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Annex V: Monsoons

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AV.1 Introduction

A monsoon refers to a seasonal transition of regimes in atmospheric circulation and precipitation in response to the annual cycle of solar insolation and the distribution of moist static energy (Wang and Ding, 2008; Wang et al., 2014; Biasutti et al., 2018). A global monsoon can be objectively identified based on precipitation contrasts in the solstice seasons to encompass all monsoon regions (Wang and Ding, 2008). In AR5, regional monsoon domains were identified starting from the definition of the global monsoon tailored over the continents and adjacent oceans, as in Kitoh et al. (2013). This annex contains the definition of the global monsoon as used in the Sixth Assessment Report (AR6; see Section AV.2); it explains the rationale for the different definition of AR6 regional monsoons compared to Fifth Assessment Report (AR5; see Section AV.3), and provides the definition and basic characteristics of each regional monsoon assessed (Section AV.4).

AV.2 Definition of the Global Monsoon

The concept of the global monsoon (GM) emerged during the second half of the 20th century. The GM represents the leading empirical orthogonal function (EOF) mode of the annual variations of precipitation and circulation in the global tropics and subtropics as a forced response of the coupled climate system to the annual cycle of solar insolation (Wang and Ding, 2008; An et al., 2015; Wang et al., 2017). GM variability represents, to a large extent, changes in the Inter-tropical Convergence Zone (ITCZ) and associated Hadley circulation (Wang et al., 2014). Changes in the GM have been attributed to both internal variability and external forcings, ranging from interannual to millennial and orbital time scales (Wang et al., 2014, 2017; An et al., 2015; Geen et al., 2020). In AR6, the global monsoon is defined as the area in which the annual range (local summer minus local winter) of precipitation is greater than 2.5 mm day^{-1} (Kitoh et al., 2013), and the domain is represented by the black contour lines in Figure AV.1. Simulation of the global monsoon and its variability is the subject of coordinated modelling experiments in the Global Monsoon Model Intercomparison Project (GMMIP; Zhou et al., 2016). Past changes, simulation and attribution, and projections of the GM are assessed in Sections 2.3.1.4.2, 3.3.3.2 and 4.4.1.4, respectively.

AV.3 Rationale for Regional Monsoon Definitions in AR6

The definition of the regional monsoons has been slightly modified in AR6 with respect to AR5, starting from the consideration that some of the continental areas identified using the global metric have a seasonality in precipitation that is not necessarily of monsoon origin. In particular, the dotted regions in Figure AV.1 located over South Africa, Central America and equatorial South America have a strong seasonality in precipitation but their qualification as monsoons is a subject of discussion. In the assessment of the regional monsoons in Sections 8.3.2.4 and 8.4.2.4, these regions are not considered as distinct regional monsoons, but they are discussed in Box 8.2, which is dedicated to changes in water cycle seasonality. The domains of the regional monsoons in AR6 are defined based on published literature and expert judgement, and accounting for the fact that the climatological summer monsoon rainy season varies across the individual monsoon regions. As shown in Figure AV.1, the AR6 regional monsoons are: South and South East Asian (Section AV.4.1), East Asian (Section AV.4.2), West African (Section AV.4.3), North American (Section AV.4.4), South American (Section AV.4.5) and Australian–Maritime Continent (Section AV.4.6) monsoons. For each region, the definition, regional justification and key features are provided, along with cross references to the main areas of assessment in AR6.

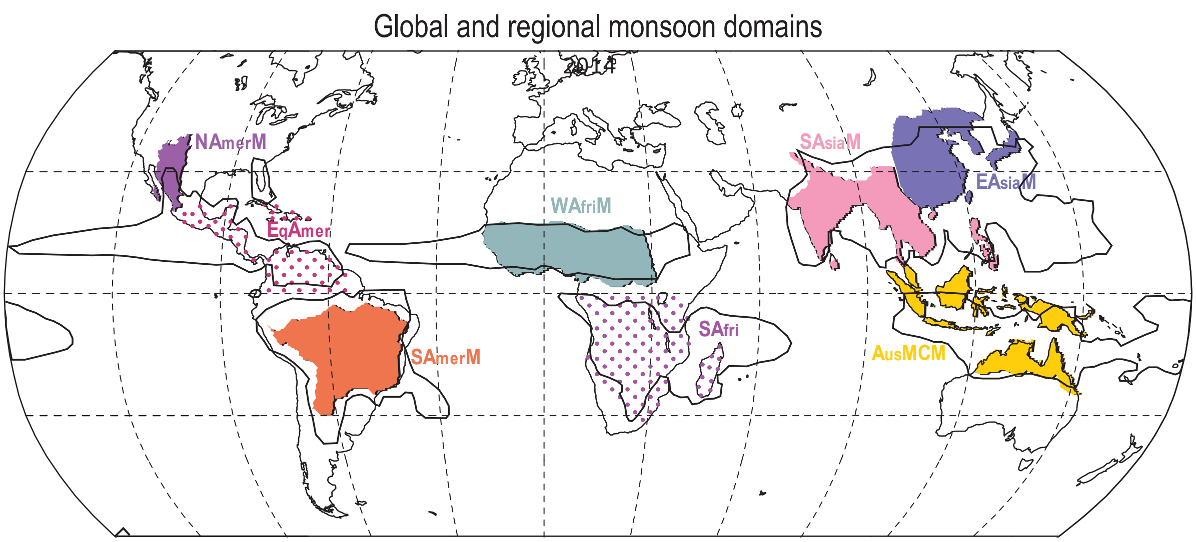

Figure AV.1 | Global and regional monsoon domains. The AR6 global monsoon area is represented by the black contour lines. The AR6 regional monsoons are: North American monsoon (NAmerM, shaded magenta), South American monsoon (SAmerM, shaded dark orange), West African monsoon (WAfriM, shaded grey), South and South East Asian monsoon (SAsiaM, shaded pink), East Asian monsoon (EAsiaM, shaded purple) and Australian–Maritime Continent monsoon (AusMCM, shaded yellow). Areas over Central America and equatorial South America and southern Africa (dotted red and dotted magenta, respectively) are highlighted but not identified as specific regional monsoons (see explanation in the main text). For each regional monsoon, the seasonal characteristics associated with each domain are specified in the main text.

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AV.4 Definition of Regional Monsoons

AV.4.1 The South and South East Asian Monsoon

The South and South East Asian monsoon (SAsiaM) is characterized by pronounced seasonal reversal of wind and precipitation. It covers vast geographical areas and several countries including India, Bangladesh, Nepal, Myanmar, Sri Lanka, Pakistan, Thailand, Laos, Cambodia, Vietnam and the Philippines (Pant and Rupa Kumar, 1997; Goswami, 2006; Gadgil et al., 2010; Shige et al., 2017), with a domain roughly extending across 60°E–110°E and 10°S–25°N as shown in Figure AV.1 (shaded pink). The SAsiaM is unique in its geographical features because of the orography surrounding the area (i.e., the Himalayas, Western Ghats and Arakan Yoma mountains, and the Tibetan Plateau to the north) and the adjacent Indian Ocean.

The SAsiaM rainy season from June to September contributes to more than 75% of the annual rainfall over much of the region, including the southern slopes of the central and eastern Himalayas (Krishnan et al., 2019b). Considering the spatial domain of the SAsiaM, monsoon precipitation maxima are located over the west coast, north-east and central/north India as well as Myanmar and Bangladesh, whereas minima are located over north-west and south-eastern India, western Pakistan, and south-eastern and northern Sri Lanka (Pant and Rupa Kumar, 1997; Gadgil et al., 2010). Prior to the SAsiaM rainy season, areas in the north-western Himalaya receive precipitation during winter and early spring from so-called 'western disturbances', which are extratropical synoptic systems originating over the Mediterranean region and propagating eastward along the subtropical westerly jet (Madhura et al., 2014; Cannon et al., 2015; Dimri et al., 2015; Hunt et al., 2018; Krishnan et al., 2019a, b).

The climatological onset of the SAsiaM occurs around 20 May over the Andaman and Nicobar Islands and covers the central Bay of Bengal around 25 May, while it simultaneously advances into mainland India from the south through Kerala (Pai et al., 2020). The normal date of monsoon onset over Kerala is 1 June and the monsoon rains progress into India both from south to north and east to west, so as to cover the entire country by 15 July (Pai et al., 2020), although with large interannual and inter-decadal variations (Ghanekar et al., 2019). Retreat of the SAsiaM typically begins from far north-west India around 1 September and withdraws completely from the country by 15 October; this is followed by the establishment of the north-east monsoon rainy season over south peninsular India from October through December (Pai et al., 2020).

The SAsiaM exhibits prominent rainfall variability on sub-seasonal, interannual and inter-decadal time scales with teleconnections to modes of climate variability (see Webster et al., 1998; Turner and Annamalai, 2012). While the gross features of the SAsiaM are simulated in GCMs, there are several large biases that have persisted over generations of climate models, in CMIP3 and CMIP5, including a large dry bias over India coupled to a lower-tropospheric circulation that is too weak (Sperber et al., 2013) and a related wet bias over the western Equatorial Indian Ocean (Bollasina and Ming, 2013). In CMIP5, cold Arabian Sea SST (sea surface temperature) biases in coupled GCMs were shown to worsen the monsoon dry bias by limiting the available moisture (Levine et al., 2013; Sandeep and Ajayamohan, 2015). Some improvements to the spatial pattern have been noted in CMIP6, particularly near orography (Gusain et al., 2020). The teleconnection to the El Niño–Southern Oscillation (ENSO) provides the main prospect for seasonal prediction and was too weak in CMIP5 (Sperber et al., 2013). The boreal summer intra-seasonal oscillation (BSISO) controls the majority of sub-seasonal variations in SAsiaM rainfall, as well as affecting the East Asian monsoon region. While CMIP5 models represented the BSISO better than CMIP3 (especially its characteristic northward propagation), the spatial pattern is still poorly simulated in most models (Sabeerali et al., 2013).

The SAsiaM is assessed in Sections 8.3.2.4.1, 8.4.2.4.1, Atlas.5.3.2 and Atlas.5.3.4. One of its main components, the Indian summer monsoon, is assessed in Section 10.6.3 as a case study on the construction of regional climate information from the distillation of multiple lines of evidence. Climate change over the Hindu Kush Himalaya is assessed in Cross-Chapter Box 10.4.

AV.4.2 The East Asian Monsoon

The East Asian monsoon (EAsiaM) is the seasonal reversal in wind and precipitation occurring over East Asia, including eastern China, Japan and the Korean Peninsula. The continental area influenced by this monsoon is roughly bounded by 102.5°E–140°E and 20°N–40°N (e.g., Wen et al., 2000; Wang and LinHo, 2002; Wang et al., 2010), and is shaded purple in Figure AV.1. Unlike other monsoons, it extends quite far north, out of the tropical belt, and it is largely influenced by subtropical systems, such as the western North Pacific subtropical high, East Asian subtropical westerly jet, and by disturbances from the mid-latitudes (Chang et al., 2000; Lee et al., 2006; Yim et al., 2008; Zhou et al., 2009; S.-S. Lee et al., 2013).

The EAsiaM manifests during boreal summer with warm and wet southerly winds, but also during boreal winter with cold and dry northerly winds (Ha et al., 2012). The winter component has been linked to the subsequent summer (e.g., Wen et al., 2000, 2016; Ge et al., 2017; Yan et al., 2020).

AV.4.2.1 The East Asian Summer Monsoon

The East Asian summer monsoon is a subtropical monsoon system (e.g., Wang and LinHo, 2002; Ding et al., 2007). It is characterized by low-level southerly winds prevailing over eastern China, Korea and Japan in boreal summer. The monsoon flow brings abundant water vapour into East Asia that converges and forms the Meiyu/Baiu/Changma rain belt over the region (Zhou et al., 2009; Jin et al., 2013; Lee et al., 2017). Rainfall onset occurs in late April/early May in the central Indochina Peninsula, and in mid-June the rainy season arrives over East Asia with the formation of the Meiyu front along the Yangtze River valley, Changma in Korea, and Baiu in Japan. Later in July, the monsoon advances up to north China, the Korean Peninsula and central Japan (Zhang et al., 2004; Yihui and Chan, 2005; Zhou et al., 2009).

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Intra-seasonal variability of the EAsiaM has been mostly related to the Madden–Julian Oscillation (MJO) (Yasunari, 1979; Zhang et al., 2009; Chen et al., 2015), to the phase of the Pacific-Japan mode (Nitta, 1987) or the Indian summer monsoon (Li et al., 2018) and to the BSISO (Kikuchi et al., 2012; Chen et al., 2015), particularly for its onset (J.-Y. Lee et al., 2013). At interannual time scales, changes in East Asian summer monsoon intensity lead to a position shift of the monsoon rain belt under the influence of ENSO (Wang et al., 2000) and the western North Pacific subtropical high (Kosaka et al., 2013; Wang et al., 2013), Arctic sea ice (Guo et al., 2014), and solar and volcanic forcing (Peng et al., 2010; Man et al., 2012; Cui et al., 2014). Variability at inter-decadal time scales is more prominent in the second half of the 20th century (Jiang, 2005; Ding et al., 2008; Wang et al., 2018), and a specific assessment about this aspect is provided in Section 8.3.2.4.2.

The basic features and interannual variations of the East Asian summer monsoon are well reproduced in climate models. For example, the climatological circulation structure is well reproduced in both atmospheric and coupled GCMs (Song and Zhou, 2013; Song et al., 2013; Jiang et al., 2016, 2020), and the relationship between the monsoon and ENSO is well represented (Sperber et al., 2013; Fu and Lu, 2017). In CMIP5 models, the main shortcomings relate to missing rainfall bands around 30°N and the northward shift of the western North Pacific subtropical high (Huang et al., 2013). Most coupled models show an inadequate strength of monsoon circulation over southern East Asia, and with little change in model performance from the Third Assessment Report (TAR) to AR6 (Jiang et al., 2016, 2020). In coupled model simulations, air–sea coupling helps improve the climatology and interannual variability of rainfall over the East Asia monsoon region (Wang et al., 2005; Song and Zhou, 2014).

The summer (June–July–August) component of the EAsiaM is assessed in Sections 8.3.2.4.2, 8.4.2.4.2, Atlas.5.1.2 and Atlas.5.1.4.

AV.4.2.2 The East Asian Winter Monsoon

The East Asian winter monsoon (EAWM) is characterized by strong north-westerlies over north-east China, Korea and Japan, and by strong north-easterlies along the coast of East Asia (Huang et al., 2003). The northerly winds extend from mid-latitude East Asia to the equatorial South China Sea (Wen et al., 2000; Wang et al., 2010). The EAWM has one component mostly linked to mid-to-high latitude circulation systems and another mostly linked to the tropical circulation and largely controlled by ENSO (Ge et al., 2012; Chen et al., 2014a). The mid-latitude component has a close relationship with autumn Arctic sea ice concentration changes (Chen et al., 2014b).

The EAWM exhibits significant variability ranging from intra-seasonal to inter-decadal time scales. Its intra-seasonal variability is suggested to be influenced by both high-latitude and subtropical Eurasian wave trains (Jiao et al., 2019). At the interannual time scale, ENSO is an important factor modulating the EAWM (Wang et al., 2000), while the relationship between them is not stable (Wang and He, 2012; Fan et al., 2020). In addition, Arctic Oscillation (Gong et al., 2001), Arctic sea ice (Ge et al., 2012), Eurasian snow (Luo and Wang, 2019), and strong volcanic eruptions (Miao et al., 2016) also play vital roles in changing the EAWM intensity. At the inter-decadal time scale, the EAWM weakened significantly in the mid-1980s, which resulted from atmospheric intrinsic quasi-stationary planetary waves (Wang et al., 2009) and external forcings (Miao et al., 2018). In the mid-2000s, the EAWM was observed to recover from its weak epoch (Wang and Chen, 2014).

The large-scale features of the EAWM are well reproduced by climate models, although the strength of monsoon circulation is underestimated. The ability of coupled models to simulate the EAWM shows little difference through the TAR to AR5, but has improved from AR5 to AR6 (Jiang et al., 2016, 2020). In CMIP5 models, reasonable simulations of the Siberian High and Aleutian Low intensities and the relationship with ENSO help improve the climatology and interannual variability of the EAWM (Gong et al., 2014).

Model simulations of the winter component of the EAsiaM are assessed in chapter Atlas.5.1.3.

AV.4.3 The West African Monsoon

The West African monsoon (WAfriM) is a seasonal reversal in wind and precipitation extending over a vast and contrasted geographical region, from the equator to the margins of the Sahara, and from the Atlantic coast inland. The WAfriM domain includes Togo, Guinea Bissau, Gambia, Senegal, Mauritania, Guinea, Sierra Leone, Liberia, Mali, Ivory Coast, Burkina Faso, Ghana, Benin, Chad, Cape Verde, northern Cameroon, Niger, Nigeria and the northern Central African Republic. It is roughly bounded by 18°W–30°E, 5°N–18°N, and is shaded grey in Figure AV.1 (e.g., Adedokun, 1978). The West African monsoon and Sahel are sometimes considered interchangeably. However, the Sahel lies in the northern part of the WAfriM region, often limited to the west and central north Africa (e.g., Nicholson et al. (2018) considered the domain to be 20°W–30°E, 10°N–20°N) or sometimes extended to the east (e.g., Giannini and Kaplan (2019) considered the domain to be 20°W–40°E, 10°N–20°N). The East African region with the Greater Horn of Africa, which includes Ethiopia, Sudan and South Sudan, lies on the fringes of both the West African and Indian monsoons (Nicholson, 2017).

The WAfriM is characterized by the northward progression from May to September of moist low-level south-westerlies from the Gulf of Guinea, meeting the dry north-easterlies (Harmattan) from the Saharan Heat Low at the inter-tropical discontinuity (e.g., Hamilton et al., 1945; Omotosho, 1988). In May and June, rainfall remains essentially along the Guinean coast with a maximum occurring near 5°N, then the rainfall maximum jumps suddenly over the Sudan–Sahel zone near 12°N, followed by a 'little dry season' over the Guinea coast (Adejuwon and Odekunle, 2006). This apparent shift is known as the West African monsoon jump and it is concomitant with the monsoon onset over the Sahel (Sultan and Janicot, 2003; Cook, 2015). Rainfall continues to progress towards the north up to about 18°N–20°N. The rainfall maximum occurs in the Sahel in August/September, followed by a rapid retreat of rainfall to the Guinean coast and a second maximum occurs over this region in October/November.

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The WAfriM features variability at different time scales. Unprecedented droughts occurred in West Africa and particularly over the Sahel from the late 1960s to the mid-1990s, and a specific assessment of the understanding of mechanisms related to these changes is provided in Sections 8.3.2.4.3 and 10.4.2.1.

At interannual time scales, tropical oceans (Atlantic, Pacific and Indian) appear to be the major drivers with their SST anomalies leading to variations in the accumulated seasonal rainfall (e.g., Lamb, 1978; Diakhaté et al. 2019). At intra-seasonal time scales, equatorial waves (e.g., Mekonnen et al., 2008; Janicot et al., 2010) and interactions with mid-latitudes and the Mediterranean (e.g., Vizy and Cook, 2009; Roehrig et al., 2011) have an effect on WAfriM activity. African Easterly Waves (AEWs) and mesoscale convective systems (MCSs), including squall lines, are the prominent weather synoptic scale aspects of the WAfriM, supplying almost all rainfall in the Sahel. The strong coupling between AEWs and MCSs has been investigated in depth, as well as their interactions with the Saharan Heat Low, the moisture supply from East Africa and the Mediterranean region or from near the equator (e.g., Diongue et al., 2002; Brammer and Thorncroft, 2017; Lafore et al., 2017). Land surface processes are known to influence WAfriM precipitation (Boone et al., 2016).

Simulation of West African climate, including the monsoon, has received specific attention under coordinated programmes in the recent past: the African Multidisciplinary Monsoon Analysis Model Intercomparison Project (AMMA-MIP), the AMMA Land-surface Model Intercomparison Project (ALMIP), the West African Monsoon Modelling and Evaluation (WAMME) project and the Coordinated Regional Downscaling Experiment (CORDEX; Raj et al., 2019). CMIP5 and CMIP6 struggle to reproduce the amplitude of observed decadal variability in the 20th century and to represent the mean climatology including the northward propagation of the monsoon (Roehrig et al., 2013; Monerie et al., 2020; Sow et al., 2020). A higher horizontal resolution improves the representation of the intensity and spatial distribution of WAfriM rainfall and related circulation, because of the effects of vegetation, orography and coastlines (Hourdin et al., 2010; Sylla et al., 2010; Xue et al., 2010; Raj et al., 2019).

The WAfriM is assessed in Sections 8.3.2.4.3 and 8.4.2.4.3. The observed Sahel and West African monsoon drought and recovery is assessed as a regional climate change attribution example in Section 10.4.2.1.

AV.4.4 The North American Monsoon

The North American monsoon (NAmerM) is a regional-scale atmospheric circulation system dominated by boreal summer precipitation over north-western Mexico and south-west USA, where it contributes to almost 70% and 40% of the total annual precipitation, respectively (Douglas et al., 1993; Higgins et al., 1997). The NAmerM domain is shaded magenta in Figure AV.1, is roughly bounded by 15°N–35°N, 100°W–115°W and is defined where the July–August–September minus June mean precipitation exceeds 25 mm month^{-1} (Douglas et al., 1993; Adams and Comrie, 1997; Barlow et al., 1998; Cook and Seager, 2013).

The identification of a regional monsoon regime in North America dates back to the 1990s (Douglas et al., 1993; Adams and Comrie, 1997; Higgins et al., 1997; Barlow et al., 1998), although consideration of the monsoonal character of the south-western USA precipitation goes back considerably further (e.g., Bryson and Lowry, 1955). The monsoonal characteristics of the region include a pronounced annual maximum of precipitation in boreal summer (June–July–August) accompanied by a surface low-pressure system and an upper-level anticyclone, although the seasonal reversal of the surface winds is primarily limited to the northern Gulf of California. In particular, the summer precipitation in the NAmerM region is dictated by the location of the upper-level anticyclone (Reed, 1933; Castro et al., 2001). The decay phase of the NAmerM is typically observed during late September to October, when convection migrates from Central to South America (Vera et al., 2006).

Mesoscale variability of the NAmerM comes from the pulsing of the Gulf of California low-level jet, the intensification/reduction of the land–sea contrast (Torres-Alavez et al., 2014) and moisture surges over the Gulf of California (Vera et al., 2006). Synoptic variability of the NAmerM is mainly associated with the activity of tropical cyclones and of easterly waves (Stensrud et al., 1997; Fuller and Stensrud, 2000). In addition, the NAmerM is strongly influenced by ENSO variability at both interannual (Higgins et al., 1999; Higgins and Shi, 2001) and decadal (e.g., Castro et al., 2001) time scales.

The region is challenging for climate modelling for several reasons, including complex topography, the importance of Mesoscale Convective Systems (MCSs), and sensitivity to SST bias (e.g., Pascale et al., 2019). Many CMIP3 and CMIP5 models with resolutions coarser than 100 km are unable to realistically resolve the topography of the NAmerM region, thus inducing biases in simulating the monsoon (Geil et al., 2013). Among other factors, these biases are due to deficiencies in the simulation of the Gulf of California summer low-level flow (Kim et al., 2008; Pascale et al., 2017) and to failures in representing properly the diurnal cycle (Risanto et al., 2019) and the decay (Bukovsky et al., 2015) of the NAmerM precipitation. Simulations at higher horizontal resolution (i.e., with at least 0.25° grid) exhibit an improved representation of the regional topography, which provides a better representation of the regional circulation and therefore of the NAmerM (Varuolo-Clarke et al., 2019).

The NAmerM is assessed in Sections 8.3.2.4.4, 8.4.2.4.4, Atlas.7.1.3 and Atlas.9.1.

AV.4.5 The South American Monsoon

The South American monsoon (SAmerM) is a regional circulation system characterized by inflow of low-level winds from the Atlantic Ocean toward South America, involving Brazil, Peru, Bolivia and northern Argentina, associated with the development of surface pressure gradients (and intense precipitation) during austral summer (December–January–February; DJF). Based on climatological precipitation intensity, the SAmerM region is roughly

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bounded by 5°S–25°S and 70°W–50°W (Zhou and Lau, 1998; Vera et al., 2006; Raia and Cavalcanti, 2008) and is shaded dark orange in Figure AV.1.

During austral spring (September–October–November; SON), areas of intense convection migrate from north-western South America to the south (Raia and Cavalcanti, 2008), forming the South Atlantic Convergence Zone (SACZ) during austral summer (Kodama, 1992; Jones and Carvalho, 2002; Vera et al., 2006). Associated with this regime, an upper-tropospheric anticyclone (the Bolivian High) forms over the Altiplano region during the monsoon onset (Lenters and Cook, 1997). The establishment of this upper-level anticyclone has been related to the transition from southerly to northerly winds and the occurrence of strong convective heating over the Amazon (Lenters and Cook, 1997; Wang and Fu, 2002). The SAmerM then retreats during austral autumn (March–April–May) with a north-eastward migration of the convection (e.g., Vera et al., 2006).

The SAmerM displays considerable variability on time scales raging from intra-seasonal to decadal (Vera et al., 2006; Marengo et al., 2012; Vuille et al., 2012; Novello et al., 2017). The Madden-Julian Oscillation (MJO, Section AIV.2.8) influences the SACZ via changes in mid-latitude synoptic disturbances (Jones and Carvalho, 2002; Liebmann et al., 2004). At interannual time scales, ENSO explains most of the SAmerM variability (e.g., Paegle and Mo, 2002; Marengo et al., 2012). Tropical Atlantic temperatures also affect the SAmerM, with reduced atmospheric moisture transport to feed the monsoon under warmer tropical North Atlantic conditions (e.g., Marengo et al., 2008; Zeng et al., 2008). In addition to SSTs, interannual variability of the SAmerM is linked to changes in land surface processes, cold-air incursions, the latitudinal location of the subtropical jet, and the Southern Annular Mode (Section AIV.2.2; e.g., Silvestri and Vera, 2003; Li and Fu, 2004; Collini et al., 2008; Yin et al., 2014). At inter-decadal time scales, the SAmerM is influenced by important modes of climate variability (e.g., Robertson and Mechoso, 2000; Paegle and Mo, 2002; Chiessi et al., 2009; Silvestri and Vera, 2009).

The general large-scale features of the SAmerM are reasonably well simulated by coupled climate models although they do not adequately reproduce maximum precipitation over the core of the monsoon, even when considering simulations under past natural forcings, such as those during the last millennium (Rojas et al., 2016; Díaz and Vera, 2018). However, CMIP5 models featured an improved representation of the SAmerM with respect to CMIP3 (Joetzjer et al., 2013; Jones and Carvalho, 2013; Gulizia and Camilloni, 2015; Díaz and Vera, 2017).

The SAmerM is assessed in Sections 8.3.2.4.5 and 8.4.2.4.5.

AV.4.6 The Australian-Maritime Continent Monsoon

The Australian-Maritime Continent monsoon (AusMCM) occurs during austral summer (December–January–February), with the large-scale shift of the ITCZ into the Southern Hemisphere. It covers northern Australia and the Maritime Continent up to 10°N (e.g., McBride, 1987; Suppiah, 1992; Robertson et al., 2011), and it corresponds to the yellow shaded region in Figure AV.1. The identification of the Australian monsoon by meteorologists dates back to the early 20th century (see review of Suppiah, 1992), with later studies providing classifications of monsoon circulation regimes (e.g., McBride, 1987) and definitions of monsoon onset (Troup, 1961; Holland, 1986; Hendon and Liebmann, 1990; Drosdowsky, 1996).

The AusMCM is characterized by the seasonal reversal of prevailing easterly winds to westerly winds and the onset of periods of active convection and heavy rainfall (Zhang and Moise, 2016). Over northern Australia, the monsoon season generally lasts from December to March and is associated with the west to north-westerly inflow of moist winds, producing convection and heavy precipitation. Over the Maritime Continent, the main rainy season south of the equator is centred on December to February with north-westerly monsoon flow at low levels.

Over Australia, the monsoon is strongly influenced by ENSO on interannual time scales: during El Niño years the monsoon onset tends to be delayed (Nicholls et al., 1982; McBride and Nicholls, 1983; Drosdowsky, 1996). This relationship breaks down after the onset of the wet season, leading to little correlation between ENSO phase and total monsoon rainfall or duration (e.g., Hendon et al., 2012). The Maritime Continent also experiences a delay in monsoon onset during El Niño years and monsoon rainfall is correlated with ENSO during the dry and transition seasons (Robertson et al., 2011). The AusMCM is also influenced by the Indian Ocean Dipole (peaking in September to November) that tends to weaken the following monsoon when in its positive phase (Cai et al., 2005).

The ability of climate models to simulate the Australian monsoon has improved in successive generations of coupled models (i.e., from CMIP3 to CMIP6, Moise et al., 2012; Brown et al., 2016; Narsey et al., 2020), with sensitivity of monsoon rainfall to the magnitude of SST biases in the Equatorial Pacific (Brown et al., 2016).

The AusMCM is assessed in Sections 8.3.2.4.6 and 8.4.2.4.6.

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