°C: (a) the evolution of actual ET under wet conditions, (b) the evolution of actual ET under dry conditions, (c) the evolution of potential ET under wet conditions, and (d) the evolution of potential ET under dry conditions. For an initial temperature of 6°C: (e) the temperature evolution under dry and wet conditions, and (f) the vapour pressure deficit under dry and wet conditions.

and airborne equipment in coordination with measurements of surface and atmospheric parameters and satellite overpasses. The primary source of data used to run the model comes from the spatially-averaged FIFE surface measurements (Betts & Ball, 1998) and the PBL radiosonde observations. The surface measurements are given at 30-min resolution, while radiosondes were launched at roughly 90-min intervals between sunrise and sunset.

The PBL model was forced with data from the FIFE day 5 June 1987. This day had a maximum net radiation of 690 W m^{-2}. An early morning radiosonde was used to estimate the initial conditions and lapse rates required by the model. In addition to the initial conditions and lapse rates, the model also required the latent and sensible heat flux as forcing variables. The latent heat flux (W m^{-2}) through the day was determined using the Penman-Monteith equation.

The surface resistance parameter, which is a measure of moisture availability, was used to simulate varying soil moisture conditions. A value of 50 s m^{-1} was used to simulate high soil moisture while a value 5000 s m^{-1} was used to simulate very low soil moisture. The results of the

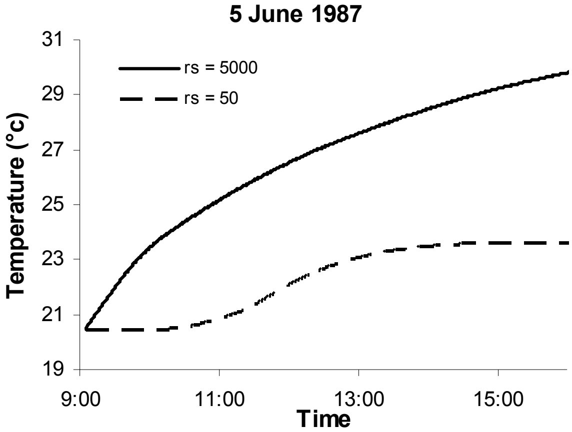

Fig. 3 Evolution of temperature under dry and wet soil moisture conditions.

simulation are shown in Fig. 3. It can be seen that the dry soil conditions led to a much greater increase in daytime temperature; the wet soil moisture conditions yielded a temperature increase of 3.1°C over the course of a day while the dry soil moisture conditions yielded a temperature increase of 8.9°C.

From these results it is clear that soil moisture influences the maximum temperature that can be reached during the daytime. Soil moisture controls the division of net radiation into latent and sensible heat. Dry soil moisture conditions lead to increased temperatures as there is less evaporation and therefore more net radiation is partitioned into sensible heat allowing for increased temperature.

CONCLUSIONS

Previous studies examining the recent drought in the MDB have confused the causality of the interaction between evaporation and temperature by suggesting that increased temperatures were responsible for increased evaporation. Here we used a simple Planetary Boundary Layer model to show that the anomalously high temperatures reached during the 2002 drought would have had only a small impact on the actual evaporation occurring in the drought. More generally, it was illustrated that soil moisture has the dominant influence on the amount of actual and potential evaporation that occurs, and that this in turn influences the temperature that can be reached during the day. Under dry conditions the lack of actual evaporation leads to an increase in the atmospheric moisture demand. In turn, this results in elevated potential evaporation, increased surface heating and ultimately the higher temperatures observed during drought. These are entirely natural consequences of hydrological drought.

REFERENCES

Betts, A. K. & Ball, J. H. (1998) FIFE surface climate and site-averaged data set: 1987–1989. *J. Atmos. Sci.* **55**, 1091–1108.

Cai, W. J. & Cowan, T. (2008) Evidence of impacts from rising temperature on inflows to the Murray-Darling Basin. *Geophys. Res. Lett.* **35**, L07701, doi:10.1029/2008GL033390.

Cai, W., Cowan, T., Briggs, P. & Raupach, M. (2009) Rising temperature depletes soil moisture and exacerbates severe drought conditions across southeast Australia. *Geophys. Res. Lett.* **36**(21), L21709, doi:10.1029/2009GL040334.

Karoly, D. J., Risbey, J. & Reynolds, A. (2003) Global warming contributes to Australia's worst drought. WWF research report, WWF Australia.

Kavetski, D. & Clark, M. (2011) Numerical troubles in conceptual hydrology: approximations, absurdities and impact on hypothesis testing. *Hydrol. Processes* **25**, 661–670, doi:10.1002/hyp.7899.

Margulis, S. A. & Entekhabi, D. (2001) A coupled land surface–boundary layer model and its adjoint. *J. Hydromet.* **2**, 274–296.

Monteith, J. L. (1965) Evaporation and environment. In: *The State and Movement of Water in Living Organisms* (ed. by G. E. Fogg), 205–234. Symp. Soc. Exp. Biol. 19, Academic, New York, USA.

Monteith, J. L. (1981) Evaporation and surface temperature. *Quart. J. Roy. Met. Soc.* **107**(451), 1–27.

Nicholls, N. (2004) The changing nature of Australian droughts. *Climatic Change* **63**, 323–336.

Quinn, P., Beven, K. & Culf, A. (1995) The introduction of macroscale hydrological complexity into land surface–atmosphere transfer models and the effect on planetary boundary layer development, *J. Hydrol.* **166**, 421–444.

Santanello, J. A., Friedl, M. A. & Kustas, W. P. (2005) An empirical investigation of convective planetary boundary layer evolution and its relationship with the land surface. *J. Appl. Met.* **44**, 917–932.

Stensrud, D. J. (2007) *Parameterization Schemes: Keys to Understanding Numerical Weather Predicition Models*. Cambridge University Press.

Tennekes, H. (1973) A model for the dynamics of the inversion above a convective boundary layer. *J. Atmos. Sci.* **30***,* 558–567.

Trier, S. B., Chen, F., Manning, K. W., LeMone, M. A. & Davis, C. A. (2008) Sensitivity of the PBL and precipitation in 12-day simulations of warm-season convection using different land surface models and soil wetness conditions. *Monthly Weather Rev.* **136**, 2321–2343.