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Rising temperature depletes soil moisture and exacerbates severe drought conditions across southeast Australia

Wenju Cai,[1] Tim Cowan,[1] Peter Briggs,[2] and Michael Raupach[2]

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[1] Over the past decade the southern catchments of the Murray Darling Basin (MDB), responsible for much of Australia's agricultural output, have experienced a severe drought (termed the "Big Dry") with record high temperatures and record low inflow. We find that during the Big Dry the sensitivity of soil moisture to rainfall decline is over 80% higher than during the World War II drought from 1937–1945. A relationship exists between soil moisture and temperature independent of rainfall, particularly in austral spring and summer. Annually, a rise of 1°C leads to a 9% reduction in soil moisture over the southern MDB, contributing to the recent high sensitivity. Since 1950, the impact from rising temperature contributes to 45% of the total soil moisture reduction. In a warming climate, as the same process also leads to an inflow reduction, the reduced water availability can only be mitigated by increased rainfall. Other implications for future climate change are discussed. **Citation:** Cai, W., T. Cowan, P. Briggs, and M. Raupach (2009), Rising temperature depletes soil moisture and exacerbates severe drought conditions across southeast Australia, *Geophys. Res. Lett.*, *36*, L21709, doi:10.1029/2009GL040334.

1. Introduction

[2] Much of the Australia's agricultural activities are carried out over southeastern Australia (SEA), including some regions where the required soil moisture level is maintained by irrigation using water from the country's longest river system, the Murray and Darling Rivers. In recent years, the southern MDB has experienced a severe drought. During this the Big Dry (1995 to present), while the multi-year mean of rainfall or subsurface soil moisture averaged over the region are not the lowest on record (Figure 1a), inflows have been hovering at historical low levels of about 10–20% of the climatological mean, record-breaking in terms of lowest annual or multi-year accumulative total [*Cai and Cowan*, 2008a]. Thus, compared with inflows during the Federation Drought (1895–1903) or the World War II (WWII) drought (1937–1945), the Big Dry is the longest and most severe on record [*Cai and Cowan*, 2008a; *Murray Darling Basin Authority* (*MDBA*), 2009]. The severity of the Big Dry has also been exacerbated by increasing air temperatures over the past 2–3 decades [*Nicholls*, 2004; *Cai and Cowan*, 2008a; *Murphy and Timbal*, 2008; *Ummenhofer et al.*, 2009].

[3] The recent decline in inflow is caused by two important factors. Firstly, austral autumn rainfall is crucial for wetting the catchments so that inflow can be efficiently generated during the main rainfall seasons of winter and spring. As a result, the annual inflow is sensitive to autumn rainfall, more so than to rainfall in other seasons [*Cai and Cowan*, 2008b]. Since 1950, autumn rainfall across SEA has shown the largest reduction of all seasons, therefore contributing to the large inflow decline. The high sensitivity of annual inflow to this "autumn wetting" mechanism highlights the importance of soil moisture in the inflow generation. Secondly, although low rainfall is accompanied by a warming anomaly due to a decreased cloud cover and by low actual evaporation, high temperatures independent from the influence of rainfall have exacerbated the severity of the recent drought [e.g., *Nicholls*, 2004; *Cai and Cowan*, 2008a]. Indeed, for the southern MDB, where most inflows are generated, higher temperatures in the Big Dry are the main difference to the WWII drought (Figure 1a). *Cai and Cowan* [2008a] showed that the seasonality of the rainfall reduction alone is unable to account for the extent of inflow reduction since the 1950s, and that a 1°C rise in temperature causes an additional 15% reduction in annual inflow. Are such impacts from rising temperature supported by an imprint in soil moisture? This is the major issue we will address.

2. Data and Method

[4] Relative soil moisture data from 1900 has been developed as a part of the Australian Water Availability Project [*Raupach et al.*, 2008]. It consists of two layers: an upper layer (or WRel1) from surface to 0.2 m, and lower layer (or WRel2) from 0.2 m to 1.5 m. They are products of a simple, robust water balance model that combines model-data information. The model uses a fixed soil type for each grid-point and a fixed monthly climatological cycle of vegetation greenness, and is validated using multi-decadal streamflow data in 200 unimpaired catchments. Thus, any changes in soil moisture are due to changes in meteorological parameters. Also deployed are monthly gridded rainfall and maximum temperature (Tmax) data across Australia from 1950–2008, from the Australian Bureau of Meteorology Research Centre [*Jones et al.*, 2006]. The soil moisture and rainfall data are referenced to the 1900–2008 climatological period, whereas Tmax anomalies are referenced to the 1950–2008 period. A test shows that using a common reference period yields almost identical results.

[5] We first establish a relationship between soil moisture and the component of temperature fluctuations that is independent of rainfall, hereafter referred to as residual Tmax. We linearly detrend all seasonal data, to ensure that any relationship is not generated by trends in both time series.

[1] CSIRO Marine and Atmospheric Research, Aspendale, Victoria, Australia.

[2] CSIRO Marine and Atmospheric Research, Canberra, ACT, Australia.

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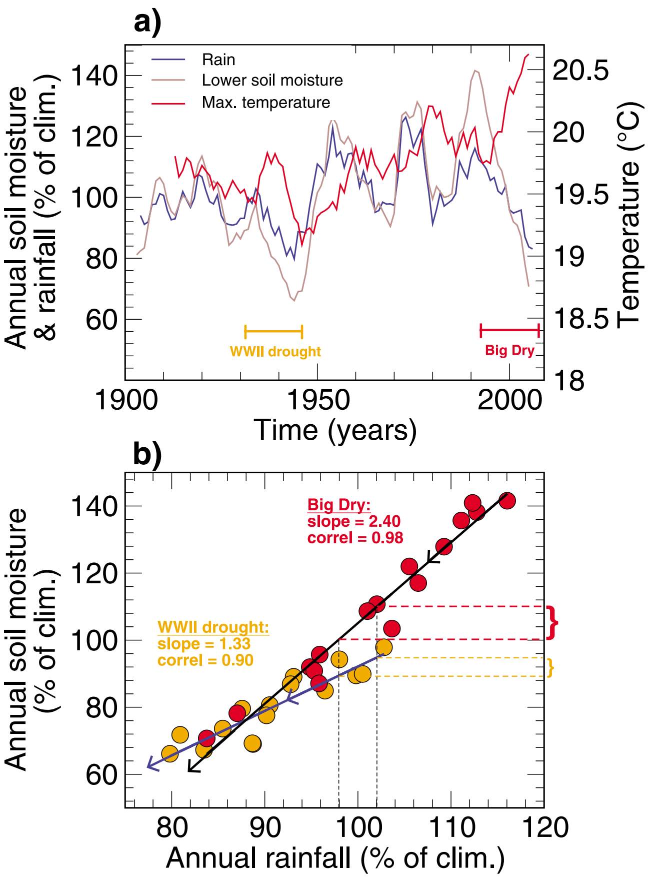

Figure 1. (a) 20th century time series of annual rain (blue curve) and subsurface soil moisture (WRel2, brown curve, shown at the 1-year lag) across the southern MDB, together with Victoria Tmax (°C, red curve), all smoothed with a 7-year running mean. (b) Scatterplot of rainfall versus WRel2 using sequences in Figure 1a showing the sensitivity of WRel2 to rainfall reduction encompassing the two drought periods starting from the time when the rainfall decline commences: WWII drought (1931–1946) and the Big Dry (1990–2005). Units of rain and WRel2 are % of climatology. The southern MDB is defined as 140°–150°E, 33°–37°S.

We then regress detrended seasonal Tmax onto detrended seasonal rainfall to obtain Tmax fluctuations that vary with rainfall, which are then subtracted from the detrended Tmax to yield detrended seasonal residual Tmax (as in *Nicholls* [2004] and *Cai and Cowan* [2008a]).

3. Seasonality of Rainfall and Relative Soil Moisture Trends

[6] Before we proceed, we examine the long-term trends (1950–2008) in rainfall, WRel1 and WRel2 for each season (Figure S1 of the auxiliary material).[1] Although the trend patterns of WRel1 and rainfall share a similar seasonal structure, with the largest reduction in autumn throughout SEA, the trend patterns in WRel2 show little seasonality, reflecting a significant cross-season "memory" due to the large soil water capacitance of the lower layer. As such, a memory of a given rainfall change, for example, the reduction in autumn rainfall, is observed in WRel2 across all seasons. In other words, unlike inflows, which are sensitive to the seasonality of a rainfall reduction, WRel2 is not. This non-sensitivity to seasonal rainfall trends is one reason why annual mean subsurface soil moisture has not reached a historical low (Figure 1a), in contrast to inflows [*Cai and Cowan*, 2008a].

[7] Water in the upper soil layer depletes relatively quickly through evaporative loss to the atmosphere and penetration to the lower soil layer. Therefore, WRel1 has a relatively short memory of rainfall, closely following variations and changes in seasonal rainfall (Figure S1). A seasonal lag correlation analysis on detrended anomalies shows that at Lag 1 (i.e., one season lag with rainfall leading WRel1), there is little correlation. After depletion, temperature fluctuations have little impact. Thus, much of the long-term trends in WRel1 are attributable to trends in rainfall (Figure S1); and, as we will show, the imprint of an influence from rising temperature resides in WRel2, where water availability persists for a longer time for temperature to exert an impact. A similar lag correlation for WRel2 confirms that significant correlation persists for many seasons through into the following year. In fact, correlation at a 1-year lag using the 7-year running mean version of an annual time series of WRel2 and rainfall (Figure 1a) is slightly greater than that at Lag 0. We adjust for this lag to maximise the coherence between the two filtered time series in our analysis.

4. Sensitivity of Soil Moisture to Rainfall Reduction

[8] Because neither rainfall nor WRel2 during the Big Dry has reached historic lows of those in the WWII drought, an impact on WRel2 from rising temperature is not immediately clear. However, the sensitivity of WRel2 to a rainfall reduction (expressed as percentage of climatology), either at Lag 0 or Lag 1, is higher during the Big Dry than during the WWII drought, as shown through a 7-year running average which suppresses variance on shorter time scales (Figure 1a). The slope (ratio of soil moisture to rainfall) for the Big Dry is 2.40, far steeper than that for the WWII drought at 1.33, a sensitivity difference of 80%. Thus a 4% reduction in rainfall (Figure 1b, red dashed lines) is associated with 9–10% soil moisture reduction for the Big Dry (Figure 1b, blue dashed lines) but only a 5–6% reduction during the WWII drought (Figure 1b, orange dashed lines).

[9] The same comparison is made using data at each grid-point to display the difference in spatial patterns between the two drought periods (Figure 2). Over eastern Australia and the southern MDB, the sensitivities are generally greater than one. This is consistent with the fact that during a drought, temperature is otherwise higher simply because of a severe lack of rainfall; higher temperatures lead to a faster rate of evaporation while the subsurface soil is still relatively wet. However the sensitivity is generally higher during the Big Dry, consistent with a broad-scale warming. One outstanding feature is the particularly high sensitivity over the Murray-Riverina and Mallee-Wimmera regions (140–

[1]Auxiliary materials are available in the HTML. doi:10.1029/2009GL040334.

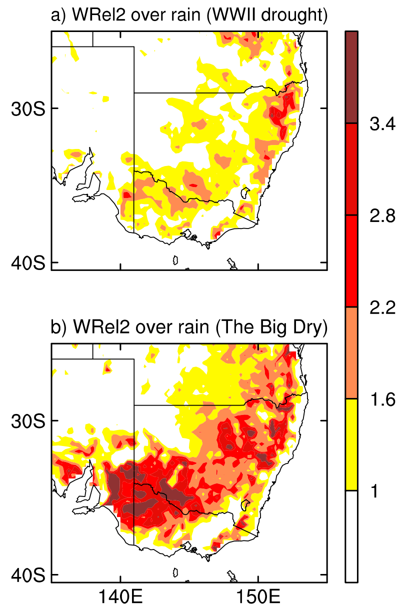

Figure 2. Ratio change of WRel2 to rainfall for the two drought periods: (a) WWII drought (1931–1946) and (b) the Big Dry (1990–2005), using a 7-year running mean.

145°E, 34–36°S), with the reduction rate of WRel2 at more than triple the rate of rainfall (Figure 2b). These regions consist of floodplain and wetland ecosystems that face an increased risk of severe environmental degradation during the Big Dry [*MDBA*, 2009]. Although the regional high sensitivity is in part attributable to the temperature impact in spring and summer, it may not be fully accounted for by the impact of rising temperature alone, as the greatest warming in SEA does not occur in these regions (discussed later). However, the annual rainfall reduction in these regions is rather large, suggesting a nonlinear impact from rising temperature, which may depend on the level of rainfall reduction, a subject beyond the scope of the present study. We focus on the cause of the generally higher sensitivity during the Big Dry.

5. Sensitivity of Soil Moisture to Residual Tmax

[10] To confirm that rising temperature has little impact on WRel1, we regress detrended seasonal grid-point WRel1 onto detrended residual Tmax to generate a sensitivity per 1°C temperature change. The associated correlations are weak and not statistically significant in most regions in any season. Separately, we calculate seasonal trends in Tmax. We then multiply the sensitivity with the trends in Tmax to give the contribution due to rising temperature. As expected, the temperature-induced trends in WRel1 are small, implying that most trends in WRel1 are due to rainfall, supporting the notion that the imprint of impacts from rising temperature resides in WRel2.

[11] By contrast, correlations between detrended WRel2 and detrended residual Tmax are statistically significant at the 95% confidence level, with considerably high regression coefficients over much of eastern Australia in all seasons (particularly in winter over southern Queensland) and across SEA in spring and summer (Figures 3a–3d). Spring is a high-rainfall season in SEA and the hydro-climatic conditions are favourable for evaporation; it is in this season when the strongest impacts from rising temperature are recorded over the Murray-Riverina and Mallee-Wimmera regions with a rate of 10–20% $°C^{-1}$ of the seasonal climatology. The large soil water capacitance (dependant on soil type) carries the significant impact into summer, when hydro-climatic conditions are similarly favourable for evaporation, though the sensitivity is somewhat lower, between 5–10% $°C^{-1}$ of the seasonal climatology over the Murray-Riverina and Mallee-Wimmera regions.

6. Impact of Rising Temperature

[12] What is the impact from temperature during the period encompassing the Big Dry? Since 1990, significant warming has occurred, with the greatest warming rate observed in spring, exceeding 2°C over some locations (Figure 3h). These short-term trends are driven by a combination of global warming and multidecadal variabil-

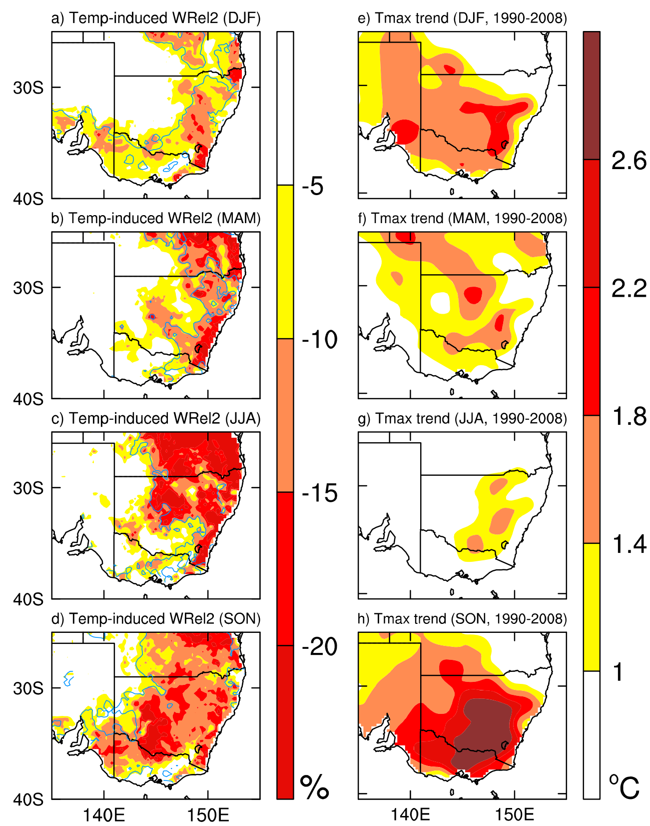

Figure 3. (a–d) Seasonal sensitivity of WRel2 to residual Tmax using detrended anomalies (1950–2008, % of climatology per 1°C), overlaid with correlation coefficients (between residual Tmax and WRel2, shown in blue) significant at the 95% confidence level; (e–h) Tmax trends encompassing the Big Dry period (1990–2008, °C per 19 years). The seasons are defined as summer (December–February, DJF), autumn (March–May, MAM), winter (June–August, JJA) and spring (September–October, SON).

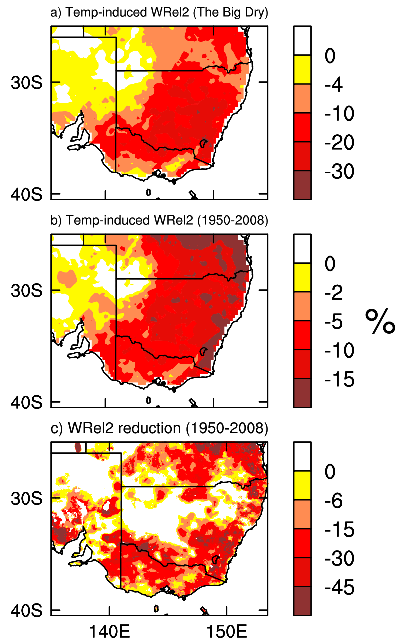

Figure 4. Tmax trend-induced annual change in WRel2 during (a) the Big Dry, and (b) 1950–2008, assuming there is a uniform warming across SEA of 1°C; (c) total trend in annual WRel2 from 1950–2008. Units for Figures 4a are % of climatology per 19 years, while 4b and 4c are % of climatology per 59 years. Note that the legends in Figures 4a and 4c are 2 and 3 times the legend values in Figure 4b, respectively.

ity. The variability component is in part due to an increasing frequency of positive Indian Ocean Dipole (pIOD) events in recent decades, which induce a strong warming anomaly over southern Australia [*Cai et al.*, 2009]. The impact of these warming trends is estimated by multiplying the grid-point seasonal sensitivities (Figures 3a–3d) with the seasonal warming trends (Figures 3e–3h), and then aggregating the product into a total annual impact, taking into account of appropriate weighting based on the seasonal climatological values. There is a 10–25% reduction across SEA (Figure 4a); averaged over the southern MDB, the reduction is 17.3% of the annual total, compared with a total trend of over 80% since 1990 using unfiltered data. Thus, the temperature trend accounts for over 20% of the total reduction during this period.

[13] In the longer term, temperature trends are more spatially homogeneous, and it is useful to construct an annual total WRel2 reduction per 1°C rise. This is done by aggregating the seasonal sensitivity (Figures 3a–3d). For a 1°C temperature rise (Figure 4b), there is generally a WRel2 reduction of more than 15% over the east Australia (e.g., Canberra), and a 5–15% reduction over the southern MDB with an areal average of over 9%. Since 1950, temperature over southern MDB has risen by approximately 1°C, therefore contributing to about 9% reduction in WRel2.

[14] How does it compare with the total WRel2 reduction over the corresponding period? The total reduction since 1950 (Figure 4c) ranges between 15–30%, with an areal average over southern MDB of about 20%. Therefore, the contribution from rising temperature is about 45% of the total decline.

7. Implications for Future Climate

[15] Our results have significant implications for future water availability in the already critically water-short regions of the MDB. In a warming climate even if rainfall does not change, more water will be needed to maintain the same soil moisture levels to cultivate the same crops. This additional water will need to come from river inflows, which will reduce by as much as 15% for a 1°C temperature rise [*Cai and Cowan*, 2008a], in part because the 9% $°C^{-1}$ WRel2 reduction that persists through all seasons will lead to a lower efficiency of the autumn wetting mechanism.

[16] The total future soil water level also depends upon the future rainfall changes. Climate models project a median annual rainfall reduction of 5–15% by 2060 over the MDB [*Shi et al.*, 2008]. The consensus is strong because the reduction is mostly in winter and spring, in which the warming pattern in the eastern Indian Ocean is pIOD-like, due to a robust greater warming over the Eurasian landmass compared to over the ocean [*Ashrit et al.*, 2001]. Further, the Southern Annular Mode trends towards a positive phase in most climate models, which also contributes to lower rainfall [*Shi et al.*, 2008; *Murphy and Timbal*, 2008]. A 15% rainfall reduction alone by 2060 translates to a 26% reduction in MDB inflows [*Cai and Cowan*, 2008a].

[17] If we assume that the relationship between rainfall and WRel2 during the WWII drought (Figure 1a) is valid in a future climate, then a 15% rainfall decrease would contribute to about a 20% reduction in WRel2 by 2060. If the relationship between Tmax and WRel2 (Figure 4b) persists into the future climate, a 2°C increase by 2060 (the mid-range A1B scenario) will reduce the subsurface soil moisture by 19%, comparable to that due to the rainfall reduction. Together, there is an average reduction in WRel2 of almost 40% by 2060 over the southern MDB. If rainfall intensity does not change in the future while mean rainfall decreases, and given that WRel2 can be charged up periodically due to its large capacitance, the reduction will manifest as an increase in the frequency of dry spells, or dry conditions that last longer.

8. Conclusions

[18] Compared with the WWII drought, multi-year averages of rainfall and subsurface soil moisture during the Big Dry are not as low, but the sensitivity of soil moisture to rainfall decline is over 80% higher. We show that a relationship exists between subsurface soil moisture variations and fluctuations of temperature not associated with rainfall over eastern Australia in all seasons, and over SEA in austral spring and summer. On an annual basis, a rise of 1°C leads to an approximate 9% subsurface soil moisture reduction in the southern MDB. Thus, in a warming climate, to maintain the same soil moisture levels in order to maintain the same crop production, more irrigation water will be needed. With the concurrent decline in inflows induced by the same processes dictating soil moisture, the increased irrigation water cannot be provided from rivers without disproportionately curtailing other uses. Our results show that since 1950, the contribution from rising temperature is

comparable to that from declining rainfall. This further strengthens the argument that rising temperatures due to the enhanced greenhouse effect and multidecadal variability have a strong impact on the future of Australia's agriculture in the southern MDB. If the relationship we identify persists into the future, a 2°C rise by 2060 will lead to a 19% subsurface soil moisture reduction. This negative effect can only be offset by an increase in rainfall, which appears to be unlikely, as most climate models are projecting a rainfall reduction. Therefore we can expect more occurrences of low soil moisture with concurrent low inflows, as observed in recent years.

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References

Ashrit, R. G., K. R. Kumar, and K. K. Kumar (2001), ENSO-monsoon relationships in a greenhouse warming scenario, *Geophys. Res. Lett.*, *28*, 1727–1730.

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

Cai, W., and T. Cowan (2008b), Dynamics of late autumn rainfall reduction over southeastern Australia, *Geophys. Res. Lett.*, *35*, L09708, doi:10.1029/2008GL033727.

Cai, W., T. Cowan, and A. Sullivan (2009), Recent unprecedented skewness towards positive Indian Ocean Dipole occurrences and its impact on Australian rainfall, *Geophys. Res. Lett.*, *36*, L11705, doi:10.1029/2009GL037604.

Jones, D., W. Wang, R. Fawcett, and I. Grant (2006), The generation and delivery of level-1 historical climate data sets, final report, Aust. Bur. of Meteorol., Melbourne, Vic.

Murphy, B. F., and B. Timbal (2008), A review of recent climate variability and climate change in southeastern Australia, *Int. J. Climatol.*, *28*, 859–879.

Murray Darling Basin Authority (MDBA) (2009), River Murray System. Drought Update, *Tech. Rep. 19*, Murray Darling Basin Authority, Canberra, ACT.

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

Raupach, M. R., P. R. Briggs, V. Haverd, E. A. King, M. Paget, and C. M. Trudinger (2008), *Australian Water Availability Project (AWAP) Component: Final Report for Phase 3*, CSIRO Mar. and Atmos. Res., Canberra, ACT.

Shi, G., J. Ribbe, W. Cai, and T. Cowan (2008), An interpretation of Australian rainfall projections, *Geophys. Res. Lett.*, *35*, L02702, doi:10.1029/2007GL032436.

Ummenhofer, C. C., M. H. England, P. C. McIntosh, G. A. Meyers, M. J. Pook, J. S. Risbey, A. S. Gupta, and A. S. Taschetto (2009), What causes southeast Australia's worst droughts?, *Geophys. Res. Lett.*, *36*, L04706, doi:10.1029/2008GL036801.

W. Cai and T. Cowan, CSIRO Marine and Atmospheric Research, PMB 1, Aspendale, Vic 3195, Australia. (wenju.cai@csiro.au)

P. Briggs and M. Raupach, CSIRO Marine and Atmospheric Research, GPO Box 3023, Canberra, ACT 2601, Australia.