Australian temperature, Australian rainfall and the Southern Oscillation, 1910-1992: coherent variability and recent changes

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The best available surface temperature (T) and precipitation (P) records for Australia dating back to 1910 have been examined to look for coherent interannual variability. P exhibits a tendency to be out of phase with daily maximum temperature, Tmax, and this results in P tending to be out of phase with both the daily average temperature, Tbar (estimated here as the average of Tmax and the daily minimum, Tmin), and the DTR (diurnal temperature range, Tmax-Tmin). The association between P and Tmin is generally weak. The (expected) increase in P associated with a positive Southern Oscillation Index is (generally) accompanied by reduced average temperatures (Tbar) and a reduced DTR, both of which primarily arise from a reduction in Tmax. When variability in both P and Tmin associated with Tmax is removed, the residual signals (P* and Tmin*) show widespread statistically significant positive correlations, consistent with the hypothesis that clouds help to reduce night-time cooling. These relationships are less clear at near-coastal sites, and absent at the island and exposed coastal sites considered.

Results from three separate ten-year integrations of the Bureau of Meteorology Research Centre’s atmospheric general circulation model were then examined. The tendency for (a) P to be out of phase with Tmax, Tbar and the DTR and (b) P* to be in phase with Tmin* over Australia on interannual time-scales was also generally evident over land elsewhere, except at high latitudes and over North Africa. An analysis of the model’s surface heat budget over land showed that this arises from associated surface short wave radiation and latent heating anomalies. The latter is generally more important over low-latitude regions where deep convection occurs, with the hierarchy reversed elsewhere. Evaporative cooling anomalies appear to be dominated by soil moisture changes. Surface long wave radiation, sensible heating and subterranean heat exchange tend to reduce the temperature change which would otherwise occur.

Recent changes in some of the relationships exhibited between observed P, T and the Southern Oscillation Index appear unusual in terms of the interdecadal variability evident in the records prior to 1972, and previous conclusions drawn on the basis of ‘all-Australia’ P and T indices were found to have broad applicability. Interrelationships between recent changes in the 20-year means of P, T and the SOI do not match the changes that might be expected on the basis of their interrelationship on interannual time-scales. Possible reasons for the changes suggested by the analysis (e.g., global warming and naturally occurring interdecadal climate variability) are discussed.

Introduction

This study focuses on relationships between interannual fluctuations in the Southern Oscillation Index (SOI),

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Australian precipitation (P) and Australian surface temperature (T), and how they may have changed since the early 1970s.

There are a number of reasons why P and T might be statistically related. For example: (a) if rainy years are
cloudy years then less sunshine will reach the surface, thereby lowering the maximum daily T (Tmax); (b) soil moisture is reduced during dry periods and so if the buffering effect of evaporation is reduced, a given surface short wave radiation (SSWR) heating anomaly will produce a larger change in T; (c) if rainy years have increased cloud at night-time, then more long wave radiation (LWR) will be trapped in the lower atmosphere, and the minimum daily T (Tmin) will rise.

If (a) and (b) dominate the relationship between P and T then it might be expected that they are out of phase. This hypothesis was tested by Coughlan (1979) who examined all annual-mean Tmax and P data available at the time for the period 1946-1975. He found this to be generally true except along the periphery of the continent and in northern Australia.

More recently, Nicholls et al. (1996a) used ‘all-Australia’ indices for P and T to independently show that high P is accompanied by a reduced diurnal temperature range, which was found to be primarily associated with reduced daytime temperatures. They suggested that these relationships might be due to cloud changes.

In this paper we will re-examine these relationships over the period 1910-1992, using the most reliable records currently available (Torok and Nicholls 1996; Lavery et al. 1992, 1997; see following section). The same observational records used by Nicholls et al. (1996 a,b) will be employed, but here they will be examined on a regional basis rather than combining them into area averages.

In the second part of this study, results obtained from three recent ten-year integrations of the Bureau of Meteorology Research Centre’s atmospheric general circulation model (Bourke et al. 1977; McAvaney et al. 1978, 1991; McAvaney and Colman 1993), or AGCM, will be presented, first to see if the model captures the anticipated relationships, but also to help quantify the relative importance of mechanisms (a) to (c).

The third and final part of this study is concerned with recent changes in the interrelationships between P, T and the SOI. Changes of this kind have important implications, as statistical methods are used to forecast seasonal climate anomalies over Australia (e.g., McBride and Nicholls 1983; Zhang and Casey 1992). If these changes are unprecedented then we may have entered a new climatic regime in which the previous methods will be less successful.

Interdecadal variability in the interannual relationships was suggested by Nicholls et al. (1996b) using ‘all-Australia’ indices for P and T. They highlighted recent changes in the relationships during the 20-year period beginning in the early 1970s. During this time, for any given value of the SOI or P, maximum daily T (Tmax) has tended to be higher than was the case previously. Likewise, for any given value of the SOI, P has tended to be greater than would have been expected on the basis of the earlier relationship.

Here we aim to quantify whether these recent changes have been unusual in terms of the variability exhibited in previous decades, and to examine the regional dependence of the changes. If they are unprecedented then perhaps we are seeing evidence for climate change, although the possibility that it is instead due to naturally occurring climate variability will be difficult to rule out on the basis of records which are less than one century long.

The paper is organised as follows. The observational data are described in the following section. The relationship between P and T in these records is then examined using both gridded and original station data. Results from the AGCM are also analysed, first to see if the same relationships are in evidence, and then to test hypotheses (a) to (c) above. An analysis of recent changes in the observed relationships and a discussion of the possible reasons for the changes precedes a summary of the main results.

**Observational records**

The precipitation data (Lavery et al. 1997) represent an expanded version of the data described by Lavery et al. (1992), which now gives an enhanced coverage by incorporating additional stations and composites of two or more neighbouring stations. The dataset consists of 341 stations covering the period 1910-1992.

The surface air temperature data (Torok 1996; Torok and Nicholls 1996) cover the period 1910-1993, and have been adjusted to account for inhomogeneities caused by station relocations and changes in exposure using objective statistical methods. An exhaustive search of the available historical documentation regarding observational practices, instrumentation, site relocation and exposure was conducted to assess the quality of both the P and T records and to choose records for compositing. It is believed that both sets are the most reliable for estimating changes in Australian climate over the past century, and have already proved useful in a number of studies (Torok, 1996; Torok and Nicholls 1996; Nicholls et al. 1996a, b; see also Nicholls and Lavery 1992; Nicholls and Kariko 1993). Note that while these datasets are the best available, imperfections at isolated stations may exist.

Both the (annual average) daily maximum (Tmax) and the daily minimum (Tmin) are available (from 224 stations), from which we have estimated the daily average, Tbar = (Tmin + Tmax)/2, and the diurnal temperature range, DTR = Tmax - Tmin. The station data were interpolated onto a regular 0.1 degree latitudinal/longi-
tudinal grid using a Barnes (1964) analysis technique and some of the resulting fields were then compared with hand-drawn analyses for particular years. The various parameters in the analysis were then tuned to give the best fit. Overall the agreement was (ultimately) found to be good, with greatest disagreement where P is large at coastal sites, where the method has excessively smoothed the data, and over Tasmania, where the data are restricted to a relatively small region for which the analysis method is not ideal. Caution should also be exercised when examining results over largely uninhabited central Australian regions, where the density of T station data is low (Torok, 1996).

A subset of the original station data was analysed for 28 stations which have both high-quality T and P data for extended periods. This reveals contrasts between inland, coastal and island stations not apparent in the gridded data.

**Inter-relationships between observed temperature and precipitation**

A number of possible relationships between P and T were suggested in the introduction on the basis of three very simple but plausible physical processes. Of course a large number of additional processes operate simultaneously, and some of these may offset or even dominate the three processes highlighted. Additional uncertainty arises from the fact that while annual rainfall totals may be dominated by a relatively small number of events, the temperature will have contributions from all days. From this perspective it is not obvious that the relationships will in fact arise.

To see if they do, correlation coefficients have been calculated for observed P and (a) Tbar, (b) Tmin, (c) Tmax, and (d) the diurnal temperature range (DTR) using data from 1910-1992. The results are presented in Fig.1. The ‘all-Australia’ indices for P and T were used to estimate the impact of autocorrelation on statistical significance. The ratio $\gamma = \sigma^2(1) / \sigma^2(0)$, where $\sigma(\tau)$ is the standard deviation at lag $\tau$, provides a measure of persistence. It was found to be 0.27 for P, 0.01 for Tmax, 0.40 for Tmin, 0.24 for Tbar and 0.15 for the DTR. Here we assume that the statistic $t = (r/f)\sqrt{N-2} / \sqrt{(1-(r/f)^2)}$ has a t-distribution (Walpole and Myers, 1985) where N is the number of data points at each grid-point (83), r is the (temporal) cross-correlation coefficient between two time series $x(t)$ and $y(t)$, and $f = \sqrt{r(1 - \gamma^2_y)}/(1 - \gamma^2_x \gamma^2_y)$ incorporates the impact of serial correlation (which is derived from the formulation given by Katz, 1988).

The largest estimate for the 95 per cent significance level is given by the two time series with the greatest autocorrelations, namely Tmin and P. Using $t_{95%} = 1.99$ (from tables) together with the figures for P and Tmin

**Fig.1** Correlation between P (precipitation) and (a) daily average temperature (Tbar), (b) daily minimum temperature (Tmin), (c) daily maximum temperature (Tmax) and (d) the diurnal temperature range (DTR) over Australia, for the period 1910-1992. All of the data were annually averaged. The 95 per cent statistical significance level is approximately 0.24 and is the same in all plots of correlations between observations.
Table 1. Correlation coefficients between precipitation (P) and Tmax, Tmin and the DTR for stations which have both P and T records. The stations have been divided into three regions: inland, coastal (or near-coastal) and islands. Coefficients have been multiplied by 100, bold and underlined indicates correlation is significant at or above the 99% level, bold the 95% level, and underlined the 90% level. P* and Tmin* have had variability coherent with Tmax removed.

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<th>(P, Tmin)</th>
<th>(P, DTR)</th>
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the entire continent. To test the fourth possibility we examined the original station data at sites where both T and P data are available. The correlations at these sites are given in Table 1. The figures indicate significance at the 99 per cent (bold and underlined), 95 per cent (bold only) and 90 per cent (underlined only) levels. Persistence was again taken into account using the method described earlier, but this time using the local P and T rather than the all-Australia indices. It was found that the impact of serial correlation was much smaller than that calculated above primarily because the local rainfall does not exhibit the same level of persistence as the indices. Greatest persistence was evident in northern Australia, but even there it reduced the 'effective' correlation coefficient by typically less than ten per cent. This means that the test described earlier is a conservative one.

Three types of stations are shown in Table 1 – inland, coastal (or near-coastal) and islands. The results are similar to those obtained with the gridded data – P tends to be out of phase with Tmax and the DTR, and statistically significant correlations between P and Tmin of both signs are in evidence. Notice that the relationship between P and Tmax is generally more prominent at the inland stations than at the coastal stations (e.g., 13 out of 14 vs 7/14 at the 99 per cent level) and that the correlations tend to be larger. The presence of a coastal contrast is not apparent in the correlation field of the gridded data – it has indeed been smoothed away. Coughlan's results suggest that the relationship between P and Tmax is absent near the coast, whereas here we see that while it is less dominant it is still in evidence. The relationship is not evident at the two island sites, nor is it evident at two of the exposed sites at Wilson's Promontory and Cape Schanck.

In the introduction we hypothesised that Tmin might tend to increase in rainy years because of increased long wave radiation trapping. Yet this relationship does not appear to be borne out by the observed results (Table 1 and Fig. 1(b)), which display both positive and negative correlations between P and Tmin. It is important to realise, however, that Tmin is set jointly by the temperature at the end of the day and by processes occurring at night-time. We have already seen that Tmax tends to be out of phase with P, and will therefore act to mask the impact of the physical processes dominating evolution during the night. So in order to test hypothesis (c) we need to first remove that part of the signal which is coherent with Tmax (assuming that it is a good proxy for the temperature at sunset). To do this we follow Nicholls et al. (1996b) and form \( P^* = P - \alpha P \cdot T_{max} \) and \( T_{min}^* = T_{min} - \alpha T_{max} \), where \( \alpha \) is the regression coefficient between the variable (P or Tmin) and Tmax. The correlations between \( P^* \) and \( T_{min}^* \) are presented in the last column of Table 1 (the individual station data) and for the gridded data in Fig. 2, where we see that it is generally positive at most sites, in support of hypothesis (c).

Fig. 2 Correlation between \( P^* \) and \( T_{min}^* \). These are the signals in P and Tmin remaining after variability coherent with Tmax has been removed. The 95 per cent statistical significance level is again approximately 0.24.

The correlations between the annual average SOI (which is based on a monthly index kindly provided by the Australian National Climate Centre, and represents the difference between normalised Tahiti and Darwin mean sea-level pressure) and (a) P, (b) Tbar, (c) Tmin, (d) Tmax and (e) the DTR are presented in Fig. 3. Greatest association with precipitation is primarily restricted to the eastern half of the continent. Positive SOI is associated with increased precipitation, and reduced maximum temperatures. These results are consistent with those obtained earlier (e.g. McBride and Nicholls 1983; Ropelewski and Halpert 1987, 1989). Halpert and Ropelewski (1992) examined the relationship between the SOI and Tbar after stratifying the data into two half-yearly periods. Their results are also consistent if we suppose that Tmax variability is more important in determining Tbar variability (which seems to be the case) and if the variability during December–June dominates the annual anomaly. Minimum temperature variability is not closely associated with fluctuations in the SOI.

Modelled relationship and the underlying mechanisms

In this section we analyse results obtained from the BMRC AGCM, first to see if the relationships are captured, but also to examine how the generally negative correlations between T and P arise. We will see that this relationship is typical of the situation in evidence over land elsewhere in the model, except at high latitudes. Here we will focus on the relationships evident between 40°S to 40°N as these are generally similar to those found over Australia.
Fig. 3 Correlation between the Southern Oscillation Index (SOI) and (a) P, (b) Tbar, (c) Tmin, (d) Tmax and (e) the DTR over Australia, for the period 1910-1992. The 95 per cent statistical significance level is again approximated by 0.24.

The model has been described in detail by Bourke et al. (1977), McAvaney et al. (1978, 1991), Bourke (1987), and McAvaney and Colman (1993). The equations of motion are solved using a spectral transform technique (Bourke et al. 1977), with rhomboidal truncation at wave number 31. Sigma coordinates are used in the vertical, with 18 unevenly spaced levels. Both the boundary layer and the vertical diffusion parametrisations are similar to those given by Louis et al. (1982). Penetrative convection is treated by a mass flux scheme (Tiedtke 1989) and shallow convection is the same as that described by Tiedtke (1984). Evaporation over the oceans is enhanced by employing a modified exchange coefficient at low wind
speeds (Miller et al. 1992). Gravity-wave drag is determined using the formulation of Palmer et al. (1986), while radiation is treated using a modified version of the Fels and Schwarzkopf (1975) scheme. The diagnostic cloud scheme calculates cloud fractions in three distinct layers, and the cloud fraction is evaluated in each of these layers from the relative humidity using simple formulae. Three different surface schemes are used, the first consisting of a two-level bucket hydrology and a three-level temperature scheme, the second with an additional moisture and temperature level, and the third incorporates the substantially more sophisticated land-surface scheme of Desborough (1997).

The model was integrated for ten years as part of the Atmospheric Model Intercomparison Project (Gates 1992), forced using observational estimates of sea-surface temperature and sea-ice distribution for the years 1979-1988. Three integrations were performed, each using a different land-surface scheme. Note that in the following discussion we will concentrate on surface temperatures, whereas the observations are for screen temperatures. Calculation of the correlation between the surface and lowest layer model temperature indicates that the two can be expected to track each other very closely.

The modelled relationships are presented in Fig. 4, where the same relationships are again in evidence. Results from the run with the simplest land-surface scheme will be presented, unless otherwise stated. Red shading indicates a correlation coefficient less than -0.3 and blue shading corresponds to a coefficient larger than +0.3. As the model was forced with daily averaged ocean temperatures with no diurnal cycle, ocean points are not considered in (b) to (d). All land points are considered. Correlations between P and all of Tmax, Tbar and the DTR are generally negative, whereas both positive and negative correlations are evident between P and Tmin. Positive correlations are in widespread evidence between P* and Tmin* (Fig. 5). Similar relationships are evident in the other two integrations with the exception of r(Tmin, P) which will be discussed below.

Fig. 4 Correlation between P and (a) Tbar, (b) Tmin, (c) Tmax and (d) the DTR in the AGCM. Only correlations exceeding 0.3 in magnitude are shaded. All land points are used (see Fig. 8 for indication of full coverage). All ocean points are masked out in (b)-(d) as the model was forced with daily averaged ocean temperatures only.

(a) AGCM: \( r(Tbar, P) \)

(b) AGCM: \( r(Tmin, P) \)

(c) AGCM: \( r(Tmax, P) \)

(d) AGCM: \( r(DTR, P) \)
Fig. 5  Correlation between P* and Tmin* in the AGCM. As in Fig. 2, P* and Tmin* represent the signal remaining after variability coherent with Tmax has been removed. Only correlations exceeding 0.3 in magnitude are shaded.

Before proceeding we need to assess the statistical significance of the modelled results. The first point to note is that similar plots were obtained in all three integrations. This increases confidence that they actually represent genuine model interrelationships, and are not merely fortuitous. Secondly, we only wish to characterise the response over 40°S to 40°N, away from high latitudes. In order to quantify the significance of the dominant relationships we therefore need to estimate the number of degrees of freedom exhibited between the various subregions within our region of interest (i.e., 40°S to 40°N), as each subregion provides an additional test of the hypothesis. Even if we suppose that there are only five independent regions over the domain (Africa, Asia, Australia, North and South America), which must surely be a conservative estimate, then the probability of having m (≤ n=5) regions with correlations generally exceeding 0.3 in magnitude is approximated by terms in the binomial expansion (p + (1-p))^n, where p is the probability that the correlation in a given subregion is merely fortuitous (e.g., Wonnacott and Wonnacott 1977). In the case of r(P, DTR) and r(P*, DTR*) m=5 and the probability that the consistent agreement is merely fortuitous is given by p^5. If we suppose that the persistence is small (which is reasonable in light of our earlier findings) and the effective number of degrees of freedom in each subregion is approximately 28 (N=2, where N is the total number of years in the three ten-year integrations) then r=0.3 corresponds to p=0.1 (using tables). So these two sets of results are significant at a level given by 0.15, which is well beyond the 99 per cent level. Alternatively we can regard each integration as a separate test, with N-2=8, p=0.43 and n=3*5, which again gives an alternative estimate above the 99 per cent level.

In the case of the correlation between P and Tbar, m=4, and the probability that this result is merely fortuitous is given by \[ C_4 = (1-p) p^4 + p^5 \], which also corresponds to a 99 per cent significance level. Even if we make allowance for a reduction in the effective number of degrees of freedom due to the presence of partially consistent SST-forced variability in the three runs by setting N=2-15, these results remain significant at a high level.

The correlation between Tmin and P in the integration with the sophisticated land-surface scheme appears different to the other two. This is illustrated in Fig. 6, where we see that there are very few correlations below -0.3. From this same perspective, therefore, the relationship between P and Tmin over 40°S-40°N is not statistically significant. Is this contrast itself statistically significant? For one run the probability that the near-total absence of negative correlations is fortuitous in a particular subregion is approximately 0.8. This figure, together with N-2=8 and n=5 gives a significance level of approximately 70 per cent. Allowing for partially consistent SST-forced variability in the three runs reduces this figure further, and so the result should be regarded as merely suggestive.

In order to help understand how the relationship between P and Tbar arises, consider the annual average surface heat budget over land:

\[ \text{SSWR} + \text{SLWR} + \text{LH} + \text{SH} + G = 0 \]

...1

where SSWR and SLWR are the (net) fluxes of short and long wave radiation, LH and SH are the turbulent fluxes of latent and sensible heat, and G is the subter-

Fig. 6  Correlation between P and Tmin in the AGCM using the land-surface scheme of Desborough (1997). Only correlations exceeding 0.3 in magnitude are shaded.
ranean heat exchange. All fluxes are positive if directed upwards away from the earth’s surface. Interannual variability in $G$ is small, generally restorative in nature, and is neglected in the following discussion.

The correlation between $T_{\text{bar}}$ and $SSWR$, $SLWR$, $SSWR+SLWR$, $LH$, and $SH$ were calculated in two of the runs (one with the simplest land scheme, the other with the most complex). Not surprisingly, $T_{\text{bar}}$ tends to increase over land as $SSWR$ into the surface increases (recall the sign convention). Similarly, $SLWR$ upward from the surface increases as $T_{\text{bar}}$ increases, indicating that changes to the upward $LWR$ flux outweigh the (probable) increase in the downwelled $LWR$ componen. $SSWR$ therefore helps to force changes in $T_{\text{bar}}$, while $SLWR$ acts to reduce the changes.

It is interesting to note that the upward latent heat flux decreases as the surface warms. This indicates that latent heat changes help to force the land-surface temperature rather than the other way around. This is in stark contrast to the situation over the tropical ocean where SST changes generally drive $LH$ changes. On the other hand, $SH$ upwards through the land surface increases as $T_{\text{bar}}$ increases, indicating that $SH$ is driven by changes in $T_{\text{bar}}$, thereby reducing fluctuations in $T_{\text{bar}}$.

In summary, $T_{\text{bar}}$ fluctuations are primarily driven by $SSWR$ and $LH$ variability, and are dampened by both $SLWR$ and $SH$ (and $G$) responses. The most obvious explanation for the out-of-phase relationship between $P$ and $T_{\text{bar}}$ (and $T_{\text{max}}$) lies with the expectation that increased $P$ is associated with increased cloudiness and reduced $SSWR$ into the land surface. This is supported by the fact that cloud cover is generally in-phase with precipitation, but out of phase with $SSWR$ into the surface.

In order to assess the relative importance of these two terms in driving temperature variations that are associated with $P$ changes, consider Fig. 7 in which we have plotted the magnitude of $LH(P)/SSWR(P)$, where $F(P)$ is defined as the covariance of $F$ and $P$, divided by the standard deviation of $P$. The results presented are from the run with the simplest land scheme, but were found to be very similar in the run with the most sophisticated scheme. At low latitudes in regions where deep convection is important $LH$ variability is more important than $SSWR$ variability, while further poleward the reverse is true. The first region extends over the ocean and into the northern mid-latitudes in the storm tracks.

The relative importance of $LH$ over $SSWR$ at low latitudes is consistent with the results obtained by Kleeman et al. (1996b) who looked at the ENSO response over the tropical Pacific Ocean. It is also interesting to note that the low latitude land areas in this plot where $LH(P) > SSWR(P)$ correspond closely to the regions of significant negative correlations between $T_{\text{min}}$ and $P$ (coloured red) in Fig. 4(b). This tendency to be out of phase is not generally evident in the observations in this region. At the same time $P^*$ and $T_{\text{min}}^*$ tend to be in phase in the model. Together these results suggest that the AGCM is overestimating the impact of $T_{\text{max}}$ on $T_{\text{min}}$, so that one or more of $LH(P)$ or $SSWR(P)$ is being overestimated in deep convective regions. The (suggestive) result discussed earlier with the more sophisticated land-surface scheme seems to indicate that it may have improved the model’s performance in this regard.

The contrasting relationships evident at high northern latitudes are probably due to the importance of snow variations there since snow will have a profound influence on the surface temperature (e.g., Watters et al. 1995). A second reason may lie with the fact that the latent heat flux is formulated in such a way that it is insensitive to soil moisture variability beyond a certain (large) threshold value. This threshold is presumably most frequently exceeded at high latitudes, where soil moisture is generally high.

Additional calculations suggest that $LH$ variability primarily arises from soil moisture changes. For example, the ratio of $LH$ fluctuations associated with soil moisture variability, to $LH$ fluctuations associated with wind-stress changes, i.e., $[LH(\text{soil moisture})/LH(\text{wind-stress})]$ (wind-speed was not readily available), generally exceeds unity south of about 30°N. This is also true in the run with the more sophisticated land-surface scheme (Desbrough 1997). These results are consistent with the results of Kleeman et al. (1996b) and Watters and Dix (1996). Kleeman et al. (1996b) showed that over the Pacific Ocean most of the $LH$ variability could be associated with variations in the difference between the saturation vapour pressure of the surface and the specific humidity of the overlying air, while Watters
and Dix (1996) showed that a regression formula relating spatial contrasts in the time mean latent heat flux was most closely associated with spatial contrasts in the time mean soil moisture. Note, however, that soil moisture (or LH) variability is not essential for the P/T relationship to arise - it simply enhances the magnitude of the change in T for a given change in P.

More recently, Watterson (1997) developed a theory relating the DTR to the sum of SSWR+LH+SH, showing that the spatial variations between these two quantities are remarkably coherent. Figure 8 indicates that the same is also true for the temporal variability. The relationship accounts for over 60 per cent of the variability evident over most of the earth away from high latitudes where the two are approximately out of phase.

**Recent changes in co-variability**

'All-Australia' indices (representing area-averaged Australian totals) for P and T were calculated by Nicholls et al. (1996a) using the methods described by Lavery et al. (1997), i.e., by combining the station data after weighting it according to the area each station was presumed to represent. Here we use the gridded data to provide an alternative estimate (the area-weighted sum of the gridded data) as a preliminary check. The resulting 'all-Australia' P indices are in very close agreement, with a correlation coefficient of 0.996.

Nicholls et al. (1996b) used the original indices to highlight recent changes in the statistical interrelationships during the last 20 years or so. This analysis is repeated here (Fig. 9), using indices based on the gridded data. The results are virtually identical to those obtained by Nicholls et al. (1996b): notice how in each scatter plot the black dots (corresponding to years 1971-1990) tend to occupy a different part of phase space, which may be indicative of a significant climatic shift perhaps due to some long-term climate vacillation or global warming.

In order to estimate how representative these indices are of regional variability they have been correlated with the gridded data. The results are presented in Fig. 10 for (a) P, (b) Tbar, (c) Tmin (d) Tmax and (e) the DTR. The

![Fig. 8](image_url) **Correlation between the DTR and SSWR+LH+SH in the AGCM.**

- **Rainfall vs. SOI**

- **Tmin vs. Rainfall**

- **Tmax vs. Rainfall**
indices are closely associated with central eastern variability, generally accounting for a smaller fraction of most near-coastal variability. This fraction is less than 20 per cent for P, and so the recent changes evident in the indices need not necessarily apply everywhere.

In order to determine where there have been unusual recent changes we performed regressions at every gridpoint for 20-year running blocks beginning in 1910 (e.g., 1910-1919, 1911-1920, ...). The regression coefficients were recorded (one for each 20-year block), and both the mean and the standard deviation of these coefficients calculated. The regression coefficients for the
period 1973-1992 (the most recent 20-year period available) were then calculated and compared to the mean of the coefficients obtained from the earlier 20-year blocks. The difference between the modern coefficients and the mean of the coefficients of the earlier periods were then compared with the standard deviation of the coefficients exhibited in the earlier period. In this way we can quantify how unusual the changes have been in terms of the earlier interdecadal variability. The use of all possible overlapping blocks is preferable to arbitrarily choosing four distinct 20-year blocks, as this could bias the magnitude of the variability. However, it is important to realise that this in no way overcomes the fundamental restriction that we have much less than 100 years of data to assess the level of naturally occurring interdecadal variability, and so the results presented in this section must be considered as suggestive only.

To illustrate the procedure, consider the regression between the SOI and P (Fig. 11), where it is supposed that $P = \alpha + \beta \cdot SOI$. The mean and standard deviation of $\alpha$ obtained from the earlier 20-year blocks are presented in Figs 11(a) and (b), while the value of $\alpha$ obtained from the regression using data from 1973-1992 only, is presented in Fig. 11(c). Notice that while Figs 11(a) and (c) are generally similar, there has been a shift to larger values over most of the continent. For example, if we concentrate on the point where the borders of Victoria, New South Wales and South Australia meet (indicated in Fig. 11(a)), we get: (a) just over 250 for the mean; (b) just over 15 for the standard deviation; and (c) approximately 350 for the modern value. Thus the latter is (well) over two standard deviations away from the mean and so we would regard this change as unusual. This is depicted by blackening in Fig. 12(a), where we see that $\alpha$ has changed to at least this degree over most of the continent.

While some of these regions overlap with regions where the observational density is low, significant change is apparent in other regions with high station density. This increases confidence that the apparent change is genuine and not merely due to observational error, although this possibility cannot be completely ruled out (Nicholls et al. 1996b). A similar plot for $\beta$ is given in Fig. 12(b). Significant change has occurred in isolated (and data-sparse) regions. Evidently the changes are primarily restricted to the constant ($\alpha$) in the regression formula rather than the gradient ($\beta$). This is consistent with the scatter plot using the indices presented in Fig. 9(a).

This same kind of analysis was then repeated on the regression between $P$ and $T_{\text{min}}$ (using $T_{\text{min}} = \alpha + \beta \cdot P$). The mean ($\alpha$) has changed significantly over most of the continent whereas the gradient ($\beta$) has not (Fig.13). This is also consistent with the scatter plot of the indices (Fig.9).

Finally, $P$ was regressed with $T_{\text{max}}$. Significant change has occurred in $\alpha$, whereas $\beta$ has changed significantly in only a few isolated regions (Fig.14). The bulk of the differences in $\alpha$ occur towards the west, although change has also occurred over about half of eastern Australia, but lies between 1 and 2 standard deviations of the mean (not shown).
Fig. 12  Regions (blackened) where the modern value of (a) $\alpha$ and (b) $\beta$ (in $P = \alpha + \beta \cdot \text{SOI}$) differs from its corresponding average by more than 2 standard deviations.

Fig. 13  Regression coefficients, $P = \alpha + \beta \cdot \text{Tmin}$: regions (blackened) indicate where the modern value of (a) $\alpha$ and (b) $\beta$ differs from the mean by more than 2 standard deviations.

Fig. 14  Regression coefficients, $P = \alpha + \beta \cdot \text{Tmax}$: regions (blackened) indicate where the modern value of (a) $\alpha$ and (b) $\beta$ differs from the mean by more than 2 standard deviations.
In summary, the earlier study performed by Nicholls et al. (1996b) using all-Australia indices adequately represents changes on a regional basis over a large part of Australia.

Discussion

The relationships between P, T and the SOI (and therefore predictive skill) can vary from decade to decade (see Allan et al. (1996) for a graphic illustration of decadal climate variability) due to (a) essentially chaotic variability which extends to decadal time-scales (e.g., Lorenz 1975; James and James 1992; Hasselmann 1976; Dix and Hunt 1995; Power et al. 1995a; Frederiksen et al. 1995; Rowell et al. 1995); (b) the existence of interdecadal modes of variability that have the potential to be predicted with some skill (Latif and Barnett 1994; Latif et al. 1997; Kleeman et al. 1996a); and (c) the evolving response to changes in the atmosphere's chemical composition due to anthropogenic influences including fossil fuel burning, land clearing and the release of sulphate aerosols (e.g., Bryan et al. 1982; Washington and Meehl 1989; Manabe et al. 1991, 1992; Mitchell and Johns 1996).

Since the early 1970s, we now expect greater rainfall for given values of the SOI, Tmin and Tmax (to a lesser extent). It is interesting to note that the change in a formulation like $P = \alpha + \beta \times SOI$ (for example) is with the constant $\alpha$ and not the gradient term $\beta$. This suggests that the interannual fluctuations in $P$ for a given SOI anomaly are more-or-less unchanged, but that there has been an overall increase in $P$. In fact Fig. 9 suggests that the SOI has actually dropped since the early 1970s. So the interrelationships between the changes in the 20-year means are quite unlike the interrelationships evident on interannual time-scales. This is consistent with climate change or a multidecadal climate fluctuation altering Australian rainfall (and T) and the SOI, but in a manner that is different to that normally associated with interannual tropical Pacific variability.

The presence of trends in Australian climate variables over the past century in Tmin, Tmax and the DTR has been noted previously (e.g. Torok and Nicholls 1996; Salinger et al. 1996; Karl et al. 1993). While trends in rainfall are more uncertain (Nicholls et al. 1996a), there appears to have been an increase in summer rainfall over eastern and northern Australia (Salinger et al. 1996). All of these trends are reflected in Fig. 9, with Tmin, Tmax and P increasing during the more recent period, but with Tmin rising more than Tmax.

These temperature changes are consistent with results obtained from coupled transient greenhouse experiments using coupled GCMs (e.g., Colman et al. 1995). There are, however, at least two difficulties with ascribing these changes to greenhouse warming. First, why are the changes not more gradual as they are in the models? More recent runs with time-dependent sulphate forcing indicate that the forcing can appear accelerated after the 1950s but this is largely confined to the northern hemisphere (e.g., G. Boer, personal communication). Secondly, and perhaps more importantly is the fact that precipitation changes over Australia from such models are very uncertain (IPCC 1996; Whetton et al. 1996). In these studies approximately half the models exhibited an increase in summer rainfall whereas the other half exhibited a decrease. Similar ambiguity arose during winter. So there isn't yet a consensus on what might actually happen to Australian rainfall under the influence of global warming. It may be that the models require further improvement especially with regard to their simulation of the ENSO phenomenon. Or perhaps longer integrations are required in order to obtain a statistically significant assessment of the actual response to greenhouse warming.

The importance of coherent and partially predictable variability in this context is equally uncertain. In the case of the north Pacific variability, evidence that it influences Australian rainfall has already been presented (Kleeman et al. 1996a; Latif et al. 1997; Power et al. 1998b), but it is unclear what impact this variability has on Australian temperature. We hope to investigate this in the not-too-distant future. Additionally, if this is dominating the changes, then why is the impact of this variability larger in recent years? Of course the variability is embedded in larger chaotic sources of variability and so this last concern cannot be used to discount north Pacific variability as a possible mechanism.

Steps to assess the relative importance of these mechanisms in driving the apparent changes seen here are ongoing. In BMRC, for example, we have recently constructed a new coupled GCM (Power et al. 1998a) which has a substantially improved oceanic component (Power et al. 1995b), and which seems to give more realistic interannual variability in the tropical Pacific that produces significant rainfall variability over Australia. This model is currently being used to estimate the response of ENSO and Australian interannual variability to global warming, and a report on these findings will be made in the near future.

Summary and conclusions

The best available precipitation (P) and surface temperature (T) records for Australia dating back to 1910 (Lavery et al. 1997; Torok, 1996; Torok and Nicholls 1996) have been examined to look for coherent interannual variability. P is generally out of phase with daily
maximum temperature, Tmax, the daily average temperature, Tbar (estimated here as the average of the Tmax and the daily minimum, Tmin), and the diurnal temperature range (DTR, Tmax-Tmin). Association between P and Tmin is generally weak. It was found, however, that this is due to the partial dependence of Tmin on the temperature at sunset. After removing the signal in both P and Tmin which is coherent with Tmax, the resulting variables (P* and Tmin*) were shown to be coherent, consistent with hypothesis 3 outlined in the introduction.

Only a small subset of the stations produced both P and T data and so the data were gridded to maximise the region over which co-variability could be analysed. The results obtained with the gridded data showed close agreement with the station data, apart from a near-coastal band along which the coherences discussed here were subdued in the station data but not in the gridded data. The coherence was not evident at the two island stations (Gablo Is., Vic., and Rottnest Is., WA) available, nor at the exposed sites of Wilson’s Promontory or Cape Schanck (both in Vic.).

Results from three ten-year integrations of an atmospheric general circulation model (McAvaney and Colman 1993) were also presented. The model was forced using observational estimates of the sea-surface temperature and sea-ice distribution for the period 1979-1988 as part of the Atmospheric Model Intercomparison Project (Gates 1992). The situation over Australia was found to typify the situation over most land points between 40°S and 40°N, with the exception of North Africa.

The model exhibited the same general tendency for P to be out of phase with Tmax, Tbar and the DTR, but not with Tmin, and for P* to be in phase with Tmin*. The surface temperature changes associated with rainfall were primarily driven by SSWR (surface short wave radiation) and LH (latent heating) changes, with the other terms in the surface heat budget generally acting to reduce the magnitude of the temperature excursions. LH was found to provide a larger source of heating than the SSWR changes (again in association with changes in P) in low latitude regions of deep convection. The hierarchy is reversed away from these regions. Temporal variability in evaporative cooling appeared to be primarily driven by soil moisture change. A possible model bias evident in the relationship between P and Tmin appeared to be corrected by the inclusion of a more sophisticated land-surface scheme (Desborough 1997). The contrasting statistical nature of the variability evident at near-coastal sites in the observational record was not apparent in the AGCM. This suggests that down-scaling methods may prove useful at such locations for constructing climate change scenarios or predicting seasonal rainfall and temperature anomalies, when using results obtained from the coupled atmosphere-ocean model mentioned earlier (Power et al. 1998a), as it contains a version of the AGCM employed here.

The third part of this study was concerned with apparent recent changes in some of the relationships exhibited between observed P, T and the Southern Oscillation Index (SOI), first noted by Nicholls et al. (1996b) using ‘all-Australia’ indices for T and P. We examined these changes on a regional basis and quantified the changes in terms of the interdecadal variability evident in the records prior to 1972. Variability between 20-year blocks (1910-1919, 1911-1920, ...) were used to show that the conclusions drawn on the basis of the indices were found to have broad applicability: regression coefficients calculated using variability since the early 1970s differ from their long-term mean values by more than two standard deviations over much of Australia. The change is primarily restricted to the mean coefficient and not the gradient (i.e., $\alpha$ in $P = \alpha + \beta * SOI$, and not $\beta$). One interpretation of this is that the interrelationships between the recent changes in PT and the SOI differ from the corresponding interrelationships evident on interannual time-scales. These changes include an increase in P for a given SOI and an increase in Tmin for a given P. Changes in Tmax, for a given P, were also significant at this level over much of the western half of the continent. Additional, less marked changes were evident towards the east. While these changes may be due to climate change, the possibility that it is instead associated with ‘naturally’ occurring interdecadal variability or perhaps even with observational problems (Nicholls et al. 1996b) cannot be ruled out on the basis of this study alone.

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