A Historical Perspective on Southeastern Australian Rainfall since 1865 Using the Instrumental Record

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ABSTRACT

The instrumental record for rainfall across Australia is regarded as being sufficiently reliable to produce national monthly gridded rainfall analyses from 1900 onward. Prior to 1900, the rainfall gauge network is much sparser. The possibility of using those nineteenth-century observations that do exist to construct an estimate of rainfall across the southeastern part of Australia (SEA) is explored by constructing a network based on 11 locations comprising either single observing sites or composites of nearby observing sites with long continuous records. It is shown that, during the period 1900–2010, the monthly rainfall reconstruction based on this network captures 98% of the variability of SEA monthly average rainfall. This network, which extends back to 1865, provides a useful view of the Federation Drought, making a comparison possible with other long-term droughts observed in SEA, around the time of the Second World War and the Millennium Drought from 1997 to 2009. A comparison of these three historical low-rainfall periods was conducted using the drought–depth–duration criteria: the ongoing decline in southeastern Australia is seen as being notably worse than the previous two historical droughts. The network also provides an insight into the decadal variability of SEA rainfall in the later part of the nineteenth century; it includes a high peak in the 1870s comparable to similar wet decadal peaks in the 1950s and 1970s. The implications of this longer perspective on the decadal variability in southeastern Australia in light of the current understanding of the ongoing rainfall deficit are discussed.

1. Introduction

Southeastern Australia (SEA), defined here as continental Australia south of 33.5°S and east of 135.5°E together with Kangaroo Island (South Australia), had between 1997 and 2009 its lowest 13-yr rainfall total in the entire 1900–2010 instrumental period. The 1997–2009 average annual total of 512.4 mm was 11.9% below the 1900–2010 average annual total of 581.7 mm. Figure 1 shows rainfall deciles across Australia for the 13-yr period 1997 to 2009, calculated using the Australian Bureau of Meteorology’s (BoM’s) high-resolution (0.05°) operational gridded monthly rainfall analyses (Jones et al. 2009). The calculation compares the 13-yr 1997–2009 rainfall total against all analogous 13-yr periods from 1900–12 to 1998–2010. Nationally, 3.8% of Australia was lowest on record in the calculation; corresponding state results were 61.4% (Victoria), 33.6% (Tasmania), and 7.8% (New South Wales). The term Millennium Drought (MD) has been used extensively by others (e.g., Leblanc et al. 2012) to describe this period of extremely low rainfall across southeastern Australia from 1997 to 2009 (i.e., the drought that was in progress on 1 January 2001); we will use the same nomenclature throughout this paper.

Across SEA (see the blue box in Fig. 3), the deficit of 69.3 mm (11.9%) is 49.7% larger than the deficit for the next driest 13-yr period. Excluding periods that overlap 1997–2009, this next driest period is 1937–49 (average annual rainfall total of 535.4 mm, and deficit of 46.3 mm or 8.0%). The slightly earlier period 1933–45 is close enough (average annual rainfall total of 535.8 mm) for these two overlapping 13-yr periods to be regarded as statistically indistinguishable. It should be noted that 1994–2006 (526.5 mm) and 1996–2008 (523.2 mm) were both drier than 1937–49. On the other hand, 2010 was quite wet across SEA, with an annual rainfall total of 809.2 mm.

Low rainfall periods similar to the MD were seen in the instrumental record during an extended period
preceding and encompassing World War II, from 1935 to 1945 (hereafter called the WWII drought or WWIID) and around the time of the federation of the Australian States (which occurred in January 1901). During the 1935–45 period, the average annual rainfall total of 518.9 mm was a severe 10.8% below the 1900–2010 average, nearly as severe a decline as that of the present drought, although over a slightly shorter duration. The duration and intensity of the MD is therefore unprecedented in terms of the period covered by the national gridded rainfall analyses. The Federation Drought (FD), straddling the start of the gridded analyses (January 1900), is too early to be included in this comparison.

The rainfall decline in SEA is happening during the cooler part of the annual cycle, affecting a continuum of months from March to October dominated by change in autumn [March–May (MAM)] (Timbal 2009). That decline is reminiscent of a similar decline that occurred in the southwest of Western Australia (SWWA) in the 1970s. Rainfall in the two regions is linked on daily to interannual timescales (Hope et al. 2010). However, the rainfall deficit in these two regions has been observed to start at different times; in the 1970s in SWWA and in the 1990s in SEA. With rainfall across much of southern Australia now in deficit (Fig. 1), that raises the possibility that the observed increase in mean sea level pressure (MSLP) (Timbal and Hope 2008) at the latitudes of southern Australia is driving the rainfall anomalies (Nicholls 2010).

Timbal and Hendon (2011) have shown that the observed rainfall deficiency across SEA could not be accounted for by naturally occurring modes of variability generated within the tropics such as the El Niño–Southern Oscillation and the Indian Ocean dipole, which were previously suggested as a plausible explanation (Ummenhofer et al. 2009), although it has been known for some time that tropical modes of variability, including those in the Indian Ocean, do not relate well with autumn rainfall in SEA (Timbal and Murphy 2007). In light of this, autumn is the critical season to understand the Millennium Drought in SEA (Timbal 2010).

Contrary to tropical modes of variability being a plausible mechanism for the MD, Timbal et al. (2010) demonstrated that the rainfall deficit observed across SEA is related to an observed strengthening of the belt of high pressure at the longitude of the eastern part of the Australian continent (Drosdowsky 2005). This belt, the subtropical ridge (STR), is the key controller of

Fig. 1. Australian rainfall deciles for the 13 years 1997 to 2009, calculated using the Bureau of Meteorology’s operational 0.05° resolution analyses.
rainfall across southern Australia. The twentieth-century evolution of the ridge has been observed to evolve alongside observed global warming (Timbal and Drosdowsky 2012). Externally forced climate model simulations confirm that coevolution was not observed by chance and reproduce a strengthening of the ridge (albeit not as strong as observed) only if anthropogenic external forcings are used (Timbal et al. 2010). In addition, Timbal et al. (2010) noted that the rainfall in SEA appears to be related to the global warming over the instrumental record, with periods of lower rainfall corresponding to periods when the global temperature of the planet was increasing, and periods with the wettest decades (e.g., 1950s and 1970s) corresponding to a period where no global warming was observed. This simple observation suggesting a possible linkage between the pace of global warming and SEA rainfall was limited because of data availability to the twentieth century when only three separate periods can be observed (i.e., warming from 1900s to 1940s, stabilization during the 1950s to 1970s, and warming since the mid-1970s).

As part of the South East Australia Climate Initiative (SEACI), the MD was compared to the WWIID in detail (Murphy and Timbal 2008; CSIRO 2010). Additionally, an attempt was made to compare both to the FD but, as previously mentioned, both the current BoM national gridded monthly rainfall analyses (Jones et al. 2009) and their predecessors (Jones and Weymouth 1997) only extend back to 1900 because of insufficiently widespread observations at the national scale prior to that. Comparisons with the FD of 1895–1902 are therefore difficult to do with consistent data and require the usage of the limited station data (Verdon-Kidd and Kiem 2009) that are available.

From these various perspectives, it is apparent that extending the observational record as far back in time as possible is a useful contribution to understanding the ongoing rainfall deficit. Trewin and Fawcett (2009) reconstructed pre-1900 Murray–Darling basin (MDB) average monthly rainfall using a network of long-reporting sites and composites thereof. That methodology has been adopted here in its essentials.

Other attempts have been made using paleodata to reconstruct past rainfall in the Australian region. In many instances, the proxies used are remote from the area and the reconstructions are based on observed teleconnections between regional rainfall and large-scale modes of variability: for example, Verdon and Franks (2006), who looked at Pacific decadal oscillation proxies to infer Australian rainfall; Lough (2007), who attempted to reconstruct far North Queensland rainfall, and McGowan et al. (2009), who rely on a single proxy in China to infer Murray–Darling outflow. Some of these attempts, while using paleoclimatic proxy data, also make use of the early instrumental record for calibration and verification purposes—for example, Gallant and Gergis (2011), who reconstruct River Murray streamflow from the late eighteenth century; also, a similar methodology is used by Gergis et al. (2012) to reconstruct SEA rainfall.

Gergis et al. (2012) is of particular interest for this study, and throughout this paper we will describe the similarities and differences between the two reconstructions as appropriate. A difference worth pointing out at this stage is that whereas we define southeastern Australia as continental Australia south of 33.5°S and east of 135.5°E together with Kangaroo Island (South Australia), Gergis et al. (2012) used a slightly larger domain: east of 135°E and south of 33°S and including Tasmania.

The structure of the paper is as follows: Section 2 describes the datasets used and the methodology. Section 3 describes the network-based reconstruction of SEA monthly and annual rainfall. In section 4, we describe the characteristics of the interannual and decadal variability with a particular focus on the new perspective for the nineteenth century (e.g., the FD and some very wet decades prior). In section 5, an in-depth comparison of the three long dry periods is performed using a drought–depth–duration concept. Conclusions arising from this work are then provided in section 6.

2. Data and methodology

We use the current BoM operational high-resolution monthly rainfall analyses (Jones et al. 2009), generated as part of the Australian Water Availability Project. These analyses are at 0.05° × 0.05° resolution, or approximately 5 km × 5 km. The analysis methodology employed in these analyses is a hybrid one, merging two-dimensional Barnes successive correction analyses (Jones and Weymouth 1997) of fractions of monthly mean rainfall and three-dimensional thin-plate smoothing spline analyses (Hutchinson 1995) of climatological monthly rainfall. The number of monthly station rainfall reports varies considerably across the period of record (1900–2011), rising from around 3000 stations per month in 1900 to around 6000 to 7000 stations per month in the second half of the 112-yr period [see Jones et al. (2009) for more detail].

A monthly SEA average rainfall time series for rainfall was calculated by averaging the high-resolution monthly analyses over continental Australia (the region east of 135.5°E and south of 33.5°E; see the box in Fig. 3). Kangaroo Island (South Australia) is also included in the averaging region. The averaging process takes into
account meridional convergence, but is otherwise a straight nonweighted averaging of the grid values.

Because the gridded averages only extend back to 1900, in this study we perform a reconstruction of the SEA monthly gridded averages using a network based on a limited number of sites with long records extending well back into the nineteenth century, with a view to capturing the monthly variability in SEA average rainfall through the Federation Drought. In doing this, we follow the methodology described by Trewin and Fawcett (2009) for the MDB. Following Trewin (2012), we use “site” to denote a single BoM observing site or station, and “location” to denote a particular place whose rainfall data are used in the reconstruction. A location therefore denotes either a single site with complete data or a composite of two or three different sites with incomplete data. Each site has a unique BoM numerical identifier or station number.

A range of possible models was considered based on 21 locations across the SEA region with data start dates ranging from 1850 to the mid-1870s. To choose the optimum model, a trade-off between the starting year of the reconstruction and the amount of variance explained was made (Fig. 2). An important reason for reconstructing at least as far back as 1870 will be seen subsequently in the results (see section 4), namely, the relative wetness of the 1870s, a perspective which differs from previous reconstructions, in particular Gergis et al. (2012). Networks were constructed starting from the simplest one based on a single location which opened in 1850. Then locations were added for the year they became available and more complex networks were developed. All locations available from a particular start year were included in the regression, whether or not their contribution was statistically significant. The number of locations included and the relative residual unexplained variance are plotted against start year (Fig. 2). The relative unexplained variance is computed as $1 - R^2$, where $R^2$ is the multivariate coefficient of determination of the regression (before the multiplicative scaling employed to remove any residual bias).

We chose the start year 1865 as providing an optimal trade-off between reconstruction skill and temporal extent of the reconstruction. In the 1865 network comprising 12 locations, just under 2% of the variation is left unexplained by the regression, and the contribution of one location (Melbourne) to the regression is not statistically significant. Its omission from the regression has hardly any impact on the fraction of variance explained. Hence the chosen network starts in 1865 and actually uses 11 locations.

The chosen network, with 11 locations, starts in 1865, explains about 98% of the monthly variance in the SEA monthly gridded rainfall totals; it comprises four single-site locations with complete records between 1865 and 2010, five composites of pairs of sites, and two composites.
of three sites. The locations used (11 in total, once Melbourne is removed) along with their contributing sites are listed in Table 1 and mapped in Fig. 3. In the case of the composites (denoted by large blue circles in Fig. 3), the first station number mentioned is taken as the primary site—in other words, the compositing process estimates a rainfall time series for the primary site.

To generate rainfall time series for composite locations, the monthly time series of the first (primary) site is linearly regressed (with a zero intercept imposed) against the monthly time series of the second site across all months with concurrent data, and the regression used to “predict” missing values in the monthly time series of the first site. For composites of three sites, this process is applied twice (i.e., iteratively), using the second site to infill gaps in the first time series, and subsequently the third site to infill gaps in the composite time series of the first and second sites. The zero intercept is used to

<table>
<thead>
<tr>
<th>Location</th>
<th>Bureau station number</th>
<th>Name (state)</th>
<th>Starting year</th>
<th>Monthly $r^2$ and (months filled)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>023000/023011</td>
<td>Adelaide (SA)</td>
<td>1850</td>
<td>0.98 (382)</td>
</tr>
<tr>
<td>2</td>
<td>086131/086125</td>
<td>Yan Yean (VIC)</td>
<td>1856</td>
<td>0.96 (24)</td>
</tr>
<tr>
<td>3</td>
<td>087046</td>
<td>Mount Buninyong (VIC)</td>
<td>1857</td>
<td>n/a</td>
</tr>
<tr>
<td>4</td>
<td>070009</td>
<td>Bukalong (NSW)</td>
<td>1858</td>
<td>n/a</td>
</tr>
<tr>
<td>5</td>
<td>066062</td>
<td>Sydney (NSW)</td>
<td>1859</td>
<td>n/a</td>
</tr>
<tr>
<td>6</td>
<td>026020/026021</td>
<td>Mount Gambier (SA)</td>
<td>1861</td>
<td>0.97 (433)</td>
</tr>
<tr>
<td>7</td>
<td>081003/081086/081083</td>
<td>Bendigo (VIC)</td>
<td>1862</td>
<td>0.95 (220), 0.91 (2)</td>
</tr>
<tr>
<td>8</td>
<td>090015</td>
<td>Cape Otway (VIC)</td>
<td>1864</td>
<td>n/a</td>
</tr>
<tr>
<td>9</td>
<td>075048/077025</td>
<td>Swan Hill (VIC)</td>
<td>1865</td>
<td>0.91 (154)</td>
</tr>
<tr>
<td>10</td>
<td>021001/021054</td>
<td>Auburn (SA)</td>
<td>1865</td>
<td>0.96 (2)</td>
</tr>
<tr>
<td>11</td>
<td>089000/079019/089085</td>
<td>Ararat (VIC)</td>
<td>1865</td>
<td>0.93 (491), 0.94 (9)</td>
</tr>
</tbody>
</table>

FIG. 3. Locations used to form the SEA network. Numbers in the figure correspond to entries in Table 1; composites made of two or three individual sites are indicated when several markers are apparent within a circle. (See Table 1 for details.) The large black box denotes the area for which the reconstruction is performed (continental Australia east of 135.5°E and south of 33.5°S including Kangaroo Island in South Australia).
preserve dry months and to prevent the infilling process from generating negative monthly rainfalls. The squares of the correlations of these regression models are shown in Table 1, together with the numbers of infilled monthly values. The first infilling may be used to infill hundreds of missing monthly values, while the second infilling is only used to infill small residual gaps. This regression-based approach amounts to the application of a homogeneity adjustment to the rainfall data of the secondary sites.

Sites in the reconstruction were chosen for a reasonable geographic spread across the study region, long records extending well into the nineteenth century (either complete records, or near-complete records with potential to be infilled via regression against a nearby site), and statistical significance in the multivariate linear regression used to model the pre-1900 rainfall variability. Candidate secondary sites for the infilling process were chosen for closeness to and a large degree of overlap with the primary site. It is interesting to note, that while 11 locations are used in this study only one site (Yan Yean, BoM station number 086131) is used by Gergis et al. (2012) to validate their paleo-reconstruction.

The monthly rainfall data for the individual sites used in the network reconstruction are taken from the Australian Data Archive for Meteorology (ADAM), managed by the BoM. As such, they have been subjected to some degree of quality control (particularly so in the case of recent data), but have not been subjected by the authors to any homogenization adjustments. Only two (Bendigo 081003 and Yan Yean 086131) of our 11 primary sites are included in the Lavery et al. (1997) high-quality rainfall dataset, which lists very few sites with pre-1865 data. That study excluded capital city sites (e.g., Adelaide and Sydney) on the basis of a perceived urbanization effect, particularly in summer storm activity; however, two of these sites are included in our reconstruction and are likely to be of very high quality as well. The quality of the rainfall data for the individual sites therefore represents a caveat on the results of this study, although it should be noted that similar caveats would also apply to the site data used in the gridded monthly analyses themselves.

Only one location (Bendigo) is common between this SEA network and the MDB network of Trewin and Fawcett (2009). This reflects both geography and data availability. In terms of geography, a substantial fraction of the SEA region does lie within the MDB, including most of northern Victoria and adjacent parts of southern New South Wales. The MDB, however, is much larger and in essence more northerly region, extending well into Queensland, thereby justifying a separate and independent reconstruction for the SEA region. Some locations used in the MDB reconstruction—Wagga Wagga (1872), Deniliquin (1870), and Wentworth (1868)—were considered in the construction of the SEA network (and actually used in the calculation summarized in Fig. 2), but their data (start years as indicated) do not extend back far enough for inclusion in the 1865 network. Other locations, such as Sydney, Adelaide, Mount Gambier, and Cape Otway, lie outside the MDB, which for most of its extent lies north and/or west of the Great Dividing Range. The coastal regions of Victoria and southern New South Wales, included in the SEA region, on the other hand lie south and/or east of the Great Dividing Range.

To reconstruct pre-1900 monthly rainfall averages across SEA, gridded monthly rainfall averages across the SEA region are computed from the high-resolution (0.05°) operational gridded monthly rainfall averages. The resulting time series is regressed against the 11 monthly rainfall location time series across the period 1900–2010. A zero intercept is imposed in the multivariate linear regression model to preserve the possibility of a completely dry month across the study region. The regression model is then used to estimate pre-1900 SEA monthly averages.

To compare the different droughts, we introduce the concept of “drought–depth–duration” (DDD) for periods ranging from 1 to 21 years. The main purpose is the ability to compare dry epochs of different duration knowing that parts of the natural and man-made environment will respond to dry periods of different durations. DDDs are computed from an annually resolved time series \( r_1, \ldots, r_n \) of length \( n \) years, which may be annual, seasonal, or even monthly rainfall. For our SEA annual rainfall reconstruction, \( n = 146 \). Over a given 21-year period \( r_{k}, \ldots, r_{k+21} \) within the time series, we calculate the following statistics:

\[
\begin{align*}
    d_1 &= \min(r_k, \ldots, r_{k+20}), \\
    d_2 &= \min((r_k + r_{k+1})/2, \ldots, (r_{k+19} + r_{k+20})/2), \\
    d_3 &= \min((r_k + r_{k+1} + r_{k+2})/3, \ldots, \frac{r_{k+20} + r_{k+19} + r_{k+20}}{3}), \\
    d_{20} &= \min((r_k + \cdots + r_{k+19})/20, (r_{k+1} + \cdots + r_{k+20})/20), \\
    d_{21} &= (r_k + \cdots + r_{k+20})/21.
\end{align*}
\]

In other words, \( d_p \) is the smallest average of \( p \) consecutive values in the 21-year period. With \( r = (r_1 + \cdots + r_n)/n \) as the average value of the entire time series, we calculate \( D_p = 100(r - d_p)/r \) as the percentage reduction below the long-term average of the driest \( p \)-year period.
in the 21-yr assessment period. The set of values \((D_1, \ldots, D_{21})\) represents the drought–depth–duration for the 21-yr period. (This concept is illustrated for several periods in Fig. 8 and results are discussed in the relevant section.) The drought–depth–duration idea could obviously be applied to periods other than 21 years (e.g., 25 or 30 years). A length of 21 years was chosen as it offers a long enough period that can affect the natural environment as well as how we interact with it (e.g., farming practices, water management, and so on).

The statistical significance of the DDD for a particular 21-yr period can be assessed in two ways, both involving Monte Carlo simulation: (i) by comparison against randomly chosen 21-yr periods and (ii) by comparison against the driest 21 years in the \(n\)-year period. In each case, we take the original time series and generate synthetic \(n\)-year time series by sampling with replacement. This assumes that there is no significant autocorrelation in the original time series needing to be taken into account. In the first way, random 21-yr periods are selected and their DDD calculated. The one-tailed probability \(\text{Pr}(D_p > D_p^*)\) under the simulation is calculated, where \(D_p^*\) is the value obtained from the original sample. In the second way, the driest 21-yr period in each synthetic time series is found and its DDD calculated, but otherwise the calculation of the one-tailed probability is the same. In both cases, it is also useful to calculate the mean DDD under the simulation. In the first way, theoretical considerations lead to the expected value of \(D_{21}\) being zero. The reason for calculating the second way is to take into account the fact that certain periods (e.g., the FD, the WWIID, and the MD) may be “cherry-picked” as being particularly dry, and this a priori selection needs to be taken into account in the estimation of statistical significance. The existence of the recent long dry period described as the MD has after all played a significant role in motivating this study (i.e., the focus on 13-yr periods as used earlier in this study arises from the dryness of the 13-yr period 1997–2009).

3. Validation of the network

Ensuring that the optimum network describes well the rainfall in SEA as estimated by the 0.05° gridded analyses is an important step. In reconstructing pre-1900 SEA monthly rainfall averages, the gridded monthly rainfall averages (1900–2010) were regressed against the 11 monthly location time series, via ordinary linear regression with a zero intercept. The multiple \(R^2\) in the regression was 0.98 (Fig. 4, left panel). The corresponding annual value was 0.998. All 11 predictors were statistically highly significant in the linear regression, with the largest of the 11 two-tailed \(p\) values for the regression coefficients being less than 0.0012. We considered and rejected the use of a Melbourne site (BoM station number 086071), with data from 1856 and located a little to the south of Yan Yean (site 2) in Fig. 3, as a potential predictor in the regression because its contribution to the regression was not statistically significant, most likely because of its close proximity to Yan Yean. The amount of annual variability captured by this reconstruction (99.8%) is higher than the nine-station rainfall network used by Gergis et al. (2012) for additional verification back to 1873 (86%). That number is different from the amount of interannual rainfall variability captured by the paleo-reconstruction in Gergis et al. (2012), which captures 33% of interannual variations in the SEA area average calculated from the Australian Water Availability Project (AWAP) rainfall grid over the 1900–88 period. That much lower number
is not surprising since this reconstruction relies solely on paleoclimatic proxies that depict modes of variability relevant to SEA rainfall. While these modes are important, the amount of explained variance is rarely above 20%–30%; in addition, there are biological reasons why such paleo-based reconstructions cannot always capture interannual variability. A more relevant comparison can be done once decadal variability is considered (see section 4). Nevertheless, it is worth noting that the largest amount of annual variability captured here with the set of locations chosen suggests that it may possible to improve the Gergis et al. (2012) paleo-reconstruction using the extended historical rainfall network presented here for calibration.

Besides annual total rainfall, it is also of interest to evaluate the ability of the optimum network to describe higher frequency (i.e., monthly total rainfall). This is particularly important to be able to analyze decadal variability and periods of low rainfall from a seasonal perspective (sections 4 and 5). Across the period 1900–2010 in the monthly reconstruction, the mean error (or bias) is $-2.9$ mm yr$^{-1}$ (the reconstruction being slightly drier than the gridded averages). We therefore remove this bias multiplicatively, resulting in a reconstruction with zero bias. Figure 4 (left panel) shows a scatterplot of the gridded and reconstructed SEA monthly rainfall totals for the period 1900–2010, while the right panel shows a scatterplot for the corresponding annual totals. Therefore the resulting network is unbiased and reproduces as much as 98% of the observed monthly rainfall variability as reproduced by the gridded averages.

Although the reconstructed network is unbiased, annual and monthly errors can at times be large, which has to be kept in mind when analyzing rainfall variability provided by the network. The cumulative distribution function of the reconstruction errors in the SEA monthly rainfall reconstruction over the period 1900–2010 is shown in Fig. 5. The reconstruction error is defined as the difference between reconstructed monthly total and the gridded monthly total, with positive (negative) values meaning the reconstruction is wetter (drier) than the gridded analysis. The corresponding annual reconstruction errors, divided by $12^{-1/2}$ to place them on a comparable scale with the monthly errors (in accordance with the central limit theorem of statistics) are also shown. The monthly reconstruction errors range from $-36.2$ to $+35.4$ mm, with 90% of the values lying between $-12.6$ mm (5th percentile) and $+11.4$ mm (95th percentile). By construction, the reconstruction is unbiased, but there is a slight skewness in the errors (the median error is $+0.5$ mm). The annual reconstruction errors range from $-61.2$ to $+67.1$ mm, with 90% of the values lying between $-46.0$ mm (5th percentile) and $+52.2$ mm (95th percentile). The median annual error is $+2.2$ mm.

The reconstruction error statistics reported in the previous paragraph are dependent or in-sample errors, in that the data used in the verification process are also used in the model construction process. We therefore
supplement these error statistics with the calculation of independent or out-of-sample reconstruction errors, calculated by cross-validation. We leave out each year’s worth of monthly data between 1900 and 2010 in turn, regenerate the 11-location 1865 model using the remaining years of data, and compute an out-of-sample reconstruction estimate (and hence a reconstruction error) for the SEA region annual rainfall for the omitted year. [The leaving out of one year at a time involves the implicit assumption that the time series of annual rainfall is not strongly autocorrelated. If the time series does have significant autocorrelation, then leaving out one year at a time would likely be insufficient to result in independence of the validation process.] The cumulative distribution function for these cross-validated error estimates is also shown in Fig. 5 (green line). Because of the size of the dataset used in the network reconstruction, the differences between the cross-validated and not cross-validated annual totals are small: the mean absolute difference between these two sets of annual totals is 0.5 mm. The range of the cross-validated errors is not strongly autocorrelated. If the time series does have significant autocorrelation, then leaving out one year at a time would likely be insufficient to result in independence of the validation process. The cumulative distribution function for these cross-validated error estimates is also shown in Fig. 5 (green line). Because of the size of the dataset used in the network reconstruction, the differences between the cross-validated and not cross-validated annual totals are small: the mean absolute difference between these two sets of annual totals is 0.5 mm. The range of the cross-validated errors is ~62.2 to +68.2 mm with 90% of them lying between −46.9 and +53.1 mm. These are slightly larger than the not cross-validated results reported in the previous paragraph.

Two additional assessments were performed, to explore the sensitivity of the reconstruction to the presence of an extremely dry period and an extremely wet period in the input data. In the first one the years 1997 to 2009, corresponding to the driest 13-yr period, were omitted from the data feeding into the reconstruction, and those omitted years were compared against the original reconstruction. The omission of the years made the reconstruction over these years wetter by 0.6 mm yr\(^{-1}\). In the second assessment the years 1946 to 1956, corresponding to the wettest 11-yr period, were omitted. The omission of the years made the reconstruction of over those years wetter by 1.7 mm yr\(^{-1}\). In both instances reconstruction errors are small and as both errors are toward the wetter side these results do not suggest a systematic underestimation of extreme periods.

Lastly, when the monthly modeling errors (calculated as modeled – gridded) are stratified against the gridded monthly totals, there is a tendency toward positive biases in the modeling errors for small monthly gridded rainfall values (up to around 50 mm month\(^{-1}\)) and increasingly negative biases for larger monthly gridded values (not shown). When this stratification is instead performed with respect to the reconstructed monthly totals, this tendency disappears.

Knowing precisely the uncertainties attached to the network reconstruction for both annual and monthly total rainfall, we now proceed to the analysis of the climate variability with a focus on the additional 35 years of data prior to 1900 that this network provides.

4. Instrumental observation of SEA rainfall variability from 1865

The annual rainfall time series for the SEA region, as reconstructed from the 11-location network, across the period 1865 to 2010, is plotted (Fig. 6) together with the 11-yr running mean (black line) and the corresponding 11-yr running mean of the high-resolution (0.05°) gridded analyses (red line). The two 11-yr running mean curves are generally very close to each other, although there are two periods (the 1950s and the 1990s) when they separate slightly. Nevertheless, Fig. 6 shows that the 11-location network adequately captures the decadal variability in SEA rainfall.

In the reconstruction, the driest year is 1982 (329.1 mm) and the wettest year is 1870 (922.2 mm). The all-years (1865–2010) mean rainfall is 586.9 mm. The driest year in the gridded averages (1910–2010) is also 1982, but since it does not go back in time past 1900, 1974 is the wettest year at 870.7 mm, with 1956 next at 867.3 mm. Both 1974 (850.2 mm) and 1956 (857.6 mm) are also very high rainfall years in the network reconstructed series, and 1956 is the wettest year in the reconstruction for the overlapping 1900–2010 period.

In terms of decadal variability, the reconstruction implies that the 1870s (average annual rainfall of 676.1 mm for 1870–80) were comparable to the mid-1940s–mid-1950s (average annual rainfall of 678.3 mm for 1946–56) for overall wetness. It is worth noting that when the gridded rainfall is used, the highest 11-yr period was also 1946–56 but with a total of 664.2 mm. It is unclear why the 11-yr running mean has such a large difference (13.9 mm) for the 1946–56 period, which is surprising considering the distribution of annual total error (Fig. 5). It is possibly due to a systematic overestimation of SEA rainfall by the network during high rainfall years and hence the same bias may exist for the very wet 1870s. Hence this result is best phrased by saying that the 1870s and the mid-1940s–mid-1950s appear to be equally wet 11-yr periods, with the difference being within the margin of error. Both periods are very wet due to a succession of La Niña events, some being major [see Garden (2009) for the 1870s]. In Gergis et al. (2012), the overlapping period of 1865 to 1900 leads to some noticeable differences: while both reconstructions agree on a wet decade in the 1890s, the 1870s are not abnormally wet in Gergis et al. (2012) whereas, as discussed above, it is one of the wettest decades in the present rainfall reconstruction using early instrumental data. This difference is surprising; since the Gergis et al. (2012) reconstruction...
reproduced 72% of the decadal variability, which is remarkable for a paleo-based reconstruction, it may reflect differences due to conditions captured outside of our target domain (e.g., Tasmania). It is worth noting, though, that Gergis et al. (2012) provide uncertainty bands in their reconstruction and our estimate is within the upper band of their uncertainties for that period [Fig. 3b in Gergis et al. (2012)].

In terms of low rainfall decades, the additional 35 years does not change what was established using the operational grids from 1900 to 2010. The 2000s [average annual rainfall of 512.8/512.0 mm (gridded/reconstructed) for 1999–2009] and the mid-1930s/mid-1940s (average annual rainfall of 519.0/518.0 mm for 1935–45) were the two driest 11-yr periods. Highlighting the high decadal variability of rainfall in this part of the world, it is worth emphasizing that the 1935–45 very dry period is immediately followed by the wettest 11-yr period in the reconstruction (1946–56). Using the network data, it is now possible to properly cover the FD. Average annual rainfall stands at 535.2 mm for 1895–1905 and is certainly the next most noticeable nonoverlapping dry period in the reconstruction but about 20 mm yr$^{-1}$ wetter than the period encompassing the MD and the WWIID.

In Gergis et al. (2012), a very sharp low point is reached in the decadal mean around 1910: it appears to be the lowest value of the entire reconstruction. However, the gridded averages do not agree well with that very sharp low point, nor do the network reconstruction presented here, which lines up very well with the gridded average in suggesting a close to average decade at the end of the earlier FD.

At 555.6 mm for 1876–86, the mid-1870s–mid-1880s appears also as a reasonably dry decade (more than 30 mm yr$^{-1}$ below the long-term mean). However, this last-mentioned period has a considerable overlap with the previously mentioned wet period of the 1870s, emphasizing the variable nature of rainfall in this region—rapid shifts between wet and dry conditions are a feature of the climate. That overall period of 1870s and 1880s was marked by strong ENSO variability with several strong La Niñas in the 1870s (Garden 2009) and in 1878–1880 (Allan et al. 1991) and a record-breaking El Niño in 1877–78 (Allan et al. 1991) and a very strong one in 1888–89 (Nicholls 1997).

Decadal variability shown for seasonal rainfall totals (Fig. 7) provided some additional insight into both the sizeable decadal variability and which seasons contribute both to the wet and dry decade discussed earlier. The very wet 1870s appear to be due to a combination of all seasons being far above the long-term mean, apart from winter. Autumn was particularly notable, with the highest 11-yr mean on record by a considerable margin. These very wet autumn years in the early part of the record.
combined with the strong autumn deficit experienced during the MD, contribute to a strong and significant downward trend for autumn rainfall.

The following dry decade of the mid-1870s–mid-1880s appears to be due to a very sharp decline of summer rainfall prior to 1880 and then winter rainfall centered on 1880. Similarly the FD is due to very low spring rainfall (lowest 11-yr running mean is reached in 1900), with record-breaking low summer rainfall during the 1900s. In both instances, the fact that the low decadal annual total rainfall is made of distinctive periods where different seasonal means reach remarkably low values are consistent with the view that these low rainfall decades were the result of chaotic behavior of the climate system. A series of well-documented El Niños did occur at the time of the FD: 1895–97 was a weak to strong event, 1899–1901 a strong one, and 1902–03 a very strong one; finally there was a strong one in 1904–05 (see http://www.bom.gov.au/climate/enso/enlist/ for more information). On the contrary, the WWIID and MD are both a combination of low rainfall during the cool part of the year (autumn–spring) with summer rainfall remaining at or only slightly below the long-term mean. While during the WWIID the contribution of all three seasons was comparable, during the MD the rainfall deficit has been largely in autumn. For both periods, a part of the rainfall deficit has been attributed to anomalies in the intensity of the STR (Timbal and Drosdowsky 2012), which appears to coevolve with the global warming of the planet (Timbal et al. 2010).

In addition to the 11-yr running means, the linear trends in the network-reconstructed time series are shown on Fig. 7. Overall, the linear trends are not very meaningful given the strong decadal to interdecadal variability. It is worth noting, however, that the trends in spring (September–November) and summer (December–February) rainfall are positive, while the trends in autumn (March–May) and winter (June–August) and the annual rainfall are negative. In relative terms, the autumn rainfall trend is clearly strongest with a contribution from both the record-breaking high autumn rainfall in the 1870s and autumn rainfall in recent years being well below previous levels. Of the five trends, the autumn trend is the only statistically significant one at any meaningful level ($p = 0.037$ two-tailed).

5. Comparison of the three historical low rainfall periods

From the analysis of interannual and decadal variability performed earlier, it appears that three low rainfall periods are worth investigating further: the Federation, World War II, and Millennium Droughts. We will start with a classical approach using fixed-term periods and then move on to the DDD approach described in the methodology section.
Table 2 compares the three major droughts in the SEA region in terms of their annual and seasonal anomalies for the driest 11-yr and 13-yr periods for each period. The driest periods are chosen as the sets of calendar years that locally minimize the average annual rainfall total and all results are based on the reconstructed network in order to properly describe the FD, which began prior to 1900. All anomalies are expressed as percentage departures from the full 146-yr period (1865–2010). Note that in the case of summer, where we used contiguous months, only 145 years are available, starting from the 1865/66 summer and finishing with the 2009/10 summer.

The three droughts show different characteristics. The FD shows the strongest reduction (in percentage terms) in spring, consistent with the role played by the large number of strong to very strong El Niños events occurring at that time. Changes in autumn and winter are relatively insignificant. Summer changes are extremely noisy, making only a small contribution for the 11-yr period (−6.1% reduction) but jumping to −12.3% for the 13-yr period. This behavior appears random and is consistent with a naturally occurring drought: low rainfall at a time during the most variable part of the annual cycle (summer), with a strong overall contribution of low spring rainfall, spring being the season with the most marked decadal variability (Timbal et al. 2010) and more clearly linked to naturally occurring tropical modes of variability (Timbal and Hendon 2011) and consistent with the historical evidence regarding ENSO variability at that time.

The WWIID shows approximately equal reductions in autumn, winter, and spring, with the summer changes being relatively insignificant. The very large spring signature for the most intense part of the WWIID (1935–45) is consistent with the documented role of tropical modes of variability for this period (Timbal and Hendon 2011). Beside the naturally occurring component of the WWIID, the severity of the WWIID was partly attributed (about 30%) to the global warming of the first half of the twentieth century and the intensification of the STR (Timbal et al. 2010). This is consistent with the seasonality of the rainfall deficit during the WWIID, which spans all the cool parts of the year when the relationship between the STR intensity and SEA rainfall is well established (Timbal and Drosdowsky 2012).

In marked contrast, the MD shows the strongest reduction in autumn (approximately equivalent to the spring reduction in the FD), with all the other seasons showing reductions of around one-third to one-half of the autumn reduction. That strong autumn signature and its relation to meridional circulation changes including the surface signature of the Hadley cell (the subtropical ridge) has been documented in detail as part of the SEACI program (CSIRO 2010).

A more in-depth comparison of the three historical low rainfall periods (FD, WWIID, and MD) has been conducted by analyzing the drought–depth–duration defined in the methodology section for periods ranging from 1 to 21 years. For each period, the 21 years with the lowest 21-yr mean rainfall encompassing each drought was chosen: these are 1895–1915 (FD), 1925–1945 (WWIID), and 1990–2010 (MD). The annual DDD is shown in Fig. 8 (bottom panel) for each of the three droughts. In addition a temporal view of the three periods is provided: accumulated monthly rainfall deficit for the three periods (Fig. 8, top panel). The temporal view provides a clear picture of the growing deficit accumulated during the MD from about month 48 to month 240 (16 years in total) with very limited recoveries [none reaching 100 mm, in contrast to the 190-mm recovery during the last 12 months (i.e., 2010)]. This is the main difference between the MD and the previous two 21-yr period where significant recoveries ranging from 12 to 36 months are noted.

Also shown are the expected or mean DDD under the Monte Carlo simulation for random or average 21-yr periods (gray line, “typical”) and the expected or mean DDD for the “driest in the entire 146 years” 21-yr period (black line, “typical worst”). The DDDs for the three droughts are larger than the typical DDD, for the obvious reason that these three drought periods have been selected because they are droughts. The

Table 2. SEA seasonal rainfall anomalies expressed as percentages relative to the long-term average for selected driest 11-yr and 13-yr periods associated with three major historical droughts.

<table>
<thead>
<tr>
<th>Period (either 13 or 11 years)</th>
<th>Annual (%)</th>
<th>Autumn (%)</th>
<th>Winter (%)</th>
<th>Spring (%)</th>
<th>Summer (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Federation Drought 1895–1907</td>
<td>−7.7</td>
<td>−1.9</td>
<td>1.3</td>
<td>−19.5</td>
<td>−12.3</td>
</tr>
<tr>
<td>Federation Drought 1895–1905</td>
<td>−8.8</td>
<td>−3.9</td>
<td>1.3</td>
<td>−23.5</td>
<td>−6.1</td>
</tr>
<tr>
<td>WWII Drought 1933–45</td>
<td>−8.8</td>
<td>−15.1</td>
<td>−9.1</td>
<td>−9.2</td>
<td>1.0</td>
</tr>
<tr>
<td>WWII Drought 1935–45</td>
<td>−11.7</td>
<td>−14.4</td>
<td>−10.7</td>
<td>−15.9</td>
<td>−4.0</td>
</tr>
<tr>
<td>Millennium Drought 1997–2009</td>
<td>−12.8</td>
<td>−23.4</td>
<td>−9.9</td>
<td>−7.8</td>
<td>−10.8</td>
</tr>
<tr>
<td>Millennium Drought 1999–2009</td>
<td>−12.8</td>
<td>−23.7</td>
<td>−9.8</td>
<td>−10.0</td>
<td>−7.5</td>
</tr>
</tbody>
</table>
comparison with the typical worst DDD in the simulations is more revealing—the MD is more intense over a wide range of assessment years (5 to 19 years) than the typical worst, in ways that the two earlier droughts are not. Only over relatively short periods (one to three years) do all three droughts fall short of the typical worst. In the 1-yr and 2-yr cases, this is because the driest and second driest years in the annual time
series (1982, 1967) are not part of these three droughts and the simulation is created by bootstrapping the actual data with replacement rather than by fitting some theoretical or parametric distribution to the actual data and sampling from that distribution.

The expected DDDs for the typical and typical worst 21-yr periods (i.e., the gray and black lines in Fig. 8) are given in Table 3, along with the statistical significances (one-tailed) for the DDD of the MD. These significances are computed in the usual way, as \( \Pr(\text{simulated } D_p > \text{observed } D_p) \). The Monte Carlo simulation used to estimate these statistical significances involved 20,000 iterations. Compared to the typical 21-yr period, the MD is statistically significant at the 5% level for periods of length seven years and longer.

Statistical significance is not achieved in the much more stringent comparison against the typical worst, although for periods from 13 to 17 years the significance is consistently below 0.20 and reaches 0.10 for 16 years. This implies that the MD is unusual in terms of the worst that could be expected in a 146-yr period, and is extreme enough to attain statistical significance at the 90% level for the 16-yr duration.

Since it was shown earlier that the network reconstruction of SEA rainfall also provided a reasonable reproduction of monthly total rainfall, the same DDD analysis was conducted on the four calendar seasons and is shown in Fig. 9.

The difference between the MD and the earlier droughts is particularly stark in autumn. The MD has a DDD worse than the typical worst in all period lengths from 1 to 21 years, which in the case of the very shortest periods is due to the driest autumn in the 146 years (2005) occurring within the MD. The earlier droughts do not exceed the typical worst in any period lengths, and the FD does not even reach the typical level over most of the periods (consistent with the fact that the average autumn rainfall for the 11-yr and 13-yr FD periods given in Table 2 was above the long-term autumn average).

In winter, the three droughts are generally less intense than the typical worst, although the two earlier droughts exceed it for the 1-yr period. This is because the driest winter (77.8 mm, 1944) occurs during the WWIID, while the second driest winter (85.4 mm, 1914) occurs during the FD. The MD does not attain typical levels for 1- to 3-yr periods, but it exceeds this intensity for 6- to 21-yr periods, and is the worst of the three droughts for 13- to 16-yr periods. The FD shows a strong abnormality in the spring DDD results, being above the typical worst for almost all periods. The driest spring in the reconstruction (65.2 mm, 1938) occurs in the WWIID, so that drought shows as worse than the typical worst for the 1-yr period. This confirms the strong influence of historical El Niños during that period (Nicholls 1997) as spring is the time of the year when tropical modes of variability impact on SEA rainfall is largest. The MD sits between the typical and typical worst for most periods.

For summer, as previously mentioned (Table 2), the FD had a strong but short period of intense summer rainfall deficiencies. Indeed, its summer DDD is above the typical worst for duration between 2 and 4 years (the mid-1900s). The MD has a typical DDD for period extending to 10 years and is marginally above for period between 10 and 20 years.

Significance levels were estimated for each season, using the same method as for the annual totals (see Table 3); results (not shown) are only briefly discussed here. For autumn rainfall, the MD attains statistical significance at the 5% level for 12 of the 21 durations and statistical significance at the 10% level for 18 of the 21 durations. Significance is not attained at the 10% level in either winter (lowest \( p \) value is 0.107 for 18-yr duration) or summer for the MD, and is only attained in spring at the 3-yr duration [due to the driest spring (2006) of the entire 146 years occurring in the MD].

### Table 3. Average DDDs and one-tailed statistical significances for the Monte Carlo simulation of SEA annual rainfall; results for average (“typical”) 21-yr periods and worst (“typical worst”) 21-yr periods are both shown. Statistical significances are for the 21-yr period (1990–2010) encompassing the MD.

<table>
<thead>
<tr>
<th>Duration</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
</tr>
</thead>
<tbody>
<tr>
<td>MD significance</td>
<td>0.39</td>
<td>0.54</td>
<td>0.27</td>
<td>0.14</td>
<td>0.10</td>
<td>0.09</td>
<td>0.04</td>
<td>0.02</td>
<td>0.03</td>
<td>0.04</td>
</tr>
<tr>
<td>MD significance</td>
<td>0.68</td>
<td>0.90</td>
<td>0.67</td>
<td>0.46</td>
<td>0.40</td>
<td>0.39</td>
<td>0.25</td>
<td>0.15</td>
<td>0.18</td>
<td>0.24</td>
</tr>
<tr>
<td>Duration</td>
<td>11</td>
<td>12</td>
<td>13</td>
<td>14</td>
<td>15</td>
<td>16</td>
<td>17</td>
<td>18</td>
<td>19</td>
<td>20</td>
</tr>
<tr>
<td>“Typical”</td>
<td>4.57</td>
<td>3.92</td>
<td>3.36</td>
<td>2.87</td>
<td>2.43</td>
<td>2.04</td>
<td>1.67</td>
<td>1.25</td>
<td>0.91</td>
<td>0.5</td>
</tr>
<tr>
<td>MD significance</td>
<td>0.03</td>
<td>0.03</td>
<td>0.02</td>
<td>0.02</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
<td>0.03</td>
<td>0.03</td>
<td>0.06</td>
</tr>
<tr>
<td>MD significance</td>
<td>0.24</td>
<td>0.22</td>
<td>0.14</td>
<td>0.19</td>
<td>0.15</td>
<td>0.10</td>
<td>0.14</td>
<td>0.35</td>
<td>0.40</td>
<td>0.44</td>
</tr>
</tbody>
</table>
6. Discussion and conclusions

The extension of SEA rainfall back to 1865 provides an interesting insight into the late part of the nineteenth century. It offers an opportunity to test and complete current understanding of the rainfall variability across SEA in particular for longer (decadal and beyond) periods. Using a concept of drought–depth–duration (DDD), the network reconstruction presented here has reinforced the historical perspective on the abnormality of the Millennium Drought (MD) that far exceeds both the Federation Drought (FD) and World War II Drought (WWIID) for an extended range of periods from short droughts of 3 years to prolonged decline of up to 20 years. Using a randomization process, it was also shown that the current situation is beyond what can be expected by chance, particularly for autumn rainfall, although in terms of annual rainfall without reaching a high significance level.

In terms of relationships with forcings within the climate system and SEA rainfall, this historical reconstruction provides an opportunity to update some recent analyses (Timbal et al. 2010; Timbal and Hendon 2011). Using a tripolar sea surface temperature index (TPI) across most of the tropical oceans, having a statistical relationship with SEA rainfall, Timbal and Hendon (2011) showed that the strength of this correlation varies on multidecadal time scales and appears to coevolve with the decadal variation of detrended global temperature. An update of their graph [Fig. 14 in Timbal and Hendon (2011)] is shown in Fig. 10 using the Hadley Centre Global Sea Ice and Sea Surface Temperature (HadISST; Rayner et al. 2006) to compute the tropical tripole over the period 1865 to 2010. The overall relationship between multidecadal variability in the strength of the coupling between the tropics and SEA rainfall and the detrended global temperature that they noted is also observed in the early part of the record—that is, during the late nineteenth century when the detrended global temperature indicated the absence of ongoing global warming, and correlation coefficients between SEA rainfall and the TPI were low compared to most of the twentieth century, certainly for both winter and spring (Fig. 10). The 30-yr correlation coefficient is re-established to meaningful levels by the 1920s for both winter and spring (i.e., for the period starting in 1890), which is consistent with the previous discussion of the role of ENSO variability during the FD.

On time scales beyond decadal variability, Timbal et al. (2010) noted that the period with very high rainfall...
during the 1950s to the 1970s, although interrupted somewhat by the drier 1960s, coincides with a period when annual global mean temperatures were relatively stable, from 1945 to 1970 [see a summary of the existing literature and relevant knowledge regarding observed global warming in Solomon et al. (2007)]. Timbal et al. (2010) argue on the basis of the rainfall–STR–global warming relationship that the very high rainfall observed in this region was due in part to the relaxation from the broader twentieth-century global warming signal. Therefore it is interesting to observe that during the later part of the nineteenth century, a period now documented by our instrumental reconstruction, high decadal variability is observed with high peaks in the 1870s (arguably close to the highest on record) and the 1890s, similar to the 1950s to 1970s. As for the 1945–70 period, no ongoing global warming was observed during the late nineteenth century, indeed the period from 1880 to 1910 was characterized by falling annual global mean temperatures. It remains, however, a limited time span to confirm (or reject) the possible relationship between the occurrence of very wet decades and the temporal evolution of the global annual mean temperatures, but the present reconstruction cannot be extended much prior to 1865, even with a coarser network. Given the limited nineteenth-century data availability, any further analysis of this possible relationship would therefore have to involve the newly developed paleoclimatic reconstructions which are emerging for SEA [e.g., Gergis et al. (2012) and other contributions to this special issue] and provide access to longer time spans for which good proxies of the global temperature of the planet exist.

While comparing our results with Gergis et al. (2012), it is obvious that they are two very different products; while historical reconstruction can provide highly reliable reconstruction (capturing nearly all the interannual variability), there is an inherent limit of the possibility to extend these records back in time; this is particularly true for Australia where until European colonization in 1788, no written records exist. On the contrary, paleo-reconstructions can provide much longer historical perspective but with a limit in the temporal accuracy (i.e., while decadal variability is well captured, interannual variability is sometimes poorly captured, particularly in data-sparse regions, which necessitates relying on records drawn from outside the target domain). However, an interesting overlap exists across the nineteenth century where extended instrumental rainfall data could provide additional overlap needed to validate and calibrate paleo-reconstructions. A closer interaction between these two research communities may yield improved estimates of past rainfall variations in the Australian region.

**Fig. 10.** The 30-yr synchronous correlation coefficient (Pearson $r$) between SEA rainfall and the TPI SST index for winter [June–August (JJA)] and spring [September–November (SON)] (left axis, reversed scale). Each 30-yr correlation coefficient is plotted against the last year in the 30-yr period. Also shown is the 30-yr running mean of the linearly detrended annual global mean temperature anomaly (dashed gray line, right axis). The linear detrending is calculated over the period 1865–2010.
This longer historical perspective allows for some possible quantification of the shift in climate baseline for the ongoing period. This is a question often asked by water managers and others affected by the changes observed during the MD (e.g., Leblanc et al. 2012). A range is provided in Table 4 for the annual mean and three calendar seasons with sizeable declines (autumn, winter, and spring). To generate this table, we calculate the difference between the mean of the 21 DDD values for the MD and the mean of the 21 expected DDD values under the Monte Carlo simulation. For a severe drought, this mean difference would be positive. We then inflate the rainfall over the MD (1990–2010) by a fixed factor and recalculate this mean difference in the DDDS. The percentages in the table are those required to reduce a mean positive difference in the actual data down to zero in the inflated data.

This range depends on the scientific view that can be established regarding how naturally occurring modes of variability may have contributed to shift the last 21 years compared to the longer 146-yr record. The choice between which end of the range is to be considered more likely depends on the available evidence regarding the role of natural variability in the ongoing decline. For autumn it can be argued that it is more likely that the shift in climate baseline is closer to the worst case than normal and hence the shift from the baseline is more likely to be close to a 17% shift. The reasons for this are that autumn is the time of the year when the natural variability is the highest (Timbal 2010) and natural variability is mostly due to random weather noise and not large-scale modes that do not influence SEA rainfall at this time of the year (Timbal and Murphy 2007). In addition, an attribution study (Timbal et al. 2010) showed that the observed case was at the extreme of the range of model projections even when forced with all existing external forcings (including anthropogenic forcings).

In winter and spring, natural variability is low (Timbal, 2010) and is largely contributed by tropical modes of variability (e.g., ENSO), which are especially important for decadal variability. Timbal and Hendon (2011) showed that tropical modes of variability in winter have had a small but positive influence on SEA rainfall, suggesting that natural variability has actually contributed to limit the drier shift in winter rainfall because of the strengthening of the STR. In addition, there is a reasonable consensus among climate models that future rainfall in SEA will be reduced (CSIRO 2010). Therefore, it is reasonable to argue that it is more likely that the shift in climate baseline is at least the magnitude of the shift from the normal case (4%–6%) since no naturally occurring modes of variability appear to have contributed and it is likely that a part of the future rainfall decline in response to anthropogenic forcing has already occurred even if it is not detectable statistically.

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