Climate variability over the last 35,000 years recorded in marine and terrestrial archives in the Australian region: an OZ-INTIMATE compilation.

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Abstract

The Australian region spans some 60 degrees of latitude and 50 degrees of longitude and displays considerable regional climate variability both today and during the Late Quaternary. A synthesis of marine and terrestrial climate records, combining findings from the Southern Ocean, temperate, tropical and arid zones, identifies a complex response of climate proxies to a background of changing boundary conditions over the last 35,000 years. Climate drivers include the seasonal timing of insolation, greenhouse gas content of the atmosphere, sea level rise and ocean and atmospheric circulation changes. Our compilation finds few climatic events that could be used to construct a climate event stratigraphy for the entire region, limiting the usefulness of this approach. Instead we have taken a spatial approach, looking to discern the patterns of change across the continent.

The data identify the clearest and most synchronous climatic response at the time of the last glacial maximum (LGM) (21 ± 3 ka), with unambiguous cooling recorded in the ocean, and evidence of glaciation in the highlands of tropical New Guinea, southeast Australia and Tasmania. Many terrestrial records suggest drier conditions, but with the timing of inferred snowmelt, and changes to the rainfall/runoff relationships, driving higher river discharge at the LGM. In contrast, the deglaciation is a time of considerable south-east to north-west variation across the region. Warming was underway in all regions by 17 ka. Post-glacial sea level rise and its associated regional impacts have played an important role in determining the magnitude and timing of climate response in the north-west of the continent in contrast to the southern latitudes. No evidence for cooling during the Younger Dryas chronozone is evident in the region, but the Antarctic cold reversal clearly occurs south of Australia. The Holocene period is a time of considerable climate variability associated with an intense monsoon in the

tropics early in the Holocene, giving way to a weakened monsoon and an increasingly El Niño-dominated ENSO to the present. The influence of ENSO is evident throughout the southeast of Australia, but not the southwest. This climate history provides a template from which to assess the regionality of climate events across Australia and make comparisons beyond our region

Keywords

Australia, tropics, temperate, arid zone, Southern Ocean, last glacial maximum, deglacial period, Holocene, INTIMATE

1. Introduction

Australia, the “Island Continent”, spans the latitudinal range of 10^{o}S to 43^{o}S, and is bordered by New Guinea and Indonesia to the north, the Pacific Ocean to the east, the Indian Ocean to the west and the Southern Ocean to the south (Fig. 1). This location means that the continent and the modern climate are juxtaposed between the heat of the equatorial tropics from the Indo-Pacific Warm Pool (IPWP) and the cool waters of the Southern Ocean. Australia’s latitudinal position results in several synoptic-scale controls on the climate including dry descending air associated with the Hadley Cell and the resulting Sub-tropical High Pressure Belt, easterly zonal flow associated with the South East Trade winds and the summer monsoon in the north and prevailing westerlies in the south (Fig. 2). The size of the mainland produces strong continentality with precipitation and temperature gradients from the coast to the dry centre.

The mean climatic conditions across Australia primarily respond to the seasonal zonal circulation by continental heating and cooling and the land-sea temperature contrast (Gimeno et al., 2010), and are driven by multiple atmospheric and oceanic influences and the interactions between them (e.g. Gallant et al., 2012). The non-stationarity, intensity and spatial variability of these climatic drivers, and the degree to which they have interacted through time, are the key criteria for addressing the reconstruction of Australian palaeoclimate.

The last 35 kyr encompasses the end of Marine Isotope Stage (MIS) 3, the transition into full glacial conditions, subsequent deglacial warming and establishment of the Holocene, and includes millennial-scale reversals (e.g. the Antarctic Cold Reversal; ACR) and centennial-scale variability of climate (e.g. the Holocene). This period, into and out of the last glacial, is of key interest to the INTegration of Ice core, Marine and Terrestrial (INTIMATE) project (Barrows et al., editorial). In this paper we bring together the records discussed in each of the four regional synthesis papers in this volume (Bostock et al., Fitzsimmons et al., Petherick et al., Reeves et al., this volume) as the 'broader Australian region'. This broader region extends from 10^{o}N encompassing New Guinea and much of Indonesia, to 65^{o}S, to include the Australian segment of the Southern Ocean, and from 100-165^{o}E (Fig. 3).

This paper provides a synthesis of major climatic events over the last 35 kyr and attempts to determine their regional distribution. We find that changing boundary conditions through time exert variable influence across the region, with regional controls such as the land:sea ratio, the monsoon and the influence of the El Niño Southern Oscillation (ENSO), and changes in the westerly winds and mean state of the Southern Ocean, important for determining the pattern of climate change.

Climate Event Stratigraphy

The original remit of the AUS-INTIMATE project was to develop a climate event stratigraphy for the Australian region (Barrows et al., editorial). From a global perspective, the phenomenon of the bi-polar see-saw (Broecker, 1998) and the continuing debate as to the relative timing of climate change between the hemispheres prompted the question of how global scale climate events are expressed in the Australian region. Clear differences are apparent between the high latitudes of both hemispheres (Blunier and Brook, 2001) indicating that a climate event stratigraphy based on a Greenland ice core record has limited meaning in the Australian region. There is potential to develop a Southern Hemisphere high-latitude stratigraphy based on Antarctic records (e.g. Pedro et al., 2011; EPICA, 2006), but it remains unclear as to how applicable this would be for the Australian region, given its distance from Antarctica. Clearly, we first need to understand how the Australian region responds to the competing influences of changing boundary conditions over such a large region.

Here we document changes across the region to better describe the driving mechanisms of climate change. We take a time-slice approach to synthesize terrestrial and marine records during the late MIS 3 period, the early glacial period, the LGM, the deglacial period including during the ACR and the early-to-mid and late Holocene. The regional compilations (e.g. Bostock et al. Fitzsimmons et al., Petherick et al., Reeves et al., this volume) examine various proxy archives from the equator to the Southern Ocean which together allow us to investigate the relative changes of the tropics versus high latitudes through time, and spatial coherence in response to global climate change across the vast Australian region. This temporal and spatial synthesis then allows for the better characterisation of the drivers of the climatic changes over the last 35 kyr and the regionality of the climate response.

1.1 Modern ocean-atmospheric climate drivers

The following section provides a brief overview of the modern ocean-atmospheric drivers of the Australian climate. We interpret climatic changes over the last 35 kyr as modifications to the major elements of the modern climate system. The oceanographic setting of the region has been discussed elsewhere in this volume (see Bostock et al., Petherick et al., Reeves et al.) and the dominant features are shown in Figure 1.

The modern climate patterns of the Australian region may be considered as including the tropics, encompassing both the perennially wet equatorial and seasonally wet monsoonal regions; the temperate zone of the east, south and south-west coasts, including Tasmania; and the arid interior, which receives little rainfall (<250 mm per annum) (see Fig. 3 for a schematic representation of this subdivision). Australia receives predominantly winter rainfall in the south, summer rainfall in the north and sporadic, winter and summer rainfall in the central region of the east coast (Gentilli, 1971). The ocean circulation patterns and the heat distribution around the continent also play a role in determining the nature and distribution of rainfall patterns across Australia. With few exceptions beyond the eastern Great Dividing Range/Eastern Escarpment, the relatively low altitude of the Australian continent results in minimal orographic influences. The interannual variation in precipitation can exceed that of annual variation, particularly in the arid zone.

The climate of Australia (Fig. 3) is determined by continental-scale atmospheric circulation patterns. In the Southern Hemisphere, warm, moist air is lifted by the Hadley Cell at the equatorial low pressure or Inter-tropical Convergence Zone (ITCZ) and descends around 30^{o}S resulting in the Sub-Tropical High Pressure Belt (STHPB). The sinking air of this STHPB dominates the modern climate of the arid interior (Sturman and Tapper, 1996). The ITCZ moves into the north of Australia during the austral summer, bringing with it the summer monsoon. To the north of Australia lies the Indo-Pacific Warm Pool (IPWP), a region where mean annual sea surface temperatures (SSTs) exceed 28^{o}C and provide a major global source of latent heat release (Gagan et al., 2004). Some of the descending air from the STHPB travels back to the equator along the surface of the Earth, creating the south easterly trade winds, which flow from the east to west across the equatorial Pacific. These trade winds in turn typically bring about deep atmospheric convection (Walker Circulation) over the tropics which result in heavy rainfall over the northeast coast of Australia during summer (November to April).

The climate patterns of the eastern half of Australia are partly modulated on an inter-annual scale by ENSO (e.g. Verdon et al., 2004). Variations in SST across the equatorial Pacific are associated with weakening (El Niño) or strengthening (La Niña) of this Walker Circulation over cycles of several years (see Diaz and Markgraf (2000) for further details). The Inter-decadal Pacific Oscillation (IPO) has similar characteristics to ENSO, but on longer timescales and affecting the wider Pacific Basin (Power et al., 1999). The IPO influences Australia via a modulation of both the magnitude and frequency of ENSO impacts leading to multi-decadal epochs that are significantly wetter or drier than others (e.g. Kiem et al., 2003; Kiem and Franks, 2004). This effect has been identified for at least the past 400 years (Verdon and Franks, 2006). Whilst other inter-annual drivers are relevant in the Australian region (e.g. Indian Ocean Dipole or Southern Annular Mode) most palaeo-proxy data are of lower resolution preventing the identification of such drivers.

South of the Hadley Cell, descending air moves southward in the Ferrel Cell, before ascending at the margins of the polar front (~60°S). The surface wind flow of the Ferrel Cell is responsible for the mid-latitude westerly wind belt or "westerlies" which are largely responsible for winter rainfall in southern Australia (Pitman et al., 2004) and play a dominant role in circulation in the Southern Ocean (Varma et al., 2010). In addition, the cold fronts embedded in the STHPB and associated with the westerlies, influence the landscape through aeolian transport of surface sediments, particularly in the arid zone (Hesse and McTainsh 1999; Hesse, 2010).

1.2 Drivers of past changes

The dominant external forcing on the global climate system over the past 35 kyr has been the changing seasonal and meridional distribution of insolation arising from cyclic changes in the Earth's orbital geometry (Berger, 1978). On glacial-interglacial timescales, feedbacks involving the global carbon cycle and atmospheric greenhouse gases have amplified the response of the climate system to this signal (e.g. Petit et al., 1999). In the Northern Hemisphere, this period has been characterised by the expansion and retreat of ice sheets in response to changes in summer insolation at high latitudes. This resulted in a long build-up of ice sheets during glacial phases and relatively rapid deglaciation, into interglacial phases. By contrast, the continents of the Southern Hemisphere did not possess ice sheets, excepting Antarctica, but the associated changes in sea level altered land:sea ratios and continental connections, and therefore altered ocean currents and regional climates. Changing temperature differentials between land and sea altered wind strengths and precipitation. The variable seasonal cycle of insolation also drove changes in the background state of the tropical Pacific Ocean, with direct consequences for the climate of the Australian region.

The Southern Ocean has played a significant role on the balance of CO_2 between the oceans and atmosphere on centennial-millennial timescales (e.g. Toggweiler, 1996; Sigman et al., 2010). The expansion of Antarctic sea ice during the glacial acted to trap CO_2 in the deep ocean, lowering atmospheric CO_2 levels. Conversely, during the deglacial when sea ice retreated, outgassing of CO_2 occurred on a large scale (e.g. Lourantou et al., 2010; Bostock et al., this volume).

On decadal to centennial timescales, changes in solar irradiance and explosive volcanism are also significant. Existing reconstructions of solar forcing only extend back as far as the early Holocene (e.g. Steinhilber et al., 2009), while reconstructions of volcanic forcing (e.g. Plummer et al., 2012) only cover the last 2,000 years. Thus the potential role of these forcings in driving climatic changes over the past 35 kyr is extremely poorly understood, and changes seen within the Australian climate over this period may therefore have been driven by unknown external events. In particular, there is some evidence that volcanic eruptions, despite being only short in duration, can be sufficient to push the climate system past tipping points into alternative states that can persist for multiple centuries (Miller et al., 2012).

2. Rationale for record inclusion

The synthesis presented here builds on the four review papers in this issue, which deal with the Australian tropical (Reeves et al.), terrestrial (Petherick et al.) and arid interior

(Fitzsimmons et al.) zones and the Southern Ocean (Bostock et al.). Our synthesis builds on an earlier effort as part of the AUS-INTIMATE project by Turney et al. (2006a). Previous regional reviews of climate have been summarised by Williams et al. (2009), who also compiled key long records with quantifications of climate change. In this review we build on this earlier works by incorporating a larger contribution from deep-sea records, covering the Holocene and incorporating new records from all regions.

The criteria for inclusion of records into this synthesis would ideally be: contininuity through the last 35 kyr, sound chronology, centennial-scale or better resolution and unambiguous and quantifiable palaeoclimate estimates. Although this has been achieved in many of the marine and speleothem records, their spatial coverage does not represent the greater Australian region. We have therefore also chosen here to also include: high-resolution short-term records (e.g. corals), discontinuous geomorphic records (e.g. fluvial, lake shore, dune, glacier) where the interpretation is robust (see Fitzsimmons et al., this volume for further discussion), as well as qualitative records (e.g. pollen), although noting the context of the site and the limitations of each record (Fig. 2).

The chronologies of the records were determined using a variety of methods. Radiocarbon is the most common method used, frequently on bulk sediment, charcoal or foraminifera for marine cores and calibrated here using INTCAL09 or MARINE09 (Reimer et al., 2009). Other methods include optically-stimulated luminescence (OSL) and thermoluminescence (TL) dating of sediments, predominantly applied to aeolian and fluvial records, exposure dating (especially ^{10}Be and ^{36}Cl) of moraines; and U/Th dating of speleothems and corals. All ages are presented as calendar years ka (thousand years) for consistency.

3. Synthesis

During the first time slice (late MIS 3), we compare climatic conditions to modern mean annual temperature and precipitation. Other time slices are described with reference to the preceding time slice. The age boundaries of these time slices are chosen to coincide with the largest changes recorded in the climate proxy records. We refer here to effective precipitation, which is influenced not only by total precipitation, but also evaporation, wind and vegetation cover. The simplified patterns of temperature and effective precipitation between regions and through time are shown in Figure 4 a-h.

35-32 ka – late Marine Isotope Stage 3

Figure 4ai,ii shows climatic conditions during the period 35-32 ka in comparison to the present day, with cooler climate in the NE tropics, arid interior and lower latitudes of the Southern Ocean, and transient or variable conditions in the other regions. Wet conditions are evident throughout most of the region due to decreased evaporation under cooler than present conditions (Galloway, 1965; Bowler and Wasson, 1984).

Leading up to the last glacial period, much of inland Australia had large permanent lakes in catchments which are presently dry or ephemeral (Bowler et al., 1976; 2012; Cohen et al., 2011, 2012; Fitzsimmons et al., this volume). Relatively high lake stands prevailed in both the temperate (Coventry, 1976; Bowler and Hamada, 1971) and tropical zones (Veth et al., 2009). This period witnessed a peak in fluvial activity across the continent, both in the north (Nanson et al., 2008; Veth et al., 2009) and in the temperate zone, particularly in the tributaries of the Murray-Darling Basin (MDB) (Bowler and Wasson, 1984; Page et al., 1996; Page and Nanson, 1996; Kemp and Rhodes, 2010). Cool and humid conditions are inferred, with increased effective precipitation (Kemp and Rhodes, 2010). However, it is important to note that although this time was wet in comparison to present, it was dwarfed by the wet intervals of Marine Isotope Stage 5 and 4 (Nanson et al., 1992, Cohen et al., 2011, 2012). Cooler conditions in the arid zone (Miller et al., 1997) and wetter conditions in the north of the continent around this time are also suggested (Wallis, 2001; van der Kaars and De Deckker, 2002). Ice core records from Antarctica show millennial-scale variability through this period, corresponding to temperature shifts of up to 2^{o}C (EPICA, 2006). These perturbations are also recorded in the Murray Canyons offshore South Australia (De Deckker et al., 2012) and in deep sea cores from the Southern Ocean (Barrows et al., 2007a; Armand and Leventer, 2010).

32-22 ka - early glacial period

Cooling began around 32 ka, with glacier advance in the Snowy Mountains (Barrows et al., 2001). Glacial conditions may have commenced in New Guinea by ~28 ka (Prentice et al., 2005). A peak in dune activity in both the central arid region (Fitzsimmons et al., this issue), and increased dust transport, most likely from the increasingly arid MDB, to the subtropics along the east coast (Petherick et al., 2008; 2009) is interpreted as indicating cooler, drier conditions. Significant vegetation change commenced, with the expansion of herbs and grasses at the expense of arboreal taxa, in southeastern Australia (Dodson, 1975; Colhoun et al., 1982; Colhoun, 2000; Kershaw et al., 2007). A drying trend was also present in the NE Indian Ocean region from 32 ka, noted in the speleothems of Flores (Lewis et al., 2011) and pollen records off Java (van der Kaars et al., 2010) and also off Cape Range, Western

Australia (van der Kaars and De Deckker, 2002). However relatively wet conditions persisted in Sumatra and West Papua (Newsome and Flenley, 1988; Hope and Tulip, 1994; van der Kaars et al., 2012).

After slightly warmer conditions in Antarctica centred around 30 ka, cooling commenced in earnest and near full glacial temperatures were achieved around 27 ka (EPICA, 2006). Strong cooling had also occurred in the Southern Ocean by ~26 ka, accompanied by a northward movement of the subtropical front by some 3-5^{o}, bringing cool, sub-Antarctic waters to the south of Australia (De Deckker et al., 2012; Bostock et al., this volume). The wet conditions of the temperate region persisted until ~25 ka, followed by drying, bringing about a dominance of grass and herb vegetation (Petherick et al., this volume). Dune activity increased within the central arid zone (e.g. Hesse et al., 2004; Fitzsimmons et al., 2007) and expanded at the desert margins during the early glacial period (Lomax et al., 2011; Fitzsimmons et al., this volume). In the tropics, dry, stable conditions are evident in the vegetation and speleothem records of Indonesia (Reeves et al., this volume). Both the lake level and the speleothem records from the tropics show a close correlation between effective precipitation and regional insolation through this period. There is no evidence for a strong monsoon penetrating the Australian mainland at this time (Devriendt, 2011; Lewis et al., 2011; Reeves et al., this volume).

There is some evidence for a short-lived expansion of rainforest taxa in north Queensland ~26-24 ka (Moss and Kershaw, 2007). This coincides with lower, oscillating lake levels in the Willandra system (Bowler et al., 2012), an increase in dust transport to the subtropics (Petherick et al., 2008), abrupt drying and increased variability in the Borneo speleothem record 26.5-25 ka (Partin et al., 2007) and warming offshore South Australia (Calvo et al.,

2007) and the Southern Ocean (Armand and Leventer, 2010). These combined factors point to a warming and drying trend in the earlier glacial period (Fig. 4bi,ii). Although there are few records from this time period that have strong enough chronological control to resolve how widespread this climatic anomaly was, in most places conditions prior to 26 ka were substantially different to those prevailing after 24 ka, when the accelerated descent into the last glacial maximum (LGM) commenced. This period is consistent with the first interstadial of the Last Glacial Cold Period in New Zealand, coincident with the Kawakawa tephra (Barrell et al, this volume; Vandergoes et al., this volume) but earlier than the brief warming in Antarctica at ~24-23 ka (EPICA, 2006).

22-18 ka – Last Glacial Maximum

The LGM, globally the interval of greatest ice sheet extent over the last full glacial cycle, encompasses both maximum glacial extent and SST minima across the region (Fig. 4ci,ii). Sea surface temperature was typically in the order of 3-6^{o}C cooler than present in the Southern Ocean (Barrows et al., 2007a). This was combined with an expansion of winter sea-ice as far north as 55^{o}S (Gersonde et al., 2005; Armand and Levanter, 2010). Cooling in the tropical oceans was less pronounced, with SST 1-3^{o}C cooler than present in the NW and 1-2.5^{o}C cooler in the NE (e.g. Stott et al., 2002; Visser et al., 2003; Barrows and Juggins, 2005; Dunbar and Dickens, 2003; Jorry et al., 2008. These cooler ocean temperatures coincided with a constriction of both the ITF and the IPWP (Reeves et al., this volume). Some of the greatest differences in SST were felt along the northwest and south coast of Australia, attributed to the weakening of the Leeuwin Current and northward movement of the subtropical front (Petherick et al., this volume).

The most significant glacier advance is centred on ~19 ka, both in Tasmania and the Snowy Mountains of southeastern Australia (Barrows et al., 2001; 2002; Kiernan et al., 2004; Mackintosh et al., 2006). Peak glacial extent occurred at the same time (20.3-19.4 ka) at Mt Giluwe in New Guinea (Barrows et al., 2011). Estimates of the maximum expanse of ice are 15 km^2 in the Snowy Mountains (Barrows et al., 2001), 1085 km^2 in Tasmania (Colhoun, 1996) and 3400 km^2 in New Guinea (Prentice et al., 2011). Cooling in the upland areas is postulated to be as much as 11^{o}C below present in New Guinea (Hope, 2009), 9^{o}C in south-eastern Australia (Galloway, 1965) and between ~6.5- 4.2^{o}C in Tasmania (Colhoun, 1985; Fletcher and Thomas, 2010) with a peak in periglacial activity at ~22 ka (Barrows et al., 2004). These estimates are based onpollen recordssnowlines and lower limits of periglacial solifluction. The extent of cooling is generally at odds with the moderate cooling observed in the oceans (Barrows et al., 2000) and highlights their buffering capacity, except where movement of the oceanic fronts brought about significant cooling (e.g. Calvo et al., 2007; Bostock et al., this issue). Some of the differences may also be due to the accuracy and precision of the temperature estimates. For example, the snowline is affected by both temperature and precipitation, solifluction can occur below the treeline, and vegetation is affected by many variables including precipitation and carbon dioxide levels.

Both cooler and drier conditions are evident in the vegetation records from the Indonesian region and the east coast of Australia, including Tasmania, with a reduction in woody taxa, in particular rainforest species, and expansion of grasses and herbs (Petherick et al., this volume, Reeves et al., this volume). Estimates of temperature from vegetation changes in the lowlands of Tasmania are in the order of 4.2^{o}C below present (Fletcher and Thomas, 2010). However, some minor rainforest persisted in refugia in the Australian subtropics (Donders et al., 2006; Petherick et al., 2008) and woodland and heath in the southwest (Dodson, 2001).

Average air temperature in the interior decreased significantly to as low as 9^{o}C below present, determined from amino-acid racemisation of emu eggshells (Miller et al., 1997). Dune activity intensified in the arid core, but was most noticeable in the semi-arid desert margins (Fitzsimmons et al., 2007) where it was better preserved (Lomax et al., 2011; Fitzsimmons et al., this volume) and into the present-day temperate zone (Hesse et al., 2003; Duller and Augustinus, 2006; Gardner et al., 2006), indicating either less vegetation, windier conditions and/or increased sediment availability with the potential expansion of the arid zone beyond its modern extent. Lakes and rivers in the northern, monsoon-influenced arid zone generally experienced lower levels and reduced flow respectively, due to an absence of the monsoon in northern Australia during the LGM (Fitzsimmons et al., this volume).

By contrast, rainfall-fed lakes in the southeastern Australian highlands, record high or oscillating levels during the LGM (Coventry, 1976; Page et al., 1994; Bowler et al., 2012) . This is coeval with significant fluvial activity in the rivers of the Riverine Plain (Page et al., 1996; Kemp and Rhodes, 2010). These conditions are believed to reflect increased runoff from seasonal snow melt and reduced vegetation within the catchments, and do not preclude a drier climatic phase (Dosseto et al., 2010; Kemp and Rhodes, 2010; Bowler et al. 2012; Fitzsimmons et al., this volume). Evidence for at least periodically wet conditions exists also in the tropical region (Nott and Price, 1999; Reeves et al., 2007; Croke et al., 2010). In addition, there is evidence for increased precipitation in South Australia, from speleothem records from Naracoorte (Ayliffe et al., 1998) as well as recurrent large floods in silt-rich floodplains accumulating in the southern Flinders Ranges (Haberlah et al., 2010), attributed to a more northerly penetration of the westerlies.

18-15 ka early deglacial period

The first evidence of warming following the LGM comes from the Coral Sea, commencing around 20 ka (Tachikawa et al., 2009). Deglacial warming in Antarctica commenced around 19 ka (Pedro et al., 2011), followed by the retreat of sea ice and increase in SST in the Southern Ocean, which occurred rapidly between 18 and 15 ka (Barrows et al., 2007a), accompanied by a dramatic increase in opal flux and atmospheric CO_2 (Armand and Levanter, 2010; Bostock et al., this volume). This warming resulted in the STF moving back to a more southerly position (Sikes et al., 2009). SST also increased off the east coast of Australia although lagged the warming in the south (Weaver et al., 2003; Petherick et al., this volume). Other than a brief decrease around 18 ka (Yokoyama et al., 2001), warming in the Indian Ocean lagged behind the Pacific Ocean, not commencing until ~15 ka (Martinez et al., 1999;). This time interval also saw glacial retreat in the Snowy Mountains around 16.8 ka (Barrows et al., 2001) and in Tasmania and New Guinea from ~18 ka, the latter becoming becoming more rapid after 15.4 ka (Barrows et al., 2011).

The rate of response in vegetation change during the deglacial period varied greatly across the region (Fig. 4di,ii). Wetter and warmer conditions are first noted in Indonesia around 17 ka and NW Australia ~15 ka with more prolonged recovery in New Guinea and NE Australia (Reeves et al., this volume). Dry conditions persisted in Borneo, as recorded in speleothems, until ~15 ka, after which time there was a rapid shift to wetter conditions (Partin et al., 2007). Vegetation records from the temperate region showed a gradual response to warmer and wetter conditions, with an increase in arboreal taxa from ~15 ka (Petherick et al., this volume).

Although there was an expansion in Lake Carpentaria from 18 ka, the northern plunge pools were considered inactive and Lake Gregory may have been dry (Reeves et al., this volume). Other lakes in the arid interior and temperate regions were also low. High lake levels are recorded in Lake Frome from 18-16 ka, although dry conditions at Lake Eyre persisted (De Deckker et al., 2010; Cohen et al., 2011, 2012). However, fluvial discharge increased in the lower MDB and in coastal rivers of NSW, possibly in response to snow-melt (Page et al., 1996; 2009; Nanson et al., 2003). Although limited dune activity persisted throughout this period within the semi/arid zone (Fitzsimmons et al., 2007; Lomax et al., 2011), conditions in the southern part of the continent are likely to have become relatively more stable and humid.

15-12 ka late deglacial period

Climate for the period from 15-12 ka was highly varied, with conflicting climatic responses across the Australian region (Fig 4ei,ii). Whilst glacier retreat was complete on the continent (Barrows et al., 2001, 2002), the re-advance of sea-ice and decrease in SST and atmospheric CO_2 in the Southern Ocean occurred between 14.5 and 13 ka during the Antarctic Cold Reversal (Pedro et al., 2011; Bostock et al., this volume). This resulted in a northward movement of the STF, again close to the southern Australian coast and suggests a lowering of SST offshore South Australia (Calvo et al., 2007). Wetter conditions are evident in western Tasmania around this time (14-11.7 ka; Fletcher and Moreno, 2011) and active river migration in the temperate zone continued throughout this period, perhaps due to snow melt, with evidence of increased precipitation in the lower MDB from 13.5 ka (Gingele et al., 2007; Petherick et al., this volume).

In contrast, SST warming commenced at 15 ka in the tropical north-west of Australia, coincident with the initial flooding of the Sunda Shelf (Bard, 1990). Warmer and wetter conditions prevailed throughout the north, with the re-invigoration of the monsoon, bringing freshwater to the Gulf of Carpentaria and recovery of woodland and forest in PNG and NE Australia (Reeves et al., this volume). Increased precipitation associated with the monsoon from ~14 ka is evident in lake level transgression in the Gregory Lakes basin (Wyrwoll and Miller 2001) and by ~12 ka at Lake Eyre (Magee et al., 2004). The speleothem record of Borneo shows a rapid decrease in $\delta^{18}O$, reflecting an increase in precipitation after 15 ka, with a reversal to drier conditions between 13.3-12.3 ka (Partin et al., 2007). Desert dune records across both the arid zone and semi-arid desert margins of the MDB show aeolian activity persisted or possibly stabilised, during this period (Fitzsimmons et al. 2007; this volume; Lomax et al., 2011).

12-8 ka early Holocene

Maximum warming in Antarctica occurred at 11.8 ka (Pedro et al., 2011). The peak in SSTs of the Southern Ocean (e.g. Barrows et al., 2007a) and ambient air temperature are coincident with the most poleward migration of the STF (Bostock et al., this volume). SST throughout the temperate and tropical regions approached modern values around 11-9 ka (Petherick et al.; Reeves et al, this volume), and saw a re-activation of the Leeuwin current extending to South Australia (Calvo et al., 2007; De Deckker et al., 2012). By 9.8 ka glaciers in New Guinea had disappeared (Barrows et al., 2011), most of the shallow shelf seas were re-established, and the land:sea ratio, particularly in the tropics, approached modern conditions. The mangrove swamps of the north and estuaries of the eastern seaboard had formed and growth of the Great Barrier Reef was re-initiated (Reeves et al., this volume).

Increasingly wet conditions in Borneo and Flores from 12 and 11 ka coincided with warming of the South China Sea (Partin et al., 2007; Griffiths et al, this volume) (Fig 4fi.ii). Wet conditions are also evidenced by the active plunge pools in the north of Australia and vegetation similar to modern in Indonesia by 11 ka and NE Australia by 9 ka, with the establishment of rainforest taxa (Reeves et al., this volume). Interestingly, although the early Holocene fluvial records of the Lake Eyre Basin see a peak in discharge compared to present, those of the Riverine Plain of the lower MDB show a significant decline compared with the deglacial, with a shift from bedload to suspended-load sedimentation, caused primarily by a reduction in discharge (Fitzsimmons et al., this volume; Petherick et al., this volume).

Vegetation records from across the temperate region east of the Great Dividing Range show a shift at ~12-11 ka to include a greater representation of rainforest taxa, indicating both warmer and wetter conditions than previous (Petherick et al., this volume). Lake levels at Keilambete and George also increased after 11 ka (Wilkins et al., in press; Fitzsimmons and Barrows, 2010) and peaked after 9.6 ka in the Wimmera (Kemp et al., 2012).

8-5 ka mid-Holocene

The mid-Holocene period represents maximum temperature in terrestrial records throughout the Australasian region, although it was expressed in different places at different times (Fig 4gi,ii). By 8 ka both sea level and SST had reached essentially modern conditions (Lewis et al., this volume) and after a decline from the high temperatures of the early Holocene, Antarctic temperatures stabilised (EPICA, 2004). In the north, the thermal maximum of the IPWP was achieved by 6.8-5.5 ka (Abram et al., 2009). Peak wet conditions occurred on Flores by 8-6 ka, but only by ~5 ka in Borneo (Reeves et al., this volume). There is an increase in composite charcoal from the north at 8 ka, associated with a switch from grass-dominant to rainforest taxa from the northeast (Mooney et al., 2011; Moss and Kershaw, 2007).

Offshore southern Australia experienced a peak in SST at ~6 ka of 18-21^{o}C (Calvo et al., 2007). An increased representation of rainforest taxa, peak in representation in the macro-charcoal record especially in the north (Mooney et al., 2011) and a peak in discharge in the fluvial records of the east coast support both warmer and wetter conditions centred around 6 to 5.5 ka (Petherick et al., this volume). Dune activity persisted in the desert margins of the MDB (Lomax et al., 2011). Increasingly humid conditions from 7 to 5 ka are evident around the Flinders Ranges and Lake Frome (Fitzsimmons et al., this volume). Both Lake George (Fitzsimmons and Barrows, 2010) and the lakes of western Victoria (Wilkins et al., in press) record high lake conditions, indicating peak effective precipitation, in the early-mid-Holocene. Although the peak warm and relatively wet conditions throughout the Australian region all occur sometime during this mid-Holocene period, the east coast appears to lag the more southerly sites.

5-0 ka late-Holocene

Increasing variability with enhanced drier conditions was the overall characteristic of much of the Australasian region through the Late Holocene, indicative of ENSO in El Niño mode (Fig 4hi,ii). This is seen in the coral (e.g. Tudhope et al., 2001; Gagan et al., 2004; McGregor and Gagan, 2004) and speleothem (Griffiths et al., 2009, 2010 a,b) records in the north, and reactivation of dunes and dust deposits in the northern, interior and temperate zones (e.g. Shulmeister and Lees, 1995; Fitzsimmons et al., 2007; Marx et al., 2009, 2011; Fitzsimmons and Barrows, 2010). Fluvial activity decreased (e.g. Cohen and Nanson, 2007), with lower lake levels across much of the continent (e.g. De Deckker, 1982; Wilkins et al. in press). In contrast, the southwest was relatively consistently warm and moist (e.g. Gouramanis et al., 2012). An increasing number of records across the continent are showing variability in the order of 1000-2000 years, attributed to El Niño in the north (e.g. McGregor and Gagan, 2004; Turney et al., 2004) and changes in the westerly wind belt in the south (e.g. Moros et al., 2009; Kemp et al., 2012; Wilkins et al., in press). Although there have been recent records which focus particularly on decadal and finer resolution of the last 2000 years in the Australasian region, these are covered in other reviews (e.g. Neukom and Gergis, 2011).

4. Discussion

The changes in climate we have documented over the last 35 kyr highlight a complex response across a continent that spans a vast array of climatic zones. The latitudinal range, together with its extensive interior and fringing mountainous areas, results in the interplay of all the major circulation systems of the low and mid latitudes over the continent. The consequences of this are often diachronous, and sometimes contradictory, responses of the landscape, vegetation and sea surface when sites are compared across this range. This patchwork response is in strong contrast to the Northern Atlantic Ocean, where huge changes in heat flux from the ocean act to synchronise the climate of Europe through time. A consequence of this is that the climate signal preserved in the ice cores of Greenland can act as a template for the stratigraphy of a large proportion of the European region (e.g. Lowe et al., 2001). No such obvious simple approach exists for Australia that can adequately describe or characterise the changes that take place. Both the magnitude and pattern of climate change vary meridionally across the Australian continent. At the sea surface, the highest magnitude temperature changes occur in the high latitudes (cf. Barrows et al., 2007b) and on land the highest changes tend to be in the highest altitudes.

A major limitation on the characterisation of Australian climate change is the lack of quantitative proxies for both temperature and precipitation across a large proportion of the continent, with exceptions being in the far north (e.g. Lewis et al., 2011; Griffiths et al., this volume) and far south (e.g. Rees et al., 2008; Fletcher and Thomas, 2010), confounded by limitations in the ability to date events precisely. Although chronologies are constantly improving, aided by increased luminescence, exposure age and radiocarbon data sets, the interpretation of many records remains difficult. Landscape responses such as lake level change, river discharge and dune activity can be produced under a range of different climate scenarios, from variable combinations of temperature and precipitation. Knowledge of temperature change across the continent is consequently mostly guided by sea-surface temperature records from around the periphery of the continent.

The changes in temperature and effective precipitation recorded across Australia during the last 35 kyr suggest large scale reorganisation of the circulation systems and major shifts in climate zones. However, current evidence indicates that the majority of these changes can be explained and accommodated using existing climate systems. Most of the changes in precipitation can be explained by the presence or absence of the monsoon in the northern half of the continent, and the meridional shifts in the westerlies and their intensity in the south. Temperature changes have not necessarily occurred in tandem with these moisture delivery changes, leading to effective precipitation at times being out of phase with major global temperature changes. This is illustrated for example during the deglaciation when both wet and arid periods are superimposed upon a dominantly warming trend.

The onset of glacial-magnitude cooling is recorded first in the high latitudes of the Australian region in the Southern Ocean, as early as 40 ka (Barrows et al., 2007a). Vandergoes et al. (2005) suggested that the early onset of cooling in the high southern latitudes relates to a minimum in Southern Hemisphere summer insolation around 30-35 ka. However, this suggestion contrasts with the lock-step relationship of temperature between the hemispheres during full glacial conditions and the Holocene. The lag in cooling of the low latitudes resulted in a stronger meridional temperature gradient than at present, which is likely to have had a response in stronger westerly and trade wind systems, delivering more water to inland lake and river systems. However, the Asian monsoon probably migrated northward between ~40-35 ka, removing the major source of inflow for the monsoon dominated north and Lake Eyre basin (e.g. Magee et al., 2004).

High resolution SST records indicate one or more interstadial periods around 30 ka (e.g. Barrows et al., 2007a), which also occur in ice core records from Antarctica (e.g., EPICA, 2006), but temperature remained well below Holocene levels. Most records indicate steady cooling from 30-20 ka. However, there is a brief reversal in the cooling trend during the period 26-24 ka recorded in Antarctic ice cores (Fig. 4bi,ii; EPICA, 2006), which produced warm and wetter conditions in the NE of Australia, and a minor interstadial in the southern temperate zone and the higher latitudes of the Southern Ocean (Petherick et al., this volume; Bostock et al., this volume). An interstadial is also recorded in Antarctic ice core records at the same time (EPICA, 2006). The cause of this event is unknown and it slightly leads Greenland Interstadial 2 in the North Atlantic (Lowe et al., 2001), indicating a relationship through the 'bipolar seesaw' (Broecker, 1998).

Rapid growth of the Northern Hemisphere ice sheets from 30-20 ka (Lambeck and Chappell, 2001) resulted in lower sea levels, which exposed the continental shelves around the Australian margin. Land bridges were formed between New Guinea and the north of Australia, forming the Sahul Shelf. The Indonesian archipelago was also largely connected, forming the Sunda Shelf. This not only altered ocean circulation, but also reduced the area of warm shallow seas. This had a major influence on the tropical climates of northern Australia because of a greater restriction on sources of tropical rainfall (Reeves et al., this volume). The exposure of the continental shelves also has an acute regional affect on moisture delivery in Bass Strait and Torres Strait.

The LGM heralded cooler, windier and drier conditions across most of the continent (Fig. 4ci,ii). The synchroneity of this response both at the sea surface (Barrows and Juggins, 2005) and on land, across a range of climatic zones, indicates a common regional control through greenhouse gas forcing of temperature. Carbon dioxide levels fell below 200 ppm before 25 ka, and remained between 180-200 ppm until 17 ka (Schmitt et al., 2012). This interval corresponds to maximum temperature depression and decline of woody taxa, despite a peak in summer insolation in the Southern Hemisphere. However, the period of maximum glaciation and periglacial activity in Australia and Papua New Guinea was only a brief interval within this period, centred at 19-22 ka, indicating that carbon dioxide cannot be the only control on surface temperature. Despite lower precipitation and presumably more northerly penetration of the westerlies, lower evaporation and increased seasonality of discharge (because of the timing of snowmelt), resulted in apparently wetter conditions in parts of inland Australia (Kemp and Rhodes, 2010; Fitzsimmons et al., this volume).

Warming in the deglacial period occurred much earlier than in the Northern Hemisphere, commencing as early as 19 ka in the south (Pedro et al., 2011; Bostock et al., this volume) and established by ~18-17 ka through much of the region (Turney et al., 2006b; Petherick et al., this volume; Reeves et al., this volume) (Fig. 4di, ii). Warming preceded increases in precipitation in most areas, particularly the northwest of Australia. The lag in this region may be attributed in part to the restricted influence of warmer tropical waters, and a less pronounced Leeuwin Current (Spooner et al., 2005, 2011; De Deckker et al., 2012). The presence of relatively high levels at Lake Frome, but dry Lake Eyre during the early deglacial period, in the absence of a monsoon may be attributed to a southerly moisture supply in the westerlies (Cohen et al., 2011, 2012).

Temperature reached near Holocene levels by 15 ka in many records. An increase in dune activity occurs at this time, perhaps as a result of temperature outpacing increases in precipitation. Because this is the last major period of dune activity, it may also be an indication of stabilisation and preservation of dunes, as much as an increase in aridity. The warming also coincides with the resumption of the monsoon increasing moisture delivery to the north (Fig. 4ei,ii) as low pressure systems developed over the continent.

With the exception of the marine records offshore South Australia (Calvo et al., 2007) the ACR (14 – 13.5 ka) is only weakly recorded in the southern part of the continent, mostly as a pause in the warming rather than as a distinct cooling event, producing wetter conditions in the temperate zone. No cooling is recorded during the Younger Dryas interval, as is expected under the 'bipolar seesaw' of interhemispheric heat exchange, nor is there a mechanism for causing a cooling at this time in the Australian region (Barrows et al., 2007b; De Deckker et al., 2012; Tibby, 2012).

Although the timing of response is varied, the early Holocene is a time of warmer and wetter conditions across the region. Conditions are wetter in the north associated with high sea-surface temperatures and moisture delivery enhanced because of flooding of the shallow shelf seas. The monsoon delivered more precipitation than at present, as a function of the pressure gradient established between northern Australia and Asia (Fig. 4fi,ii). The exception to this would be the Riverine Plain in the lower Murray-Darling Basin, where the trend is towards a decrease in river discharge. This is associated with a southerly migration of the westerly wind belt, restricting moisture supply west of the Great Dividing Range, coupled with increasing temperatures, decreasing snow cover and subsequent melt in the upland areas which fed the rivers.

The establishment of maximum sea level (~8-7.5 ka; Lewis et al, this volume) is coincident with many regions recording maximum temperature and precipitation. This period sees the reinitiation of the coral reefs on the Great Barrier Reef and the flooding and infilling of the tropical estuaries with sediment trapped by mangroves (Reeves et al., this volume). In some areas, particularly the temperate zone, vegetation change and establishment of woody taxa lags behind changes in climate and rise in CO_2, presumably owing to slow migration and re-establishment processes in forest ecosystems. Compared with the early Holocene, the trend at ~5 ka is for cooling conditions in the NE and Southern Ocean and drying in the interior (Fig. 4 gi,ii). The flooding of the broader continental shelf along the northeast coast, may have buffered the climatic response during this interval; however the driving mechanism remains unresolved. During the late Holocene,. a weakening of the Walker Circulation, and thus a weakening of the easterly trade winds over the tropical Pacific Ocean, has been attributed to a weaker Asian summer monsoon system due to decreasing insolation at northern mid-latitudes during the boreal summer (Zheng et al., 2008). This created conditions more favourable for the development of El Niño events, resulting in a progressive increase in El Niño frequency (Phipps and Brown, 2010). The El Niño mode starts to dominate from 7-5 ka (Rodbell et al., 1999; Tudhope et al., 2001; Moy et al., 2002; Gagan et al., 2004) and this has the effect of producing drier conditions in northern and eastern Australia.

The Late Holocene sees an increase in climatic variability, particularly with regards to moisture availability (Fig. 4 hi,ii) largely attributed to a stronger expression of El Niño (Moros et al., 2009). This is over-printed by the interaction with the Indian Ocean Dipole and a weakening of the monsoon (Shulmeister and Lees, 1995; Shulmeister, 1999). Thus, the late Holocene records warm conditions, with greater El Niño frequency, increased fire and extended droughts, alternating with shorter, wet La Niña events (e.g. Moy et al., 2002; Moros et al., 2009; Mooney et al., 2011), a pattern which persists today .

5. Further work

There are clear gaps in the types and distribution of data across the greater Australian region and areas where the chronology could be significantly improved. As a palaeoclimate community there is still a lack of consensus with regards to the hydrological balance and role of groundwater in key time intervals such as the LGM. Quantitative estimates are required on wind strength, evaporation and seasonality; parameters that are currently largely unconstrained. The Australian region still lacks sufficient sub-millennial-scale resolution terrestrial records to identify climatic leads and lags and there are many locations across the continent, such as the majority of Western Australia and the Northern Territory, coastal southeastern Australia and terrestrial southwest Queensland, where data is non-existent. Lastly, the role of humans in modifying the biophysical environment, which influences the interpretation of climate, is one area that is poorly known for much of the last 35 kyr.

The OZ-INTIMATE project aimed to consolidate the climate records from the broader Australian region over the past ~35 kyr and compare the response of each of the four major climatic regions to global climatic drivers. Future work will need to engage more actively with climate modelling to test the longevity of dynamical mechanisms of the climate system of the modern day and how these have manifested in the past. In addition, targeted palaeo-proxy data gathering must continue, with a focus on filling the spatial gaps outlined above, developing well-resolved chronologies and quantitative reconstructions. This combined approach will enhance our ability to predict the evolution of the climate system and the impacts of future changes on the climate of the Australian region.

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Figures

Figure 1. Map of average modern sea surface temperature and the major oceanographic features of the Australian region. The major currents are shown in solid lines: IPWP = Indo-Pacific Warm Pool, ITF = Indonesian Throughflow, SEC = South Equatorial Current, LC = Leeuwin Current (with dashed line showing extent during La Niña), EAC = East Australian Current, ACC = Antarctic Circumpolar Current. The mean positions of the major fronts are shown in dotted lines: TF = Tasman Front, STF = Subtropical Front, SAF = Subantarctic Front, PF = Polar Front.

Figure 2. Map showing the major features of the modern climate system of the Australian region. ITCZ = Inter-tropical Convergence Zone, ENSO = El Niño Southern Oscillation, IOD = Indian Ocean Dipole, SAM = Southern Annular Mode.

Figure 3. Map of the greater Australian region, with site localities (white dots) considered here in this study. The shadings are a stylistic representations of the four climate regions referred to as tropics (yellow terrestrial, purple marine), arid interior (orange), temperate (green terrestrial, royal blue offshore) and Southern Ocean (lower rectangle). The reader is directed to the other papers within this volume, which deal with each of these regions independently, for more information on the detailed boundaries (i.e. Reeves et al., Fitzsimmons et al., Petherick et al., Bostock et al.). The inset graphs show summer and winter insolation at a) 20^{o}S and b) 40^{o}S, respectively.

Figure 4. Schematic representation in the changes in temperature (i) and effective precipitation (ii) through time. Change here is considered as a trend in relation to the previous time slice, that is; are conditions hotter or wetter, for example, than the previous period. For temperature, red = hot, blue = cold, purple = no change. For precipitation, green = wet, orange = dry, yellow = no change. Grey indicates no data are available. Please refer to figure 2 for the division of the climate zones represented here by boxes, with reference to latitude and longitude. The time slices are a = c. 32 ka, b = c. 25 ka, c = c. 20 ka, d = c. 17 ka, e = c. 14 ka, f = c. 9 ka, g = c. 5 ka, h = c. 2 ka

Figure
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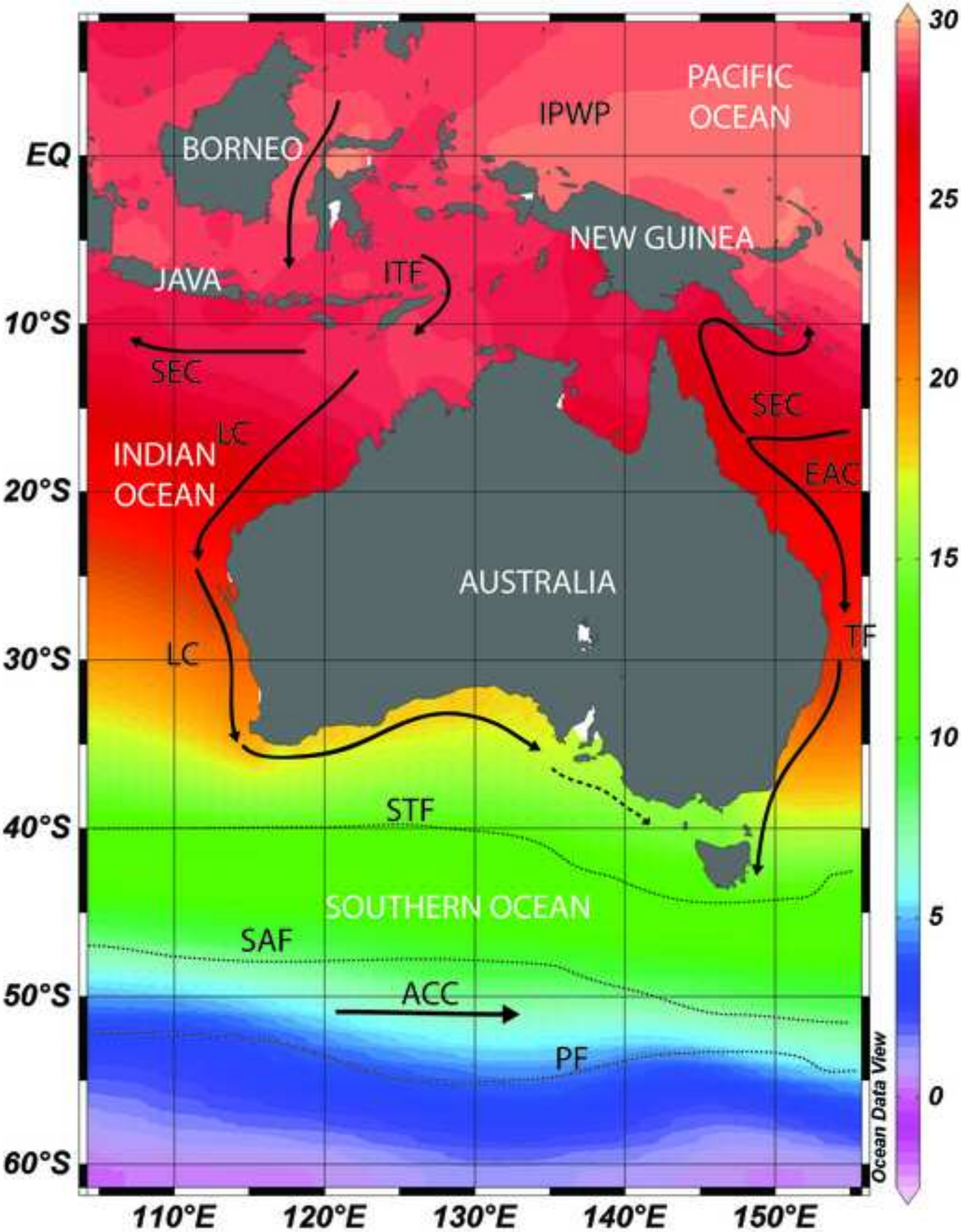

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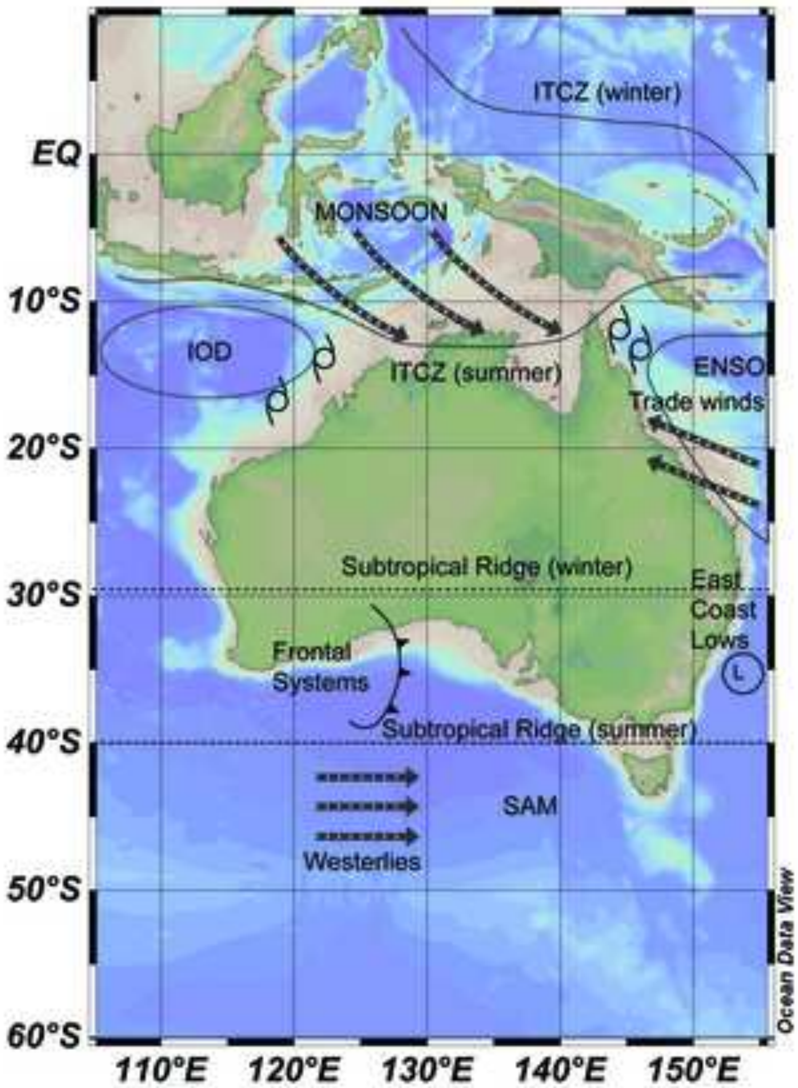

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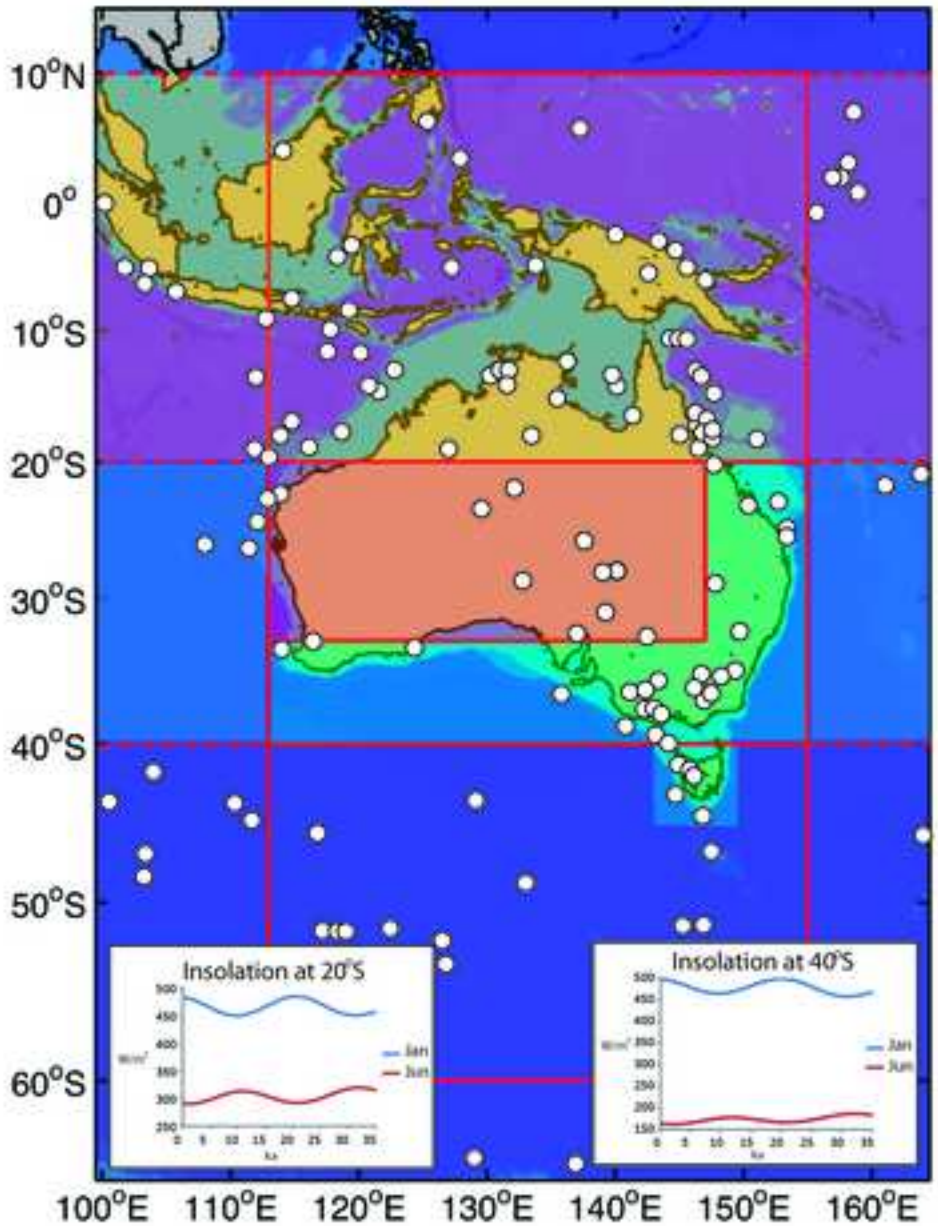

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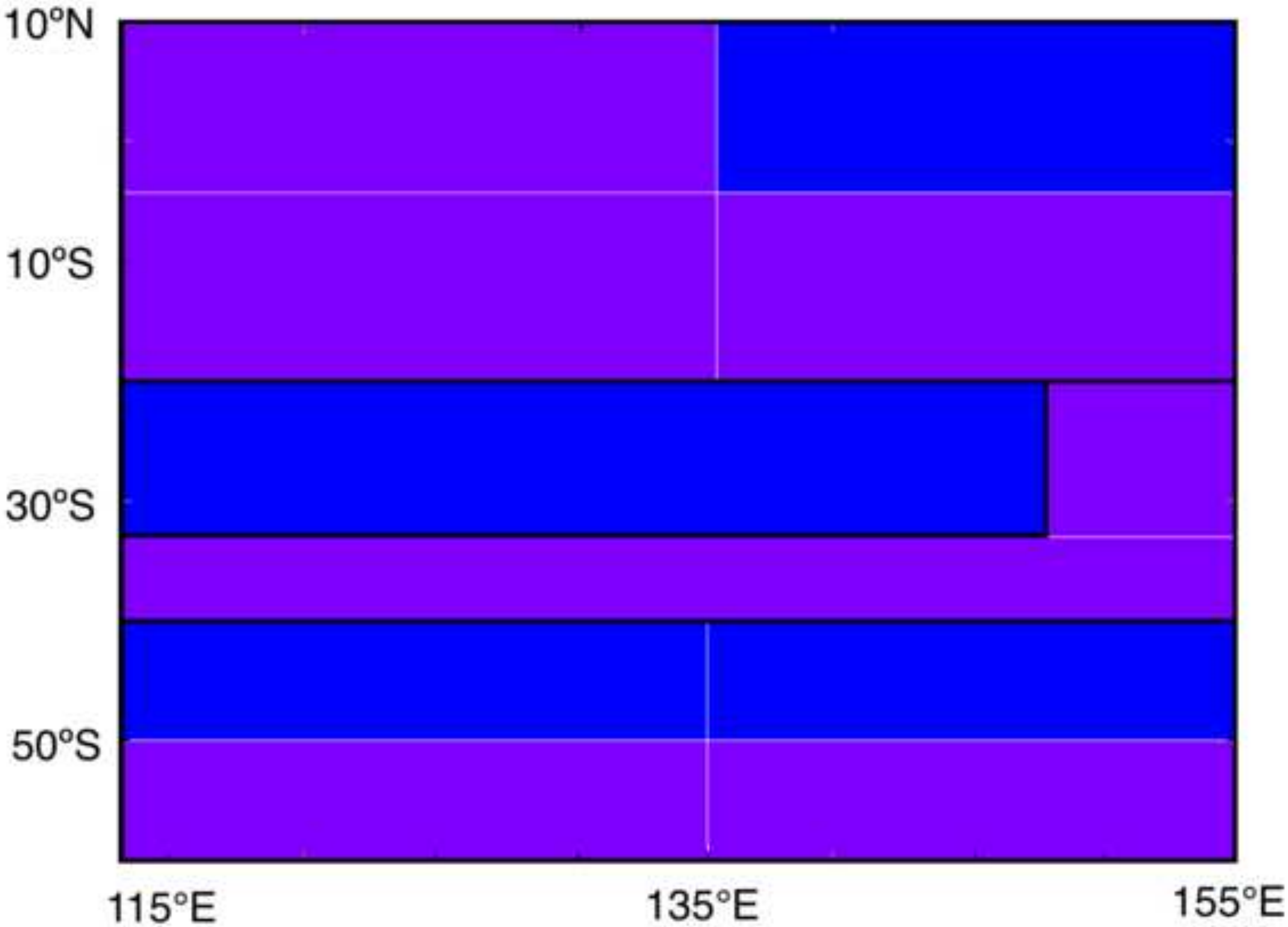

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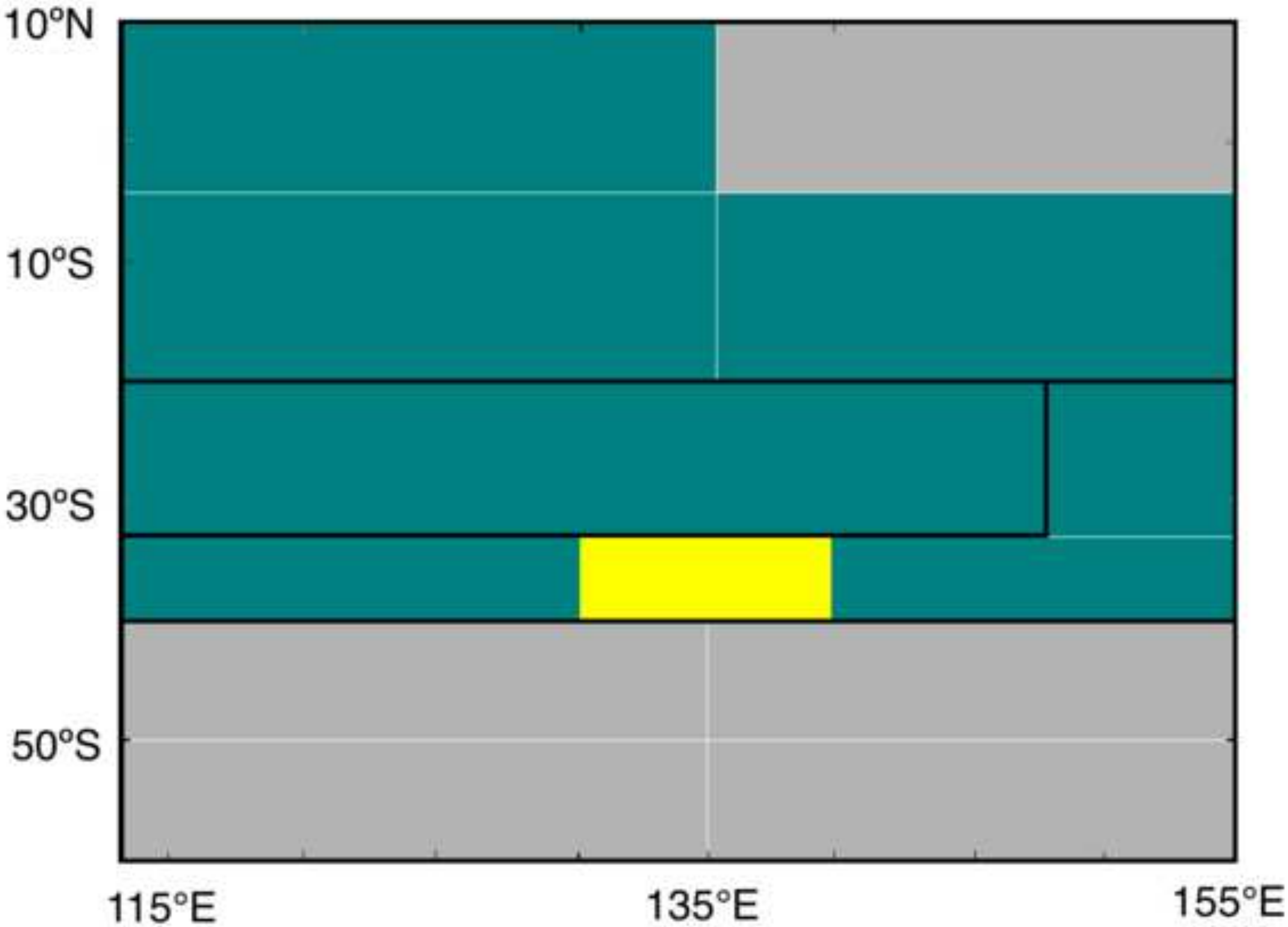

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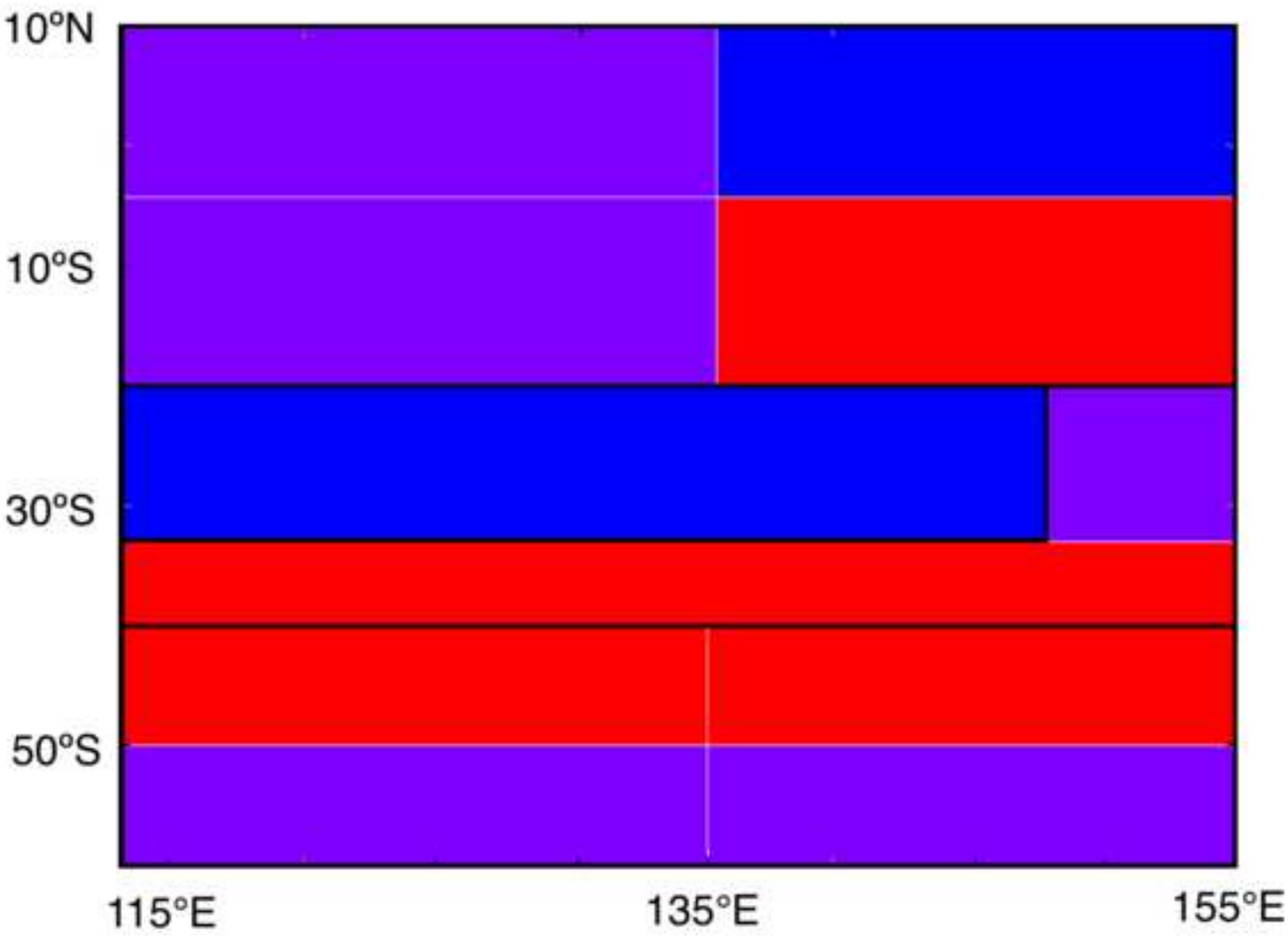

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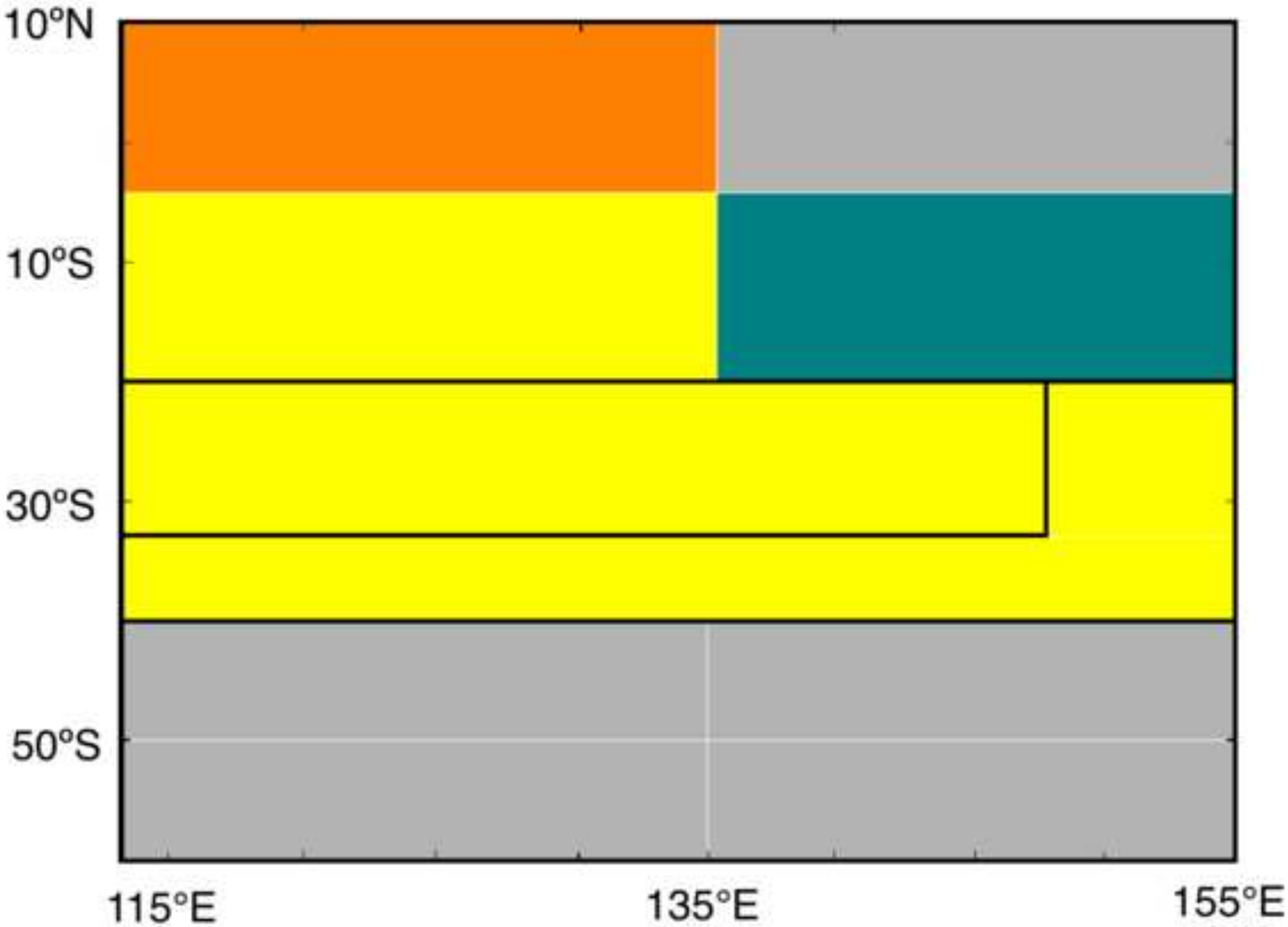

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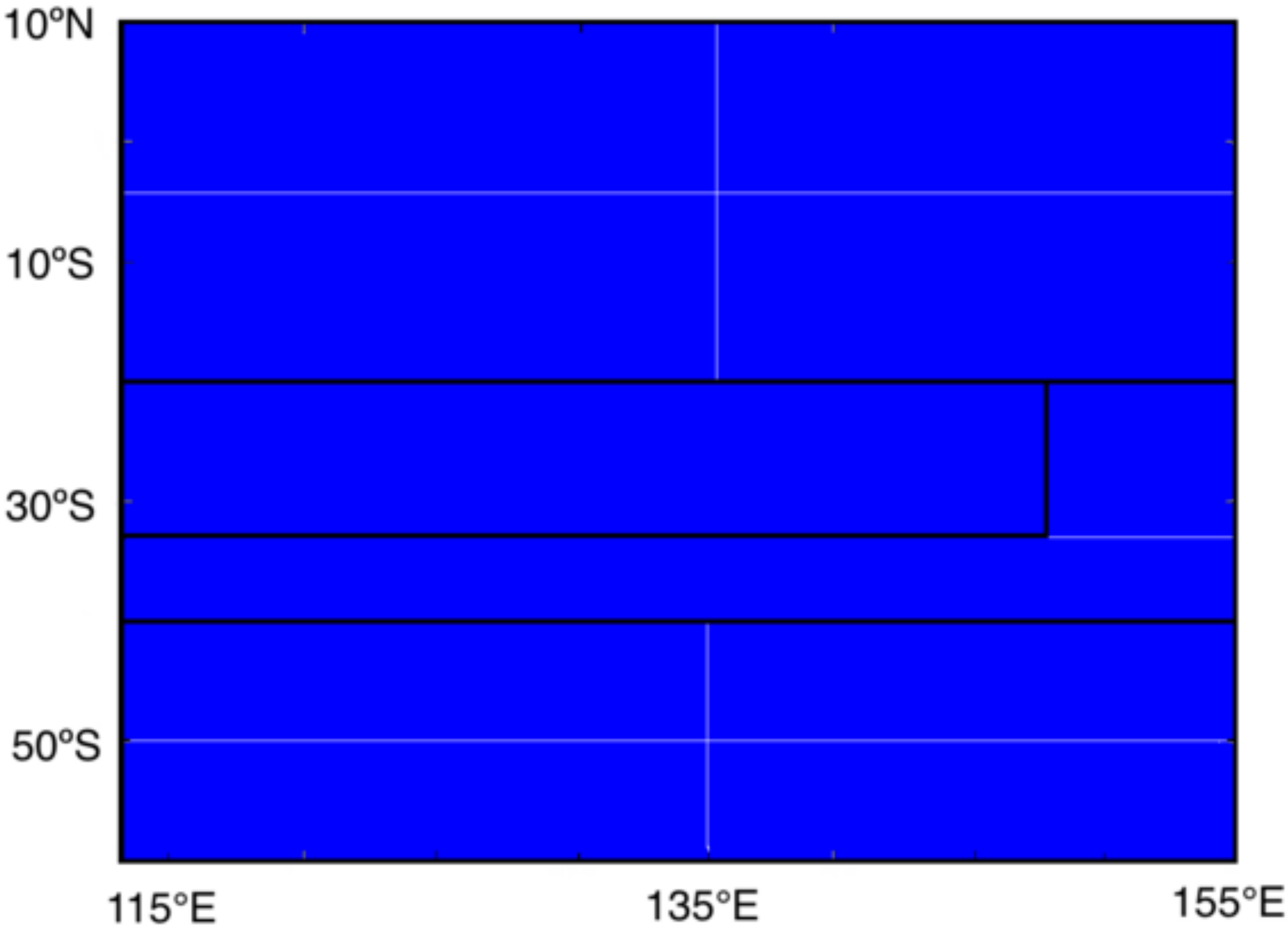

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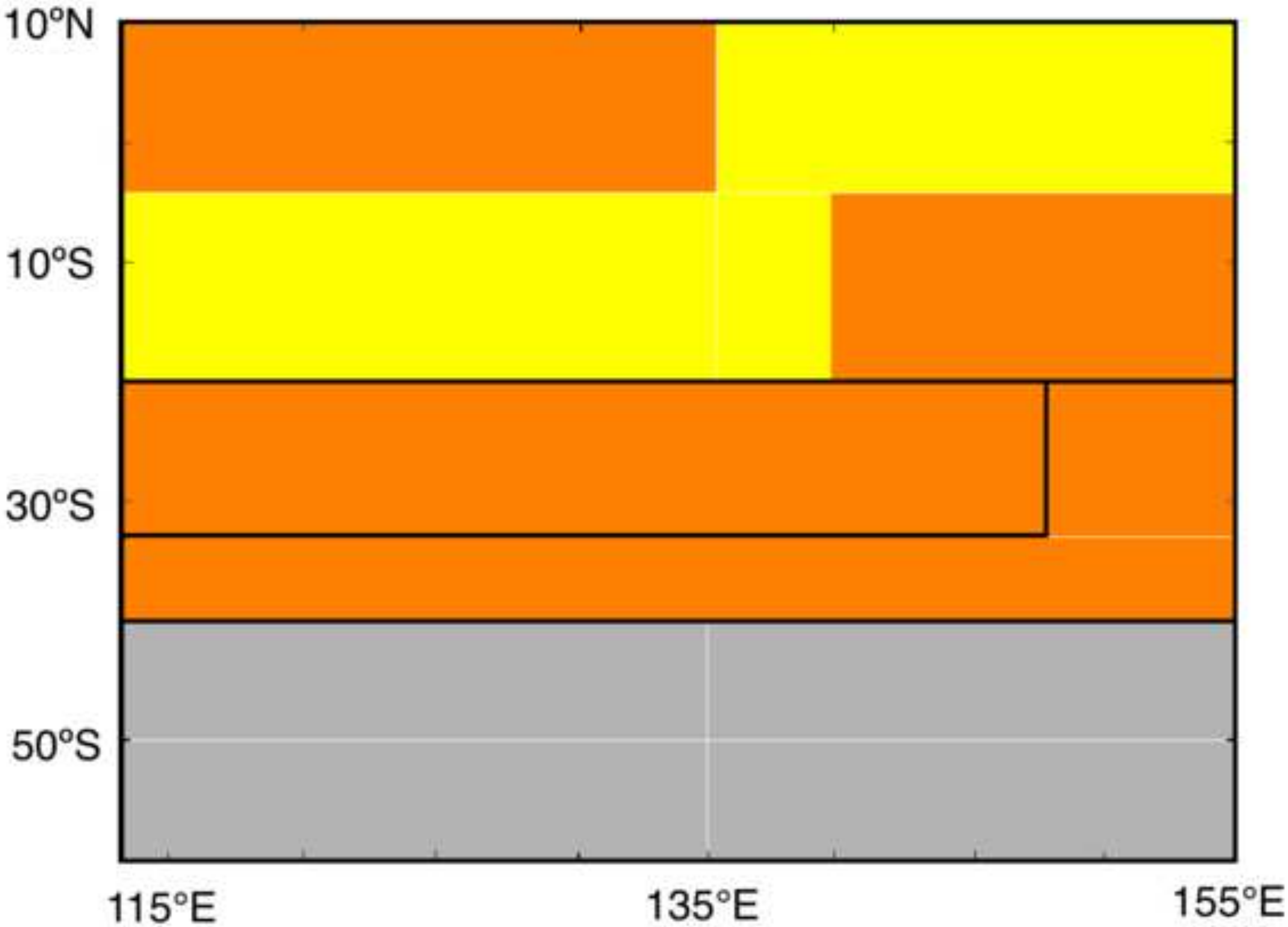

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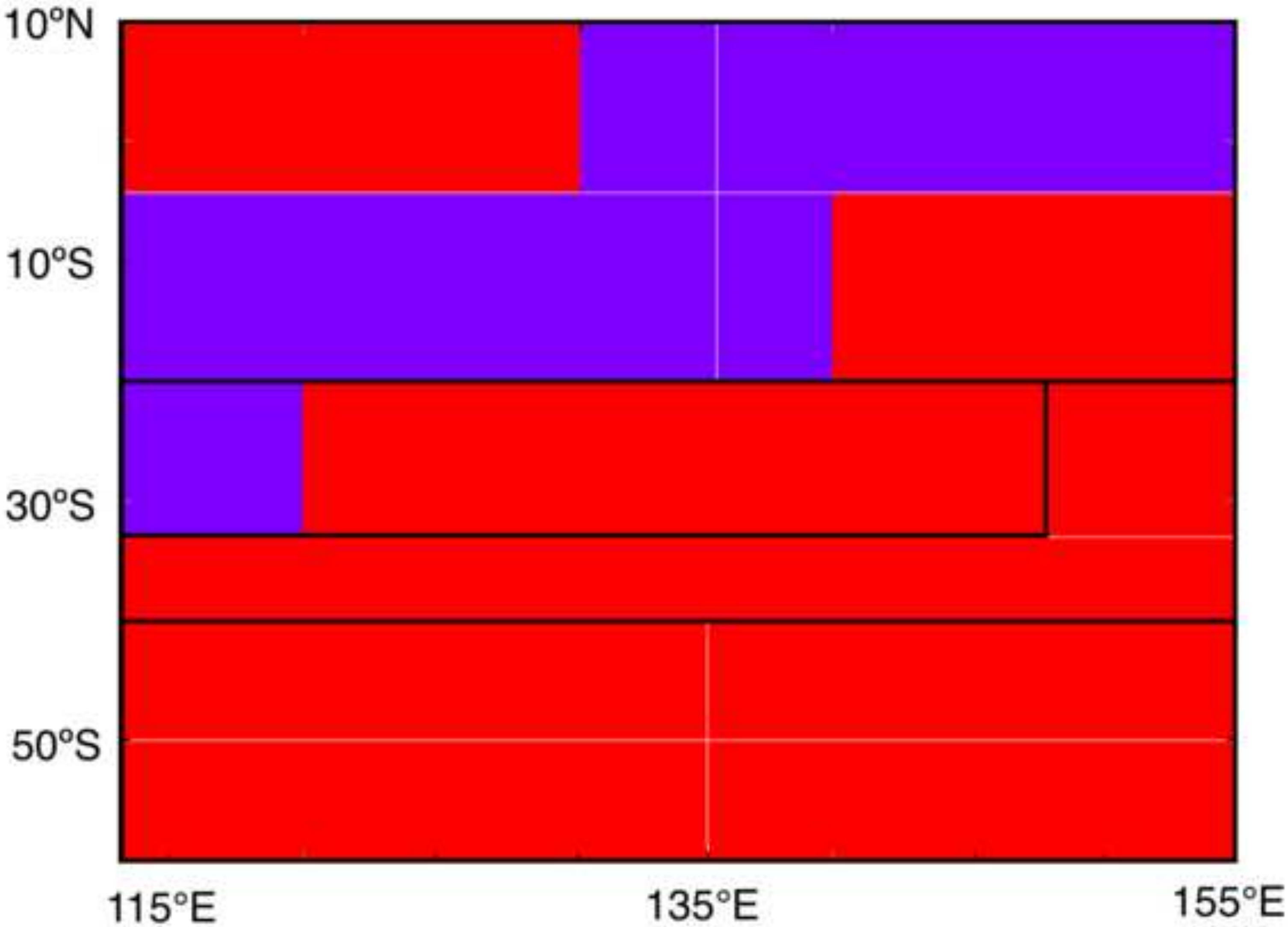

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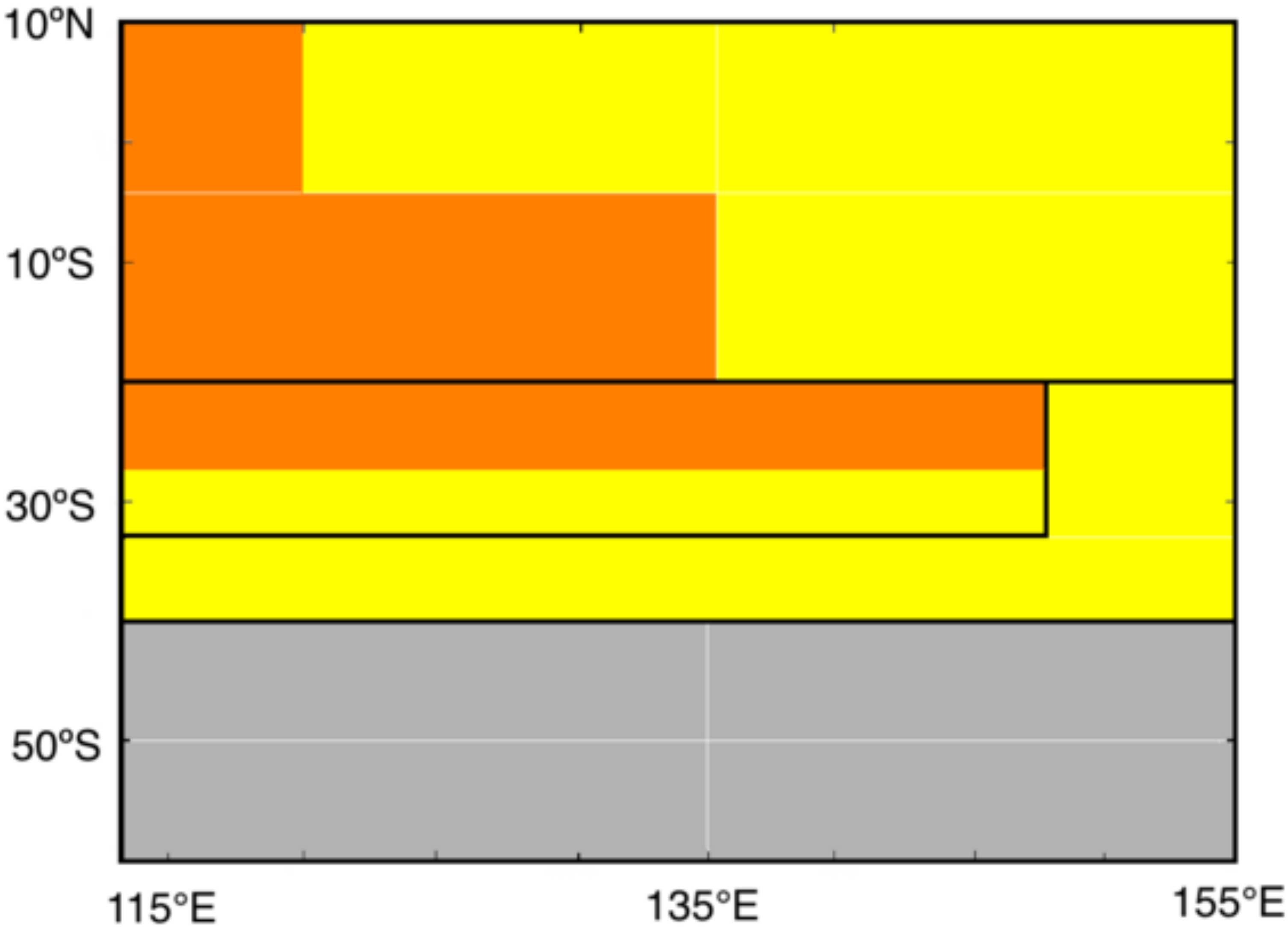

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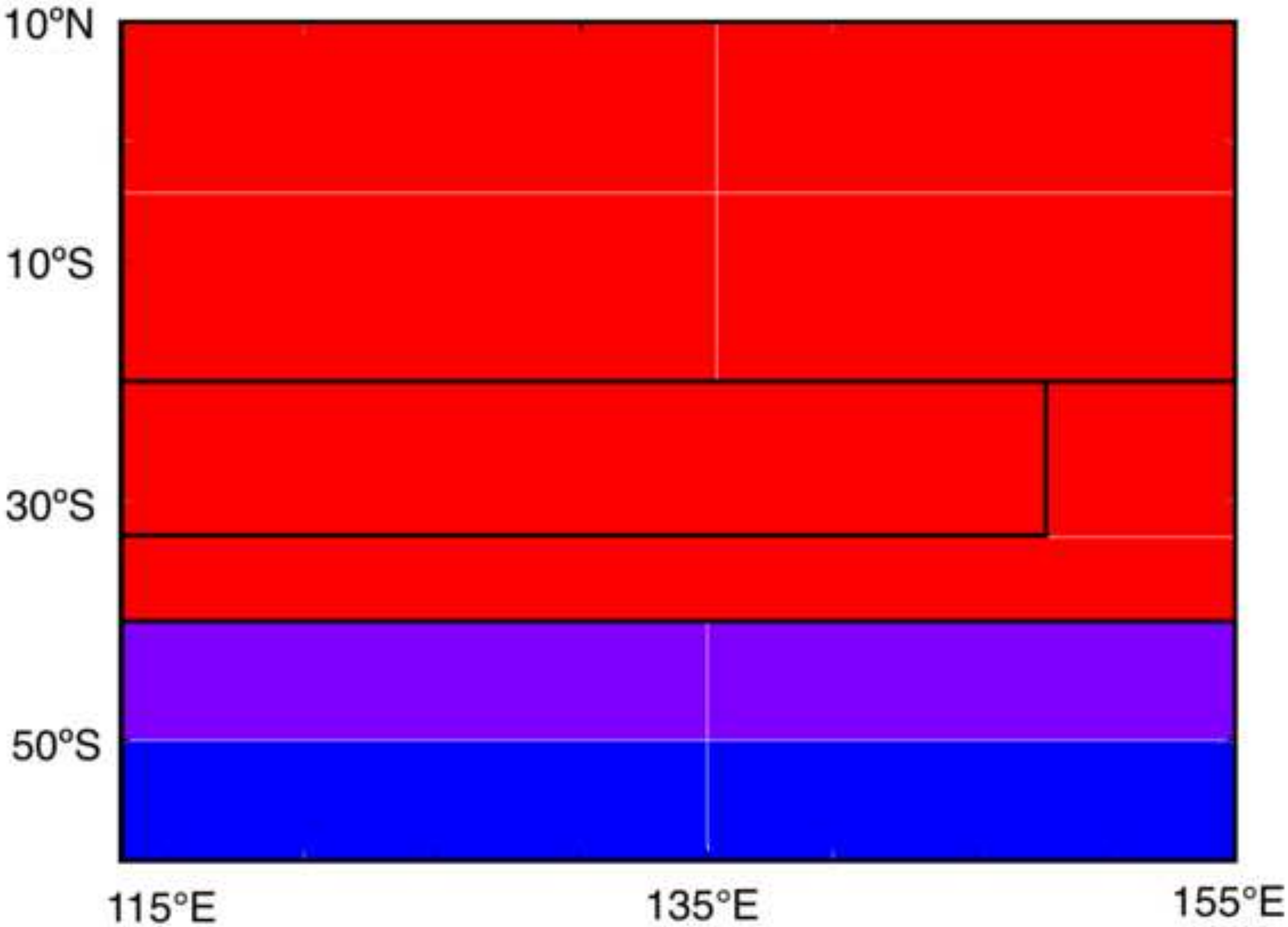

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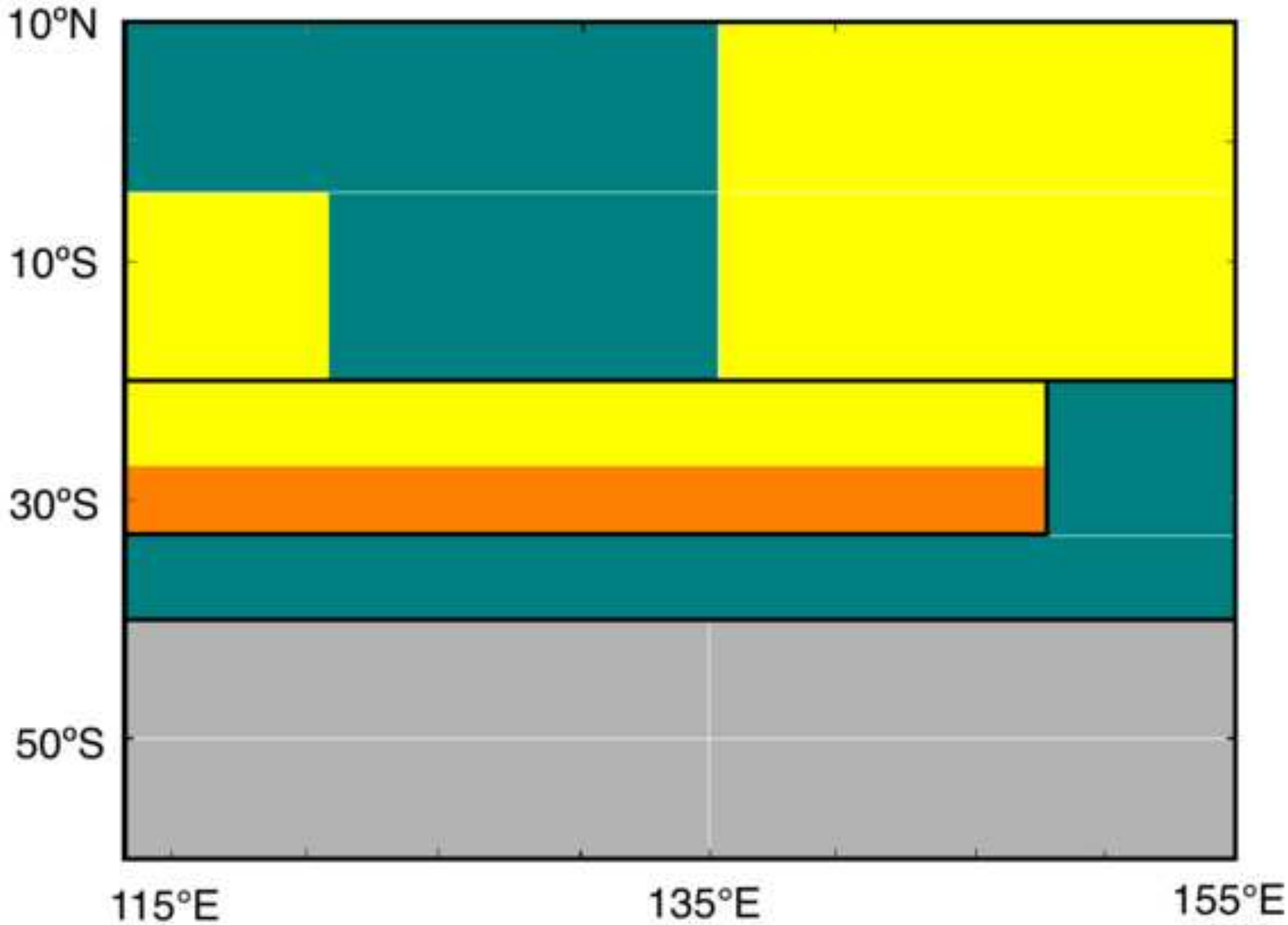

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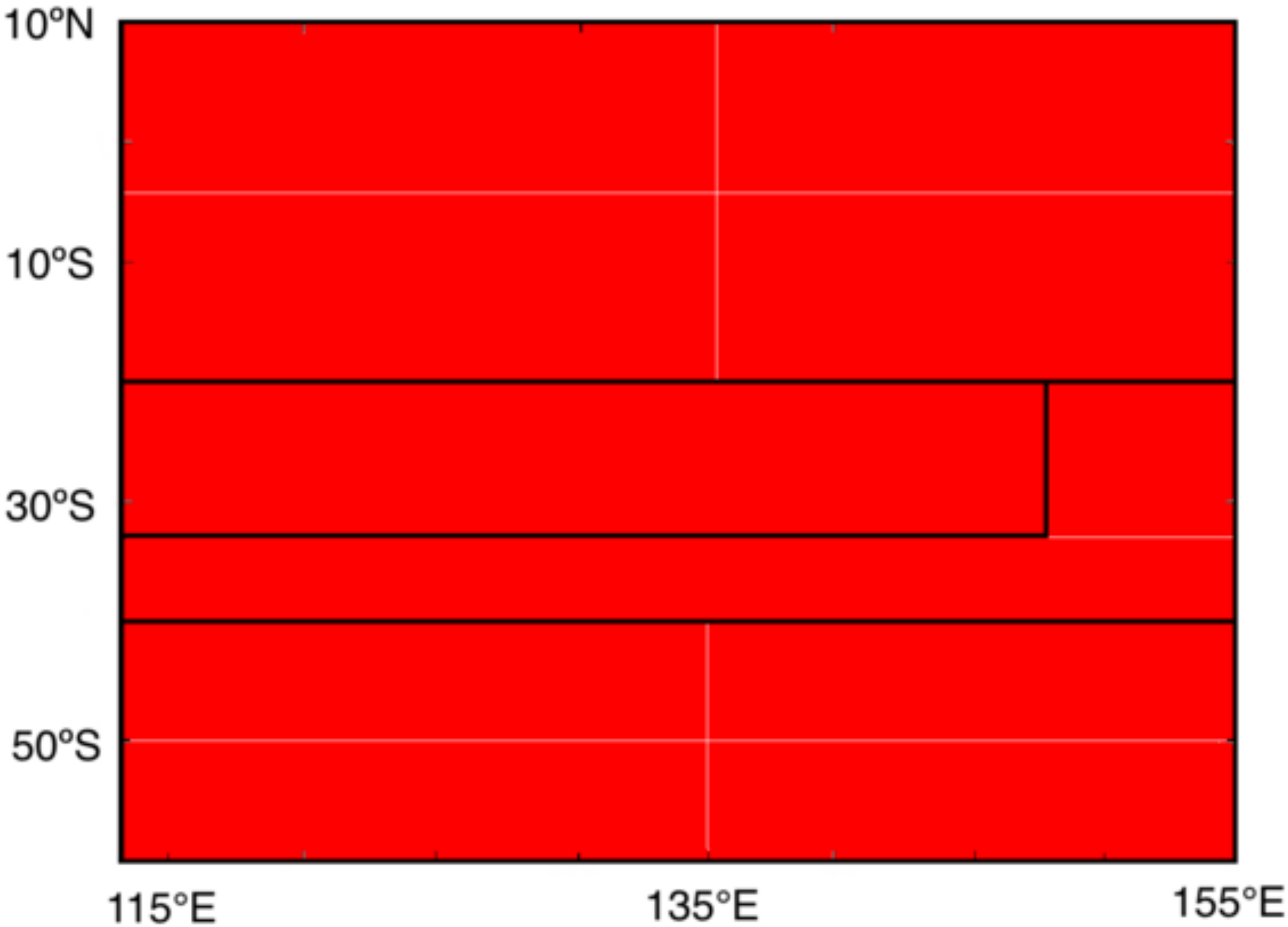

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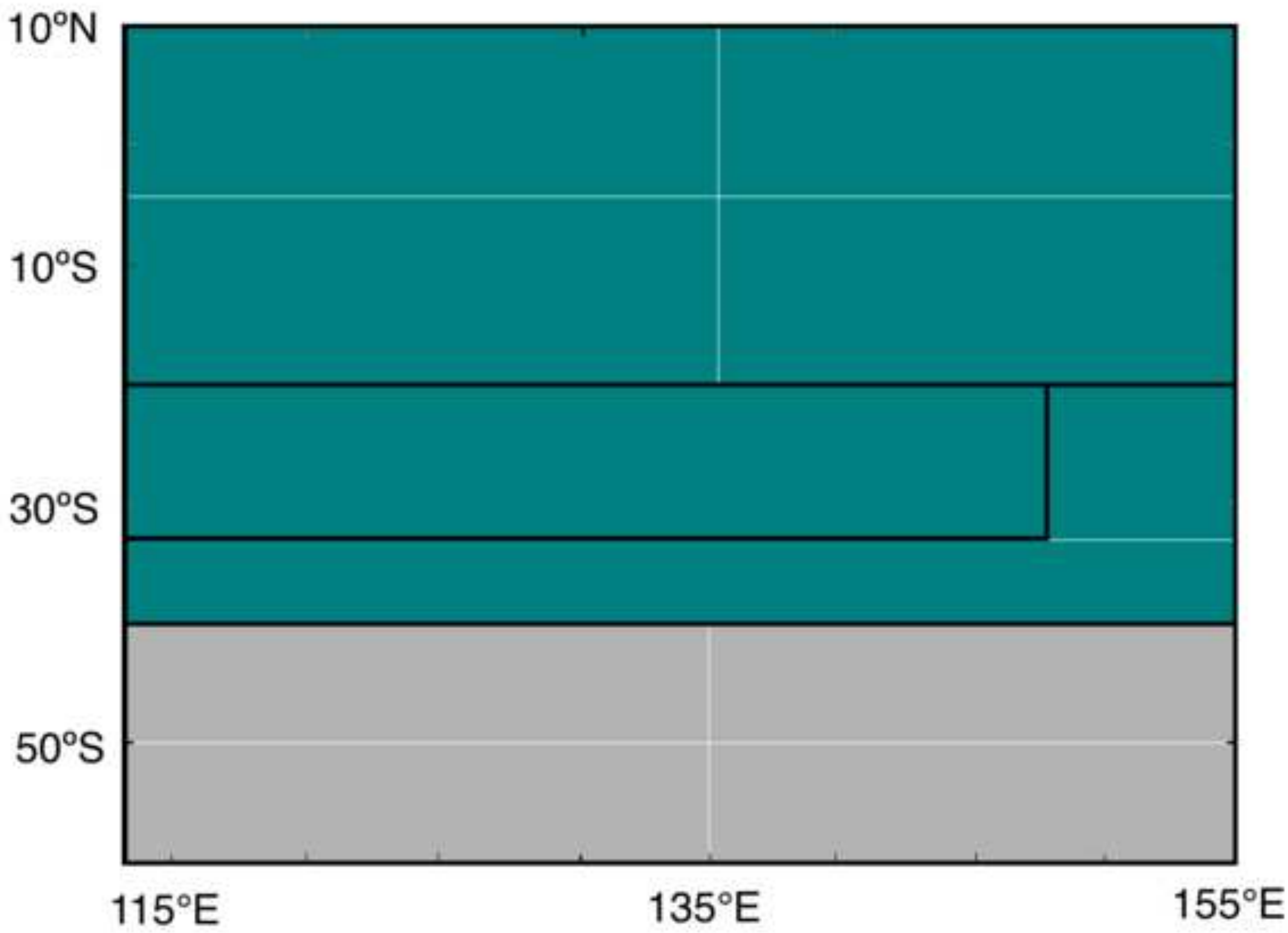

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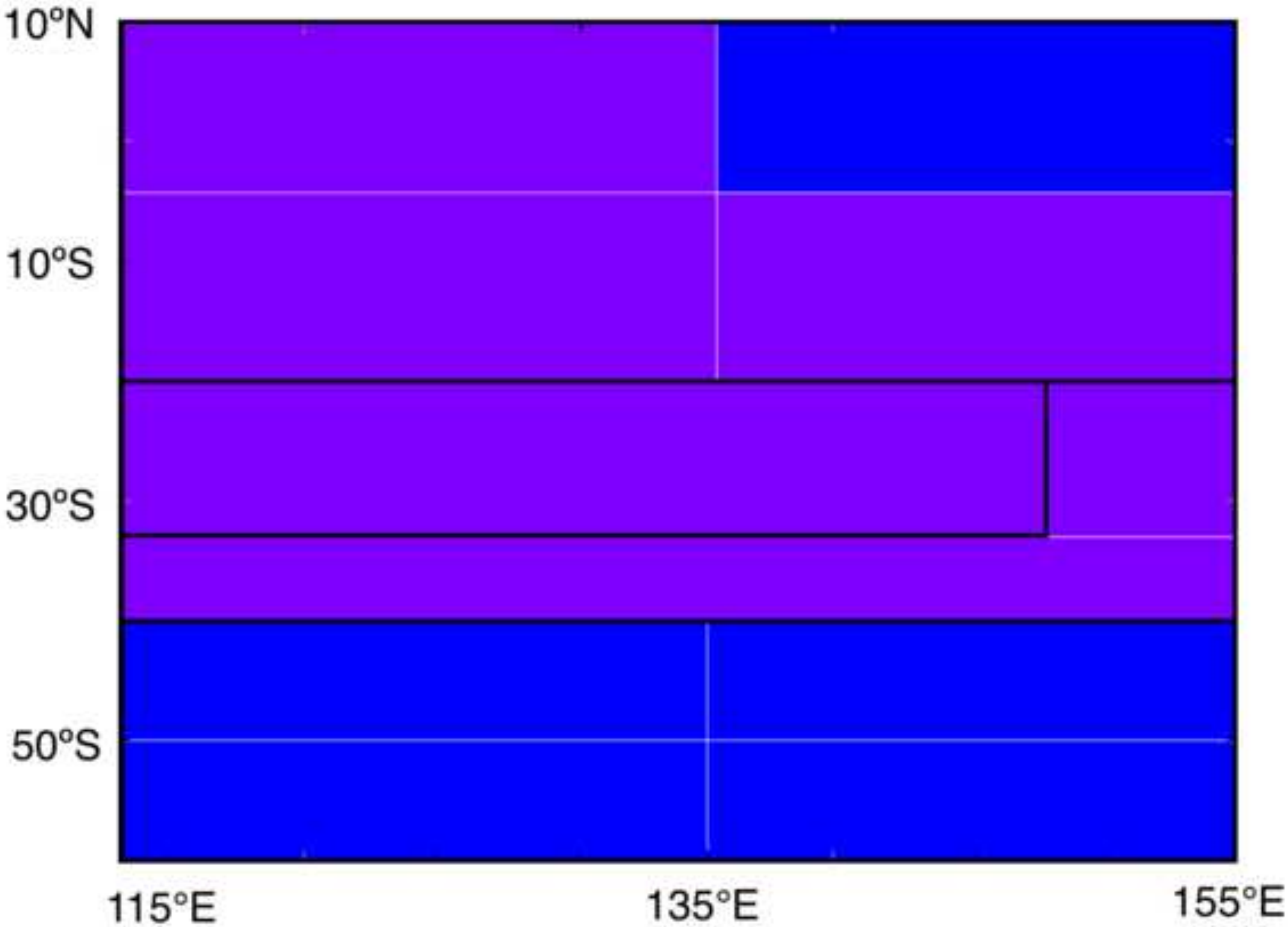

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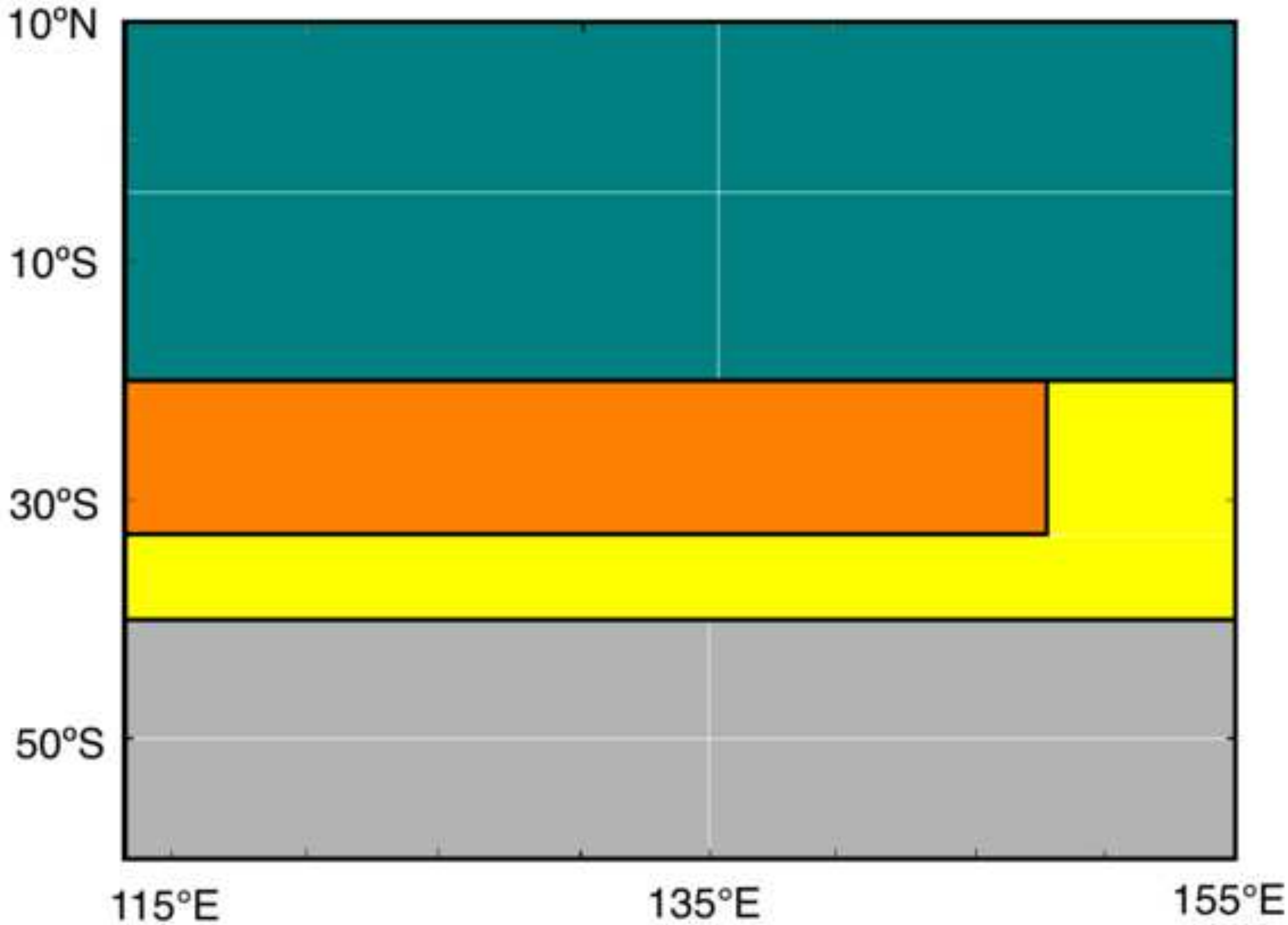

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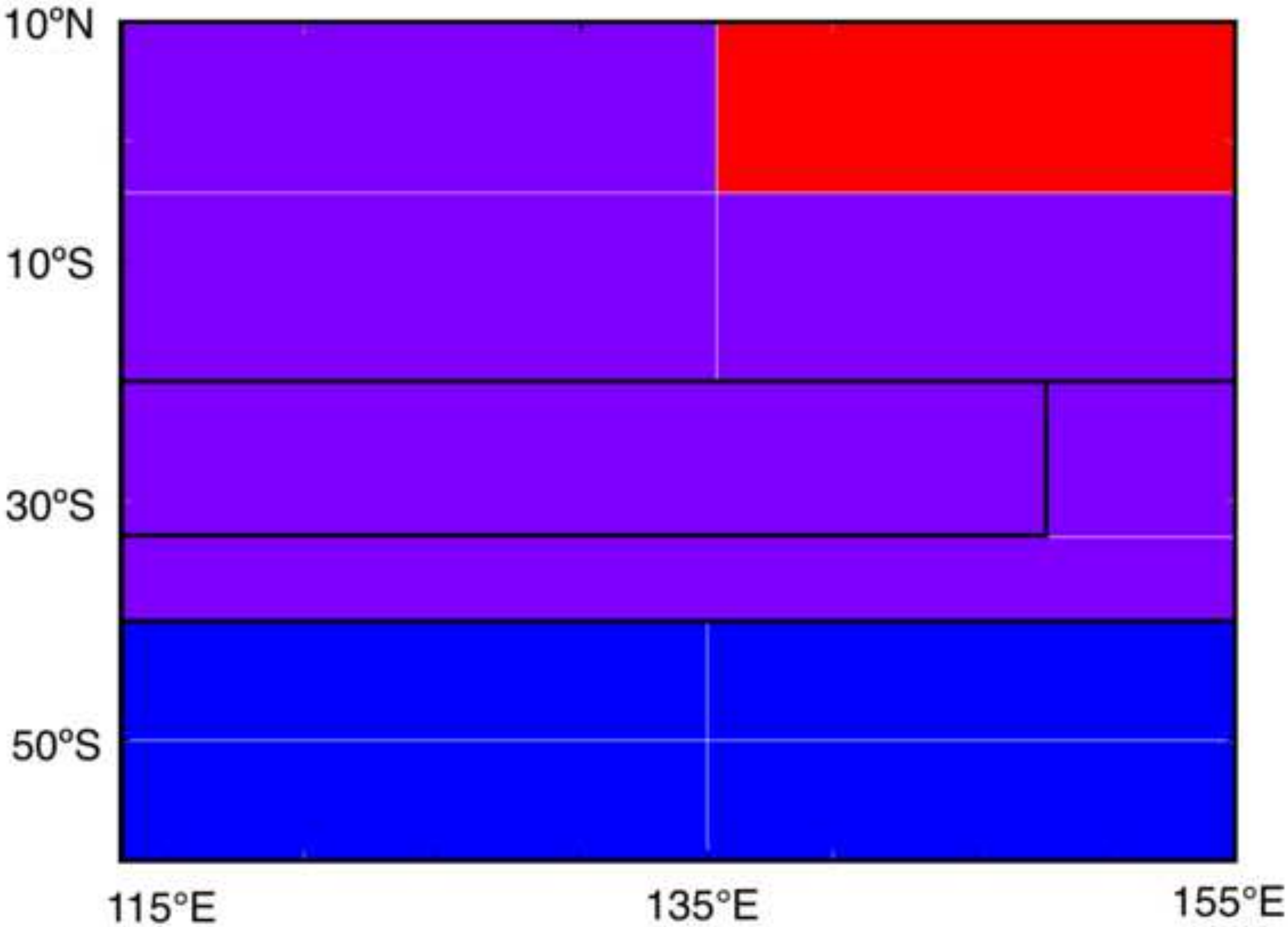

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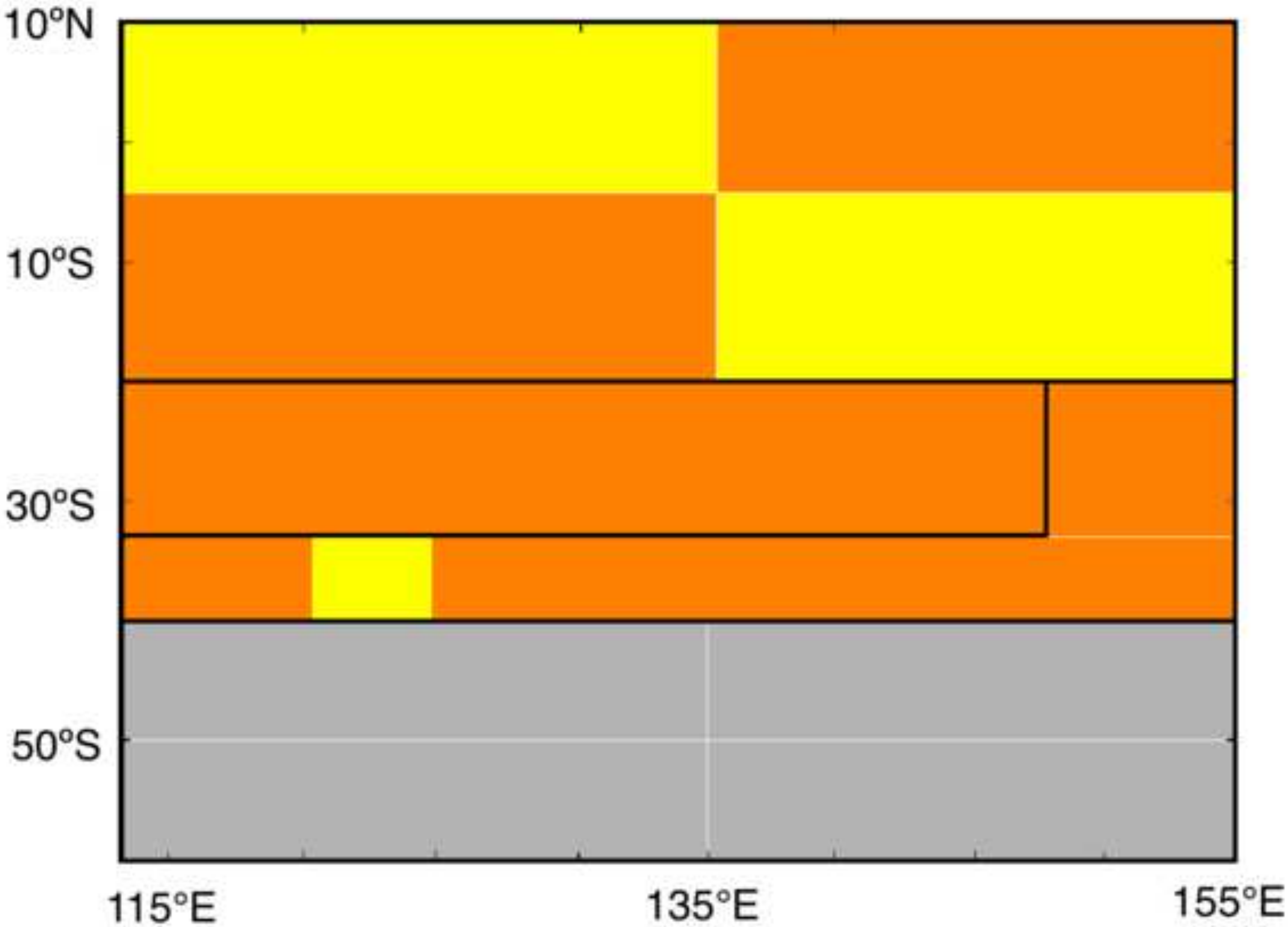
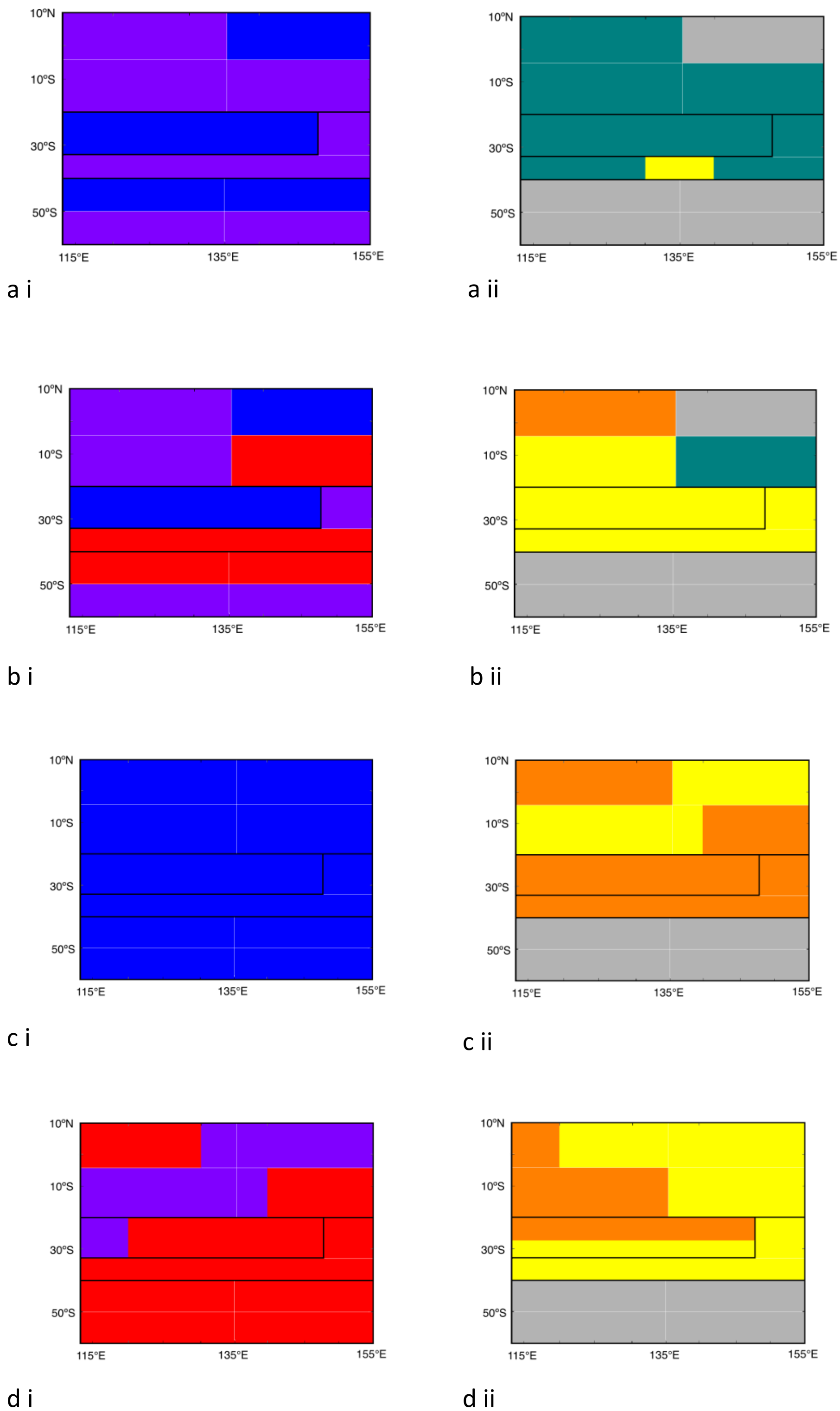
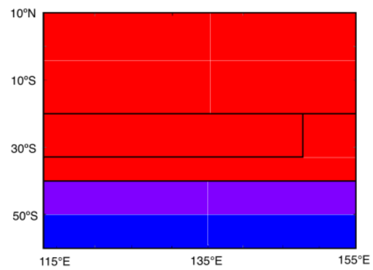


Figure 4 (layout)





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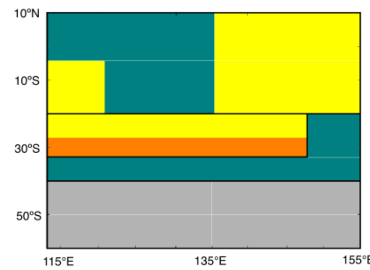

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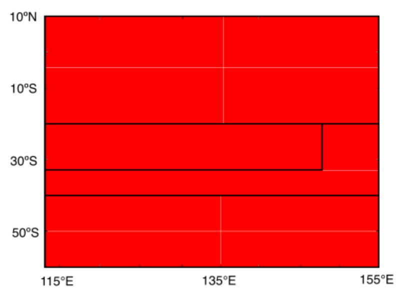

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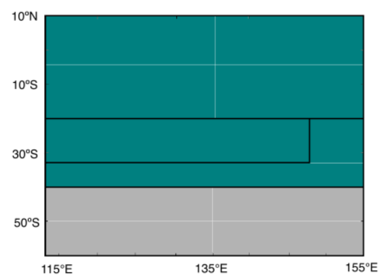

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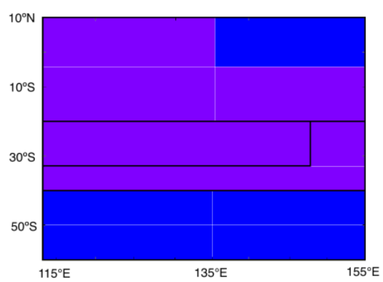

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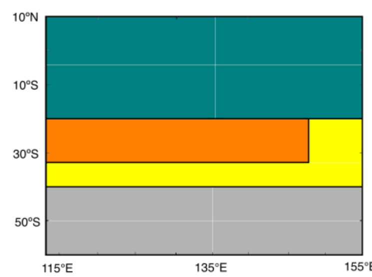

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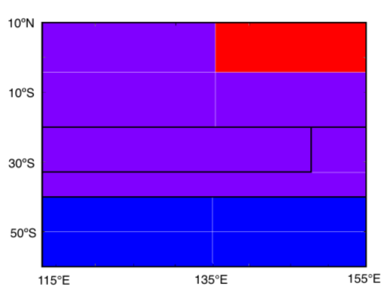

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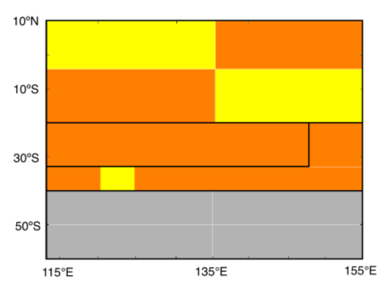

h ii