Understanding hydroclimate processes in the Murray-Darling Basin for natural resources management

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Abstract

Isolating the causes of extreme variations or changes in the hydroclimate is difficult due to the complexities of the driving mechanisms, but it is crucial for effective natural resource management. In Australia's Murray-Darling Basin (MDB), ocean-atmosphere processes causing hydroclimatic variations occur on time scales from days to centuries, all are important, and none are likely to act in isolation. Instead, interactions between all hydroclimatic drivers, on multiple time scales, are likely to have caused the variations observed in MDB instrumental records. A simplified framework is presented to assist natural resource managers in identifying the potential causes of a hydroclimatic event. The framework condenses an event into its fundamental elements, including its spatial and temporal signal and small-scale evolution. The climatic processes that are potentially responsible for the event are then examined to determine possible causes. The framework was applied to an event occurring from 1997–2010 when the southern MDB experienced prolonged and severe dry conditions, providing insights into possible causal mechanisms that are consistent with recent studies. The framework also assists in identifying uncertainties and gaps in our understanding that need to be addressed.

1. Introduction

Periods of prolonged, severe drought and moisture surplus are challenging for natural resources management/managers (NRM). The events that are at the very edge of past experience are of particular concern as they push the limits upon which standard management practices are based, making socioeconomic and some natural systems vulnerable. The unprecedented impacts of these extreme events can sometimes necessarily force a somewhat *ad hoc* management response, which may be successful but may also lead to undesirable outcomes (i.e. maladaptation). Recent periods of prolonged wet and dry conditions have provided further cause for thought as NRM debate the causes of these extreme conditions, whether they are linked to anthropogenic climate change, whether similar events may become more prevalent in the future, and even whether such periods should be considered the new norm.

Several extreme climatic events have recently occurred in Australia's largest food-producing region, the Murray–Darling Basin (MDB). Much of the region experienced ongoing drought from around late-1996 to mid-2010 [*Murphy and Timbal*, 2008], which was followed by a rapid transition to extremely wet conditions across the MDB, the north in particular, from late 2010 to the time of publication of this article. Management practices were tested throughout the sometimes-unprecedented nature of both periods and there were several instances when water resource managers, agriculturalists and biodiversity managers (hereafter collectively referred to as NRM) increasingly questioned whether the conditions in parts of the MDB contained an anthropogenic component.

To understand the causes of the recent changes to the MDB climate, and potential future changes that include the effects of anthropogenic climate change, several national research initiatives were established. These initiatives included the Murray Darling-Basin Sustainable Yields Project (MDBSY; www.csiro.au/partnerships/MDBSY); the South Eastern Australian Climate Initiative (SEACI; www.mdbc.gov.au/subs/seaci); the Indian Ocean Climate Initiative (IOCI; www.ioci.org.au); and the CSIRO-BoM Climate Change in Australia project (www.climatechangeinaustralia.gov.au). The collective research from these, and other, initiatives has provided invaluable information that is highly relevant to the NRM community.

However, increasing the body of knowledge has the potential to be confusing for NRM. For example, the climate science community has identified a number of physical processes that influence variations in the MDB hydroclimate, which are often relayed to the NRM community as climate indices (i.e. simplified metrics of these atmosphere-ocean processes). Unavoidably, the number of these indices, and opinions as to what is important and what is not, increases as more research is conducted, adding uncertainty and complicating the information for those relying on it for risk assessment.

Given the difficulties NRM might have utilizing the large amount of climate information available, this paper has two primary objectives. The first is to synthesize knowledge about the processes governing climatic variations across the MDB. This synthesis will distil up-to-date information in a concise form so that it may be interpreted more easily than compiling information from self-guided literature surveys. *Murphy and Timbal* [2008] have previously presented a synthesis of the

85 major processes driving inter-annual and shorter time scale climatic variations in
86 southeast Australia. While useful for the southern MDB, *Murphy and Timbal's* [2008]
87 paper did not describe the importance of these and other climate processes for the
88 northern MDB. Furthermore, information on decadal and longer-scale climate
89 processes, interactions between climate drivers and interactions between climate and
90 hydrological processes were not included in their review but are highly relevant for
91 NRM in the MDB.

92 The second objective is to identify a broad framework that allows NRM to
93 scrutinize existing climate knowledge in order to better isolate the potential
94 cause/causes of a climatic event. It is hoped this may make information relating to
95 hydroclimatic processes more useful in risk management applications. Such a
96 framework simplifies a complex problem and so inherently includes a large degree of
97 uncertainty as it is derived from a mostly chaotic system. So, while the framework
98 does not provide absolute answers, we argue that it establishes a more focused set of
99 considerations for NRM. Furthermore, the framework highlights areas where the
100 knowledge base is lacking (i.e. areas with large uncertainty or actual knowledge
101 gaps), thus identifying areas of future research that, if undertaken, will also prove
102 useful for NRM.

103 To fulfil these objectives, the paper is presented as follows: first, a
104 hydroclimatology for the MDB provides context. The synthesis on climate processes
105 is then presented. This includes describing processes acting on inter-annual and
106 shorter scales as an update *Murphy and Timbal* [2008] and an explanation on the
107 drivers of decadal and longer-scale variability, which has not previously been
108 presented. We then show analysis for an emerging area of the science that, we argue,

is critical to understanding and predicting wet and dry cycles in the MDB, namely interactions between climate processes. Finally, the framework for highlighting the potential cause/causes of climatic events for NRM is presented and applied to the case of the prolonged drought that affected the southern MDB for over 13 years from the mid 1990s.

2. Hydroclimatology of the Murray-Darling Basin: 1900–2010

The range of hydroclimatic variability on which NRM practices are based has traditionally been defined using historical data. By global standards the climate records in the MDB are long, spanning 111 years from 1900–2010 with some individual stations going back to the mid-1800s. Here, we give an up-to-date overview of the hydroclimatology of the MDB based on these instrumental observations.

Spanning from the subtropics (24°S) to the middle latitudes (38°S), approximately 90% of the MDB is classified as either arid or semi-arid [*Maheshwari et al.*, 1995]. The topography of the Basin consists mostly of extensive plains and low undulating areas, with much of the Basin less than 200 metres above sea level. The landscape varies throughout the catchment from the sub-tropical rainforests of the eastern uplands, to the temperate, sub-alpine regions of the southeast and the hot, dry semi-arid plains in the west [*Barros and Bowden*, 2008].

Annual rainfall averaged across the MDB is approximately 470 mm per year (calculated over the period 1900 – 2010). However, the spatial variations in annual mean rainfall across the MDB are large (Figure 1). Rainfall is highest in mountainous

areas in the southeast where the mean annual totals can exceed 1500 mm, and lowest in the west of the Basin where they are typically less than 300 mm/year. Averaged across the Basin, evaporation (estimated using pan evaporation) is approximately four times the annual mean rainfall. Geographically, evaporation ranges from approximately 1200 mm/year in the southeast of the Basin to around 2400 mm/year in the west [*BoM*, 2010].

Over the 111 years from 1900–2010, MDB rainfall has exhibited large inter-annual to multi-decadal fluctuations (Figure 2). Averaged across the MDB, conditions were broadly drier during the first half of the 20th Century with Basin-averaged rainfall deficits frequently exceeding 100 mm/year, equivalent to rainfall being at least 20% below average (Figure 2). Several periods of multi-year rainfall deficits occurred, including the "Federation Drought" (c.1895 – 1902) and the "World War II Drought" (c.1937 – 1945) [*Ummenhofer et al.*, 2009b; *Verdon-Kidd and Kiem*, 2009b]. Though 19 years in the instrumental record had above average rainfall during the first half of the 20th century (Figure 2), there was only one multi-year wet period prior to 1950 (i.e. 1916 to 1917 when MDB rainfall was at least 20% above average).

The period from 1950 to 2010 was dominated by large decadal-scale variations in rainfall. Two distinct periods during the 1950s and the 1970s had several years with rainfall considerably greater than average and major floods occurred during 1952, 1956 and 1974. Multi-year periods with below average annual rainfall marked the 1960s, early 1980s, and mid-1990s to early 2010. However, there were also some notable floods during these dry epochs, for example, 1990 across much of eastern Australia and more recently 2008 and 2009 in the northern MDB.

155 The dry conditions from the mid-1990s to early 2010 were dubbed 'The Big
156 Dry' or the 'Millennium Drought' (hereafter referred to as 'The Big Dry' only)
157 [*Ummenhofer et al.*, 2009b; *Verdon-Kidd and Kiem*, 2009b] (Figure 3). This drought
158 was mainly confined to the southern MDB and was dominated by autumn and early
159 winter rainfall deficits stemming from both a decrease in the number of rain days and
160 a reduction in the intensity of daily rainfall events [*Verdon-Kidd and Kiem*, 2009b].
161 The drought was broken in spectacular fashion in 2010, which was the wettest year on
162 record averaged across the MDB. Much of this stemmed from late winter and spring
163 rainfall, which led to annual rainfall being 70% above average (Figure 2).

164

165 3. Hydroclimate processes of the Murray-Darling Basin

166 We direct readers to the review paper by *Murphy and Timbal* [2008] for a
167 broad overview of the key processes relevant to climate variations in the southern
168 MDB on inter-annual and shorter time scales, including the mechanisms by which
169 they occur and their climatological characteristics (e.g. frequency, persistence etc.).
170 Here, we expand on these topics and include information relevant to the northern
171 MDB and interactions between hydroclimate driving mechanisms that are relevant to
172 the whole MDB. The northern and southern MDB are defined as the halves of the
173 Basin north and south of 33°S. To date, less focus has been given to understanding
174 climate processes in the northern MDB by the climate science community and a
175 synthesis has not previously been presented. An explicit discussion of relevant daily-
176 scale processes, a synthesis of the most recent findings on inter and intra-annual
177 processes since *Murphy and Timbal* [2008] was published, and a discussion on the

inter-decadal processes, which has not previously been presented for the MDB are provided.

3.1 Daily to intra-annual processes

The types of weather systems that produce and inhibit rainfall in the MDB are summarized in Figure 4. The origins of these systems vary regionally with the most distinct differences occurring from north to south. The importance of a particular type of weather system to seasonal rainfall totals is heavily dependent on the location. There can be substantial differences in the relative contributions from particular weather systems to seasonal accumulations over distances of several hundred kilometres only [*Verdon-Kidd and Kiem*, 2009a].

The rainfall accumulations associated with different occurrences of the same type of weather system also vary, and there are differences in the relative amount of rainfall each type of system tends to produce [*Pook et al.*, 2006]. For example, tropical cyclones are more likely to produce higher rainfall totals in the northern MDB compared to frontal systems in the southern MDB. The typical amount of rainfall produced depends on the intensity and duration of each system. Their contribution to the climate depends on those factors and also on the frequency at which those systems impact on an area.

The make up of the daily-scale rainfall distribution is particularly important for water resources management. For example, the impact of a one-month rainfall total of 100 mm on a hydrological system would be very different depending on whether the rainfall had fallen evenly throughout the month, or in a single event [*McIntosh et al.*, 2007; *Verdon-Kidd and Kiem*, 2009b].

Rainfall in the southern MDB mostly stems from extra-tropical weather systems [*Wright*, 1989; *Qi et al.*, 1999; *Pook et al.*, 2006; *Risbey et al.*, 2008]. Figure 4 illustrates some of these systems, such as cold fronts, cut-off low-pressure systems and cloud bands. Cut-off low pressure systems contribute the most rainfall from a single type of system, between approximately 25% and 50% of rainfall to the region [*Wright*, 1989; *Pook et al.*, 2006]. Cut-off lows are also responsible for over 80% of the rainfall that occurs on heavy rain days (described as days with over 25 mm or rain) during these months [*Pook et al.*, 2006].

Frontal systems that interact with tropical air masses (e.g. cloud bands that form from a surface or middle troposphere pressure trough entraining tropical moisture, such as the Northwest cloud band in Figure 4), those that do not (e.g. the cold fronts in Figure 4), and post-frontal rainfall provide smaller, but significant, contributions to southern MDB rainfall totals, although the influence of each varies regionally [*Wright*, 1989; *Verdon-Kidd and Kiem*, 2009a]. Post-frontal rainfall is only a significant contributor close to the mountain range spanning the southeast MDB, where orographic uplift is the primary mechanism causing rainfall. In the southern MDB, frontal rain contributes between 25% and 40% of monthly rainfall totals, with this contribution decreasing moving further north [*Pook et al.*, 2006]. *Verdon-Kidd and Kiem* [2009a] attribute sizeable proportions of rainfall in the state of Victoria in the southern MDB to systems such as pre-frontal and inland troughs that produce cloud bands. However, *Wright* [1997] and *Pook et al.* [2006] state that such systems are generally responsible for less than 10% of total rainfall in this region.

In the northern MDB, rainfall mostly stems from tropical systems, or from interactions between tropical and extra-tropical systems [*Wright*, 1997; *Sturman and*

Tapper, 2005]. Meridional troughs that extend through the subtropics of eastern Australia, sometimes called 'easterly-dips' (Figure 4), can be associated with widespread rain via cloud bands, particularly from April to October. Cloud bands that are initiated from convergence caused by the trough alone account for some 25-40% of rainfall in the northern half of the MDB [*Wright*, 1997]. There is a similar contribution from the interaction between these tropical cloud bands with extra-tropical features such as cold fronts [*Wright*, 1997].

Strong cold fronts can penetrate as far north as the Australian tropics. When they occur during the onset and breakdown of the northern wet season (approximately November – March), they may trigger thunderstorms along their leading edge, bringing very localized, heavy rain to areas in the northern MDB [*Sturman and Tapper*, 2005]. From December – April, tropical cyclones that have moved inland from the east Australian coast can occasionally contribute large rainfall totals to the northern MDB.

High-pressure systems are the features that primarily inhibit rainfall [*Sturman and Tapper*, 2005]. Over the MDB, these may constitute part of the climatological belt of high-pressure, known as the sub-tropical ridge (STR), or be independent of it. Persistent and stationary high-pressure systems that are removed from the STR are responsible for the phenomenon of atmospheric blocking, where rain bearing weather systems are diverted around the immobile blocking high-pressure system (e.g. Figure 4). In the southern MDB, atmospheric blocking and the STR that are situated directly over the region are the primary systems responsible for rainfall suppression [*Pook et al.*, 2006; *Risbey et al.*, 2008; *Verdon-Kidd and Kiem*, 2009a]. In addition to high-pressure systems that inhibit rainfall, some mechanisms that usually trigger rainfall

(e.g. a cold front) will fail if atmospheric moisture is not available [*Drost and England*, 2008].

Intra-seasonal climate variations across the MDB have been linked to the Madden-Julian Oscillation (MJO), which describes an eastward propagating region of enhanced tropical convection [*Hendon and Liebmann*, 1990]. Typically, the impacts of the MJO on intra-seasonal MDB rainfall vary sub-regionally, seasonally and with the phase of the MJO [*Wheeler et al.*, 2009]. However, across the MDB, the strongest impacts are during the winter and spring months. See *Murphy and Timbal* [2008] and *Wheeler et al.* [2009] for a description of the impacts of the MJO on the southern MDB and regional Australian circulation respectively.

Intra-annual wet and dry periods in the MDB are primarily linked to shifts in the atmospheric circulation inherent to the seasonal cycle. In the northern MDB (at subtropical latitudes), rain bearing systems are more prevalent during the warmer months (approximately November – April), due to the greater influence of tropical weather systems such as those shown in Figure 4a. The opposite is true for the southern MDB, where extra-tropical weather systems dominate (Figure 4b) and rainfall mostly occurs during the cooler months (approximately May – October).

3.2 Inter-annual processes: an update

The El Niño – Southern Oscillation (ENSO) can be responsible for over 20% of local annual rainfall variations in the MDB [*Pittock*, 1975; *McBride and Nicholls*, 1983; *Ropelewski and Halpert*, 1987; *Nicholls*, 1988; *Power et al.*, 1998; *Risbey et al.*, 2009].

Seasonally, winter, spring and summer MDB rainfall variations are most strongly associated with ENSO events [*McBride and Nicholls*, 1983; *Risbey et al.*, 2009], although the winter relationships may also be due to Indian Ocean influences and their coincidence with ENSO (discussed further below). The effects of ENSO in the MDB also include magnified fluctuations in streamflow volumes compared to rainfall [*Chiew et al.*, 1998; *Wooldridge et al.*, 2001; *Verdon et al.*, 2004b], elevated flood risk during La Niña events [*Kiem et al.*, 2003], and increased risk of drought [*Kiem and Franks*, 2004; *Barros and Bowden*, 2008] and bushfire [*Verdon et al.*, 2004a] during El Niño events.

The intensity of wet and dry periods in the MDB is not directly proportional to the strength of ENSO events, where the strength of an ENSO event is determined based on the size of the departure from normal sea-surface temperature (SST) or sea-level pressure (SLP) conditions. In fact, the rate at which the state of the Pacific changes from one phase of ENSO to another appears to be just as important as the strength of the ENSO event [*Quinn et al.*, 1978; *Stone and Auliciems*, 1992; *Stone et al.*, 1996; *Kiem and Franks*, 2001]. This is demonstrated by the fact that some of the strongest El Niños have not been associated with severe, widespread drought (*e.g.* 1997/1998), while other weaker El Niño events have coincided with particularly intense drought (*e.g.* 2002/2003) [*Wang and Hendon*, 2007; *Brown et al.*, 2009]. The non-linear relationship between Australian rainfall variations and the magnitude of ENSO has been attributed to interactions with random atmospheric noise [*Wang and Hendon*, 2007], interactions with other climate drivers [*Meyers et al.*, 2007; *Verdon-Kidd and Kiem*, 2009a; *Kiem and Verdon-Kidd*, 2010], and the location of the SST anomalies associated with ENSO [*Wang and Hendon*, 2007].

Larkin and Harrison [2005] and *Wang and Hendon* [2007] showed that Australian rainfall is particularly sensitive to ENSO-like behaviour in the central Pacific. A pattern of central equatorial Pacific sea surface temperature variations has recently been identified by *Ashok et al.* [2007] and defined as an 'ENSO Modoki' ('Modoki' is a Japanese word meaning 'a similar but different thing'). An El Niño Modoki is characterised by warm central Pacific waters flanked by anomalously cool SSTs to the west and east, separating the Walker Circulation into two distinct circulations. Conversely, a La Niña Modoki is characterised as having cool central Pacific waters that are flanked by warmer than normal SSTs.

The location, seasonality and magnitude of the regional Australian climate response induced by an ENSO Modoki event can be stronger than for a traditional ENSO event [*Ashok et al.*, 2009; *Cai and Cowan*, 2009; *Taschetto and England*, 2009; *Taschetto et al.*, 2009] and *Cai and Cowan* [2009] also associated La Niña Modoki events with elevated autumn rainfall in the MDB (i.e. a season not typically influenced by traditional ENSO events). However, the work on ENSO Modoki is in its infancy and there have been conflicting conclusions in the literature as to the impacts of El Niño Modoki on the Australian climate. For example, *Taschetto and England* [2009] reported strong associations between El Niño Modoki events and reduced rainfall over parts of Australia during autumn, but *Cai and Cowan* [2009] detected no significant correlations during such events. *Lee and McPhaden* [2010] also recently identified a significant increase in the intensity of central Pacific El Niño events since 1982. Given the potential links to the MDB climate, the role of Central Pacific ENSO events must be investigated further.

Inter-annual variations in southern MDB winter (JJA) and spring (SON) rainfall have been linked to Indian Ocean SST anomalies [*Nicholls*, 1989; *Simmonds*, 1990; *Drosdowsky*, 1993; *Ashok et al.*, 2000; *Drosdowsky and Chambers*, 2001; *Verdon and Franks*, 2005] and the Indian Ocean Dipole (IOD) [*Saji et al.*, 1999; *Ashok et al.*, 2003]. Verdon and Franks (2005) also showed strong relationships between east Indian Ocean SST anomalies and the northern MDB hydroclimate (in some cases stronger than the Indian Ocean-southern MDB relationships). This is contrary to the commonly accepted understanding that the influence of the Indian Ocean on the northern MDB is weaker than in the southern MDB. Either way the influence of the Indian Ocean on MDB hydroclimate needs further inverstigation, particularly with respect to the northern MDB.

The IOD is characterized by SST anomalies of opposite sign in the east and west of the Indian Ocean Basin, which are coincident with large-scale anomalous circulation patterns. During the phase of the IOD associated with cool east and warm west Indian Ocean SST anomalies, low winter rainfall over the southern MDB is likely, and vice versa for the opposite phase of the IOD [*Saji et al.*, 1999; *Ashok et al.*, 2003; *Meyers et al.*, 2007]. However, several studies show a similar modulation of rainfall with eastern Indian Ocean SSTs only [*Verdon and Franks*, 2005; *Cai and Cowan*, 2008; *Nicholls*, 2009], suggesting that the influence of the SST gradient (combined western/central and east Indian Ocean SSTs) on southeast Australian rainfall is perhaps not as important as the state of eastern Indian Ocean SSTs alone [*Nicholls*, 1989; *Verdon and Franks*, 2005].

Several recent studies have also supplied convincing evidence for the non-existence of an equatorial IOD as an independent mode of climate variability and have

342 argued that it is the result of random variations in the east and west of the Indian
343 Ocean Basin. The two nodes of the IOD are sometimes in phase (i.e. not always
344 negatively correlated), indicating the lack of a consistent dipole structure [*Dommenget*
345 *and Latif*, 2001]. Moreover, *Dommenget* [2007] and *Dommenget and Jansen* [2009]
346 have shown that the dipole structure can be reconstructed by applying the same
347 statistical technique used to decompose the mode to random noise, suggesting the
348 IOD is simply an artefact of this technique and not a physical structure. Several
349 studies have also suggested that the appearance of an IOD is simply a combination of
350 a byproduct of ENSO and random variations [*Nicholls*, 1984; *Cadet*, 1985; *Chambers*
351 *et al.*, 1999; *Allan et al.*, 2001]. This interaction is discussed further in Section 4.

352 Along with the weight of evidence suggesting the existence of the IOD as
353 erroneous, its links to Australian rainfall have recently been debated. *Ummenhofer et*
354 *al.* [2009a] used a climate model to demonstrate that the eastern pole of the IOD does
355 not regulate Australian precipitation on its own and that a slightly stronger response
356 occurs when there is a forced SST differential between the southern and eastern
357 Indian Ocean. Other studies have shown a stronger response between Australian
358 rainfall and eastern Indian Ocean SSTs [*Verdon and Franks*, 2005; *Cai and Cowan*,
359 2008; *Nicholls*, 2009] and SSTs to the north of Australia [*Smith and Timbal*, 2010]. In
360 light of these recent findings it seems that though the Indian Ocean likely has an effect
361 on the MDB climate, the existence of a dipole mechanism as an atmosphere/ocean
362 interaction is questionable. Indeed the strongest evidence suggests southern MDB
363 rainfall variations are more likely associated with SST variability in the eastern Indian
364 Ocean. We stress that, on the balance of current evidence, the Indian Ocean does play
365 a role in regulating the MDB hydroclimate but it is likely that this role is as an
366 intermediary driver only. What is in question is the modulation via an east-west dipole

(i.e. IOD) mechanism and its dependence on or independence of the Indian Ocean teleconnections from ENSO.

Variations in the Southern Annular Mode (SAM) effectively describe variations in the position of the Southern Hemisphere mid-latitude storm track [*Karoly*, 1990; *Thompson and Wallace*, 2000; *Thompson and Solomon*, 2002]. The SAM has links to MDB rainfall that vary regionally and seasonally [*Gillett et al.*, 2006; *Hendon et al.*, 2007; *Meneghini et al.*, 2007]. With a poleward contraction of the mid-latitude storm track, far southern sections of the MDB are more likely to experience lower rainfall during winter [*Hendon et al.*, 2007] due to southward displacement of rain-bearing cold fronts and cyclones. However, during the spring and summer months, anomalously poleward SAM induces changes to the local circulation that draw moist easterly winds inland and increase the likelihood of rainfall across much of the MDB, particularly in the eastern half of the Basin [*Meneghini et al.*, 2007].

The weather systems described in Section 3.1 regulate inter-annual variations in MDB rainfall as their frequency, persistence and intensity changes from year-to-year. For example, the STR changes in average strength and location from year-to-year. A stronger STR has been linked to rainfall suppression in southern Australian rainfall during autumn and early winter, while the position of the STR further north or south than normal is associated with summer rainfall variations [*Larsen and Nicholls*, 2009].

Atmospheric blocking was also previously described as a regional synoptic process affecting daily rainfall, and its importance as an inter-annual driver of the southern Australia climate has recently been recognised [*Risbey et al.*, 2009]. The

position and intensity of Southern Hemisphere blocking undergoes variations from year-to-year and decade-to-decade [*Trenberth and Mo*, 1985]. Specifically, the prevalence of blocking systems in key regions has been linked to fluctuations in the frequency of cut-off low pressure systems [*Risbey et al.*, 2008], which are responsible for a significant portion of rainfall in the southern MDB [*Pook et al.*, 2006]. The frequent occurrence of blocking highs centred near 140°E (longitude close to the southwest MDB) favours rainfall in the southern MDB [*Risbey et al.*, 2008; *Risbey et al.*, 2009].

3.3 Inter-decadal processes

In the Pacific Ocean, a coherent pattern of SST and SLP variability operating on multi-decadal timescales has been identified [*Zhang et al.*, 1997]. In the North Pacific, this variability is commonly termed the Pacific Decadal Oscillation (PDO) [*Mantua et al.*, 1997; *Mantua and Hare*, 2002]. *Power et al.* [1999b] refer to a similar Pacific Basin-wide mode of variability as the Interdecadal Pacific Oscillation (IPO). Importantly, the PDO and IPO time series are highly correlated and represent variable epochs of warming (i.e. positive phase) and cooling (i.e. negative phase) in both hemispheres of the Pacific Ocean [*Mantua et al.*, 1997; *Folland et al.*, 2002; *Franks*, 2002]. In fact, *Folland et al.* [1999] suggested that the IPO can be regarded as the Pacific wide manifestation of the PDO.

The IPO and PDO have been described as a sustained ENSO-like pattern of Pacific climate variability [*Zhang et al.*, 1997]. However, as noted by Mantua and Hare [2002], two characteristics distinguish the PDO (and IPO) from ENSO: the persistence of PDO/IPO epochs (15-30 years) and the fact that the climatic fingerprint

414 of the PDO is most dominant in the north Pacific sector with a secondary signature in
415 the tropics (whereas the opposite is true for ENSO).

416 Links between the IPO/PDO phenomena and climate variability in Australia
417 include decadal and annual-scale fluctuations in rainfall, maximum temperature, water
418 volume transport and wheat crop yield [*Power et al.*, 1999a; *Kiem et al.*, 2003; *Kiem*
419 *and Franks*, 2004; *Verdon et al.*, 2004b]. The IPO/PDO primarily influences the
420 eastern Australian climate, which includes the MDB, during the austral spring,
421 summer and autumn by inducing variations in the South Pacific Convergence Zone,
422 which tends to be active during these months [*Folland et al.*, 2002].

423 The IPO/PDO regulates the eastern Australian climate indirectly by
424 modulating both the magnitude and frequency of ENSO impacts [*Power et al.*, 1999b;
425 *Kiem et al.*, 2003; *Verdon et al.*, 2004b; *Cai and Cowan*, 2009]. When the IPO/PDO
426 is in a warm phase, the relationship between ENSO and Australian rainfall is
427 weakened, while it is strengthened during the cool phase [*Power et al.*, 1999a]. The
428 greatest effect of this modulation is a magnified response of rainfall and streamflow to
429 La Niña events during a cool IPO/PDO phase. During the cool (i.e. negative)
430 IPO/PDO phase, wet events are likely to be wetter and more frequent than during a
431 neutral or warm IPO/PDO phase, elevating flood risk in the MDB [*Kiem et al.*, 2003;
432 *Verdon et al.*, 2004b]. Conversely, during the warm (i.e. positive) IPO/PDO phase wet
433 events are less frequent and not as wet as they are during the IPO/PDO cool phase
434 which results in an increased risk of drought across the MDB and other parts of
435 eastern Australian [*Kiem and Franks*, 2004; *Verdon-Kidd and Kiem*, 2009a]. *Verdon*
436 *and Franks* [2006] confirmed that the relationships between IPO/PDO phase and the
437 frequency of ENSO events is consistent over the past 450 years by examining

paleoclimate reconstructions of the two climate modes. While Lough [2007] showed that the relationship between ENSO, IPO and rainfall/streamflow in northeast Queensland is consistent for at least the last 400 years.

Also important to note is the strong and statistically significant relationships between decadal rainfall in the MDB and SSTs in the Tasman Sea north of New Zealand and south of 15°S – wetter decades are associated with warmer SSTs in the Tasman Sea [*Power et al.*, 1999a]. This Tasman SST-MDB rainfall relationship is also evident on inter-annual time scales. The effect of this association on regional circulation patterns is not yet understood. However, it is reasonable to suppose that larger rainfall totals would be caused by warmer, moist air being transported to, and converging over, the Australian mainland [*White et al.*, 2003].

Decadal and multi-decadal scale variations in SSTs and surface pressures around Australia that are reminiscent of an ENSO pattern have been linked to multi-year summer-time droughts in eastern Australia [*White et al.*, 2003; *White et al.*, 2004]. Such droughts may be due to random fluctuations of ENSO [*Kestin et al.*, 1998] or they may be due to some other as yet unknown process(es). The established statistical relationships suggest a connection between decadal-scale SST variability and fluctuations in MDB rainfall. However, the physical mechanisms which cause the decadal-scale SST variability and the links between changes in SST and regional-scale variability in synoptic patterns (and rainfall) are currently unknown and require further investigation.

In addition to the natural variability described above, anthropogenic influences will very likely also be responsible for future multi-decadal scale climate variations in the MDB [*CSIRO and BoM*, 2007; *IPCC*, 2007]. Anthropogenically-induced changes

462 in some aspects of the regional Australian climate have already been detected [*Karoly*
463 *and Braganza*, 2005]. There are a wide range of projected impacts for the MDB
464 climate (i.e. 'wetter and warmer' or 'drier and warmer') and unfortunately, there is no
465 clear signal in the direction of these changes across the MDB.

466 The uncertainties primarily stem from large differences between future rainfall
467 changes in the MDB projected by global climate models in response to future
468 greenhouse gas emission scenarios. While these projections consistently show
469 increasing temperatures across the MDB over the 21st century, they do not show
470 consistent changes in rainfall [*Sun et al.*, 2011]. One regional exception is the
471 southern MDB, for which many climate models consistently project drying. However,
472 the magnitude of this drying varies considerably between models [*Chiew et al.*, 2008]
473 and the dynamical mechanisms behind this drying have not yet been identified. Given
474 the potential for anthropogenic warming to be associated with significant negative
475 impacts across the MDB [*Alexander and Arblaster*, 2009; *BoM*, 2010], the uncertainty
476 associated with climate model projections must be addressed so that the contribution
477 of anthropogenic forcing to inter-annual, multi-decadal and longer changes in MDB
478 can be properly assessed. This numerous sources of uncertainty associated with global
479 climate model projections for the MDB are discussed further in Section 5.

480

481 # 4. Interactions between hydroclimatic processes: a missing link in our
482 # understanding?

483 ## 4.1 Interactions between climate processes

As single processes alone account for less than 20% of monthly rainfall variability [*Risbey et al.*, 2009], it is unlikely that the climate drivers influential on the MDB are independent of each other or act to drive the MDB climate in isolation [*Risbey et al.*, 2008; *Verdon-Kidd and Kiem*, 2009b; *Kiem and Verdon-Kidd*, 2010]. Deducing the interactions between climate drivers and their subsequent impact on the Australian hydroclimate is only an emerging area of research, but one that is crucial to better capturing the causes of variations in the MDB climate.

There is increasing evidence that the mean state of the atmosphere/ocean system, which is partially modulated by the inter-annual and inter-decadal processes previously described, plays a part in regulating the occurrence of daily weather events. For example, the large-scale climate processes described in Section 3 have been linked to variations in the positions of the jet streams, regulation of moisture availability and other aspects of the atmosphere/ocean system, as well as variations in weather systems [*Risbey et al.*, 2008; *Verdon-Kidd and Kiem*, 2009a].

The characteristics of weather systems, including their location, frequency of occurrence and intensity, show dependence on intra-seasonal, inter-annual and decadal drivers of climate such as the MJO, ENSO and the IPO/PDO [*Nicholls and Kariko*, 1993; *Pezza et al.*, 2008]. The intensity of cyclones and anticyclones in the mid-latitudes increases and their frequency decreases when there is a strong warm phase of the PDO [*Pezza et al.*, 2007]. *Verdon-Kidd and Kiem* [2009a] also found that warm phases of the IPO decrease the likelihood of wet weather events in parts of the MDB, while cool phases increase this likelihood, and vice versa for dry weather events.

Variations in the prevalence of atmospheric blocking have been linked to the state of ENSO, with blocking frequency and associated rainfall becoming more prominent during La Niñas and less prominent during El Niños [*Risbey et al.*, 2008]. Weather features such as pre-frontal troughs and 'easterly-dip'-type patterns are more prominent during La Niñas, and broad high pressure systems across southeastern Australia and a northward retraction and weakening of easterly troughs are more prominent during El Niños [*Verdon-Kidd and Kiem*, 2009a]. The wetter and drier phases of SAM and Indian Ocean SSTs are similarly linked to weather systems that are more or less likely to enhance or suppress rainfall in the MDB [*Hendon et al.*, 2007; *Verdon-Kidd and Kiem*, 2009a; *Kiem and Verdon-Kidd*, 2010].

We present analysis in Figure 5 demonstrating the connections between climate processes operating on inter-annual and daily time scales. *Pook et al.* [2006] found that at least 80% of April–October heavy rain days (defined as days with accumulations above 25 mm) were associated with cut-off low-pressure systems in a domain covering the far southwest MDB. They also showed that cut-off low-pressure systems strongly modulate inter-annual rainfall there. Given that Section 3.2 outlined the modulation of inter-annual rainfall in the MDB by ENSO, SAM and the Indian Ocean, a logical hypothesis is that these remote drivers partially act by modulating the number of cut-off low-pressure systems that influence the region. This influence was examined here by using heavy rain days as a proxy for cut-off lows.

The number of heavy rain days was calculated at seven stations for every April–October period from 1970–2002, the same period used by *Pook et al.* [2006] for their analysis. All stations fall within the *Pook et al.* [2006] domain and come from the *Lavery et al.* [1992] high-quality daily rainfall network. The average number of

heavy rain days during years in a particular extreme state of the SAM, ENSO and eastern Indian Ocean SSTs were compared to neutral years for each index. The SAM, ENSO and eastern Indian Ocean SSTs were represented by the SAM index of *Marshall* [2003], the Southern Oscillation Index (SOI) of *Troup* [1965] and the eastern pole of the Dipole Mode Index (DMIeast) of *Saji et al.* [1999], generated from the global HadISST sea-suface temperature data set [*Rayner et al.*, 2003]. A positive (negative) phase of each driver was defined if the standardized index was above (below) 0.5 (-0.5) standard deviations, giving about 30% of years in each of the positive or negative phases. All indices were standardized over the 1970–2002 period. To determine if the mean frequency of heavy rain days during a particular phase of a driver was statistically significantly different from neutral years, a 95% confidence interval was generated from neutral years. The mean number of rain days from 1000 bootstrap replicates of *n* neutral years were used, where *n* was the number of years when the inter-annual driver was in its positive or negative phase.

Every station in the domain had a statistically significant link between April–September heavy rain days and ENSO, with a higher than average number of heavy rainfall days during La Niña years and a lower than average number during El Niño years (Figure 5). This result was the same whether SOI or SSTs from the Nino3.4 region were used to represent ENSO and is consistent with the links described between ENSO and cut-off lows in the region by *Pook et al.* [2006]. All but one station had a lower than average number of heavy rain days during negative SAM events, with three stations showing a statistically significant difference. These results may reflect the influence of the SAM during the austral spring, where negative SAM is associated with decreased rainfall in the region [*Hendon et al.*, 2007]. The majority of stations also had significantly more heavy rain events during years with warmer

than normal eastern Indian Ocean SSTs and fewer heavy rain events when these SSTs were cooler than normal. The results here suggest that all three major drivers of interannual climate variations in the region partially regulate rainfall by changing the frequency of heavy daily rainfall events, which in turn are likely to be via modulation of the frequency of cut-off low pressure systems.

The above analysis showed that large-scale climate drivers play a role in regulating daily weather events. However, the large-scale drivers themselves also undergo decadal and multi-decadal scale variations. These variations may be largely independent or they may co-occur, indicating relationships between processes that, in turn, affect regional climates.

Figure 6 shows fluctuations in the relationships between indices representing ENSO, SAM, Indian Ocean SST variability and STR location and intensity for running 30-year periods from 1905–2004. The indices were chosen so the longest period possible could be analysed. The SOI represents ENSO variations and an extended reconstruction of the Marshall [2003] SAM index by *Jones et al.* [2009b] was used as the Marshall [2003] index is available from 1958 only. The eastern pole of the DMI [*Saji et al.*, 1999] represents eastern Indian Ocean SSTs. Despite the fact that we are reluctant to accept the IOD as a physical atmosphere/ocean processes for the reasons outlined in Section 3, we also include the DMI (representing the IOD) in our analysis for a later discussion surrounding evidence that any large-scale climate variations in the Indian Ocean may be partially linked to ENSO. The latitude and intensity of the STR [*Drosdowsky*, 2005] is also assessed as an inter-annual driver of MDB rainfall variations. All indices were detrended by removing the line-of-best-fit

579 over the 1905–2004 period, computed using linear regression, to highlight
580 relationships on inter-annual time scales only.

581 The fluctuations in the relationships between inter-annual drivers of the MDB
582 climate vary between pairs of drivers and season. For example, summer (December–
583 February) STR intensity has a statistically significant relationship with SAM that has
584 remained fairly consistent from 1905–2004. Conversely, the relationship between the
585 inter-annual variations in ENSO and the eastern Indian Ocean SSTs were barely
586 statistically significant in the 1930s (correlations around -0.35), while the 30 years
587 centred on 1970 had a correlation of almost -0.9. These indicate large changes in the
588 relationship between ENSO and the eastern Indian Ocean during the 20th Century.
589 There are consistent statistically significant relationships between SAM and STR
590 intensity from winter through to summer (June–February) and between SAM and STR
591 latitude during winter and spring (June–November), perhaps highlighting broader
592 scale fluctuations in the Hadley circulation. Interestingly, there has been a consistent
593 strengthening of the relationship between the SAM and STR latitude during the
594 austral autumn (March–May) where correlations in the early 1900s were insignificant
595 and negative and were statistically significant and positive from around 1960 (Figure
596 6). As the indices were detrended prior to computation of the correlations, the
597 strengthening of this relationship over time is probably not associated with the trends
598 that have occurred in both processes since around this time [*Marshall*, 2003; *Larsen*
599 *and Nicholls*, 2009]. The trend may reflect poor data coverage for calculation of the
600 SAM prior to the 1960s. However, if this were the case, a step-function, rather than a
601 monotonic trend might be expected.

Previous authors have shown how interactions between various large-scale climate drivers impact on the MDB climate [*Risbey et al.*, 2009; *Silvestri and Vera*, 2009; *Verdon-Kidd and Kiem*, 2009a]. We have also previously argued that the IOD is not a physical atmosphere/ocean interaction (Section 3) with many studies providing evidence that it is a combination of a byproduct of ENSO and stochastic variations [*Nicholls*, 1984; *Cadet*, 1985; *Chambers et al.*, 1999; *Allan et al.*, 2001]. The extreme phases of ENSO and IOD are also often 'phase locked' meaning that there is a high probability that a La Niña will occur with warm, and an El Niño with cool, eastern Indian Ocean SSTs [*Meyers et al.*, 2007]. As these coinciding phases are indicative of wet or dry conditions for both ENSO and the IOD, the result is often an amplification of the rainfall signature [*Meyers et al.*, 2007]. The synchronicity between extreme ENSO and IOD variations suggests dependence between the two and *Allan et al.* [2001] reported significant lag correlations in ENSO and IOD indices.

Using the IOD and ENSO indices previously described from 1905–2004, we found statistically significant correlations of -0.57 between spring indices, and -0.59 when a lag correlation was calculated between the winter (June-August) SOI and the DMI in the subsequent spring (September-November). Both correlations are strongly indicative of dependence between the two during the time when the Indian Ocean has its greatest influence on MDB rainfall. Furthermore, temporal variations in correlations between September–November SOI and DMI for running 30-year periods from 1905–2004 show consistent statistical significance from at least 1960, when data is most reliable (Figure 6). Moreover, when the SOI was correlated against the eastern pole of the DMI only (i.e. eastern Indian Ocean SSTs) there were statistically significant relationships during winter, spring and summer.

However, others have presented evidence that Indian Ocean SST anomalies can occur irrespective of the state of the tropical Pacific Ocean [*Nicholls*, 1989; *Saji et al.*, 1999; *Webster et al.*, 1999]. *Saji et al.* [1999] report that the IOD is the most active during the austral winter (JJA) and that there is a lack of statistically significant correlations between winter IOD and ENSO indices, potentially indicating independence. *Webster et al.* [1999] demonstrated that the Indian Ocean can exhibit strong ocean-atmosphere-land interactions that are self-maintaining and capable of producing large perturbations that are independent of ENSO. *Fischer et al.* [2005] used a GCM to study the triggers of the IOD and showed that two types of IOD may actually exist, one entirely independent of ENSO and the other a consequence of tropical Pacific conditions (i.e. ENSO). This is supported by the fact that the IOD is not always phase locked with ENSO and that there are few incidences where a La Niña coincides with a positive IOD and vice versa (i.e. few incidences where El Niño coincides with negative IOD). Regardless, none of these studies provide a physical mechanism for the IOD as a climate process that is independent of ENSO. As such, we continue to argue for the non-existence for an IOD and like previous studies, suggest that any apparent 'IOD' is probably the result of combined ENSO influences and stochastic variations, or random variations alone. However, given the conflictions in the literature, more work needs to be undertaken to establish if this is the case. As mentioned previously (Section 3.2) this is not to say that the Indian Ocean is not an important influence on MDB hydroclimate but that the influence is more likely to be driven by eastern Indian Ocean SSTs rather than the east-west dipole (i.e. IOD). This is demonstrated by Verdon and Franks (2005) who found that the relationship between eastern Indian Ocean SSTs and eastern Australian hydroclimate still holds when ENSO events are removed.

651 The interaction between SAM and ENSO and their combined impact on
652 Australian rainfall has received little attention in the literature. From correlation
653 analysis performed here, the relationship between SAM and ENSO is not significant
654 during any season, except perhaps in more recent years during the austral autumn
655 (Figure 6). Correlations between ENSO and SAM using data from 1905–2004 at lags
656 of one to three months were generally not statistically significant. Again, a potential
657 exception is during the austral autumn (March–May) when the correlation between
658 SAM and ENSO lagged by one and two months was marginally significant at the 95%
659 level ($r = 0.21$ and $r = 0.22$ respectively, $n = 100$). However, neither driver shows a
660 statistically significant relationship with rainfall in the MDB during this season
661 [*Hendon et al.*, 2007; *Meneghini et al.*, 2007; *Risbey et al.*, 2009]. In general, the
662 evidence presented here suggests that SAM and ENSO are largely independent modes
663 of variability. However, *Karoly* [1989] described covariations between the Southern
664 Hemisphere middle latitude circulation and ENSO events, which perhaps indicates a
665 relationship between SAM and ENSO in summer.

666 Despite the fact that ENSO and SAM appear to act in an independent manner,
667 there are competing effects between the two that influence MDB seasonal rainfall.
668 *Kiem and Verdon-Kidd* [2009; 2010] reported that dry conditions during autumn are
669 more likely if an El Niño event occurs in combination with a positive SAM. Previous
670 La Niña events have not necessarily been associated with above-average rainfall in
671 the southern MDB [*Gallant and Karoly*, 2009], which may be due to interactions with
672 coinciding SAM events that place the storm track further south than normal [*Kiem*
673 *and Verdon-Kidd*, 2009; 2010].

674 An example of the impacts of competing SAM and ENSO phases on MDB
675 rainfall are illustrated in Figure 7. The rainfall anomalies were generated using the
676 operational Australian Bureau of Meteorology AWAP rainfall grids [*Jones et al.*,
677 2009a] from 1905–2004. The *Jones et al.* [2009b] SAM index, described previously
678 as an extension of the *Marshall* [2003] index, was used to maximise the number of
679 years available for analysis. As previously, the SOI represents ENSO variability. A
680 positive SAM or ENSO event was defined when the standardized indices (relative to
681 the 1905–2004 period) exceeded 0.5 standard deviations. However, the results were
682 similar when thresholds of between 0.5 and 1 standard deviation were applied and
683 when SSTs from the Niño3.4 region were used to represent ENSO.

684 In all seasons, when combinations of the extreme phases of each driver co-
685 occurred the SAM typically enhanced or tempered the impact of ENSO, depending on
686 the phase of the SAM. However, the extent of this influence depended on the season.
687 Specifically, the strength of the relationship between MDB rainfall and each process
688 tended to dictate the influence each had on seasonal rainfall. For example, neither the
689 SAM nor ENSO, when considered in isolation, have a strong relationship with
690 autumn (March–May) rainfall in the MDB but when SAM-ENSO interactions are
691 analysed some autumn rainfall variability is explained (e.g. *Kiem and Verdon-Kidd*,
692 2009, 2010]. In comparison, winter (June–August) rainfall has a stronger relationship
693 with both SAM and ENSO [*Risbey et al.*, 2009], making competing influences more
694 noticeable (Figure 7). During the austral winter the southern MDB was typically drier,
695 while the northern MDB was wetter, for a co-occurring La Niña and positive SAM
696 event, compared to a La Niña event occurring in isolation. Conversely, the drier
697 conditions in the southern MDB during past El Niño events have been enhanced by
698 positive SAM and tempered by negative SAM.

699 Note that the frequencies at which combined events occurred were small in
700 several cases. So, producing composites from *n* randomised years tested the statistical
701 significance, where *n* was equal to the number of years used to produce each
702 composite (see Figure 7). Statistical significance was defined if the value of the
703 composite being tested was larger/smaller than 97.5% of the randomised composites.

704 During combined positive SAM and El Niño, and negative SAM and La Niña,
705 the rainfall anomalies in Figure 7 were statistically significant across most of the
706 MDB (not shown). Interestingly, during JJA, the extreme phases of ENSO during
707 neutral SAM had few statistically significant rainfall anomalies in the southern MDB.
708 Thus, across the MDB, particularly the southern half, there is significantly enhanced
709 drying and wetting respectively when positive SAM/El Niño or negative SAM/La
710 Niña phases co-occur.

711 For combined positive SAM/La Niña events, the rainfall anomalies were not
712 statistically significantly different in the southern MDB. However, when La Niña
713 combined with negative or neutral SAM conditions it was statistically significantly
714 wetter in this region, indicating that positive SAM may have some tempering effect.
715 Similarly, for combined negative SAM/El Niño years, significant drying was limited
716 to the far northeast of the Basin. However, for neutral SAM/El Niño, significant
717 drying extended across the northern and eastern two-thirds of the Basin and across
718 most of the MDB for a positive SAM/El Niño combination.

719 Interactions between climate processes have also been demonstrated between
720 drivers operating on decadal and inter-annual time scales. The IPO/PDO regulates the
721 eastern Australian climate indirectly by modulating both the magnitude and frequency
722 of ENSO impacts [*Power et al.*, 1999b; *Kiem et al.*, 2003; *Verdon et al.*, 2004b; *Cai*

723 *and Cowan*, 2009]. When the IPO/PDO is in a warm phase, the relationship between
724 ENSO and Australian rainfall is weakened, while it is strengthened during the cool
725 phase [*Power et al.*, 1999a]. The greatest effect of this modulation is a magnified
726 response of rainfall and streamflow to La Niña events during a cool IPO/PDO phase.
727 During the cool (i.e. negative) IPO/PDO phase, wet events are likely to be wetter and
728 more frequent than during a neutral or warm IPO/PDO phase, elevating flood risk in
729 the MDB [*Kiem et al.*, 2003; *Verdon et al.*, 2004b]. Conversely, during the warm (i.e.
730 positive) IPO/PDO phase wet events are less frequent and not as wet as they are
731 during the IPO/PDO cool phase which results in an increased risk of drought across
732 the MDB and other parts of eastern Australian [*Kiem and Franks*, 2004; *Verdon-Kidd*
733 *and Kiem*, 2009a].

734 Given the temporal fluctuations in the relationships between the major climate
735 drivers, further work is needed to understand the mechanisms that cause these
736 fluctuations and how they impact on regional climates, such as the MDB. The long-
737 term variations could stem from random variations [*Kestin et al.*, 1998], interactions
738 with unknown climate processes on numerous time scales [*Power et al.*, 1999a], other
739 external forcings [*Meehl et al.*, 2006], or a combination of all or some of these. Also
740 unclear at this stage is how large-scale climate processes such as ENSO, Indian Ocean
741 SSTs, SAM, STR and IPO/PDO might be affected by anthropogenic climate change.

742 **4.2 Hydrological drivers and their interactions with climate processes**

743 Interactions influencing the MDB hydroclimate (e.g. rainfall, temperature,
744 runoff, streamflow etc.) are not limited to those between climate processes.
745 Hydrological and land surface processes also play an important role in regulating
746 hydroclimatic variability in the MDB. Rainfall and surface temperature persistence

747 are related to soil moisture, which can retain some memory of the recent climate
748 [*Timbal et al.*, 2002]. Antecedent land surface conditions play an important role in
749 runoff and streamflow variability [*Chiew et al.*, 1998; *Kiem and Verdon-Kidd*, 2009].
750 *CSIRO* [2008] also suggests that it is likely that after a prolonged dry period, there is
751 less connectivity between the subsurface storage and the river system, and therefore
752 significant amounts of rainfall and recharge are required to fill the subsurface storage
753 before runoff can occur.

754 It is likely that vegetation feedbacks also play a role in regulating regional
755 climate and that extensive land clearance across the Australian continent has
756 contributed to some long-term climatic trends [*Murphy and Timbal*, 2008]. Rainfall
757 recycling feedbacks (and the memory the system has of previous rainfall) is poorly
758 understood because the soil moisture-vegetation-rainfall feedbacks are complex.
759 However, it is believed to be important in determining downwind rainfall in
760 continental areas [*Koster et al.*, 2004]. This is an area of ongoing research that also
761 has, as an added complexity, the difficulties and computational limitations associated
762 with integrating land surface models with climate models.

763

764 **5. A framework for focussing climate information for natural**
765 **resource management**

766 Knowledge of the MDB hydroclimate, and its driving mechanisms, assists
767 NRM in developing appropriate management strategies. As this environment changes
768 management strategies must be adapted. Given that the MDB has large natural climate
769 cycles, and more recently the possibility of anthropogenic influences, it is critical that

NRM is better able to isolate the potential causes of extremes or changes in the climate.

There are numerous, complex processes modulating the MDB hydroclimate (see Sections 3 and 4) and sifting through this vast amount of knowledge can be an arduous task. So, we present a broad framework that summarizes the relevant climatic information and assists in isolating the possible causes of a climatic event. It is important to note that this is a simplified framework for a complex problem and will not provide definitive answers. This is because the processes previously presented do not equal the sum of knowledge required for complete understanding of a climate system that is substantially chaotic and involves complex non-linear and non-stationary interactions – as well as several significant knowledge gaps. However, the framework does allow the user to qualitatively estimate the likelihood of a cause of a climatic event. We stress that this framework is not intended to be a decision-making tool – only to better focus understanding.

The framework breaks down a climatic event into its fundamental components, isolating i) spatial signatures, ii) temporal signatures and iii) small-scale evolution. Spatial information identifies the regional coverage of the event, including where it occurred and over what size area. Temporal signals identify the persistence of the event, including any seasonality or recurrence. For example, a single season might experience an extreme climatic event but this may reoccur for multiple years. Small-scale evolution deconstructs the event to isolate the small-scale features. For example, the make-up of the daily rainfall distribution causing a seasonal anomaly – is the deficit due to a lack of rain days, or a decrease in rainfall intensity or

magnitudes? Small-scale evolution also includes assessing changes in weather systems.

Following the characterisation of an event using the above framework, the event's features can be compared to those typical of the known climatic drivers described in Sections 3 and 4. This comparison allows the user to rank the possible mechanisms, where the ranks are based on the consistency between a feature and the known impacts of a climatic process. Allowances must be made where the consistency between the fundamental elements of the event and the climatic driver are unknown or the uncertainty is large. The consideration that there are likely to be processes (or interactions and feedbacks) of which we are currently unaware that are affecting the event must also be made. These caveats are discussed in the case study presented below.

5.1 Case study: identifying potential causes of the Big Dry (c. 1997–2010)

The framework presented above is now applied to a case study, namely, the Big Dry (c. 1997–2010). Here, we focus solely on the rainfall deficits to simplify the framework for presentation. However, formal event attribution for NRM should be applied to multiple variables (i.e. incorporate knowledge of other changes such as soil moisture, temperatures etc.). The major characterising features of the Big Dry are identified below and each feature is tagged 1–7 for later comparison with climate processes.

1. Spatial signatures

The rainfall deficits were mainly confined to the southern MDB[1]. The largest deficits were in the far southeast and southwest of the region (Figure 3). Note that drought was also experienced in the northern MDB from 1997-2006, however, for this

817 exercise the northern MDB drought is considered to be a different event to the Big
818 Dry, which is characterised by rainfall deficits in the southern MDB from 1997-2010.

819 *2. Temporal signatures*

820 The 13-year rainfall deficit was primarily driven by consistently lower than
821 normal rainfall during autumn[2]. However, several individual years also experienced
822 deficits during winter and spring[3] [*Verdon-Kidd and Kiem*, 2009b]. While deficits
823 generally did not last more than a few consecutive seasons, there was a distinct lack of
824 periods with very high rainfall[4] [*Gallant and Karoly*, 2010].

825 *3. Small-scale evolution*

826 There was a lack of high one-day rainfall totals[4] [*Murphy and Timbal*, 2008;
827 *Verdon-Kidd and Kiem*, 2009b], which is consistent with a reduction in the amount of
828 rainfall associated with cut-off low pressure systems over this period[5] [*Pook et al.*,
829 2006; *McIntosh et al.*, 2007] and an absence of persistent pre-frontal troughs that aid
830 the penetration of rain producing cold fronts into the southern MDB[6] [*Verdon-Kidd*
831 *and Kiem*, 2009a; *Alexander et al.*, 2010]. The recent drying has also coincided with
832 an increase in regional surface pressure[7] [*Drosdowsky*, 2005; *Timbal and Murphy*,
833 2007; *Larsen and Nicholls*, 2009; *Williams and Stone*, 2009; *Alexander et al.*, 2010].
834 The Big Dry has also been associated with abnormally high temperatures [*Nicholls*,
835 2004; *Gallant and Karoly*, 2009]. However, the causal mechanisms behind the
836 relationship between elevated temperatures and the hydroclimate (e.g. evaporation
837 and streamflow) are largely unknown when compared with changes in other
838 atmospheric conditions such as radiation, humidity and wind [*Donohue et al.*, 2010].
839 This is another area in need of further study.

As the drought primarily occurred on inter-annual and longer time scales with symptomatic small-scale changes (i.e. weather systems), Table 1 present simplified summaries of climate processes (from Section 3) relevant to these time scales. Table 2 then compares the seven main features of the Big Dry to this information, thus focusing causal candidates. Note that while the simplified comparison in Table 2 identifies where links between drought features and climate processes are unknown, it does not incorporate the uncertainties described in Sections 3 and 4, which are very important (see later discussion). Further, longer term trends in some inter-annual processes have recently been identified including an increase in the intensity of the STR [*Drosdowsky*, 2005; *Timbal and Murphy*, 2007; *Larsen and Nicholls*, 2009; *Williams and Stone*, 2009] and a poleward retraction of the mid-latitude storm belt [*Frederiksen et al.*, 2011] which is consistent with trends towards a more positive SAM [*Marshall*, 2003]. As such, these trends have also been included and assessed in Table 2 as potential causal mechanisms of the Big Dry.

Based on the number of consistencies, unknowns and inconsistencies between each drought feature and climate process indicated in Table 2 it is then possible to rank the most likely causal mechanism(s) of the Big Dry with the highest ranks given to those processes with the most consistencies and least inconsistencies. Note this method is biased towards those processes that are most well understood and as such processes that included many unknowns should not necessarily be ruled out as possible causes.

Using the above approach the most likely causal mechanisms behind the Big Dry were (in order); a southerly shift in the storm track; ENSO/SAM interactions, increases in STR intensity; anthropogenic climate change; more positive SAM; and

the IPO/PDO. Less likely was Indian Ocean variability and ENSO, when considered as isolated processes, mainly due to the inconsistencies in the seasonality of the most persistent rainfall deficits. While it is likely that these processes contributed to a worsening of the drought, on their own they were probably not responsible for the majority of the rainfall deficits [*Verdon-Kidd and Kiem*, 2009b]. Also worth noting, though there is no evidence of this in existing literature, is the possibility that the Big Dry was caused by interactions between IPO, ENSO and SAM. This inference is based on the following reasoning: IPO modulates the magnitude and frequency of ENSO impacts, the impacts of ENSO are also modulated by SAM and vice versa, therefore it is possible that IPO also influences SAM impacts (either directly or via IPOs modulation of ENSO impacts). This hypothesis is currently under investigation.

The high ranking of ENSO/SAM interactions supports the point made previously (see Section 4, Figure 5, Figure 6, Figure 7), that all climate processes have some degree of interaction and should not be taken as isolated causes. For example, the three La Niña events occurring during the Big Dry were been associated with unusually low rainfall in the context of historical events [*Gallant and Karoly*, 2009]. *Kiem and Verdon-Kidd* [2009] attribute this to a compounding influence from a positive SAM and this is consistent with Figure 7. The lack of La Niña events during the 1990s (i.e. the initiation of Big Dry) is also consistent with the IPO being in a positive phase [*Kiem et al.*, 2003].

The rankings from Table 2 are also in agreement with *Nicholls* [2009] and *Verdon-Kidd and Kiem* [2009b] who suggested that a consistently positive SAM since the 1990s, associated with a poleward retraction of the mid-latitude storm track [*Frederiksen et al.*, 2011], may be responsible for the recent trends in southeast

888 Australian rainfall. However, there are some conflicts between these results and other
889 studies that show there are no significant associations between southeast Australian
890 autumn rainfall and the SAM [*Hendon et al.*, 2007; *Timbal and Murphy*, 2007; *Risbey*
891 *et al.*, 2009]. However, the critical difference is that the studies finding no significant
892 associations have only examined the SAM in isolation as opposed to considering
893 interactions with other processes (e.g. Figures 6 and 7), which potentially
894 enhance/suppress typical impacts [*Kiem and Verdon-Kidd*, 2010].

895 The association between the intensity of the STR and rainfall variations has
896 also previously been suggested as a potential cause [*Timbal and Murphy*, 2007;
897 *Timbal et al.*, 2007; *Murphy and Timbal*, 2008; *Larsen and Nicholls*, 2009]. Though
898 plausible, it is unclear whether an increase in STR intensity is the ultimate cause.
899 Indeed, Figure 6 shows positive relationships between STR intensity and SAM during
900 autumn and winter that indicate that the SAM is associated with up to 44% of the
901 variance in STR intensity (the largest rainfall declines have been reported during late
902 autumn and early winter [*Murphy and Timbal*, 2008; *Verdon-Kidd and Kiem*, 2009b]).

903 All hypotheses of the cause of the Big Dry have a common similarity. That is,
904 all involve decadal scale variations in the major climate drivers and how they
905 influence southern MDB rainfall, and that the causes are possibly linked, either via
906 interactions with each other or via some ultimate cause, which may include
907 anthropogenic climate change [*Thompson and Solomon*, 2002; *Arblaster and Meehl*,
908 2006]. Indeed, the possible climate processes responsible for the drought indentified
909 in Table 2 are probably a combination of both natural and anthropogenic sources.
910 However, although anthropogenic climate change was identified as a possible or

911 partial cause of the drought there are very large uncertainties stemming from many
912 sources that surround this statement.

913 The most obvious source of uncertainty is that there are numerous aspects of
914 the climate system that are still not well understood and probably many more that are
915 yet to be identified. This is particularly true of decadal and longer-scale processes for
916 which knowledge is fundamentally limited by a lack of observations (e.g. the MDB
917 only has 110 years, or 11 independent decades, of climate records). This makes it
918 particularly difficult to gauge the true extent of decadal-scale variations in MDB
919 hydroclimate (i.e. to put the Big Dry into context) and to differentiate between the
920 impacts of natural inter-decadal to centennial variability and anthropogenic climate
921 change. This can be overcome by continuing two major research streams.

922 The first involves complementing the limited instrumental record
923 (approximately 100 years) by extending the historical record using various sources of
924 palaeoclimate information [*Verdon and Franks*, 2006; *Lough*, 2007; *Gallant and*
925 *Gergis*, 2011]. *Verdon and Franks* [2006] estimated variations in east Australian
926 rainfall using reconstructions of the PDO for the past 400 years. More recently,
927 *Gallant and Gergis* [2011] and *Gergis et al.* [2011] developed an experimental
928 reconstruction of River Murray streamflow and southeast Australian rainfall
929 respectively, and showed that the streamflow deficits associated with the Big Dry
930 were very rare, but not necessarily unprecedented, during the period 1783–1988. In
931 both studies, palaeo records indicate there have been large decadal-scale variations in
932 the east Australian hydroclimate prior to the instrumental record. Palaeoclimate
933 reconstructions are useful for estimating the bounds of past climate variability.
934 However, interpreting the physical processes behind the pre-instrumental variations is

935 difficult. This is something that can only be properly studied via complementary
936 modelling studies.

937 To date, much of our understanding of the processes driving the MDB climate
938 has been established through the identification of consistent patterns and modes of
939 variation in the atmosphere/ocean system. As such, the second approach to
940 understanding decadal-scale hydroclimate variations, which as mentioned is difficult
941 with limited instrumental data, is via studies utilising climate models that represent
942 the fundamental physics of the atmosphere/ocean system. However, this presents a
943 further problem, as climate models can still not sufficiently model all these processes
944 that are known to drive regional hydroclimatic variability. A subsequent issue is that
945 their teleconnections with regional climates are not well captured. Thus, while climate
946 models may suggest a particular future change that is consistent with a present-day
947 trend, it is difficult to ascertain whether the mechanisms behind these trends in model-
948 space are consistent with reality. This is a knowledge gap that requires immediate
949 attention.

950 Given the lack of understanding about the physics behind both anthropogenic
951 and natural processes, we suggest that if a framework similar to that presented here is
952 utilised by NRM, that a weighting should be applied to each process that is inversely
953 proportional to the uncertainty (i.e. the higher the uncertainty the less weight given to
954 that causal process). This weighting factor may be arbitrary or based on some
955 quantitative measure of uncertainty associated with each process. Developing such
956 weighting factors is a complex process and is outside the scope of this paper. So at
957 present, the best we can do given the current state of knowledge is to propose how the
958 processes we understand, and the interplay between them, fit in the context of this

959 event (Table 2). However, it is important to remember that the certainty of our
960 attribution is proportional to our understanding and hence, is limited by it.

961 ## 6. Concluding remarks

962 This paper demonstrates advances in our understanding of the climate
963 processes influencing the MDB, however, it also highlights that there is much that is
964 not well understood. The framework presented in Section 5 is designed to improve
965 understanding to better focus NRM. So, in conclusion we summarise the key areas of
966 research that we believe will assist in improving the reliability of hydroclimatic
967 insights and forecasts for the MDB (on seasonal to decadal scales) and better
968 quantification of climate related risk for NRM.

969 *i) Expanding data records – instrumental and palaeoclimate-proxies*

970 A maximum of 60 to 110 years of instrumental data are available in the MDB
971 and other regions that regulate its climate (e.g. the Southern Ocean). As such, reliable
972 estimates of climate variations on multi-decadal and longer time scales are not
973 possible due to limited degrees of freedom. This is a problem for water management,
974 which often assumes that the instrumental hydroclimate records are representative of
975 current and future climatic conditions. Given the lack of long- term instrumental data,
976 research investment into expansion of the instrumental record and the use of palaeo
977 records is recommended.

978 *ii) Improving understanding of interactions between climate drivers*

979 In Section 4.1, we performed some simple analysis to highlight interactions
980 between the processes driving the MDB climate on multiple time-scales. From this
981 analysis, it is clear that a better understanding of climate driver interactions on all time

scales is required. Until such knowledge is gained it is difficult to: i) know how to assess their impacts relative to one another; ii) determine the extent of their independence and how their contributions vary spatially and temporally; iii) discriminate between impacts associated with natural and anthropogenic climate variations; iv) develop (and/or improve) climate models to have greater confidence in future climate projections – it is unclear how climate models (either statistical or dynamical) can simulate climate process that we do not yet properly understand.

iii) Improving understanding of interactions between climatic and hydrological processes

Section 4.2 described the limited body of work examining interactions between ocean-atmosphere circulation patterns and hydrological/land-surface processes. Given its potential importance, further examination of i) connections between antecedent hydrological conditions and/or land cover change and regional climates; ii) the mechanisms driving the apparent relationship between increased temperature and decreased streamflow; iii) the possibility of a rainfall recycling phenomena [*Koster et al.*, 2004] in the MDB, is required.

iv) Improving the representation of climate processes in climate models

Climate models provide a complementary way of examining variations in MDB hydroclimate. However, improving how climate models simulate the processes that drive the MDB hydroclimatic variability is crucial, as is the need for regionally specific information. Improved modelling capabilities will assist with the previous three research aims and vice-versa.

The research areas outlined above focus on improving understanding of natural climatic variations, the processes behind these, and possible future changes. These research areas require urgent investigation as it is clear that there is still a lot we do not understand with respect to historical climate patterns and causal processes in the MDB – and even less is known about the future.

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FIGURE CAPTIONS

Figure 1. The Murray-Darling Basin (dashed outline) overlaid on mean annual January – December (1900– 2010) observed rainfall (mm), from the Australian Bureau of Meteorology Australian Water Availability Project rainfall grids [*Jones et al.*, 2009a]. Adapted from Draper and Mills [2008].

Figure 2. Area-average annual (January – December) rainfall (thin grey line) and decadal mean rainfall (thick black line) for the Murray-Darling Basin from 1900 to 2010, generated from the Australian Water Availability Project rainfall grids [*Jones et al.*, 2009a]. The dashed line indicates the mean annual rainfall total over the 1900 – 2010 mean (471 mm).

Figure 3. The 13-year June 1997 to May 2009 rainfall totals shown as deciles compared to all other 13-year June to May periods from June 1900 to January 2009. Figure provided by the Australian Bureau of Meteorology.

Figure 4. The key weather systems that affect the Murray-Darling Basin during the austral (a) summer (DJF) and (b) winter (JJA). The southern half of the Basin is primarily affected by extra-tropical systems and receives the majority of its annual rainfall during the cooler months (May–October). The northern half of the Basin is primarily affected by tropical systems and interactions between tropical and extra-tropical systems and receives the majority of its annual rainfall during the warmer months (November–April). The systems that tend to suppress rainfall are shown in red.

Figure 5. The squares represent the difference in the 6-monthly (April–September) mean number of heavy rain days (defined as days with over 25 mm of rainfall)

between years experiencing neutral and extreme phases of ENSO, SAM and Indian Ocean SST variations. The mean number of rain days during years in the positive (top row) and negative (bottom row) phases of ENSO (left column), SAM (centre column) and eastern Indian Ocean SSTs (right column) are shown. Blue indicates that a station has more, and red that a station has less heavy rain days than in neutral years. A solid square indicates this difference is significant at the 95% level based on a confidence interval generated from 1000 bootstrap replicates of neutral years.

Figure 6. Running 30-year correlations between pairs of seasonal indices representing the drivers of inter-annual climate variability in the MDB. SAM and STR intensity (solid black line), SAM and STR latitude (dashed black line), SAM and ENSO (solid yellow line), ENSO and IOD (solid blue line), ENSO and the eastern pole of the IOD (dashed blue line), ENSO and STR intensity (solid red line), and ENSO and STR latitude (dashed red line) are shown. A t-distribution with a sample size of 30 years was used to estimate the bounds of statistical significance and is indicated by the thin dashed lines ($t = \pm 0.35$).

Figure 7. Composite mean percentage winter (June–August) rainfall anomalies in the MDB during co-occurring phases of the SAM and ENSO from 1905–2004. A positive or negative phase of SAM and ENSO was defined when the standardized anomalies (relative to 1905–2004) of each index exceeded ±0.5 standard deviations. Red colours indicate rainfall deficits and blue, rainfall surplus. The number, *n*, in each panel indicates the number of years from which the composite was computed, this is, the number of years in which the particular phases of ENSO and SAM co-occurred. Rainfall data is from the AWAP high-quality Australian rainfall grids [*Jones et al.*, 2009a].

Table 1. The seasonality of the inter-annual and longer hydroclimate processes for the Murray-Darling Basin is identified. The black circles indicate the months in which previous studies (see Section 3) have reported an influence of the process stated in the left-half column on the Murray-Darling Basin's climate. The months shaded in orange (March, April, May, June) are those that experienced large and persistent rainfall deficits during the Big Dry (c. 1997–2010).

	J	**F**	**M**	**A**	**M**	**J**	**J**	**A**	**S**	**O**	**N**	**D**
Inter-annual processes												
El Niño – Southern Oscillation	●	●	●			●	●	●	●	●	●	●
Indian Ocean SSTs						●	●	●	●	●	●	
Southern Annular Mode	●	●				●	●	●	●	●	●	●
Sub-Tropical Ridge Intensity				●	●	●	●					
Sub-Tropical Ridge Position	●	●										●
Tropical North Australia SSTs									●	●	●	
ENSO/SAM interactions	●	●	●	●	●	●	●	●	●	●	●	●
Inter-decadal – centennial												
Inter-decadal Pacific Oscillation	●	●				●	●	●	●	●	●	●
Anthropogenic climate change	●	●	●	●	●	●	●	●	●	●	●	●

Table 2. Identifies the characteristics of the Big Dry that are consistent with the typical influences of known climate processes. These include similarities in spatial and temporal features and associated small-scale influences e.g. the influence of an inter-annual process on weather systems. The seven major defining characteristics of the drought identified in Section 5 are stated below the table. A tick indicates that a characteristic feature of the drought is consistent with current understanding of how the processes in the far-left column influence the regional climate, a cross indicates an inconsistency, and a dash indicates that the link between that characteristic and the climate process is unknown.

Characteristic drought feature	**1**	**2**	**3**	**4**	**5**	**6**	**7**
Inter-annual processes							
El Niño – Southern Oscillation	✓	×	✓	–	–	–	×
Indian Ocean SSTs	✓	×	✓	–	–	–	–
Southern Annular Mode	✓	×	✓	–	–	✓	–
Sub-Tropical Ridge Intensity	✓	✓	✓	–	–	–	✓
Sub-Tropical Ridge Position	✓	×	×	–	–	–	✓
Tropical North Australia SSTs	✓	×	✓	–	–	–	–
ENSO/SAM interactions	✓	✓	✓	✓	–	✓	–
Inter-decadal – centennial							
Inter-decadal Pacific Oscillation	✓	×	✓	✓	–	–	–
Anthropogenic climate change	✓	✓	✓	–	–	–	✓
Trends							
Trend in STR intensity	✓	✓	✓	–	–	–	✓
Southerly shift in storm track	✓	✓	✓	–	–	✓	✓
Warmer Indian Ocean SSTs	✓	×	✓	–	–	–	–

Major defining characteristics of the Big Dry.
[1] Impacts the southern Murray-Darling Basin (south of approximately 33°S)
[2] Persistent late autumn and early winter (MAMJ) rainfall deficits
[3] Occasional winter/spring rainfall deficits
[4] Lack of high rainfall periods and high one-day rainfall totals
[5] Reduction in rainfall stemming from cut-off low pressure systems
[6] Absence of rain-bearing troughs
[7] Increases in surface pressure

1433 **Figure 1**

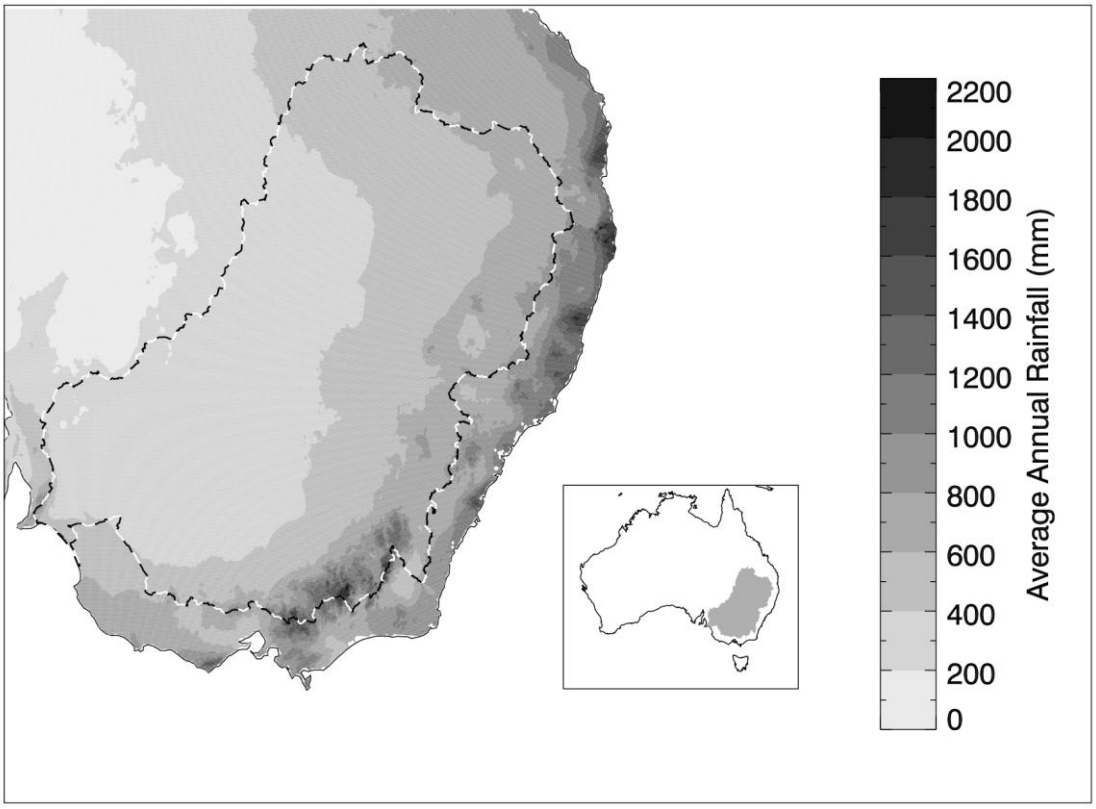

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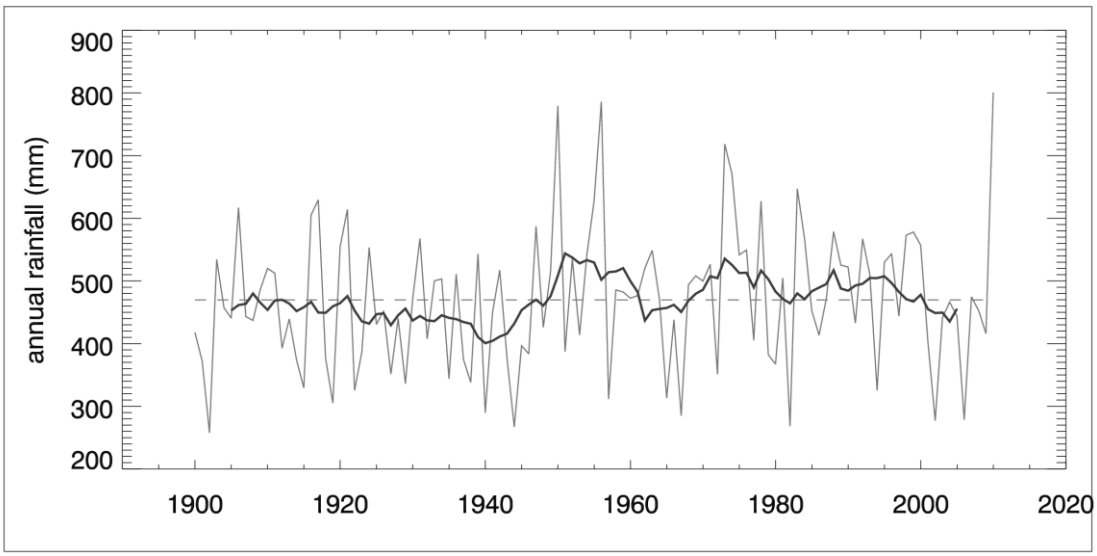
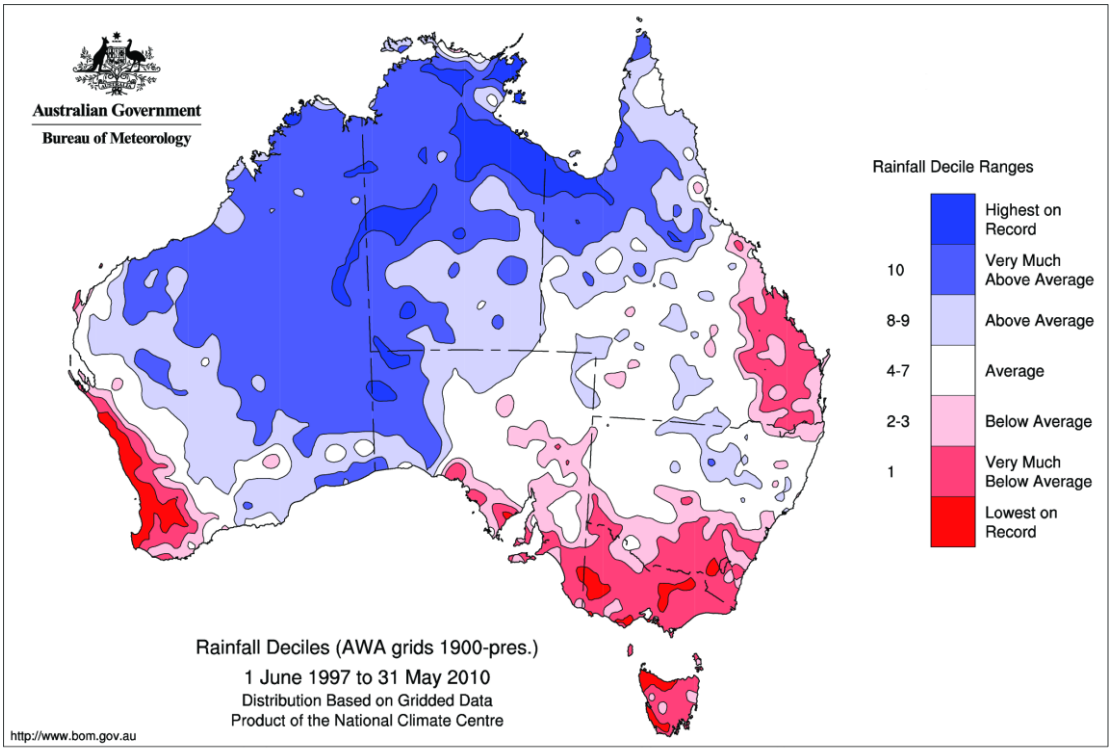
Figure 2


annual rainfall (mm)
900
800
700
600
500
400
300
200
1900
1920
1940
1960
1980
2000
2020

1439 **Figure 3**



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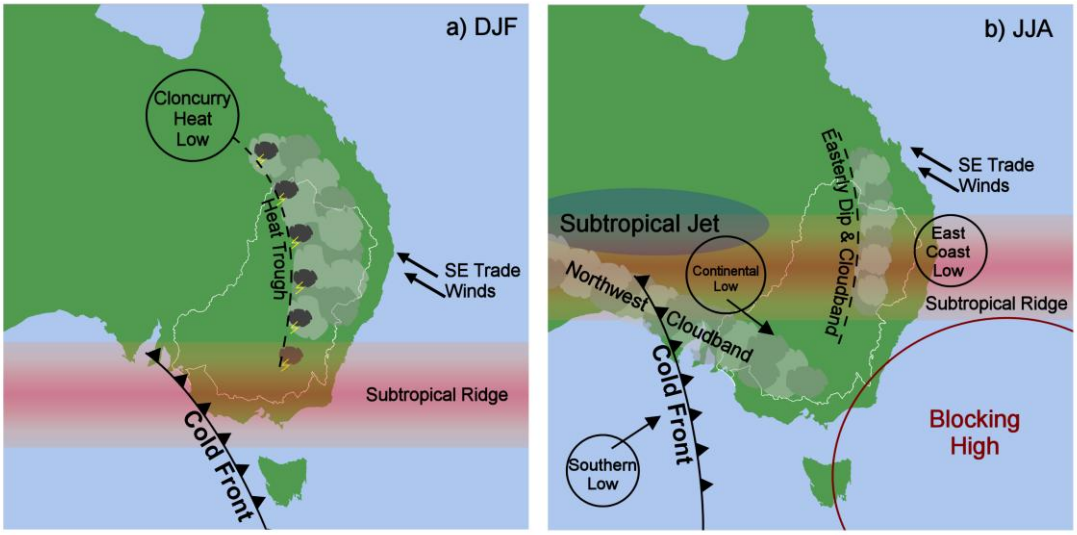
1443 **Figure 4**



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1446 **Figure 5**

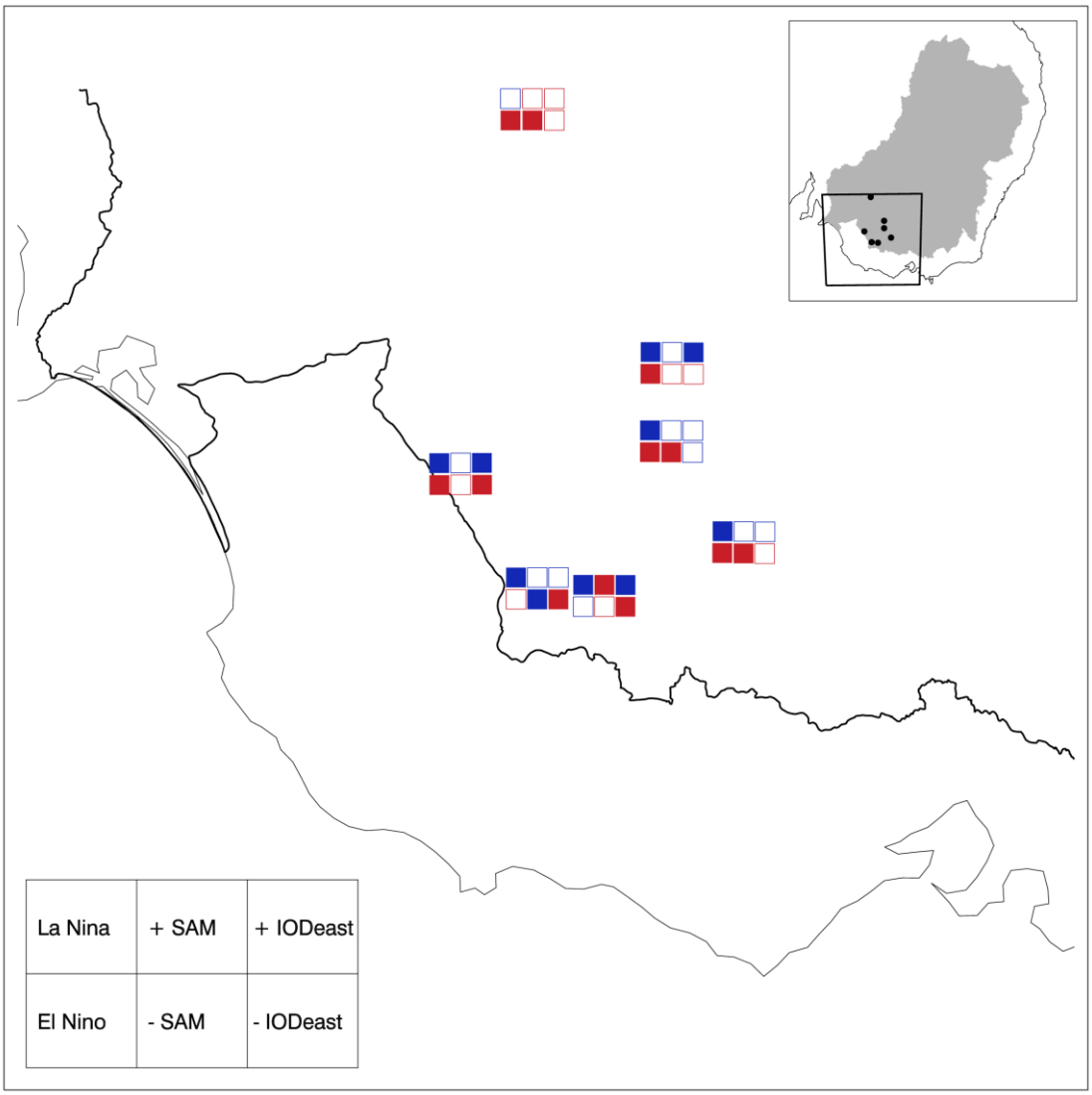

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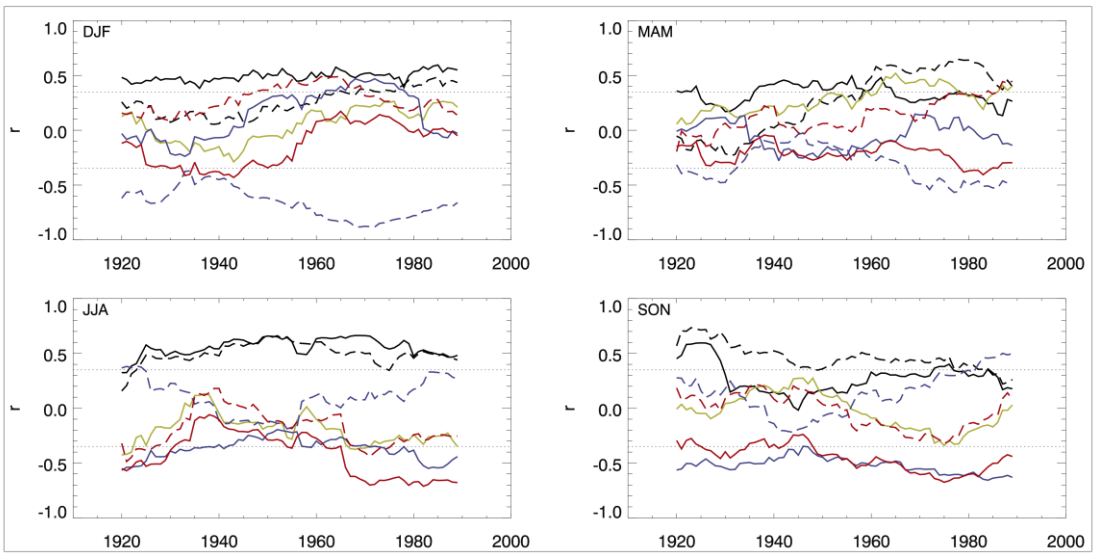
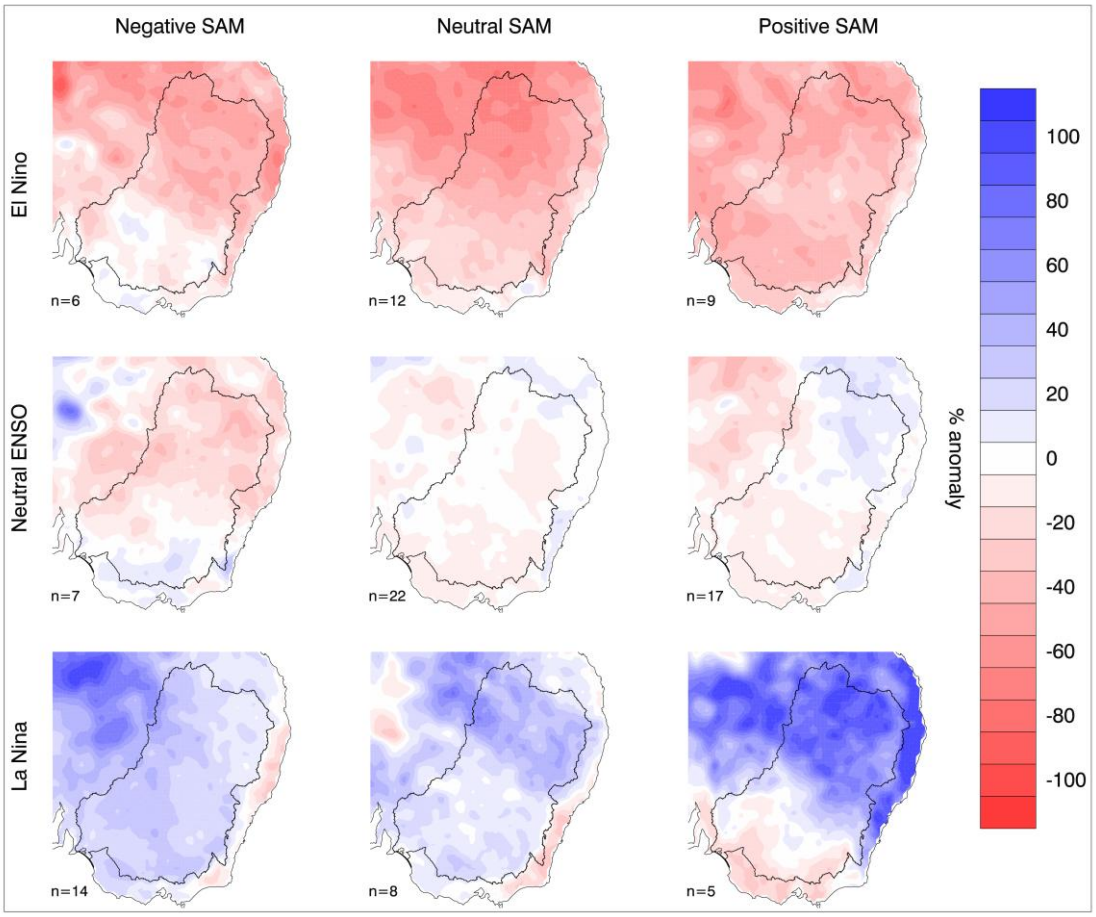
Figure 6



1452 **Figure 7**



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