confidence that the Indian summer monsoon circulation will weaken, but this is compensated by increased atmospheric moisture content, leading to more precipitation. For the East Asian summer monsoon, both monsoon circulation and precipitation are projected to increase. There is *low confidence* that over the Maritime Continent boreal summer rainfall will decrease and boreal winter rainfall will increase. There is *low confidence* that changes in the tropical Australian monsoon rainfall are small. There is *low confidence* that Western North Pacific summer monsoon circulation changes are small, but with increased rainfall due to enhanced moisture. There is *medium confidence* in an increase of Indian summer monsoon rainfall and its extremes throughout the 21st century under all RCP scenarios. Their percentage change ratios are the largest and model agreement is highest among all monsoon regions.

There is *low confidence* in projections of American monsoon precipitation changes but there is high confidence in increases of precipitation extremes, of wet days and consecutive dry days. It is *likely* that precipitation associated with the NAMS will arrive later in the annual cycle, and persist longer. Future changes in the timing and duration of the SAMS are also *likely*, but details of these changes remain uncertain. There is *high confidence* in the expansion of SAMS, resulting from increased temperature and humidity. There is *low confidence* in precipitation changes in SAMS but it is *likely* that precipitation extremes of wet days and consecutive dry days increase.

Based on how models represent known drivers of the West African monsoon, there is *low confidence* in projections of its future development based on CMIP5. Confidence is *low* in projections of a small delay in the onset of the West African rainy season with an intensification of late-season rains.

14.3 Tropical Phenomena

14.3.1 Convergence Zones

Section 7.6 presents a radiative perspective of changes in convection (including the differences between GHG and aerosol forcings), and Section 12.4.5.2 discusses patterns of precipitation change on the global scale. The emphasis here is on regional aspects of tropical changes. Tropical convection over the oceans, averaged for a month or longer, is organized into long and narrow convergence zones, often anchored by SST structures. Latent heat release in convection drives atmospheric circulation and affects global climate. In model experiments where spatially-uniform SST warming is imposed, precipitation increases in these tropical convergence zones (Xie et al., 2010b), following the 'wet-get-wetter' paradigm (Held and Soden, 2006). On the flanks of a convergence zone, rainfall may decrease because of the increased horizontal gradient in specific humidity and the resultant increase in dry advection into the convergence zone (Neelin et al., 2003).

While these arguments based on moist atmospheric dynamics call for changes in tropical convection to be organized around the climatological rain band, studies since AR4 show that such changes in a warmer climate also depend on the spatial pattern of SST warming. As a result of the SST pattern effect, rainfall change does not generally project onto the climatological convergence zones, especially for the annual mean. In CMIP3/5 model projections, annual rainfall change over tropical oceans follows a 'warmer-get-wetter' pattern, increasing where the SST warming exceeds the tropical mean and vice versa (Figure 14.8, Xie et al., 2010b; Sobel and Camargo, 2011; Chadwick et al., 2012). Differences among models in the SST warming pattern are an important source of uncertainty in rainfall projections, accounting for a third of inter-model variability in annual precipitation change in the tropics (Ma and Xie, 2013).

Figure 14.8 presents selected indices for several robust patterns of SST warming for RCP8.5. They include: greater warming in the northern than the southern hemisphere, a pattern favouring rainfall increase at locations north of the equator and decreases to the south (Friedman et al., 2013); enhanced equatorial warming (Liu et al., 2005) that anchors a pronounced rainfall increase in the equatorial Pacific; reduced warming in the subtropical Southeast Pacific that weakens convection there; decreased zonal SST gradient across the equatorial Pacific (see Section 14.4) and increased westward SST gradient across the equatorial Indian Ocean (see Section 14.3.3) that together contribute to the weakened Walker cells.

Changes in tropical convection affect the pattern of SST change (Chou et al., 2005) and such atmospheric and oceanic perturbations are inherently coupled. The SST pattern effect dominates the annual rainfall change while the wet-get-wetter effect becomes important for seasonal mean rainfall in the summer hemisphere (Huang et al., 2013). This is equivalent to an increase in the annual range of precipitation in a warmer climate (Chou et al., 2013). Given uncertainties in SST warming pattern, the confidence is generally higher for seasonal than annual mean changes in tropical rainfall.

[INSERT FIGURE 14.8 HERE]

Figure 14.8: Upper panel: Annual-mean precipitation percentage change (δP/P in green/gray shade and white contours at 20% intervals), and relative SST change (colour contours at intervals of 0.2°C; negative shaded) to the tropical (20°S–20°N) mean warming in RCP8.5 projections, shown as 23 CMIP5 model ensemble mean. Lower panel: SST warming pattern indices in the 23-model RCP8.5 ensemble, shown as the 2081–-2100 minus 1986–2005 difference. From left: Northern (EQ-60°N) minus Southern (60°S-EQ) Hemisphere; equatorial (120°E–60°W, 5°S–5°N) and Southeast (130°W–70°W, 30°S–15°S) Pacific relative to the tropical mean warming; zonal SST gradient in the equatorial Pacific (120°E–180°E minus 150°W–90°W, 5°S–5°N) and Indian (50°E–70°E, 10°S–10°N minus 90°E–110°S, 10°S–EQ) Oceans. Rightmost: spatial correlation (*r*) between relative SST change and precipitation percentage change (δP/P) in the tropics (20°S–20°N) in each model. (The spatial correlation for the multi-model ensemble mean fields in the upper panel is 0.63). The circle and error bar indicate the ensemble mean and ±1 standard deviation, respectively. The upper panel is a CMIP5 update of Ma and Xie (2013), and see text for indices in the lower panel.

14.3.1.1 Inter-Tropical Convergence Zone

The Inter-Tropical Convergence Zone (ITCZ) is a zonal band of persistent low-level convergence, atmospheric convection, and heavy rainfall. Over the Atlantic and eastern half of the Pacific, the ITCZ is displaced north of the equator due to ocean-atmosphere interaction (Xie et al., 2007) and extratropical influences (Kang et al., 2008; Fučkar et al., 2013). Many models show an unrealistic double-ITCZ pattern over the tropical Pacific and Atlantic, with excessive rainfall south of the equator (Section 9.4.2.5.1). This bias needs to be kept in mind in assessing ITCZ changes in model projections, especially for boreal spring when the model biases are largest.

The global zonal mean ITCZ migrates back and forth across the equator following the sun. In CMIP5, seasonal-mean rainfall is projected to increase on the equatorward flank of the ITCZ (Figure 14.9). The co-migration of rainfall increase with the ITCZ is due to the wet-get-wetter effect while the equatorward displacement is due to the SST pattern effect (Huang et al., 2013).

[INSERT FIGURE 14.9 HERE]

Figure 14.9: Seasonal cycle of zonal-mean tropical precipitation change (2081–2100 in RCP8.5 minus 1986–2005) in CMIP5 multimodel ensemble (MME) mean. Eighteen CMIP5 models were used. Stippling indicates that more than 90% models agree on the sign of MME change. The red curve represents the meridional maximum of the climatological rainfall. Adapted from Huang et al. (2013).

14.3.1.2 South Pacific Convergence Zone

The South Pacific Convergence Zone (SPCZ , Widlansky et al., 2011) extends south-eastward from the tropical western Pacific to French Polynesia and the Southern Hemisphere mid-latitudes, contributing most of the yearly rainfall to the many South Pacific island nations under its influence. The SPCZ is most pronounced during austral summer (DJF).

Zonal and meridional SST gradients, trade wind strength, and subsidence over the eastern Pacific are important mechanisms for SPCZ orientation and variability (Takahashi and Battisti, 2007; Lintner and Neelin, 2008; Vincent et al., 2011; Widlansky et al., 2011). Many GCMs simulate the SPCZ as lying east-west, giving a 'double-ITCZ' structure and missing the southeastward orientation (Brown et al., 2012a).

The majority of CMIP models simulate increased austral summer mean precipitation in the SPCZ, with decreased precipitation at the eastern edge of the SPCZ (Brown et al., 2012a; Brown et al., 2012b). The position of the SPCZ varies on interannual to decadal time scales, shifting northeast in response to El Niño (Folland et al., 2002; Vincent et al., 2011). Strong El Niño events induce a zonally-oriented SPCZ located well northeast of its average position, while more moderate ENSO (Section 14.4) events are associated with movement of the SPCZ to the northeast or southwest, without a change in its orientation.

Models from both CMIP3 and CMIP5 that simulate the SPCZ well show a consistent tendency towards much more frequent zonally-oriented SPCZ events in future (Cai et al., 2012b). The mechanism appears to be associated with a reduction in near-equatorial meridional SST gradient, a robust feature of modelled SST response to anthropogenic forcing (Widlansky et al., 2013). An increased frequency of zonally-oriented SPCZ events would have major implications for regional climate, possibly leading to longer dry spells in the southwest Pacific.

14.3.1.3 South Atlantic Convergence Zone

The South Atlantic Convergence Zone (SACZ) extends from the Amazon region through southeastern Brazil towards the Atlantic Ocean during austral summer (Cunningham and Cavalcanti, 2006; Carvalho et al., 2011; de Oliveira Vieira et al., 2013). Floods or dry conditions in southeastern Brazil are often related to SACZ variability (Muza et al., 2009; Lima et al., 2010; Vasconcellos and Cavalcanti, 2010). A subset of CMIP models simulate the SACZ (Vera and Silvestri, 2009; Seth et al., 2010; Yin et al., 2012) and its variability as a dipolar structure (Junquas et al., 2011; Cavalcanti and Shimizu, 2012).

A southward displacement of SACZ and intensification of the southern centre of the precipitation dipole are suggested in projections of CMIP3 and CMIP5 models (Seth et al., 2010; Junquas et al., 2011; Cavalcanti and Shimizu, 2012). This displacement is consistent with the increased precipitation over Southeastern South America, south of 25°S, projected for the second half of the 21st century, in CMIP3, CMIP5 and regional models (Figure AI.34, Figure 14.21). It is also consistent with the southward displacement of the Atlantic subtropical high (Seth et al., 2010) related to the southward expansion of the Hadley cell (Lu et al., 2007). Pacific SST warming and the strengthening of the PSA-like wave train (Section 14.6.2) are potential mechanisms for changes in the dipolar pattern resulting in SACZ change (Junquas et al., 2011). This change is also supported by the intensification and increased frequency of the low level jet over South America in future projections (Soares and Marengo, 2009; Seth et al., 2010).

14.3.2 Madden Julian Oscillation

The Madden-Julian Oscillation (MJO) is the dominant mode of tropical intraseasonal (20–100 days) variability (Zhang, 2005). The MJO modulates tropical cyclone activity (Frank and Roundy, 2006), contributes to intraseasonal fluctuations of the monsoons (Maloney and Shaman, 2008), and excites teleconnection patterns outside the tropics (L'Heureux and Higgins, 2008; Lin et al., 2009). Simulation of the MJO by GCMs remains challenging, but with some improvements made in recent years (Section 9.5.2.3).

Possible changes in the MJO in a future warmer climate have just begun to be explored with models that simulate the phenomenon. In the Max Planck Institute Earth System Model, MJO variance increases appreciably with increasing warming (Schubert et al., 2013). The change in MJO variance is highly sensitive to the spatial pattern of SST warming (Maloney and Xie, 2013). In light of the low skill in simulating MJO, and its sensitive to SST warming pattern, which in itself is subject to large uncertainties, confidence is *low* in assessing how MJO will change in a warmer climate.

14.3.3 Indian Ocean Modes

The tropical Indian Ocean SST exhibits two modes of interannual variability (Schott et al., 2009; Deser et al., 2010b): the Indian Ocean Basin (IOB) mode featuring a basin-wide structure of the same sign, and the Indian Ocean dipole (IOD) mode with largest amplitude in the eastern Indian Ocean off Indonesia, and weaker anomalies of the opposite polarity over the rest of the basin (Box 2.5). Both modes are statistically significantly correlated with ENSO (Section 14.4). CMIP models simulate both modes well (Section 9.5.3.4.2).

The formation of IOB is linked to ENSO via an atmospheric bridge and surface heat flux adjustment (Klein et al., 1999; Alexander et al., 2002). Ocean-atmosphere interactions within the Indian Ocean are important for the long persistence of this mode (Izumo et al., 2008; Wu et al., 2008; Du et al., 2009). The basin mode affects the termination of ENSO events (Kug and Kang, 2006), it induces coherent atmospheric anomalies in the summer following El Nino (Xie et al., 2009), including supressed convection (Wang et al., 2003) and reduced tropical cyclone activity (Du et al., 2011) over the Northwest Pacific, and anomalous rainfall over

East Asia (Huang et al., 2004).

IOD develops in July-November and involves Bjerknes feedback between zonal SST gradient, zonal wind and thermocline tilt along the equator (Saji et al., 1999; Webster et al., 1999). A positive IOD event (with negative SST anomalies off Sumatra) is associated with droughts in Indonesia, reduced rainfall over Australia, intensified Indian summer monsoon, increased precipitation in East Africa, and anomalous conditions in the extratropical Southern Hemisphere (Yamagata et al., 2004). Most CMIP3 models are able to reproduce the general features of the IOD, including its phase lock onto the July-November season, while detailed analysis of CMIP5 simulations are not yet available (Section 9.5.3.4.2)

Basin-mean SST has risen steadily for much of the 20th Century, a trend captured by CMIP3 20th century simulations (Alory et al., 2007). The SST increase over the North Indian Ocean since about 1930 is noticeably weaker than for the rest of the basin. This spatial pattern is suggestive of the effects of reduced surface solar radiation due to Asian brown clouds (Chung and Ramanathan, 2006) and it affects Arabian Sea cyclones (Evan et al., 2011b). In the equatorial Indian Ocean, coral isotope records off Indonesia indicate a reduced SST warming and/or increased salinity during the 20th century (Abram et al., 2008). An easterly wind change especially during July-October has been observed over the past six decades, a result consistent with a reduction of marine cloudiness in the east and a decreasing precipitation trend over the maritime continent (Tokinaga et al., 2012). Atmospheric reanalysis products have difficulty representing these changes (Han et al., 2010).

The projected changes over the equatorial Indian Ocean include easterly wind anomalies, a shoaling thermocline (Vecchi and Soden, 2007a; Du and Xie, 2008) and reduced SST warming in the east (Stowasser et al., 2009), a result confirmed by CMIP5 multi-model analysis (Zheng et al., 2013; Figure 14.10). The change in zonal SST gradient, in turn, reinforces the easterly wind change, indicative of a positive feedback between them as envisioned by Bjerknes (1969). This coupled pattern is most pronounced during July-November, and is broadly consistent with the observed changes in the equatorial Indian Ocean.

[INSERT FIGURE 14.10 HERE]
Figure 14.10: September to November changes in a 22-model CMIP5 ensemble (2081–2100 in RCP8.5 minus 1986–2005 in historical run). (a) SST (colour contours at 0.1°C intervals) relative to the tropical mean (20°S–20°N), and precipitation (shading and white contours at 20 mm per month intervals). (b) Surface wind velocity (m s^{-1}), and sea surface height deviation from the global mean (contours, cm). Over the equatorial Indian Ocean, ocean-atmospheric changes form Bjerknes feedback, with the reduced SST warming and suppressed convection in the east. Updated with CMIP5 from Xie et al. (2010b).

In one CMIP3 model, the IOB mode and its capacitor effect persist longer, through summer into early fall towards the end of the century (2081–2100, Zheng et al., 2011). This increased persistence intensifies ENSO's influence on the Northwest Pacific summer monsoon. The confidence level of this relationship is *low* due to the lack of multi-model studies.

The IOD variability in SST remains nearly unchanged in CMIP future projections (Ihara et al., 2009; Figure 14.11a) despite the easterly wind change that lifts the thermocline (Figure 14.10b) and intensifies thermocline feedback on SST in the eastern equatorial Indian Ocean. The global increase in atmospheric dry static stability weakens the atmospheric response to zonal SST gradient changes, countering the enhanced thermocline feedback (Zheng et al., 2010). The weakened atmospheric feedback is reflected in a decrease in IOD variance in both zonal wind and the thermocline depth (Zheng et al., 2013; Figure 14.11b-c).

[INSERT FIGURE 14.11 HERE]
Figure 14.11: CMIP5 multi-model ensemble mean standard deviations of interannual variability for September to November in pre-industrial (PiControl; blue bars) and RCP8.5 (red) runs: (a) the Indian Ocean dipole index defined as the western (50°E–70°E, 10°S–10°N) minus eastern (90°E–110°E, 10°S–0°) SST difference; (b) zonal wind in the central equatorial Indian Ocean (70°E–90°E, 5°S–5°N); and (c) sea surface height in the eastern equatorial Indian Ocean (90°E–110°E, 10°S–0°). The standard deviation is normalized by the pre-industrial (PiControl) value for each model before ensemble average. Blue box-whisker plots show the 10th, 25th, 50th, 75th, and 90th percentiles of 51-year windows for PiControl, representing natural variability. Red box-whisker plots represent inter-model variability for RCP8.5 (red), based on the nearest rank. Adapted from Zheng et al. (2013).

14.3.4 Atlantic Ocean Modes

The Atlantic features a northward-displaced ITCZ (Section 14.3.1.1), and a cold tongue that develops in boreal summer. Climate models generally fail to simulate these characteristics of tropical Atlantic climate (Section 9.5.3.3). The biases severely limit model skill in simulating modes of Atlantic climate variability and in projecting future climate change in the Atlantic sector. In-depth analysis of the CMIP5 projections of Atlantic Ocean Modes has not yet been fully explored, but see Section 12.4.3.

The inter-hemispheric SST gradient displays pronounced interannual to decadal variability (Box 2.5, Figure 2), referred to as the Atlantic meridional mode (AMM; Servain et al., 1999; Chiang and Vimont, 2004; Xie and Carton, 2004). A thermodynamic feedback between surface winds, evaporation, and SST (WES; Xie and Philander, 1994) is fundamental to the AMM (Chang et al., 2006). This mode affects precipitation in northeastern Brazil by displacing the ITCZ (Servain et al., 1999; Chiang and Vimont, 2004; Xie and Carton, 2004), and Atlantic hurricane activity (Vimont and Kossin, 2007; Smirnov and Vimont, 2011).

The Atlantic Niño mode represents interannual variability in the equatorial cold tongue, akin to ENSO (Box 2.5, Figure 2). Bjerknes feedback is considered important for energizing the mode (Zebiak, 1993; Carton and Huang, 1994; Keenlyside and Latif, 2007). This mode affects the West Africa Monsoon (Vizy and Cook, 2002; Giannini et al., 2003).

Over the past century, the Atlantic has experienced a pronounced and persistent warming trend. The warming has brought detectable changes in atmospheric circulation and rainfall patterns in the region. In particular, the ITCZ has shifted southward and land precipitation has increased over the equatorial Amazon, equatorial West Africa, and along the Guinea coast, while it has decreased over the Sahel (Deser et al., 2010a; Tokinaga and Xie, 2011; see also Sections 2.5 and 2.7). Atlantic Niño variability has weakened by 40% in amplitude from 1960 to 1999, associated with a weakening of the equatorial cold tongue (Tokinaga and Xie, 2011).

The CMIP3 20th century climate simulations generally capture the warming trend of the basin-averaged SST over the tropical Atlantic. A majority of the models also seem to capture the secular trend in the tropical Atlantic SST inter-hemispheric gradient and, as a result, the southward shift of the Atlantic ITCZ over the past century (Chang et al., 2011).

Many CMIP3 model simulations with the A1B emission scenario show only minor changes in the SST variance associated with the AMM. However, the few models that give the best AMM simulation over the 20^{th} Century project a weakening in future AMM activity (Breugem et al., 2006), possibly due to the northward shift of the ITCZ (Breugem et al., 2007). At present, model projections of future change in AMM activity is considered highly uncertain because of the poorly simulated Atlantic ITCZ. In fact, uncertainty in projected changes in Atlantic meridional SST gradient limits the confidence in regional climate projections surrounding the tropical Atlantic Ocean (Good et al., 2008).

A majority of CMIP3 models forced with the A1B emission scenario project no major change in Atlantic Niño activity in the 21st Century, while a few models project a sizable decrease in future activity (Breugem et al., 2006).

CMIP5 projections show an accelerated SST warming over much of the tropical Atlantic (Figure 12.11). RCP8.5 projections of the inter-hemispheric SST gradient change within the basin, however, are not consistent among CMIP5 models as future GHG increase dominates over the anthropogenic aerosol effect.

14.3.5 Assessment Summary

There is *medium confidence* that annual rainfall changes over tropical oceans follow a ‘warmer-get-wetter’ pattern, increasing where the SST warming exceeds the tropical mean and vice versa. One third of inter-model differences in precipitation projection are due to those in SST pattern. The SST pattern effect on precipitation change is a new finding since AR4.

The wet-get-wetter effect is more obvious in the seasonal than annual rainfall change in the tropics.

Confidence is generally higher in seasonal than in annual mean changes in tropical precipitation. There is *medium confidence* that seasonal rainfall will increase on the equatorward flank of the current ITCZ; that the frequency of zonally-oriented SPCZ events will increase, with the SPCZ lying well to the northeast of its average position during those events; and that the SACZ shifts southwards, in conjunction with the southward displacement of the South Atlantic subtropical high, leading to an increase in precipitation over southeastern South America.

Due to models' ability to reproduce general features of IOD and agreement on future projections, there is *high confidence* that the tropical Indian Ocean is *likely* to feature a zonal pattern with reduced (enhanced) warming and decreased (increased) rainfall in the east (west), a pattern especially pronounced during August-November. The Indian Ocean dipole mode will *very likely* remain active, with interannual variability unchanged in SST but decreasing in thermocline depth. There is low confidence in changes in the summer persistence of the Indian Ocean SST response to ENSO and in ENSO's influence on summer climate over the Northwest Pacific and East Asia.

The observed SST warming in the tropical Atlantic represents a reduction in spatial variation in climatology: the warming is weaker north than south of the equator; and the equatorial cold tongue weakens both in the mean and interannual variability. There is *low confidence* in projected changes over the tropical Atlantic - both for the mean and interannual modes, because of large errors in model simulations of current climate.

There is *low confidence* in how MJO will change in the future due to the poor skill of models in simulating MJO and the sensitivity of its change to SST warming patterns that are themselves subject to large uncertainties in the projections.

14.4 El Niño-Southern Oscillation

The El Niño-Southern Oscillation (ENSO) is a coupled ocean-atmosphere phenomenon naturally occurring at the inter-annual time scale over the tropical Pacific (See Box 2.5, Supplementary Material Section 14.SM.2, and Figure 14.12).

[INSERT FIGURE 14.12 HERE]

Figure 14.12: Idealized schematic showing atmospheric and oceanic conditions of the tropical Pacific region and their interactions during normal conditions, El Niño conditions, and in a warmer world. (a) Mean climate conditions in the tropical Pacific, indicating SSTs, surface wind stress and associated Walker circulation, the mean position of convection, and the mean upwelling and position of the thermocline. (b) Typical conditions during an El Niño event. SSTs are anomalously warm in the east; convection moves into the central Pacific; the trade winds weaken in the east and the Walker circulation is disrupted; the thermocline flattens and the upwelling is reduced. (c) The likely mean conditions under climate change derived from observations, theory and coupled GCMs. The trade winds weaken; the thermocline flattens and shoals; the upwelling is reduced although the mean vertical temperature gradient is increased; and SSTs (shown as anomalies with respect to the mean tropical-wide warming) increase more on the equator than off. Diagrams with absolute SST fields are shown on the left, diagrams with SST anomalies are shown on the right. For the climate change fields, anomalies are expressed with respect to the basin average temperature change so that blue colours indicate a warming smaller than the basin mean, not a cooling (Collins et al., 2010).

14.4.1 Tropical Pacific Mean State

SST in the western tropical Pacific has increased by up to 1.5°C per century, and the warm pool has expanded (Liu and Huang, 2000; Huang and Liu, 2001; Cravatte et al., 2009). Studies disagree on how the east-west SST gradient along the equator has changed, some showing a strengthening (Cane et al., 1997; Hansen et al., 2006; Karnauskas et al., 2009; An et al., 2011) and others showing a weakening (Deser et al., 2010a; Tokinaga et al., 2012), because of observational uncertainties associated with limited data sampling, changing measurement techniques, and analysis procedures. Most CMIP3 and CMIP5 models also disagree on the response of zonal SST gradient across the equatorial Pacific (Yeh et al., 2012).

The Pacific Ocean warms more near the equator than in the subtropics in CMIP3 and CMIP5 projections (Liu et al., 2005; Gastineau and Soden, 2009; Widlansky et al., 2013; Figure 14.12) because of the difference in evaporative damping (Xie et al., 2010b). Other oceanic changes include a basin-wide thermocline shoaling (Vecchi and Soden, 2007a; DiNezio et al., 2009; Collins et al., 2010; Figure 14.12), a weakening of surface currents, and a slight upward shift and strengthening of the equatorial undercurrent (Luo and Rothstein, 2011; Sen Gupta et al., 2012). A weakening of tropical atmosphere circulation during the twentieth century was documented in observations and reanalyses (Vecchi et al., 2006; Zhang and Song, 2006; Vecchi and Soden, 2007a; Bunge and Clarke, 2009; Karnauskas et al., 2009; Yu and Zwiers, 2010; Tokinaga et al., 2012) and in CMIP models (Vecchi and Soden, 2007a; Gastineau and Soden, 2009). The Pacific Walker circulation, however, intensified during the most recent two decades (Mitas and Clement, 2005; Liu and Curry, 2006; Mitas and Clement, 2006; Sohn and Park, 2010; Li and Ren, 2011; Zahn and Allan, 2011; Zhang et al., 2011a), illustrating the effects of natural variability.

14.4.2 El Niño Changes over Recent Decades and in the Future

The amplitude modulation of ENSO at longer timescales has been observed in reconstructed instrumental records (Gu and Philander, 1995; Wang, 1995; Mitchell and Wallace, 1996; Wang and Wang, 1996; Power et al., 1999; An and Wang, 2000; Yeh and Kirtman, 2005; Power and Smith, 2007; Section 5.4.1), in proxy records (Cobb et al., 2003; Braganza et al., 2009; Li et al., 2011c; Yan et al., 2011), and is also simulated by coupled GCMs (Lau et al., 2008; Wittenberg, 2009). Some studies have suggested that the modulation was due to changes in mean climate conditions in the tropical Pacific (An and Wang, 2000; Fedorov and Philander, 2000; Wang and An, 2001, 2002; Li et al., 2011c), as observed since the 1980s (An and Jin, 2000; An and Wang, 2000; Fedorov and Philander, 2000; Kim and An, 2011). Since the late 1990s, the maximum SST warming during El Niño has been frequently observed in the central Pacific (Figure 14.13; Ashok et al., 2007; Kao and Yu, 2009; Kug et al., 2009; Section 9.5.3.4.1 and Supplementary Material Section 14.SM.2; Yeh et al., 2009), with global impacts that are distinct from 'standard' El Niño events where the maximum warming is over the eastern Pacific (Kumar et al., 2006a; Ashok et al., 2007; Kao and Yu, 2009; Hu et al., 2012b). During the past century, an increasing trend in ENSO amplitude was also observed (Li et al., 2011c; Vance et al., 2012), possibly caused by a warming climate (Zhang et al., 2008; Kim and An, 2011) although other reconstructions in this data-sparse region dispute this trend (Giese and Ray, 2011).

[INSERT FIGURE 14.13 HERE]
Figure 14.13: Intensities of El Niño and La Niña events for the last 60 years in the eastern equatorial Pacific (Niño3 region) and in the central equatorial Pacific (Niño4 region), and the estimated linear trends, obtained from ERSSTv3.

Long coupled GCM simulations show that decadal-to-centennial modulations of ENSO can be generated without any change in external forcing (Wittenberg, 2009; Yeh et al., 2011), with epochs of extreme ENSO behaviour lasting decades or even centuries. The modulations result from nonlinear processes in the tropical climate system (Timmermann et al., 2003), the interaction with the mean climate state (Ye and Hsieh, 2008; Choi et al., 2009; Choi et al., 2011; Choi et al., 2012), or from random changes in ENSO activity triggered by chaotic atmospheric variability (Power and Colman, 2006; Power et al., 2006). There is little consensus as to whether the decadal modulations of ENSO properties (amplitude and spatial pattern) during recent decades are due to anthropogenic effects or natural variability. Instrumental SST records are available back to the 1850s, but good observations of the coupled air-sea feedbacks that control ENSO behaviour – including subsurface temperature and current fluctuations, and air-sea exchanges of heat, momentum, and water – are available only after the late 1970s, making observed historical variations in ENSO feedbacks highly uncertain (Chen, 2003; Wittenberg, 2004).

[INSERT FIGURE 14.14 HERE]
Figure 14.14: Standard deviation of Niño3 SST anomalies from CMIP5 model experiments. PI, 20C, RCP4.5, RCP8.5 indicate pre-industrial control experiments, 20th century experiments, and 21st century experiment from the RCP4.5 and RCP8.5. Open dot and solid black line indicate multi-model ensemble mean and median, respectively, and the cross mark is 20th observation, respectively. Thick bar and thin outer bar indicate 50% and 75% percentile ranges, respectively.

CMIP5 models show some improvement compared to CMIP3, especially in ENSO amplitude (Section 9.5.3.4.1). Selected CMIP5 models that simulate well strong El Niño events show a gradual increase of El Niño intensity, especially over the central Pacific (Kim and Yu, 2012). CMIP3 models suggested a westward shift of SST variability in future projections (Boer, 2009; Yeh et al., 2009). Generally, however, future changes in El Niño intensity in CMIP5 models are model-dependent (Guilyardi et al., 2012; Kim and Yu, 2012; Stevenson et al., 2012), and not significantly distinguished from natural modulations (Stevenson, 2012; Figure 14.14). Because the change in tropical mean conditions (especially the zonal gradient) in a warming climate is model dependent (Section 14.4.1), changes in ENSO intensity for the 21st century (Solomon and Newman, 2011; Hu et al., 2012a) are uncertain (Figure 14.14). Future changes in ENSO depend on competing changes in coupled ocean-atmospheric feedback (Philip and Van Oldenborgh, 2006; Collins et al., 2010; Vecchi and Wittenberg, 2010), and on the dynamical regime a given model is in. It is *very likely*, however, that ENSO will remain the dominant mode of natural climate variability in the 21st century (Collins et al., 2010; Figure 14.14). Due to enhanced moisture availability (Section 12.4.4) ENSO-induced rainfall variability on regional scales is therefore also *likely* to intensify.

14.4.3 Teleconnections

There is little improvement in the CMIP5 ensemble relative to CMIP3 in the amplitude and spatial correlation metrics of precipitation teleconnections in response to ENSO, in particular within regions of strong observed precipitation teleconnections (equatorial South America, the western equatorial Pacific and a southern section of North America; Langenbrunner and Neelin, 2013). Scenario projections in CMIP3 and CMIP5 showed a systematic eastward shift in both El Niño- and La Niña-induced teleconnection patterns over the extratropical Northern Hemisphere (Meehl and Teng, 2007; Stevenson et al., 2012), which might be due to the eastward migration of tropical convection centres associated with the expansion of the warm pool in a warm climate (Muller and Roeckner, 2006; Müller and Roeckner, 2008; Cravatte et al., 2009; Kug et al., 2010), or changes in the mid-latitude mean circulation (Meehl and Teng, 2007). Some models produced an intensified ENSO teleconnection pattern over the North Atlantic region in a warmer climate (Müller and Roeckner, 2008; Bulic et al., 2012) and a weakened teleconnection pattern over the North Pacific (Stevenson, 2012). It is unclear whether the eastward shift of tropical convection is related to longitudinal shifts in El Niño maximum SST anomalies (see Supplementary Material Section 14.SM.2) or to changes in the mean state in the tropical Pacific. Some coupled GCMs, which do not show an increase in the central Pacific warming during El Nino in response to a warming climate, do not produce a substantial change in the longitudinal location of tropical convection (Müller and Roeckner, 2008; Yeh et al., 2009).

[INSERT FIGURE 14.15 HERE]
Figure 14.15: Changes to sea level pressure teleconnections during DJF in the CMIP5 models. (a) SLP anomalies for El Niño during the 20th century. (b) SLP anomalies for La Nina during the 20th century. (c) SLP anomalies for El Niño during RCP4.5. (d) SLP anomalies for La Niña during RCP4.5. Maps in (a)-(d) are stippled where more than 2/3 of models agree on the sign of the SLP anomaly ((a),(b): 18 models; (c),(d): 12 models), and hatched where differences between the RCP4.5 multi-model mean SLP anomaly exceed the 60th percentile (red-bordered regions) or are less than the 40th percentile (blue-bordered regions) of the distribution of 20th century ensemble means. In all panels, El Niño (La Niña) periods are defined as years having DJF NINO3.4 SST above (below) one standard deviation relative to the mean of the detrended time series. For ensemble-mean calculations, all SLP anomalies have been normalized to the standard deviation of the ensemble-member detrended NINO3.4 SST. (e) Change in the "centre of mass" of the Aleutian Low SLP anomaly, RCP4.5 - 20th century. The Aleutian Low SLP centre of mass is a vector with two elements (lat, lon), and is defined as the sum of (lat, lon) weighted by the SLP anomaly, over all points in the region 180°E–120°E, 40°N–60°N having a negative SLP anomaly during El Niño.

14.4.4 Assessment summary

ENSO shows considerable inter-decadal modulations in amplitude and spatial pattern within the instrumental record. Models without changes in external forcing display similar modulations, and there is little consensus on whether the observed changes in ENSO are due to external forcing or natural variability (see also Section 10.3.3 for an attribution discussion).

It is *very likely* that ENSO will remain the dominant mode of interannual variability with global influences in the 21st century, and due to changes in moisture availability ENSO-induced rainfall variability on regional scales will intensify. There is *medium confidence* that ENSO-induced teleconnection patterns will shift eastward over the North Pacific and North America. There is *low confidence* in changes in the intensity and spatial pattern of El Niño in a warmer climate.

14.5 Annular and Dipolar Modes

The North Atlantic Oscillation (NAO), the North Pacific Oscillation (NPO), and the Northern and Southern Annular Modes (NAM and SAM) are dominant modes of variability in the extra-tropics. These modes are