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Climatic records over the past 30 ka from temperate Australia – a synthesis from the Oz-INTIMATE workgroup

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ABSTRACT

Temperate Australia sits between the heat engine of the tropics and the cold Southern Ocean, encompassing a range of rainfall regimes and falling under the influence of different climatic drivers. Despite this heterogeneity, broad-scale trends in climatic and environmental change are evident over the past 30 ka. During the early glacial period (~30–22 ka) and the Last Glacial Maximum (~22–18 ka), climate was relatively cool across the entire temperate zone and there was an expansion of grasslands and increased fluvial activity in regionally important Murray–Darling Basin. The temperate region at this time appears to be dominated by expanded sea ice in the Southern Ocean forcing a northerly shift in the position of the oceanic fronts and a concomitant influx of cold water along the southeast (including Tasmania) and southwest Australian coasts. The deglacial period (~18–12 ka) was characterised by glacial recession and eventual disappearance resulting from an increase in temperature deduced from terrestrial records, while there is some evidence for climatic reversals (e.g. the Antarctic Cold Reversal) in high resolution marine sediment cores through this period. The high spatial density of Holocene terrestrial records reveals an overall expansion of sclerophyll woodland and rainforest taxa across the temperate region after ~12 ka, presumably in response to increasing temperature, while hydrological records reveal spatially heterogeneous hydro-climatic trends. Patterns after ~6 ka suggest higher frequency climatic variability that possibly reflects the onset of large scale climate variability caused by the El Niño/Southern Oscillation.

1. Introduction

The temperate zone of Australia, as defined here, occupies ~20–45°S, excluding the arid interior. The zone extends from (modern) subtropical Queensland in the North, southward along the eastern seaboard (New South Wales (NSW) and Victoria), coastal South Australia, Tasmania and southern Western Australia (WA) (Fig. 1a). The subtropics are included in this synthesis, because during the Last Glacial they experienced a drier, cooler climate more similar to the modern temperate zone rather than the tropics (e.g. Donders et al., 2006; Petherick et al., 2008). The response of temperate zone Australia to global climatic drivers over the past 30 ka is poorly understood. To a large extent this is due to the relative lack of sites on land with continuous palaeoenvironmental records, many of which have poor chronological

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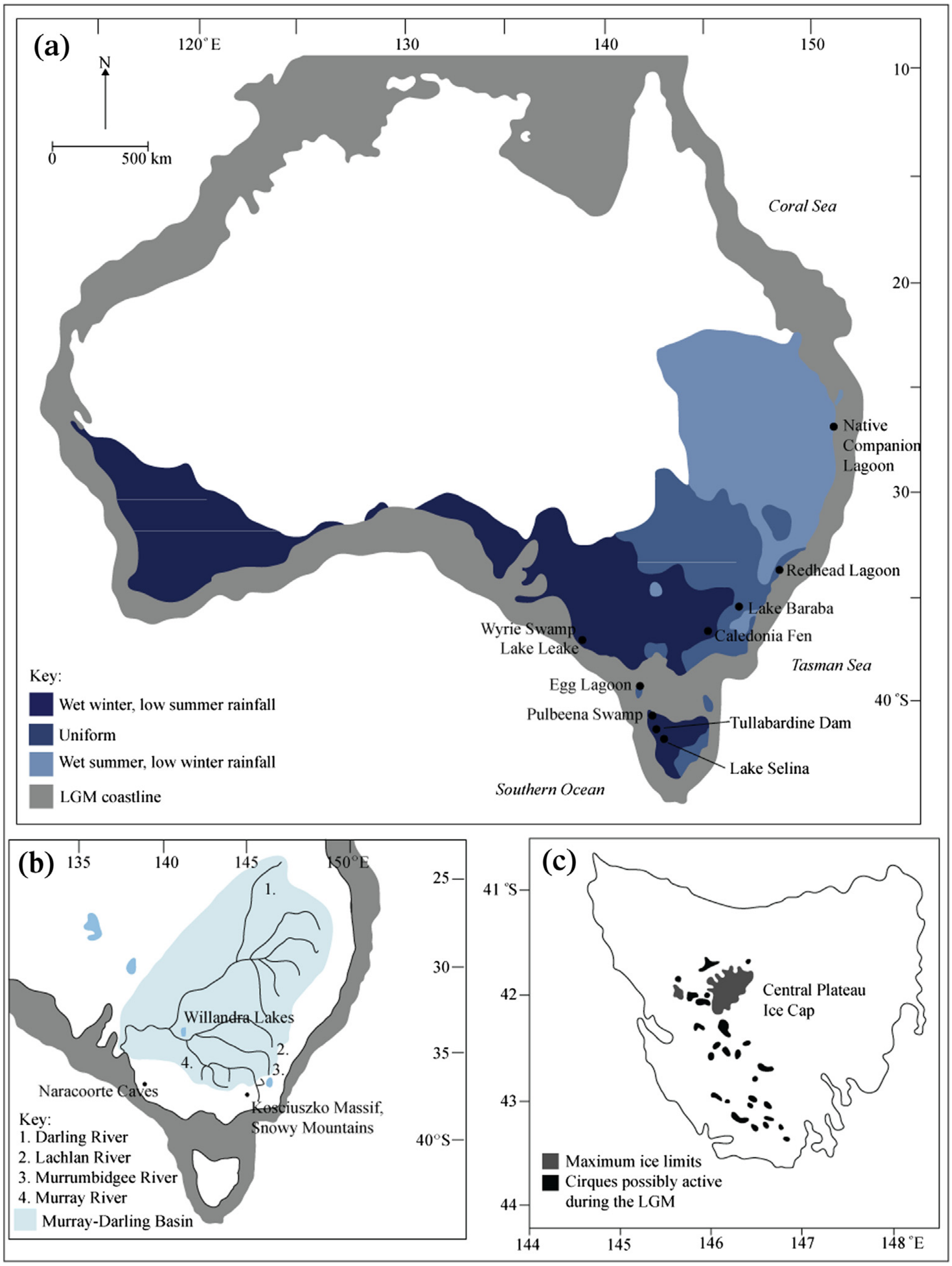

Fig. 1. Location of continuous records of climatic and environmental variability encompassing the period 30–0 ka in eastern Australia, (a) overlain on rainfall seasonality in the temperate zone (adapted from BOM, 2005) with the estimated LGM coastline (adapted from Gillespie, 2002), (b) the hydrologic and fluvial record sites in the Murray–Darling Basin along with the location of the Naracoorte Caves and the Kosciuszko Massif, and (c) close-up of Tasmania showing the extent of glaciation during the late Pleistocene (adapted from Barrows et al., 2002).

control (Turney et al., 2006). There are large spatial gaps between datasets, with a noticeable underrepresentation of long records from southwest Australia (Figs. 1 and 2).

The past 30 ka are comprised of the early Last Glacial period (~30–22 ka), the Last Glacial Maximum (LGM: ~22–18 ka), the deglacial period (~18–12 ka) and the Holocene (~12–0 ka). Early work hypothesised that the LGM was one of the most extreme climatic events affecting temperate Australia within this time period, with a 9 °C reduction in temperature of the warmest month at 35°S (Galloway, 1965), resulting in glaciation, depressed snowlines and altitudinally lower periglacial limits. The LGM of temperate Australia was postulated not only to have been cold, but also relatively dry (Galloway, 1965; Bowler, 1976). Subsequent work suggests that conditions were also cooler across temperate eastern Australia immediately prior to the LGM (e.g. Colhoun et al., 1999; Forbes et al., 2007; Kershaw et al., 2007; Petherick et al., 2008), with wetter conditions than present implied by both higher lake levels in some lacustrine systems (Wasson and Donnelly, 1991; Harrison, 1993) and increased fluvial activity (Page et al., 1996). Importantly, these studies hinted at a possible correlation between temperature and effective precipitation, a pattern which was observed across continental Australia, leading to the hypothesis that the last full glacial cycle saw the wettest interstadials of the Quaternary (Nanson et al., 1992; Kershaw and Nanson, 1993).

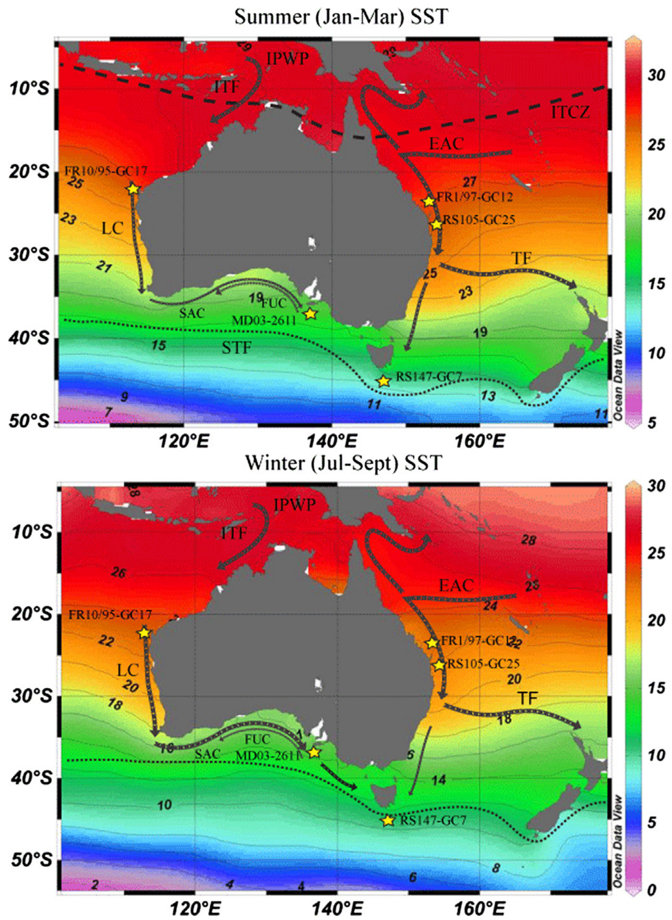

Fig. 2. Map of the modern SST for Austral summer (Jan–Mar) and Austral winter (Jul–Sept) with the position of the main surface currents and subtropical front. The location of the main cores described in the text and in this figure is shown by a yellow star. EAC – East Australian Current, LC – Leeuwin Current, FUC – Flinders Undercurrent, TF – Tasman Front, STF – Subtropical Front, ITCZ – Inter-tropical Convergence Zone.

However, more recent studies suggest a more complex picture of spatio-temporal variability in temperature and precipitation patterns, as evidenced by the large variation seen in more recent compilations of palaeoenvironmental data from the Australian region (Turney et al., 2006; Williams et al., 2009).

An increase in sampling and dating intensity over the last few decades has increased our ability to constrain periods of significant climate change over the last 30 ka. Recent improvements in the precision, and the increased application, of geochronological techniques such as optically stimulated luminescence (OSL), cosmogenic and U/Th dating have also enabled more reliable age control, allowing for the construction of chronologic frameworks for landscape response to palaeoenvironmental change. This paper is part of a series published in this special issue, under the auspices of the INTIMATE (INTegration of Ice, MArine and TErrestrial records) initiative. The INTIMATE project is a core programme of the INQUA Palaeoclimate Commission (Walker et al., 2001). Initially based in the North Atlantic, the aim of INTIMATE was to produce a synthesis of ice-core, marine and terrestrial proxy data from the region for the period 30–8 ka (Walker et al., 2001). More recently other regions, including Australasia (OZ-INTIMATE), have been funded to develop similar syntheses across the globe (refer to Reeves et al., 2013).

It is our aim here to synthesise palaeoenvironmental data from temperate Australia over the last 30 ka in an attempt to highlight (1) regional trends and (2) existing gaps in space and time that require urgent attention. We review records from both marine and terrestrial environments with the aim of identifying the timing, nature and duration of palaeoenvironmental change and elucidating the potential climatic drivers of these changes. By compiling multiple independent proxies (e.g. marine, fluvial, lacustrine, speleothem, glacial) from across temperate Australia, we hope to develop a more robust picture of environmental change that, importantly, will enable the identification of critical areas with little to no palaeoenvironmental data and enables us to generate a broad framework of environmental and climatic change through the last 30 ka that can guide future palaeoclimatic research for the region.

Several important reviews of past climatic variability in temperate Australia have previously been published. Two of the earliest reviews were published by Bowler et al. (1976) and Chappell and Grindrod (1983) (part of the Climate Mapping of Australia and New Zealand (CLIMANZ) project). More recently, in contribution to an earlier iteration of OZ-INTIMATE, Turney et al. (2006) produced a review of climatic and environmental variability for Australia based on several key records for the period 30–8 ka. Williams et al. (2009) also produced a review of climatic patterns in greater Australia for the period 35–10 ka. The synthesis we present here builds on the earlier work of Bowler et al. and the CLIMANZ project and it differs in a two important ways from the reviews of both Turney et al. (2006) and Williams et al. (2009). First, we encompass a broader temporal range (i.e. the inclusion of the Holocene); and second, we focus specifically on the temperate region of Australia, as opposed to the broader spatial treatments of these earlier reviews. Further, by interrogating trends within the temperate zone alone, we hope to identify, characterise and contextualise environmental changes within this zone over the last 30 ka, enabling the formation of a coherent picture of regional trends that can be used to generate testable hypotheses for future investigations.

Here we divide the past 30 ka into the following time periods: early Last Glacial period (~30–22 ka), LGM (22–18 ka), deglacial period (18–12 ka), early Holocene (12–6 ka) and late Holocene (6–0 ka). All dates are expressed as years B.P. unless stated otherwise.

2. Regional setting

The temperate zone is broadly characterised by cool wet winters and warm dry summers, and can be further classified based on of the distribution of rainfall throughout the year (Fig. 1a) (Australian Bureau of Meteorology (BOM), 2005):

1. Wet winter, low summer rainfall: southwest WA, southern SA, southern NSW, most of Victoria, most of Tasmania.
2. Uniform rainfall: central and southeast NSW, southeast Victoria, southeast Tasmania.
3. Wet summer, low winter rainfall: central and southeast Queensland, northeast NSW.

The southern Australian continent (south of ~35°S) is generally characterised by a winter dominant rainfall regime (BOM, 2005) resulting from the incursion of the moisture-laden southern westerly winds (SWW) inland to southern Australia, as the subtropical high pressure system that dominates continental Australia migrates north in response to the Northern Hemisphere insolation maxima (Fig. 2). In contrast, the northern extent of the temperate zone experiences a summer dominant rainfall regime resulting from the poleward migration of the Inter-tropical Convergence Zone (ITCZ) in response to the Southern Hemisphere solar maxima and the deliverance of rain by the southeast trade winds.

The tension zone between northern and southern temperate zones is marked by an interplay between areas dominated by summer easterly derived moisture and winter westerly derived moisture (Fig. 1a) (BOM, 2005). The eastern coast of temperate Australia (viz. Southern Queensland, NSW and Victoria) comes under the influence of the rain-bearing East Coast Low pressure systems during the winter (Fig. 2) (Hopkins and Holland, 1997; BOM, 2007). The frequency of East Coast Lows is associated with the position of the subtropical high, and is strongly correlated with the Southern Oscillation Index (Hopkins and Holland, 1997).

2.1. Modern oceanography

The offshore East Australian Current (EAC) is a warm, south flowing boundary current along the east coast of Australia (Fig. 3) (Boland and Church, 1981). The EAC intensifies, accelerates and deepens as it follows the coastline until ~32°S, where the main flow of the current (primarily the upper layer) separates and flows east across the Tasman Sea to form the Tasman Front (Andrews et al., 1980). A more minor flow, including the deeper layers of the EAC, continues south of ~32°S along the coast often as a series of deep eddies (Fig. 3) (Wyrtki, 1962; Ridgway and Godfrey, 1994) as far south as the subtropical front (STF) which sits over the South Tasman Rise oscillating between 45°S and 47°S in winter and summer, respectively (Orsi et al., 1995; Belkin and Gordon, 1996; Rintoul et al., 1997).

In contrast, on the west coast of Australia the Leeuwin Current (LC) is an anomalous south flowing boundary current (most eastern boundary currents in the Southern Hemisphere flow north) (Fig. 3) that transports warm, low salinity water from the Indo-Pacific Warm Pool and the Central Indian Ocean Gyre and is sourced from the Indonesian Throughflow (ITF) (Friedman, 1979; Nof et al., 2002; Sprintall et al., 2002). The LC is seasonal, with stronger flows during the Austral winter. The LC flows along the coast of Western Australia becoming well defined from Northwest Cape, northwestern Australia to Cape Leeuwin, the southwestern tip of the continent; a distance of ~5500 km. The LC is restricted from continuing south from the southwest tip of Australia by the presence of the STF between 38 and 40°S (Orsi et al., 1995). The warm water of the LC can extend around the south coast of Australia during the winter contributing to the South Australian Current (SAC), and can flow as far as the west coast of Tasmania, driven by the SWW. During the summer, the waters of the LC are sourced from the Southern Ocean, resulting in colder SSTs than in winter (Ridgway and Condie, 2004; Middleton and Bye, 2007). The subsurface Flinders Undercurrent (FUC) originates in the Southern Ocean and flows east–west along southern Australia (Moros et al., 2009).

2.2. Modern drivers of climatic variability in terrestrial temperate Australia

2.2.1. El Niño/Southern Oscillation (ENSO)

The two phases of the ENSO, El Niño and La Niña, have markedly different impacts on both spatial and temporal variability in rainfall patterns in northern and eastern Australia (Risbey et al., 2003). During El Niño phases the central and eastern Pacific Ocean becomes anomalously warm, with above average pressure in the Indonesian–North Australia region, while the central Pacific experiences below average pressure. During El Niño events, northern and eastern Australia (extending as far south as Tasmania) experience lower than average rainfall (Manins et al., 2001; Hill et al., 2009) and can result in major drought episodes (Risbey et al., 2003). Conversely, the opposing La Niña mode results in anomalously high rainfall across much of temperate Australia (Kotwicki and Allan, 1998).

The ENSO operates on a range of timescales, from inter-annual (e.g. 2–7 years) variability, to millennial (e.g. 2000 years) and semi-precessional timescales (e.g. 11.9 ka) (Heusser and Sirocko, 1997; Clement et al., 2001; Tudhope et al., 2001; Moy et al.,

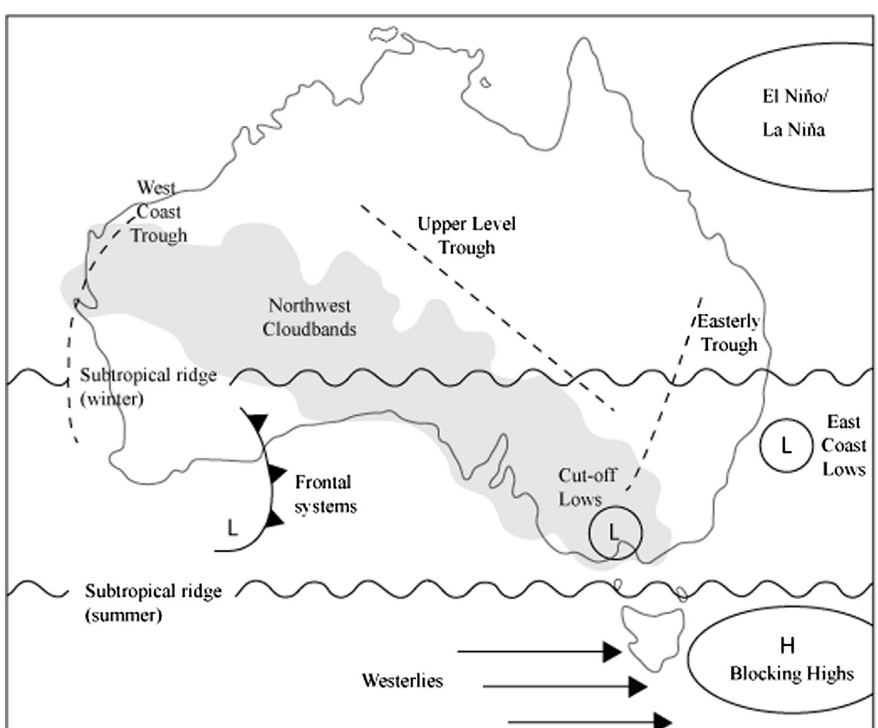

Fig. 3. Modern drivers of climate variability in temperate Australia (adapted from BOM, 2005; Risbey et al., 2009).

2002; Crimp and Day, 2003; Turney et al., 2004). Although relatively little is known about the variability of the ENSO on these longer timescales, time series analysis of *Porites* corals from Papua New Guinea indicate no significant changes in the cyclicity of the ENSO over the past 130 ka (Moy et al., 2002). However, the *Porites* records do suggest major changes in the amplitude of the ENSO signal through time (Moy et al., 2002) e.g. a mid-Holocene (ca 7–4 ka) intensification of ENSO to modern conditions is evident from a number of records (e.g. McGlone et al., 1992; Shulmeister and Lees, 1995; Moy et al., 2002; Gagan et al., 2004).

2.2.2. Pacific Decadal Oscillation (PDO)

The Pacific Decadal Oscillation (PDO) is the decadal-scale oscillations of sea surface temperatures in the North Pacific that has a significant influence on Australian rainfall (Mantua and Hare, 2002; Stone et al., 2003). PDO phases are classified as either "warm" (associated with below average rainfall in Australia) or "cool" (associated with above average rainfall in Australia) (Crimp and Day, 2003; Stone et al., 2003). The influence of the PDO on Australian rainfall can be modulated by interactions with ENSO, with, for example, 25–44% of modern summer rainfall variability in Queensland attributed to a combination of these two modes (Crimp and Day, 2003; Stone et al., 2003). This influence extends to the River Murray to the south, where rainfall, river flow and surface temperature have been shown to correlate with interactions between the PDO and the ENSO (Power et al., 1999).

2.2.3. Southern westerly winds (SWW) and the Southern Annular Mode (SAM)

The SWW are a key feature of the meteorology of the Southern Hemisphere, governing precipitation in the mid to high latitudes (Garreaud et al., 2009) and being implicated in global CO_2 variations (Toggweiler et al., 2006; Rojas et al., 2009; Moreno et al., 2010). In addition, the SWW drive oceanic circulation patterns such as the upwelling of deepwater in Southern Ocean (Hurrell et al., 1998; Varma et al., 2010) and the leakage of the Agulhas Current, southern Africa into the South Atlantic (Biastoch et al., 2009).

Depressions and rain-bearing troughs associated with the SWW play an important role in the climate of Australia (Pitman et al., 2004) and cold fronts associated with the SWW can impart an influence over landscape dynamics via, for example, wind erosion and transport of surface sediments (Hesse and McTainsh, 1999). The leading inter-annual mode of the SWW, the Southern Annular Mode (SAM)/Antarctic Oscillation (AAO), results in either a northward (negative SAM/AAO) or southward (positive SAM/AAO) shifts in the mean position of the SWW, and, as such, is significantly correlated with precipitation in southern temperate Australia (Hendon et al., 2007). Moreover, like the ENSO/PDO relationship elucidated above (Section 2.2.2), SAM/AAO is modulated by interactions with ENSO, with complimentary trends in these modes amplifying the climate effects of each mode (Fogt and Bromwich, 2006).

3. Proxy records from temperate Australia

3.1. Palaeoecological change based on lake, peat and wetland records

3.1.1. Vegetation reconstructions

Temperate Australia has yielded a number of continuous pollen records that span the last 30 ka (Fig. 1a, Table 1). Age control for these lake, peat and wetland records was determined using radiocarbon dates (predominantly on bulk organics) and calibrated here using the IntCal09 curve (Reimer et al., 2009). Whilst it is possible to infer broad scale climate trends from the available pollen data, our ability to infer finer scale climate trends from such data for this time period is complicated by a number of poorly understood factors that influence fossil pollen composition (Turney et al., 2006; Williams et al., 2009). In addition, it is difficult to assess the extent of the spatial range (i.e. local, regional) that the pollen is representing. Pollen dispersal rates, production rates, taphonomic processes and perhaps most importantly, the large number of species that contribute to dominant pollen taxa means that climatic inferences from this data remain largely qualitative.

Recent advances in the development of transfer functions between climate variables and the modern range of pollen taxa have produced seemingly more realistic quantified pollen-based estimates of temperature and precipitation (Cook and Van der Kaars, 2006; Fletcher and Thomas, 2010a). For example, Fletcher and Thomas (2010a) produced quantitative estimates of temperature over the past 135 ka based on pollen archives from Tullabardine Dam and Lake Selina in maritime western Tasmania (Fig. 1a) that correlate well with proximal marine SST record reconstructed by Barrows et al. (2007) that suggests some degree of reliability.

3.1.2. Fire regimes

The quantification of charcoal in sedimentary sequences provides important palaeoenvironmental information about past fire activity (Power et al., 2008; Conedera et al., 2009). Fire is a globally important environmental factor that plays a key role across a range of landscapes (e.g. Power et al., 2008; Bowman et al., 2009) and sedimentary charcoal records can provide valuable insight into factors such as climatic variability, biomass burning, fuel availability, anthropogenic land use and/or complex interactions between all of these variables (Fletcher and Thomas, 2010b). The analysis of microscopic charcoal fragments (typically <100 μm) is frequently used in Australia, as tallying microscopic charcoal fragments can be done in tandem with pollen counting (e.g. Newsome and Pickett, 1993; Dodson and Lu, 2000; Black et al., 2006; Williams et al., 2006; Kershaw et al., 2007).

3.1.3. Other biological indicators

Biological indicators such as diatoms, ostracods, chironomids, beetles and insects in lake sediments have all been used to provide insight into past climatic variability at a number of sites in temperate Australia (Fig. 1a, Table 1) (Edwards et al., 2006; Rees et al., 2008; Gouramanis et al., 2010, 2012; Porch, 2010; Rees and Cwynar, 2010; Haynes et al., 2011; Kemp et al., 2012). For example, ostracods have been used extensively as proxies for palaeosalinity in both freshwater and marine environments (De Deckker, 1982; Chivas et al., 1985, 1986a; Yassini and Kendrick, 1988). The longest ostracod records are from the Victorian sites Tower Hill (D'Costa et al., 1989) and Lake Wangoom (Edney et al., 1990). These studies used ostracods in conjunction with pollen and charcoal (and at Tower Hill, diatoms) to develop records of palaeoenvironmental variability extending to ~21 ka and ~51 ka respectively.

Diatoms were initially identified as proxies for past and modern lake salinity in the 1990s (Reid et al., 1995), and have since been used in Holocene environmental reconstructions in western Victoria (Gell, 1997) and the Coorong region in South Australia (Edwards et al., 2006). Despite a growing number of datasets for use as modern analogues (Rees et al., 2008; Haynes et al., 2011; Saunders, 2011), quantified reconstructions of past conditions based on bio-indicators are relatively few, and, with the exception of Tower Hill and Lake Wangoom, generally have not been extended beyond the Holocene (Tibby et al., 2007).

The chief limitation of using transfer functions between modern datasets and archived biological indicators are the complex, site-specific relationships between species abundance and the

Table 1
Key records of climatic and environmental variability encompassing the past 30 ka from temperate Australia, listed by latitude (north to south). Ages are ka BP.

Site	Location	Length of record	Setting	Proxies	Reference
FR1/97-GC12	23°34′S, 153°13′E	30 ka	Marine	$\delta^{18}O$ *G. Ruber*	Bostock et al. (2006)
RS105-GC25	26°35′S, 153°51′E	22 ka	Marine	$\delta^{18}O$ *G. Ruber*	Troedson and Davies (2001) and Bostock et al. (2006)
Native Companion Lagoon	27°45′S, 153°25′E	40 ka	Lacustrine	Dust, pollen, charcoal, geochemistry, particle size analysis	McGowan et al. (2008) and Petherick et al. (2008, 2009)
Ulungra Springs	31°43′S, 149°6′E	30 ka – late Holocene	Lacustrine	Pollen, charcoal	Dodson and Wright (1989)
Riverine Plain, MDB	NSW and southern QLD	>30 ka	Fluvial	Palaeohydrology (bedload, source-bordering dunes)	Page et al. (1996, 2001) and Kemp and Rhodes (2010)
Redhead Lagoon	32°59′S, 151°43′E	ca 75 ka	Lacustrine	Pollen, charcoal	Williams et al. (2006)
Willandra Lakes	33°40′S, 143°00′E	>30 ka	Lacustrine	Palaeohydrology (lunette stratigraphy)	Bowler (1998) and Bowler et al. (2003, 2012)
Lake Baraba	34°13′S, 150°13′E	>43 ka	Lacustrine	Pollen, charcoal, LOI	Black et al. (2006)
Lake Tyrrell	35°20′S, 142°50′E	>30 ka	Lacustrine	Palaeohydrology, microbial lipids	Stone (2006) and Bray et al. (2012)
Snowy Mountains	36°20′S, 148°15′E	n/a	Glacial moraines	Cosmogenic dating (^{10}Be)	Barrows et al. (2001)
Bega Swamp	36°31′S, 149°31′E	Holocene	Lacustrine	Charcoal, pollen	Green et al. (1988) and Donders et al. (2007)
MD03-2611	36°44′S, 136°33′E	33 ka	Marine	$\delta^{18}O$ *G. bulloides*, Alkenone SST.	Calvo et al. (2007), Gingele et al. (2007) and Moros et al. (2009)
Caledonia Fen	37°20′S, 146°44′E: 1280 m a.s.l.	140 ka	Lacustrine	Pollen, charcoal, magnetic susceptibility	Kershaw et al. (2007)
Naracoorte caves	37°20′S, 140°48′E	500 ka	Speleothem compilation	$^{230}Th/^{234}U$	Ayliffe et al. (1998), Brown and Wells (2000), Desmarchelier et al. (2000), Moriarty et al. (2000), Macken et al. (2011), and St Pierre et al. (2012)
Lake George	37°25′S, 139°59′E	>30 ka	Lacustrine	Lake shorelines (palaeohydrology), pollen, ostracods, charcoal	Galloway (1965), Coventry (1976), Singh et al. (1981), De Deckker (1982), Singh and Geissler (1985), Lees and Cook (1991), and Fitzsimmons and Barrows (2010)
Lake Leake	37°37′S, 140°35′E	50 ka	Lacustrine	Pollen	Dodson (1974, 1975)
Wyrie Swamp	37°37′S, 140°21′E	50 ka	Lacustrine	Pollen	Dodson (1977)
Lake Keilambete	38°12′S, 142°53′E	ca 30 ka	Lacustrine	Hydrologic modelling, ostracods,	Bowler and Hamada (1971), Chivas et al. (1985), Mooney (1997), Jones et al. (1998), and Wilkins et al. (2012)
Egg Lagoon	39°38′S, 143°59′E: 17 m a.s.l.	ca 50 ka	Lacustrine	Molluscs, pollen, charcoal	D'Costa et al. (1993)
Pulbeena Swamp	40°51′S, 145°7′E	ca 50 ka	Lacustrine	Pollen	Colhoun et al. (1982)
Lynds Cave	41°34′S, 146°14′E	9.2–5.1 ka	Speleothem	$\delta^{18}O$, $\delta^{13}C$, $^{234}U/^{238}U$, physical property	Xia et al. (2001)
Tullabardine Dam	41°41′S, 145°39′E	>43.8 ka	Lacustrine	Pollen, charcoal	Colhoun and van der Geer (1986)
Lake Selina	41°53′S, 145°36′E	ca 70 ka	Lacustrine	Pollen, charcoal	Colhoun et al. (1999)
Central Ice Plateau	42°40′S, 146°10′E	n/a	Glacial moraines	Cosmogenic dating (^{10}Be and ^{36}Cl)	Barrows et al. (2002)
RS147-GC7	45°09′S, 146°17′E	100 ka	Marine	$\delta^{18}O$ *G. bulloides* Alkenone SST.	Sikes et al. (2009)

environmental variable being estimated (e.g. salinity) (Tibby et al., 2007). In addition, uncertainty in interpretation can arise due to the sensitivity of the bio-indicators to other environmental conditions (e.g. turbidity) (Tibby et al., 2007; Huntley, 2012). Insect assemblages may also respond to a range of environmental conditions including the presence of prey, local vegetation and soil conditions (Porch and Elias, 2000). Despite these issues, recent studies indicate that site-specific reconstructions of past conditions based on bio-indicator transfer functions can produce robust results (Tibby et al., 2007; Kemp et al., 2012).

3.2. Palaeohydrology recorded in rivers and lakes

Geomorphic records from river and lake systems provide long but often discontinuous records of palaeohydrological changes (Fig. 1b). Table 1 lists key lacustrine records from the temperate zone, both from the coastal catchments and the southern MDB, and including the endorheic Lake George catchment which straddles the Great Divide between the inland and fluvial catchments. There are as yet no robust hydrologic records for the temperate zone of Western Australia.

The catchment size of temperate Australian fluvial systems ranges from relatively short coastal systems (<20,000 km^2), to the sub-continental Murray–Darling Basin (MDB: 1.01×10^{-6} km^2). The MDB occupies approximately one sixth of the continent and lies within both summer- and winter-dominant rainfall regions, and consequently preserves evidence of past influence of both precipitation regimes on geomorphic archives within the catchment. The headwaters of the MDB primarily lie within the temperate and subtropical Great Dividing Range of eastern Australia, and its tributaries flow through the temperate zone and semi-arid interior, before discharging into the Southern Ocean (Fig. 1b). The MDB records (predominantly from the southern basin) reflect the alternation in dominance of fluvial and aeolian processes that have been interpreted in terms of hydro-climatic regimes. Records from eastern coastal catchments comprise discontinuous floodplain pockets and upland swamps, and provide data on both intrinsic and extrinsic drivers of change (Nanson et al., 2003; Cohen and Nanson, 2007; Cheetham et al., 2010). Consequently the rivers of the MDB represent a substantial contribution of water inflow to the semi-arid interior of the continent, and have been the focus of most of the research to date.

Evidence of past fluvial activity has been reconstructed from bedload and palaeochannel deposits, as well as from transverse source-bordering dunes which reflect high sediment availability from increased sandy bedload transport within the MDB rivers (Bowler and Wasson, 1983; Nott and Price, 1991; Page et al., 1996, 2001). The pivotal paper of Bowler and Wasson (1983) provided for the first time an overview of environmental variability in the MDB over the past 30 ka by comparing glacial, aeolian, fluvial and lacustrine environments. The chronology for such records is predominantly based on chronologies derived from OSL, thermoluminescence (TL) and radiocarbon (^{14}C) dating of sediment deposits (Bowler and Wasson, 1983; Page et al., 1996, 2001; Cohen and Nanson, 2007; Cheetham et al., 2010). In addition, prior channel dimensions have been used to reconstruct past discharge characteristics and infer changes in hydrological balance through time in these systems. One potential complication in the interpretation of fluvial archives as palaeohydrological proxies is the potential for a considerable lag in response time of sediment discharge associated with soil and vegetation development and/or destabilisation in the headwater source areas (Vandenberghe, 1995).

The lake systems of temperate Australia comprise both depositional sinks and deflationary regimes. Depositional sinks provide higher resolution records of salinity and lake hydrology, and often occur within smaller catchments with shorter (Holocene) histories. Examples include Lake Keilambete and the Wimmera lakes in southwestern Victoria (Chivas et al., 1985, 1986a, 1986b; Jones et al., 1998; Kemp and Rhodes, 2010; Bowler et al., 2012; Kemp et al., 2012; Wilkins et al., 2012) and those presented as part of the CLIMANZ compilation (Chappell and Grindrod, 1983). Deflationary environments are more common in the case of inland lakes, particularly those which receive intermittent or no fluvial inflow, such as the highland-fed Willandra lakes within the semi-arid zone (Fig. 1b).

Strandlines surrounding lakes, and transverse aeolian lunettes which form downwind from them, provide crucial sources of palaeohydrological information, particularly in the case of deflationary regimes where palaeoecological data are not preserved within the lake basins. Prime examples of shoreline records include Lake George, which straddles the Great Dividing Range and lies adjacent the headwaters of the Lachlan and Murrumbidgee Rivers within the MDB tributaries (Fig. 1b) (Coventry, 1976; Fitzsimmons and Barrows, 2010), and the Willandra Lakes system within the MDB, fed by a distributary of the Lachlan River and lying within the semi-arid zone (Bowler, 1998; Bowler et al., 2003). Both of these systems provide detailed geomorphic evidence of past lake level transgression and regression, and are closely linked to changes in major river systems.

3.3. Marine records

A number of marine records have been published from temperate Australia (Fig. 2 and Table 1). These records predominantly use nanno and microfossil assemblages and the modern analogue technique to reconstruct mean annual sea surface temperatures (SST) from a variety of biological proxies, including: planktonic foraminifera (Barrows and Juggins, 2005), coccolithophores (Wells et al., 1994) and diatoms (Nees et al., 1999) (Fig. 2). The use of statistical techniques like the modern analogue to estimate SST is, like most methods that employ modern assemblages to infer past conditions, subject to a number of caveats, such as preferential diagenesis or dissolution of fossils, winnowing or focussing of sediment by currents, the species have not altered their habitat, and that SST has been the principal driver in system dynamics through the reconstruction period. Alkenones (U^{k}_{37}) have also been analysed on several cores as an independent method measuring the geochemical remains of coccolithophores to reconstruct SST (Pelejero et al., 2006; Calvo et al., 2007; Sikes et al., 2009). There are two different alkenone-based SST calibrations (Prahl et al., 1988; Sikes and Volkman, 1993). The Prahl et al. (1988) calibration is based on North Pacific samples, and provides the best correlations for warm ($\geq$25 °C) and cool SSTs ($\leq$12 °C). The Sikes and Volkman (1993) calibration is based on samples from the Southern Ocean, south of Australia, and provides good correlation for summer SSTs (5–25 °C). Given the latter is based on our region we feel it is more suitable for the records presented here, and as such, we use the Sikes and Volkman (1993) calibration.

Marine sediment cores provide relatively continuous records, with an occasional hiatus or disruptions from turbidites. Unfortunately, the majority of cores around temperate Australia provide low resolution climate records, as they have relatively low pelagic and hemipelagic sedimentation rates (<2 cm/ka). This is due to predominance of oligotrophic subtropical surface waters and the limited terrigenous flux received from the largely arid and flat continental landmass of Australia. However, there are a several higher resolution (>5 cm/ka) cores from the Murray Canyons (fed by the River Murray), southern Australia (Fig. 2) (e.g. MD03-2607, MD03-2611) (Calvo et al., 2007; Gingele et al., 2007; Moros et al., 2009; De Deckker et al., 2012; Lopes dos Santos et al., 2012).

For this review, we have primarily focussed on 5 continuous, high resolution (>5 cm/ka) marine records, but with reference to many more, to determine the major palaeoceanographic changes over the last 30 ka. Some of the marine records also provide information on what happened on land, including terrestrial pollen records (Harle, 1997), river sediment supply (a proxy for fluvial discharge) and aeolian activity (Hesse, 1994; Gingele et al., 2004, 2007). The chronology of the selected cores (Fig. 2 and Table 1) has been determined from multiple calibrated radiocarbon dates measured on planktic foraminifera. The radiocarbon dates were corrected for local reservoir ages ranging from 300 to 500 years (from nearby sites in the Marine Reservoir Correction Database; see individual references for details) and calibrated here to calendar years using Marine09 calibration curve (Reimer et al., 2009). This assumes that there is a constant reservoir age throughout the last 30 ka. We have labelled the LGM as the period from 22 to 18 ka, a period that incorporates the maximum stable oxygen isotope ($\delta^{18}O$) and minimum SST in the cores.

3.4. Glacial deposits

Based on interpretation of geomorphology, stratigraphic sequences and exposure dating of glacial deposits, glaciation in Australia during the early glacial period and LGM was limited to Tasmania and the Snowy Mountains (Fig. 1c, Table 1) (Colhoun, 1985a; Barrows et al., 2001, 2002).

The timing of maximum glacial extent in Tasmania was determined using ^{14}C dating on the outwash of the former Dante glacier, King Valley, western Tasmania (Fitzsimons and Colhoun, 1991; Colhoun and Fitzsimons, 1996; Colhoun et al., 1996). More recently, exposure dating of glacial deposits has provided direct ages for moraines. Exposure dating is especially useful for directly dating inorganic glacial and periglacial material (Barrows et al., 2001, 2002, 2004). Barrows et al. (2002) dated 18 late Pleistocene moraines in 8 glaciated areas in the Snowy Mountains and Tasmania using cosmogenic ^{36}Cl and ^{10}Be with further local studies by Mackintosh et al. (2006) and Kiernan et al. (2010).

3.5. Speleothems

While the value of speleothems as high-resolution archives of palaeoenvironmental information is well-established, speleothem records from temperate Australia are relatively sparse (Fig. 1b, Table 1) (Moriarty et al., 2000; Fairchild and Treble, 2009). Two key sites in temperate Australia returned a Holocene record from Lynds Cave in cool temperate Tasmania (Xia et al., 2001) and the much studied mid- to late-Pleistocene records from the Naracoorte Caves in South Australia (Fig. 1b) (Ayliffe et al., 1998; Brown and Wells, 2000; Desmarchelier et al., 2000; Moriarty et al., 2000; Forbes et al., 2007; Macken et al., 2011; St Pierre et al., 2012). The composite Naracoorte Caves record provides a composite register of changes in speleothem deposition rate that have been used to infer changes in effective precipitation (Ayliffe et al., 1998).

4. Synthesis: the climate and environment of temperate Australia ~30–0 ka

During the Last Glacial, eustatic sea level ranged between 60 and 120 m lower than modern, with lowest sea levels occurring during the LGM (Clark and Mix, 2002; Yokoyama et al., 2006). As such, the terrestrial sites discussed in this synthesis lay further from the coast than in modern times, albeit this distance varied greatly site to site. Consequently, each site may have been influenced by considerably different drivers of environmental variability during the early Last Glacial period and LGM than during the Holocene (Fig. 1a). Furthermore, under lowered sea levels, Tasmania was joined to mainland Australia by the Bassian Plain, which underlies the present day Bass Strait (Fig. 1a and b) (Lambeck and Chappell, 2001). The first oceanic encroachment on to this plain occurred after ~17.5 ka (from the west), with complete isolation of Tasmania by ~14 ka (Lambeck and Chappell, 2001).

4.1. Early glacial period ~30–22 ka

The early glacial period overlaps with the end of Marine Isotope Stage 3 (MIS3: ~60–24 ka) and the start of MIS2 (~24–11 ka). Planktonic foraminifera during this period indicates a slow decrease, or consistently cold SST off the coast of eastern Australia and a simultaneous increase in $\delta^{18}O$ suggests that global ice volume was increasing at the poles (Fig. 4) (Troedson and Davies, 2001; Barrows and Juggins, 2005; Bostock et al., 2006; De Deckker et al., 2012). The LC and EAC were both relatively weak, the STF was significantly closer to southern Australia, and there was considerably more influence over the Australian continent by Subantarctic waters (Bostock et al., 2006; Calvo et al., 2007; Sikes et al., 2009; De Deckker et al., 2012). A combination of increased nutrients from Subantarctic waters and the deposition of high amounts of dust and terrigenous flux into the oceans at this time increased offshore productivity (Hesse, 1994; Gingele et al., 2007; Sikes et al., 2009; De Deckker et al., 2012; Bostock et al., 2013).

River and lake records indicate higher energy fluvial activity and higher lake levels around 30 ka (e.g. Lake George, Coventry, 1976, Fig. 5). At approximately the same time (~32 ka) the first glaciers appeared in the Snowy Mountains since MIS4 (Barrows et al., 2001). The Lake Keilambete record suggests generally high lake levels at the same time, although this is only constrained by two conventional radiocarbon dates (Bowler and Hamada, 1971) and warrants further investigation. Large channels developed on all rivers feeding the Lachlan, Murrumbidgee, Murray and Goulburn Rivers (collectively the Riverine Plains) (Bowler, 1978; Page et al., 1994, 1996, 2009; Page and Nanson, 1996; Hamilton, 2006; Kemp and Rhodes, 2010). The large palaeochannels preserved in the MDB and dated at 35–25 ka (the Gum Creek Phase) and are interpreted to reflect high rates of lateral channel activity, driven by high sediment loads derived from denuded and destabilised headwater catchments leading up to the coldest period during the LGM (Page et al., 1996, 2001). The same Gum Creek Phase has also been recognised in alluvial deposits on a number of coastal rivers east of the Great Dividing Range in NSW (Nanson et al., 2003).

Estimates from meander wavelengths and reconstructed channel cross-sections in the MDB suggest that bankfull discharges were between 5 and 10 times larger than present between 35 and 25 ka (the Gum Creek Phase) (Page and Nanson, 1996; Kemp and Spooner, 2007; Kemp and Rhodes, 2010). On the Riverine Plains, lakes fed by the major river systems were deep or overflowing (Page et al., 1994) and records from the Willandra Lakes suggest a period of high lake levels at Lakes Mungo and Arumpo just prior to 30 ka, followed by oscillating lake levels persisting after this time and throughout the LGM (Bowler, 1998; Bowler et al., 2003, 2012; McIntosh and Barrows, 2011). This fluvio–lacustrine activity has been argued to represent increased runoff within the catchments in response to increased snowmelt in the highland headwaters, combined with reduced vegetation cover (Kemp and Rhodes, 2010). During the early Last Glacial period, the Willandra Lakes record shows a progressive drop in lake levels, intermittent with periods of clay deflation during what is referred to as the Upper Mungo unit (Bowler et al., 2012).

Periods of increased formation rates from 35 to 20 ka in the Naracoorte Cave speleothems suggest a phase of increased effective precipitation that correspond with the timing of pre-LGM fluvial conditions across South East Australia (Ayliffe et al., 1998; Darrénougué et al., 2009). In contrast, increased sediment deposition in Robertson Cave, also part of the Naracoorte Cave system, has been used as evidence of drier and colder conditions from 30 ka (Forbes et al., 2007). Increased regional- and local-scale aridity at ~30 ka is indicated by a peak in aeolian sediment flux in the Native Companion Lagoon record, southeast Queensland (Petherick et al., 2008).

Broadly speaking, a cooler, drier climate from ~30 ka is also indicated by dramatic shifts in vegetation assemblages (i.e. increased representation of herbaceous taxa at the expense of arboreal taxa) (e.g. Dodson, 1975; Harle, 1997; Colhoun et al., 1999; Kershaw et al., 2007; Builth et al., 2008; Petherick et al., 2008). It should be noted that the colder, drier conditions of glacial stages means that there were complex feedbacks between climate, vegetation and atmospheric CO_2 levels (Levis et al., 1999). As such, vegetation may have been influenced by lowered CO_2 levels (which favour C_4 grasses) rather than climate (Levis et al., 1999). As yet, this influence is not quantifiable.

Southeast Queensland and northern NSW observe a more open landscape, albeit with a slightly greater representation of arboreal taxa (predominantly Casuarinaceae) compared with sites further south (Black et al., 2006; Williams et al., 2006; Moss et al., 2013). The records from Fraser and North Stradbroke Islands, subtropical Queensland, also record the presence of rainforest taxa throughout the early Last Glacial period (Donders et al., 2006; Petherick et al., 2008; Moss et al., 2013). Further south, southern and western Victoria, South Australia and the exposed Bassian Plain developed an open and virtually treeless landscape (Dodson, 1975, 1977; D'Costa et al., 1989, 1993; Edney et al., 1990; Van De Geer et al., 1994; Harle, 1997; Kershaw et al., 2007). An exception is the record from 30 ka Lake Surprise, southern Victoria, which although dominated by Asteraceae, shows a peak in *Eucalyptus* during the LGM (Builth et al., 2008).

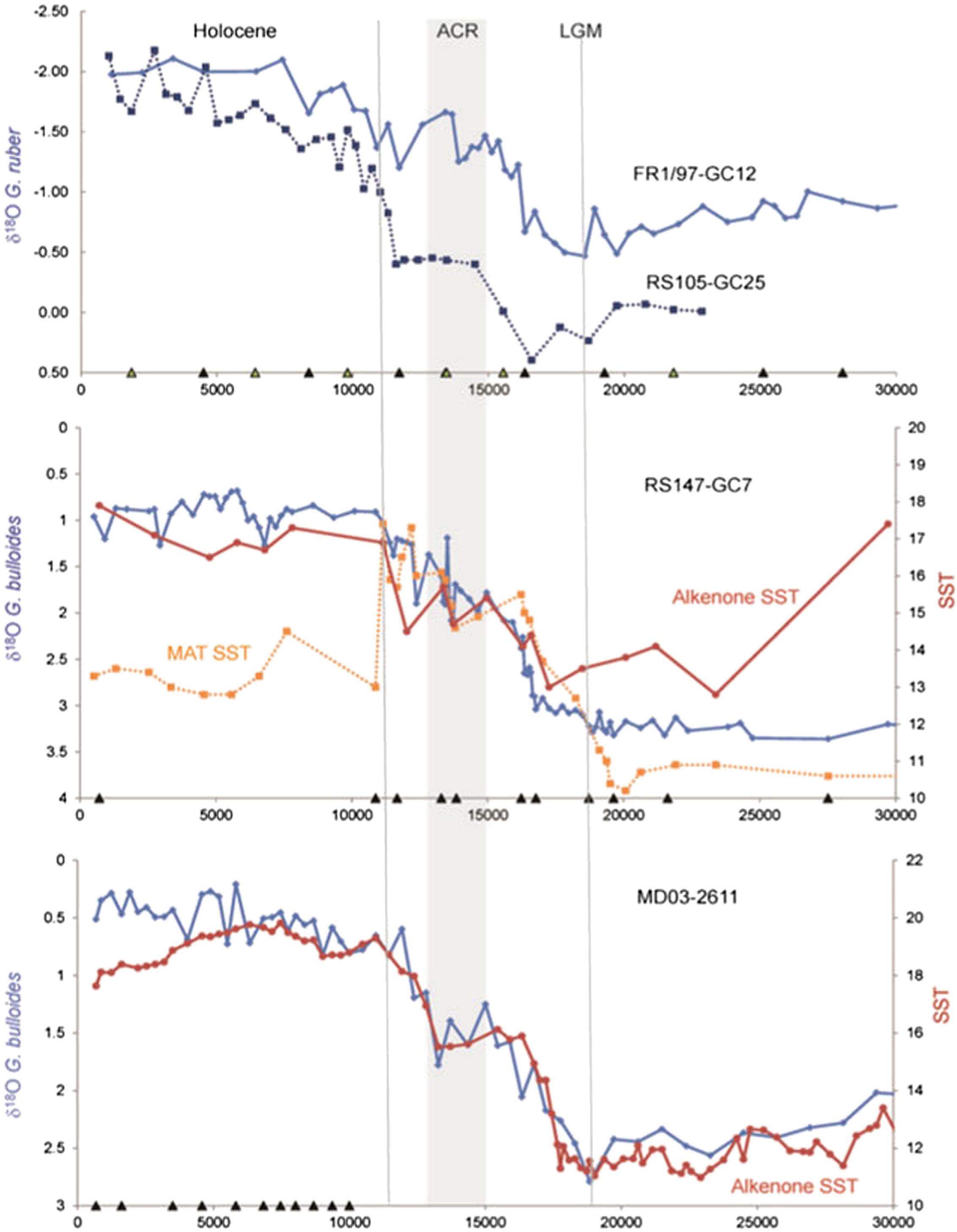

Fig. 4. Top graph shows the $\delta^{18}O$ *G. ruber* records of FR1/97-GC12 (Bostock et al., 2006) and RS105-GC25 (Troedson and Davies, 2001). The middle graph shows the $\delta^{18}O$ *G. bulloides* (blue) and foraminiferal assemblage MAT SST (orange) and alkenone SST (red) from core RS147-GC7 (Sikes et al., 2009). The bottom graph shows the $\delta^{18}O$ *G. bulloides* (blue) and alkenone SST (red) from MD03-2611 (Calvo et al., 2007; Gingele et al., 2007). The black triangles highlight the radiocarbon samples (green triangles for RS105-GC7). The LGM determined by maximum in $\delta^{18}O$ in most of the cores at ∼19 ka is highlighted by a dashed grey line. The ACR is highlighted by a grey box from 12.5 to 15 ka, the start of the Holocene is shown by a grey dashed line at ∼11.5 ka.

An open landscape is also inferred from Tasmanian pollen records, which are characterised by an increase in the importance of alpine/subalpine herbland/shrubland plant assemblages (Colhoun et al., 1982, 1999; Colhoun, 1985b). Charcoal records generally indicate relatively high biomass burning until ca 24 ka, decreasing into the LGM (Mooney et al., 2011).

The differences in vegetation assemblages between the pollen records may be a reflection of regional heterogeneity in response to limiting climate variables. For example, it is likely that precipitation was not a limiting factor for the development of broad vegetation types (i.e. forest, grassland, alpine, etc) in the high rainfall mountainous zone of western Tasmania throughout the last 30 ka

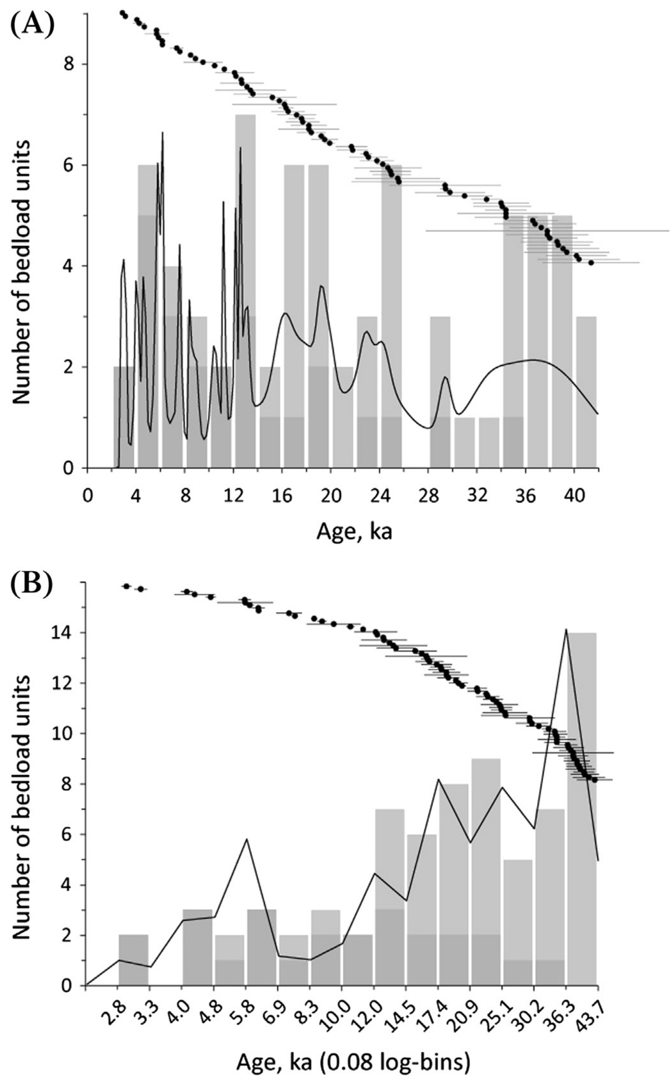

Fig. 5. Compilation of OSL, TL and ^{14}C-dated bedload units from rivers in the Murray–Darling Basin plotted as frequency distributions of depositional ages ($n = 73$); (A) Ranked age plot with 1σ errors, histogram (linear 2 ka bins) and kernel density estimate (0.2 ka bandwidth) plotted as non-dimensional probability, and (B) The same data separated into 0.08 logarithmic bins, which expand with age as indicated, and kernel density estimate (0.08 log ka bandwidth) plotted as non-dimensional probability. ^{14}C dates are denoted by dark-grey shading in the histograms (Bowler, 1978; Bowler et al., 1978; Page et al., 1991; Page and Nanson, 1996; Ogden et al., 2001; Kemp and Spooner, 2007; dated units derived from; Power et al., 2008; Kemp and Rhodes, 2010). Ages <2 ka are excluded and the weighted mean age is plotted for bedload units with multiple dates.

(Colhoun et al., 1999), responding, rather, to temperature fluctuations (Colhoun et al., 1999; Fletcher and Thomas, 2010a) and fire (Fletcher and Thomas, 2010b), with precipitation changes through this time implicated in compositional changes within vegetation types (Fletcher and Moreno, 2011, 2012). Conversely, records on the mainland, where moisture is and has been a limiting factor for vegetation, pollen records are likely to have responded more strongly to changes in moisture availability (i.e. precipitation, surface runoff, seasonal snowmelt) (Colhoun et al., 1999).

4.2. Last Glacial Maximum (LGM) ~22–18 ka

The LGM was characterised by limited glacial advance in the Tasmanian highlands and the Snowy Mountains, with the maximum extent of glaciation occurring synchronously between these distant regions at ~19 ka (Colhoun et al., 1994, 1996; Colhoun and Fitzsimons, 1996; Barrows et al., 2001, 2002). The highlands of western and central Tasmania were the most extensively glaciated regions of Australia, with an estimated 450 cirques and two ice caps covering ~108 km^2 (Fig. 1c).

Arboreal pollen is virtually absent in pollen records from western Tasmania during the LGM and estimates of maximum temperature depression range from between 6.5 °C using qualitative inferences of maximum tree-line depression from pollen records (Colhoun, 1985a; Colhoun et al., 1999) to between 3.7 and 4.2 °C using quantitative pollen-based transfer functions (Fletcher and Thomas, 2010a). The latter closely matches estimates of maximum LGM SST depression immediately offshore from the pollen sites (Barrows and Juggins, 2005; Lopes dos Santos et al., 2012) and is consistent with a persistence of the present day coupling of ocean and atmosphere temperature regimes in this oceanic locale (Fletcher and Thomas, 2010a).

Evidence from $\delta^{18}O$ of the planktonic foraminifera *Globigerinoides ruber* in marine cores suggest that SSTs off Sydney were as much as 3–5 °C lower during the LGM (Bostock et al., 2006), while temperatures in the subtropics were only 2 °C lower (Troedson and Davies, 2001). The larger offset in $\delta^{18}O$ between cores FR1/97-GC12 from the southern Great Barrier Reef and RS105-GC25 off Noosa in southeast Queensland (Fig. 2) equivalent to a 3 °C drop in temperature, is interpreted as the separation of the EAC from the coast at ~26°S (Bostock et al., 2006). The more northerly position of the EAC separation and the Tasman Front is also supported by SST estimates from foraminiferal assemblages (using the MAT), along the Lord Howe Rise and the Tasman Sea, which indicate that cores between 25 and 33°S experienced SSTs 3–5 °C lower during the LGM glacial than present (Martinez, 1994; Martinez et al., 1999; Barrows and Juggins, 2005).

In the south, offshore southern Victoria and Tasmania, foraminiferal assemblages and $\delta^{18}O$ also suggest a decrease in SSTs of as much as 3–5 °C (Fig. 2) (Passlow et al., 1997; Barrows and Juggins, 2005; Sikes et al., 2009), while alkenone-based reconstructions indicate SSTs were reduced by 5–7 °C compared to present (Pelejero et al., 2006; Sikes et al., 2009; De Deckker et al., 2012). This difference between the foraminiferal modern analogue technique SSTs and alkenone SSTs estimates may be due to the seasonal preferences of foraminifera which are dominant in spring and autumn (King and Howard, 2001) and the coccolithophores from which alkenones are derived that bloom in summer (Sikes et al., 2009). SST change is not uniform across the South Tasman Rise region and this is interpreted as a northerly shift of the STF (presently found at 47°S in summer and 45°S in winter) which was likely butting up against the south coast of Tasmania at 43.5°S during the LGM (Passlow et al., 1997; Sikes et al., 2009).

On the west coast of Australia sedimentary cores suggest a 6–9 °C decrease in SST ages and a thickening of the mixed layer during the LGM from foraminiferal assemblages (Martinez et al., 1999). This is caused by a decline in the strength of the LC, caused by a reduced ITF leading to the increased dominance of the cooler West Australian Undercurrent (Wells et al., 1994; Wells and Okada, 1996; Okada and Wells, 1997; Martinez et al., 1999; Takahashi and Okada, 2000; De Deckker et al., 2002; Barrows and Juggins, 2005; Spooner et al., 2011; Reeves et al., 2013). There is some evidence for pulses of upwelling bringing nutrient rich water to the west coast of Australia during the LGM (Wells et al., 1994; Rogers and De Deckker, 2007), but it is unlikely that there was ever a fully developed upwelling system comparable to other eastern boundary currents in the Southern Hemisphere. The latitudinal limit of the reduced LC during the LGM is equivocal with Barrows and Juggins (2005) using foraminiferal assemblages to suggest that the LC probably reached as far as 32°S, while Takahashi and

Okada (2000) used calcareous nannofossils to show that it did not reach as far south as 24°S. This may again be due to the seasonality of these organisms and the LC stronger during winter than summer when the coccolithophores bloom. It is certainly unlikely that the warm waters of the LC reached South Australia, as foraminiferal MAT and alkenones suggest SSTs were 5–7 °C lower than present in this region (Calvo et al., 2007). It may have been that the STF was as far north of 36°S during the glacial off the south coast of Australia (where the STF currently sits between 38 and 40°S) (Fig. 2) and thus this region was bathed in cool Subantarctic waters (Calvo et al., 2007; De Deckker et al., 2012).

Hydrological records in the MDB suggest that the LGM was characterised by declining river channel activity (Page et al., 1991, 1996; Kemp and Rhodes, 2010) on the Lachlan and Murrumbidgee Rivers, but remained substantial on the lower Darling River (Bowler et al., 1978), although the latter has its headwaters in the subtropics and flows through the semi-arid zone. The Willandra Lakes, fed by a distributary of the Lachlan River, experienced oscillating lake levels consistent with seasonal fluvial activity, sufficient to fill lakes several hundred kilometres distant from their source (Bowler et al., 2012), and potentially relating to seasonal snowmelt from the glaciated highlands (Kemp and Rhodes, 2010). Recent work by Dosseto et al. (2010) suggests increased sediment production and reduced residence time in the southern MDB during the LGM, possibly indicating high seasonal flow. It was also hypothesised that Lake George filled during the LGM in response to lowered evaporation rates (Galloway, 1965), although this has yet to be confirmed through direct dating.

Generally, pollen records from mainland Australia indicate vegetation in the LGM similar to during the early Last Glacial period (i.e. increased presence of Poaceae, Asteraceae), indicating a relatively cool climate. Increased representation of Asteraceae (Asteraceae Tubuliflorae and *Tubulifloridites pleistocenicus*) is commonly interpreted as being indicative of glacial conditions in a number of pollen records from temperate Australia (Colhoun and van der Geer, 1986; Williams et al., 2006; Kershaw et al., 2007; Builth et al., 2008; Petherick et al., 2008). During the LGM, Asteraceae is identified as far north as North Stradbroke Island, which is now a coastal, subtropical environment. The absence of Asteraceae in the modern flora of North Stradbroke provides further evidence for a significantly different range of climate variables (e.g. temperature, precipitation) during the LGM than today (Petherick et al., 2008).

Charcoal records from temperate Australia are variable during the LGM, but generally show relatively high abundance (Turney et al., 2000). For example, records from Tasmania (e.g. Lake Selina (Colhoun et al., 1999), Egg Lagoon (D'Costa et al., 1993)) and Victoria (e.g. Lake Surprise (Builth et al., 2008), Lake Wangoom (Edney et al., 1990)) generally indicate high levels of burning. Conversely, generally low charcoal abundance is recorded in New South Wales (e.g. Lake Baraba (Black et al., 2006), Lake George (Singh and Geissler, 1985)), which may indicate a dry climate, low biomass availability or decreased anthropogenic burning. The Australasian composite charcoal record of Mooney et al. (2011) also indicates low biomass burning during the LGM, which they suggest is a common feature of glacial stages during the late Pleistocene. Analysis of the composite charcoal record from sites in the subtropical high belt (STH) region of Australia (roughly correlating with the temperate zone) shows no significant differences in the LGM than through the rest of the record (excluding the late Holocene) (Fig. 6).

4.3. *The deglacial ~18–12 ka*

Globally, the deglacial period was a time of significant environmental change and sea level rise. In temperate Australia, increasing temperatures are indicated by the rapid retreat of ice in the Snowy Mountains and Tasmania which commenced from 18 ka (Barrows et al., 2001; Mackintosh et al., 2006). Deglaciation was virtually complete by 16 ka, with a short-lived re-advance at Mt Twynam in the Snowy Mountains at 16.8 ka and at a similar time in Tasmania (Barrows et al., 2001; Mackintosh et al., 2006).

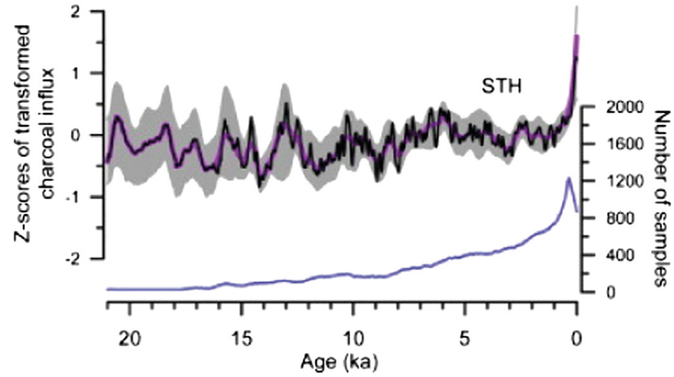

Fig. 6. Macrocharcoal composite from the modern subtropical high pressure belt (STH: 25°S–45°S, 100°E–177°W, broadly temperate Australasia). The data have been smoothed using a window of 400 years (purple line) and a window of 100 years (bold black line) to emphasize the centennial-scale variability. The bootstrapped confidence intervals are based on a 400-year smoothing of the curves. The number of observations contributing to the record within each 400-year window is also shown (blue line) (Mooney et al., 2011).

Rapidly increasing SST and decreasing $\delta^{18}O$ between 18 and ~15 ka is reflected in marine core MD03-2611, Murray Canyons, offshore South Australia (Fig. 4) (Weaver et al., 2003; Calvo et al., 2007). There is a slight delay in the decrease in $\delta^{18}O$ at RS147-GC7 on the Tasman Rise (Sikes et al., 2009), which does not start to decline rapidly until ~17 ka, while cores RS105-GC25 (Troedson and Davies, 2001) and FR1/97-GC12 (Bostock et al., 2006) do not decrease until after 16.5 ka (Fig. 4). The alkenones from core MD03-2611 suggest that the increase in SST lags behind the $\delta^{18}O$, while the modern analogue technique SST increase in core RS147-GC7 leads the $\delta^{18}O$ signal by several thousand years (Fig. 4) (Weaver et al., 2003; Calvo et al., 2007), potentially implying that the spring SST warmed before the summer SST.

Increased post-LGM temperatures are reflected in changes in the pollen records. A phase of substantial re-assortment of vegetation across the temperate zone of Australia saw the re-emergence of arboreal taxa (e.g. Casuarinaceae, *Eucalyptus*) at the expense of herbs and shrubs (e.g. Asteraceae, Chenopodiaceae) in temperate Australia, indicating the onset of warmer and/or wetter conditions (MacPhail, 1979; Markgraf et al., 1986; Harle, 1997; Hopf et al., 2000; Williams et al., 2006; Darrénougué et al., 2009). Rainforest taxa became more prolific in western Tasmania from ~14 ka (MacPhail, 1979; Colhoun et al., 1999; Fletcher and Moreno, 2011).

Amid the general trend of increasing temperature and/or precipitation, there is some indication of short-lived returns to near-glacial conditions during the deglacial. For example, reanalysis of the Lake Vera pollen record from western Tasmania reveals fluctuating moisture trends during the deglacial (MacPhail, 1979), with increased relative moisture between ~14 and 11.7 ka coincident with decreased charcoal abundance across western Tasmanian sites (Fletcher and Moreno, 2011, 2012). Vegetation assemblages from North Stradbroke Island indicate a return to drier condition ~15–12 ka (Moss et al., 2013). This timing co-incides with the Antarctic Cold Reversal (ACR: ~14–12.5 ka; (Blunier et al., 1997)), as recorded in Antarctic ice cores. A slight reversal/stabilisation between 15 and 12.5 ka also occurs in $\delta^{18}O$ and SST in several of the marine cores off southern Australia (Fig. 2; Calvo et al., 2007; Sikes et al., 2009; Lopes dos Santos et al., 2012). This is likely due to decline in the LC and SAC and the increased influence of the Flinders Undercurrent fed by Subantarctic waters.

It has been suggested that there was a cooling in the South Australian Bight during the Younger Dryas (YD: ~12.9–11.6 ka) (Lowell and Kelly, 2008) from increases in $\delta^{18}O$ signal of planktonic foraminifera (a combination of ice-volume, temperature and salinity) and a high concentration of iron in the cores (Andres et al., 2003). However, there is no evidence for the YD from SST proxy data in other marine records (Calvo et al., 2007; De Deckker et al., 2012; Lopes dos Santos et al., 2012), which suggests this may be a response to salinity rather than temperature or ice-volume. Furthermore, a recent review of records from eastern Australia indicates no conclusive evidence for significant climatic variability during the YD chronozone in the region (Tibby, 2012).

At Lake George, shoreline and ostracod evidence suggests high lake levels around ~14 ka and again at ~10 ka, although a lack of ages between these peaks cannot confirm whether these conditions persisted throughout (De Deckker, 1982; Fitzsimmons and Barrows, 2010). High bankfull discharges are indicated by large, actively migrating channels on the Murray, Goulburn and Murrumbidgee rivers, occurring from ~20 to 13 ka and was termed the Yanco Phase by Page et al. (1996, 2009). A similar period of fluvial activity has been recognised by Nanson et al. (2003) as occurring on the coastal rivers of NSW. Towards the end of this period, a general increase in precipitation in the Murray River Basin is suggested by higher proportions of fluvial illite in Murray Canyon sediments after 13.5 ka (Gingele et al., 2007), consistent with an increased input of SWW derived rainfall.

4.4. Holocene ~12–0 ka

The Holocene in the Australian temperate zone was characterised by a relatively warm and wet climate, as indicated by almost all of the proxy records. Superimposed on the "warm and wet" Holocene interglacial are peaks and troughs in effective precipitation at varying times, indicating variability which while not as substantial as that occurring during glacial–interglacial transitions is nevertheless notable. Given the more subtle climate variation of the Holocene, differences between reconstructed climate trends is to be expected. Hence, this discussion focuses on composite records or those that draw inferences from a variety of locations. In a similar fashion to the Australian late Pleistocene record, the majority of Holocene data is derived from the eastern part of the continent from sites that derive their moisture from proximity to the ocean.

4.4.1. Early to mid Holocene ~12–6 ka

On the east coast of Australia, there is evidence from stable isotopes for early Holocene re-invigoration and strengthening of the EAC and the initiation of modern ocean circulation in the southwest Pacific (Bostock et al., 2006). RS147-GC7 also shows a maximum in MAT SST (between 13.4 and 11 ka BP) and a peak in alkenone SST in the early Holocene, indicating this is the warmest period for both spring and summer and suggesting that STF was at its most southerly summer position at 47.5°S (Sikes et al., 2009). Early Holocene peaks in SST are also evident in other cores (Martinez et al., 1999; Calvo et al., 2007; Moros et al., 2009; De Deckker et al., 2012).

Core MD03-2611 offshore South Australia provides a detailed Holocene SST record which has a dip between 10 and 7.5 ka (Calvo et al., 2007), probably caused by the influx of weak stratified, relatively cold water of the Flinders Undercurrent evidenced by an increase in the abundance of the upwelling planktic foraminifera *Globigerina bulloides* (Moros et al., 2009). A sharp rise in *Globigerina rubescens and Globigerina tenella* occurs after 7.5 ka in MD03-2611. In addition, Gingele et al. (2007) determined higher Murray River discharge offshore South Australia between 9.5 and 7.5 ka which co-incides with increasing lake levels in Western Victoria (Jones et al., 1998) although the maximum lake levels occur at the end of this period (~7.5 ka; Wilkins et al., 2012). From the fluvial sedimentary record of south-eastern Australia, Nanson et al. (2003) and Cohen and Nanson (2007) recognised the early to mid-Holocene (10–4.5 ka) as a period with enhanced river discharge, which they termed the Nambucca Phase.

A peak in temperature at ~8–7 ka is indicated by the Lynds Cave, north-central Tasmania (Xia et al., 2001). In NSW pollen records (e.g. Redhead Lagoon, Blue Mountains composite records), a peak in regional precipitation is recorded at ~10–8 ka (Williams et al., 2006). Many pollen records reflect another abrupt change in vegetation at the start of the Holocene, indicating a period of significant environmental change, indicated by the further expansion of arboreal taxa (e.g. Kershaw et al., 2007; Moss et al., 2013). Increases in both thermophyllous and hygrophyllous pollen taxa from ~12 ka indicates the establishment of warmer and possibly wetter conditions across a broad swath of environments (Dodson and Wright, 1989; Williams et al., 2006; Kershaw et al., 2007; Kiernan et al., 2010; Moss et al., 2013). The expansion of (principally) forest taxa occurs at the expense of herbaceous taxa, such as *T. pleistocenicus*, that indicate cool/cold glacial conditions at a range of sites across temperate Australia (e.g. D'Costa et al., 1989; Edney et al., 1990; D'Costa et al., 1993; D'Costa and Kershaw, 1995; Williams et al., 2006). At higher altitude, *Eucalyptus* pollen at Caledonia Fen (Fig. 1a) increased abruptly from 5% to 60% at ~11 ka, indicating an increase in temperature and/or effective precipitation at that site (Kershaw et al., 2007).

The onset of the Holocene is characterised by high levels of biomass burning in Australasia, which is suggested to be typical of interglacial stages during the late Pleistocene (Mooney et al., 2011). This increase in burning is not noticeable in the composite STH records, but there is an increasing trend from the early to mid Holocene, with high biomass burning in southernmost Australia (including Tasmania), but low biomass burning in NSW (Fig. 6) (Mooney et al., 2011).

4.4.2. Mid to late Holocene ~6–0 ka

SSTs on the South Tasman Rise decline from an early Holocene maximum, to modern SSTs and modern STF summer extent (~47°S) in the late Holocene (Sikes et al., 2009). The planktonic foraminifera $\delta^{18}O$ data (which is a combination of SST and salinity during the Holocene) from most cores in this region, however, continue to decline from the initial minimum in the early Holocene throughout the mid-late Holocene, albeit at a much slower rate than the deglacial period (Troedson and Davies, 2001; Bostock et al., 2006; Calvo et al., 2007; Sikes et al., 2009). Sea surface temperature reconstructed from the high resolution MD03-2611 record reaches a maximum at 6 ka, which is then followed by a steady decrease in summer SST in the late Holocene (Calvo et al., 2007). During this SST decline from the mid Holocene, there are several ~1500 yr cycles in the $\delta^{18}O$ *G. bulloides* (Moros et al., 2009).

At Lake George, lake level rise took place during the mid-Holocene (~6 ka), followed by a phase of fluctuating lake levels responding to seasonal or longer term dry conditions, and finally regression, from ~2 to 0.3 ka (De Deckker, 1982; Fitzsimmons and Barrows, 2010). From 4.5 ka onwards, reduced discharges and/or increased discharge variability lead to the vertical accretion of many fine-grained floodplains in coastal NSW valleys (Cohen and Nanson, 2007; Cheetham et al., 2010), whilst other sedimentary records (e.g. cut-and-fill floodplains) have been subject to ongoing intrinsically driven erosion and aggradation (Cheetham et al., 2010). A short-lived phase of aeolian dust deposition at Lake George, indicating dry lake conditions, took place around 1 ka, and was followed by a short-lived high lake phase during ~0.6–0.3 ka (Fitzsimmons and Barrows, 2010).

Although confounded by a legacy of spatial patchiness and poor temporal control, a number of earlier studies highlight the probable dry phases since 6 ka, augmenting the notion of a variable hydroclimate over this time. These include: a drier than modern phase between 6 and 5 ka (Harrison, 1993; Pickett et al., 2004; Black et al., 2006); a substantial drying in between ~3 and 1 ka (Ahmad, 1996) and a series of studies that argue for episodic drier/wet phases and/or a general trend towards drier conditions between 5 and 0 ka (Bowler, 1976; Bowler et al., 1976; Prahl et al., 1988; Wasson and Donnelly, 1991; McGrath and Boyd, 1998; Stanley and De Deckker, 2002; Gingele et al., 2007; McGowan et al., 2008).

Estuarine and near-coastal records from southern WA suggest a relatively dry climate at ~6–4 ka (Semeniuk and Searle, 1985; Semeniuk, 1986; Yassini and Kendrick, 1988), although these data do not correlate well and the inferred climate changes are relatively minor. Vegetation reconstructions from southern WA (which at the time of publication has only Holocene pollen records) show extensive spatial variability between sites. For example, Churchill (1968) found for wetter conditions 6–5 ka, 5–2.5 ka drier conditions (maximum aridity at 3.2 ka) and 2.5–0 ka increasing effective precipitation to modern conditions. By contrast, Dodson and Lu (2000) found evidence for increasing effective precipitation from 4.8 ka. Gouramanis et al. (2012) found that the mid to late Holocene was characterised by periods of significant wetting and drying, associated with changes in effective precipitation. Disparities between pollen records has led to the suggestion that vegetation in southern WA is not responding to climate, but rather to local factors such as fire regimes and species composition (Newsome and Pickett, 1993; Itzstein-Davey, 2004).

Pollen and charcoal records from the rest of the temperate zone suggest that the mid to late Holocene was characterised by increased sub-millennial scale variability, with an overall increase in the frequency and/or intensity of fires after ~6 ka (Kershaw et al., 2007; McGowan et al., 2008; Fletcher and Moreno, 2011). At 6–5 ka, the composite charcoal record of Mooney et al. (2011) shows increased biomass burning in the northern temperate zone and decreased biomass burning in the south and Tasmania, essentially the opposite to conditions between 8.5 and 7.5 ka (Mooney et al., 2011).

5. Summary

The modern temperate zone is a diverse region encompassing a range of climate regimes. Despite this, it is possible to synthesise available palaeoenvironmental records to identify broad trends in climatic and environmental variability over the past 30 ka (Table 2; Fig. 7). Proxy records indicate cool conditions during the early Last Glacial period (Fig. 7a). High lake levels and increased fluvial activity in the southern MDB indicate relatively wet conditions, possibly due to seasonal snowmelt runoff in the river headlands (Bowler, 1978; Page et al., 1994, 1996, 2009; Page and Nanson, 1996; Hamilton, 2006; Kemp and Rhodes, 2010). By contrast, the pollen records generally indicate dry conditions (Dodson, 1975; Harle, 1997; Colhoun et al., 1999; Kershaw et al., 2007; Builth et al., 2008; Petherick et al., 2008), which is also indicated in the Robertson Cave speleothem record (Forbes et al., 2007) and the Native Companion Lagoon record of aeolian sedimentation (Petherick et al., 2008).

This disparity may reflect the poor degree to which runoff relationships are constrained. We have a limited understanding of the correlation between rainfall and runoff and their association with the vegetation records. Poor chronological control of the timing of deposition may also be a factor. Alternatively, the pollen records may be responding to temperature and/or lower atmospheric CO_2, rather than moisture availability.

However, wetter and drier conditions in the lead up to the LGM are not necessarily mutually exclusive. For example, increased seasonality (i.e. wet winter, dry summer) would allow high lake levels and fluvial activity to occur at the same time as the vegetation assemblages (which may not respond rapidly to climatic variability) reflecting increasing aridity. Alternatively, fluctuating wet and dry conditions may be an indication of longer-term climatic variability e.g. the ENSO.

Table 2
Summary of paleoenvironmental implications of the temperate Australia region over the last 30 ka.

Age	Marine	Geomorphic	Pollen	Glacial
Early Last Glacial period 30–22 ka	SST 3–9 °C lower than modern during the glacial. Equatorward shift in the mean position of the STF, which resulted in cool, Subantarctic waters surrounding southern Australia.	35–25 ka: Gum Creek Phase: Pluvial conditions. Lakes Keilambete and George full. Development of large river channels on the Riverine Plain. High sediment loads. Lake Mungo full 30–26 ka and 24–20 ka.	Generally increased presence of grasses and herbs indicating a relatively cool, dry climate.	Extensive glaciation in Tasmanian highlands. Limited (~15 km^2) ice cover on the mainland. Major glacial advance at ~32 ka.
LGM 22–18 ka	SST 3–9 °C lower than modern during the glacial. Equatorward shift in the mean position of the subtropical front, which resulted in cool, Subantarctic waters surrounding southern Australia.	Drier conditions. 20–17 ka: high lake levels at Mungo, Keilambete and George, perhaps due to high (cf. modern) runoff associated with seasonal snowmelt.	Vegetation dominated by grass and herb taxa. Reduced presence of trees. Temperature 4–6 °C lower than modern. Reduced effective precipitation.	The most significant glacial advance at ~19.1 ka.
Deglacial period 18–12 ka	Rapid increase in SST sometime between 18 and 15 ka, corresponding with MWP-1A. Stabilisation in SSTs 15 to 12.5 ka corresponding with the ACR. STF positioned polewards.	20–13 ka: Yanco phase. Large migrating channels across the Riverine Plain, indicating increased fluvial activity. Increased precipitation after 13.5 ka.	Increased representation of arboreal taxa indicating increased effective precipitation and temperatures.	Reduced extent of glaciated areas. Re-advance in Snowy Mountains at ~16.8 ka.
Holocene 12–0 ka	Early Holocene peak in SSTs. 10–7.5 ka dip in SST. Maximum SSTs at 6 ka, then steady decrease to modern.	12–7 ka: Change from bedload to dominantly suspended-load sedimentation on the rivers of the MDB signals widespread re-vegetation of the highlands and a return to full interglacial conditions. 8–4 ka: Nambucca phase –mid-HCO. From 6 ka increased drying of lakes to modern conditions.	Increased representation of rainforest taxa, indicating increased effective precipitation. Some evidence for a mid-HCO and possible increased aridity in the late Holocene, but disagreements between records suggests caution with interpretation.	Continental Australia and Tasmania ice-free.

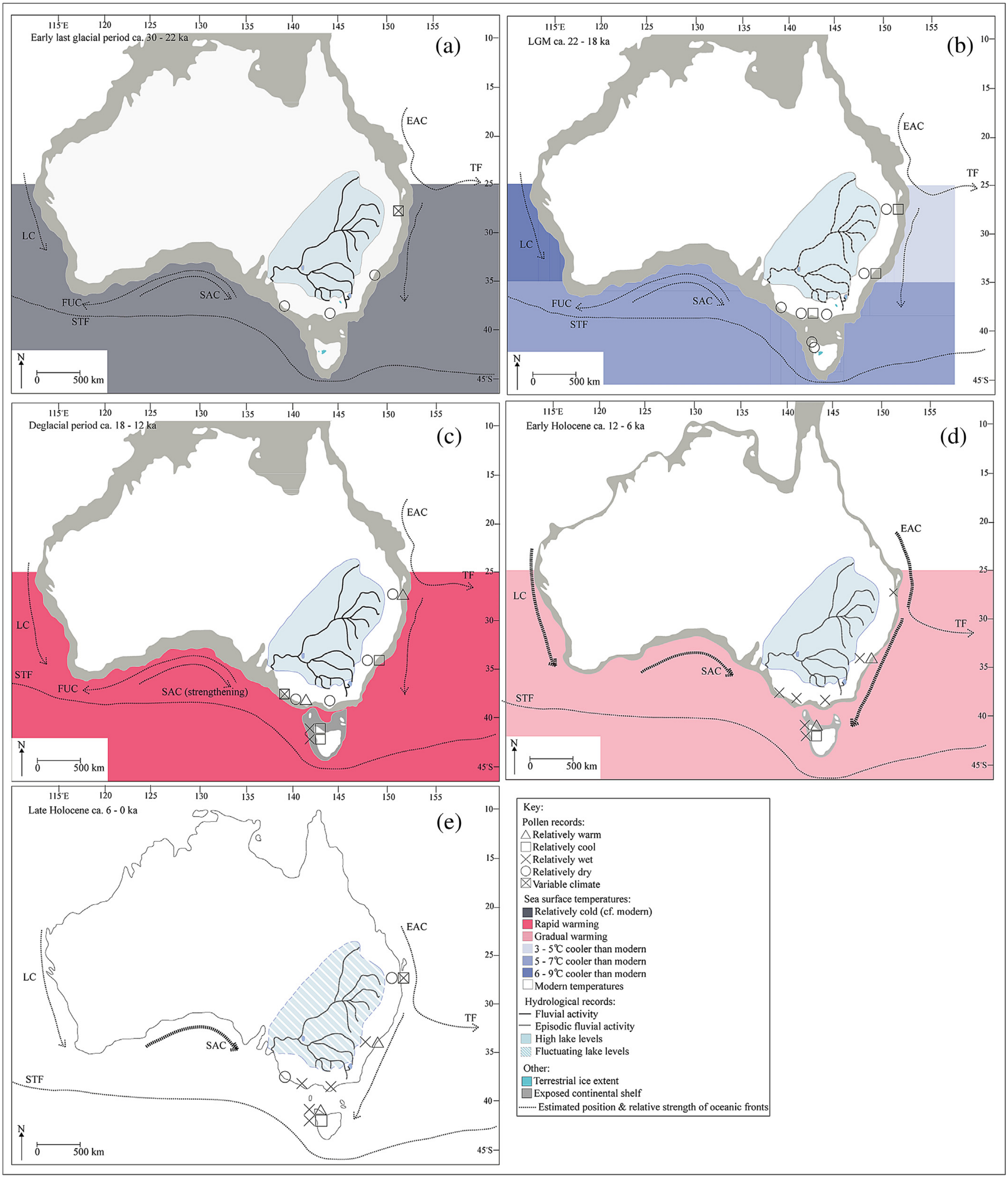

Fig. 7. Comparison of climatic conditions indicated by proxy records from temperate Australia for the periods (a) early Last Glacial period, (b) LGM, (c) the deglacial period, (d) early to mid-Holocene and (e) mid to late Holocene, indicating the position of oceanic fronts and currents (Martinez, 1994; Passlow et al., 1997; Martinez et al., 1999; Troedson and Davies, 2001; Barrows and Juggins, 2005; Bostock et al., 2006; Pelejero et al., 2006; Sikes et al., 2009), relative SSTs (Passlow et al., 1997; Bostock et al., 2006; Calvo et al., 2007; Moros et al., 2009; Sikes et al., 2009; De Deckker et al., 2012), the changing extent of the Australian coastline (Lambeck and Chappell, 2001; Gillespie, 2002), glacial extent (Barrows et al., 2001, 2002), hydrological activity (Page et al., 1991, 1996; Kemp and Rhodes, 2010; Cohen et al., 2011) and climatic conditions inferred from pollen records from Native Companion Lagoon (Petherick et al., 2008), Redhead Lagoon (Williams et al., 2006), Caledonia Fen (Kershaw et al., 2007); Lake Leake (Dodson, 1974, 1975), Lake Wyrie (Dodson, 1977), Lake Selina (Colhoun et al., 1999), Pulbeena Swamp (Colhoun et al., 1982). Where STF = subtropical front, TF = Tasman Front, FUC = Flinders Undercurrent, EAC = East Australian Current.

Proxy records generally indicate that the LGM was the coldest and driest phase in the past 30 ka (Fig. 7b). Maximum glacial advance occurred synchronously between Tasmania and the Snowy Mountains at ~19 ka (Barrows et al., 2001, 2002). Marine records indicate cooler SSTs than modern during the LGM, although these range latitudinally from 2 °C (subtropics) to 7 °C (southern Australia), which is interpreted as reflecting the more equatorward position of the EAC separation and the Tasman Front and the northward movement of the STF during the LGM (Troedson and Davies, 2001; Bostock et al., 2006). Pollen records show the prevalence of grasses and herbs, suggesting a relatively open landscape (Dodson, 1975; Harle, 1997; Colhoun et al., 1999; Kershaw et al., 2007; Builth et al., 2008; Petherick et al., 2008).

Declining river channel activity is indicated in the Murrumbidgee and Lachlan Rivers (Page et al., 1991, 1996; Kemp and Rhodes, 2010). However, a marked increase in bedload activity is indicated in the southern Darling River (Bowler et al., 1978). Whilst there are few bedload ages recorded between 22 and 20 ka (Fig. 4a), a more general peak in fluvial activity is recorded between 24 and 17.5 ka (Fig. 4b), coeval with high lake levels in the MDB. Evidence for increased bedload transport has been proposed to be due to increased runoff from seasonal snowmelt and reduced vegetation cover (Kemp and Rhodes, 2010). This would suggest a major change in the water balance with increased surface-water runoff despite indications for lower mean annual temperature and potentially mean annual precipitation.

During the deglacial period, vegetation reconstructions show the expansion of arboreal taxa at the expense of herbs and grasses which implies an increase in temperature and/or precipitation, consistent with increased fluvial activity and high lake levels in the MDB (Fig. 7c). Wetter conditions in western Tasmania and SW Victoria between ~14 and 12 ka occur simultaneously with drier conditions in parts of the MDB and eastern Tasmania, reflecting an increase in the influence of the SWW across the southern temperate zone (Fletcher and Moreno, 2012).

During the early deglacial period, decline in $\delta^{18}O$ in the northern marine cores leads the South Australia core (Weaver et al., 2003; Bostock et al., 2006; Calvo et al., 2007; Sikes et al., 2009), suggesting that perhaps there is a delay in the cold water and ice melt signal reaching the northern cores via the tropics transported in the South Pacific gyre. There is a clear ACR present in most of the high-resolution sediment records, with no definitive evidence for the YD, reflecting the importance of Antarctica and the Southern Ocean in temperate Australia's climate.

At the end of the deglacial and in to the early Holocene the tropical influence (e.g. trade winds) dominates, with the warmest SSTs in the past 30 ka. The interglacial oceanographic circulation is re-initiated, with the stronger, and warm, EAC and LC flowing south down the east and west coasts of Australia and seasonally along the south coast (Fig. 7d) (Bostock et al., 2006). Vegetation records indicate the further expansion of sclerophyll woodland taxa and rainforest at the commencement of the Holocene, indicating increased temperatures and/or increased effective precipitation (Dodson and Wright, 1989; Williams et al., 2006; Kershaw et al., 2007; Kiernan et al., 2010; Moss et al., 2013). A decrease in SWW flow across the southern temperate zones between 11 and 8 ka is indicated by a decrease in moisture in areas where precipitation is positively correlated with SWW wind-speed in the modern climate, with a reversal of this trend occurring between 8 and 6 ka (Fletcher and Moreno, 2012). An increase in climatic variability and a spatial patterning of moisture balance changes (Fig. 7e) through the last 6 ka consistent with ENSO suggests that SE Australia responded rapidly to the onset of ENSO variability in the equatorial Pacific Ocean (Moy et al., 2002). This is supported by marine records, which suggest that from the mid- to late-Holocene the variability in the records was most likely caused by ENSO.

While this increased spatial and temporal variability occurs at a time of increasing sub-millennial ENSO variability (Moy et al., 2002; Yan et al., 2011), disentangling the complex array of factors that can influence charcoal and pollen signatures at shorter timescales, such as climate, ecology and human activity, requires a more strategic targeting of sites and the development of multiple independent proxies. Nevertheless, it is possible to develop a broad picture of climate change over this period from the available evidence. The millennial-scale synthesis of Fletcher and Moreno (2012) for selected palaeoenvironmental data from the SWW zone of influence suggests a breakdown of the pattern consistent with a dominant SWW control of the southern Australian hydroclimate. These authors contend, via an intra-hemispheric comparison of palaeo-moisture reconstructions, that the onset of an amplified ENSO system after ~6 ka (Moy et al., 2002) shifted the multi-millennial scale patterning of moisture regimes in the Southern Hemisphere from one dominated by the zonally symmetric SWW system, to one dominated by the zonally asymmetric ENSO pattern.

The synthesis of the temperate records from Australia shows that it is possible to gain insight into broader scale climatic and environmental variability without losing the intricacies of individual records. There is always the danger of ignoring the heterogeneous nature of proxy records when attempting to look at regional to continental scale variability. The role of local- and regional-scale responses to different climate variables is crucial for understanding broader-scale patterns. As shown in Fig. 7, it is possible to identify periods of significant environmental change without sacrificing details in individual records. Furthermore, the comparison of multiple, independent proxies is crucial for developing more accurate measures with realistic errors.

6. Future work

Urgent attention needs to be focussed on a number of key shortcomings in the current palaeoenvironmental dataset for temperate Australia. Most critical is the undertaking of data collection at finer temporal and spatial scales in key climatic locations. Low sedimentation and often times extremely low productivity is a significant obstacle to producing fine-temporal scale analyses in Australia, although the advent of high-precision and fine-scale non-destructive core-scanning techniques (such as ITRAX and reflectance spectroscopy) show significant promise. In-step with finer temporal resolution is a commitment to establishing robust chronologies, through larger numbers of ages and increased spatial cover of datasets. Shifting from single proxy analyses to the development of more multi-proxy datasets, targeting the spatial gaps in the current state of knowledge will also yield significant fruit that will bring Australia on par with the impressive northern datasets. As noted by Turney et al. (2006) in their OZ-INTIMATE contribution, there is an urgent need for more quantitative datasets and better geochronological control, neither of which has really been advanced further. Proxies that have the potential for providing quantitative estimates of environmental variables should be the focus in the future e.g. pollen, ostracods, diatoms and chironomids.

Although south-eastern Australia is relatively well-represented, there is an urgent need for records extending beyond the Holocene from the subtropics and eastern Tasmania. The temperate zone of Western Australia is grossly under represented. In addition, more high resolution marine records, particularly from the east coast, are needed. There is large potential to more adequately develop a stronger understanding of the modern climatic drivers and landscape response to climatic change in order to use geomorphological records as proxies. Finally, palaeo-data has significant potential for setting quantitative boundary conditions for global or regional climate models and there is an urgent need for the incorporation of this data into climate and vegetation models in order to gain greater insight into the drivers of climatic variability during the late Quaternary.

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