A number of components or phenomena within the climate system have been proposed as potentially exhibiting threshold behaviour. Crossing such thresholds can lead to an abrupt or irreversible transition into a different state of the climate system or some of its components.

*Abrupt climate change* is defined in this IPCC Fifth Assessment Report (AR5) as a large-scale change in the climate system that takes place over a few decades or less, persists (or is anticipated to persist) for at least a few decades and causes substantial disruptions in human and natural systems. There is information on potential consequences of some abrupt changes, but in general there is *low confidence* and little consensus on the likelihood of such events over the 21st century. Examples of components susceptible to such abrupt change are the strength of the Atlantic Meridional Overturning Circulation (AMOC), clathrate methane release, tropical and boreal forest dieback, disappearance of summer sea ice in the Arctic Ocean, long-term drought and monsoonal circulation. (5.7, 6.4.7, 12.5.5; Table 12.4)

A change is said to be *irreversible* if the recovery time scale from this state due to natural processes is significantly longer than the time it takes for the system to reach this perturbed state. Such behaviour may arise because the time scales for perturbations and recovery processes are different, or because climate change may persist due to the long residence time of a carbon dioxide (CO$_2$) perturbation in the atmosphere (see TFE.8). Whereas changes in Arctic Ocean summer sea ice extent, long-term droughts and monsoonal circulation are assessed to be reversible within years to decades, tropical or boreal forest dieback may be reversible only within centuries. Changes in clathrate methane and permafrost carbon release, Greenland and Antarctic ice sheet collapse may be irreversible during millennia after the causal perturbation. (5.8, 6.4.7, 12.5.5, 13.4.3, 13.4.4; Table 12.4)

**Abrupt Climate Change Linked with AMOC**

New transient climate model simulations have confirmed with *high confidence* that strong changes in the strength of the AMOC produce abrupt climate changes at global scale with magnitude and pattern resembling past glacial Dansgaard–Oeschger events and Heinrich stadials. Confidence in the link between changes in North Atlantic climate and low-latitude precipitation has increased since the IPCC Fourth Assessment Report (AR4). From new paleoclimate reconstructions and modelling studies, there is *very high confidence* that a reduced strength of the AMOC and the associated surface cooling in the North Atlantic region caused southward shifts of the Atlantic Intertropical Convergence Zone and affected the American (north and south), African and Asian monsoons. (5.7)

The interglacial mode of the AMOC can recover (*high confidence*) from a short-lived freshwater input into the subpolar North Atlantic. Approximately 8.2 ka, a sudden freshwater release occurred during the final stages of North America ice sheet melting. Paleoclimate observations and model results indicate, with *high confidence*, a marked reduction in the strength of the AMOC followed by a rapid recovery, within approximately 200 years after the perturbation. (5.8.2)

Although many more model simulations have been conducted since AR4 under a wide range of future forcing scenarios, projections of the AMOC behaviour have not changed. It remains *very likely* that the AMOC will weaken over the 21st century relative to 1850-1900 values. Best estimates and ranges for the reduction from the Coupled Model Intercomparison Project Phase 5 (CMIP5) are 11% (1 to 24%) for the Representative Concentration Pathway RCP2.6 and 34% (12 to 54%) for RCP8.5, but there is *low confidence* on the magnitude of weakening. It also remains *very unlikely* that the AMOC will undergo an abrupt transition or collapse in the 21st century for the scenarios considered (*high confidence*) (TFE.5, Figure 1). For an abrupt transition of the AMOC to occur, the sensitivity of the AMOC to forcing would have to be far greater than seen in current models, or would require meltwater flux from the Greenland ice sheet greatly exceeding even the highest of current projections. Although neither possibility can be excluded entirely, it is *unlikely* that the AMOC will collapse beyond the end of the 21st century for the scenarios considered, but a collapse beyond the 21st century for large sustained warming cannot be excluded. There is *low confidence* in assessing the evolution of AMOC beyond the 21st century because of limited number of analyses and equivocal results. (12.4.7, 12.5.5)

**Potential Irreversibility of Changes in Permafrost, Methane Clathrates and Forests**

In a warming climate, permafrost thawing may induce decomposition of carbon accumulated in frozen soils which could persist for hundreds to thousands of years, leading to an increase of atmospheric CO$_2$ and/or methane (CH$_4$)

(continued on next page)
concentrations. The existing modelling studies of permafrost carbon balance under future warming that take into account at least some of the essential permafrost-related processes do not yield consistent results, beyond the fact that present-day permafrost will become a net emitter of carbon during the 21st century under plausible future warming scenarios (low confidence). This also reflects an insufficient understanding of the relevant soil processes during and after permafrost thaw, including processes leading to stabilization of unfrozen soil carbon, and precludes any quantitative assessment of the amplitude of irreversible changes in the climate system potentially related to permafrost degassing and associated feedbacks. {6.4.7, 12.5.5}

Anthropogenic warming will very likely lead to enhanced CH$_4$ emissions from both terrestrial and oceanic clathrates. Deposits of CH$_4$ clathrates below the sea floor are susceptible to destabilization via ocean warming. However, sea level rise due to changes in ocean mass enhances clathrate stability in the ocean. While difficult to formally assess, initial estimates of the 21st century feedback from CH$_4$ clathrate destabilization are small but not insignificant. It is very unlikely that CH$_4$ from clathrates will undergo catastrophic release during the 21st century (high confidence). On multi-millennial time scales, such CH$_4$ emissions may provide a positive feedback to anthropogenic warming and may be irreversible, due to the difference between release and accumulation time scales. (6.4.7, 12.5.5)

The existence of critical climate change driven dieback thresholds in the Amazonian and other tropical rainforests purely driven by climate change remains highly uncertain. The possibility of a critical threshold being crossed in precipitation volume and duration of dry seasons cannot be ruled out. The response of boreal forest to projected climate change is also highly uncertain, and the existence of critical thresholds cannot at present be ruled out. There is low confidence in projections of the collapse of large areas of tropical and/or boreal forests. (12.5.5)

**Potential Irreversibility of Changes in the Cryosphere**

The reversibility of sea ice loss has been directly assessed in sensitivity studies to CO$_2$ increase and decrease with Atmosphere–Ocean General Circulation Models (AOGCMs) or Earth System Models (ESMs). None of them show evidence of an irreversible change in Arctic sea ice at any point. By contrast, as a result of the strong coupling between surface and deep waters in the Southern Ocean, the Antarctic sea ice in some models integrated with ramp-up and ramp-down atmospheric CO$_2$ concentration exhibits some hysteresis behaviour. (12.5.5)

At present, both the Greenland and Antarctic ice sheets have a positive surface mass balance (snowfall exceeds melting), although both are losing mass because ice outflow into the sea exceeds the net surface mass balance. A positive feedback operates to reduce ice sheet volume and extent when a decrease of the surface elevation of the ice sheet induces a decreased surface mass balance. This arises generally through increased surface melting, and therefore applies in the 21st century to Greenland, but not to Antarctica, where surface melting is currently very small. Surface melting in Antarctica is projected to become important after several centuries under high well-mixed greenhouse gas radiative forcing scenarios. (4.4, 13.4.4; Boxes 5.2, 13.2)

Abrupt change in ice sheet outflow to the sea may be caused by unstable retreat of the grounding line in regions where the bedrock is below sea level and slopes downwards towards the interior of the ice sheet. This mainly
Technical Summary

applys to West Antarctica, but also to parts of East Antarctica and Greenland. Grounding line retreat can be triggered by ice shelf decay, due to warmer ocean water under ice shelves enhancing submarine ice shelf melt, or melt water ponds on the surface of the ice shelf promoting ice shelf fracture. Because ice sheet growth is a slow process, such changes would be irreversible in the definition adopted here. (4.4.5; Box 13.2)

There is high confidence that the volumes of the Greenland and West Antarctic ice sheets were reduced during periods of the past few million years that were globally warmer than present. Ice sheet model simulations and geological data suggest that the West Antarctic ice sheet is very sensitive to subsurface ocean warming and imply with medium confidence a West Antarctic ice sheet retreat if atmospheric CO$_2$ concentration stays within, or above, the range of 350–450 ppm for several millennia. (5.8.1, 13.4.4; Box 13.2)

The available evidence indicates that global warming beyond a threshold would lead to the near-complete loss of the Greenland ice sheet over a millennium or longer, causing a global mean sea level rise of approximately 7 m. Studies with fixed present-day ice sheet topography indicate that the threshold is greater than 2°C but less than 4°C (medium confidence) of global mean surface temperature rise above pre-industrial. The one study with a dynamical ice sheet suggests the threshold is greater than about 1°C (low confidence) global mean warming with respect to pre-industrial. Considering the present state of scientific uncertainty, a likely range cannot be quantified. The complete loss of the Greenland ice sheet is not inevitable because this would take a millennium or more; if temperatures decline before the ice sheet has completely vanished, the ice sheet might regrow. However, some part of the mass loss might be irreversible, depending on the duration and degree of exceedance of the threshold, because the ice sheet may have multiple steady states, due to its interaction with regional climate. (13.4.3, 13.4.4)

**TS.4.6 Water Cycle**

Since the AR4, new evidence has emerged of a detectable human influence on several aspects of the water cycle. There is medium confidence that observed changes in near-surface specific humidity since 1973 contain a detectable anthropogenic component. The anthropogenic water vapour fingerprint simulated by an ensemble of climate models has been detected in lower tropospheric moisture content estimates derived from Special Sensor Microwave Imager (SSM/I) data covering the period 1988–2006. An anthropogenic contribution to increases in tropospheric specific humidity is found with medium confidence. (2.5, 10.3)

Attribution studies of global zonal mean terrestrial precipitation and Arctic precipitation both find a detectable anthropogenic influence. Overall there is medium confidence in a significant human influence on global scale changes in precipitation patterns, including increases in NH mid-to-high latitudes. Remaining observational and modelling uncertainties and the large effect of internal variability on observed precipitation preclude a more confident assessment. (2.5, 7.6, 10.3)

Based on the collected evidence for attributable changes (with varying levels of confidence and likelihood) in specific humidity, terrestrial precipitation and ocean surface salinity through its connection to precipitation and evaporation, and from physical understanding of the water cycle, it is likely that human influence has affected the global water cycle since 1960. This is a major advance since AR4. (2.4, 2.5, 3.3, 9.4.1, 10.3, 10.4.2; Table 10.1; FAQ 3.2)

**TS.4.7 Climate Extremes**

Several new attribution studies have found a detectable anthropogenic influence in the observed increased frequency of warm days and nights and decreased frequency of cold days and nights. Since the AR4 and SREX, there is new evidence for detection of human influence on extremely warm daytime temperature and there is new evidence that the influence of anthropogenic forcing may be detected separately from the influence of natural forcing at global scales and in some continental and sub-continental regions. This strengthens the conclusions from both AR4 and SREX, and it is now very likely that anthropogenic forcing has contributed to the observed changes in the frequency and intensity of daily temperature extremes on the global scale since the mid-20th century. It is likely that human influence has significantly increased the probability of occurrence of heat waves in some locations. See TFE.9 and TFE.9, Table 1 for a summary of the assessment of extreme weather and climate events. (10.6)

Since the AR4, there is some new limited direct evidence for an anthropogenic influence on extreme precipitation, including a formal detection and attribution study and indirect evidence that extreme precipitation would be expected to have increased given the evidence of anthropogenic influence on various aspects of the global hydrological cycle and high confidence that the intensity of extreme precipitation events will increase with warming, at a rate well exceeding that of the mean precipitation. In land regions where observational coverage is sufficient for assessment, there is medium confidence that anthropogenic forcing has contributed to a global-scale intensification of heavy precipitation over the second half of the 20th century. (7.6, 10.6)
this acceleration has been reported (Gehrels et al., 2006, 2008, 2011; Kemp et al., 2009, 2011), but Gehrels and Woodworth (2013) have concluded that these mismatches can be reconciled within the observational uncertainties. Combined with the instrumental evidence (see Section 3.7) and with inferences drawn from archaeological evidence from 2000 years ago (Lambeck et al., 2004b), rates of sea level rise exceeded the late Holocene background rate after about 1900 (high confidence) (Figure 5.17).

Regionally, as along the US Atlantic coast and Gulf of Mexico coast, the salt-marsh records reveal some consistency in multi-decadal and centennial time scales deviations from the linear trends expected from the GIA signal (see e.g., panels (a) and (b) in Figure 5.17) (van de Plassche et al., 1998; González and Törnqvist, 2009; Kemp et al., 2011) but they have not yet been identified as truly global phenomena. For the past 5 millennia the most complete sea level record from a single location consists of microatoll evidence from Kirritimati (Christmas Island; Pacific Ocean) (Woodroffe et al., 2012) that reveals with medium confidence that amplitudes of any fluctuations in GMSL during this interval did not exceed approximately ±25 cm on time scales of a few hundred years. Proxy data from other localities with quasi-continuous records for parts of this pre-industrial period, likewise, do not identify significant global oscillations on centennial time scales (Figure 5.17).

5.7 Evidence and Processes of Abrupt Climate Change

Many paleoclimate archives document climate changes that happened at rates considerably exceeding the average rate of change for longer-term averaging periods prior and after this change (see Glossary for other definition of Abrupt Climate Change). A variety of mechanisms have been suggested to explain the emergence of such abrupt climate changes (see Section 12.5.5). Most of them invoke the existence of nonlinearities or, more specifically, thresholds in the underlying dynamics of one or more Earth-system components. Both internal dynamics and external forcings can generate abrupt changes in the climate state. Documentation of abrupt climate changes in the past using multiple sources of proxy evidence can provide important benchmarks to test instability mechanisms in climate models. This assessment of abrupt climate change on time scales of 10 to 100 years focuses on Dansgaard-Oeschger (DO) events and iceberg/meltwater discharges during Heinrich events, especially the advances since AR4 in reconstructing and understanding their global impacts and in extending the record of millennial-scale variability to about 800 ka.

Twenty-five abrupt DO events (North Greenland Ice Core Project members, 2004) and several centennial-scale events (Capron et al., 2010b) occurred during the last glacial cycle (see Section 5.3.2). DO events in Greenland were marked by an abrupt transition (within a few decades) from a cold phase, referred to as Greenland Stadial (GS) into a warm phase, known as Greenland Interstadial (GI). Subsequently but within a GI, a gradual cooling preceded a rapid jump to GS that lasted for centuries up to millennia. Thermal gas-fractionation methods (Landais et al., 2004; Huber et al., 2006) suggest that for certain DO events Greenland temperatures increased by up to 16°C ± 2.5°C (1 standard deviation) within several decades. Such transitions were also accompanied by abrupt shifts in dust and deuterium excess, indicative of reorganizations in atmospheric circulation (Steffensen et al., 2008; Thomas et al., 2009). Reconstructions from the subtropical Atlantic and Mediterranean reveal concomitant SST changes attaining values up to 5°C (e.g., Martrat et al., 2004; Martrat et al., 2007).

In spite of the visible presence of DO events in many paleoclimate records from both hemispheres, the underlying mechanisms still remain unresolved and range from internally generated atmosphere–ocean–ice sheet events (Timmermann et al., 2003; Ditllevsen and Ditllevsen, 2009), to solar-forced variability (Braun et al., 2008; Braun and Kurths, 2010). However, given the lack of observational evidence for a direct linear modulation of solar irradiance on DO time scales, (Muscheler and Beer, 2006), solar forcing is an improbable candidate to generate DO events. There is robust evidence from multiple lines of paleoceanographic information and modelling that DO variability is often associated with AMOC changes, as suggested by climate models of varying complexity (Ganopolski and Rahmstorf, 2001; Arzel et al., 2009) and marine proxy records (Piotrowski et al., 2005; Kissel et al., 2008; Barker et al., 2010; Roberts et al., 2010); but also potential influences of sea-ice cover (Li et al., 2010b), atmosphere circulation and ice sheet topography (Wunsch, 2006) have been proposed.

The widespread presence of massive layers of ice-rafted detritus in North Atlantic marine sediments provide robust evidence that some DO GS, known as Heinrich stadials, were associated with iceberg discharges originating from the Northern Hemispheric ice sheets. During these periods global sea level rose by up to several tens of meters (Chappell, 2002; Rohling et al., 2008b; Siddall et al., 2008; González and Dupont, 2009; Yokoyama and Esat, 2011), with remaining uncertainties in timing and amplitude of sea level rise, stadial cooling and ocean circulation changes relative to the iceberg discharge (Hall et al., 2006; Arz et al., 2007; Siddall et al., 2008; González and Dupont, 2009; Sierra et al., 2009; Hodell et al., 2010). Internal instabilities of the Laurentide ice sheet can cause massive calving and meltwater events similar to those reconstructed from proxy records (Calov et al., 2002, 2010; Marshall and Koutnik, 2006). Alternatively, an initial weakening of the AMOC can lead to subsurface warming in parts of the North Atlantic (Shaffer et al., 2004) and subsequent basal melting of the Labrador ice shelves, and a resulting acceleration of ice streams and iceberg discharge (Alvarez-Solas et al., 2010; Marcott et al., 2011). At present, unresolved dynamics in ice sheet models and limited proxy information do not allow us to distinguish the two mechanisms with confidence.

Since AR4, climate model simulations (Liu et al., 2009b; Otto-Bliesner and Brady, 2010; Meniel et al., 2011; Kageyama et al., 2013) have further confirmed the finding (high confidence) that changes in AMOC strength induce abrupt climate changes with magnitude and patterns resembling reconstructed paleoclimate-proxy data of DO and Heinrich events.

Recent studies have presented a better understanding of the global imprints of DO events and Heinrich events, for various regions. Widespread North Atlantic cooling and sea-ice anomalies during GS induced atmospheric circulation changes (high confidence) (Krebs and Timmermann, 2007; Clement and Peterson, 2008; Kageyama et al., 2010; Merkel et al., 2010; Otto-Bliesner and Brady, 2010; Timmermann et
al., 2010) which in turn affected inter-hemispheric tropical rainfall patterns, leading to drying in Northern South America (Peterson and Haug, 2006), the Mediterranean (Fletcher and Sánchez Goñi, 2008; Fleitmann et al., 2009), equatorial western Africa and Arabia (Higginson et al., 2004; Ivanovich et al., 2005; Weldeab et al., 2007a; Mulitza et al., 2008; Tjallingii et al., 2008; Itambi et al., 2009; Weldeab, 2012), wide parts of Asia (Wang et al., 2008; Cai et al., 2010) (see Figure 5.4e) as well as in the Australian-Indonesian monsoon region (Mohtadi et al., 2011). Concomitant wetter conditions have been reconstructed for southwestern North America (Asmerom et al., 2010; Wagner et al., 2010) and southern South America (Kanner et al., 2012) (Figure 5.4h).

Moreover, atmospheric circulation changes have been invoked (Zhang and Delworth, 2005; Xie et al., 2008; Okumura et al., 2009) to explain temperature variations in the North Pacific that varied in unison with abrupt climate change in the North Atlantic region (Harada et al., 2008, 2012; Pak et al., 2012). Other factors that may have contributed to North Pacific climate anomalies include large-scale Pacific Ocean circulation changes (Saenko et al., 2004; Schmittner et al., 2007; Harada et al., 2009; Okazaki et al., 2010) during phases of a weak AMOC. Recent high-resolution ice core studies (EPICA Community Members, 2006; Capron et al., 2010a, 2010b, 2012; Stenni et al., 2011) show that Antarctica warmed gradually for most GS, reaching maximum values at the time of GS/Gl transitions, which is in agreement with the bipolar seesaw concept (Stocker and Johnsen, 2003; Stenni et al., 2011). A recent global temperature compilation (Shakun et al., 2012), Southern Ocean temperature records (Lamy et al., 2007; Barker et al., 2009; De Deckker et al., 2012), evidence from SH terrestrial records (Kaplan et al., 2010; Putnam et al., 2011) and transient climate model experiments (Menviel et al., 2011) provide multiple lines of evidence for the inter-hemispheric character of millennial-scale variability during the last glacial termination and for DO events (high confidence).

Newly available marine records (Martrat et al., 2007; Grützner and Higgs, 2010; Margari et al., 2010; Kleiven et al., 2011), Antarctic WMGHG records (Loulergue et al., 2008; Schilt et al., 2010) and statistical analyses of Antarctic ice core data (Siddall et al., 2010; Lambert et al., 2012) combined with bipolar seesaw modelling (Siddall et al., 2006; Barker et al., 2011) document with high confidence that abrupt climate change events, similar to the DO events and Heinrich stadials of the last glacial cycle, occurred during previous glacial periods extending back about 800 ka and, with medium confidence, to 1100 ka.

5.8 Paleoclimate Perspective on Irreversibility in the Climate System

For an introduction of the concept of irreversibility see Glossary.

5.8.1 Ice Sheets

Modelling studies suggest the existence of multiple equilibrium states for ice sheets with respect to temperature, CO₂ concentration and orbital forcing phase spaces (DeConto and Pollard, 2003; Calov and Ganopolski, 2005; Ridley et al., 2010). This implies a possibility of irreversible changes in the climate-cryosphere system in the past and future.

The existence of threshold behaviour in the EAIS is consistent with an abrupt increase in Antarctic ice volume at the Eocene/Oligocene boundary, 33 Ma, attributed to gradual atmospheric CO₂ concentration decline on geological time scale (Pagani et al., 2005b; Pearson et al., 2009) (Figure 5.2, Section 5.2.2). Ice sheet models produce a hysteresis behaviour of the EAIS with respect to CO₂ concentrations, leading to EAIS glaciation when CO₂ concentration declined to 600–900 ppm (DeConto and Pollard, 2003; Langebroek et al., 2009) and deglaciation for CO₂ above 1200 ppm (Pollard and DeConto, 2009).

Proxy records suggest that the WAIS might have collapsed during last interglacials (Naish et al., 2009b; Vaughan et al., 2011) and was absent during warm periods of the Pliocene when CO₂ concentration was 350 to 450 ppm (see Section 5.2.2.2) and global sea level was higher than present (see Section 5.6.1). These reconstructions and one ice sheet model simulation (Pollard and DeConto, 2009) suggest that WAIS is very sensitive to the subsurface ocean temperature. This implies, with medium confidence, that a large part of the WAIS will be eventually lost if the atmospheric CO₂ concentration stays within, or above, the range of 350 to 450 ppm for several millennia.

Observational evidence suggest that the GIS was also much smaller than today during the MPWP (see Sections 5.6.1 and 5.2.2), consistent with the results of simulations with ice sheet models (Dolan et al., 2011; Koenig et al., 2011). Ice sheet model simulations and proxy records show that the volume of the GIS was also reduced during the past interglacial period (Section 5.6.2). This supports modelling results that indicate temperature or CO₂ thresholds for melting and re-growth of the GIS may lie in close proximity to the present and future levels (Gregory and Huybrechts, 2006; Lunt et al., 2008) (Section 5.6.1) and that the GIS may have multiple equilibrium states under present-day climate state (Ridley et al., 2010).

Therefore, proxy records and results of model simulations indicate with medium confidence that the GIS and WAIS could be destabilized by projected climate changes, although the time scales of the ice sheets response to climate change are very long (several centuries to millennia).

5.8.2 Ocean Circulation

Numerous modelling studies demonstrate that increased freshwater flux into the North Atlantic leads to weakening of the AMOC. Results of EMICs (Rahmstorf et al., 2005) and coupled GCMs also suggest that AMOC may have multiple equilibrium states under present or glacial climate conditions (Hawkins et al., 2011; Hu et al., 2012). Experiments with climate models provide evidence that the sensitivity of the AMOC to freshwater perturbation is larger for glacial boundary conditions than for interglacial conditions (Swingedouw et al., 2009) and that the recovery time scale of the AMOC is longer for LGM conditions than for the Holocene (Bitz et al., 2007).

The abrupt climate-change event at 8.2 ka permits the study of the recovery time of the AMOC to freshwater perturbation under near-modern boundary conditions (Rohling and Pälike, 2005). Since AR4, new proxy records and simulations confirm that the pattern of surface-ocean and atmospheric climate anomalies is consistent with a reduction in the strength of the AMOC (Figure 5.18a, b, d). Available
proxy records from the North Atlantic support the hypothesis that freshwater input into the North Atlantic reduced the amount of deep and central water-mass formation, Nordic Seas overflows, intermediate water temperatures and the ventilation state of North Atlantic Deep Water (Figure 5.18c, d) (McManus et al., 2004; Ellison et al., 2006; Kleiven et al., 2008; Bamberg et al., 2010). A concomitant decrease of SST and atmospheric temperatures in the North Atlantic and in Greenland has been observed (Figure 5.18a, b) with the climate anomaly associated with the event lasting 100–160 years (Daley et al., 2011). The additional freshwater that entered the North Atlantic during the 8.2 ka event is estimated between 1.6·10^{14} m$^3$ and 8·10^{14} m$^3$ (von Grafenstein et al., 1998; Barber et al., 1999; Clarke et al., 2004). The duration of the

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**Figure 5.18** | Compilation of selected paleoenvironmental and climate model data for the abrupt Holocene cold event at 8.2 ka, documenting temperature and ocean-circulation changes around the event and the spatial extent of climate anomalies following the event. Published age constraints for the period of release of freshwater from glacier lakes Agassiz and Ojibway are bracketed inside the vertical blue bar. Vertical grey bar denotes the time of the main cold event as found in Greenland ice core records (Thomas et al., 2007). Thick lines in (a–d) denote 5-point running mean of underlying data in thin lines. (a) Black curve: North Greenland Ice Core Project (NGRIP) $\delta^{18}$O (temperature proxy) from Greenland Summit (North Greenland Ice Core Project members, 2004). Red curve: Simulated Greenland temperature in an 8.2 ka event simulation with the ECBilt-CLIO-VECODE model (Wiersma et al., 2011). Blue curve: Simulated Greenland temperature in an 8.2 ka event simulation with the CCSM3 model (Morrill et al., 2011). (b) North Atlantic/Nordic Seas sea surface temperature (SST) reconstructions, age models are aligned on the peak of the cold-event (less than 100-year adjustment). Blue curve: Nordic Seas (Risebrobakken et al., 2011). Black curve: Gardar Drift south of Iceland (Ellison et al., 2006). (c) Deep- and intermediate-water records. Black curve: Sortable silt (SS µm) record (overflow strength proxy) from Gardar Drift south of Iceland (Ellison et al., 2006), Atlantic intermediate water temperature reconstruction (Bamberg et al., 2010). (d) Black curve: $\delta^{13}$C (deep water ventilation proxy) at 3.4 km water depth south of Greenland (Kleiven et al., 2008). Age model is aligned on the minimum overflow strength in (c) (less than 100-year adjustment). Modelled change in the strength of the Atlantic Ocean meridional overturning circulation (AMOC)—Green curve: an 8.2 ka event simulation with the GiSS model (LeGrande et al., 2006). Red curve: an 8.2 ka event simulation with the ECBilt-CLIO-VECODE (v. 3) model (Wiersma et al., 2011). Blue curve: an 8.2 ka event simulation with the CCSM3 model (Morrill et al., 2011). (e) Spatial distribution of the 4-member ensemble mean annual mean surface temperature anomaly ($^\circ$C) compared with the control experiment from model simulations of the effects of a freshwater release at 8.2 ka (based on Morrill et al., 2013a). White dots indicate regions where less than 3 models agree on the sign of change. Coloured circles show paleoclimate data from records resolving the 8.2 ka event: purple = cold anomaly, yellow = warm anomaly, grey = no significant anomaly. Data source and significance thresholds are as summarized by Morrill et al. (2013b). (f) Same as (e) but for annual mean precipitation anomalies in %. Coloured circles show paleoclimate data from records resolving the 8.2 ka event: purple = dry anomaly, yellow = wet anomaly, grey = no significant anomaly.
meltwater release may have been as short as 0.5 years (Clarke et al., 2004), but new drainage estimates indicate an up to 200 year-duration in two separate stages (Gregoire et al., 2012). A four-model ensemble with a one-year freshwater perturbation of 2.5 Sv only gives temperature anomalies half of what has been reconstructed and with a shorter duration than observed, resulting from unresolved processes in models, imprecise representation of the initial climate state or a too short duration of the freshwater forcing (Morrill et al., 2013a). These marine-based reconstructions consistently show that the recovery time scale of the shallow and deep overturning circulation is on the order of 200 years (Ellison et al., 2006; Bamberg et al., 2010) (Figure 5.18c, d), with one record pointing to a partial recovery on a decadal time scale (Kleiven et al., 2008). Both recovery time scale and sensitivity of the AMOC to the freshwater perturbation are generally consistent with model experiments for the 8.2 ka event using coarse-resolution models, GCMs and eddy permitting models (LeGrande and Schmidt, 2008; Spence et al., 2008; Li et al., 2009). The recovery of temperatures out of the cold anomaly appears overprinted with natural variability in the proxy data, and is more gradual in data than in the AOGCM experiments (Figure 5.18c, d). In summary, multiple lines of evidence indicate, with high confidence, that the interglacial mode of the AMOC can recover from a short-term freshwater input into the subpolar North Atlantic.

The characteristic teleconnection patterns associated with a colder North Atlantic Ocean as described in Section 5.7 are evident for the 8.2 ka event in both models and proxy data (Figure 5.18e, f).

5.8.3 Next Glacial Inception

Since orbital forcing can be accurately calculated for the future (see Section 5.2.1), efforts can be made to predict the onset of the next glacial period. However, the glaciation threshold depends not only on insolation but also on the atmospheric CO₂ concentration (Archer and Ganopolski, 2005). Models of different complexity have been used to investigate the response to orbital forcing in the future for a range of atmospheric CO₂ levels. These results consistently show that a glacial inception is not expected to happen within the next approximate 50 kyr if either atmospheric CO₂ concentration remains above 300 ppm or cumulative carbon emissions exceed 1000 Pg C (Loutre and Berger, 2000; Archer and Ganopolski, 2005; Cochelin et al., 2006). Only if atmospheric CO₂ content was below the pre-industrial level would a glaciation be possible under present orbital configuration (Loutre and Berger, 2000; Cochelin et al., 2006; Kutzbach et al., 2011; Vettoretti and Peltier, 2011; Tzedakis et al., 2012a). Simulations with climate–carbon cycle models show multi-millennial lifetime of the anthropogenic CO₂ in the atmosphere (see Box 6.1). Even for the lowest RCP 2.6 scenario, atmospheric CO₂ concentrations will exceed 300 ppm until the year 3000. It is therefore virtually certain that orbital forcing will not trigger a glacial inception before the end of the next millennium.

5.9 Concluding Remarks

The assessments in this chapter are based on a rapidly growing body of new evidence from the peer-review literature. Since AR4, there exists a wide range of new information on past changes in atmospheric composition, sea level, regional climates including droughts and floods, as well as new results from internationally coordinated model experiments on past climates (PMIP3/CMIP5). At the regional scale proxy-based temperature estimates are still scarce for key regions such as Africa, India and parts of the Americas. Syntheses of past precipitation changes were too limited to support regional assessments.

Precise knowledge of past changes in atmospheric concentrations of well-mixed GHGs prior to the period for which ice core records are available remains a strong limitation on assessing longer-term climate change. Key limitations to our knowledge of past climate continues to be associated with uncertainties of the quantitative information derived from climate proxies, in particular due to seasonality effects, the lack of proxy records sensitive to winter temperature, or the precise water depth at which ocean proxies signals form. Moreover, methodological uncertainties associated with regional, hemispheric or global syntheses need to be further investigated and quantified.

Despite progress on developing proxy records of past changes in sea ice it is not yet possible to provide quantitative and spatially coherent assessments of past sea ice cover in both polar oceans.

While this assessment could build on improved reconstructions of abrupt climate changes during glacial periods, key questions remain open regarding the underlying cause of these changes. Large uncertainties remain on the variations experienced by the West and East Antarctic ice sheets over various time scales of the past. Regarding past sea level change, major difficulties are associated with deconvolving changes in ocean geodynamic effects, as well as for inferring global signals from regional reconstructions.

The PMIP3/CMIP5 model framework offers the opportunity to directly incorporate information from paleoclimate data and simulations into assessments of future projections. This is an emerging field for which only preliminary information was available for AR5.

Acknowledgements

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12.5.5 Potentially Abrupt or Irreversible Changes

12.5.5.1 Introduction

This report adopts the definition of abrupt climate change used in Synthesis and Assessment Product 3.4 of the U.S. Climate Change Science Program CCSP (CCSP, 2008b). We define abrupt climate change as a large-scale change in the climate system that takes place over a few decades or less, persists (or is anticipated to persist) for at least a few decades, and causes substantial disruptions in human and natural systems (see Glossary). Other definitions of abrupt climate change exist. For example, in the AR4 climate change was defined as abrupt if it occurred faster than the typical time scale of the responsible forcing.

A number of components or phenomena within the Earth system have been proposed as potentially possessing critical thresholds (sometimes referred to as tipping points (Lenton et al., 2008)), beyond which abrupt or nonlinear transitions to a different state ensue. The term irreversibility is used in various ways in the literature. The AR5 report defines a perturbed state as irreversible on a given time scale if the recovery time scale from this state due to natural processes is significantly longer than the time it takes for the system to reach this perturbed state (see Glossary). In that context, most aspects of the climate change resulting from CO₂ emissions are irreversible, due to the long residence time of the CO₂ perturbation in the atmosphere and the resulting warming (Solomon et al., 2009). These results are discussed in Sections 12.5.2 to 12.5.4. Here, we also assess aspects of irreversibility in the context of abrupt change, multiple steady states and hysteresis, i.e., the question whether a change (abrupt or not) would be reversible if the forcing was reversed or removed (e.g., Boucher et al., 2012). Irreversibility of ice sheets and sea level rise are also assessed in Chapter 13.
In this section we examine the main components or phenomena within the Earth system that have been proposed in the literature as potentially being susceptible to abrupt or irreversible change (see Table 12.4). Abrupt changes that arise from nonlinearities within the climate system are inherently difficult to assess and their timing, if any, of future occurrences is difficult to predict. Nevertheless, progress is being made exploring the potential existence of early warning signs for abrupt climate change (see e.g., Dakos et al., 2008; Scheffer et al., 2009).

### 12.5.5.2 The Atlantic Meridional Overturning

EMICs for which the stability has been systematically assessed by suitably designed hysteresis experiments robustly show a threshold beyond which the Atlantic thermohaline circulation cannot be sustained (Rahmstorf et al., 2005). This is also the case for one low-resolution ESM (Hawkins et al., 2011). However, proximity to this threshold is highly model dependent and influenced by factors that are currently poorly understood. There is some indication that the CMIP3 climate models may generally overestimate the stability of the Atlantic Ocean circulation (Hofmann and Rahmstorf, 2009; Drijfhout et al., 2010). In particular, De Vries and Weber (2005), Dijkstra (2007), Weber et al. (2007), Huisman et al. (2010), Drijfhout et al. (2010) and Hawkins et al. (2011) suggest that the sign of net freshwater flux into the Atlantic transported through its southern boundary via the overturning circulation determines whether or not the AMOC is in a mono-stable or bi-stable state. For the pre-industrial control climate of most of the CMIP3 models, Drijfhout et al. (2010) found that the salt flux was negative (implying a positive freshwater flux), indicating that they were in a mono-stable regime. However, this is not the case in the CMIP5 models where Weaver et al. (2012) found that the majority of the models were in a bi-stable regime during RCP integrations. Observations suggest that the present day ocean is in a bi-stable regime, thereby allowing for multiple equilibria and a stable ‘off’ state of the AMOC (Bryden et al., 2011; Hawkins et al., 2011).

In addition to the main threshold for a complete breakdown of the circulation, others may exist that involve more limited changes, such as a cessation of Labrador Sea deep water formation (Wood et al., 1999). Rapid melting of the Greenland ice sheet causes increases in freshwater runoff, potentially weakening the AMOC. None of the CMIP5 simulations include an interactive ice sheet component. However, Jungclaus et al. (2006), Mikolajewicz et al. (2007), Driesschaert et al. (2007) and Hu et al. (2009) found only a slight temporary effect of increased melt water fluxes on the AMOC, that was either small compared to the effect of enhanced poleward atmospheric moisture transport or only noticeable in the most extreme scenarios.

Although many more model simulations have been conducted since the AR4 under a wide range of forcing scenarios, projections of the AMOC behaviour have not changed. Based on the available CMIP5 models, EMICs and the literature, it remains very likely that the AMOC will weaken over the 21st century relative to pre-industrial. Best estimates and ranges for the reduction from CMIP5 are 11% (1 to 24%) in RCP2.6 and 34% (12 to 54%) in RCP8.5 (Weaver et al., 2012) (see Section 12.4.7.2, Figure 12.35). But there is low confidence in the magnitude of the weakening. Drijfhout et al. (2012) show that the AMOC decrease per degree global mean temperature rise varies from 1.5 to 1.9 Sv (10^6 m^3 s^-1) for the CMIP5 multi-model ensemble members they considered depending on the scenario, but that the standard deviation in this regression is almost half the signal.

The FIO-ESM model shows cooling over much of the NH that may be related to a strong reduction of the AMOC in all RCP scenarios (even RCP2.6), but the limited output available from the model precludes an assessment of the response and realism of this response. Hence it is not included the overall assessment of the likelihood of abrupt changes.

### Table 12.4 | Components in the Earth system that have been proposed in the literature as potentially being susceptible to abrupt or irreversible change. Column 2 defines whether or not a potential change can be considered to be abrupt under the AR5 definition. Column 3 states whether or not the process is irreversible in the context of abrupt change, and also gives the typical recovery time scales. Column 4 provides an assessment, if possible, of the likelihood of occurrence of abrupt change in the 21st century for the respective components or phenomena within the Earth system, for the scenarios considered in this chapter.

<table>
<thead>
<tr>
<th>Change in climate system component</th>
<th>Potentially abrupt (AR5 definition)</th>
<th>Irreversibility if forcing reversed</th>
<th>Projected likelihood of 21st century change in scenarios considered</th>
</tr>
</thead>
<tbody>
<tr>
<td>Atlantic MOC collapse</td>
<td>Yes</td>
<td>Unknown</td>
<td>Very unlikely that the AMOC will undergo a rapid transition (high confidence)</td>
</tr>
<tr>
<td>Ice sheet collapse</td>
<td>No</td>
<td>Irreversible for millennia</td>
<td>Exceptionally unlikely that either Greenland or West Antarctic ice sheets will suffer near-complete disintegration (high confidence)</td>
</tr>
<tr>
<td>Permafrost carbon release</td>
<td>No</td>
<td>Irreversible for millennia</td>
<td>Possible that permafrost will become a net source of atmospheric greenhouse gases (low confidence)</td>
</tr>
<tr>
<td>Clathrate methane release</td>
<td>Yes</td>
<td>Irreversible for millennia</td>
<td>Very unlikely that methane from clathrates will undergo catastrophic release (high confidence)</td>
</tr>
<tr>
<td>Tropical forests dieback</td>
<td>Yes</td>
<td>Reversible within centuries</td>
<td>Low confidence in projections of the collapse of large areas of tropical forest</td>
</tr>
<tr>
<td>Boreal forests dieback</td>
<td>Yes</td>
<td>Reversible within centuries</td>
<td>Low confidence in projections of the collapse of large areas of boreal forest</td>
</tr>
<tr>
<td>Disappearance of summer Arctic sea ice</td>
<td>Yes</td>
<td>Reversible within years to decades</td>
<td>Likely that the Arctic Ocean becomes nearly ice-free in September before mid-century under high forcing scenarios such as RCP8.5 (medium confidence)</td>
</tr>
<tr>
<td>Long-term droughts</td>
<td>Yes</td>
<td>Reversible within years to decades</td>
<td>Low confidence in projections of changes in the frequency and duration of megadroughts</td>
</tr>
<tr>
<td>Monsoonal circulation</td>
<td>Yes</td>
<td>Reversible within years to decades</td>
<td>Low confidence in projections of a collapse in monsoon circulations</td>
</tr>
</tbody>
</table>
It is unlikely that the AMOC will collapse beyond the end of the 21st century for the scenarios considered but a collapse beyond the 21st century for large sustained warming cannot be excluded. There is low confidence in assessing the evolution of the AMOC beyond the 21st century. Two of the CMIPS models revealed an eventual slowdown of the AMOC to an off state (Figure 12.35). But this did not occur abruptly.

As assessed by Delworth et al. (2008), for an abrupt transition of the AMOC to occur, the sensitivity of the AMOC to forcing would have to be far greater than that seen in current models. Alternatively, significant ablation of the Greenland ice sheet greatly exceeding even the most aggressive of current projections would be required (Swingedouw et al., 2007; Hu et al., 2009). While neither possibility can be excluded entirely, it is unlikely that the AMOC will collapse beyond the end of the 21st century because of global warming based on the models and range of scenarios considered.

12.5.5.3 Ice Sheets

As detailed in Section 13.4.3, all available modelling studies agree that the Greenland ice sheet will significantly decrease in area and volume in a warmer climate as a consequence of increased melt rates not compensated for by increased snowfall rates and amplified by positive feedbacks. Conversely, the surface mass balance of the Antarctic ice sheet is projected to increase in most projections because increased snowfall rates outweigh melt increase (see Section 13.4.4). Irreversibility of ice sheet volume and extent changes can arise because of the surface-elevation feedback that operates when a decrease of the elevation of the ice sheet induces a decreased surface mass balance (generally through increased melting), and therefore essentially applies to Greenland. As detailed in Section 13.4.3.3, several stable states of the Greenland ice sheet might exist (Charbit et al., 2008; Ridley et al., 2010; Langen et al., 2012; Robinson et al., 2012; Solgaard and Langen, 2012), and the ice sheet might irreversibly shrink to a stable smaller state once a warming threshold is crossed for a certain amount of time, with the critical duration depending on how far the temperature threshold has been exceeded. Based on the available evidence (see Section 13.4.3.3), an irreversible decrease of the Greenland ice sheet due to surface mass balance changes appears very unlikely in the 21st century but likely on multi-centennial to millennial time scales in the strongest forcing scenarios.

In theory (Weertman, 1974; Schoof, 2007) ice sheet volume and extent changes can be abrupt because of the grounding line instability that can occur in coastal regions where bedrock is retrograde (i.e., sloping towards the interior of the ice sheet) and below sea level (see Section 4.4.4 and Box 13.2). This essentially applies to West Antarctica, but also to parts of Greenland and East Antarctica. Furthermore, ice shelf decay induced by oceanic or atmospheric warming might lead to abruptly accelerated ice flow further inland (De Angelis and Skvarca, 2003). Because ice sheet growth is usually a slow process, such changes could also be irreversible in the definition adopted here. The available evidence (see Section 13.4) suggests that it is exceptionally unlikely that the ice sheets of either Greenland or West Antarctica will suffer a near-complete disintegration during the 21st century. More generally, the potential for abrupt and/or irreversible ice sheet changes (or the initiation thereof) during the 21st century and beyond is discussed in detail in Sections 13.4.3 and 13.4.4.

12.5.5.4 Permafrost Carbon Storage

Since the IPCC AR4, estimates of the amount of carbon stored in permafrost have been significantly revised upwards (Tarnocai et al., 2009), putting the permafrost carbon stock to an equivalent of twice the atmospheric carbon pool (Dolman et al., 2010). Because of low carbon input at high latitudes, permafrost carbon is to a large part of Pleistocene (Zimov et al., 2006) or Holocene (Smith et al., 2004) origin, and its potential vulnerability is dominated by decomposition (Eglin et al., 2010). The conjunction of a long carbon accumulation time scale on one hand and potentially rapid permafrost thawing and carbon decomposition under warmer climatic conditions (Zimov et al., 2006; Schuur et al., 2009; Kuhry et al., 2010) on the other hand suggests potential irreversibility of permafrost carbon decomposition (leading to an increase of atmospheric CO$_2$ and/or CH$_4$ concentrations) on time scales of hundreds to thousands of years in a warming climate. Indeed, recent observations (Dorrepaal et al., 2009; Kuhry et al., 2010) suggest that this process, induced by widespread permafrost warming and thawing (Romanovsky et al., 2010), might be already occurring. However, the existing modelling studies of permafrost carbon balance under future warming that take into account at least some of the essential permafrost-related processes (Khvorostyanov et al., 2008; Wania et al., 2009; Koven et al., 2011; Schaefer et al., 2011; MacDougall et al., 2012; Schneider von Deimling et al., 2012) do not yield coherent results beyond the fact that present-day permafrost might become a net emitter of carbon during the 21st century under plausible future warming scenarios (low confidence). This also reflects an insufficient understanding of the relevant soil processes during and after permafrost thaw, including processes leading to stabilization of unfrozen soil carbon (Schmidt et al., 2011), and precludes a firm assessment of the amplitude of irreversible changes in the climate system potentially related to permafrost degassing and associated global feedbacks at this stage (see also Sections 6.4.3.4 and 6.4.7.2 and FAQ 6.1).

12.5.5.5 Atmospheric Methane from Terrestrial and Oceanic Clathrates

Model simulations (Fyke and Weaver, 2006; Reagan and Moridis, 2007; Lamarque, 2008; Reagan and Moridis, 2009) suggest that clathrate deposits in shallow regions (in particular at high latitude regions and in the Gulf of Mexico) are susceptible to destabilization via ocean warming. However, concomitant sea level rise due to changes in ocean mass enhances clathrate stability in the ocean (Fyke and Weaver, 2006). A recent assessment of the potential for a future abrupt release of methane was undertaken by the U.S. Climate Change Science Program (Synthesis and Assessment Product 3.4 see Brooke et al., 2008). They concluded that it was very unlikely that such a catastrophic release would occur this century. However, they argued that anthropogenic warming will very likely lead to enhanced methane emissions from both terrestrial and oceanic clathrates (Brooke et al., 2008). Although difficult to formally assess, initial estimates of the 21st century positive feedback from methane clathrate destabilization are small but not insignificant (Fyke and Weaver, 2006; Archer, 2007; Lamarque, 2008). Nevertheless, on multi-millennial time scales, the positive feedback to anthropogenic
warming of such methane emissions is potentially larger (Archer and Buffett, 2005; Archer, 2007; Brooke et al., 2008). Once more, due to the difference between release and accumulation time scales, such emissions are irreversible. See also FAQ 6.1.

12.5.5.6 Tropical and Boreal Forests

12.5.5.6.1 Tropical forests

In today’s climate, the strongest growth in the Amazonian forest occurs during the dry season when strong insolation is combined with water drawn from underground aquifers that store the previous wet season’s rainfall (Huete et al., 2006). AOGCMs do not agree about how the dry season length in the Amazon may change in the future under the SRES A1 scenario (Bombardi and Carvalho, 2009), but simulations with coupled regional climate/potential vegetation models are consistent in simulating an increase in dry season length, a 70% reduction in the areal extent of the forest by the end of the 21st century using the SRES A2 scenario, and an eastward expansion of the Catinga vegetation (Cook and Vizy, 2008; Sorensson et al., 2010). In addition, some models have demonstrated the existence of multiple equilibria of the tropical South American climate–vegetation system (e.g., Oyama and Nobre, 2003). The transition could be abrupt when the dry season becomes too long for the vegetation to survive, although the resilience of the vegetation to a longer dry period may be increased by the CO₂ fertilization effect (Zelazowski et al., 2011). Deforestation may also increase dry season length (Costa and Pires, 2010) and drier conditions increase the likelihood of wildfires that, combined with fire ignition associated with human activity, can undermine the forest’s resiliency to climate change (see also Section 6.4.8.1). If climate change brings drier conditions closer to those supportive of seasonal forests rather than rainforest, fire can act as a trigger to abruptly and irreversibly change the ecosystem (Malhi et al., 2009). However, the existence of refugia is an important determinant of the potential for the re-emergence of the vegetation (Walker et al., 2009).

Analysis of projected change in the climate–biome space of current vegetation distributions suggest that the risk of Amazonian forest dieback is small (Malhi et al., 2009), a finding supported by modelling when strong carbon dioxide fertilization effects on Amazonian vegetation are assumed (Rammig et al., 2010). However, the strength of CO₂ fertilization on tropical vegetation is poorly known (see Box 6.3). Uncertainty concerning the existence of critical thresholds in the Amazonian and other tropical rainforests purely driven by climate change therefore remains high, and so the possibility of a critical threshold being crossed in precipitation volume cannot be ruled out (Nobre and Borma, 2009; Good et al., 2011b, 2011c). Nevertheless, there is still some question as to whether a transition of the Amazonian or other tropical rainforests into a lower biomass state could result from the combined effects of limits to carbon fertilization, climate warming, potential precipitation decline in interaction with the effects of human land use.

12.5.5.6.2 Boreal forest

Evidence from field observations and biogeochemical modelling make it scientifically conceivable that regions of the boreal forest could tip into a different vegetation state under climate warming, but uncertainties on the likelihood of this occurring are very high (Lenton et al., 2008; Allen et al., 2010). This is mainly due to large gaps in knowledge concerning relevant ecosystemic and plant physiological responses to warming (Niinemets, 2010). The main response is a potential advancement of the boreal forest northward and the potential transition from a forest to a woodland or grassland state on its dry southern edges in the continental interiors, leading to an overall increase in herbaceous vegetation cover in the affected parts of the boreal zone (Lucht et al., 2006). The proposed potential mechanisms for decreased forest growth and/or increased forest mortality are: increased drought stress under warmer summer conditions in regions with low soil moisture (Barber et al., 2000; Dulamsuren et al., 2009, 2010); desiccation of saplings with shallow roots due to summer drought periods in the top soil layers, causing suppression of forest reproduction (Hogg and Schwarz, 1997); leaf tissue damage due to high leaf temperatures during peak summer temperatures under strong climate warming; and increased insect, herbivory and subsequent fire damage in damaged or struggling stands (Dulamsuren et al., 2008). The balance of effects controlling standing biomass, fire type and frequency, permafrost thaw depth, snow volume and soil moisture remains uncertain. Although the existence of, and the thresholds controlling, a potential critical threshold in the boreal forest are extremely uncertain, its existence cannot at present be ruled out.

12.5.5.7 Sea Ice

Several studies based on observational data or model hindcasts suggest that the rapidly declining summer Arctic sea ice cover might reach or might already have passed a tipping point (Lindsay and Zhang, 2005; Wadhams, 2012; Livina and Lenton, 2013). Identifying Arctic sea ice tipping points from the short observational record is difficult due to high interannual and decadal variability. In some climate projections, the decrease in summer Arctic sea ice areal coverage is not gradual but is instead punctuated by 5- to 10- year periods of strong ice loss (Holland et al., 2006; Vavrus et al., 2012; Dösscher and Koenigk, 2013). Still, these abrupt reductions do not necessarily require the existence of a tipping point in the system or further imply an irreversible behaviour (Amstrup et al., 2010; Lenton, 2012). The 5- to 10-year events discussed by Hol-land et al. (2006) arise when large natural climate variability in the Arctic reinforces the anthropogenically-forced change (Holland et al., 2008). Positive trends on the same time scale also occur when internal variability counteracts the forced change until the middle of the 21st century (Holland et al., 2008; Kay et al., 2011; Vavrus et al., 2012).

Further work using single-column energy-balance models (Merryfield et al., 2008; Eisenman and Wettlaufer, 2009; Abbot et al., 2011) yielded mixed results about the possibility of tipping points and bifurcations in the transition from perennial to seasonal sea ice cover. Thin ice and snow covers promote strong longwave radiative loss to space and high ice growth rates (e.g., Bitz and Roe, 2004; Notz, 2009; Eisenman, 2012). These stabilizing negative feedbacks can be large enough to overcome the positive surface–albedo feedback and/or cloud feedback, which act to amplify the forced sea ice response. In such low-order models, the emergence of multiple stable states with increased climate forcing is a parameter-dependent feature (Abbot et al., 2011; Eisenman, 2012). For example, Eisenman (2012) showed with a single-column energy-bal ance model that certain parameter choices that cause thicker ice or warmer ocean under a given climate forcing make the model more prone to bifurcations and hence irreversible behaviour.
The reversibility of sea ice loss with respect to global or hemispheric mean surface temperature change has been directly assessed in AOGCMs/ESMs by first raising the CO$_2$ concentration until virtually all sea ice disappears year-round and then lowering the CO$_2$ level at the same rate as during the ramp-up phase until it reaches again the initial value (Armour et al., 2011; Boucher et al., 2012; Ridley et al., 2012; Li et al., 2013b). None of these studies show evidence of a bifurcation leading to irreversible changes in Arctic sea ice. AOGCMs have also been used to test summer Arctic sea ice recovery after either sudden or very rapid artificial removal, and all had sea ice return within a few years (Schröder and Connolley, 2007; Sedláček et al., 2011; Tietsche et al., 2011). In the Antarctic, as a result of the strong coupling between the Southern Ocean’s surface and the deep ocean, the sea ice areal coverage in some of the models integrated with ramp-up and ramp-down atmospheric CO$_2$ concentration exhibits a significant lag relative to the global or hemispheric mean surface temperature (Ridley et al., 2012; Li et al., 2013b), so that its changes may be considered irreversible on centennial time scales.

Diagnostic analyses of a few global climate models have shown abrupt sea ice losses in the transition from seasonal to year-round Arctic ice-free conditions after raising CO$_2$ to very high levels (Winton, 2006b; Ridley et al., 2008; Li et al., 2013b), but without evidence for irreversible changes. Winton (2006b, 2008) hypothesized that the small ice cap instability (North, 1984) could cause such an abrupt transition. With a low-order Arctic sea ice model, Eisenman and Wettlaufer (2009) also found an abrupt change behaviour in the transition from seasonal ice to year-round ice-free conditions, accompanied by an irreversible bifurcation to a new stable, annually ice-free state. They concluded that the cause is a loss of the stabilizing effect of sea ice growth when the ice season shrinks in time. The Arctic sea ice may thus experience a sharp transition to annually ice-free conditions, but the irreversible nature of this transition seems to depend on the model complexity and structure.

In conclusion, rapid summer Arctic sea ice losses are likely to occur in the transition to seasonally ice-free conditions. These abrupt changes might have consequences throughout the climate system as noted by Vavrus et al. (2011) for cloud cover and Lawrence et al. (2008b) for the high-latitude ground state. Furthermore, the interannual-to-decadal variability in the summer Arctic sea ice extent is projected to increase in response to global warming (Holland et al., 2008; Goosse et al., 2009). These studies suggest that large anomalies in Arctic sea ice areal coverage, like the ones that occurred in 2007 and 2012, might become increasingly frequent. However, there is little evidence in global climate models of a tipping point (or critical threshold) in the transition from a perennially ice-covered to a seasonally ice-free Arctic Ocean beyond which further sea ice loss is unstoppable and irreversible.

12.5.5.8 Hydrologic Variability: Long-Term Droughts and Monsoonal Circulation

12.5.5.8.1 Long-term Droughts

As noted in Section 5.5.5, long-term droughts (often called megadroughts, see Glossary) are a recurring feature of Holocene paleoclimate records in North America, East and South Asia, Europe, Africa and India. The transitions into and out of the long-term droughts take many years. Because the long-term droughts all ended, they are not irreversible. Nonetheless transitions over years to a decade into a state of long-term drought would have impacts on human and natural systems. AR4 climate model projections (Milly et al., 2008) and CMIP5 ensembles (Figure 12.23) both suggest widespread drying and drought across most of southwestern North America and many other subtropical regions by the mid to late 21st century (see Section 12.4.5), although without abrupt change. Some studies suggest that this subtropical drying may have already begun in southwestern North America (Seager et al., 2007; Seidel and Randel, 2007; Barnett et al., 2008; Pierce et al., 2008). More recent studies (Hoerling et al., 2010; Seager and Vecchi, 2010; Dai, 2011; Seager and Naik, 2012) suggest that regional reductions in precipitation are due primarily to internal variability and that the anthropogenic forced trends are currently weak in comparison.

While previous long-term droughts in southwest North America arose from natural causes, climate models project that this region will undergo progressive aridity as part of a general drying and poleward expansion of the subtropical dry zones driven by rising GHGs (Held and Soden, 2006; Seager et al., 2007; Seager and Vecchi, 2010). The models project the aridification to intensify steadily as RF and global warming progress without abrupt changes. Because of the very long lifetime of the anthropogenic atmospheric CO$_2$ perturbation, such drying induced by global warming would be largely irreversible on millennium time scale (Solomon et al., 2009; Frölicher and Joos, 2010; Gillett et al., 2011) (see Sections 12.5.2 and 12.5.4). For example, Solomon et al. (2009) found in a simulation where atmospheric CO$_2$ increases to 600 ppm followed by zero emissions, that the 15% reduction in precipitation in areas such as southwest North America, southern Europe and western Australia would persist long after emissions ceased. This, however, is largely a consequence of the warming persisting for centuries after emissions cease rather than an irreversible behaviour of the water cycle itself.

12.5.5.8.2 Monsoonal circulation

Climate model simulations and paleo-reconstructions provide evidence of past abrupt changes in Saharan vegetation, with the ‘green Sahara’ conditions (Hoelzmann et al., 1998) of the African Humid Period (AHP) during the mid-Holocene serving as the most recent example. However, Mitchell (1990) and Claussen et al. (2003) note that the mid-Holocene is not a direct analogue for future GHG-induced climate change since the forcings are different: a increased shortwave forcing in the NH summer versus a globally and seasonally uniform atmospheric CO$_2$ increase, respectively. Paleoclimatic examples suggest that a strong radiative or SST forcing is needed to achieve a rapid climate change, and that the rapid changes are reversible when the forcing is withdrawn. Both the abrupt onset and termination of the AHP were triggered when northern African summer insolation was 4.2% higher than present day, representing a local increase of about 19 W m$^{-2}$ (deMenocal et al., 2000). Note that the globally averaged radiative anthropogenic forcing from 1750 to 2011 (Table 8.6) is small compared to this local increase in insolation. A rapid Saharan greening has been simulated in a climate model of intermediate complexity forced by a rapid increase in atmospheric CO$_2$, with the overall extent of greening depending on the equilibrium atmospheric CO$_2$ level reached (Claussen et al., 2003).
Abrupt Saharan vegetation changes of the Younger Dryas are linked with a rapid AMOC weakening which is considered very unlikely during the 21st century and unlikely beyond that as a consequence of global warming.

Studies with conceptual models (Zickfeld et al., 2005; Levermann et al., 2009) have shown that the Indian summer monsoon can operate in two stable regimes: besides the ‘wet’ summer monsoon, a stable state exists which is characterized by low precipitation over India. These studies suggest that any perturbation of the radiative budget that tends to weaken the driving pressure gradient has the potential to induce abrupt transitions between these two regimes.

Numerous studies with coupled ocean–atmosphere models have explored the potential impact of anthropogenic forcing on the Indian monsoon (see also Section 14.2). When forced with anticipated increases in GHG concentrations, the majority of these studies show an intensification of the rainfall associated with the Indian summer monsoon (Meehl and Washington, 1993; Kitoh et al., 1997; Douville et al., 2000; Hu et al., 2000; May, 2002; Ueda et al., 2006; Kripalani et al., 2007; Stowasser et al., 2009; Cherchi et al., 2010). Despite the intensification of precipitation, several of these modelling studies show a weakening of the summer monsoon circulation (Kitoh et al., 1997; May, 2002; Ueda et al., 2006; Kripalani et al., 2007; Stowasser et al., 2009; Cherchi et al., 2010). The net effect is nevertheless an increase of precipitation due to enhanced moisture transport into the Asian monsoon region (Ueda et al., 2006). In recent years, studies with GCMs have also explored the direct effect of aerosol forcing on the Indian monsoon (Lau et al., 2006; Meehl et al., 2008; Randles and Ramaswamy, 2008; Collier and Zhang, 2009). Considering absorbing aerosols (black carbon) only, Meehl et al. (2008) found an increase in pre-monsoonal precipitation, but a decrease in summer monsoon precipitation over parts of South Asia. In contrast, Lau et al. (2006) found an increase in May–June–July precipitation in that region. If an increase in scattering aerosols only is considered, the monsoon circulation weakens and precipitation is inhibited (Randles and Ramaswamy, 2008). More recently, Bollasina et al. (2011) showed that anthropogenic aerosols played a fundamental role in driving the recent observed weakening of the summer monsoon. Given that the effect of increased atmospheric regional loading of aerosols is opposed by the concomitant increases in GHG concentrations, it is unlikely that an abrupt transition to the dry summer monsoon regime will be triggered in the 21st century.

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