

CARLOS DIEGO DE SOUSA GURJÃO

**SISTEMA DE MONÇÕES TROPICAIS E MUDANÇAS DE VEGETAÇÃO EM
PERÍODOS INTERGLACIAIS PASSADOS**

Tese apresentada à Universidade Federal de Viçosa, como parte das exigências do Programa de Pós-Graduação em Meteorologia Aplicada, para obtenção do título de *Doctor Scientiae*.

Orientador: Flávio Barbosa Justino

Coorientadoras: Gabrielle Ferreira Pires
Mônica C. Alves Senna

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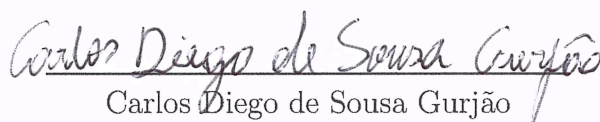
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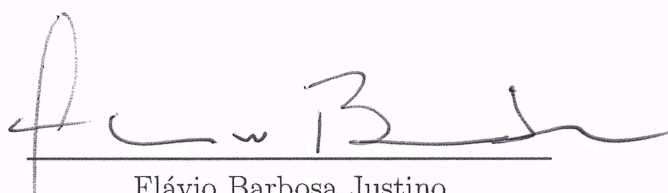
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RESUMO

GURJÃO, Carlos Diego de Sousa, D.Sc., Universidade Federal de Viçosa, junho de 2022. **Sistema de Monções Tropicais e Mudanças de Vegetação em Períodos Interglaciais Passados.** Orientador: Flávio Barbosa Justino. Coorientadoras: Gabrielle Ferreira Pires e Mônica Carneiro Alves Senna.

Os sistemas de monção e a vegetação dinâmica do Hemisfério Sul foram investigados com base em experimentos climáticos conduzidos pelos modelos acoplados ICTP-CGCM (SPEEDY-NEMO) e CCM3/IBIS para os períodos interglaciais Holoceno-Médio (6 mil anos antes do presente - AP), MIS5e (127 mil anos AP), MIS11c (409 mil anos AP) e MIS31 (1,072 milhões de anos AP). A variabilidade interanual e a intensidade dos eventos de monção são analisadas a partir de índices de vortacidade e processos de interação oceano-atmosfera para monções da África, Austrália e América do Sul. Comparado ao clima atual, as análises de correlação demonstraram fraca influência do Oceano Pacífico equatorial na monção de verão austral para o MIS5e, MIS11c e MIS31. Exceção é notada para a monção da América do Sul, onde o Niño_{3.4} exerce um papel mais proeminente nesses intervalos distantes, em consonância com o Atlântico tropical em particular durante o MIS5e. Experimentos de sensibilidade indicam, ainda, que as monções são forçadas não apenas por fatores externos, como o efeito dominante da insolação, mas também por anomalias climáticas como a temperatura da superfície do mar (TSM) das bacias equatoriais do Atlântico e do Pacífico. Após verificar a influência dos sistemas de monções do Hemisfério Austral, o modelo CCM3/IBIS foi executado a partir de saídas de TSM climatológicas do Speedy-Nemo. Para os períodos Holoceno-Médio, MIS5e e MIS11c, notou-se que as áreas apresentam uma redução significativa da vegetação ombrófila, sendo trocada por extensas áreas de savanas e pastagens, em partes da região sul e central da África, e grande parte da Amazônia. Em outro período importante, os resultados mostraram que houve mudanças na cobertura vegetal devido à forçante de insolação durante o MIS31 quando comparado a outros interglaciais, mas o impacto foi maior na Austrália e na região central da América do Sul. Por outro lado, o clima tropical austral se tornou mais seco devido as anomalias negativas de TSM, induzidas pelas bacias do Atlântico e do Pacífico Tropical, reduzindo assim a precipitação continental. Demonstra-se ainda que a resposta da vegetação ao longo da região tropical do Hemisfério Sul, não está fortemente ligada apenas às variações orbitais, mas modulada pelas mudanças internas da TSM do Atlântico e do Pacífico. Ao longo das regiões de estudo, em particular, durante o Holoceno-Médio, MIS5e e MIS11c, a sazonalidade da insolação foi reduzida no Hemisfério Sul, levando ao resfriamento anômalo das bacias oceânicas tropicais, e consequentemente, resultou na migração da zona de precipitação de verão que tem como o principal sistema

meteorológico a Zona de Convergência Intertropical (ZCIT) para latitudes tropicais do Hemisfério Norte. Portanto, combinado com o aumento das temperaturas continentais de verão em baixas latitudes e um período prolongado de seca foram suficientes para que o forçamento levasse a uma retração lenta, mas constante das florestas tropicais do Congo e da Amazônia causando um efeito de extrema aridez.

Palavras-chave: Vegetação. Temperatura da Superfície do Mar. Modelagem Acoplada. Paleoclima.

ABSTRACT

GURJÃO, Carlos Diego de Sousa, D.Sc., Universidade Federal de Viçosa, June, 2022. **Tropical Monsoonal System and Vegetation Changes in Past Interglacial Periods.** Adviser: Flávio Barbosa Justino. Co-advisers: Gabrielle Ferreira Pires and Mônica Carneiro Alves Senna.

Based on coupled climate experiments conducted with the coupled ICTP-CGCM model focusing on interglacial stages Mid-Holocen (6 ka), MIS5e (127 ka), MIS11c (409 ka) and MIS31 (1,072 ka), the Austral summer monsoonal system is investigated. The interannual variability and intensity of monsoon events are analysed from vorticity indices and air-sea interaction processes for Africa, Australia and South America monsoons. Results demonstrated with respect to present day conditions, an orbital driven decrease in precipitation in summer, but slightly shift of the onset and demise periods of monsoons. During the interglacial stages, cooling occurs in the Southern Hemisphere whereas Northern Hemisphere substantially warms, inducing remarkable changes in the position of the oceanic subtropical high pressure systems and equatorial convergence zone. These mechanisms contribute to periods of drought, with reduced precipitation rate over sectors of the Amazon and Northeastern Brazil, northern Australia and southern Africa. Monsoonal rainfall shows different responses to precessional forcing, as well as the relationship between the monsoon and Niño 3.4 differs among the interglacial stages. Compared to current climate, correlation analyses have demonstrated weaker influence of the equatorial Pacific Ocean on the Austral summer monsoon for the MIS5e, MIS11c and MIS31. Exception is noticed for the South America monsoon, where the Niño 3.4 exerts a more prominent role in these distant intervals, in consonance with the tropical Atlantic in particular during the MIS5e. Results are based on a suite of coupled climate simulations conducted with the ICTP-CGCM (Speed-Nemo), which provides the boundary conditions for the CCM3/IBIS model. The CCM3/IBIS model was run from outputs Speed-Nemo of the global ocean and individualized Atlantic, Pacific, and Indian ocean basins forcing, in addition to orbital parameters, and greenhouse gases. For the interglacials, MH, MIS 5e and 11c, it has been found that areas areal a significant reduction in the tropical evergreen forest, but more extensive savanna and grasslands, over parts of the southern and central African region, and into Amazon region. In another important period, the results showed that have been changes that of vegetation cover due to insolation forcing during MIS31 when compared to other interglacials, but the impact was greater in Australia and central South America. On the other hand, the southern tropical climate became drier due to negative SST anomalies, induced by the Atlantic and Tropical Pacific basins, and thus reduced continental precipitation. It is further demonstrated that the response of the vegetation

across the tropical region of the SH, is not strongly linked only to orbital variations, but modulated by the internal changes of Atlantic and Pacific SST. In particular, during the MH, MIS5e and MIS11c, the seasonality of insolation was reduced in the SH, leading to the cooling of the southern tropical ocean basins, resulting in the migration of the summer precipitation zone to boreal latitudes. Therefore, combined with increased low-latitude summer temperatures and a prolonged dryer period were sufficient for the forcing led to a slow retraction, but steady of the rainforests of Congo and Amazonia causing an effect of extreme aridity.

Keywords: Vegetation. Sea Surface Temperature. Dynamic Vegetation Model. Interglacial.

LISTA DE SIGLAS E ABREVIATURAS

AGCM		Atmospheric General Circulation Mode
AFR		Africa
AMZ		Amazonia
AP		Antes do Presente
ASM		Asian Summer Monsoon
ATL		Atlantic Ocean
AUS		Australia
AUSMS		Australian Monsoon System
CAF		Central Africa
CC		Correlation Coefficients
CGSS		Cool Grasslands
CGSH		Cool Grass Shrublands
CO₂		Carbon Dioxide
COMI		Cool Mixed Forest
CTRL		Control
CTRF		Cool Temperate Rainforest
CCM3		Community Climate Model, version 3
CCSM3		Community Climate System Model, version 3
CGCM		Coupled Global Atmosphere-Ocean Circulation Model
DESE		Desert
DJF		December, January and February
EASM		East Asian Summer Monsoon
ENSO		El Niño Southern Oscillation
ET		Evapotranspiration
ET₀		Potential Evapotranspiration

GCM		Global Circulation Models
GHGs		Greenhouse Gases
I		Index
IBIS		Integrated Biosphere Simulator
ICTP		International Centre for Theoretical Physics
IGBP		International Geosphere Biosphere Programme
IND		Indian Ocean
INDO		Indonesia
ITCZ		Intertropical Convergence Zone
JASO		July, August, September and October
JJA		June, July and August
LGM		Last Glacial Maximum
LOVECLIM		the atmosphere (ECBilt), the ocean and sea ice (CLIO), the terrestrial biosphere (VECODE), the oceanic carbon cycle (LOCH) and the Greenland and Antarctic ice sheets (AGISM)
MAMJ		March, April, May and June
MH		Mid-Holocene
MIS		Marine Isotope Stage
MJJA		May, June, July and August
MOC		Meridional Overturning Circulation
MCGA		Modelos de Circulação Geral da Atmosfera
NCAR		National Center for Atmospheric Research
MIS		Marine Isotope Stage
NDJF		November, December, January and February
NCAR		National Center for Atmospheric Research
NEB		Northeastern Brazil
NEMO		Nucleus for European Modeling of the Ocean

NH		Northern Hemisphere
P		Annual Rainfall
PAC		Pacific Ocean
PFTs		Plant Functional Types
R		Reconstructions
Rn		Radiative Budget
SAF		South Africa
SAM		South America
SACZ		South Atlantic Convergence Zone
SAFSM		South African Monsoon System
SH		Southern Hemisphere
SLP		Sea Level Pressure
SAMS		South American Monsoon System
SPEEDY		Simplified Parameterizations, primitivE-Equation DYnamics
SST		Sea Surface Temperature
STEP		Steppe
TAV		Tropical Atlantic Variability
TDFO		Tropical Dry Forest
TOA		Top of Atmosphere
TRFO		Tropical Rainforest
TSFO		Tropical Seasonal Forest
TSM		Temperatura da Superfície do Mar
UNEP		United Nations Environmental Programme
WTRF		Warm Temperate Rainforest
WEFO		Warm Temperate Evergreen Broadleaf Forest
WAMF		Warm Temperate Mixed Forest

SUMÁRIO

1	INTRODUÇÃO GERAL	12
2	DELINEAMENTO EXPERIMENTAL	26
3	ARTIGO 1	37
4	ARTIGO 2	78
5	CONCLUSÃO	120
	ANEXO A	124
	APÊNDICE A	129

INTRODUÇÃO GERAL

A variabilidade da precipitação continental é um fator importante quando há uma preocupação com os cenários futuros do clima, principalmente quando se analisa as monções de verão de regiões tropicais. Um levantamento dos climas passados voltados a períodos interglaciais, podem nos fornecer informações detalhadas da circulação, precipitação, temperatura em superfície e vegetação global, do ponto de vista da variabilidade climática. Considerados como épocas quentes, a sensibilidade dessas variáveis climatológicas podem estar relacionados a fatores externos como a incidência de radiação no topo da atmosfera devido a mudanças nos parâmetros orbitais, bem como da variabilidade da temperatura da superfície do mar (TSM).

Nas latitudes médias da Europa e da América do Norte, o conceito de período interglacial como um intervalo de tempo durante o qual as condições eram pelo menos tão quentes quanto o atual interglacial (Holoceno) era tradicionalmente baseado em evidências do tipo de cobertura vegetal, indicando o desenvolvimento de floresta decídua temperada. Contudo, com o avanço do conhecimento da relação entre a resposta da vegetação dinâmica e a TSM, em diferentes períodos interglaciais permitirão uma maior compreensão da interação entre eles, e ainda podem apontar para o efeito potencial das variações climáticas futuras no padrão de vegetação. Atualmente, estudos que descrevam a influência dos diferentes processos de interação são escassos, como por exemplo, de que modos os Oceanos Pacífico e Atlântico norte e sul e suas áreas equatoriais induziram modificações nos padrões de vegetação, como demonstrado por Pereira et al. (2017).

Todavia, sabe-se que a relação entre a vegetação global e o clima é de imensa complexidade. Por exemplo, o clima da América do Sul é fortemente dependente dos padrões da TSM dos Oceanos Pacífico e Atlântico, e quaisquer alterações climáticas resultarão em mudanças na vegetação (Nobre and Srukla, 1996). Pereira et al. (2017) verificaram que o clima da região sul da América do Sul (incluindo a Patagônia) é altamente dependente das variações da TSM nos extratropicais, embora também esteja correlacionado com o Pacífico equatorial. Deve-se destacar que outros fatores, como por exemplo mudanças na quantidade de radiação solar devido a efeitos orbitais terrestre causem mudanças na distribuição da

vegetação. Neste sentido, há evidências que mostram que a distribuição da vegetação na Terra foi substancialmente diferente a depender da era geológica. Durante o último máximo glacial (Last Glacial Maximum - LGM), há aproximadamente 21000 anos Antes do Presente (AP), alguns refúgios vegetacionais ainda podiam ser encontrados na América do Sul, mas a área atual de florestas ombrófilas foi reduzida, havendo predomínio de gramíneas e savanas Behling (2002).

O desenvolvimento e uso de modelos de vegetação dinâmica globais são necessários para analisar respostas transientes dos ecossistemas terrestres às variações no clima, uso da terra e aumento da concentração do CO₂ atmosférico (Steffen et al., 1992). Os ecossistemas terrestres desempenham papéis importantes na modulação da concentração do CO₂ na atmosfera e na atenuação dos efeitos do aquecimento global (Le Quéré et al., 2015).

Maley (1997), argumenta que a transição do Holoceno Médio (cerca de 7000 a 3800 anos AP) para o Holoceno Tardio (aproximadamente 3800 AP), foi marcado por grandes mudanças climáticas. Estas mudanças teriam sido introduzidas por variações orbitais da Terra e desta forma por variações na quantidade de energia disponível na superfície do planeta.

Estudos usando modelos acoplados biosfera-atmosfera indicam que a região amazônica pode sofrer um declínio significativo na precipitação em um cenário de aquecimento global, levando a perda rápida da floresta tropical e, conseqüentemente, uma redução adicional das chuvas (Cox et al., 2004; Betts et al., 2004). No entanto, esses resultados contrastam fortemente com os resultados dos estudos de modelagem sobre a resposta da vegetação tropical ao CO₂ elevado (Cox et al., 2013). Algumas simulações indicam que o aumento de CO₂ será acompanhado por um aumento da cobertura vegetal: este processo iniciaria um ciclo de retroalimentação hidrológico, levando a aumentos nas precipitações e na umidade do solo (Levis et al., 2000). As trocas de CO₂ que ocorrem na interface oceano-atmosfera são uma das componentes mais importantes do ciclo do carbono antropogênico e natural global (IPCC, 2013). Entretanto, deve ser levado em consideração Em escalas de tempo longas (multi-decenais a milhares de anos), a transferência do CO₂ com origem em águas profundas é um fator crítico para regular a concentração desse gás na atmosfera (Raupach et al., 2007).

Os modelos numéricos podem ser utilizados para entender o comportamento da atmosfera, bem como do oceano, levando em consideração diversos aspectos, tais como a transferência de energia entre os oceanos e a atmosfera, através do fluxo de calor, vapor d'água, momentum e concentração de CO_2 . Visualiza como a vegetação tropical responde as mudanças na TSM, em períodos interglaciais, torna-se importante porque pode ser um indicativo de como o sistema climático responderá a cenários futuros.

A influência dos padrões de aquecimento da TSM é notório em estudos de modelagem climática e da mudança da vegetação, como por exemplo a retração da Floresta Amazônica (Cox et al., 2000). Baseando-se nos experimentos conduzidos por Pereira et al. (2014), será de importância ímpar analisar a resposta global de vários tipos de biomas as mudanças da TSM. Entretanto, devemos nos perguntar: quão marcante será a mudança da vegetação em grande escala nos trópicos, principalmente sobre a América do Sul (AS)? Uma vez que serão realizados testes de sensibilidades que envolvem alterações nos parâmetros orbitais da Terra (excentricidade, obliquidade e precessão), em que chamaremos de simulações MIS, por representar características semelhantes ao Marine Isotope Stage.

1.1 Objetivos

Com base em modelagens numéricas acopladas, o objetivo deste trabalho será examinar em detalhes a variabilidade intrasazonal das monções do Hemisfério Sul, mais precisamente das regiões Central e Sul da África, Austrália e América do Sul. Além disso, será investigado o efeito da variabilidade espaço-temporal da TSM sobre a vegetação dinâmica em períodos interglaciais, usando modelagem acoplada vegetação-clima. Especificamente, o estudo procura responder as seguintes questões:

- O modelo ICTP-AGCM (Speedy-Nemo) será capaz de reproduzir o sistema de monções do Hemisfério Sul durante os interglaciais?
- O modelo CCM3/IBIS será capaz de reproduzir o clima e a vegetação em épocas remotas quando comparados a paleo-reconstruções?
- Como a vegetação global respondeu a uma potencial mudança de TSM nas bacias oceânicas individualizadas durante os períodos interglaciais?

O restante deste primeiro capítulo é dedicado a revisão bibliográfica tomada como base para este trabalho. Serão abordados assuntos relacionados ao entendimento paleoclimático e interações vegetação-oceano-atmosfera. No segundo capítulo será descrito o delineamento experimental, o conjunto de dados utilizados neste estudo, a escolha dos modelos de circulação geral da atmosfera e testes de sensibilidade climática. O terceiro capítulo, intitulado "Southern Hemisphere Monsoonal System During Superinterglacial Stages: MIS5e, MIS11c and MIS31", foi submetido na revista científica *Climate Dynamics*. O quarto capítulo é intitulado "Orbitally-driven Vegetation Changes During Past Interglacials: the Role of Sea Surface Temperature".

1.2 Revisão bibliográfica

Nesta sessão será apresentada a revisão bibliográfica tomada como base para este trabalho. Antes de tudo, uma breve definição de variabilidade natural do clima será apresentada, enfatizando os fatores externos e internos.

O entendimento da variabilidade orbital sobre escalas de tempo remotas é um tema central na teoria orbital. O planeta Terra está em plano orbital ao redor do Sol, chamada de Eclíptica. Os três componentes mais importantes da teoria orbital são precessão, obliquidade e excentricidade.

Existem diversas formas que afetam os padrões de sazonalidade da radiação solar incidente no topo da atmosfera e, conseqüentemente, a variabilidade climática, principalmente quando estão associadas a configuração orbital. Sabe-se que a órbita da Terra é associadas a força gravitacional imposta pelo Sol e pelos outros corpos celestes do Sistema Solar que influenciam três parâmetros orbitais distintos da Terra com diferentes periodicidades que são chamados de Ciclos de Milankovitch (Milankovitch, 1930; Berger, 1978)(Figura 1.1). Esses efeitos dependem das posições relativas entre o Sol, a Terra e os outros planetas. De acordo com a teoria, os parâmetros excentricidade, precessão e obliquidade, são capazes de alterar o clima da Terra em escalas de tempo milenar. Com a gradativa alteração desses parâmetros, a quantidade de energia que atinge o topo da atmosfera sofre variações, assim como também a que atinge a superfície do planeta. As descrições dos ciclos de Milankovitch são dadas em sequência:

- variações da forma elíptica da órbita da Terra em torno do Sol (excentricidade);
- mudanças na inclinação do eixo de rotação da Terra em relação ao plano elíptico (obliquidade);
- oscilação do eixo da Terra como um pião (precessão).

A periodicidade da excentricidade pode variar em torno de 100.000 à 400.000 anos, enquanto as variações na obliquidade ocorrem com periodicidades em torno de 40.000 anos, e 20.000 anos para a precessão.

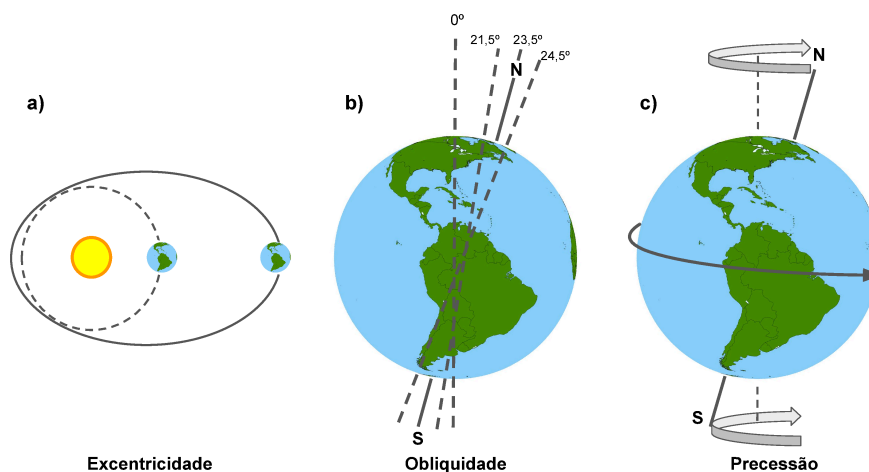


Figura 1.1 – Variações dos parâmetros orbitais: (a) excentricidade, (b) obliquidade e (c) precessão. Adaptada de Harper (2007)

1.2.1 Sistemas de Monção do Hemisfério Sul

O regime de monção é caracterizado por alternância marcante nos períodos secos e chuvosos durante o ano. No Hemisfério Sul a chuva de monção tende a ocorrer durante os meses que caracterizam o verão austral, predominando em novembro, dezembro, janeiro e fevereiro. Durante o inverno austral, quando ocorre a estação seca, a componente zonal do vento é invertida sobre o continente, soprando ar frio e seco para o oceano. No entanto, os impactos causados por esse sistema geram extremos climáticos, pois além de causar inundações e danos notórios à população em geral, principalmente à sociedade mais vulnerável em áreas urbanas devido às chuvas torrenciais; o contrário causa ausência de recursos hídricos e calor extremo durante a seca.

A Figura 1.2 mostra as amplitudes do primeiro harmônico da precipitação para o ciclo de doze meses, sendo de janeiro a dezembro, utilizando-se dos metadados de precipitação inferidos pela reanálise do ERA5 Hersbach (2016) para o período de 1979 a 2020 (Figura 1.2a), e para um período de 100 anos do acoplado Speedy-Nemo (Figura 1.2b). O destaque da precipitação se dá pelos altos valores de precipitação sobre a faixa tropical, onde mostra a Zona de Convergência Intertropical (ZCIT).

Durante os meses do verão austral (NDJF), a radiação de onda curta que atinge a superfície tende a aumentar no HS, vindo a provocar o aquecimento, tanto oceânica, assim como continental, e consequentemente, favorece o deslocamento meridional do ramo

ascendente da Célula de Hadley do HN ao HS (Zhang et al., 2015) (ver Figura A2 do material suplementar para mais detalhes). Logo, nota-se a distribuição da precipitação, sendo estes mais intensos sobre a região equatorial. As chuvas de monção da América do Sul (AS), África do Sul (AFS) e Austrália (AUS), mostram picos durante quadriênio (NDJF), em que 70% da precipitação anual é contribuída durante essa temporada (Figura 1.2c-d-e). No entanto, a fase chuvosa das monções do Hemisfério Austral tendem a perdurar até meados de abril, dando início a estação seca que vai até setembro. Isso sugere o enfraquecimento da monção austral. A explicação mais clara do motivo associado ao desencadeamento da diminuição (aumento) da precipitação e enfraquecimento (fortalecimento) das monções no HS (HN) é devido a redução (aumento) de insolação que atinge o topo da atmosfera, assim como da superfície do planeta devido à configuração astronômica da época.

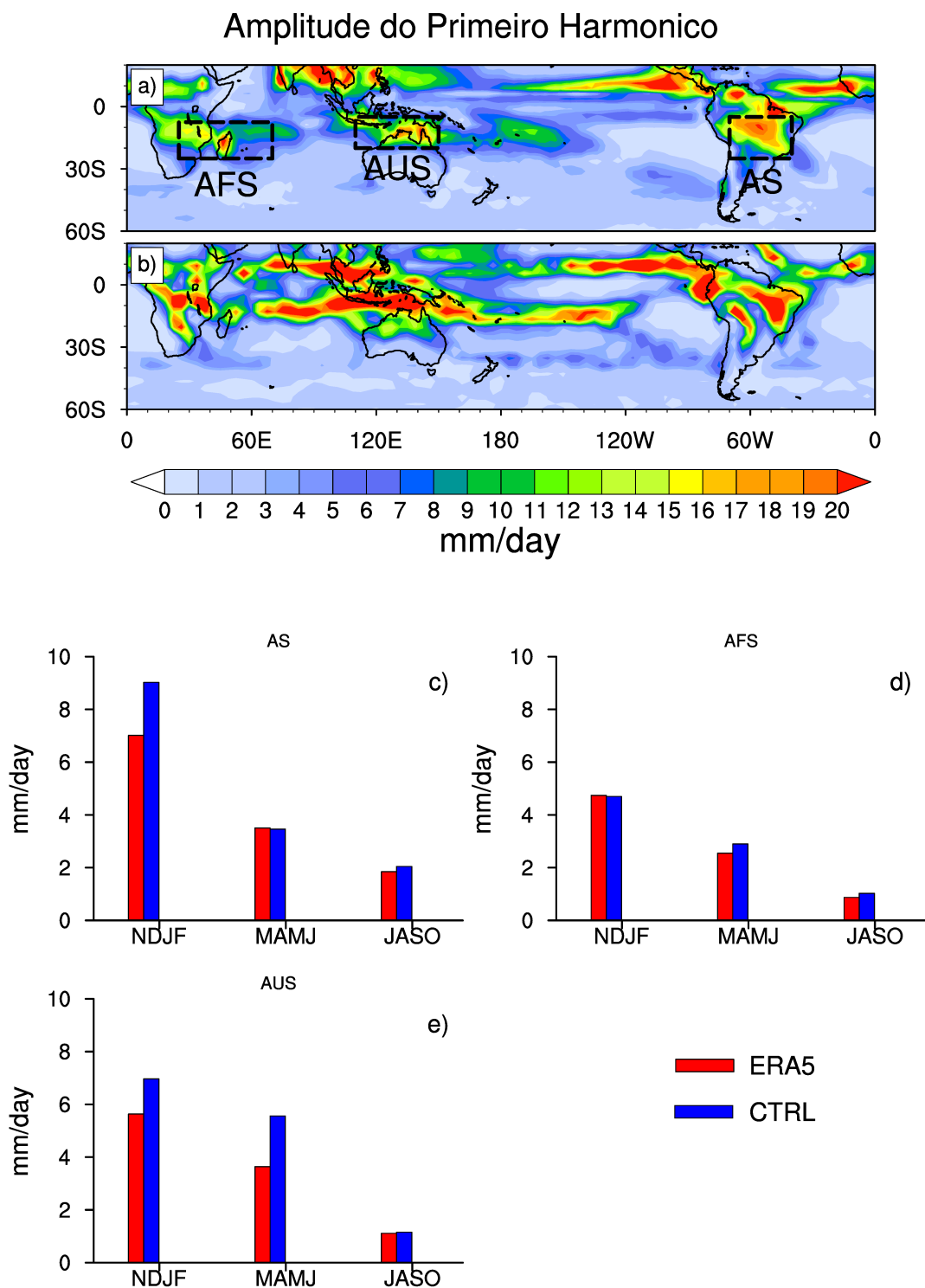


Figura 1.2 – Amplitudes do primeiro harmônico do ciclo de precipitação (mm/dia) anual de janeiro a dezembro. (a) ERA5 e (b) Simulação Controle (CTRL), modelo acoplado Speedy-Nemo (apreviações em inglês para *Simplified Parameterizations, primitivE-Equation DYnamics - Nucleus for European Modeling of the Ocean*); Distribuição sazonal da precipitação das regiões de Monção do Hemisfério Sul; (c) América do Sul (AS, 5S-25S,70W-40W), (d) Sul da África (AFS, 7.5S-25S,25E-70E) e (e) Austrália (AUS, 5S-20S,110E-150E).

1.2.2 Os Interglaciais Holoceno-Médio, MIS5e, MIS11c e MIS31

O Holoceno Médio apresenta uma grande semelhança com os interglaciais MIS-5e e MIS-11c, no que se trata de vegetação sobre a América do Sul. Porém, em relação a monção de verão, através dos arquivos geológicos e modelagem numérica aqui discutidos, notou-se que durante o Holoceno Médio a monção da América do Sul era uma pouco mais intensa do que os interglaciais Eemian e MIS-11c.

MIS-5 é um estágio de isótopo marinho no registro da temperatura geológica, entre 80.000 e 130.000 AP, apresentando concentrações de CO₂ da ordem de 290 ppm constante. Assim, o Holoceno-Médio, é comparado com o MIS5 (Lüthi et al., 2008). Considerado como uma época quente em que as médias global de temperaturas foram estimadas em torno de 0.5-1°C, e que também apresentou redução significativa de gelo na Groelândia em razão das altas temperaturas médias do verão boreal das altas latitudes, essas que se aproximavam de 4°C a mais quando comparado aos dias atuais Otto-Bliesner et al. (2006); Hoffman et al. (2017). Nikolova et al. (2013), utilizando de dois modelos de diferentes complexidades, sendo o CCSM3 (Community Climate System Model 3) e o LOVECLIM (LOch-Vecode-Ecbilt-CLio- agIsM Model), também verificaram a resposta da temperatura a forçante orbital durante o Emian (MIS5e). O forte aquecimento foi mais perceptível durante o verão do Hemisfério Norte, principalmente sobre o continente do que o Oceano devido as grandes diferenças entre as capacidades térmicas. Essa diferença entre os dois modelos, segundo os autores, pode estar relacionada ao fato de que os feedbacks clima-vegetação estão ausentes no CCSM3 devido à falta de um modelo dinâmico de vegetação. Por outro lado, pode estar relacionado ao fato de que LOVECLIM leva a superestimar a precipitação em torno de 30°N e a temperatura nos trópicos Goosse et al. (2010) e, portanto, faz com que amplifique a resposta da vegetação. Isso também se refletiu na comparação entre modelos para o clima do Holoceno médio (Braconnot et al., 2007). A expansão da área vegetada durante MIS5e no Sahara pelo LOVECLIM, resulta da migração da ZCIT para o norte, bem como da advecção de umidade, de acordo com registros proxy Jolly et al. (1998), em que mostra um Saara mais úmido e com uma maior expansão da vegetação. Em contrapartida, há um aumento de deserto sobre a Austrália devido a diminuição da precipitação.

MIS-11c é caracterizado por anomalias positivas de TSM em altas latitudes e forte circulação termohalina. O MIS-11c é considerado o período interglacial mais quente dos últimos 500.000 anos, com concentrações de CO₂ em torno de 280 ppm constante (Candy et al., 2014). Além do fator externo como resposta para o aquecimento anômalo do HN durante o MIS-11c, há também evidências de que o derretimento de geleiras durante esse interglacial tem associação com fatores internos como a Circulação de Revolvimento Meridional do Atlântico (do inglês, *Atlantic meridional overturning circulation - AMOC*) (Rachmayani et al., 2017). A partir de registros paleoclimáticos terrestres como sedimentos de lagos antigos, loess (sedimentos depositados pelo vento) e cavernas, alguns autores Cheng et al. (2016); Meckler et al. (2012) sugerem que o MIS-11c foi considerado quente com abundante paisagem de biomassa e altas concentrações de CO₂ e CH₄, assemelhando-se ao Holoceno-Médio.

O MIS-31 foi um período com fortes períodos interglaciais, também coincide com uma das combinações orbitais mais extremas do Pleistoceno, com altos valores de obliquidade e excentricidade, produzindo fortes anomalias de radiação durante verões de altas latitudes, alternando entre os hemisférios. Uma estimativa recente de CO₂ do Pleistoceno Médio com base em isótopos de boro (Hönisch et al., 2009), sugeriram níveis de cerca de 300 ppm constante no momento do MIS31. De acordo com DeConto et al. (2007), houve um aquecimento da TSM na Antártica, em que o modelo simulou em torno de 2-5°C e, consequentemente, o derretimento do gelo marinho, no qual foi impulsionado por respostas diretas e indiretas da insolação durante o MIS31.

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DELINEAMENTO EXPERIMENTAL

Este capítulo tem como propósito apresentar a região de estudo, os conjuntos de dados e descrever os procedimentos e metodologias utilizados na elaboração dessa tese de doutorado. Antes de tudo, optou-se em utilizar os metadados do ERA5 que fazem parte da nova geração de reanálises para fins de validação do modelo. Portanto, apresentar os conjuntos de modelos (Tabela 2.1), descrever os procedimentos e metodologias que serão utilizados na elaboração deste estudo. Contudo, o trabalho proposto tem como ferramenta a modelagem climática acoplada.

Tabela 2.1 – Características dos produtos de reanálises do Era-Interim e ERA5.

Reanálises	Era-Interim	ERA5
Intervalo de Tempo	1979 - 2020	1979 - presente
Domínio	Global	Global
Resolução do Modelo	TL255 (80km)	TL639 (31km)
Resolução Espacial	0,75°x0,75°	0,25°x0,25°
Níveis Verticais	Híbrido Sigma (60 níveis)	Híbrido Sigma (137 níveis)
Esquema de Assimilação	4D-Var	4D-Var
Referência	Dee et al. (2011)	Hersbach (2016)

Nos dias atuais ainda há escassez de medições e coleta de dados de forma direta dentro do continente e oceano. No entanto, ferramentas como os produtos de reanálises são fundamentais para estudar a variabilidade climática das monções tropicais para fins de comparação com períodos paleoclimático e cenários futuros. Contudo, as reanálises do ERA5 foram desenvolvidos a partir da nova geração de produtos do Centro Europeu de Previsões do Tempo de Médio Prazo (ECMWF) Hersbach (2016), sendo interpolados a uma grade de 0,25°x0,25°, apresentando um domínio global, 137 níveis verticais do tipo Híbrido Sigma e esquema de assimilação 4D-Var (Tabela 2.1). A vantagem de usar o novo banco de dados é que ele apresenta uma série de melhorias quando comparado ao antecessor, o ERA-Interim (Dee et al., 2011). É importante destacar que para a análise da circulação dinâmica, dados de reanálise de médias mensais do ERA5, correspondendo o período de janeiro de 1979 a dezembro de 2020. Neste trabalho foram empregados os seguintes parâmetros meteorológicos: componentes zonal e meridional do vento em 850

hPa e 250 hPa, temperatura do ar em superfície (T_{air}), precipitação (mm/day), TSM e altura geopotencial de 250 hPa.

2.0.1 Testes de Sensibilidades

Os experimentos interglaciais apresentados nesta ocasião foram realizados nos modelos Speedy-Nemo e CCM3/IBIS. O modelo acoplado Oceano-Atmosfera tem uma configuração computacionalmente eficiente que permite realizar simulações "ensemble" e milenares. Já o CCM3/IBIS, é um modelo atmosférico, no entanto, com a mesma eficiência em realizar simulações. Outro aspecto positivo é que o modelo simula a vegetação dinâmica. Portanto, para o desenvolvimento deste estudo, foram realizados vários experimentos de sensibilidade climática com alterações dos parâmetros orbitais (excentricidade, obliquidade e precessão) da Terra e concentrações de gases de efeito estufa como CO_2 , CH_4 e NO_2 , associando as características da época que remetem à condições semelhantes aos interglaciais Holoceno-Médio, MIS-5e, MIS-11c e MIS-31 (Tabela 3.1).

Tabela 2.2 – Configurações das concentrações de efeito estufa (CO_2 (ppmv), CH_4 (ppbv), e N_2O (ppbv)) e parâmetros orbitais utilizados nos experimentos CTRL, Holoceno-Médio, MIS5e, MIS11c e MIS31

Experimentos	Data	CO_2	CH_4	N_2O	Ecc.	Obl.	Prec.
CTRL	Presente	380	801	289	0.016	23.43	101.37
Holoceno-Médio	6 ka BP	280	700	275	0.018	24.10	0.87
MIS5e	127 ka BP	287	724	262	0.039	24.04	272.92
MIS11c	409 ka BP	285	713	285	0.019	23.78	265.34
MIS31	1.072 ka BP	325	800	288	0.056	23.89	289.79

A simulação de controle (CTRL), conduzida com o modelo do (Centro Internacional de Física Teórica - International Centre for Theoretical Physics - Atmospheric General Circulation Model - ICTP-AGCM) é uma representação estatística do estado climático atual conforme destacado em Justino et al. (2019). A concentração de CO_2 em nosso clima de CTRL é de 380 ppm constante, pois caracteriza as emissões antropogênicas de CO_2 até o ano de 1950, mas não inclui o acelerado aumento que ocorreu no final do século XX. Para os cenários climáticos interglaciais, as simulações foram conduzidas por um período de tempo necessário de ajuste devido ao fato de que as condições iniciais e de fronteira não estarem em equilíbrio com o sistema ao qual serão forçados, ou seja, será utilizada uma

climatologia dos últimos 500 anos de cada simulação para analisar as variações atmosféricas e oceânicas, com 5 ensembles. A escolha por um período maior de análise está relacionada ao tempo necessário para a parte oceânica entrar em equilíbrio, bem como obter uma representação mais realista da vegetação e de suas situações climáticas.

A influência dos parâmetros orbitais no clima do planeta são baseados na Teoria de Milankovitch, em que a excentricidade, precessão e a obliquidade da Terra influenciam o nosso clima.

Para se certificar de que os dados do modelo acoplado Speedy-Nemo são acurados para os estudos, foram necessários testes de sensibilidade que alteram os parâmetros orbitais da Terra de acordo com a Teoria de Milankovitch (Milankovitch, 1930).

No capítulo que envolve o modelo de vegetação dinâmica, foram utilizados experimentos similares aos conduzidos por (Pereira et al., 2014). Estritamente falando, dois grupos de simulações climáticas serão avaliados. Experimentos forçados pela TSM-CTRL mantendo a distribuição global, e simulações forçadas pela modificação da TSM para diferentes bacias oceânicas do Atlântico Sul (ATL), Pacífico Sul (PAC) e Índico (IND). Os padrões de TSM combinados são usados em cada simulação como mostra o diagrama esquemático das simulações (Figura 2.1).

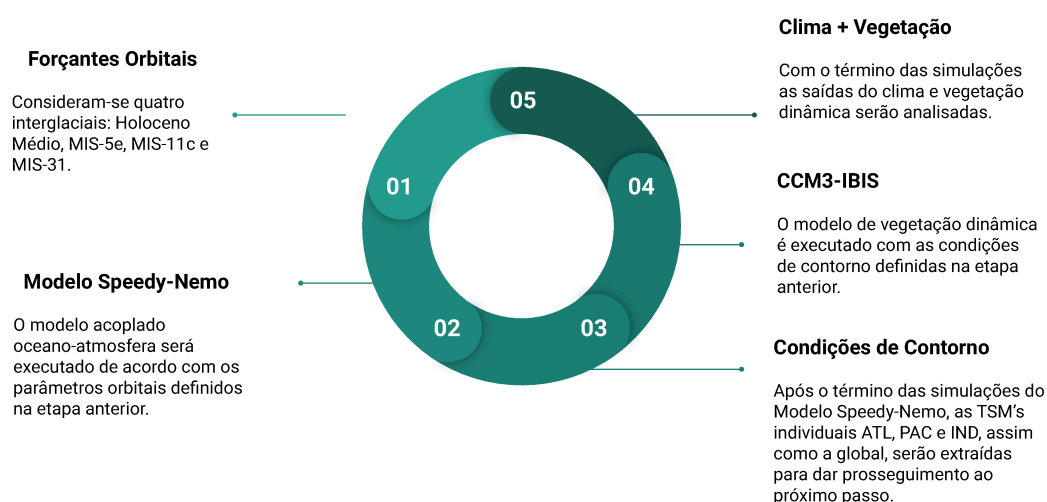


Figura 2.1 – Ilustração esquemática das etapas de simulação pelos modelos acoplados oceano-atmosfera (Speedy-Nemo) e vegetação-atmosfera (CCM3/IBIS). Parâmetros orbitais: Presente; 6 ka = Holoceno Médio; 127 ka = MIS5e; 409 ka = MIS11c e 1072 ka = MIS31.

As bacias individualizadas PAC, ATL e INDI, são baseados em uma combinação de TSMs climatológicas junto com o CTRL. Para evitar a descontinuidade, um fator de relaxamento é aplicado à interface entre as TSMs de acordo com uma função senoidal. Essas simulações são executadas por 50 anos. Os três setores do Oceano global que afetam fortemente o clima são mostrados na Figura 2.2.

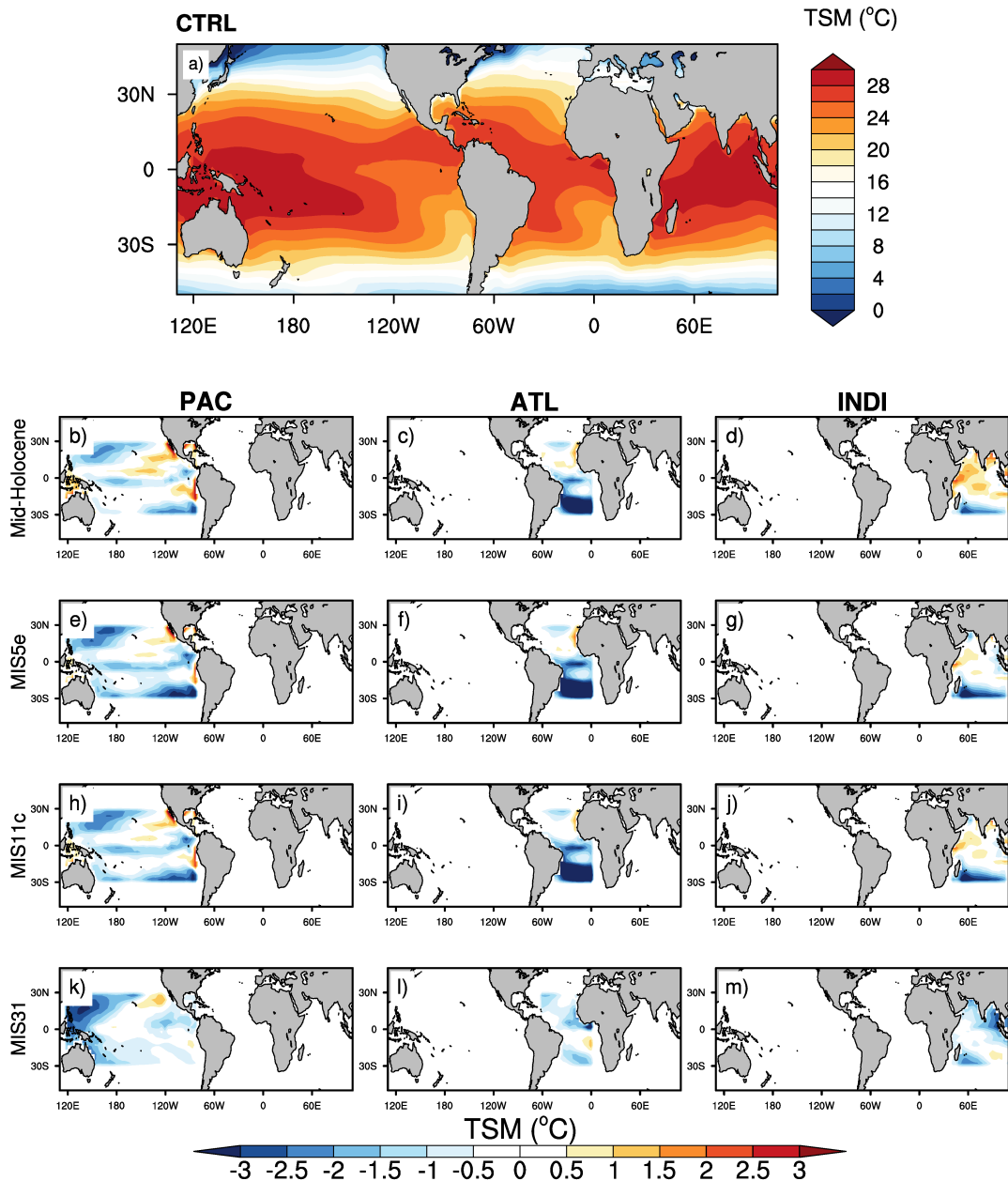


Figura 2.2 – As condições de contorno referentes as TSMs são representadas pelo CTRL = TSM global (a), Pacífico Tropical = PAC, 30°N-30°S e 120°E-280°W (b-e-h-k); Atlântico Tropical = ATL, 30°N-30°S e 300°E-360°W (c-f-i-l) e Índico Tropical = INDI, 30°N-30°S e 40°E-110°W (d-g-j-m). Os interglaciais são representados nas linhas do painel como Holoceno-Médio (b-c-d), MIS5e (e-f-g), MIS11c (h-i-j) e MIS31 (k-l-m).

2.1 Descrição dos Modelos

Os Modelos de Circulação Geral da Atmosfera (MCGA) disponibilizam de representações numéricas dos diferentes componentes do sistema climático e suas interações. Os MCGA incluem códigos numéricos que simulam, por exemplo, a atmosfera, o oceano, a superfície terrestre e o gelo marinho. Atualmente, os MCGA são usados para uma variedade de propósitos, desde estudos da dinâmica do clima até projeções decenais e centenárias do clima futuro (Valcke, 2013). A principal limitação dos MCGA é o alto custo computacional porque acarreta em grande número de execuções, bem como de longas simulações.

Um dos modelos climáticos acoplados oceano-atmosfera que será utilizado é o ICTP-AGCM (Tabela 2.3), constituído pelo modelo climático global atmosférico SPEEDY, versão 41 Kucharski et al. (2006, 2016) acoplado ao modelo oceanico Nucleus for European Modeling of the Ocean (NEMO) Madec et al. (2015), com o modelo acoplado OASIS3 (Valcke, 2013). O componente atmosférico é executado na resolução horizontal T30 e têm oito níveis na vertical. O modelo inclui parametrizações fisicamente baseadas de condensação em grande escala, convecção superficial e profunda, radiação de ondas curtas e ondas longas, fluxos superficiais de momento, calor e umidade e difusão vertical (Molteni, 2003; Dee et al., 2015).

O NEMO é um modelo oceânico de equações primitivas de nível-z baseado nas aproximações hidrostática e de Boussinesq. Esta versão utiliza uma resolução horizontal de 2° e um refinamento tropical de $0,5^\circ$. A componente oceânica possui 31 níveis verticais com espessuras de camada variando de 10 m na superfície a 500 m no fundo do oceano, sendo 16 níveis nos 200 m superiores. Detalhes adicionais do ICTP-CGCM são descritos por Justino et al. (2015) e Kucharski et al. (2016).

Tabela 2.3 – Descrição dos Modelos ICTP-AGCM e CCCM3/IBIS

	ICTP-CGCM	NEMO	CCM3-IBIS
Características			
Domínio	Global	Regional / Global	Global
Resolução Espacial	$3,75^\circ$ (lat) x $3,75^\circ$ (lon)	2° (lat) x 2° (lon)	$2,8^\circ$ (lat) x $2,8^\circ$ (lon)
Níveis verticais	8 níveis (925 – 30 hPa)	31 níveis (10 – 5000 m)	18 níveis (925 – 30 hPa)
Coordenadas	Sigma	Hibrido-sigma Z	Sigma

Neste estudo, foi usado um modelo de biosfera global conhecido Integrated Biosphere Simulator (IBIS) que representa processos biofísicos, ecológicos e hidrológicos em várias escalas espaciais, desde a microescala a global (Foley et al., 1996b; Kucharik et al., 2000) (Tabela 2.3). O IBIS é um modelo global de vegetação dinâmica que simula mudanças na composição e estrutura da vegetação em resposta a condições ambientais (Delire et al., 2002). Classificado como um modelo de vegetação dinâmica que compreende processos diversificados como interações entre a superfície e atmosfera, o modelo de superfície constitui de dois dosséis de vegetação, sendo árvores de dossel superior, além de arbustos e gramíneas no dossel inferior. As trocas energéticas (fluxos de radiação de ondas curtas e longas; calor latente e sensível) e o momento entre a atmosfera e superfície dependem do estado de cada dossel. No IBIS, os processos físicos e ecológicos são representados em uma hierarquia de intervalos de tempos horários, mensais e anuais.

As mudanças na dinâmica da vegetação, incluindo os Tipos Funcionais de Plantas (PFTs) são determinadas no final de cada ano, exceto para a fenologia foliar de PFTs decíduos que é atualizada diariamente. As PFTs estão agregadas as diferenciadas características fisiológicas, como as árvores e gramíneas (fisionomia básica), perenes e decíduas (hábito das folhas), fotossintéticas (C3 e C4) e o formato da folha (folha larga ou aguda).

A competição entre os PFTs é responsável pela estrutura de dois tipos de estratos: árvores de dossel superior captam radiação de ondas curtas com mais facilidade devido a altura da copa, mas gramíneas (plantas herbáceas) têm acesso preferencial a água em razão de suas raízes terem maiores proporções nas primeiras camadas do solo. Entretanto, plantas lenhosas tem o poder de absorção de água das camadas mais profundas do que as plantas herbáceas ao simular o balanço hídrico. Portanto, a cobertura vegetal de uma célula de grade individual caracteriza-se pelo somatório de área foliar e biomassa favorecida por cada PFT.

- o módulo de superfície da terra usa duas camadas de vegetação;
- o módulo da vegetação opera com passo de tempo diário;
- o módulo do carbono calcula a fotossíntese;
- a dinâmica da vegetação é calculada no passo de tempo anual.

Duas camadas de vegetação são representadas: tipos funcionais de plantas arbóreas (árvores) de dossel superior e PFTs herbáceas de dossel inferior. Os PFTs na camada superior da vegetação são capazes de capturar a luz primeiro e sombrear o dossel da vegetação inferior; entretanto, a vegetação da camada inferior é capaz de absorver a umidade do solo primeiro. Diferentes parâmetros de fisiologia da planta (ou seja, fotossíntese e parâmetros de respiração) e parâmetros de dinâmica da vegetação são adotados para diferentes PFTs para calcular a biomassa viva e o índice de área foliar com um conjunto de equações de processo bioquímico. Diferentes fenologias de vegetação também estão presentes em cada PFT. A competição entre PFTs é impulsionada por diferenças no balanço anual de carbono, que resultam de diferentes estratégias ecológicas (Foley et al., 1996b).

O segundo modelo atmosférico que será utilizado é o CCM3 (Community Climate Model, version 3), desenvolvido no NCAR (National Center for Atmospheric Research) dos Estados Unidos. Foi integrado com resolução de 2,8° de latitude e longitude (T42), e, na vertical, apresenta um sistema de coordenadas híbrido pressão-sigma com 18 níveis (Kiehl et al., 1998). Este modelo é acoplado com uma versão atualizada do modelo de superfície conhecido como Integrated Biosphere Simulator (IBIS) (Foley et al., 1996a).

O modelo acoplado CCM3-IBIS é usado para produzir as interações bidirecionais entre a superfície da terra e a atmosfera, importante para o estudo da distribuição da vegetação, função do ecossistema e feedbacks do clima no contexto de mudanças climáticas global e mudanças no uso da terra (Foley et al., 2000). Portanto, nesse estudo o modelo CCM3/IBIS fará uso das condições de fronteira proveniente do ICTP-CGCM que já foi rodado para o equilíbrio com os períodos interglaciais. A comparação dos resultados entre o CCM3-IBIS e o ICTP-CGCM permitirá a avaliação da capacidade do último, que tem uma complexidade menor no que tange a simular vegetação, em reproduzir padrões de vegetação sob condições climáticas tão distintas das atuais.

A validação das simulações climáticas (em termos de clima e vegetação) será feita através da comparação com reconstruções de superfície. Estas são as simulações bases a serem empregadas no CCM3/IBIS como condições iniciais de superfície.

Os modelos de vegetação dinâmica são de importância impar para o entendimento da biosfera planetária, Embora, assim como todo modelo numérico, há incertezas, como já tem

sido relatado pela comunidade científica. Pelo fato de serem configurados em esquemas de tipo PFTs apresentam anti-realismo no que diz respeito a vegetação dinâmica. No entanto, os esquemas do tipo de planta funcional obtiveram resultados positivos de desempenhos em simulações da vegetação dinâmica global sob condições de mudanças climáticas, assim como reconstrução de paleo-vegetação (Yang et al., 2015).

Uma das razões mais importantes para a reconstrução de modelos de dinâmica da vegetação é que eles podem ajudar a prever cenários quando um determinado ambiente sofre pequenas mudanças e/ou perturbações que podem provocar mudanças drásticas na vegetação (Pausas, 1999).

Com passar do tempo, e desenvolvimento de modelos avançados da dinâmica de vegetação global, assim como de seus acoplamentos com a superfície terrestre, os modelos climáticos melhoraram suas capacidades de representar de forma mais realista o sistema dinâmico acoplado da interação superfície-atmosfera, levando em conta a resposta dos ecossistemas terrestres quanto seu impacto na variabilidade natural do clima e/ou por meio de forçantes antrópicas. A resposta da vegetação dinâmica com variabilidade climática vem sendo estudada há várias décadas. Permito-me citar, como exemplo, Charney (1975), em que o autor verificou que o efeito da redução da vegetação sobre o clima pode ter desencadeado um feedback positivo, contribuindo para a persistência da seca do século XX no Saara.

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ARTIGO 1

Southern Hemisphere Monsoonal System During Superinterglacial Stages: MIS5e, MIS11c and MIS31

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Abstract: Based on coupled climate experiments conducted with the coupled ICTP-CGCM model focusing on interglacial stages MIS5e (127 ka), MIS11c (409 ka) and MIS31 (1,072 ka), the Austral summer monsoonal system is investigated. The interannual variability and intensity of monsoon events are analysed from vorticity indices and air-sea interaction

processes for Africa, Australia and South America monsoons. Results demonstrated with respect to present day conditions, an orbital driven decrease in precipitation in summer, but slightly shift of the onset and demise periods of monsoons. Sensitivity experiments indicate, furthermore, that the monsoons are forced not only by external factors such as the dominant effect of insolation, but also by distant climate anomalies, such as surface temperature of the equatorial Atlantic and Pacific basins. During the interglacial stages, cooling occurs in the Southern Hemisphere whereas Northern Hemisphere substantially warms, inducing remarkable changes in the position of the oceanic subtropical high pressure systems and equatorial convergence zone. Regionally, these mechanisms contribute to periods of drought, with reduced precipitation rate over sectors of the Amazon and Northeastern Brazil, northern Australia and southern Africa. Monsoonal rainfall shows different responses to precessional forcing, as well as the relationship between the monsoon and Niño 3.4 differs among the interglacial stages. Compared to current climate, correlation analyses have demonstrated weaker influence of the equatorial Pacific Ocean on the Austral summer monsoon for the MIS5e, MIS11c and MIS31. Exception is noticed for the South America monsoon, where the Niño 3.4 exerts a more prominent role in these distant intervals, in consonance with the tropical Atlantic in particular during the MIS5e.

Keywords: Climate changes. Sea Surface Temperature. Interglacials. Marine Isotope Stages

3.1 Introduction

Modelling studies that investigate the impact of changes in seasonality on global heat content provide an opportunity to understand past and future climate features, in particular the monsoonal system, which is highly dependent on surface thermal conditions (Diaz and Boos, 2021; Zeng and Zhang, 2020; Deng et al., 2018; Yin and Berger, 2012). Global precipitation is primarily dominated by summer monsoon characteristics, which have been found to be highly correlated with the annual cycle of solar insolation, as well as associated with changes in radiative forcing due to increased greenhouse gases (Geen et al., 2020; Li and Ting, 2017; Trenberth et al., 2000). Studies have demonstrated that global hydrological cycle becomes stronger in a warmer climate (Held and Soden, 2006), but additional scrutiny is needed to verify hydroclimatic changes under precessional inter-hemispheric seesaw temperature anomalies; with warmer Boreal latitudes and Austral cooling, as characteristics of interglacial stages (Justino et al., 2017; Melles et al., 2012; Braconnot et al., 2008; Lisiecki and Raymo, 2005). Indeed, past changes in the radiative forcing have been responsible for weakening the SH monsoons in the mid-Holocene as a result of decreased net energy input (D’Agostino et al., 2020).

Among interglacial stages, the Marine Isotope Stages (MIS) MIS5e (127 ka), MIS11c (409,000 ka) and MIS31 (1072 ka), for instance, are marked by Northern Hemispheric increases in summer oceanic and terrestrial temperatures. The MIS5e also known as the Eemian, experienced sudden changes in the dynamics of the atmosphere and ocean in response to strong insolation in the Northern Hemisphere (NH) but weaker in the Southern Hemisphere (SH) (Yin and Berger, 2012; Berger and Loutre, 1991; Members, 2006; Siccha et al., 2015; Bard et al., 1990). Additional findings also indicate that the northward migration of the Intertropical Convergence Zone (ITCZ) during the Eemian, favored more intense NH summer monsoon with respect to the present climate (Montoya et al., 2000).

The MIS11c (~ 409 ka BP) also experienced global high temperatures dictated by increases in Boreal insolation. However, the Asian summer monsoon (ASM) during MIS11c, as revealed by data from the Yongxing cave, China, does not differ remarkably from monsoon characteristics in the Late Holocene (Zhao et al., 2019). It should be noted that the intensity of the MIS11c warming, is regionally dependent (Yin and Berger, 2012; Oppo

et al., 1998; Rousseau et al., 1992), raising questioning on how MIS11c warming or slight cooling may have induced changes in local precipitation features. For instance, the East Asian summer monsoon (EASM) responds differently to strong summer insolation during the MIS13, contributing to strengthen significantly the summer precipitation in northern China but barely changing that in southern China (Yin et al., 2014). Additional effort might indeed be pursued to clarify these issues based on coupled modelling experiments, carried under drastically modified astronomical forcing.

The MIS31 represents one of the last interglacials ever recorded, being considered the warmest marine isotope stage interval that contributed to substantial melting of NH polar glaciers compared to the present day (Lisiecki and Raymo, 2005; Melles et al., 2012). Such changes also impacted the seasonality of NH atmospheric and oceanic circulation patterns, and El NIÑO Southern Oscillation (ENSO) intensification and periodicity, besides significant changes in the Meridional Overturning Circulation (MOC). Insofar as monsoonal patterns are concerned, it is found that the Indian monsoon was enhanced but no correlation with the ENSO is identified in the MIS31 climate, differing from conditions delivered by today's climate. It has also been found that monsoonal precipitation for this interglacial is more closely connected to hemispherical features than to the tropical–extratropical climate interaction (Justino et al., 2019). Oliveira et al. (2017) argue that low-latitude insolation forcing plays an important role during the MIS31 by inducing a mild and humid climate regime in the Mediterranean region, with reduced seasonality.

Despite large effort in characterizing in detail those past climatic features, the spatio-temporal patterns of SH monsoon related to changes in external factors is urgent needed. These monsoon characteristics are more noticeable during the austral summer, in which the inversion of the zonal wind component from east to west are highly correlated to significant increases in precipitation rates (Yim et al., 2014; Geen et al., 2020). The seasonal zonal wind shifts in the lower troposphere is a response to the thermal contrast between ocean and continent, which causes rainy summers and dry winters. Moreover, in some locations monsoonal precipitation is associated with climatic extremes, because leads to flash floods and notorious damage to the population, especially to most vulnerable people, in urban and agricultural zones (Ávila Díaz et al., 2016; Kundzewicz et al., 2014). The weakening of the monsoon on the other hand leads to the absence of water resources and extreme heat

during drought events (Deng et al., 2018; Ha et al., 2020). It is worth noting that most of agricultural activities in South America, Africa and Australia are highly dependent on monsoonal driven-precipitation. Therefore, investigation of the SH monsoon system under very different climates, as delivered during past interglacial, are very useful to foresee adaptative measures to cope with future changes in the anthropocene.

Thus, past interglacial temperature contrasts between land and oceanic regions, and the out-of-phase Sea Surface Temperature (SST) pattern induced by the precessional cycle, between the Northern and Southern Hemispheres might have dictated remarkable changes in seasonal precipitation. The SH monsoon is evaluated here based on a series of coupled climate model simulations conducted for different interglacial stages, including all forcing factors such as changes in orbital parameters and greenhouse gases characteristics of the epoch. In section 2 is describe the coupled model specifications and experimental modelling setup. Section 3 discuss results under different perspectives: (1) the control climate is compared to ERA5, (2) changes in the radiative components and temperature are analysed; and (3) the impact of those changes on atmospheric dynamics responsible for anomalous magnitude, onset and cessation of monsoonal precipitation during the interglacials, are shown. In session 4, conclusions are summarised in the light of previous investigations and paleo-proxies perspectives.

3.2 ICTP-CGCM and modeling experiment design

The present study applies the Global Coupled Atmosphere-Ocean Circulation Model developed at the International Centre for Theoretical Physics (ICTP-CGCM), which consists of the global atmospheric climate model "SPEEDY" (version 41) coupled to the Nucleus for European Modeling of the Ocean (NEMO) model (Madec et al., 1998; Madec, 2008), with the coupler OASIS3 (Valcke, 2013). In computational terms, SPEEDY proofs to be effective in reproducing the main characteristics of the climate system of tropical and extratropical latitudes (Molteni, 2003; Kucharski et al., 2006; Justino et al., 2021). The atmospheric component runs at eight vertical levels in T30 horizontal resolution (Kucharski et al., 2016). NEMO is an ocean model adapted for regional and global ocean circulation studies, based on 31 vertical levels with thicknesses ranging from 10 to 5000 meters of

the ocean floor, with 16 levels in the first 200 meters. The current version uses a tri-polar ORCA2 configuration with a horizontal grid resolution of 2° and a tropical refinement of 0.5° (Madec, 2012).

To investigate interglacial intervals three experiments integrated from 500 years after the spin-up were carried out (Table 3.1). Greenhouse gas concentrations, as well as the orbital parameters have been set according to (Coletti et al., 2015; Bereiter et al., 2015; Lüthi et al., 2008). The study centered the interglacial time at three fixed periods, 127 ka (MIS5e), 409 ka (MIS11c) and 1,080 Ma (MIS31), for this reason, the orbital forcing is characteristics of the vernal equinox as the control experiment (CTRL) climate. The CTRL is conducted under current orbital forcing and atmospheric conditions as described in (Justino et al., 2017, 2019). For all experiments the first 100 years of the simulation are not included in the analyses of results. Verification of the CTRL climate in reproducing present day conditions is based on comparison with the ERA5 database (Hersbach, 2016) from 1979 to 2020, insofar as characteristics of the tropical climatology of precipitation, zonal component (u), SST and Sea Level Pressure (SLP) are concerned.

3.3 Results

3.3.1 Differences between the CTRL simulation and MIS experiments

3.3.1.1 CTRL radiative components and temperature

Based on Table 3.1, it is noticed that no remarkable changes of the atmospheric composition exist among the interglacials, but they favour slightly cooling conditions with respect to CTRL, due to lower atmospheric CO_2 concentration and weaker greenhouse capacity of the atmosphere, during those past interglacial. These features contrast with modifications of the Earth's orbit, in particular the precessional cycle. Largest differences with respect to CTRL are noticed in the MIS31 resulting in very distinct climate (Justino et al., 2021; Coletti et al., 2015).

Figure 3.1 shows zonally averaged radiative components at the top of atmosphere (TOA) based on ERA5 and the CTRL simulation. This is important because main differences between the MIS experiments and CTRL primarily derives from changes in the radiative

Table 3.1 – Orbital configurations and greenhouse gases concentrations utilized in the CTRL, MIS5e, MIS11c and MIS31 experiments.

Experiment	Date	CO ₂ (ppmv)	CH ₄ (ppbv)	N ₂ O(ppbv)	Ecc.	Obl.	Prec.
CTRL	Present	380	801	289	0.01671	23.438	101.37
MIS5e	127 ka BP	287	724	262	0.03938	24.040	272.92
MIS11c	409 ka BP	285	713	285	0.01932	23.781	265.34
MIS31	1.072 ka BP	325	800	288	0.05597	23.898	289.79

forcing. Thus, verifying the ICTP-CGCM capability to reproduce ERA5 demonstrates whether the model is suitable for the proposed investigation.

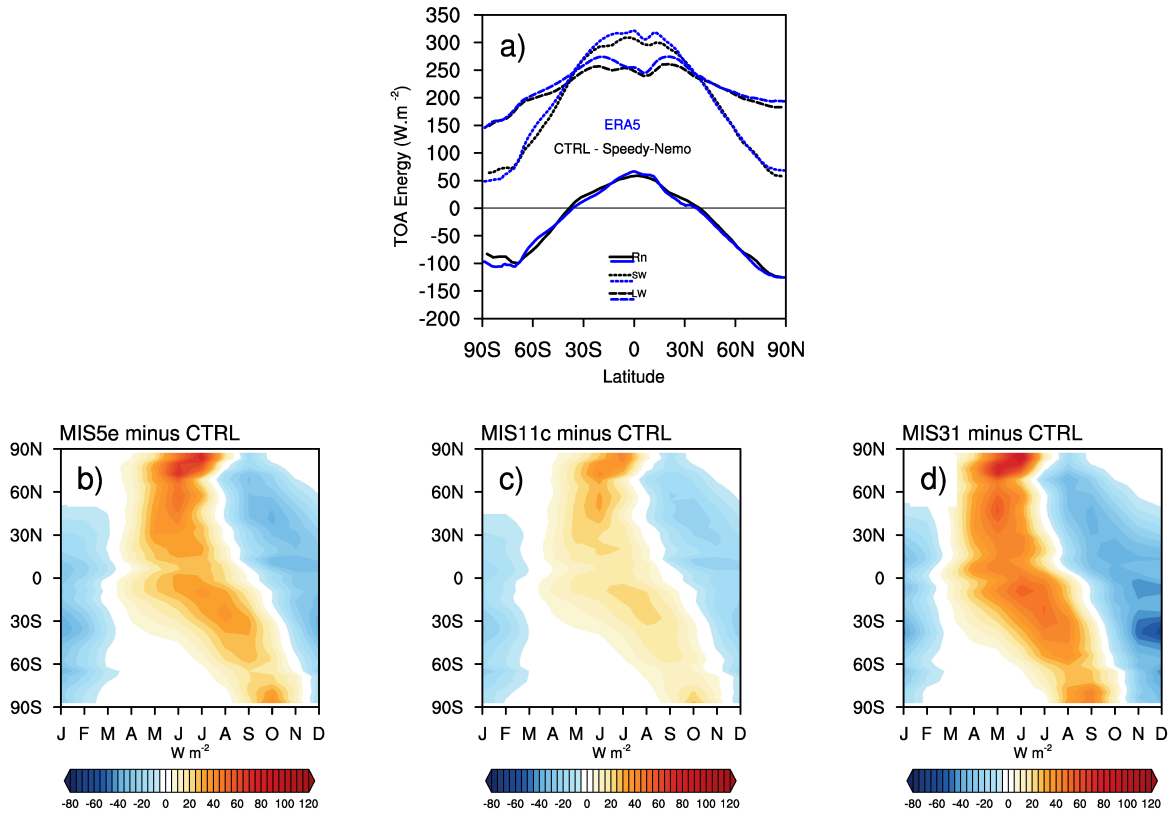


Figure 3.1 – Zonal mean energy budget components (Wm^{-2}) in the top-of-atmosphere (TOA)(a). (b) shows differences of shortwave radiation at the top of the atmosphere (W.m^{-2}) between MIS-5E minus CTRL, MIS-11C minus CTRL (c) and MIS-31 minus CTRL (d). Values in (a) are based on ERA5 reanalysis and the ICTP-CGCM coupled model.

Overall, the coupled model is able to represent the energy balance at the TOA and surface. However, the ICTP-CGCM underestimates ERA5 values about by 5% at the TOA (Fig. 3.1a). Good match between ERA5 and the model is found in the zonal distribution, but the presence of the Intertropical Convergence Zone (ITCZ) in the equatorial region,

leads to more pronounced differences across 10°S and 10°N . Atmospheric models running under lower resolution show drawbacks in reproducing meso-scale convective system in the tropics leading to slightly modified radiative balance. In the polar regions and extratropics as expected (Fig. fig1a), more energy is emitted to space via longwave radiation than is provided by the shortwave component. In general grounds, the modelled radiative budget (R_n) is well represented by assuming the meridional gradient of the radiative components, that in fact, is responsible for the large-scale atmospheric and oceanic circulations. It should be emphasised that estimates of the radiative balance has been for many years a challenging task associated with improvements of coupled models (Wild, 2020).

Figure 3.1b,c,d show differences of the short wave radiation at the top of the atmosphere, during the interglacials MIS5e, MIS11c and MIS31 with respect the CTRL climate. The influence of modified orbital parameters leads from June to November in the SH tropics, to an increase of 60 Wm^{-2} in the MIS5e and 20 Wm^{-2} in MIS11c (Fig. 3.1b-c). The most significant increase is, however, observed in NH where the shortwave radiation is larger than CTRL in summer and extends throughout the hemisphere. The MIS31 is among the interglacial periods the most affected by orbital changes from March to October. It should be stressed that from November to March, the SH experiences a drop in the amount of incoming energy, from October to February, which is the dominant period of the monsoonal features (Fig. 3.1b-d). This reveals that the general attribution for warmer climates in past interglacial should be taken in light of individual hemispheres and seasonal dependence.

Figure 3.2 shows that modified orbital parameters and subsequent changes of incident solar radiation result in anomalous near surface air temperatures. Differences during the interglacials with respect to current climate, are characterized by cooling from November to February (NDJF) across the continents (Fig. 3.2a-c-e). This gradually turns to warmer conditions in Boreal spring with largest anomalies found in summer, which are reproduced by annual mean anomalies (Fig. 3.2b-d-f).

The SH summer monsoons, which are dominant in NDJF months, can be severely affected by surface thermal changes not only causing a reduction in the amount of rainfall, but also by inducing distinct onset and demise periods. Surface condition are determinant for SH monsoons because soil moisture and evapotranspiration play an important role at early stages of the monsoon (Collini et al., 2008). In particular, when water vapour

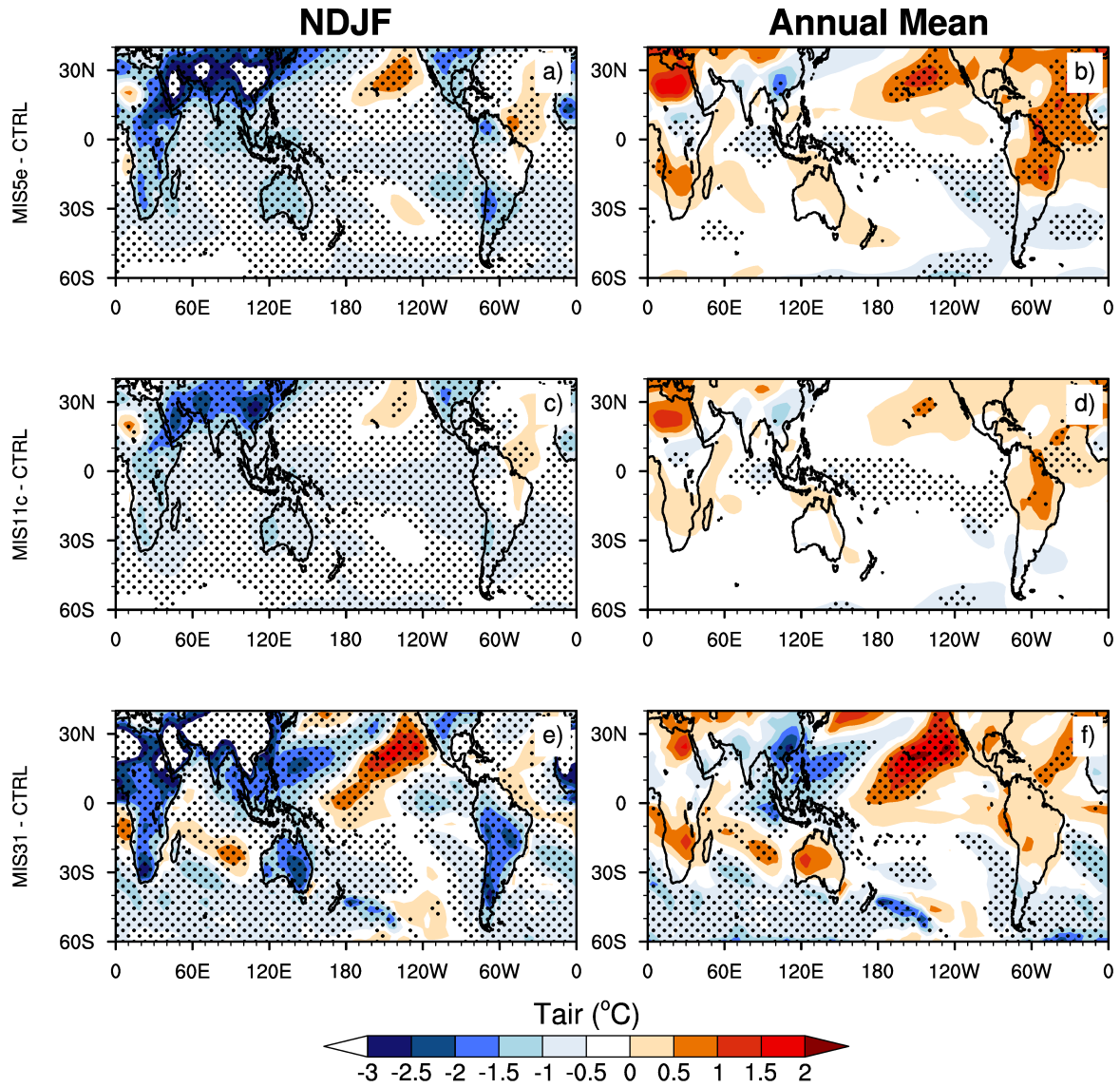


Figure 3.2 – NDJF mean surface temperature differences ($^{\circ}\text{C}$) between MIS5e and CTRL (a), MIS11c and CTRL(c), and MIS31 and CTRL (e). (b-d-f) is the same as (a-c-e) but for mean annual differences. Dotted regions are statistically significant at the 95% confidence level.

advection from surrounded monsoonal domain is weak. Indeed, lower temperatures are also a consequence of reduce energy and evaporation from the surface, causing a weakening in the convection of the monsoon system (D’Agostino et al., 2020).

These changes across the tropical region, which is an important source of humidity to the subtropics, can directly impact monsoonal characteristics. The substantial warming delivered over the continents as shown by annual time averaged (Fig. 3.2b-f), induces a northward migration of the ITCZ, as well as a weakening of the southern trade winds (Justino et al., 2019). It is not speculative to argue that these anomalous features will

be associated with modifications in the thermal ocean-land characteristics, impacting the monsoonal system.

3.3.1.2 Annual distribution of precipitation and the SH monsoonal characteristics

To identify the SH monsoons spatial domain, the amplitude of the first harmonic of precipitation has been applied (Fig.3.3a,b). The first harmonic characterizes regions where the annual cycle is dominant explaining more than 80% of the variance. This fits nicely with monsoonal features. Figure 3.3a depicts the first harmonic amplitude of precipitation from the 12 months climatology, based on a 100-year CTRL simulation. Harmonic analyses also known as Fourier transformation, are computed on a series of sine and cosine functions which may be used to verify differences in the precipitation magnitude between the rainy and dry seasons, for instance. When the seasonal cycle is dominant, the 1st harmonic explains most of the variance, which is the case in monsoonal regions.

The ICTP-CGCM is capable to reasonable simulate the amplitude over the South African Monsoon System (SAFMS), the Australian Monsoon System (AUSMS) and the South American Monsoon System (SAMS) regions (Fig. 3.3a). The CTRL experiment (Fig. 3.3b), however, overestimates the 1st harmonic amplitudes by 5 mm/day in particular across SAMS, in comparison to values delivered by ERA5 dataset (Fig.3.3a,b). Drawbacks in simulating tropical dynamics in coupled models, arise due to limitations in reproducing ocean-atmosphere interaction, as compared to processes in the extratropics, because based on intermediate complexity atmospheric components, convection, turbulent heat fluxes and diabatic processes are more difficult to be parameterized (Flato et al., 2014; Neelin et al., 1994). (Marengo et al., 2012)

Furthermore, it should be mentioned that the CTRL climate discussed here shows differences from ERA5, in part because time intervals used are not the case. The CTRL is based on a 100 years simulation forced with initial conditions correspondent to 1950. Certainly, these different intervals drive slightly distinct climates, also due to the ICTP-CGCM interannual and interdecadal variability, which may differ with respect to ERA5 conditions computed from 1979 to 2020.

The SH monsoonal regions, present a well-defined annual cycle, with rainy summers

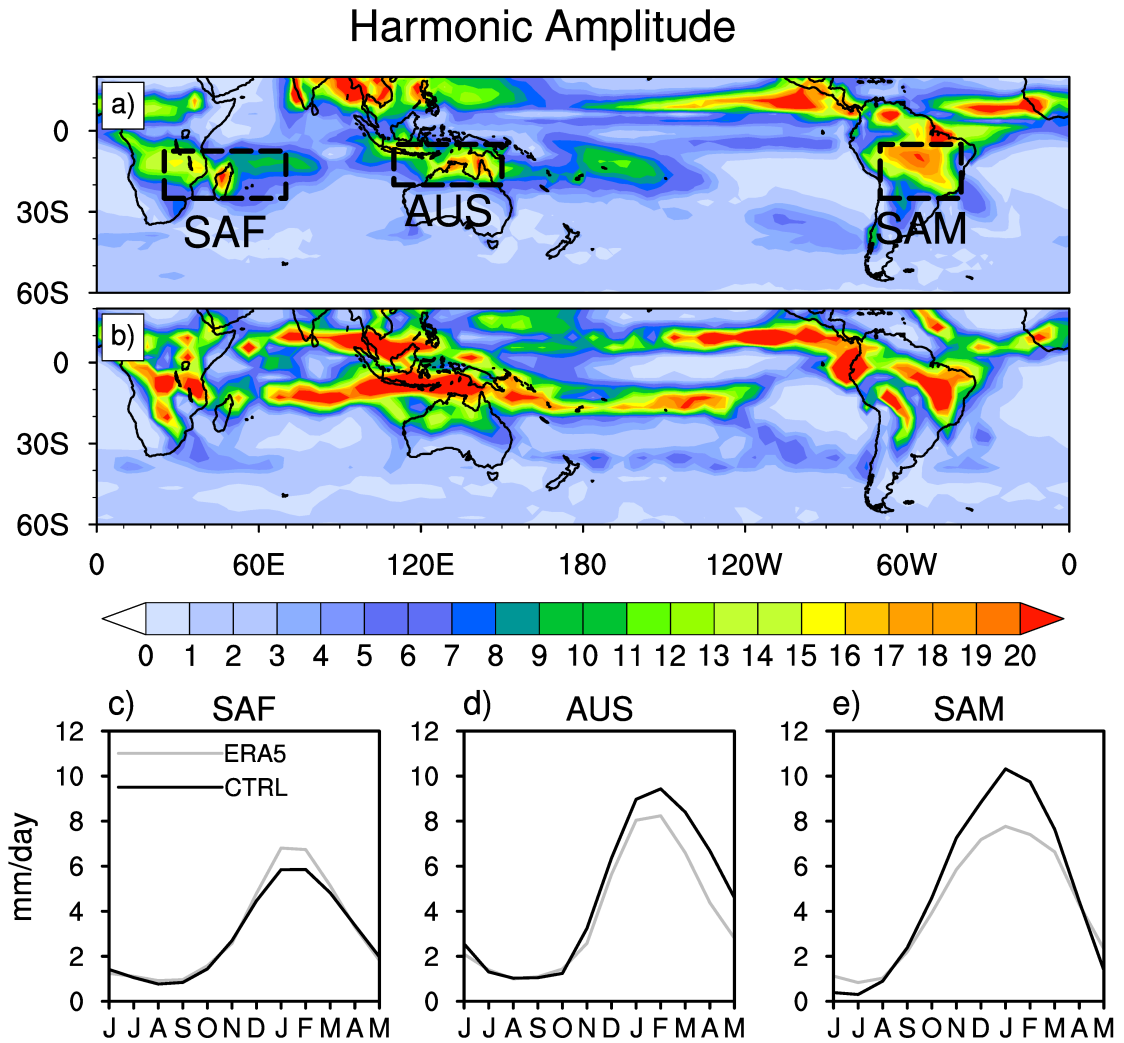


Figure 3.3 – Amplitude of the 1st harmonic based on ERA5 (a), and the CTRL simulation (b). (d) and (e) show the climatological annual cycle of areal averaged precipitation across the SAFSM, AUSSM and SASM domains (mm/day).

(DJFM) and dry winters (MJJA), as shown in Figure 3.3c-d-e. The CTRL climate delivers with respect to ERA5, higher precipitation as reproduced by areal averages over SAMS and AUSMS but lower values are found for SAFMS. Among the monsoons, higher daily precipitation amount is delivered for the AUSMS (≈ 10 mm/day), whereas lower values are observed in the SAFMS (≈ 6 mm/day), during the peak month. This is highlighted from November to March. Regionally, 70% of annual rainfall occurs during the NDJF months related to the monsoons. ERA5 shows along the SAMS, SAFMS and AUSMS domains, magnitudes by about 7, 9 and 8 mm/day (Fig. 3.3c-e).

Table 3.2 – Correlation coefficients between the regional monsoon precipitation in the rectangular domains and respective vorticity indices. Bold values indicate statistical significance at the 95% level.

Monsoon Domain	Monsoon (index)	ERA5 CC	CTRL CC	MIS5e CC	MIS11c CC	MIS31 CC
SAFSM (0-20°S,10°E-50°E)	$U_{850}(5^{\circ}\text{S}-15^{\circ}\text{S}, 20^{\circ}\text{E}-50^{\circ}\text{E}) - U_{850}(20^{\circ}\text{S}-30^{\circ}\text{S}, 30^{\circ}\text{E}-55^{\circ}\text{E})$	0,52	0,74	0,76	0,78	0,77
AUSSM (0-20°S,105°E-160°E)	$U_{850}(0-15^{\circ}\text{S}, 90^{\circ}\text{E}-130^{\circ}\text{E}) - U_{850}(20^{\circ}\text{S}-30^{\circ}\text{S}, 100^{\circ}\text{E}-140^{\circ}\text{E})$	0,93	0,86	0,72	0,74	0,73
SASM (5°S-25°S,70°W-40°W)	$U_{850}(5^{\circ}\text{S}-20^{\circ}\text{S}, 90^{\circ}\text{E}-130^{\circ}\text{E}) - U_{850}(20^{\circ}\text{S}-30^{\circ}\text{S}, 100^{\circ}\text{E}-140^{\circ}\text{E})$	0,80	0,37	0,50	0,65	0,31

Correlation analyses show reasonable correspondence between ERA5 and the CTRL run (Table 3.2), based on monsoon indices as represented by the zonal circulation and the areal averaged rainfall, values up to 0.93 and 0.74 are found for AUSMS and SAFMS, but much lower is delivered for SAMS.

This indicates that insofar as the SAMS is concerned, the vorticity index in the CTRL run is not able to individually represent precipitation changes related to the monsoon. The presence of the Andes, the recurrent cold fronts and the Amazonian contribution to the SAMS through evapotranspiration, are complex features that may weak the correlation between the vorticity index and SAMS precipitation. It may be argue that nearby oceanic processes exert an important role to define the SAM. These processes involve SST changes in the southwestern Atlantic Ocean, which impact in two-ways interaction the strength of the South Atlantic Convergence Zone (SACZ), by modifying the maritime water vapor and heat transports onto continental regions (Pezzi et al., 2022; Jorgetti et al., 2014). The SST impact on the monsoonal system is discussed in more detail in the following sections.

3.4 Monsoonal precipitation

In the interglacial experiments, changes in insolation during the interglacial lead to drastically modified monsoonal precipitation amounts (Fig. 3.4). Precipitation anomalies related with interglacial periods (Fig. 3.4a-f) deliver distinct pattern associated to each

monsoon domain.

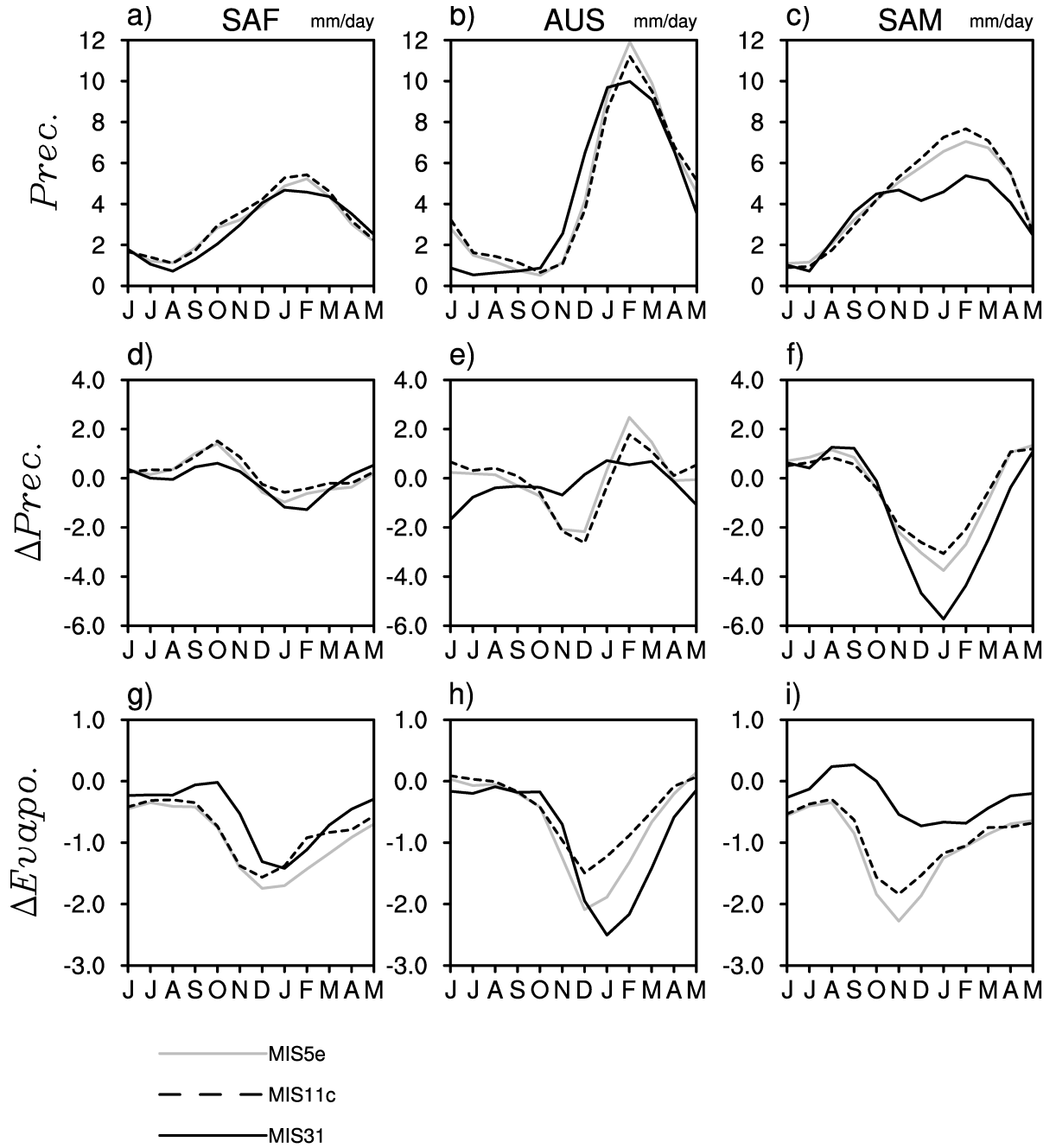


Figure 3.4 – Climatological annual cycle of precipitation for SAF (a), AUS (b), and SAM (c). (d-e-f) show differences of precipitation ($\Delta Prec.$) between the MIS5e, MIS11c, and MIS31 with respect to CTRL conditions. (g-h-i) are the same as (d-e-f) but for evapotranspiration (mm/day, $\Delta Evapo.$).

Results show that increased precipitation occurs in the dry season (June, July and August) in particular for SAMS and SAFMS (Fig. 3.4c,f). These features are enhanced in the MIS11c and MIS5e, and may represent a slightly shift in the rainy period over the two monsoons domains. In AUSMS, larger changes are depicted from October to February,

with a drop in rainfall (Fig. 3.4e), however, therefrom there exist an enhancement of the monsoonal precipitation until April. The interglacial driven by strongest orbital radiative forcing, the MIS31, delivers remarkable changes with respect to the CTRL, with substantial reduction of precipitation in the SAMS, by more than 60% (Fig. 3.4f). Reduction is also found for the other SH systems, the weakening of the SAFMS precipitation occurs during the summer months, whereas for the AUSMS drier conditions are found from June to December with respect to the CTRL climate (Fig. 3.4d,e).

Changes in surface conditions such as shown by evapotranspiration (ET, Fig. 3.4g-i) in line with those depicted by precipitation in the SAFMS, indicating the importance of interaction between surface processes and precipitation, in particular soil moisture. Analyses for the other two SH monsoon do not demonstrate a close relationship between ET and precipitation. In the AUSMS, enhanced monsoon rainfall is accompanied by reduced ET, which lead to assume that the AUSMS during these interglacial, experiences large contribution from water vapor transport originating from outside the monsoonal domain. The pattern over the SAMS shows a time lag relationship, in which the peak in reduction of ET leads by two months negative anomalies of precipitation in the MIS11c and MIS5e. In the case of MIS31, changes in ET seems to influence in lower degree modification of precipitation, with respect to the other interglacials (Fig. 3.4g-i).

The Hovmöller diagram (Fig. 3.5) based on the 850hPa zonal wind shows onset and demise of the rainy season related to the monsoon, through the inversion of the zonal wind; eastward winds in the dry season and westward winds more dominant for the rainy season. In general, changes in atmospheric circulation via the inversion of the zonal wind are well simulated by the ICTP-CGCM (Fig. 3.5a-b-c).

In terms of individual monsoons, close pattern is depicted for the SAFMS and AUSMS, by indicating an east-west wind shift during the summer months (Fig. 3.5a-b). A different pattern emerges for the SAMS, where wind transition is not clearly seen. However, in the SAMS wind features are longitudinal depend, and confined to the eastbound of the monsoon from 60°-30°W (Fig. 3.5c). Despite reasonably reproducing the monsoon annual cycle, the ICTP-CGCM does not relate the wind features to precipitation in the SAMS domain. This is associated with the topographic barrier imposed by the Andes, which channels throughout the year the easterly Atlantic flow, hampering the seasonal shift, as

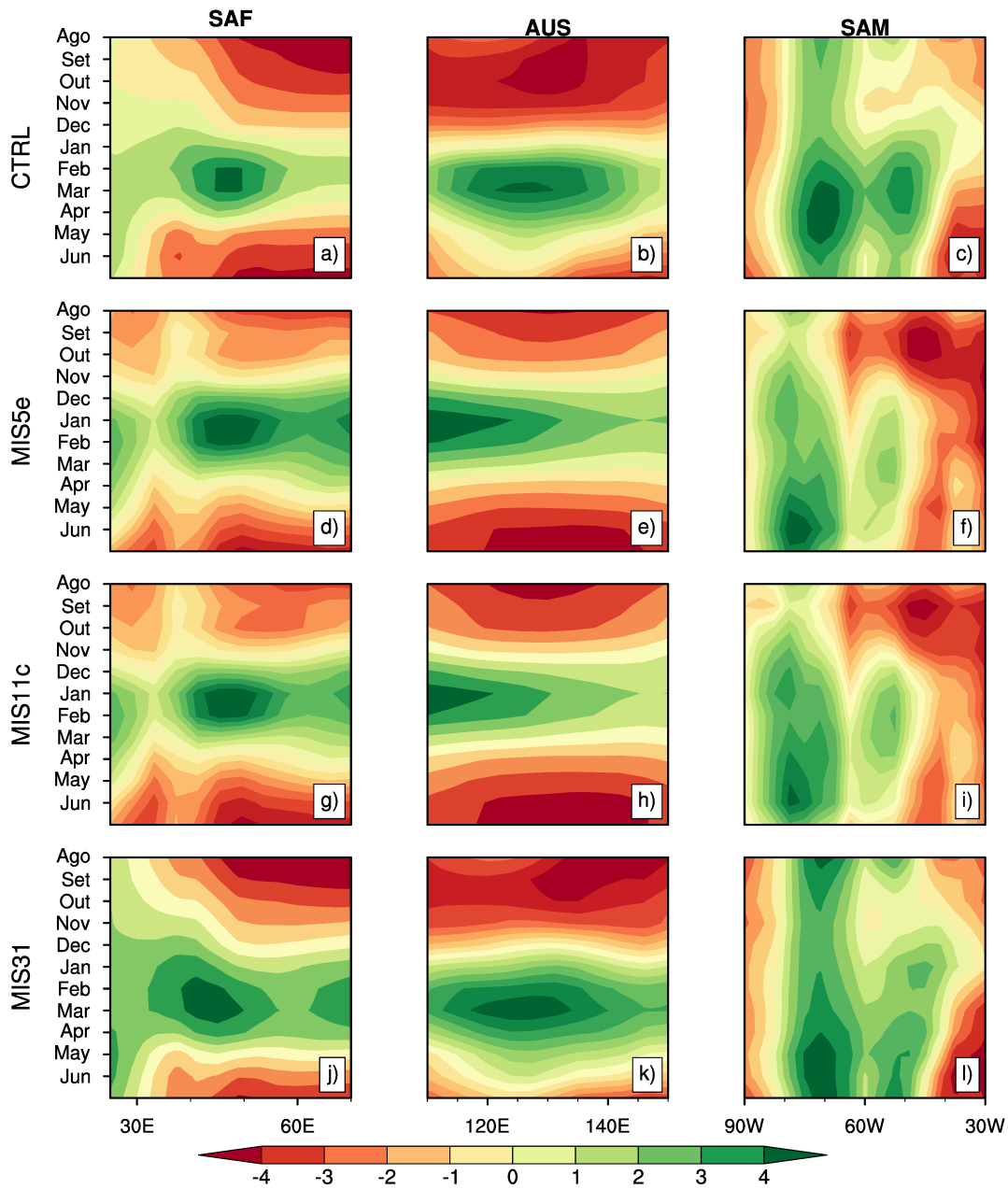


Figure 3.5 – Longitude-time cross-section of climatological zonal wind (m/s) at 850 hPa averaged between latitudes 7.5°S–25°S (SAF), 5°S–20°S (AUS) and 5°S–25°S (SAM).

discussed by (Nogués-Paegle et al., 2002).

The MIS5e and MIS11c deliver a semi-annual component that is most evident in the AUSMS domain. This characteristic found in the SAFMS and AUSMS, is absent in the MIS31, which exhibits a well defined annual cycle (Fig. 3.5j,k). Indeed, stronger annual cycle in the MIS31 has been found by Justino et al. (2019). Weaker interannual ENSO may acts in line to enhance the orbital forcing. The SAMS is different because it experiences

large changes related to the implementation of the orbital forcing (Fig. 3.5f,i,j). For this monsoon, the MIS5e and 11C differ substantially from the CTRL with increased zonal wind shear, and subsequently increased convergence over 45°W , from December to March (Fig. 3.5j,i). Moreover, it is found drastically modified circulation from August to November, with respect to CTRL. Strong westerly flow is evident in the MIS5e and MIS11c climates from $60\text{--}30^{\circ}\text{W}$. This results from intensified sub-tropical Atlantic high.

In order to evaluate in more detail mechanisms responsible for interglacial monsoonal changes, it is shown in Figure 3.6, the correlation pattern between the SAFMS, AUSMS and SAMS vorticity indices and SST changes. It highlights that similarities between the SAFMS and AUSMS under CTRL conditions, do not propagate through interglacials stages (Fig. 3.6a-h). The MIS5e exhibits a pattern, in particular across the Pacific Ocean, that is between CTRL and the other more distant interglacials, MIS11c and MIS31. For instances, both monsoon seem to be well correlated with Pacific and Indian Ocean SST in the CTRL climate, however, in MIS11c and MIS31, the SAFMS is highly related to Indian Ocean only (Fig. 3.6c,d). In the AUSMS case, changes over the warm pool Pacific region are dominant insofar as SST are concerned because weak correlation are found over the NIÑO (Equatorial Pacific) region (Figs. 3.6g,h).

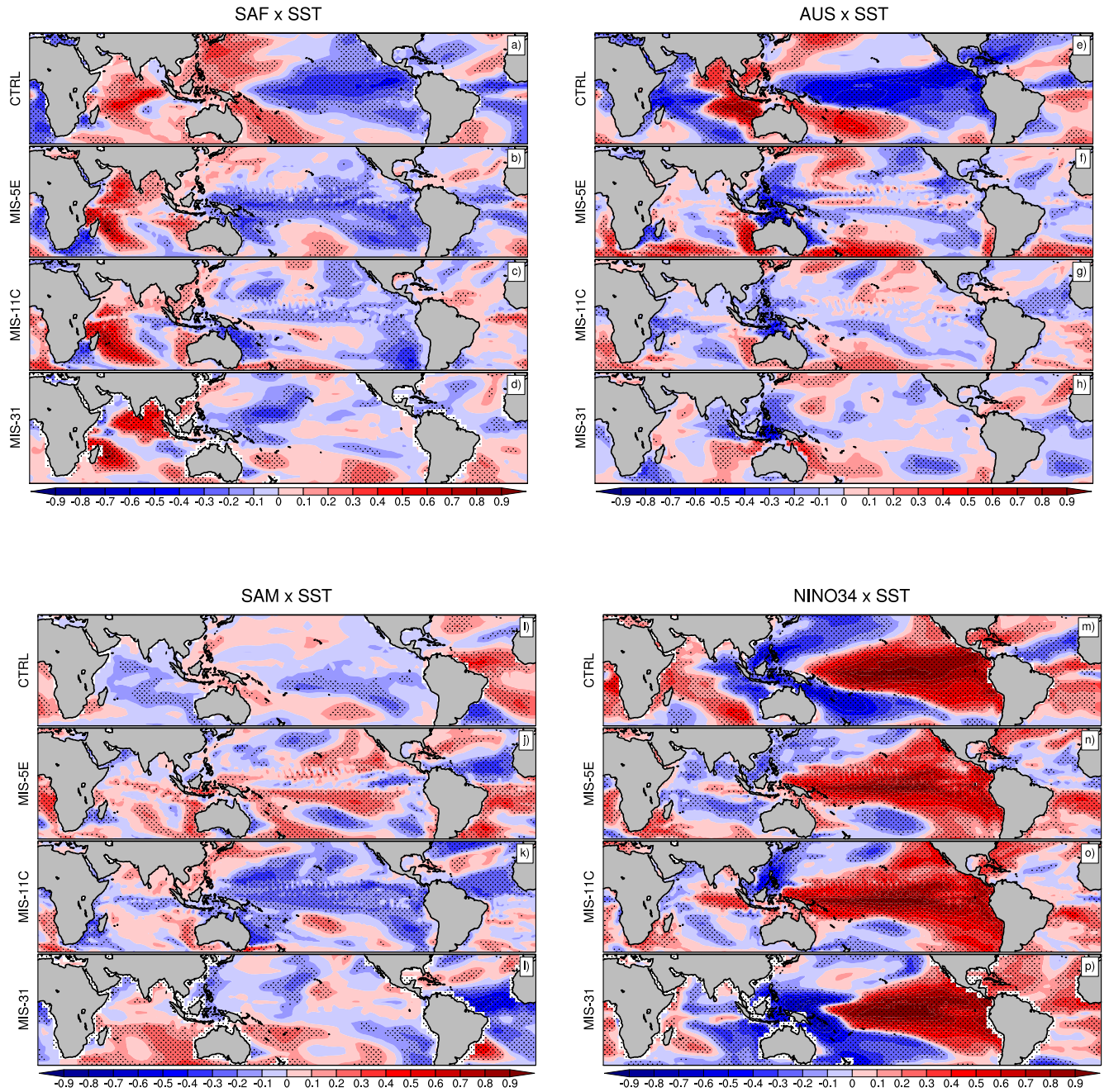


Figure 3.6 – Correlation coefficient between regional summer monsoons precipitation indices and SST for SAF (a-d), AUS (e-h) and SAM (i-l). (m-p) display the correlation between the Niño 3.4 and SST anomalies. Also shown is the 95% significant areas (dotted grids).

As demonstrated by previous discussion, the SAMS (Fig. 3.6i-l) differs substantially from the other monsoons. The Atlantic Ocean exerts an important role which is dependent upon the interglacial investigated. Indeed, in CTRL conditions higher positive correlations are found between precipitation and warmer Atlantic SST, under La Niña-like background

(Fig. 3.6i). Turning to MIS5e, the presence of the Atlantic variability (TAV, (Cabos et al., 2019) is enhanced with respect to CTRL, with a well defined dipole (Fig. 3.6j).

This SST correlation pattern indicates a southward position of the ITCZ, due to northern Atlantic cooling. Positive correlation also in the southwestern Atlantic is other feature that indicates intensification of the ZCAS. The relationship between SST and SAMS precipitation for the MIS11c and MIS31 demonstrated that the positive correlation is weaker in the equatorial southern Atlantic in the MIS11C, whereas an overall anticorrelation is found in the MIS31 (Fig. 3.6i). In general, the MIS31 exhibits contrasting values with respect to the CTRL (Fig. 3.6i,l), insofar as the SAM is concerned.

The eastern tropical Pacific exhibits a strong relationship with continental tropical rainfall, particularly due to ENSO characteristics. Table 3.3 indicates that during El Niño events, monsoonal precipitation of the three analyzed sectors is reduced with larger changes for AUSM in both ERA5 and in CTRL. These significant correlations, however significant, also reveal that NINO34 can drive interannual variability over the summer monsoons for the MIS5e and MIS11c, with exception of the AUSSM sector. On the other hand, the AUSSM is negatively correlated with ENSO in MIS31 at a 95% confidence level. Although, the spatio-temporal variability patterns of rainfall are different for each interglacial, it depends on the intensity of individual ENSO events. Because NINO34 may be projected as a remote forcing, it changes local thermodynamic and dynamic processes influencing climate and weather at long distances (McPhaden et al., 2006).

Table 3.3 – Correlation coefficients between Niño 3.4, TAV indices and the regional monsoon precipitation. Bold values indicate statistical significance at the 95% level.

		ERA5		CTRL		MIS5e		MIS11c		MIS31	
Austral Summer (DJFM)											
Index		Niño 3.4 TAV		Niño 3.4 TAV		Niño 3.4 TAV		Niño 3.4 TAV		Niño 3.4 TAV	
Monsoon Domain											
SAFSM		-0.40	-0.25	-0.36	0.26	-0.33	0.14	-0.27	0.1	-0.14	0.13
AUSSM		-0.70	0.26	-0.56	0.16	-0.05	0.01	-0.04	-0.14	-0.30	0.04
SASM		-0.30	0.30	-0.20	0.18	-0.29	-0.33	-0.23	-0.11	-0.03	-0.15

Also in this study, we examined the correlations between SH continental monsoon rainfall and Tropical Atlantic Variability (TAV). Under CTRL conditions most correlations are not significant. Turning to the interglacial behavior, the analyses demonstrate correlations

are weaker with respect to ENSO, but significant over the SAMS in the MIS5e. Interesting is that correlations between TAV and SAMS (-0.33) are higher than correlations based on ENSO (-0.29) (Table 3).

3.5 Spatial patterns of Precipitation and SLP

Figures 3.7 and 3.8 show differences in precipitation and SLP (interglacial minus control) overlaid with significant correlation (dotted) between the regional monsoon vorticity index; and precipitation and SLP.

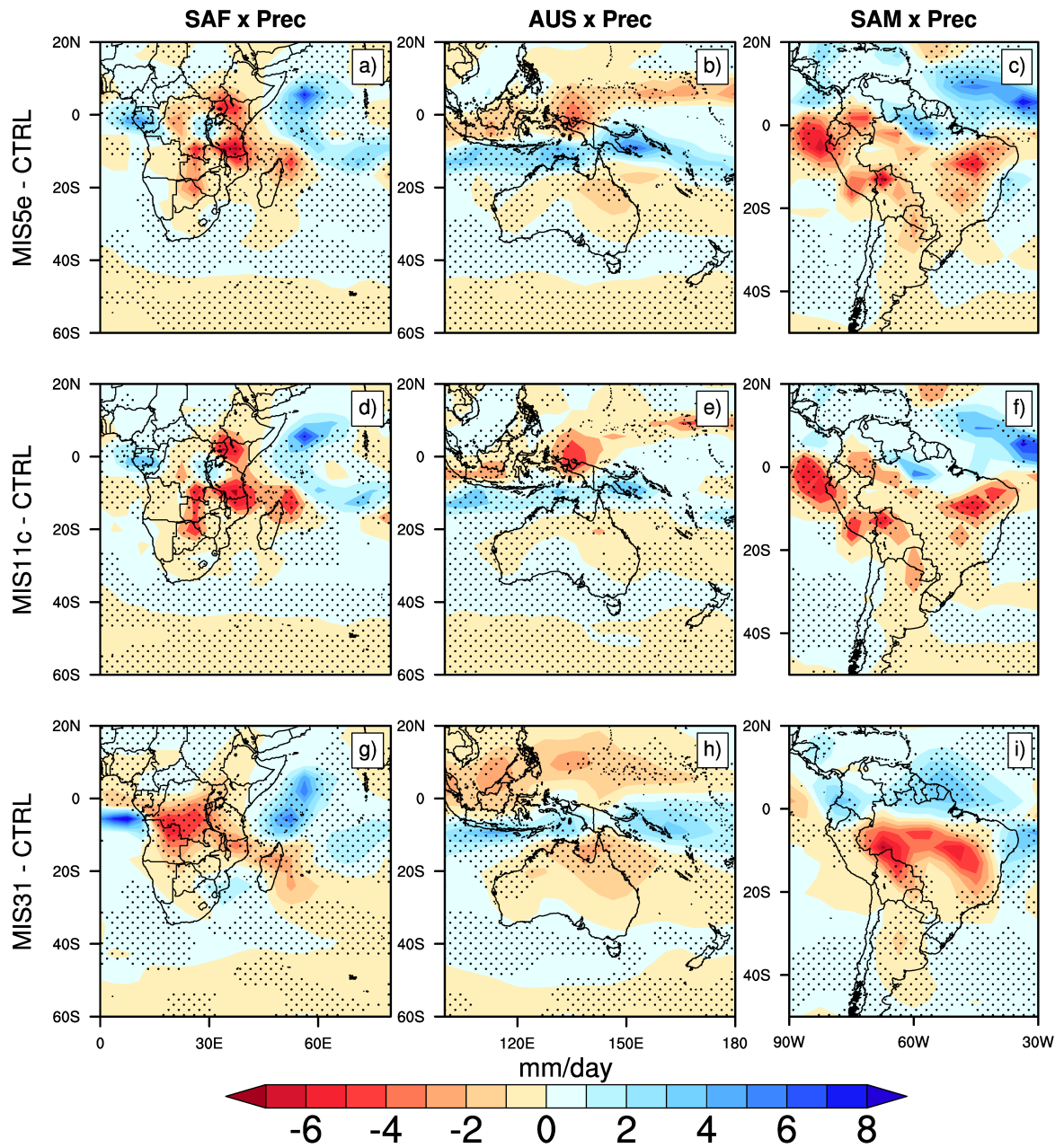


Figure 3.7 – Precipitation anomalies between MIS5e minus CTRL (a-b-c); MIS11c minus CTRL (d-e-f) and MIS31 minus CTRL (g-h-i). Dots indicate significant correlation between the regional monsoonal vorticity index and precipitation.

Analyses of SLP are important because may indicate regions of low and high pressure systems within monsoon domains in the lower troposphere. This can be further used to find convergence and divergent flows, which may induce or suppress precipitation.

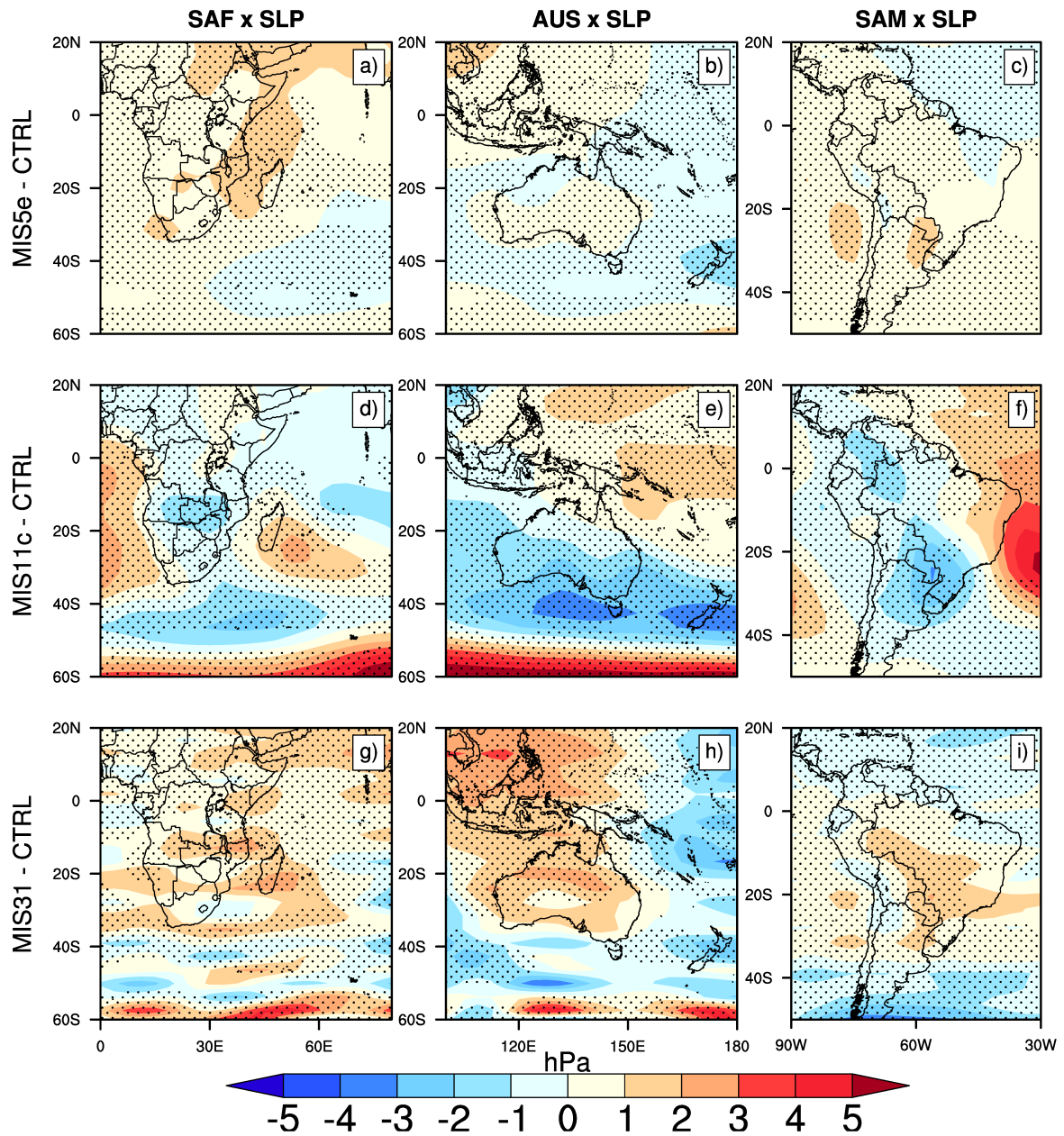


Figure 3.8 – Same as figure 3.7 but for SLP.

During the DJFM season for MIS5e and MIS11c, there was a reduction in rainfall in southern/southeast Africa and much of South America with values from -2 to -5 mm/day (3.7(a-f)), with respect to CTRL. Precipitation increases are found in the Indonesian Archipelago and southern Australia by up to 4 mm/day (Fig. 3.7(b-e)). The dominant effect of the orbital forcing in all 3 interglacial is highlighted as reduced precipitation across most of continental regions. In the MIS31, reduced rainfall in Africa and South America show values between -2 mm/day and -6 mm/day (Fig. 3.7g-i). Rainfall increases in northern Amazon and semiarid Northeast Brazil (3 mm/day), but there is no evidence

that this is linked to monsoonal changes. It should be noted that precipitation changes in Australia are lesser affected by the implementation of the orbital forcing. Most anomalies range between ± 2 mm/day, and in the MIS31, are not statistically significant.

Analyses of differences of SLP may clarify causes of precipitation changes. The SLP anomalies between the interglacials experiments (MIS5e, MIS11c e MIS31) and CTRL, exhibit statistically significant increased pressure in line with precipitation reduction over SAFMS, AUSMS and SAMS regions (Figures 3.8). Differences of MIS11c and CTRL, on the other hand, show slight reduction in SLP across southern Australia and south-central Brazil (Figures 3.8e,f). Those low pressure anomalies induce precipitation (Fig. 3.7) in Australia by enhancing convection and upward motion. It has to be noticed that this drop in SLP during the MIS11c in southern Brazil/South America, theoretically would increase the frequency of subtropical frontal systems, passing onto the SAMS region, but the high pressure anomaly associated with the subtropical Atlantic high, leads to a blocking situation hampering the subtropical system to move northwards. Thus, leading to dryness in most of Brazil during the interglacial. In the MIS31, most striking feature is related to cooling in the western Pacific by up to -3°C , in particular over the AUSMS domain. This results in positive SLP anomalies characteristics of permanent El Niño conditions (Fig. 3.2c,e and 3.8h), and reduced monsoonal precipitation (Fig. 3.7h).

3.6 Intercomparison between modeled precipitation and proxies data

Paleoclimate simulations are able to simulate Earth's climate in distant past, but comparison with reconstruction are still necessary to reveal model-proxies differences, in regions where models may struggle to represent particular atmospheric/oceanic features. Thus, indicating climate mechanisms that should be addressed in more detail (Braconnot et al., 2012). Justino et al. (2017) provides an evaluation of modeled MIS31 surface temperature and proxies, in which demonstrated that the ICTP-CGCM is able to reproduce reconstructions, differing only by $\pm 1^{\circ}\text{C}$, but larger values up to 3°C are found in the extratropics. These large differences have been attributed to reduced NH seaice cover, and therefore higher SSTs.

The three interglacials are warmer (colder) in the NH (SH) than the CTRL climate

(Table 3.4). This seesaw feature is related to changes in the magnitude of seasonal insolation. It is also worth mentioning that the ICTP-CGCM model simulates lower temperatures than reconstructions primarily in the Austral extratropical region for the MIS5e and MIS11c. On the other hand, these two interglacials show warmer temperatures in the NH ($2-6^{\circ}\text{C}$). This peculiar warming in the NH during MIS5e and MIS11c may indicate the model characteristics in reproducing the seaice extent, highlighted by increased cover in the SH but widespread reductions in sea ice across the NH. These changes in high latitudes are able to modify the meridional thermal contrast, impacting the atmospheric circulation and local turbulent oceanic fluxes, which can result in distinct SSTs.

Table 3.4 – Maximum values of SST records and modeling results for each interglacial. R represents values of reconstruction. $\Delta_{SST} (^{\circ}\text{C})$ shows differences between ICTP-CGCM results and R.

Site	Coordinates	MIS5e R Δ_{SST}	MIS11c R Δ_{SST}	MIS31 R Δ_{SST}	References
Lake E	67°N-172°E	14.5 [-0.32]	12.2 [1.3]	14.3 [-1.8]	Melles et al. (2012)
ODP 982	57°N-15°W	16.2 [0.71]	15.0 [1.49]	13.8 [-3.0]	Lawrence et al. (2009)
DSDP 552s	56°N-23°W	15.1 [2.31]	16.4 [0.05]		Ruddiman et al. (1986)
DSDP607	41°N-31°W	29.1 [1.4]		17.5 [-0.6]	Raymo et al. (1996)
306-U1313	41°N-32°W			18.0 [-1.1]	Naafs et al. (2013)
ODP 1020	41°N-126°W	14.1 [6.41]	14.0 [6.37]		Herbert et al. (2001)
DSDP 607s	56°N-32°W	25.1 [-0.53]	26.8 [-3.42]		Ruddiman et al. (1989)
ODP 1012	32°N-118°W	19.5 [-0.05]	19.1 [5.9]		Liu et al. (2005)
ODP 1146	19°N-116°E	27.3 [3.94]	26.8 [3.73]	26.0 [-1.0]	Herbert et al. (2010a)
ODP 722	16°N-60°E	27.7 [3.51]	27.5 [3.36]	27.0 [1.0]	Herbert et al. (2010b)
ODP 1143	9°N-113°E	28.8 [3.14]	28.3 [3.0]	28.3 [-0.8]	Li et al. (2011)
ODP 871	5°N-172°E			29.3 [-0.4]	Dyez and Ravelo (2014)
HY04	4°N-95°W	27.2 [1.34]	26.3 [2.19]		Horikawa et al. (2010)
MD97-2140	2°N-141°E	29.5 [0.14]	29.5 [0.58]		de Garidel-Thoron et al. (2005)
ODP 806B	0.3°N-159°E	29.6 [1.03]	30.2 [-0.08]		Medina-Elizalde and Lea (2005)
ODP 847	0-95°W			25.6 [-0.6]	Medina-Elizalde et al. (2008)
ODP 849	0-110°W			25.8 [-0.8]	McClymont and Rosell-Melé (2005)
ODP 846	3°S-90°W	25.1 [0.69]	24.0 [2.49]	24.3 [0.5]	Herbert et al. (2010c)
MD-06-301	23°S-166°E			25.0 [-1.1]	Russon et al. (2011)
ODP 1087	31°S-15°E			18.0 [-0.3]	McClymont et al. (2005)
ODP 1123	41°S-171°W	17.7 [-0.61]	19.3 [-3.25]	16.0 [0.8]	Crundwell et al. (2008)
ODP 1090	43°S-9°E	17.1 [-2.16]	13.9 [0.14]	11.5 [-1.7]	Martínez-García et al. (2010)
MD06-2986	43°S-168°E	18.0 [-3.12]	18.1 [-2.51]		Hayward et al. (2012)
DSDP594	45°S-175°E	18.3 [-4.9]	17.5 [-5.85]		Schaefer et al. (2005)
PS75/034-2	54°S-80°W	10.3 [-0.03]	8.8 [2.19]		Ho et al. (2012)

Reconstruction also exhibits caveats because in some cases they may reproduce long

term changes that are dictated by a particular seasonal strength, such as demonstrated by (Liu et al., 2014). This argumentation is not used to justify differences between model and reconstructions, but usefully serves to indicate that reconstruction represent local conditions that can be smoother in a grid box representation. Hydroclimatic reconstructions from MIS5e and MIS11c are shown in Table 3.5 and Figure 3.9, across the 5°N - 33° domain. This allows to identifying the precipitation and evaporation relationship, and therefore atmospheric humidity in both time slices, with respect to CTRL climate. By evaluating proxies for individual latitudinal bands divided by 10° shows that between 5°N - 5°S , drier conditions prevailed during NDJF months in the MIS5e, with exception of African sites in reconstruction and model results (red circles in Fig. 3.9).

These conditions are present not only during the monsoonal season but it is distributed throughout the year (Fig. 3.9a,b,c). From 6°S to 15°S most reconstructions are marked by increased humidity (blue circles in Fig. 3.9, differing from simulated values, that during the monsoons show dryer conditions, in particular in central Africa and Australia. Agreement is found over Indonesian Archipelago, east Africa and the Andes. However, by comparing reconstructions and model results, good agreement between these data is found for the MAMJ and JASO months, which indicates that reconstructions primary reproduce enhanced precipitation in Australia, Andes and eastern Africa, from March to October during the MIS5e (3.9b,c). Southward to 16°S most proxies are dominated by wetter conditions in Australia and southern Africa.

Some conditions persist along the months, such as the weakening of the ITCZ, that is supported by both modelled precipitation and reconstructions. Brazil and equatorial eastern Africa are also characterized by dryer and wetter conditions, respectively. Little information is available for MIS11c regarding moisture availability, proxies from Africa and South America allow us to infer that during MIS11c conditions were wetter than the current climate. For MIS31 there is scarce information which hampers to derive conclusions from comparison between model outputs in this study. But Justino et al. (2019) argue that the link between the Niño 3.4 and the Australian monsoon is weakened with respect to the CTRL characteristics, and indicates that changes in AUSM during the MIS31 are more closely connected to hemispherical features than to the equatorial climate interaction.

Table 3.5 – Hydroclimate reconstruction and modeled values for MIS5e, MIS11c and MIS31.
Values indicate wet or dry conditions in reconstructions.

Site	Coordinates	MIS5e	MIS11c	MIS31	References
GeoB4411-2	5°N - 44°W	Wet			Govin et al. (2014)
Gunung Mulu	4°N - 114°E	Dry			Carolin et al. (2016)
KS 84067	4°N - 4°W	Wet			Frédoux (1994)
MD03-2707	2°N - 9°E	Dry			Weldeab et al. (2007)
TR163-19	2°N - 90°W	Dry			Lea et al. (2000)
TR163-22	0.5°N - 92°W	Dry			Lea et al. (2006)
Lake Magadi	0.1°S - 36°E	Wet			Owen et al. (2018)
MD01-3340	0.3°S - 128°E	Dry			Dang et al. (2015)
ODP 1239	0.4°S - 82°W	Wet	Wet		Rincón-Martínez et al. (2010)
Lake Naivasha	0.5°S - 36°E	Wet			Trauth et al. (2001)
M16772	1°S - 11°W	Dry			Abrantes (2003)
Lake Challa	3°S - 37°E	Dry			Moernaut et al. (2010)
ODP 1077	5°S - 10°E	Dry			Uliana et al. (2002)
Cueva del Diamante	5°S - 77°W	Wet			Cheng et al. (2013)
SO139-74KL	6°S - 103°E	Dry			Lückge et al. (2009)
MD98-2152	6°S - 104°E	Wet			Windler et al. (2019)
GeoB1401-4	6°S - 9°E	Wet			Gingele et al. (1998)
GeoB1008-3	6°S - 10°E	Wet			Govin et al. (2014)
DPDR-I, DPDR-II	7°S - 108°E	Wet			Van der Kaars and Dam (1995)
MD05-2925	9°S - 151°E	Wet			Lo et al. (2017)
ODP 1229	10°S - 77°W	Wet			Contreras et al. (2010)
GLAD7-MAL05-1	11°S - 34°E	Dry	Wet		Ivory et al. (2016)
MD01-2378	13°S - 121°E	Wet			Kawamura et al. (2006)
Altiplano	16°S - 68°W	Wet			Placzek et al. (2013)
Lake Titicaca, LT01-2B	16°S - 70°W	Dry			Gosling et al. (2008)
Lynch's Crater	17°S - 145°E	Wet			Moss and Kershaw (2007)
MD96-2094	19°S - 9°E	Dry			Stuut et al. (2002)
Gregory(Mulan)	20°S - 127°E	Wet			Bowler et al. (2001)
Salar de Uyuni	20°S - 67°W	Dry			Fritz et al. (2004)
Kalahari - Makgadikgadi	20°S - 25°E	Wet			Burrough et al. (2009)
M125-55-7/8	20°S - 38°W	Dry			Hou et al. (2020)
Coastal Cordillera	21°S - 70°W	Dry			Ritter et al. (2019)
GeoB3911-1	21°S - 36°E	Dry			Dupont and Kuhlmann (2017)
MD00-2361	22°S - 113°E	Dry			Stuut et al. (2014)
MD08-3167	23°S - 12°E	Wet			Collins et al. (2014)
Pretoria Saltpan	25°S - 28°E	Wet			Partridge et al. (1997)
MD96-2098	25°S - 13°E	Dry			Daniau et al. (2013)
MD96-2048	26°S - 34°E	Dry			Caley et al. (2018)
Frern Gully Lagoon	27°S - 153°E	Wet			Kemp et al. (2020)
Lake Eyre	28°S - 137°E	Wet			Magee et al. (2004)
ODP1085	29°S - 13°E		Wet		Dickson et al. (2010)
KT-LE	29°S - 137°E	Wet			Miller et al. (2016)
Frome	30°S - 139°E	Wet			Miller et al. (2016)
CD154-10-06P	31°S - 9°E	Dry			Simon et al. (2015)
PA	32°S - 137°E	Dry			Miller et al. (2016)
Darling	33°S - 144°E	Wet			Miller et al. (2016)

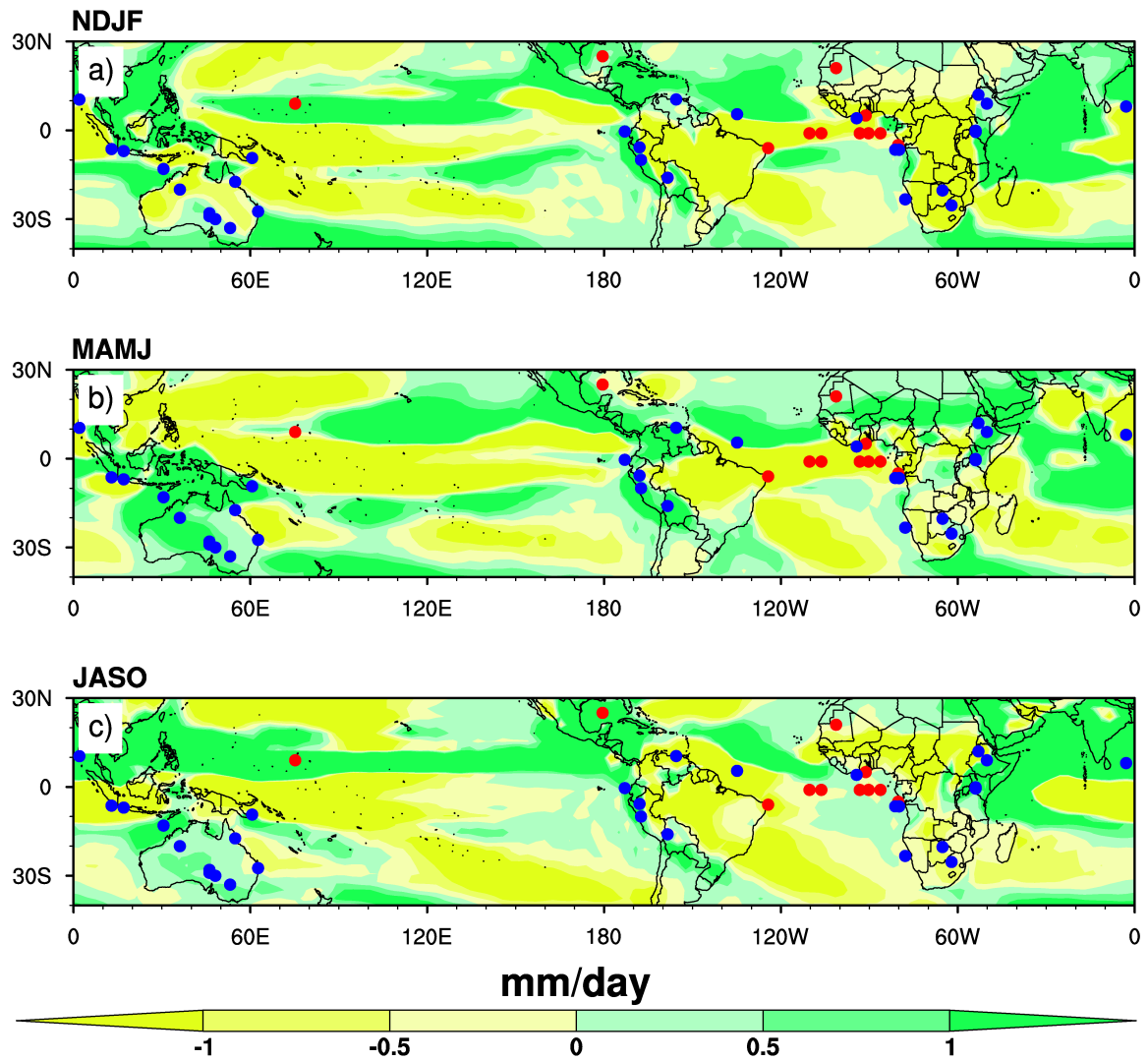


Figure 3.9 – Map showing precipitation differences between MIS5e and CTRL simulations overlaid by proxy records (wet-blue dots and dry-red dots) during the MIS5e, with respect the present day; for NDJF a), MAMJ b) and JASO c)

3.7 Concluding remarks

This study demonstrates that the Austral summer monsoons experienced strong changes during the MIS5e, MIS11c and the MIS31 intervals. This results due to changes in orbital insolation, which further modify the SSTs and global teleconnection patterns. Indeed, changes in the precessional cycle triggered significant cooling across Southern Oceans. In the SAMS region, for example, the MIS5e and MIS11c exhibit a decrease in the transport of moisture from the Amazon to southeastern South America. The contribution

of continental dry air mass quantitatively favored the decrease of summer monsoon rainfall, and the intensified reduction of SST during these interglacials, provided a northward displacement of the SASM, as well as of the SACZ.

Changes of monsoonal precipitation have also been delivered in terms of the month with maximum occurrence. Despite weaker precipitation in summer, the MIS5e and MIS11c stages are dominated by enhanced rainfall in the SH winter and early spring, for the SAFM, and winter/autumn for the AUSM. Thus, reduced precipitation is related to colder conditions during the SH summer, in particular over the continents. Correlation analyses indicate that large-scale processes, such as related to ENSO during the interglacials are weaker with respect to CTRL, in exception of increased significant correlation values over the SAMS.

Comparison between palaeoreconstructions and model simulations revealed a good agreement. It is worth noting that the palaeoproxies deliver the main response to a particular forcing, but does not indicate the preferential season of occurrence. Hence, for analysing the modeled precipitation it is necessary to figure out modeled monthly distribution, because as demonstrated in Figure 3.9, large differences in rainfall between the modelling experiments and CTRL, insofar as monsoonal precipitation is concerned, are found not only for the summer season, but substantial changes have also been found throughout the year.

Climate simulations aiming at reproducing the monsoonal system usefully serve to point out that enhanced Northern Hemisphere warming, can modify the temporal and spatial pattern of Southern Hemisphere precipitation. In a palaeoclimate perspective, this may be understood as analogues of conditions during Dansgaard–Oeschger events by means of the two hemispheres seesaw (Stocker and Johnsen, 2003). At some extent, global warming that also shows much warmer Boreal climate may also deliver modified SH monsoonal precipitation in line with interglacial conditions, as simulated here. Thus, Earth's future climate can be explored in detail based on climate experiments for past intervals stages. as based on coupled models of intermediate complexity as currently applied.

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ARTIGO 2

Orbitally-Driven Vegetation Changes During Past Interglacials: The Role of Sea Surface Temperature

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Abstract: Climatic feedbacks associated with orbitally driven Sea Surface Temperature (SST) lead to a profound impact on the Southern Hemisphere (SH) vegetation cover during the Mid-Holocene, MIS5e, MIS11c, and MIS31 interglacials. Results are based on a suite of coupled climate simulations conducted with the ICTP-CGCM (Speed-Nemo), which provides the boundary conditions for the CCM3/IBIS vegetation model. The CCM3/IBIS model was run from outputs Speed-Nemo of the global ocean and individualized Atlantic, Pacific, and Indian ocean basins forcing, in addition to orbital parameters, and greenhouse gases. For the interglacials, MH, MIS 5e and 11c, it has been found that areas areal a

significant reduction in the tropical evergreen forest, but more extensive savanna and grasslands, over parts of the southern and central African region, and into northern South America (Amazon region). In another important period, the results showed that have been changes that of vegetation cover due to insolation forcing during MIS31 when compared to other interglacials, but the impact was greater in Australia and central South America. On the other hand, the southern tropical climate became drier due to negative SST anomalies, induced by the Atlantic and Tropical Pacific basins, and thus reduced continental precipitation. It is further demonstrated that the response of the vegetation across the tropical region of the SH, is not strongly linked only to orbital variations, but modulated by the internal changes of Atlantic and Pacific SST. Over the study regions, in particular, during the MH, MIS5e and MIS11c, the seasonality of insolation was reduced in the SH, leading to the cooling of the southern tropical ocean basins, resulting in the migration of the summer precipitation zone to boreal latitudes. Therefore, combined with increased low-latitude summer temperatures and a prolonged dryer period were sufficient for the forcing led to a slow retraction, but steady of the rainforests of Congo and Amazonia causing an effect of extreme aridity.

Keywords: Tropical Vegetation. Interglacials. Sea Surface Temperature.

4.1 Introduction

During the last Pleistocene, the vegetation cover has undergone changes due to climate variability such as changes in the Earth's orbit, changes in ocean temperature and greenhouse gases (GHGs) (Lee et al., 2021; De Boer et al., 2021; Ganopolski and Calov, 2011; Lüthi et al., 2008). Previous studies using global circulation models (GCM) already indicated that vegetation-atmosphere feedbacks in response to orbital parameters were responsible for climate change during the last Pleistocene and Holocene (Foley et al., 2000; Claussen, 1994; Foley et al., 1994). Several approaches including paleoreconstructions have explored changes in global vegetation of past interglacial, such as during the MIS 5e and MH, which occurred by about 6000, 120000 years before present. However, the MIS 11c and 31 stages, that took place by 411000 years ago, and 1.08M have not been investigated in detail on a global perspective based on global climate simulations (Shugart and Woodward, 2011). The ocean plays an important role in vegetation-atmosphere interactions, as the distributions of rainfall and tropical vegetation cover are strongly influenced by the variability of SST (Kim et al., 2021). In fact, oceanic changes on a geological time scale, mediated by interglacial periods favor ecosystem changes worldwide, and due to modifications in climate feedbacks (Vázquez-Rivera and Currie, 2015) can shed light on Earth climate, that is currently experiencing remarkable changes in its thermal structure .

Changes in vegetation dynamics and climate variability have already been discussed by global political and scientific authorities in recent years (IPCC, 2013, 2014; Callaghan et al., 2020), once the impacts caused by climate change made the ecosystem more vulnerable (Brandt et al., 2017; Gao et al., 2018), the shrink of tropical vegetation leads to the reduction of evapotranspiration, resulting in the destabilization of the hydrological cycle. Despite the direct existence of the relationship between the main parameters of climate (temperature and precipitation) and vegetation cover (Wu et al., 2016), investigate the responses of vegetation dynamics to climate change associated with the past, offers us the opportunity to understand the distribution of landscape dynamics (Jamieson et al., 2012). According to a recent study, it indicated that photosynthetic activity will continue to decrease under conditions of increasing temperature and water deficit, especially in regions

with climates vulnerable to drought (Adepoju et al., 2019). Consequently, the climate impact on populations that live in a situation vulnerable to poverty and that depend on subsistence agriculture, especially rural populations, will be much greater because there are still no strategic adaptation plans for future extreme events (Olsson et al., 2014).

Based on several climate modeling experiments, previous studies have reported past climate variations involving vegetation-precipitation during the Middle Holocene, in which precession-induced insolation dominated that epoch, causing SST elevation in the global tropical region, northward migration from African summer monsoons and increased undergrowth in Sahara and Sahel regions (Collins et al., 2013; Kropelin et al., 2008; Zhao et al., 2005), as well as the strengthening of the South American monsoon (Prado et al., 2013).

Paleovegetation studies conclude that climate changes and increase in CO₂ concentrations were responsible for changes in vegetation dynamics and composition across the globe (Nolan et al., 2018; Prentice et al., 2011). However, the oceanic component mostly governs energy storage in the climate system due to its high specific heat value and physical-chemical properties. Through coupled modeling, Braconnot et al. (2007) showed that African monsoon rain during the MH (≈ 6 kyr ago) depends on ocean feedbacks, ie a strong meridional temperature gradient of the Tropical Atlantic can induce changes in precipitation. The response to vegetation cover is induced feedbacks between the ocean and the atmosphere (Braconnot, 2004). In a study focused on the Last Glacial Maximum (LGM) (21.000 ka BP), Kubatzki and Claussen (1998), concluded that the climatic conditions of the SH are determined by ocean basins.

In this context, we selected four interglacials that have similarities in their orbital configurations, also because they portray the variability of climatic conditions and vegetation classification, namely, MH and the Marine Isotope Stages (MIS) 5e, 11c and 31 (1.072 million years). Interglacials MH and MIS's 5e, 11c and 31 are marked by the strong meridional gradient of global SST anomalies that are induced by the seasonal distribution of insolation at the top of the atmosphere due to variation in long-term orbital forcing (Berger, 1978, 2021; Smulsky, 2021). Therefore, for these four periods there is an analogous factor between them, which is the seasonal cycle of insolation being more intense in Northern Hemisphere (NH) and attenuated in SH due to the effect of the orbital

parameter of precession (Berger, 1988).

MH a was a period with a weaker seasonal cycle in the SH in response to the variation of orbital parameters, strong meridional gradient of SST from the Tropical Atlantic, contributing to the displacement of the ITCZ to the North Atlantic (Shin et al., 2006), low CO₂ when compared to the present day (Raynaud et al., 1993), in addition to having been important in the variation of the climate system (Berger, 1978), mainly in the modifications of the tropical vegetation of the SH (Braconnot et al., 2019).

MIS5e (127 kyr ago) also known as Eemian, experienced strong prolonged climate changes due to oceanic and atmospheric feedbacks in response to the seasonal insolation cycle (Yin and Berger, 2012; Berger and Loutre, 1991; Members, 2006; Siccha et al., 2015; Bard et al., 1990). When compared to global averages from SST paleoreconstructions, MIS5e was considered the warmest interglacial in the last 800 kyr ago (of PAGES, 2016). Although they have different nomenclatures, their definitions also differ. While MIS5e was based on marine oxygen isotopes, Eemian was derived from the change in global vegetation cover (Shackleton et al., 2003).

MIS11c (409 kyr ago) was also considered an interglacial with negative SST anomalies during the SH Austral Summer, but positive in the NH, from paleoreconstructions (Kandiano et al., 2017). Changes in equatorial cross meridional heat transport may have caused anomalous cooling in the South Atlantic during MIS11c (Riveiros et al., 2013). In a recent survey, Tzedakis et al. (2022) observed that the insolation forcing was weak when compared to other interglacials, however its prolonged duration was sufficient to reduce the Greenland and Antarctic sea ice sheets, causing sea level rise and smoothed release of CO₂.

MIS 31 was considered a super interglacial, being responsible for the melting of glaciers in Greenland when compared to the present day (Lisiecki and Raymo, 2005; Melles et al., 2012). Such changes in the climate system during MIS31 are a response to more extreme orbital combinations, that is, high values of precession and eccentricity were able to generate strong SST anomalies at high latitudes and reduction of sea ice during the boreal summer (Oliveira et al., 2017; Yin and Berger, 2012). MIS31 exhibits characteristics analogous to Eemian and MIS11 due to the reduction of the Antarctic ice sheet in response to the

$\sim 5^{\circ}\text{C}$ increase in SST in sectors of West Antarctica (Pollard and DeConto, 2009; Villa et al., 2012; Teitler et al., 2015). From simulations using a coupled model of intermediate complexity, an increase above 1°C was estimated during the northern summer (Justino et al., 2019).

In this paper, we investigate the dynamic vegetation response of tropical continental regions of the SH to SST variation of the different southern ocean basins, using two coupled CGCM models, the ICTP-CGCM (Speedy-Nemo) and the CCM3/IBIS. The study analyzes the interaction between atmosphere-continent-ocean, focusing on the parameters of precipitation, temperature and dynamic vegetation, generated from the simulations of the global dynamic vegetation model that were forced through the boundary conditions (SST) of individualized Speedy-Nemo basins, from the orbital forcing of interglacials MH e MIS's 5e, 11c e 31. To verify if the dynamic vegetation corresponds to the Mid-Holocene, we used the CCM3/IBIS simulation outputs to compare with the paleovegetation reconstructions of the BIOME6000 project (Prentice et al., 2000, 2011). The specific questions addressed here were the following: on the continental regions of the SH, specifically in monsoon areas, what is the magnitude of the changes suffered by the dynamic vegetation? In addition to the orbital factor, could climate variation on a millennial scale related to OSH changes have impacted the vegetation cover of tropical wet and dry environments? How does the vegetation pattern respond to combinations of climate system changes and individualized ocean basins?

4.2 Models and experiments

4.2.1 Models description

The present work applies the Global Coupled Atmosphere-Ocean Circulation Model from the International Center for Theoretical Physics (ICTP-CGCM), which consists of the global atmospheric climate model "SPEEDY" (versão 41), coupled to the Nucleus for European Modeling of the Ocean (NEMO) (Madec et al., 1998; Madec, 2008; Valcke, 2013). In computational terms, SPEEDY proves to be effective in reproducing the main characteristics of the climate system of tropical and extratropical latitudes (Molteni, 2003; Kucharski et al., 2006; Justino et al., 2021). The atmospheric component runs at eight

vertical levels in T30 horizontal resolution (Kucharski et al., 2016) (See Table 4.1).

Table 4.1 – Models description of this study

	ICTP-CGCM	NEMO	CCM3-IBIS
Characters			
Domain	Global	Regional / Global	Global
Spatial Resolution	3,75° (lat) x 3,75° (lon)	2° (lat) x 2° (lon)	2,8° (lat) x 2,8° (lon)
Vertical Levels	8 levels (925–30 hPa)	31 levels (10–5000 m)	18 levels (925–30 hPa)
Coordinates	Sigma	Hybrid-sigma Z	Hybrid-sigma

In order to study the impact of interglacial boundary conditions on vegetation changes, the CCM3-IBIS coupled model is used to reproduce the bidirectional interactions between the land surface and the atmosphere, which is crucial for characterizing global vegetation distribution, ecosystem function and climate feedbacks in the context of global climate and land use changes (Foley et al., 2000).

The CCM3-IBIS is based on the atmospheric model CCM3 (Community Climate Model, version 3), developed at the National Center for Atmospheric Research (NCAR, USA). It has been integrated with 2.8° latitude and longitude (T42) resolution, and 18-level pressure-sigma hybrid coordinate system Kiehl et al. (1998). This model is coupled with a 2.6.4 version of the surface model known as the Integrated Biosphere Simulator (IBIS) Foley et al. (1996, 2000); Kucharik et al. (2000). IBIS is a global dynamic vegetation model that simulates changes in vegetation composition and structure in response to environmental conditions Delire et al. (2002).

4.2.2 The experiments

The experiments consist of the control simulation (CTRL) conducted for the climate of 20th century (Figure 4.1), integrated from 500 years after the spin-up of the ICTP-CGCM model, taking as the reference climate state. Four sensitivity experiments have been performed to characterize intervals representing interglacials stages, MH, and MIS 5e, MIS 11c and MIS 31. For all experiments, the first 100 years of the simulation are not included in the current analyses of the results. These simulations apply boundary conditions characteristics of orbital changes and greenhouse gases concentration (CO_2 , CH_4 and N_2O),

as described in Coletti et al. (2015); Bereiter et al. (2015); Lüthi et al. (2008); Justino et al. (2017, 2019).

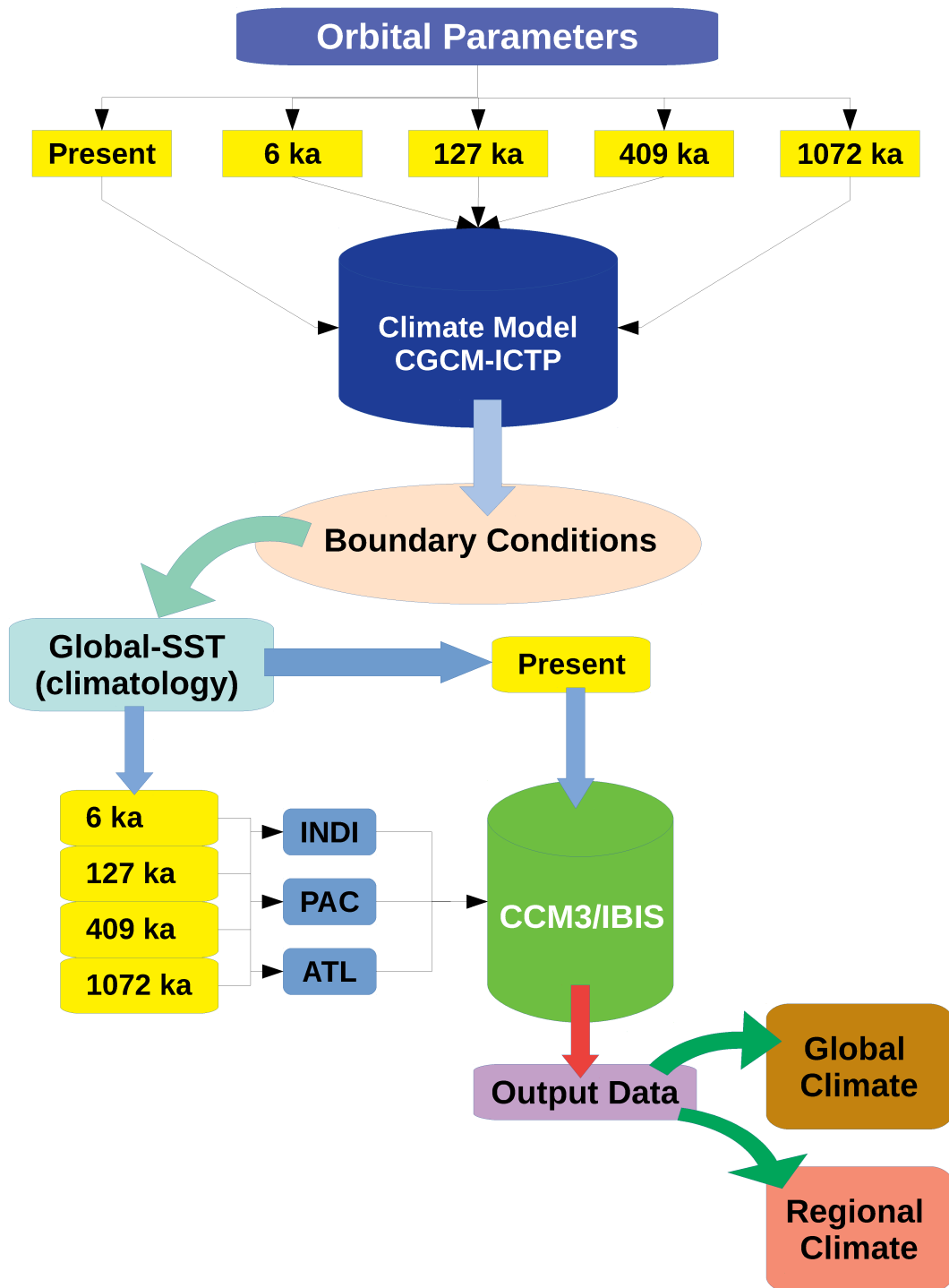


Figure 4.1 – Schematic illustration of global model systems to study climate change past by CGCM-ICTP in the framework of the stages of the integration. The second model illustrates the stages from coupled model atmospheric-vegetation CCM3/IBIS. Orbital Parameters: Present; 6 ka = Mid-Holocene; 127 ka = MIS5e; 409 ka = MIS11c and 1072 ka = MIS31. INDI = Tropical Indian Ocean, 30°N-30°S and 40°E-110°W; PAC = Tropical Pacific, 30°N-30°S and 120°E-280°W and ATL = Tropical Atlantic, 30°N-30°S and 300°E-360°W. See text and tables for more detailed descriptions.

After achieving the quasi-equilibrium state for those individual runs, additional experiments have been carried out with CCM3/IBIS forced by the CTRL, MH, MIS5e, MIS11c and MIS31 SST boundary conditions characteristics of these interglacial eras. For the third and final stage, a set of 12 forced simulations of the aforementioned interglacials were conducted, but using individualized ocean basins, following the criteria used by Pereira et al. (2014), namely: the Tropical Indian Ocean (IND, 30°N-30°S and 40°E-110°W), Tropical Pacific (PAC, 30°N-30°S and 120°E-280°W) and Tropical Atlantic (ATL, 30°N-30°S and 300°E-360°W). It is worth mentioning that each experiment consists of 1 round of 50 years. However, the first 10 years are ignored, and the last 40 years of each simulation are analyzed, in which the spin-up time of each experiment has been reached to obtain a more realistic representation of the climatology and vegetation. For more detail, the characteristics of the rounds are summarized in Table 4.2, and the experimental design is summarized in the Fig. 4.1.

Table 4.2 – Orbital configurations and greenhouse gases concentrations utilized in the CTRL, MH, MIS-5e, MIS-1c and MIS-31 experiments.

Experiment	Date	CO ₂ (ppmv)	CH ₄ (ppbv)	N ₂ O(ppbv)	Ecc.	Obl.	Prec.
CTRL	Present day	380	801	289	0.01671	23.438	101.37
MH	6 ka BP	280	700	275	0.01868	24.105	0.87
MIS5e	127 ka BP	287	724	262	0.03938	24.040	272.92
MIS11c	409 ka BP	285	713	285	0.01932	23.781	265.34
MIS31	1072 ka BP	325	800	288	0.05597	23.898	289.79

Verification of the CCM3/IBIS capacity to reproduce the dominant land cover type under present day conditions, is based on comparison with annual land cover map derived from MODIS dataset (MOD12C1) for year 2000 to 2020, with a resolution of 0.05 degrees (Hodges, 2002). This product contains the global vegetation classification scheme derived from the IGBP (International Geosphere Biosphere Programme).

4.3 Results and discussion

4.3.1 Surface temperature and precipitation differences

Time averaged surface temperature and precipitation differences are shown in Figure 4.2. Temperature differences between the interglacial stages and CTRL demonstrate that

negative anomalies are dominant features in the tropical ocean basins (Figure 4.2a-b-c), which are more pronounced in MIS 5e and 11c (Figure 4.2b-c) due to weak insolation over the SH in response to changes in precession. It has to mentioned that across the Indian Ocean and in most of the MIS31 the SST are higher than in the CTRL experiment.

Turning to the pattern over land, positive temperature anomalies are shown over northern SAM, CAF and SAF, and northern Australia in MH, MIS5e and MIS11c. Despite the warmer ocean, MIS31 over the Amazon and Australia does not shown significant changes of temperatures with respect to the CTRL run (Figure 4.2d). However, the pattern over Africa is characterized by warmer SH and colder NH, indicating the highly influential effect of precession.

Precipitation differences also shown in Figure 4.2 indicate an overall reduction in continental rainfall in the north-central SAM and south-central Africa, but with a slight increase over Australia, in the MH, MIS5e and MIS11c intervals (Figure 4.2e-f-g). Increased precipitation is also observed in Australia during the MIS31 (Figure 4.2h). The combination of climatic feedback such as the decrease in insolation and the reduction of the horizontal temperature gradient between the oceanic basins and continents in the SH, lead to the weakening of the monsoons in Africa, Australia and SAM, hampering the transport of water vapor from the ocean to the mainland during the rainy season. These changes in precipitation associated with the decrease, in turn, tend to reduce evapotranspiration, and consequently favor a positive feedback. The annual reduction in precipitation over the tropical region of the SAM may be associated with the weakening of evapotranspiration which in turn leads to the desintensification of deep convection in the Amazon (Wright et al., 2017). In summary, the effects of the rainfall pattern on Amazonia during interglacials depend on the magnitude of the differential warming between continent and ocean, once this warming becomes attenuated, the ascending branch tends to be weakened, carrying less water vapor to the high levels of the troposphere and reducing the release of latent heat through condensation.

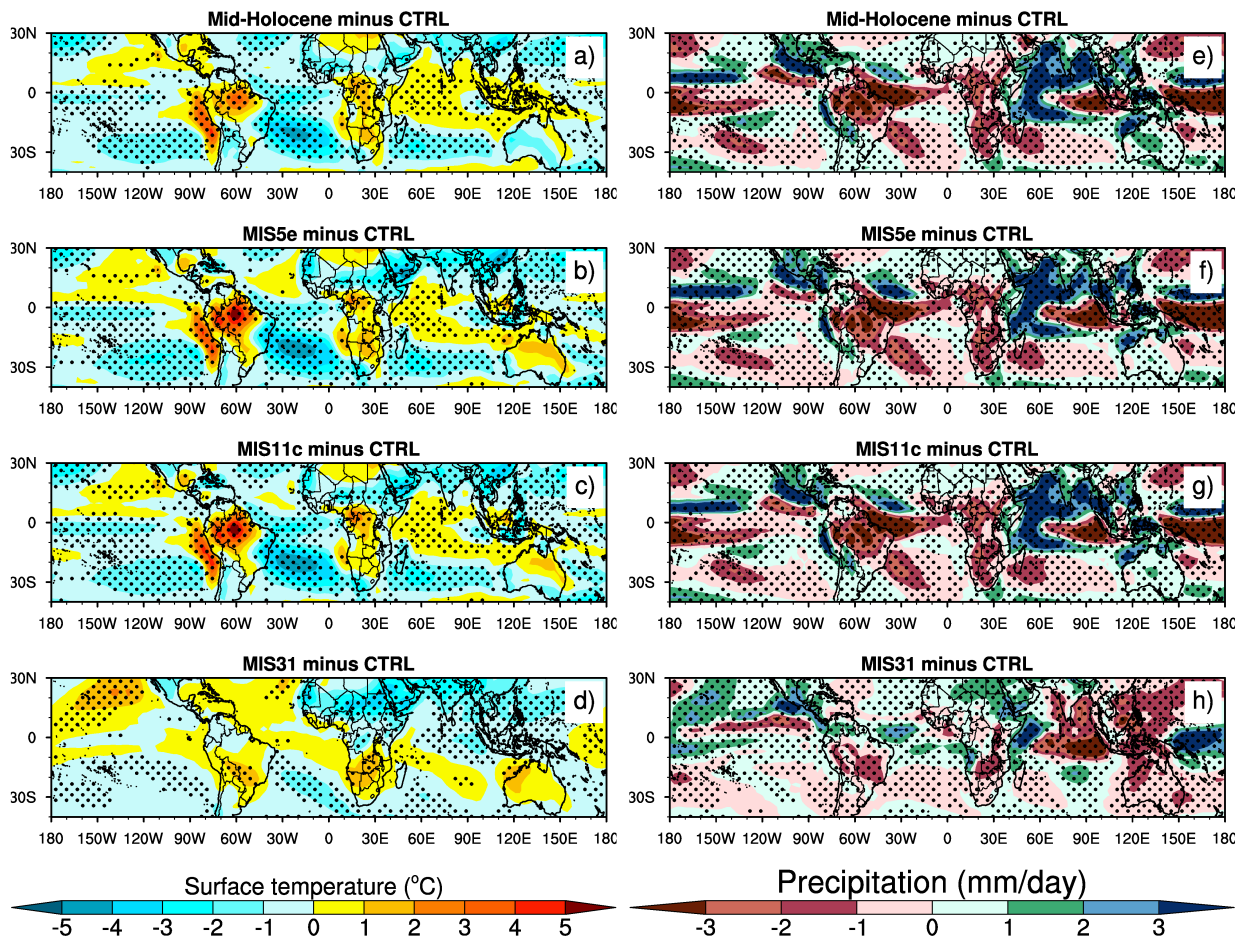


Figure 4.2 – Spatial distribution of annual mean surface temperature difference and annual mean precipitation difference between interglacials experiments compared to SST and precipitation from CTRL; MH minus CTRL (a,e), MIS5e minus CTRL (b,f), MIS11c minus CTRL (c,g) and MIS31 minus CTRL (d,h). Black dots represent correspond to statistically significant anomalies at the 95% confidence interval.

4.3.2 Changes in cover land

The climate response to SST changes revealed that precipitation plays a key role in land cover distribution. Figure 4.3 shows the spatial distribution of the aridity index, which may be treated as an indication of dry climates over the Africa, SAM and the Australia sectors. The aridity index (I_{UNEP}) is evaluated in this study to identify regions that are affected by water stress, as provided by the relationship between precipitation

and evapotranspiration. The I_{UNEP} is defined by the equation

$$I_{UNEP} = \frac{P}{ET_0} \quad (4.1)$$

where, P is the average annual rainfall provided by CCM3/IBIS, and ET_0 is the annual potential evapotranspiration, calculated using the Thornthwaite (1948) and adjusted by Penman (Penman, 1948). The spatial distribution adopted in this study is based on the classification of the aridity index similar to the current climate (Feng and Fu, 2013). Dry and wet climates are defined by I_{UNEP} lower than 0.65, hyperarid ($I_{UNEP} < 0.05$), arid ($0.05 \leq I_{UNEP} < 0.2$), semi-arid ($0.2 \leq I_{UNEP} < 0.5$), dry-subhumid ($0.5 \leq I_{UNEP} < 0.65$) and humid ($I_{UNEP} > 0.65$).

In the CTR run, the CCM3/IBIS model is able to reproduce arid and semi-arid areas located in tropical and subtropical regions, such as Southern Africa, Australia and Northeast Brazil (NEB) and Southern SAM (Figure 4.3). The most arid lands are located in the Sahara/Sahel and central Australia, as expected. By analyzing the I_{UNEP} delivered by past climate through differences in the spatial distribution between the experiments and CTRL run, it is noticed that positive temperature anomalies ranging from 1-3°C over the continents (Figure 4.2), favoring the expansion of dryland over the Congo region, Southern Africa, Indonesia and much of the Amazon during MH, MIS5e and MIS11c (Figure 4.3b-c-d). The prolonged drought over the tropical region can cause changes in large-scale meteorological systems such as the ITCZ. From supplementary materials, Nilsson-Kerr et al. (2021) disponibility a summary of the available global compiled hydroclimate proxy data for MIS5e. A comparison between CCM3/IBIS experiments and hydroclimate proxy indicates a dry continent over the Amazon region in North Brazil. Therefore, the main spatial patterns of aridity are well established in response to decreased insolation in southern hemisphere summer during MIS5e.

In addition to the macroscale effect, there is also a decrease in local convection due to subsidence, which may cause suppression of precipitation not only locally, but also in remote regions. Despite these changes being on past distant intervals and representing the natural variability of the climate system, studies aimed at understanding deforestation through coupled modeling Nobre et al. (2009), suggest that the reduction of rainfall over the Amazon depends on a set of complex factors, related to the weakening of the Equatorial

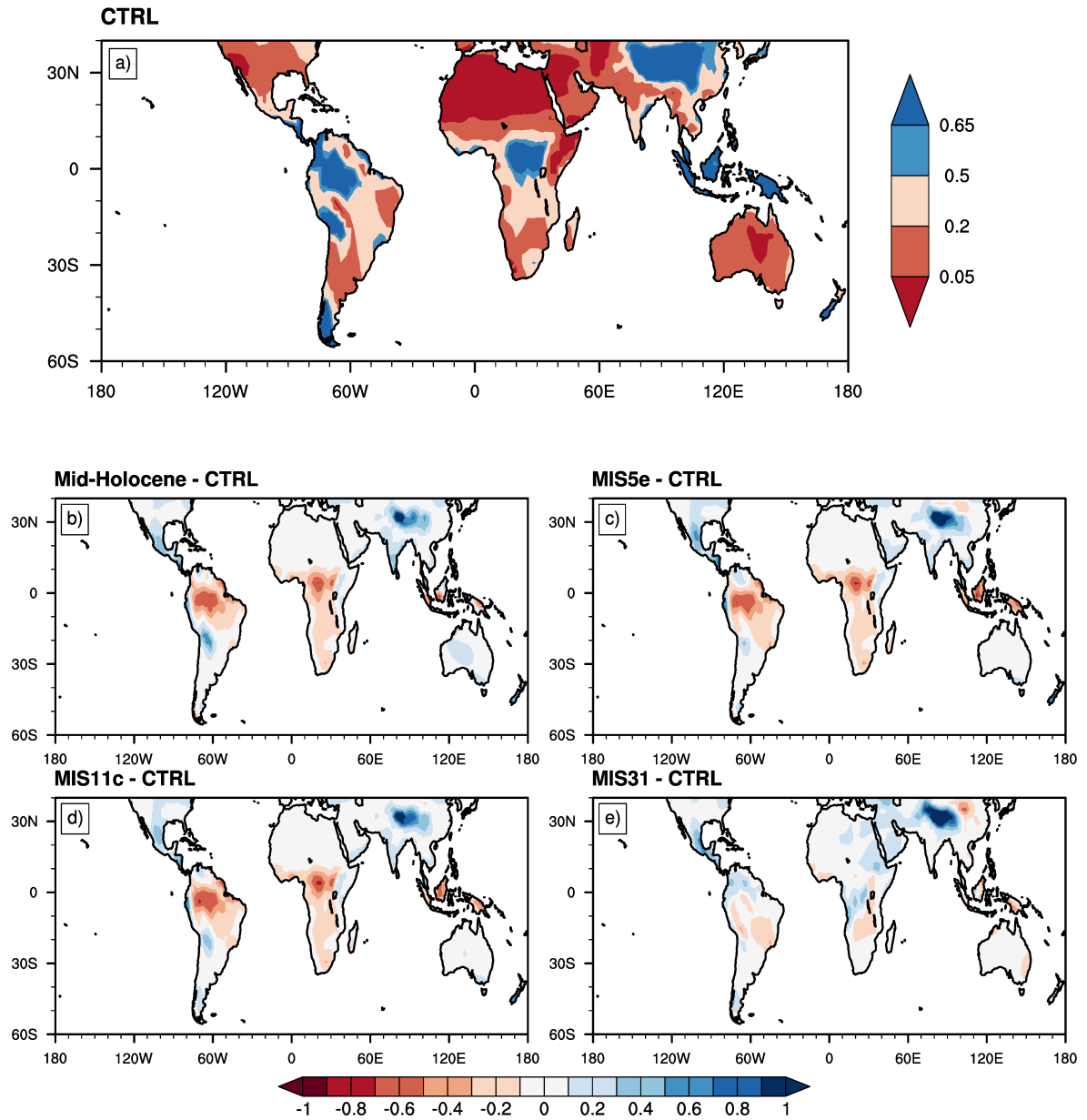


Figure 4.3 – Aridity index to CTRL (a), MH minus CTRL (b), MIS5e minus CTRL (c), MIS11c minus CTRL (d) and MIS31 minus CTRL (e).

Pacific temperature gradient, reduction of deep convection in the Amazon and changes in land use, which favors the weakening of the ascending branch of the Hadley Cell in the latitudes between 10°S-0 (Supplementary Fig. A2). MIS31 does not show substantial changes for the aridity index (Fig. 4.3e). However, dryer condition may be found over South America and subtropical Africa.

It might be expected that these changes in the hydrological cycle as reproduced by the I_{UNEP} , lead to changes in vegetation patterns during the past interglacial eras. Higher

continental temperatures increases the potential evapotranspiration that can reduce soil moisture, if a lack of precipitation takes place. Theoretically, dry soil conditions will result in the presence of much sparse vegetation affected by enhanced seasonal changes.

4.3.3 Global patterns of vegetation

Figure 4.4 shows the distribution of observed vegetation from the MODIS data and the dynamic vegetation simulated by the CCM3-IBIS model based on the CTRL simulation. It has to be mentioned that MODIS vegetation applies a mathematical algorithm to convert radiances in physical vegetation properties, and delivers mean conditions to the year 2010. The CCM3-IBIS model, on the other hand, simulates the potential vegetation related to mean climate conditions, and does not include the anthropogenic influence associated to deforestation. Thus, regions with evergreen forests in South America and Africa according to MODIS cover small areas with respect to CCM3-IBIS.

Overall, CCM3-IBIS is able to reproduce the main features of the African, Australian and South American (SAM) biomes when compared to the MODIS dataset. At the regional scale differences are noted. The simulated vegetation over Central Africa shows grassland, evergreen tropical and deciduous tropical forests, while the observed vegetation map shows savanna and mixed forest in parts of the Congo region. The model simulated tropical evergreen forest across the Indonesia region, corroborating the satellite-based vegetation. The land cover over Australia slightly differs from observed vegetation; MODIS shows most mixed and deciduous forests and savannas, CCM3-IBIS simulates boreal evergreen forest, mixed forest and savannas; over the northern portion of Australia the simulated landcover is dense shrubland, but the observed vegetation map shows savanna and grassland. In South America, the dominant region of the tropical evergreen forest of the Amazon is simulated, in some areas over Venezuela and Colombia, the MODIS dataset shows savanna, but the CCM-IBIS reproduced tropical deciduous forest.

Turning to interglacials experiments, the South America vegetation experiences significant changes during MH, MIS5e and MIS11c (Fig.4.4 c-d-e), as a consequence of the drop in temperatures and air humidity over the southern tropical region, with exception of MIS31 (Fig.4.4 f). During the MH, MIS's 5e and 11c dominant biomes such as evergreen

and deciduous forest over the Amazon have been replaced by savannas and grasslands, mainly in the west/south of Amazon, which are sectors affected by the South American monsoon, as well as reduced atmospheric CO₂ levels during those past intervals weakens the physiological effect related to fertilization. Based on simulations to reconstruct the global carbon cycle, Schurgers et al. (2006) demonstrated that large dominance of areas covered by savanna over Amazon in response to a dry climate during the Holocene and Eemian.

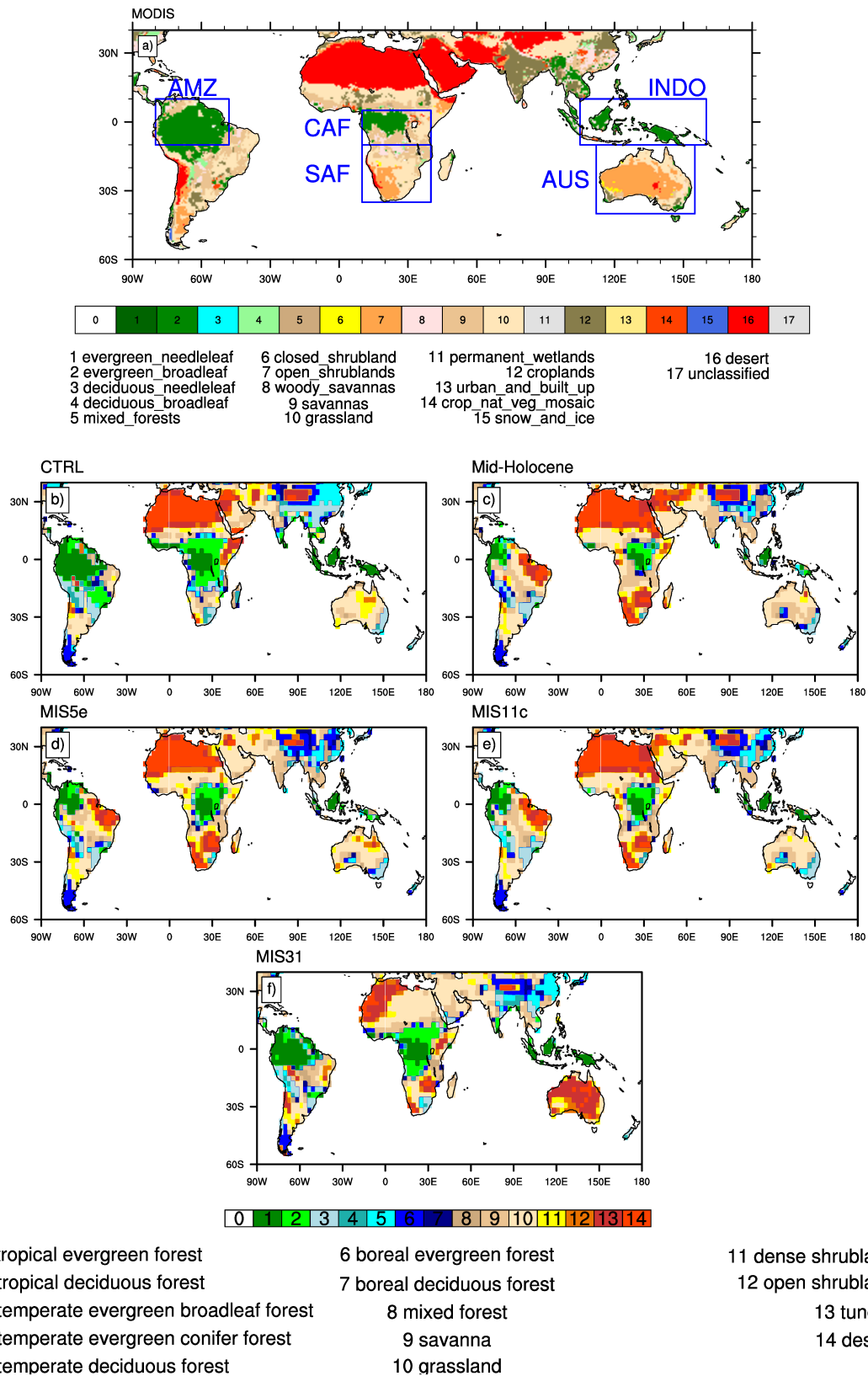


Figure 4.4 – Land cover map derived from MODIS data (MOD12C1) (a), and defined in this study at T42 spatial resolution from CTRL by CCM3/IBIS (b), MH (c), MIS5e (d), MIS11c (e) and MIS31 (f), experiments. Regions delimited in blue boxes refer to domains in which percentage classification vegetations, precipitation e temperature are averaged, as analyzed in this paper are shown; AMZ: Amazonia (10°N-10°S and 50°W-80°W); CAF: Central Africa (5°N-10°S and 10°E-40°E); SAF: South Africa (10°S-35°S and 10°E-40°E); INDO: Indonesia (10°N-10°S and 105°E-160°E); AUS: Australia (10°S-40°S and 112°E-155°E).

Ledru et al. (2001); Ledru (2002); Prado et al. (2013) based on pollen inferences found an expansion of savanna and pasture over eastern Amazon and central Brazil between the Holocene and Middle Holocene periods.

Other extreme scenario is also shown in the NEB, where the caatinga biome is replaced by deserts in the MIS5e, MH and MIS11c interglacials (Fig.4.4 c-d-e). This suggests that the shift from the caatinga biome to the desert may be associated with a dry climate and reduced atmospheric CO₂ concentration. These changes can be attributed to the intensification of feedback mechanisms through the anomalies that generate diabatic cooling, as discussed in past studies (Zebiak, 1986). For southern Brazil, the model was able to simulate grass vegetation that may be related to the pampa, as well as a much smaller arboreal fraction over the Southeast when compared to CTRL. Overall, our results corroborate the results of previous studies such as Behling et al. (2005), where it was found that pollens from São Francisco de Assis, Rio Grande do Sul, indicate grasses with forest refugees, associated with hot and dry climates during the Holocene.

Changes in vegetation dynamics during MIS31 are also remarkable. For example, Australia was totally desertified, part of the Sahara changed from desert to savannah, and tropical forest increase in Tropical Africa. The Amazon perennial forest remains dominant as simulated by the CTR run, however, savanna begins to expand between southern Amazonia and northern central South America (Fig.4.4f).

Figure 4.5 shows vegetation reconstruction from the BIOME6000 project for MH. As verified and discussed earlier in Figure 4.4c, our simulation matches well to the grassland domain and tropical deciduous forest between Central and Southern Africa. In addition, good correspondence is also found with savanna expansion across the Congo region and Australia. The retreat of the tropical evergreen forest is not shown by the inferred pollen data, but the BIOME6000 dataset covers a small region and therefore it is not feasible to draw considerations for the entire Amazon domain.

4.3.4 Change in vegetation areal coverage

Analyses for vegetation types changes during the interglacials demonstrated that parts of the CAF region under CTRL conditions is covered by about 35% of evergreen tropical

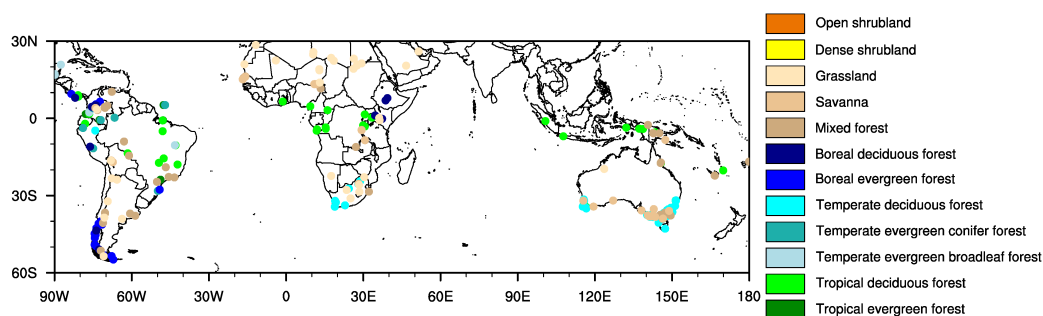


Figure 4.5 – Biome mid-holocene palaeovegetation data from BIOME 6000 (Prentice et al., 2000). Coloured circles represent vegetation classification each, respectively. See Table An1 in Supplement.

forest, 34% of deciduous forest, mixed forest 3% and 9% of grassland (Table 4.3, Fig. 4.6 a). During MH, MIS5e and MIS11c, there is a systematical reduction of forests by up to 60%, 50%, whereas there is an increase of grassland in particular during the MH. Savannas that is almost absent in the CTRL climate over the CAF region, experiences increases for all interglacial stages, due to enhanced climate seasonality with well defined months governed by low precipitation. Paleohydrological analyzes indicate that the Sahara/Sahel regions during the MH large extent of grass and savannah vegetation (Lézine et al., 2011) were present in agreement with the simulated results shown in Table 4.3.

Over the SAF domain, the CTRL delivers 11% of perennial forest and 14% of the Atlantic forest (Fig. 4.6 b). There were no discrepant changes among the interglacials, where the dominant savanna and grassland vegetation remained conserved with small variations. It should be stressed that climate conditions during the MH, MIS5e, MIS11c does not support evergreen forests (Fig. 4.2 and Table 4.3). The MIS31 differs from other interglacial and shows very similar features as delivered by the CTRL. It is noticed that precipitation changes are small which are accompanied by lower temperatures.

In our CTRL simulation Indonesia (INDO) was dominated by tropical evergreen forest (Fig. 4.6 c), thus it reproduced the observed vegetation very well, although there was a reduction in the MH interglacials (53%), MIS5e (50%), MIS11c (34%) and MIS31 (10%), being replaced by Mixed forest during MH (21%), MIS5e (14%) and MIS11c (21%), except for MIS31.

The reproduction of Savanna and grassland dominance over Australia (Fig.4.6 d), in

the CTRL experiment, was similar to observed vegetation. Despite having successfully predicted the dominant biomes, CCM3/IBIS was not able to reproduce the tropical evergreen forest cover over southeastern Australia, but was replaced by 9% Temperate evergreen broadleaf forest. Even so, the differences are few between the biomes of the paleoclimatic periods and the CTRL simulation, conserving the Temperate evergreen broadleaf forest over the same region of Australia in the MH, MIS5e and MIS11c, except in the MIS31, where there was a reduction of 82% of the savanna and 78% of pasture, while it showed coverage of 16% of tundra and 48% of desert.

CCM3/IBIS overestimated the Tropical evergreen forest (Amazon rain forest) biome by 12% over northern SAM when compared to observed vegetation, but warm paleoclimatic periods showed a reduction of 76% (MH), 58% (MIS5e), 64% (MIS11c) and 9% (MIS31), when compared to the CTRL experiment. However, there was an increase in savanna and pasture, mainly in the periods MH (81% and 95%), MIS5e (80% and 93%) and MIS11c (75% and 94%).

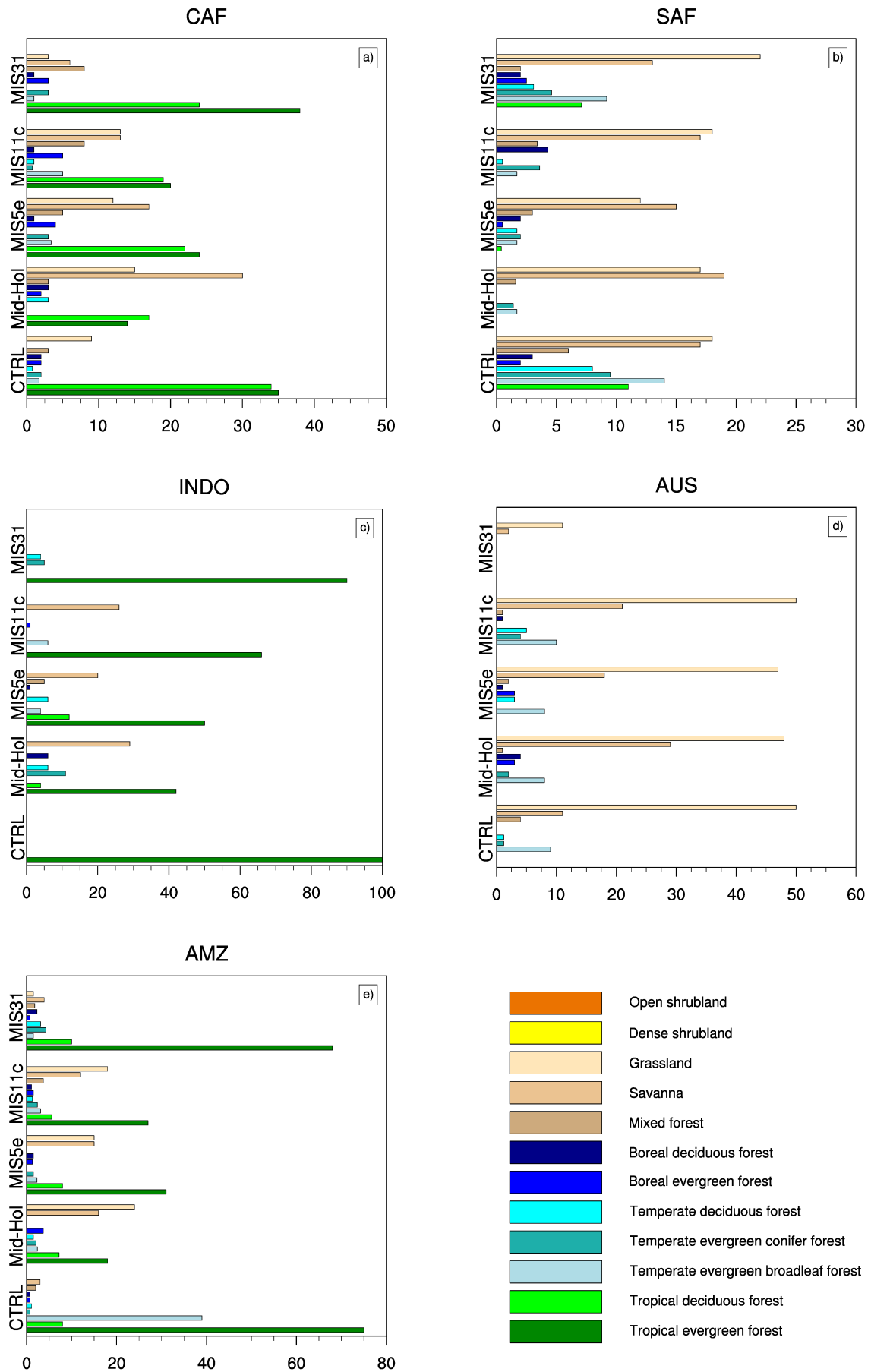


Figure 4.6 – Bar chart of percent (%) change from biome type of the dominant classifications over each study region. a) CAF, b) SAF, c) INDO, d) AUS and e) AMZ.

Table 4.3 – Percent change in the number of pixels from biome type

Biome type	CTRL	MH	MIS5e	MIS11c	MIS31
CAF					
Tropical evergreen forest	35%	14%	24%	20%	38%
Tropical deciduous forest	34%	17%	22%	19%	24%
Mixed Forest	3%	3%	5%	8%	8%
Savanna	0.1%	30%	17%	13%	6%
Grassland	9%	15%	12%	13%	3%
SAF					
Tropical deciduous forest	11%	0.1%	0.4%	0.1%	7%
Temperate evergreen broadleaf forest	14%	1%	1%	1%	9%
Temperate evergreen conifer forest	9%	1%	2%	3%	4%
Savanna	17%	19%	15%	16%	13%
Grassland	18%	17%	12%	18%	22%
AUS					
Tropical evergreen forest	0	0	0	0	0
Savanna	11%	29%	18%	21%	2%
Grassland	50%	48%	47%	50%	11%
Tundra	3%	0%	7%	1%	16%
Desert	0%	1%	1%	0%	48%
IDA					
Tropical evergreen forest	100%	47%	50%	66%	90%
Tropical deciduous forest	0	4%	12%	0	0
Temperate evergreen conifer forest	0	12%	0%	0	6%
Mixed forest	0	21%	14%	21%	0
AMZ					
Tropical evergreen forest	75%	18%	31%	27%	68%
Tropical deciduous forest	8%	7%	8%	5%	10%
Mixed forest	2%	0.1%	0.1%	3%	2%
Savanna	3%	16%	15%	12%	4%
Grassland	0.1%	24%	15%	18%	1%

4.3.5 Effect of moisture transport on vegetation

To better describe the interaction between vegetation dynamics and terrestrial precipitation, in Figure 4.7 the spatial fields of horizontal moisture advection at the 850 hPa level for the summer are shown (first column, Fig.4.7 a-c-e-g-i) and winter (second column, Fig.4.7 b-d-f-h-j) austral. The CTRL simulation clearly shows the circulation pattern associated with the trade winds entering the tropical continents of Africa, Australia and South America (Figure4.7a-b). The maximum moisture advection during summer and winter is concentrated over the Amazon and southern South America, southern Africa and northern Australia, with the ocean as a source of moisture, except in the Congo region, where it

has a moisture transport regulated by the tropical Atlantic during winter.

The humid air masses that are over the Amazon during the austral summer and winter seasons can be explained by the combination of air masses from the Atlantic Ocean and the evapotranspiration of the Amazon Forest itself. This justifies the pattern of vegetation associated with tropical forest due to the high rate of annual precipitation.

The strong changes in moisture transport are shown with the differences between interglacials and CTRL (Figure 4.7, MH c-d, MIS5e e-f, MIS11c g-h and MIS31 i-j). The similarities between the MH, MIS5e and MIS11c periods are highlighted by the minimum concentrations of moisture advection that can be observed with negative signs in parts of the NEB and central Amazon region during the DJF and JJA stations (Figure 4.7, c-d,e-f,g-h), associated with the weakening of the western zonal component, which is the main parameter for the transport of moisture in the tropical region, and which is part of the SAM monsoon circulation, mainly during the summer. Although it is not entirely clear whether this drives dynamic vegetation modification. The weakening of westerly winds and the reduction of tropical precipitation from South America was attributed to orbital forcing that caused insolation attenuation in the SH during the early Holocene (Cruz et al., 2009). In order to validate the global model simulations through pollen records, Prado et al. (2013) analyzed the South American monsoon precipitation field based on PMIP3/CMIP5 during the MH, where they could see that the response to the reduction of precipitation in some sectors of Brazil is related to a weakening of the South American monsoon due to the influence of the configuration orbital, since the intensity of insolation on the surface in the austral summer was more attenuated for that time when compared to the pre-industrial period.

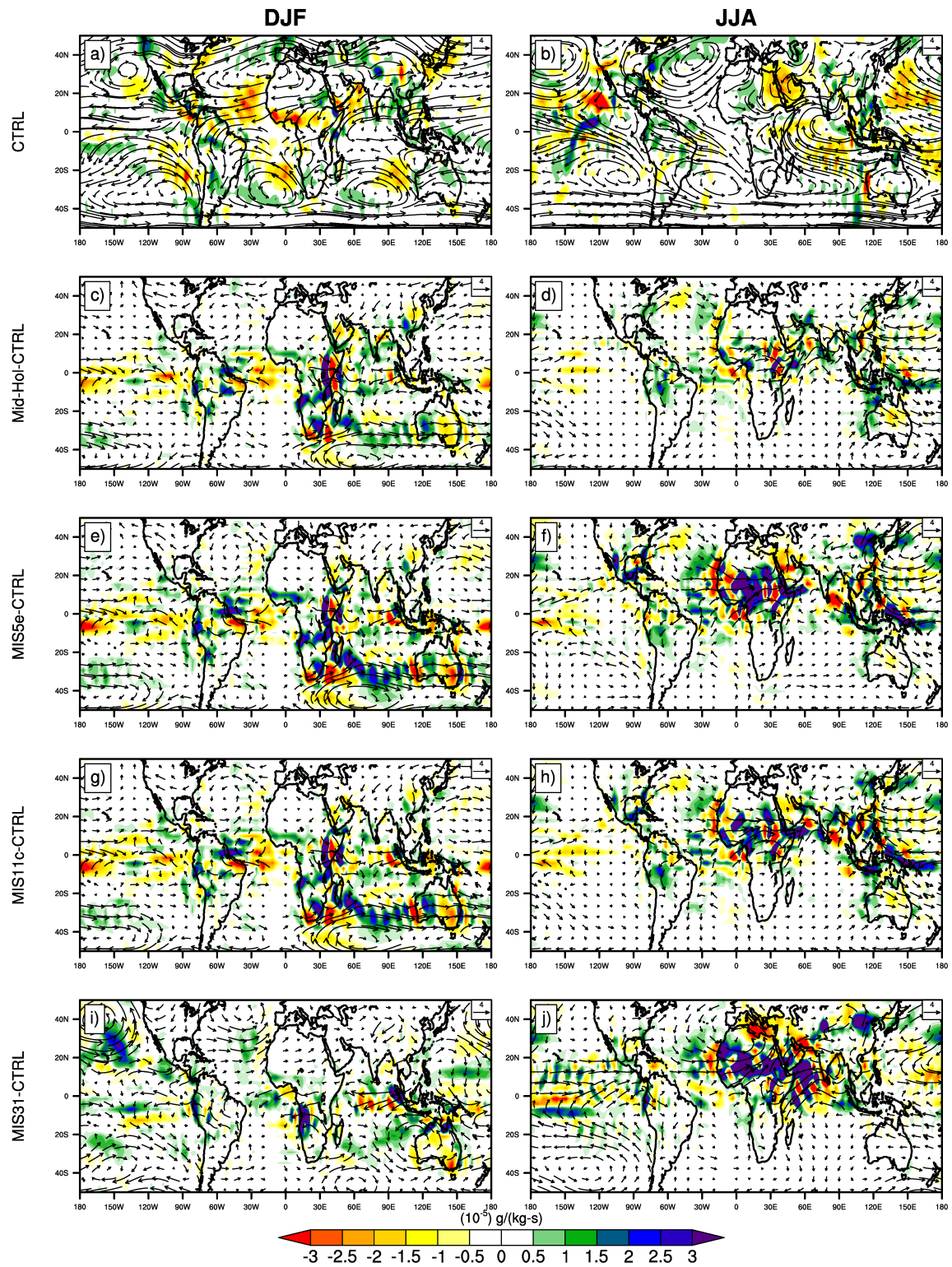


Figure 4.7 – Map of horizontal moisture flux advection of South American, African and Australia region. First column (austral summer) and second column (austral winter); CTRL (a-b), MH minus CTRL (c-d), MIS5e minus CTRL (e-f), MIS11c minus CTRL (g-h) and e MIS31 minus CTRL (i-j).

Similar signs of drying during the two seasons can be seen in the extreme south of

Africa due to the transport of dry air from the west coast. This suggests that the waters off the west coast of Africa, being prone to upwelling, reduce convection due to dry downward movements. However, there was an increase in the flow of moisture between the Congo and Sahara region, coming from the evaporation of the North Atlantic due to the positioning of the ITCZ. Braconnot et al. (1999), for example, showed from a modeling set coupled to a biome model for the MH, oceanic feedbacks were responsible for the increase in moisture advection from the Tropical Atlantic, and concluded that the North African vegetation is responsible for the increase of local evapotranspiration. This explains the expansion of Savanna and grassland over the desert during the interglacials analyzed here. Through paleoclimatic records over North Africa, from marine sediments and lakes, during the MH an increase in pasture cover was observed due to the increase in the annual rate of precipitation (Larrasoana et al., 2003; McGee et al., 2013; DeMenocal et al., 2012).

During the MH, MIS5e and MIS11c interglacials, weakened moisture transport from the eastern Tropical Indian Ocean is present over western Australia in the austral summer (Fig.4.7 c-e-g), having a cohesive relationship with the increase in precipitation over the same sector (Figure 4.2e-f-g). In contrast, a high pressure center that is between the southern Tropical Indian Ocean and western Australia during the summer at MIS31 (Fig.4.7 i), causes subsidence and coincides with negative precipitation anomalies (Fig.4.2 h).

For MIS31, the South American monsoon circulation is weakened, driven by the intensification of the zonal east wind in the DJF period (Figure4.7i). This is also apparent over the monsoon regions of Africa and Australia; however, a maximum advective moisture concentration is observed between the Sahara/Sahel coming from the North Atlantic during the boreal summer, in response to the migration from the ITCZ to the NH (Figure4.7j).

4.3.6 Climate response under individualized ocean basin changes

Precipitation responses over continental areas of Africa, Australia and South America, caused by changes in individual ocean basins such as PAC, ATL and INDI, during interglacials are shown in Figure 4.8. For the MH, the results indicate that the PAC has no association with the precipitation of the AFR, AUS and SAM sectors (Fig. 4.8a). On

the other hand, the ATL was the basin that most influenced the significant reduction in SAM and AFR precipitation (Fig. 4.8b), although the INDI caused a reduction over Southern Africa (Fig. 4.8c). As shown in Fig. 4.8d-f, g-i e j-l, the PAC induces a reduction in precipitation only in South Africa, except for MIS31, where southeastern SAM and Australia also had similar responses. For the three MIS interglacials 5e, 11c and 31, the SAM and AFR are totally dependent on the ATL. This suggests that much of the cooling of the equatorial Atlantic causes a reduction in precipitation, as the ITCZ migrates to the Boreal Hemisphere in response to the variation of this ocean basin and acting on the photosynthetic processes of the Tropical Forest. Under climatological conditions, for all interglacials, INDI has simulated significant rainfall reduction in parts of southern Africa and Australia.

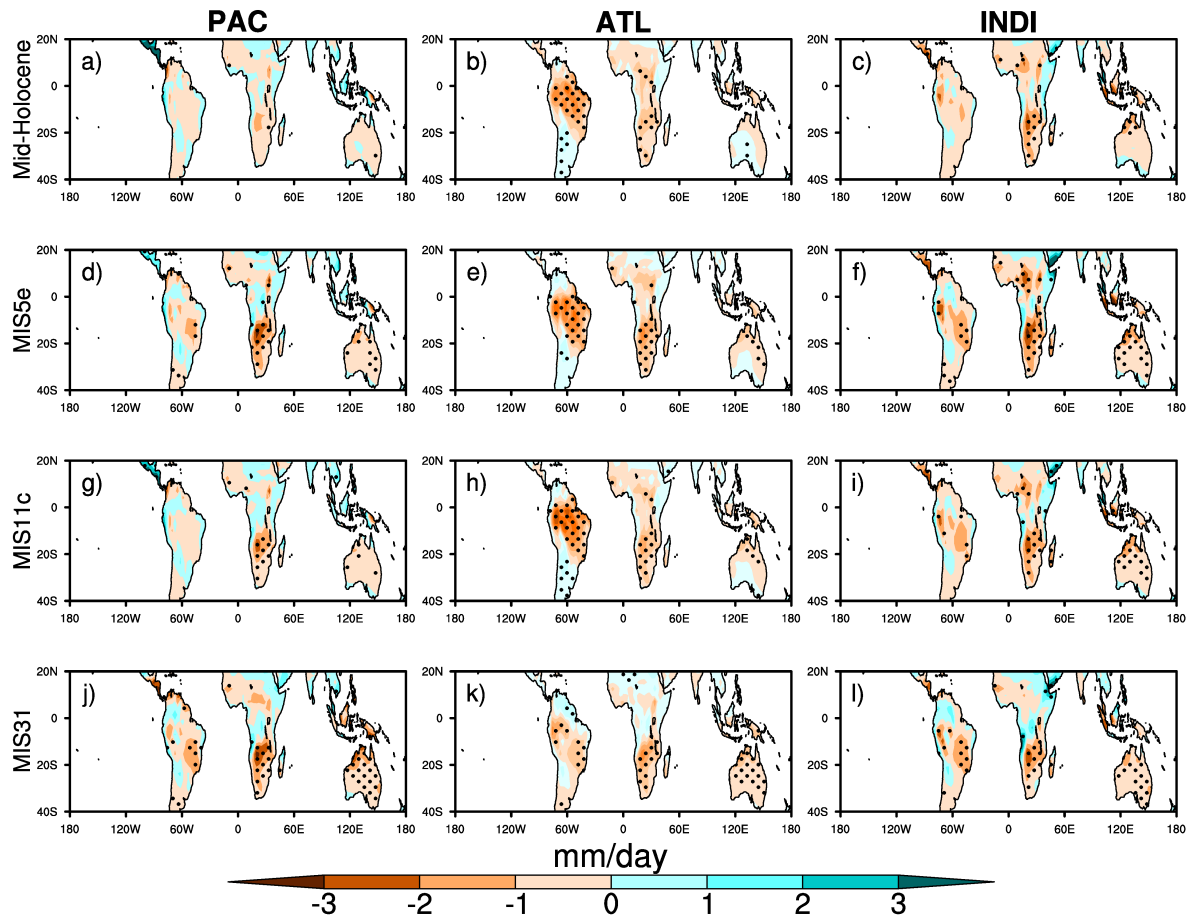


Figure 4.8 – Continental distribution of precipitation (mm/day) anomalies for respective rows and columns representing the oceanic basins regions forced in MH (a-b-c), MIS5e (d-e-f), MIS11c (g-h-i) and MIS31 (j-k-l) experiments. Black dots represent areas where the results are statistically significant (at a 95% confidence level).

When calculating annual averages of precipitation over the areas of CAF, SAF, AUS, IND and AMZ, strong connections were found with variations in SST (see precipitation values for climatological conditions on Figure A3 in Supplement). For the CAF region, during the MH, MIS5e and MIS11c, there was a reduction in the average annual precipitation around 18%, 15% and 15,5% (-0.71, -0.57 e -0.59 mm/day) (Fig. 4.9a-f-k). Although the Congo region is strongly influenced by the ITCZ during the austral summer and autumn, this negative variation in precipitation from these three interglacials is a response to the anomalous cooling of the ATL SST, as the cold water reduces convection and increases the formation of stratus clouds, causing the ITCZ to migrate to the Boreal Atlantic Ocean. Despite changes in precipitation during the closest interglacials, precipitation in MIS31 turns out to be very close to climatology (CTRL) when the model is forced by the ATL and PAC experiments (Fig.4.9j). The reduction of annual precipitation was also shown on the SAF with 30%, 29%, 26% and 11% (-0.65, -0.63, -0.56 and -0.25 mm/day) in MH, MIS5e, MIS- 11c and MIS31 (Fig. 4.9a-d-g-j). This water deficit is associated with changes in IND SST patterns, since the differences between IND_{MH} minus CTRL simulations show reductions of 27% as well as 24% for IND_{31} minus CTRL; MIS's 5e and 11c, showed suave rainfall when the model was forced with the CAP's SST anomalies.

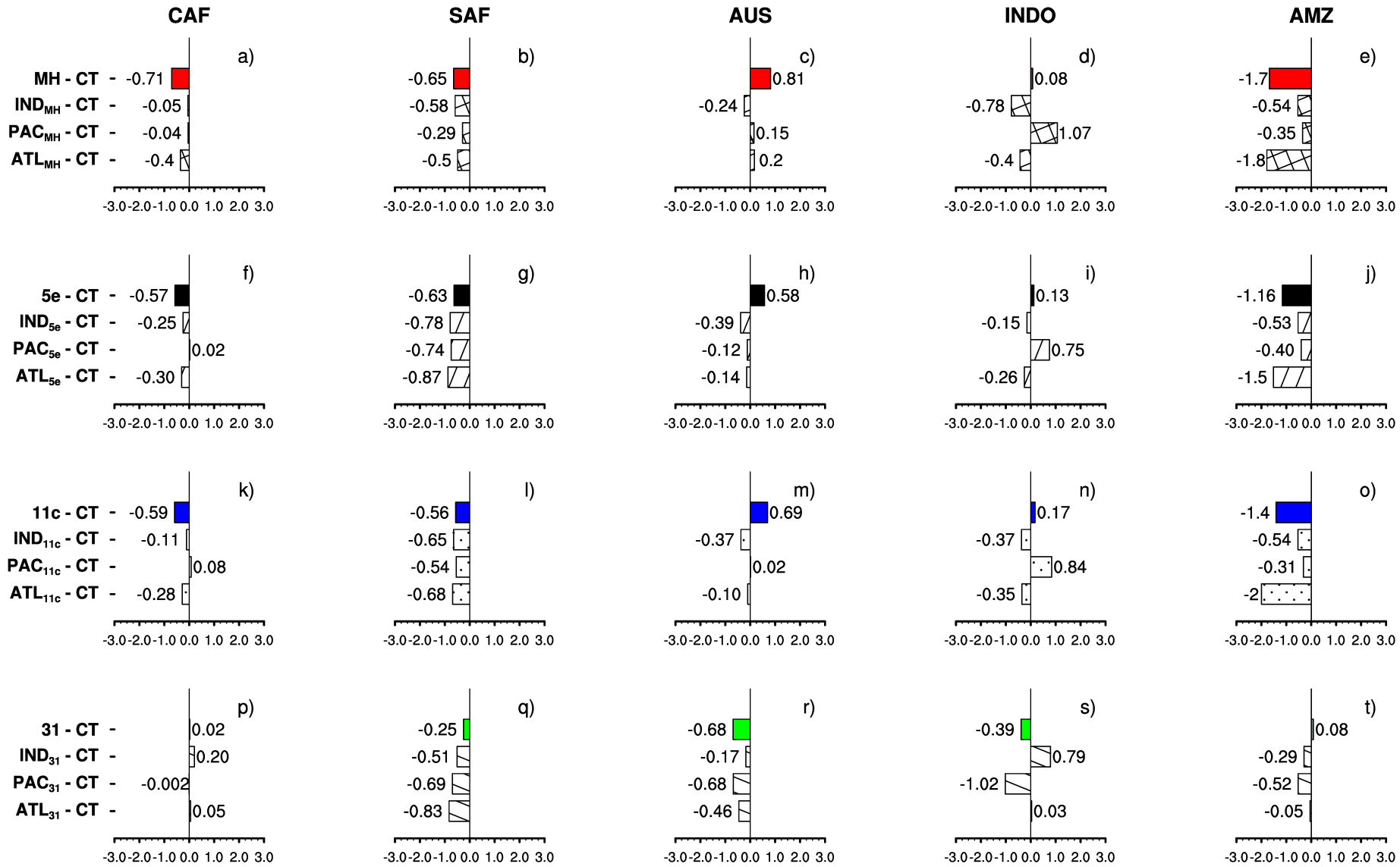


Figure 4.9 – Areal averaged precipitation (mm/day) CAF, SAF, AUS, IND and AMZ, region for the SST anomaly sensitivity experiments.

The increase of 38%, 27% and 32% (+0.81, +0.58 and +0.69 mm/day) is shown over AUS, in MH, MIS5e and MIS11c, in relation to climatology (Fig. 4.9c-h-m), although MIS31 has a reduced average annual rainfall of 32% (-0.68 mm/day) (Fig. 4.9r), these changes occurred in all interglacials are associated with the SST anomalies that were forced by the PAC basin. The Indonesia region (INDO) showed a weak increase when compared to the climatology of annual precipitation, around 1%, 2% and 2.5% (+0.08, 0.13 and 0.17 mm/day) in MH, MIS5e and MIS11c (Fig. 4.9d-i-n), except for MIS31, where there was a 5% reduction (-0.39 mm/day) (Fig. 4.9s).

The average annual precipitation is lower over the AMZ region when we observe the periods MH, MIS's 5e e 11c (Fig.4.9e-j-o), where there was a reduction of 30%, 20% and 25% (-1.7, -1.16 and -1.4 mm/day). However, a weak increase of almost 2% (+0.08 mm/day) is observed in MIS31 (Fig.4.9t). These changes in the hydrological cycle over the AMZ are due to variations in SST in the ATL basin, because when quantifying the differences between the ATL_{MH} , ATL_{5e} and ATL_{11c} experiments with CTRL, there were decreases of 32%, 27% and 35%. In the analysis of the regional mean temperature of the AMZ, the experiments MIS-5e, MH and MIS-11c, exhibit heating of almost 3°C, which is also attributed by the ATL forcing (see Figure A4 in Supplement). This suggests that oceanic energy transport leads to an inter-hemispheric ITCZ shift (Marshall et al., 2014), mainly to the hemisphere that provides a greater exchange of energy between the ocean and atmosphere (Chiang and Friedman, 2012). On the other hand, MIS31 shows similarity with current temperatures (Fig.4.9h), in addition to the contribution of SST forcing through the PAC basin.

4.4 Conclusions

Although the orbital parameter has a strong influence on the planet's natural variability, especially on the precipitation variable, this factor is not the only factor that explains the sensitivity of the CCM3/IBIS model.

In fact, changes in the SST of the ocean basins were affected by radiative forcing due to the precessional factor during interglacials, resulting in the displacement of the ITCZ to the NH, in addition to feedbacks from the climate system (Chiang and Bitz, 2005). During the austral summer, the annual precipitation rate is reduced over South America, southern Africa and northern Australia in response to cold SST anomalies and high precession. Although our results do not provide an in-depth analysis of the dynamic mechanisms associated with ITCZ, our interpretation is that the reduction in the flow of moisture from the Atlantic to the Amazon caused an imbalance in the annual precipitation rate of the AMZ, and consequently, making it a climate dry and prolonged, leading to the reduction of the tropical evergreen forest biome to the expansion of savanna and grasslands.

The high precession and obliquity values during summer and winter were not enough to strengthen the South American monsoon (Fig.A1a-d), although there was an induction of an anomalous low locally over Amazonia during the MH and MIS's 5e and 11c, the moisture transport inferred by the North Atlantic was not sufficient to maintain deep convection. The precipitation decrease coincides with the anomalous high pressure center between the South Atlantic basin and the South American continent induced intense cooling between the Equatorial and Southern Atlantic. Under all these conditions mentioned, despite the periods studied being of remote time scale, that is, millenary, the natural variability without anthropogenic influence and stabilized CO₂ levels, which leads us to suggest that the model CCM3/IBIS was able to produce a "dieback" of the Amazon forest (Parry et al., 2022; Malhi et al., 2009), that is, the extensive region inhabited by a vast perennial vegetation tends to extinction due to drastic reductions in precipitation and long periods of drought. The opposite effect can be observed between the Sahara/Sahel and Congo regions during boreal summer (Fig.A1), where a wide anomalous low is induced during all interglacials, leading to increased continental humidity due to humid winds from the North Atlantic.

Although CCM3/IBIS performs well in reproducing the current climate, the lack of

pollen paleo-reconstruction data makes it difficult to analyze interglacial periods prior to MIS5e. What we can conclude is that the dynamic vegetation distributions that we analyzed are well related to SST changes. However, we cannot explain that the continental tropical rains that occur over the SH monsoon sectors are totally associated with SST changes. Roberts and Hopcroft (2020) suggest that changes in heat transport that occur in the ocean are more associated with climate balance, although there is a link between the location of the ITCZ and atmospheric feedbacks. On the other hand, in a recent study, Venancio et al. (2022) through paleoreconstructions, suggest that the main causes for ITCZ migration during abrupt MIS5 events were modulated by oceanic heat transport and variations in NH ice cover.

The impact of extreme aridity was sufficient to cause extreme drought over the tropical regions of Congo and Amazonia during the MH, MIS5e and MIS11c interglacials. The analysis presented here suggests that the equatorial convection of these sectors depends on sources of moisture from the processes of evapotranspiration and advection of humid air from warm SSTs. However, the effects of the precipitation deficit on the continent causes a long-term retraction of the perennial vegetation, since the vertical water vapor transport became attenuated due to the decrease in local evapotranspiration, and therefore, these changes made the driest and most extreme conditions. Drier conditions for a tropical forest can result in the adverse effects of evapotranspiration that are critical to the distribution of convective activity. Although our simulations show an increase in temperature in regions over the Amazon rainforest, it was not possible to intensify the convective cell of precipitation.

The relationship between $\text{Ni}\tilde{\text{N}}\text{O}_{34}$ and precipitation over the study regions was relatively weak and not significant across all interglacials (see Table A1 in the Supplement). Although there is no influence on the interannual variations of the precipitation pattern, it is worth mentioning that the CCM3/IBIS model was forced from the climatological boundary conditions, and therefore, this may not accurately exhibit the significant relationship, which suggests that oceanic forcing tropical Pacific in ENSO years is not entirely clear when it is correlated with SH continental precipitation. On the other hand, as shown in this study, the SST of oceanic basins play an important role in determining the continental climatological pattern, as well as in the internal mechanisms of precipitation

and in the pattern of vegetation. Although tropical regions are influenced both by climate and vegetation in response to Pacific and Atlantic SSTs, Nobre et al. (2009) combined simulations for scenarios with changing vegetation cover from a coupled model to investigate whether vegetation can play a role in modulating the ocean. The collaborators verified that in the scenario of forest removal there was a weakening of the deep convection of the Amazon, following a series of events, such as a reduction in the trade winds and meridional circulation. In a recent study focused on the current period, Towner et al. (2021) showed that the response of the hydrological cycle over Amazonia depends on the ENSO phase, but overall it can play an important role in vegetation, i.e. the probability of savanna occurrence and tropical deciduous rainforest are less related to Equatorial Pacific anomalies.

The results of our study provide important insights into how SST changes of individualized ocean basins alter southern tropical precipitation as well as dynamic vegetation during remote warm periods. What we want to express is that without forcing these ocean basins it would not be possible to highlight the sensitivity of the CCM3/IBIS model to the response of the vegetation pattern. Despite the absence of these records, the classification of the dynamic vegetation of MH, MIS5e and MIS11c were very similar.

4.5 Acknowledgements

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CONCLUSÃO

Este estudo demonstra que as monções de verão do Austral sofreram fortes mudanças durante os intervalos MIS5e, MIS11c e MIS31. Isso ocorre devido a mudanças na insolação orbital, que modificam ainda mais os padrões globais de TSM e teleconexão. De fato, mudanças no ciclo precessional desencadearam um resfriamento significativo nas bacias oceânicas do Hemisfério Sul (HS). Na região do Sistema de Monção da América do Sul (SMAS), por exemplo, o MIS5e e o MIS11c mostraram uma redução do transporte de umidade da Amazônia para o sudeste da América do Sul (AS). A contribuição da massa de ar seco continental favoreceu quantitativamente a diminuição das chuvas de monção de verão, e a redução intensificada da TSM durante esses interglaciais, proporcionou um deslocamento para norte do SMAS, bem como da Zona de Convergência do Atlântico Sul. Apesar da precipitação mais fraca no verão, os estágios MIS5e e MIS11c são dominados pelo aumento da precipitação no inverno e início da primavera do Hemisfério Sul, para o Sistema de Monção da África do Sul (SMAF), e inverno/outono para o Monção da Austrália (MAUS). Assim, a redução da precipitação está relacionada com as condições mais frias durante o verão HS, em particular nos continentes. As análises de correlação indicam que os processos de grande escala, como os relacionados ao El Niño Oscilação Sul (ENOS) durante os interglaciais, são mais fracos em relação ao CTRL, com exceção do aumento dos valores de correlação significativos sobre o SMAS.

A comparação entre paleoreconstruções e simulações de modelos revelou uma boa concordância. Vale ressaltar que as paleoreconstruções dão a resposta principal a um forçamento particular, mas não indicam a estação preferencial de ocorrência. As simulações climáticas com o objetivo de reproduzir o sistema de monções servem para apontar que a intensidade do aquecimento do Hemisfério Boreal pode modificar o padrão temporal e espacial da precipitação do HS. Em uma perspectiva paleoclimática, isso pode ser entendido como análogos das condições durante os eventos de Dansgaard–Oeschger. Até certo ponto, o aquecimento global que também mostra um clima boreal muito mais quente também pode fornecer precipitação de monção HS modificada de acordo com as condições interglaciais, conforme foi simulado neste estudo. Assim, o clima futuro da Terra pode ser explorado em

detalhes com base em experimentos climáticos para estágios de intervalos passados.

Embora o parâmetro orbital tenha forte influência na variabilidade natural do planeta, principalmente na variável precipitação, esse fator não é o único fator que explica a sensibilidade do modelo CCM3/IBIS. As mudanças das bacias oceânicas foram afetadas pela forçante de insolação em razão do fator precessional durante os períodos quentes paleoclimáticos, levando ao deslocamento da ZCIT para o Hemisfério Norte. A taxa de precipitação anual foi reduzida sobre a América do Sul, sul da África e norte da Austrália em resposta das anomalias frias de TSM e alta precessão. Embora nossos resultados não forneçam uma análise aprofundada dos mecanismos dinâmicos associados à ZCIT, nossa interpretação é que a redução do fluxo de umidade do Atlântico para a Amazônia causou um desequilíbrio na taxa anual de precipitação da Amazônia, proporcionando um clima seco e prolongado, levando à redução do bioma floresta tropical para a expansão de savanas e pastos.

A redução da taxa de precipitação coincide com o centro anômalo de alta pressão entre a bacia do Atlântico Sul e o continente Sul-Americano induzido por intenso resfriamento entre o Atlântico Equatorial e o Atlântico Sul. Os altos valores de precessão e obliquidade durante o verão e inverno não foram suficientes para fortalecer a monção Sul-Americana (Fig.A1a-d), embora tenha ocorrido a indução de uma baixa anômala localmente sobre a Amazônia durante o Holoceno-Médio e MIS's 5e e 11c, o transporte de umidade inferido pelo Atlântico Norte não foi suficiente para manter a convecção profunda. Sob todas essas condições mencionadas, apesar dos períodos estudados serem de escala milenar, a variabilidade natural sem influência antrópica e níveis de CO₂ estabilizados, nos leva a sugerir que o modelo CCM3/IBIS foi capaz de produzir um "dieback" da floresta amazônica, ou seja, a extensa região habitada por uma vasta vegetação perene tende à "desaparecer" devido as drásticas reduções de precipitação e longos períodos de seca. O efeito oposto pode ser observado entre as regiões do Norte da África, mais precisamente no Saara/Sahel e Congo durante o verão boreal (Fig.A1), onde uma ampla baixa anômala é induzida durante todos os interglaciais, levando ao aumento da umidade continental devido aos ventos úmidos do Atlântico Norte.

Embora o CCM3/IBIS tenha um bom desempenho na reprodução do clima atual, a falta de dados de paleoreconstrução de pólen dificulta a análise de períodos interglaciais

anteriores ao MIS5e. O que podemos concluir é que as distribuições dinâmicas da vegetação que analisamos estão bem relacionadas às mudanças de TSM. No entanto, não podemos explicar que as chuvas tropicais continentais que ocorrem sobre os setores de monções continentais do HS estejam totalmente associadas às mudanças de TSM, embora haja uma ligação entre a localização da ZCIT e os feedbacks atmosféricos e oceânicos.

O impacto da extrema aridez foi suficiente para causar seca extrema nas regiões tropicais do Congo e da Amazônia durante os interglaciais MH, MIS5e e MIS11c. A análise aqui apresentada sugere que a convecção equatorial desses setores depende de fontes de umidade provenientes dos processos de evapotranspiração e advecção de ar úmido das bacias oceânicas. No entanto, os efeitos do déficit de precipitação no continente provoca uma retração a longo prazo da vegetação perene, uma vez que o transporte vertical de vapor d'água se tornou atenuado devido à diminuição da evapotranspiração local e, portanto, essas mudanças tornaram as condições mais secas e extremas. Condições mais secas para uma floresta tropical podem resultar em efeitos adversos da evapotranspiração que são críticos para o "gatilho" da atividade convectiva.

Pelas análises com o CCM3/IBIS, a relação entre Niño₃₄ e precipitação nas regiões de estudo foi relativamente fraca e não significativa em todos os interglaciais (ver Tabela A1 no Apêndice). Embora não haja influência nas variações interanuais do padrão de precipitação, vale ressaltar que o modelo CCM3/IBIS foi forçado a partir das condições de contorno climatológicas de TSM, e portanto, isso pode não apresentar com precisão a relação significativa, o que sugere que o forçamento oceânico tropical do Pacífico sem a escolhas de anos de El Niño/La Niña não esteja totalmente claro quando correlacionado com a precipitação continental HS. Por outro lado, como mostrado nesta tese, a TSM das bacias oceânicas desempenha um papel importante na determinação do padrão climatológico continental, bem como nos mecanismos internos de precipitação e no padrão da vegetação.

Os resultados de nosso estudo fornecem informações importantes sobre como as mudanças de TSM de bacias oceânicas individualizadas alteram a precipitação tropical do sul, bem como a vegetação dinâmica durante períodos quentes e remotos. O que queremos expressar é que sem as condições de contorno, ou seja, as bacias oceânicas, não seria possível evidenciar a sensibilidade do modelo CCM3/IBIS à resposta do padrão de vegetação. Apesar da ausência desses registros, a classificação da vegetação dinâmica do Holoceno-Médio, MIS5e

e MIS11c foram muito semelhantes.

ANEXO A

Tabela An1 – Biome reconstructed for South America. Code biome: TRFO (Tropical rainforest); TSFO (Tropical seasonal forest); TDFO (Tropical dry forest); WTRF (Warm temperate rainforest); WEFO (Warm temperate evergreen broadleaf forest); CTRF (Cool temperate rainforest); WAMF (Warm temperate mixed forest); COMI (Cool mixed forest); CGSS (Cool grasslands); STEP (Steppe); DESE (Desert); CGSH (Cool grass shrublands).

Site	Lat	Lon	Biome 6k
Lake Åsa3	-62.62	-61.13	STEP
Harberton	-54.88	-67.16	CTRF
Puerto del Hambre	-53.59	-70.92	STEP
La Misión	-53.5	-67.83	CTRF
Punta Arenas	-53.15	-70.97	CTRF
Meseta Latorre 1	-51.52	-72.05	CGSH
Meseta Latorre 2	-51.44	-72.03	TDFO
Torres del Paine	-50.98	-72.66	COMI
Dichan	-49.66	-73.88	CTRF
Puerto Eden	-49.13	-74.41	CTRF
Laguna Six Minutes	-46.43	-74.33	CTRF
Laguna Stibnite	-46.43	-74.43	CGSH
Laguna Stibnite	-46.43	-74.38	CTRF
Laguna Lincoln	-45.34	-74.07	CTRF
Laguna Lofel	-44.85	-74.43	CTRF
Estero Huitanque	-43.61	-73.82	COMI
Mayol	-42.64	-73.75	CTRF
Puchilco	-42.63	-73.62	CTRF
Mallin Book	-41.33	-71.58	CTRF
Puerto Octay PM13	-40.93	-72.90	CTRF
Primavera	-40.66	-71.18	TDFO
Caunahue	-40.00	-72.00	CTRF

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Tabela An1 – continued from previous page

Site	Lat	Lon	Biome 6k
Salina Anzotegui	-39.06	-63.77	STEP
Cueva Haichol	-38.58	-70.66	STEP
Cerro La China	-37.84	-58.64	TDFO
Empalme Querandies	-37.00	-60.65	TDFO
Veranada Pelan	-36.88	-70.38	TDFO
Vaca Lauquen	-36.83	-71.08	CGSH
Veranada Vulkanpickel	-36.68	-70.41	TDFO
Salina 2	-32.25	-69.33	STEP
Serra do Rio Rastro	-28.55	-49.55	CGSH
Morro da Igreja	-28.18	-49.86	WTRF
Serra da Boa Vista	-27.70	-49.15	CTRF
Rano Raraku Bore 3	-27.16	-109.28	TDFO
Rano Aroui	-27.08	-109.40	CGSH
Serra Campos Gerais	-24.66	-50.21	TDFO
Aguilar	-23.83	-65.75	STEP
Rio da Curuá	-23.83	-48.83	TRFO
Tumbre 2	-23.31	-67.78	STEP
Aguas Calientas	-23.08	-67.42	CGSH
Lagoa da Caço	-22.97	-43.43	TDFO
Morro de Itapeva	-22.78	-45.63	TDFO
Salitre	-19.00	-46.78	TDFO
Lago do Pires	-17.95	-42.21	TSFO
Wasa Mayu	-17.54	-65.91	CGSH
Crominia	-17.28	-49.45	TSFO
Mt. Blanco	-17.02	-67.35	CGSS
Lake Huinãmimarca	-16.50	-69.00	CGSH
Cerro Calvario	-16.50	-68.50	CGSH
Rio Kaluyo	-16.43	-68.13	STEP

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Tabela An1 – continued from previous page

Site	Lat	Lon	Biome 6k
Chacaltaya 1	-16.36	-68.13	CGSH
Cumre Unduavi	-16.33	-68.03	CGSS
Aguads Emendadas	-15.56	-47.58	TSFO
Amarete	-15.23	-68.98	CGSH
Cotapampa	-15.21	-69.11	CGSH
Katantica	-14.8	-69.18	CGSH
Laguna Chaplin	-14.50	-61.05	TDFO
Laguna Bella Vista	-13.58	-61.56	TSFO
Laguna Jeronimo	-11.78	-75.21	CGSH
Laguna Pomacocha	-11.75	-75.50	CGSH
Laguna Tuctua	-11.66	-75.00	WTRF
Laguna Milloc	-11.56	-76.35	CGSH
Laguna Junin2	-11.00	-76.18	COMI
Laguna Huatacocha	-10.76	-76.55	CGSH
Rio São Francisco	-10.46	-43.00	TSFO
Saquinho	-10.44	-43.23	WEFO
Katira	-9.00	-63.00	TDFO
Carajas	-5.00	-48.00	TSFO
Ciudad Universitaria X	-4.75	-74.18	WAMF
Lake Surucucho	-3.75	-78.95	WTRF
Ayauch	-2.09	-78.13	TSFO
Lago Crispim	-0.8	-48.00	TSFO
Lagoa da Curuça2	-0.76	-47.85	TRFO
Mariñame-II	-0.66	-72.03	WTRF
Monica-1	-0.60	-72.50	TRFO
Cayambe	-0.03	-78.03	CGSH
La Pata	0.25	-66.66	WTRF
Lagoa das Patas	0.26	-66.68	TSFO

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Tabela An1 – continued from previous page

Site	Lat	Lon	Biome 6k
Piusbi	1.66	-77.89	TRFO
Pitalito	1.75	-76.50	TSFO
Piagua	2.30	-76.50	WEFO
Pantano de Genagra	2.50	-76.50	WEFO
Rio Timbio	2.50	-76.50	WEFO
Loma Linda	3.22	-73.35	TDFO
Lago Agua Sucia	3.46	-73.54	STEP
El Gobernador	3.95	-75.00	WTRF
La Guitarra	4.00	-74.28	CTRF
La Primavera	4.00	-74.13	COMI
Carimagua	04.04	-74.14	CGSS
La Rabona	04.05	-74.25	CTRF
El Piñal	04.09	-70.40	TDFO
Alsacia	04.09	-74.11	COMI
Andabobos	04.09	-74.15	CGSH
De la América	4.33	-74.00	CTRF
Laguna Angel	4.45	-70.54	TDFO
Libano	4.50	-75.50	COMI
de Pedro Palo III	4.50	-74.41	WTRF
Paramo Palacio	4.76	-73.88	CGSH
Greja	4.86	-73.70	CGSH
Sardinas	4.95	-69.45	TDFO
Herrera	5.00	-73.91	COMI
Agua Blanca	5.0	-74.45	CTRF
El Abra II	05.02	-73.96	CTRF
Paramo de Peña Negra	05.09	-74.09	COMI
ODP site 932	5.18	-47.03	WTRF
Comprida	5.18	-47.63	TRFO

Continued on next page

Tabela An1 – continued from previous page

Site	Lat	Lon	Biome 6k
Geral	5.18	-47.53	TRFO
Paramo de Laguna Verde	5.25	-74.00	CGSH
Fúquene II	5.50	-73.87	CTRF
Cienaga del Visitador	6.13	-72.83	CGSH
Valle de Lagunillas	6.50	-72.34	CTRF
La Chonta	8.00	-82.00	COMI
La Yeguada,	8.43	-80.78	TSFO
Valle Laguna Victoria	8.80	-70.79	CGSH
Paramo de Miranda	8.91	-70.85	CGSH
Panama	9.00	-80.87	TSFO
Paramo Piedras Blancas	9.16	-70.83	CGSH
Lago de las Morrenas	9.50	-83.49	CTRF
Lake Valencia	10.32	-67.75	TDFO
Lago Quexil	16.92	-89.88	WEFO
Lake Peten-Itza	17.25	-90.00	WEFO
Quila	19.30	-99.20	WAMF
Lake Texcoco	19.44	-99.12	WAMF
Chalco Lake	19.50	-99.00	WAMF
Lake Pátzcuaro	19.58	-101.58	WAMF
San Jose Chulchaca	20.86	-90.13	WEFO
Lake Coba	20.86	-87.55	WEFO

APÊNDICE A

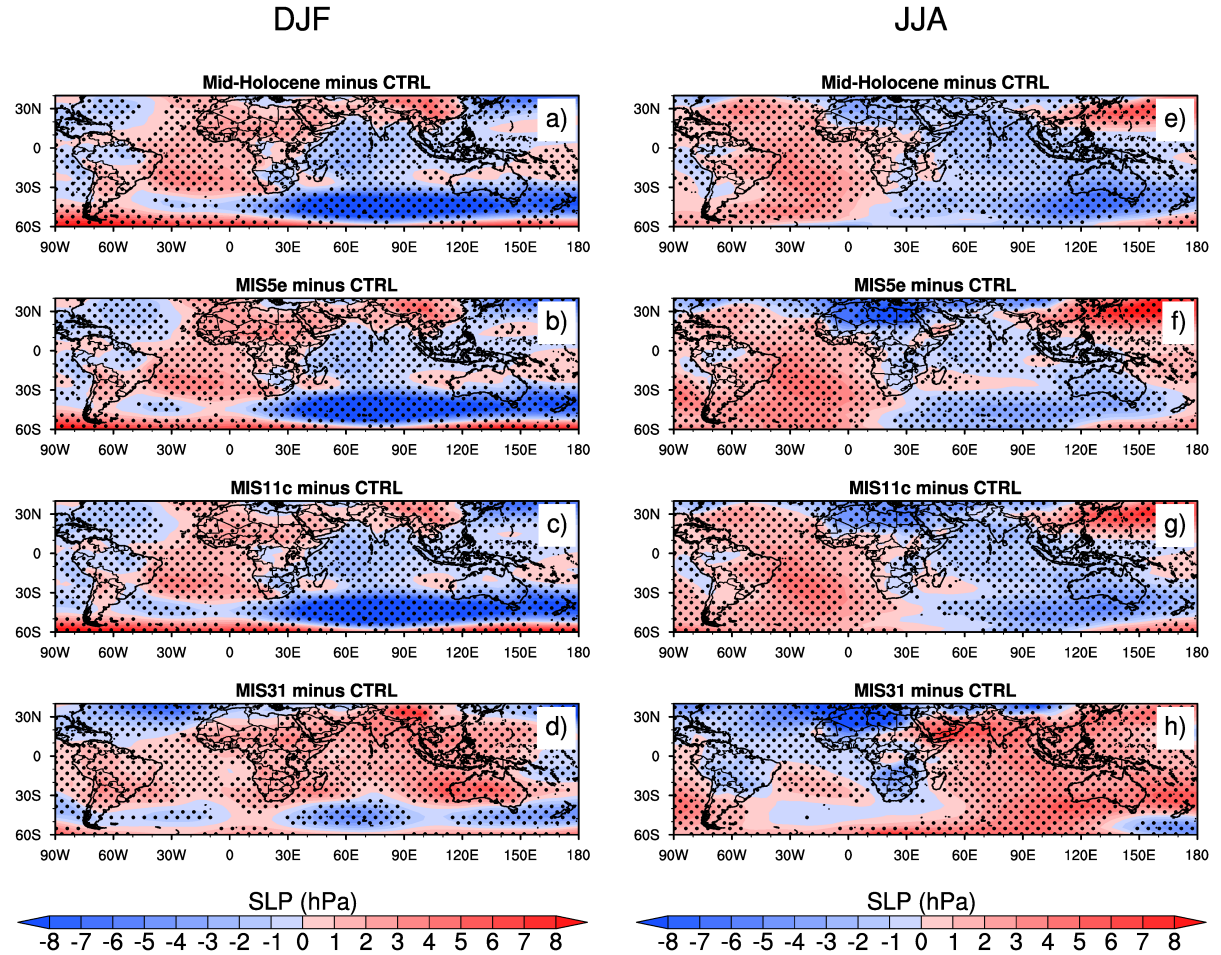


Figura A1 – Spatial distribution of season (DJF and JJA) mean surface SLP difference between interglacials experiments; MH minus CTRL (a,e), MIS5e minus CTRL (b,f), MIS11c minus CTRL (c,g) and MIS31 minus CTRL (d,h). Black dots represent correspond to statistically significant anomalies at the 95% confidence interval.

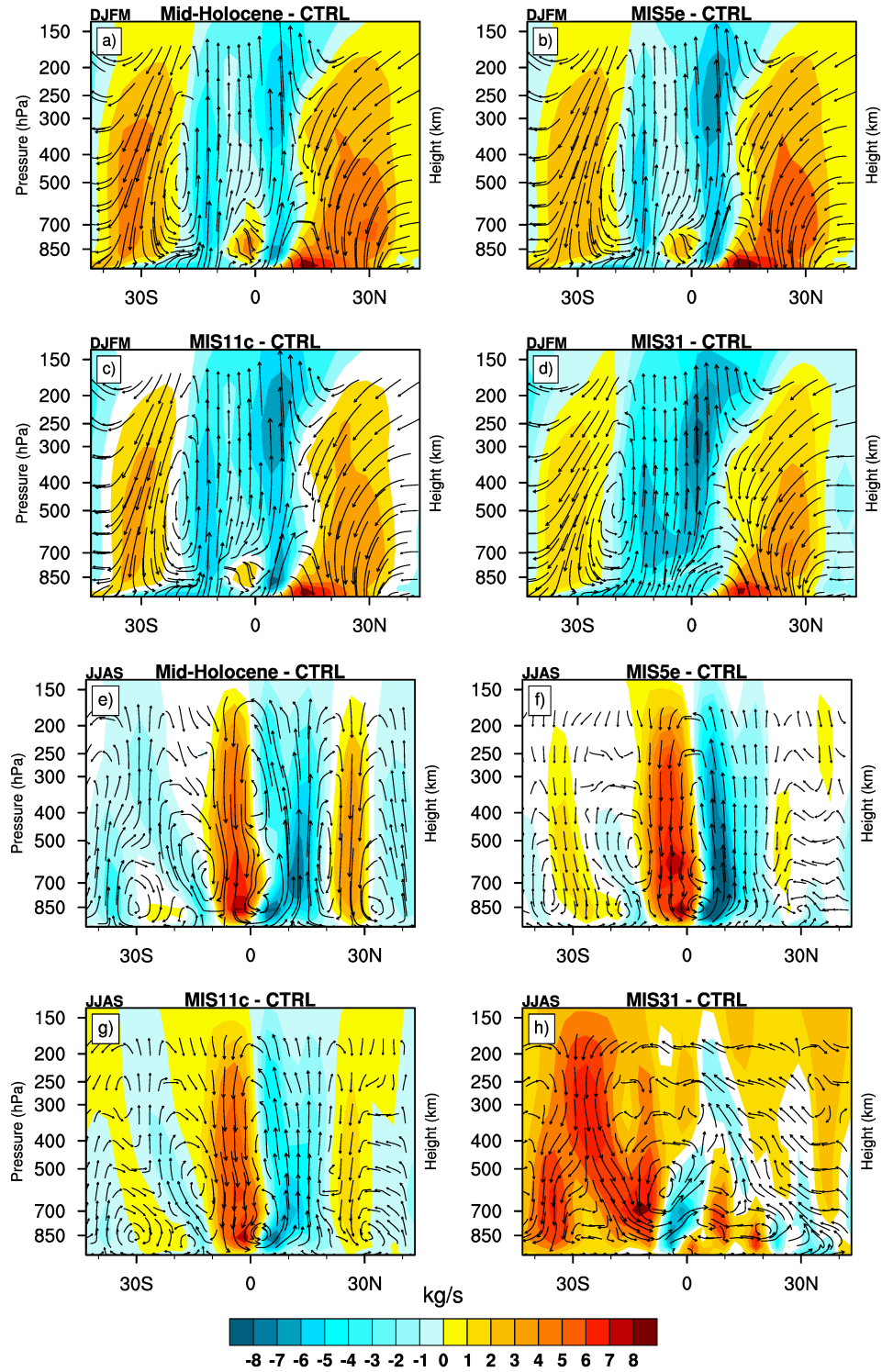


Figura A2 – Hadley circulation from the difference between summer (DJFM) composite anomalies MH-CTRL (a), MIS5e-CTRL (b), MIS11c-CTRL (c) and MIS5e-CTRL (d); and winter (JJAS) MH-CTRL (e), MIS5e-CTRL (f), MIS11c-CTRL (g) and MIS5e-CTRL (h). Color bar indicates the value of mass stream function of mean meridional circulation. The contours intervals represents 10^9 kg/s.

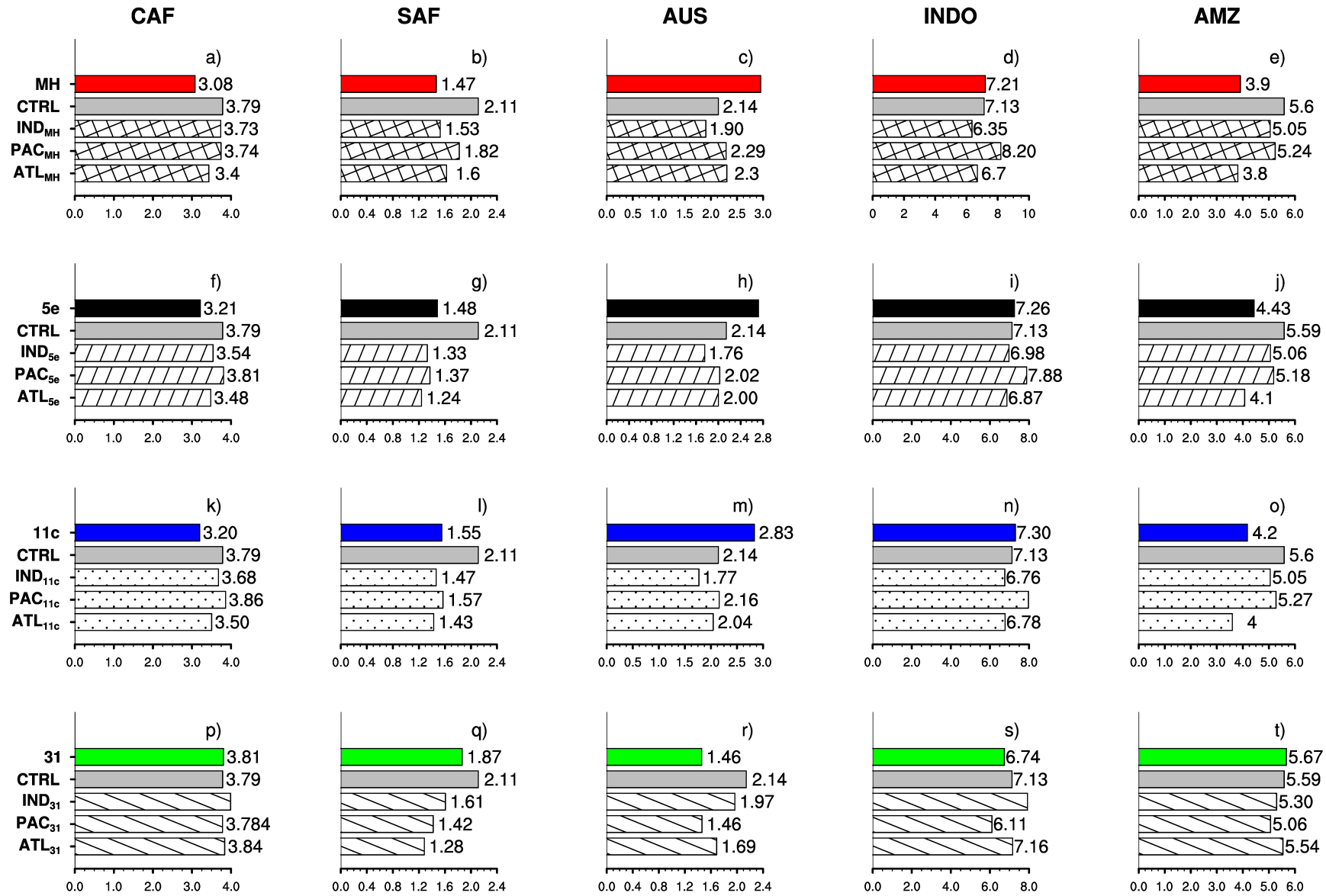


Figura A3 – Areal averaged precipitation (mm/day) CAF, SAF, AUS, IND and AMZ, region for the SST sensitivity experiments.

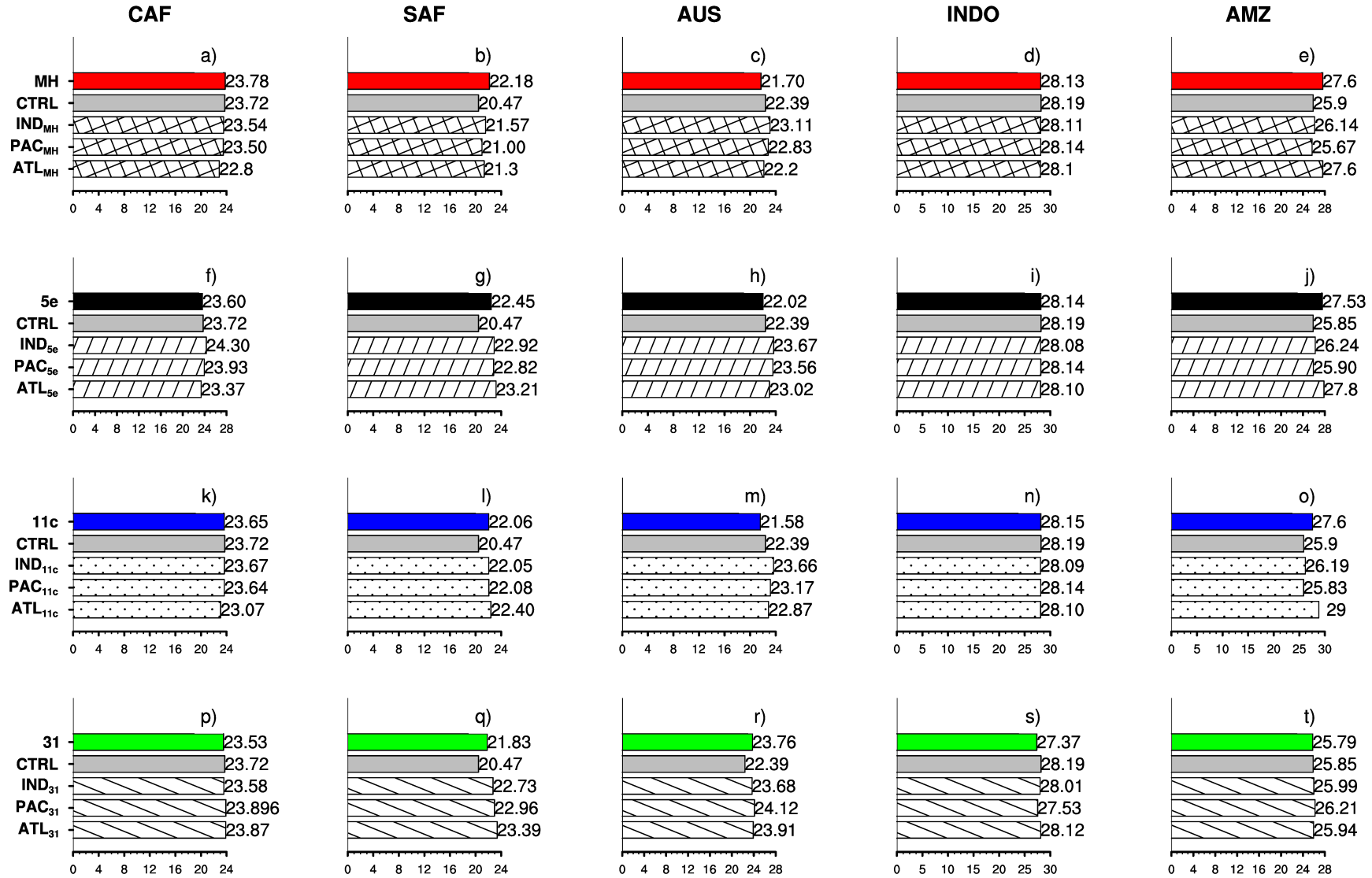


Figura A4 – Areal averaged temperature (mm/day) CAF, SAF, AUS, IND and AMZ, region for the SST sensitivity experiments.

Tabela A1 – Correlation coefficients of mean time series between the Niño3.4 index and the precipitations over areas of the tropical regional. Bold values indicate statistical significance at the 95% level.

	MH	MIS5e	MIS11c	MIS31
Austral Summer (DJFM)				
Index	Niño ₃₄			
Precipitation Domain				
CAF	-0.39	-0.24	0.34	0.12
SAF	-0.24	-0.12	-0.11	0.13
AUS	-0.35	-0.1	0.17	-0.47
INDO	-0.23	-0.14	0.29	-0.25
AMZ	-0.19	-0.13	-0.1	0.17