Chapter 11: Regional Climate Projections

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Supplementary Material

The following supplementary material is available on CDRom and in on-line versions of this report.

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Executive Summary

Increasingly reliable regional climate change projections are now available for many regions of the world due to advances in modeling and our understanding of the physical processes of the climate system. A number of important themes have emerged:

- Warming over many land areas is greater than global annual mean warming due to less water availability for evaporative cooling and a smaller thermal inertia as compared to the oceans;

- Warming generally increases spatial variability of precipitation, contributing to a reduction of rainfall in the subtropics and an increase in higher latitudes and parts of the tropics. The precise location of boundaries between regions of robust increase and decrease remains uncertain and this is commonly where AOGCM projections disagree;

- The poleward expansion of the subtropical highs, combined with the general tendency towards reduction in subtropical precipitation, creates especially robust projections of a reduction in precipitation on the poleward edges of the subtropics. Most of the regional projections of reductions in precipitation in the 21st century are associated with areas adjacent to these subtropical highs;

- There is a tendency for monsoonal circulations to result in increased precipitation due to enhanced moisture convergence, despite a tendency towards weakening of the monsoonal flows themselves. However, many aspects of tropical climatic responses remain uncertain.

AOGCMs remain the primary source of regional information on the range of possible future climates. A clearer picture of the robust aspects of regional climate change is emerging due to improvement in model resolution, the simulation of processes of importance for regional change, and the expanding set of available simulations. Advances have been made in developing probabilistic information on regional scales from the AOGCM simulations, but these methods remain in the exploratory phase. There has been less development extending this to downscaled regional information. However, downscaling methods have matured since the TAR (IPCC, 2001) and have been more widely applied, although, only in some regions has large-scale coordination of multi-model downscaling climate change simulations been achieved.

Regional climate change projections presented here are assessed drawing on information from four potential sources: AOGCM simulations; downscaling of AOGCM-simulated data using techniques to enhance regional detail; physical understanding of the processes governing regional responses; and recent historical climate change.

Previous chapters describe observed climate change on regional scales (Chapter 3) and compare global model simulations with these changes (Chapter 9). We can use comparisons of model simulations of temperature change with observations to help constrain future regional temperature projections. Regional assessments of precipitation change rely primarily on convergence in both global and downscaling models along with physical insights. Where there is near unanimity among models with good supporting physical arguments, as is more typical for middle and higher latitudes, these factors encourage stronger statements as to the likelihood of a regional climate change. In some circumstances physical insights alone clearly indicate the direction of future change.

The summary likelihood statements on projected regional climate are as follows:

- *Temperature projections:* These are comparable in magnitude to those of the TAR and confidence in the regional projections is now higher due to a larger number and variety of simulations, improved models, a better understanding of the role of model deficiencies, and more detailed analyses of the results. Warming, often greater than the global mean, is very likely over all landmasses;

- *Precipitation projections:* Overall patterns of change are comparable to those of TAR, with greater confidence in the projections for some regions. Model agreement is seen over more and larger regions. For some regions there are grounds for stating that the projected precipitation changes are likely or very likely. For other regions confidence in the projected change remains weak;

- *Extremes:* There is a large increase in the available analyses on changes in extremes. This allows for a more comprehensive assessment for most regions. The general findings are in line with the assessment made in TAR and now have a higher level of confidence derived from multiple sources of information. The most notable improvements in confidence relate to the regional statements concerning heat waves, heavy precipitation, and droughts. Despite these advances, specific analyses of models are not available in some regions which is reflected in the robust statements on extremes. In particular, projections concerning extreme events in the tropics remain uncertain. The difficulty in projecting the distribution of tropical cyclones adds to this uncertainty. Changes in extra-tropical cyclones are dependent on details in regional atmospheric circulation response, some of which remain uncertain.

The following summarizes the robust findings of the projected regional change over the 21st century. Supporting narratives are provided in Sections 11.2 to 11.9. These changes are assessed as likely to very likely taking into account the uncertainties in climate sensitivity and emission trajectories (in the SRES B1/A1B/B2 range) discussed in earlier Chapters.

All land regions:
It is very likely that all land regions will warm in the 21st century.

Africa:
Warming is very likely to be larger than the global annual-mean warming throughout the continent and in all seasons, with drier subtropical regions warming more than the moister tropics. Annual rainfall is likely to decrease in much of Mediterranean Africa and Northern Sahara, with a greater likelihood of decreasing rainfall as one approach the Mediterranean coast. Rainfall in southern Africa will likely decrease in much of the winter rainfall region and western margins. There will likely be an increase in annual mean rainfall in East Africa. It is unclear how rainfall in the Sahel, the Guinean Coast, and the Southern Sahara will evolve.

Mediterranean and Europe:
Annual mean temperatures in Europe are likely to increase more than the global mean. Seasonally, the largest warming in likely to be in northern Europe in winter and in the Mediterranean area in summer. Minimum winter temperatures are likely to increase more than the average in northern Europe. Maximum summer temperatures are likely to increase more than the average in southern and central Europe. Annual precipitation is very likely to increase in most of northern Europe and decrease in most of the Mediterranean area. In central Europe, precipitation is likely to increase in winter but decrease in summer. Extremes of daily precipitation will very likely increase in northern Europe. The annual number of precipitation days is very likely to decrease in the Mediterranean area. Risk of summer drought is likely to increase in central Europe and in the Mediterranean area. Snow season is very likely to shorten, and snow depth is likely to decrease in most of Europe.

Asia:
All of Asia is very likely to warm during this century . Warming is likely to be well above the global mean in Central Asia, Tibetan Plateau and Northern Asia, above in Eastern Asia and South Asia, and similar to the global mean in Southeast Asia. Precipitation in winter will very likely increase in Northern Asia and the Tibetan Plateau, and likely increase in Eastern Asia and the southern parts of Southeastern Asia. Precipitation in summer will likely increase in Northern Asia, East Asia, South Asia and most of Southeast Asia, but it will likely decrease in Central Asia. It is very likely that heat waves / hot spells in summer will be of longer duration, more intense, and more frequent in East Asia. Fewer very cold days are very likely in East Asia and South Asia. There will very likely be an increase in frequency of intense precipitation events in parts of South Asia, and in East Asia. Extreme rainfall and winds associated with tropical cyclones are likely to increase in East Asia, Southeast Asia and South Asia.
North America:
The annual mean warming is likely to exceed the global mean warming in most areas. Seasonally, warming is likely to be largest in winter in northern regions and in summer in the south-west. Minimum winter temperatures are likely to increase more than the average in northern North America. Maximum summer

temperatures are likely to increase more than the average in the south-west. Annual-mean precipitation is very likely to increase in Canada and north-east USA, and likely to decrease in the south-west. In southern Canada, precipitation is likely to increase in winter and spring but decrease in summer. Snow season length and snow depth are very likely to decrease in most of North America except in the northernmost part of Canada where maximum snow depth is likely to increase.

Central and South America:
The annual mean warming is likely to be similar to the global mean warming in Southern South America but larger than the global mean warming in the rest of the area. Annual precipitation is likely to decrease in most of Central America and in the Southern Andes, although changes in atmospheric circulation may induce large local variability in precipitation response in mountainous areas. WINTER precipitation in Tierra del Fuego and summer precipitation in south-eastern South America is likely to increase. Winter precipitation in Tierra del Fuego and summer precipitation in south-eastern South America is likely to increase. It is uncertain how annual and seasonal mean rainfall will change over northern South America, including the Amazon forest. However, there is qualitative consistency among the simulations in some areas (rainfall increasing in Ecuador and northern Peru, and decrease in the northern tip of the continent and in southern northeast Brazil).

Australia - New Zealand:
Warming is likely to be larger than that of the surrounding oceans, but comparable to the global mean. The warming is less in the south, especially in winter, with the warming in the South Island of New Zealand likely to remain less than the global mean. Precipitation is likely to decrease in Southern Australia in winter and spring. Precipitation is very likely to decrease in Southwestern Australia in winter. Precipitation is likely to increase in the west of the South Island of New Zealand. Changes in rainfall in Northern and Central Australia are uncertain. Increased mean wind speed is likely across the South Island of New Zealand, particularly in winter. Increased frequency of extreme high daily temperatures in Australia and New Zealand, and decrease in the frequency of cold extremes is very likely. Extremes of daily precipitation will very likely increase, except possibly in areas of significant decrease in mean rainfall (southern Australia in winter and spring.). Increased risk of drought in southern areas of Australia is likely.

Polar regions:
The Arctic is very likely to warm during this century more than the global mean. Warming is projected to be largest in winter and less in summer. Annual Arctic precipitation is very likely to increase. It is very likely that the relative precipitation increase is largest in winter and smallest in summer. Arctic sea ice is very likely to decrease in its extent and thickness. It is uncertain how the Arctic Ocean circulation will change. The Antarctic is likely to warm and the precipitation will likely increase over the continent. It is uncertain to what extent the frequency of extreme temperature and precipitation events will change in the polar regions.

Small Islands:
Sea levels will likely rise on average during the century around the small islands of the Caribbean Sea, Indian Ocean and Northern and Southern Pacific Oceans. The rise will likely not be geographically uniform but large deviations among models make regional estimates across the Caribbean, Indian and Pacific Oceans uncertain. All Caribbean, Indian Ocean, as well as North and South Pacific islands are very likely to warm during this century. The warming is likely to be somewhat smaller than the global annual mean. Summer rainfall in the Caribbean is likely to decrease in the vicinity of the Greater Antilles but changes elsewhere and in winter are uncertain. Annual rainfall is likely to increase in the northern Indian Ocean with increases likely in the vicinity of the Seychelles in December-January-February, and in the vicinity of the Maldives in June-July-August while decreases are likely in the vicinity of Mauritius in June-July-August. Annual rainfall is likely to increase in the equatorial Pacific, while decreases are projected by most models just east of French Polynesia in December-January-February.

11.1 Introduction

Increasingly reliable regional climate change projections are now available for many regions of the world due to advances in modeling and our understanding of the physical processes of the climate system. AOGCMs remain the foundation for projections while downscaling techniques now provide valuable additional detail. AOGCMs cannot provide information at scales finer than their computational grid (typically of the order of 200km) and processes at the unresolved scales are important. Providing information at finer scales can be achieved through using high resolution in dynamical models or empirical statistical downscaling (SD). Development of downscaling methodologies remains an important focus. Downscaled climate change projections that are tailored to specific needs are only now starting to become available.

11.1.1 Summary of TAR

The assessment of regional climate projections in the TAR (IPCC, 2001; Chapter 10) was largely restricted to GCM-derived temperature with limited precipitation statements. The major assessment of temperature change was that it is very likely all land areas will warm more than the global average (with the exception of Southeast Asia and South America in JJA), with amplification at high latitudes. The changes in precipitation assessed to be likely were: an increase over northern mid-latitude regions in winter and over high-latitude regions in both winter and summer; in DJF, an increase in tropical Africa, little change in Southeast Asia, and a decrease in Central America; an increase or little change in JJA over South Asia and a decrease over Australia and the Mediterranean region. These projections were almost entirely based on analysis of 9 coarse-resolution AOGCMs that had performed transient experiments for the 20th century with the specifications for the A2 and B2 emission scenarios. TAR Chapter 10 noted that studies with regional models indicate that changes at finer scales may be substantially different in magnitude from these large sub-continental findings.

Information available for assessment regarding climate variability and extremes at the regional scale was too sparse for it to be meaningful to draw it together in a systematic manner. However, some statements of a more generic nature were made. It was assessed that the variability of daily to interannual temperatures is likely to decrease in winter and increase in summer for mid-latitude Northern Hemisphere land areas; daily high temperature extremes will likely increase; future increase in mean precipitation will very likely lead to an increase in variability. In some specifically analysed regions it was assessed that extreme precipitation may increase and there were indications that droughts or dry spells may increase in occurrence in Europe, North America and Australia.

11.1.2 Introduction to Regional Projections

Assessments of climate change projections are provided here on a region by region basis. The discussion is organized according to the same continental-scale regions used in WGII in the AR4 and in earlier assessments: *Africa, Europe and Mediterranean, Asia, North America, Central and South America, Australia-New Zealand, Polar Regions, and Small Islands.* While the topics covered vary somewhat from region to region, each section includes a discussion of key processes of importance for climate change in that region, relevant aspects of model skill in simulating current climate, and projections of future regional climate change based on global models and downscaling techniques.

Each of these continental-scale regions encompasses a broad range of climates and is too large to be used as a basis for conveying quantitative regional climate change information. Therefore, each is subdivided into a number of sub-continental or oceanic regions. The sub-continental regions as defined in Table 11.1 are the framework for developing specific regional or sub-continental robusts statements of projected change. Area-averaged temperature and precipitation changes are presented from the coordinated set of climate model simulations archived at PCMDI (subsequently called the multi-model dataset or MMD). The regions are very close to those initially devised by Giorgi and Francesco (2000) with some minor modifications similar to those of Ruosteenoja et al. (2003). They have simple shapes and are no smaller than the horizontal scales on which current AOGCMs are useful for climate simulations (typically judged to be roughly a thousand kilometres).

These regional averages have some deficiencies for discussion of the AOGCM projections. In several instances, the simple definition of these boxes results in spatial averaging over regions in which precipitation is projected to increase and decrease. There are also sub- regions where the case can be made for a robust and physically plausible hydrological response, information about which is lost in the regional averages. Partially to help in discussing these features, we also use maps of temperature and precipitation responses, interpolated to a 128 (longitude) x 64 (latitude) grid typical of many of the lower resolution atmospheric models in the MMD.

In the regional discussion to follow, the starting points are temperature and precipitation. Changes in temperature are introduced in each continental section by plotting for each of the regions the evolution of the range of projected decadal mean change for the A1B scenario through the 21st century (simulations hereafter referred to as MMD-A1B). These are put into context of observed changes in the 20th century by plotting the observed changes and how well the models reproduce these. This summary information is displayed for continental regions in Box 11.1 which also contains details of how the figures were constructed. The equivalent figures for the individual regions of each continental-scale region are then displayed in the following sections. These are constructed in the same way as Box 11.1, Figure 1. The 20th century parts of these figures are also displayed in Chapter 9, section 9.4 where more details on their construction are provided.The discussion on precipitation provides a limited view of hydrological changes. Supplementary material Figure S11.1 expands on this issue by comparing the annual mean responses in precipitation and in precipitation minus evaporation over the 21st century in the MMD-A1B projections. Over North America and Europe, for example, the region of drying in the sense of precipitation minus evaporation is shifted polewards compared to the region of reduced precipitation. A summary of the more significant hydrological cycle changes from the regional discussions is presented in Box 11.1.

Box 11.1: Summary of Regional Responses

As an introduction to the more detailed regional analysis presented in this chapter, we illustrate in Box 11.1 Figure 1 how continental scale warming is projected to evolve in the 21st century using the MMD models. We also put this warming into the context of the observed warming during the 20th century by comparing results from that subset of the models incoporating a representation of all known forcings with the observed evolution (see Chapter 9.4 for more details). Thus for the six continental region we display: 1) the observed time series of the evolution of decadally-averaged surface air temperature from 1906–2005 as an anomaly from the 1901–1950 average; 2) the range of the equivalent anomalies derived from 20th century simulations by the MMD models that contain a full set of historical forcings; 3) the evolution of the range of this anomaly in MMD-A1B projections between 2000 and 2100, and 4) the range of the projected anomaly for the last decade of the 21st century for the B1, A1B, and A2 scenarios. For the observed part of these graphs, the decadal averages are centred on the decade boundaries, i.e. the last point is for 1996–2005, whereas for the future period they are centred on the decade mid-points, i.e. the first point is for 2001–2010. The width of the shading and the bars represents the 5–95% range of the model results. To construct the ranges all simulations from the set of models involved were considered independent realisations of the possible evolution of the climate given the forcings applied. This involved 58 simulations from 14 models for the observed period and 47 simulations from 18 models for the future. Important in this representation is that the models' estimate of natural climate variability is included and thus the ranges include both the potential mitigating and amplifying effect of variability on the underlying signal. In contrast, the bars representing the range of projected change at the end of the century are constructed from ensemble mean changes from the models and thus provide a measure of the forced response. These bars were constructed from decadal mean anomalies from 21 models using A1B scenario forcings, from the 20 of these model that used the B1 forcings and the 17 that used the A2. The bars for the B1 and A2 scenarios were scaled to approximate ranges for the full set of models. The scaling factor for B1 was derived from the ratio between its range and the A1B range of the corresponding 20 models. The same procedure was used to obtain the A2 scaling factor. Only 18 models were used to display the ranges of projected temperature evolution as the control simulations for the other 3 had a drift of >0.2°C per century which precludes clearly defining the decadal anomalies from these models. However, anomalies from all 21 models were included in calculating the bars in order to provide the fullest possible representation of projected changes in the MMD. Comparison of these different representations shows that the main messages from the MMD about projected continental temperature change are insensitive to the choices made. Finally, results are not shown here for Antarctica because the observational record is not long enough to provide the relevant information for the first part of the 20th century. Results of a similar nature to those shown here using the observations that are available are presented in Section 11.8.

[INSERT BOX 11.1, FIGURE 1 HERE]

Box11.1, Figure 2 serves to illustrate some of the more significant hydrological changes, with the two panels corresponding to the months of December-January-February and June-July-August. The backdrop to these figures is the fraction of the GCMs (out of the 21 considered for this purpose) that predict an increase in mean precipitation in that grid cell (using the A1B scenario and comparing the period 2080–2099 with the control 1980–1999). Aspects of this pattern are examined more closely in the separate regional discussions. Robust findings on regional climate change for mean and extreme precipitation, drought and snow are highlighted on the figure with further detail in the accompanying notes.

[INSERT BOX 11.1, FIGURE 2 HERE]

Table 11.1 provides detailed information for each region generated from the MMD-A1B models focusing on the change in climate between the 1980–1999 period in the 20th century integrations and the 2080–2099 period. The distribution of the annual and seasonal mean surface air temperature response and percentage change in precipitation are described by the median, the 25% and 75% values (half of the models lie between these two values) and the maximum and minimum values in the model ensemble. Information on model biases in these regional averages for the 1980–1999 simulations is provided in Supplementary material Table S11.1 in a similar format. Maps of biases are referred to in some of the following and are included in Supplementary material as well. Data sources used in these comparisons are listed in the table and figure captions where these biases are displayed.

Most of the discussion focuses on the A1B scenario. The global mean near-surface temperature responses (between the period 1980–1999 of the 20th century integrations and the period 2080–2099) in the ensemble mean of the MMD models are in the ratio 0.69:1:1.17 for the B1:A1B:A2 scenarios. The local temperature responses in nearly all regions closely follow the same ratio, as discussed in Chapter 10 and as illustrated in Supplementary material Figures S11.2-4. Therefore, little is gained by repeating discussion of the A1B scenario for the other scenarios. The ensemble mean local precipitation responses also approximately scale with the global mean temperature response, although not as precisely as the temperature itself. Given the substantial uncertainties in hydrological responses, the generally smaller signal/noise ratio, and the similarities in the basic structure of the GCM precipitation responses in the different scenarios, a focus on A1B seems justified for the precipitation as well. The overall regional assessments, however, do rely on all available scenario information.

Given the dominantly linear response of the models, the 2080–2099 period allows the greatest clarity of the background climate change underlying the interannual and decadal variability. In the ensemble mean AOGCM projections there is no indication of abrupt climate change, nor does the literature on individual models provide any strong suggestions of robust nonlinearities. Some local temporal non-linearities are to be expected, for example as the sea ice boundary retreats from a particular location in the Arctic. While the possibility exists that changes of more abrubt character could happen, such as major ocean circulation or land surface/vegetation change, there is little basis to judge the plausibility of these factors (see Chapter 10). Therefore, we base our discussion on this linear picture.

Table 11.1 also provides some simple estimates of the signal-to-noise ratio. The signal is the change in 20 year means of seasonal or annual mean temperature or precipitation. The noise is an estimate of the internal variability of 20 year means of seasonal or annual mean temperature or precipitation, as generated by the models. The signal-to-noise ratio is converted into the time interval that is required before the signal is *clearly discernable*, assuming that the signal grows linearly over the century as the average rate in the ensemble mean A1B projection. We define clearly discernible in this context as distinguishable with 95% confidence. As an example, the annual mean precipitation increase in Northern Europe (NEU) (Table 11.1) is clearly discernible in these models after 45 years, meaning that the 20 year average from 2025–2044 will be greater than the 20 year mean over 1980–1999 with 95% confidence, accounting only for the internal variability in the models and no other sources of uncertainty. In contrast, the annual temperature response in South-East Asia (SEA) rises above the noise by this measure after only 10 years, implying that the average temperature over the period 1990–2009 is clearly discernible in the models from the average over the control 1980–1999 period. This measure is likely an overestimate of the time of emergence of the signal as compared to that obtained with more refined detection strategies (of the kind discussed in Chapter 9). This noise estimate is solely based on the models and must be treated with caution, but it would be wrong to assume that models always underestimate this internal variability. Some models overestimate and some underestimate the amplitude of ENSO, for example, thereby over- or under-estimating the most important source of interannual variability in the tropics. On the other hand, few models capture the range of decadal variability of rainfall in West Africa for example (Hoerling, et. al. 2006; Chapter 8, Section 8.4).

Also included in Table 11.1 is an estimate of the probability of *extremely warm, extremely wet, and extremely dry seasons*, for the A1B scenario and for the time period 2080–2099. An *extremely warm* summer is defined as follows. Examining all of the summers simulated in a particular realization of a model in the 1980–1999 control period, one can compute the warmest of these 20 summers, as an estimate of the temperature of the warmest 5% of all summers in the control climate. One then examines the period 2080–2099, and determines what fraction of the summers exceed this warmth. This is referred to as the probability of extremely warm summers. The results are tabulated after averaging over models, and similarly for both extremely low and extremely high seasonal precipitation amounts. Values smaller (larger) than 5% indicate a decrease (increase) in the frequency of extremes. This follows the approach in Weisheimer annd Palmer (2005) except that we compare each model's future with its own 20th century to help avoid distortions due to differing biases in the different models. The results are shown in Table 11.1 only when 14 out of the 21 models are in agreement as to the sign of the change in frequency of extremes. For example, in Central North America (CNA), 15% of the summers in 2080–2099 in the A1B scenario are projected to be extremely dry, corresponding to a factor of 3 increase in the frequency of these events. In contrast, in many regions and seasons, the frequency of extreme warmth is 100%, implying that all seasons in 2080–2099 are warmer than the warmest season in 1980-1999, according to every model in this ensemble.

In each continental section a figure is provided summarizing the temperature and precipitation responses in the MMD-A1B projection for the last two decades of the 21st century. These figures portray a multi-model mean comprising individual models or model ensemble means where ensembles exist. Also shown is the simple statistic of the number of these models that show agreement in the sign of the precipitation change. The annual mean temperature and precipitation responses in each of the 21 separate AOGCMs are provided in Supplementary material Figures S11.5–11.12 and S11.13–11.20, respectively.

[INSERT TABLE 11.1 HERE]

Recent explorations of multi-model ensemble projections seek to develop probabilistic estimates of uncertainties and are provided in the Supplementary material Table S11.2. This information is based on the approach of Tebaldi et al. (2004, 2005), see also section 11.10.2.

11.1.3 Some Unifying Themes

The basic pattern of the projected warming as is described in Chapter 10 is little changed from previous assessments. Examining the spread across the MMD models, temperature projections in many regions are strongly correlated with the global mean projections, with the most sensitive models in global mean temperature often the most sensitive locally. Differing treatments of regional processes and the dynamical interactions between a given region and the rest of the climate system are responsible for some spread. However, a substantial part of the spread in regional temperature projections is due to differences in the sum of the feedbacks that control transient climate sensitivity (see also Chapter 10).

The response of the hydrological cycle is controlled in part by fundamental consequences of warmer temperatures and the increase in water vapor in the atmosphere (Chapter 3). Water is transported horizontally by the atmosphere from regions of moisture divergence (particularly in the subtropics) to regions of convergence. Even if the circulation does not change, these transports will increase due to the increase in vapor. We see the consequences of this increased moisture transport in the global response of precipitation described in Chapter 10, where, on average, precipitation increases in the intertropical convergence zones, decreases in the subtropics, and increases in sub-polar and polar regions. Over North America and Europe, the pattern of subpolar moistening and subtropical drying dominates the 21st century projections. This pattern is also described in Chapter 9.5.4, which assesses the extent to which this pattern is visible over land during the 20th century in precipitation observations and model simulations. Regions of large uncertainty often lie near the boundaries between these robust moistening and drying regions, with boundaries placed differently by each model.

High resolution model results indicate that in regions with strong orographic forcing, some of these large scale findings can be considerably altered locally. In some cases this may result in changes in the opposite direction to the more general large scale behaviour. In addition, large area and grid box average projections for precipitation are often very different from local changes within the area (Good and Lowe, 2006). These issues demonstrate the inadequacy of inferring the behaviour at fine-scales from that of large-area averages.

Another important theme in the 21st century projections is the poleward expansion of the subtropical highs, and the poleward displacement of the midlatitude westerlies and associated storm tracks. This circulation response is often referred to an enhanced positive phase of the Northern or Southern Annular Mode, or when focusing on the North Atlantic, the positive phase of the North Atlantic Oscillation. In regions without strong orographic forcing, superposition of the tendency towards subtropical drying and poleward expansion of the subtropical highs creates especially robust drying responses on the poleward boundaries of the 5 subtropical oceanic high centers in the South Indian, South Atlantic, South Pacific, North Atlantic and, less robustly, the North Pacific (where a tendency towards El-Niño-like conditions in the Pacific in the models tends to counteract this expansion). Most of the regional projections of strong drying tendencies over land in the 21st century are immediately downstream of these centers (Southwestern Australia, the Western Cape Provinces of South Africa, the central Andes, the Mediterranean, and Mexico). The robustness of this large-scale circulation signal is discussed in Chapter 10, while Chapters 3, 8, and 9 describe the observed poleward shifts in the late 20th century and the ability of models to simulate these shifts.

The retreat of snow and ice cover are important for local climates. The difficulty of quantifying these effects in regions of substantial topographic relief is a significant limitation of global models (see Box 11.3) and is improved with dynamical and statistical downscaling. The drying effect of an earlier spring snowmelt, and, more generally, the earlier reduction in soil moisture (Manabe and Wetherald, 1987) is a continuing theme in discussion of summertime continental climates.

The strong interactions between sea surface temperature gradients and tropical rainfall variability provides an important unifying theme for tropical climates. Models can differ in their projections of small changes in tropical ocean temperature gradients and in the simulation of the potentially large shifts in rainfall that are related to these oceanic changes. Chou and Neelin (2004) provides a guide to some of the complexity involved in diagnosing and evaluating hydrological responses in the tropics. With a few exceptions the spread in projections of hydrological changes is still too large to make strong statements about the future of tropical climates on regional scales (see also Chapter 10 section 10.3). Projected tropical precipitation changes are large in many AOGCMs, so uncertainty as to the regional pattern of these changes should not be taken as evidence that these changes are likely to be small.

Assessments of the regional and sub-regional climate change projections have primarily been based on the GCM projections summarized in Table 11.1 and an analysis of the biases in the GCM simulations, regional downscaling studies available for some regions with either physical or statistical models or both, and reference to plausible physical mechanisms.

To assist the reader in placing the various regional assessments in a global context, Box 11.1 displays many of the detailed assesments documented in the following regional sections. Likewise, an overview of projected changes in various types of extreme weather statistics is summarised in Table 11.2. This table not only contains information from the assessments within this chapter, but also holds information from Chapter 10. Thus the details of the assessment that lead to each individual statements can all be found in either Chapter 10, or the respective regional sections, and links for each statement are identifiable from Table 11.2.

11.2 Africa

Assessment of projected climate changes for Africa:

> All of Africa is very likely to warm during this century. The warming is very likely to be larger than the global, annual-mean warming throughout the continent and in all seasons, with drier subtropical regions warming more than the moister tropics.
>
> Annual rainfall is likely to decrease in much of Mediterranean Africa and Northern Sahara, with the likelihood of a decrease in rainfall increasing as one approaches the Mediterranean coast. Rainfall in southern Africa will likely decrease in much of the winter rainfall region and on western margins There will likely be an increase in annual mean rainfall in East Africa. It is uncertain how rainfall in the Sahel, the Guinean Coast and the Southern Sahara will evolve in this century.
>
> The MMD models have significant systematic errors in and around Africa (excessive rainfall in the south, southward displacement of the Atlantic ITCZ, insufficient upwelling off the West Coast) making it difficult to assess the consequences for climate projections. The absence of realistic variability in the Sahel in most 20th century simulations casts some doubt on the reliability of coupled models in this region. Vegetation feedbacks and feedbacks from dust aerosol production are not included in the global models. Possible future land-surface modification is also not taken into account in the projections. The extent to which current regional models can successfully downscale precipitation over Africa is unclear, and limitations of empirical downscaling results for Africa are not fully understood. There is insufficient information on which to assess possible changes in the spatial distribution and frequency of tropical cyclones impacting Africa.

11.2.1 Key Processes

The bulk of the African continent is tropical or subtropical with the central phenomenon being the seasonal migration of the tropical rain belts. Small shifts in the position of these rain belts result in large local changes in rainfall. There are also regions on the northern and southern boundaries of the continent with winter rainfall regimes governed by the passage of mid-latitude fronts, that are therefore sensitive to a poleward displacement of the storm tracks. This is evident from the correlation between South African rainfall and the Southern Annular Mode (Reason and Rouault, 2005) and between North African rainfall and the North Atlantic Oscillation (Lamb and Peppler, 1987). Troughs penetrating into the tropics from mid-latitudes also influence warm season rainfall, especially in Southern Africa, and can contribute to a sensitivity of warm season rains to a displacement of the circulation (Todd and Washington, 1999). Any change in tropical cyclones distribution and intensity will affect the southeast coastal regions, including Madagascar (Reason and Keibel, 2004).

The factors that determine the Southern boundary of the Sahara and rainfall in the Sahel have attracted special interest because of the extended drought experienced by this region in the 1970’s and 80’s. The field has moved steadily away from explanations for rainfall variations in this region as primarily due to land-use changes and towards explanations based on changes in sea surface temperatures (SSTs). The early SST perturbation AGCM experiments (Palmer, 1986; Rowell, et al., 1995) are reinforced by the results from the most recent models (Giannini et al., 2003; Hoerling et al., 2006; Lu and Delworth, 2005). The north-south inter-hemispheric gradient, with colder Northern Hemisphere oceans conducive to an equatorward shift and/or a reduction in Sahel rainfall, is important. This has created interest in the possibility that aerosol cooling localized in the Northern Hemisphere could dry the Sahel (Rotstayn and Lohmann, 2002). See also Section 9.5.4.3.1. However, temperatures over other oceanic region, including the Mediterranean (Rowell, 2003), are also important.

In Southern Africa, changing SSTs rather than changing land-use patterns are possibly more important in controlling warm season rainfall variability and trends. Evidence has been presented for strong links with Indian Ocean temperatures (Hoerling et al., 2006). The warming of the troposphere over South Africa, possibly a consequence of warming of the Indo-Pacific, has been linked with the increase in days with stable inversion layers over southern Africa (Freiman and Tyson, 2000; Tadross et al., 2005b, 2006) in the late-20th century.

In addition to the importance of ocean temperatures, vegetation patterns help shape the climatic zones throughout much of Africa (e.g., G. Wang and Eltahir, 2000; Paeth and Henre, 2004; Maynard and Royer, 2004a; see also Box 11.4). In the past, land-surface changes have primarily acted as feedbacks generated by the underlying response to SST anomalies and vegetation changes are thought to provide a positive feedback with climate change. This is plausible as recent work suggests that land-surface feedbacks may also play an important role in both intra-seasonal variability and rainy season onset in Southern Africa (New et al., 2003; Tadross et al., 2005ab; Anyah and Semazzi, 2004).

The MMD models prescribe vegetation cover; they would likely respond more strongly to large-scale forcing if they predicted vegetation, especially in semi-arid areas. The possibility of multiple stable modes of African climate due to vegetation/climate interactions has been raised, especially in the context of discussions of the very wet Sahara during the mid-Holocene 6-8 thousand years ago (Foley et al., 2003; Claussen et al., 1999). One implication is that centennial-timescale feedbacks associated with vegetation patterns may have the potential to make climate changes over Africa less reversible.

11.2.2 Skill of Models in Simulating Present and Past Climates

There are biases in the simulations of African climate that are systematic across the MMD models, with 90% of models overestimating precipitation in Southern Africa by more than 20% on average (and in some cases by as much as 80%) over a wide area often extending into equatorial Africa. The temperature biases over land are not considered large enough to directly affect the credibility of the model projections. (See Supplementary Material Figure S11.21 and Table S11.1)

The intertropical convergence zone in the Atlantic is displaced equatorward in nearly all of these AOGCMs. Ocean temperatures are too warm by an average of 1 to 2°C in the Gulf of Guinea and typically by 3°C off the southwest coast in the region of intense upwelling which is clearly too weak in many models. In several of the models there is no West African Monsoon as the summer rains fail to move from the Gulf onto land but most of the models do have a monsoonal climate albeit with some distortion. Moderately realistic interannual variability of SSTs in the Gulf of Guinea and the associated dipolar rainfall variations in the Sahel and the Guinean Coast is, by the criteria of Cook and Vizy (2006), only present in 4 of the 18 models examined. Tennant (2003) describes biases in several AGCMs, such as the equatorward displacement of the midlatitude jet in austral summer, a deficiency that persists in the most recent simulations (Chapter 8).

Despite these deficiencies, atmospheric GCMs can simulate the basic pattern of rainfall trends in the second half of the 20th century if given the observed SST evolution as boundary conditions, as described in the multi-model analysis of Hoerling et al. (2006) and the growing literature on the interannual variability and trends in individual models (e.g., Rowell et al., 1995; Bader and Latif, 2003; Giannini et al., 2003; Kamga et al., 2005; Haarsma et al., 2005; Lu and Delworth, 2005). However, there is less confidence in the ability of AOGCMs to generate interannual variability in the SSTs of the type known to affect African rainfall, as evidenced by the fact that very few AOGCMs produce droughts comparable in magnitude to the Sahel drought of the 1970's and 1980's (Hoerling et al., 2006). There are exceptions, but what distinguishes these from the bulk of the models is not understood.

The very wet Sahara 6–8 thousand years ago is thought to have been a response to the increased summer insolation due to changes in the Earth's orbital configuration. Modelling studies of this response provide background information on the quality of a model's African monsoon, but the processes controlling the response to changing seasonal insolation may be different from those controlling the response to increasing greenhouse gases. The fact that GCMs have difficulty in simulating the full magnitude of the mid-Holocene wet period, especially in the absence of vegetation feedbacks, may indicate a lack of sensitivity to other kinds of forcing (Jolly et al., 1996; Kutzbach et al., 1996).

Regional climate modelling has mostly focused on southern Africa where the models generally improve on the climate simulated by global models. For example, Engelbrecht et al. (2002) and Arnell et al. (2003) both simulate excessive rainfall in parts of southern Africa, reminiscent of the bias in the MMD. Hewitson et al. (2004) and Tadross et al. (2006) note strong sensitivity to the choice of convective parameterisation, and to changes in soil moisture and vegetative cover (Tadross et al. 2005b; New et al., 2003), reinforcing the view (Rowell, et al, 1995) that land-surface feedbacks enhance regional climate sensitivity over Africa's semi-arid regions. Over West Africa the number of RCM investigations is even more limited (Jenkins et al., 2002; Vizy and Cook, 2002). The quality of the 25-year simulation undertaken by Paeth et al. (2005) is encouraging, emphasizing the role of regional SSTs and changes in the land surface in forcing West African rainfall anomalies. Several recent AGCM time-slice simulations focusing on tropical Africa show good simulation of the rainy season (Coppola and Giorgi, 2005; Oouchi et al., 2006; Caminade et al., 2006).

Hewitson and Crane (2005) have developed empirical downscaling for point-scale precipitation at sites spanning the continent, as well as a 0.1° resolution grid over South Africa. The downscaled precipitation forced by reanalysis data provide a close match to the historical climate record, including regions such as the eastern escarpment of the sub-continent that have proven difficult for RCMs.

11.2.3 Climate Projections

11.2.3.1 Mean Temperature

[INSERT FIGURE 11.1 HERE]

The differences in near surface temperature between years 2080–2099 and the years 1980–1999 in MMD-A1B projections, averaged over the West African (WAF), East African (EAF), South African (SAF) and Saharan (SAH) sub-regions are provided in Table 11.1, with the time evolution displayed in Figure 11.1. The Mediterranean coast is discussed together with Southern Europe in Section 11.3. In all 4 regions and in all seasons, the median temperature increase lies between 3°C and 4°C, roughly 1.5 times the global-mean response. Half of the models project warming within about 0.5°C of these median values. The distributions estimated by Tebaldi et al. (2004, 2005) (also Supplementary material Table S11.2) have a very similar half width, but reduce the likelihood of the extreme high limit as compared to the raw quartiles in Table 11.1. There is a strong correlation across these AOGCMs between the global-mean temperature response and the response in Africa. The signal/noise ratio is very large for these 20-year mean temperatures and 10 years is typically adequate to obtain a *clearly discernible* signal, as defined in Section 11.1.2. Regionally-averaged temperatures averaged over the period 1990–2009 are clearly discernible from the 1980–1999 averages.

The upper panels in Figure 11.2 show the geographical structure of the ensemble-mean projected warming for the A1B scenario in more detail. Smaller values of projected warming, near 3°C, are found in equatorial and coastal areas and larger values, above 4°C, in the Western Sahara. The largest temperature responses in North Africa are projected to occur in June-July-August, while the largest responses in Southern Africa occur in September-October-November. But the seasonal structure in the temperature response over Africa is modest as compared to extratropical regions. The basic pattern of projected warming has been robust to changes in models since the TAR, as indicated by comparison with Hulme et al. (2001).

To date there is insufficient evidence from RCMs to modify the large-scale temperature projections from GCMs, although Tadross et al. (2005b) project changes in the A2 scenario for southern Africa which are lower than that in the forcing GCM and near the low end of the spread in the MMD models, likely due to a weaker drying tendency than in most of the global models.

[INSERT FIGURE 11.2 HERE]

11.2.3.2 Mean Precipitation

Figure 11.2 and Table 11.1 illustrate some of the robust aspects of the precipitation response over Africa in the MMD-A1B projections. The fractional changes in annual-mean precipitation in each of 21 models are provided in Supplementary material Figure S11.13. With respect to the most robust features (drying in the Mediterranean and much of Southern Africa, and increases in rainfall in East Africa), there is qualitative agreement with the results in Hulme et al. (2001) and Ruosteenoja et al. (2003) summarizing results from the models available at the time of the TAR.

The large-scale picture is one of drying in much of the subtropics and an increase (or little change) in the tropics, increasing the rainfall gradients. This is a plausible hydrological response to a warmer atmosphere, a consequence of the increase in water vapour and the resulting increase in vapour transport in the atmosphere from regions of moisture divergence to regions of moisture convergence (see Chapter 9 and Section 11.2.1).

The drying along Africa's Mediterranean coast is a component of a larger scale drying pattern surrounding the Mediterranean and is discussed further in the following section on Europe. A 20% drying in the annual mean is typical along the African Mediterranean coast in A1B by the end of the 21st century. Drying is seen throughout the year and is generated by nearly every MMD model. The drying signal in this composite extends into the Northern Sahara, and down the West coast as far as 15°N. The processes involved include increased moisture divergence as well as a systematic poleward shift of the storm tracks affecting the winter rains, with positive feedback from decreasing soil moisture in summer (see Section 11.3).

In Southern Africa a similar set of processes produces drying which is especially robust in the extreme southwest in winter, a manifestation of a much broader scale poleward shift in the circulation across the South Atlantic and Indian oceans. However, the drying is subject to the caveat that strong orographic forcing my result in locally different changes (as discussed in Box 11.3). With the exception of the winter rainfall region in the south west, the robust drying in winter corresponds to the dry season over most of the subcontinent and does not contribute to the bulk of the annual-mean drying. More than half of the annual-mean reduction occurs in the spring and is mirrored in some RCM simulations for this region (see below). To an extent this can be thought of as a delay in the onset of the rainy season. This springtime drying suppresses evaporation, contributing to the springtime maximum in the temperature response.

The increase in rainfall in East Africa, extending into the Horn of Africa, is also robust across the ensemble of models, with 18 of 21 models projecting an increase in the core of this region, east of the Great Lakes. This East African increase was also evident in Hulme et al. (2001) and Ruosteenoja et al. (2003). The Guinean coastal rain belts and the Sahel do not show as robust a response. A straight average across the ensemble results in modest moistening in the Sahel with little change on the Guinean coast. The composite model has a weak drying trend in the Sahel in the 20th century that does not continue in the future projections (Hoerling, et al., 2006; Biasutti and Giannini, 2006), implying that the weak 20th century drying trend in the composite model is unlikely to be forced by greenhouse gases, but is more likely forced by aerosols, as in Rotstayn and Lohmann (2002), or a result of low-frequency internal variability of the climate.

Individual models generate large, but disparate, responses in the Sahel. Two outliers are GFDL/CM2.1, which projects very strong drying in the Sahel and throughout the Sahara, and MIROC3.2_midres which shows a very strong trend towards increased rainfall in the same region (see Supplementary Figure S11.13). Cook and Vizy (2006) find moderately realistic interannual variability in the Gulf of Guinea and Sahel in both models. While the drying in the GFDL model is extreme within the ensemble, it generates a plausible simulation of 20th century Sahel rainfall trends (Held et al., 2005; Hoerling et al., 2006) and an empirical downscaling from AOGCMs (Hewitson and Crane, 2006) shows a similar response (see below). More research is needed to understand the variety of modelled precipitation responses in the Sahel and elsewhere in the tropics. Progress is being made in developing new methodologies for this purpose (e.g., Chou and Neelin, 2004; Lintner and Chiang, 2005; Chou et al., 2006), leading to better appreciation of the sources of model differences. Haarsma et al. (2005) describe a plausible mechanism, associated with increasing land-ocean temperature contrast and decreasing surface pressures over the Sahara, which contributes to the increase in Sahel precipitation with warming in some models.

It has been argued (e.g., Paethe and Hense, 2004) that the partial amelioration of the Sahel drought since the 90's may be a sign of a greenhouse-gas driven increase in rainfall, providing support for those models that moisten the Sahel into the 21st century (e.g., Kamga et al., 2005; Haarsma et al., 2005; Maynard et al., 2002). However, it is premature to take this partial amelioration as evidence of a global warming signature, given the likely influence of internal variability on the inter-hemispheric SST gradients that influence Sahel rainfall, as well as the influence of aerosol variations.

Table 11.1 provides information on the spread of model projected precipitation change in the four African sub-regions. The regions/seasons for which the central half (25–75%) of the projections are uniformly of one sign are: EAF where there is an increase in DJF, MAM, SON, and in the annual mean; SAF where there is a decrease in austral winter and spring; and SAH where there is a decrease in boreal winter and spring. The Tebaldi et al. (2004, 2005) Bayesian estimates (Supplementary Material Table S11.2) do not change this distinction between robust and non-robust regions/seasons. The time required for emergence of a clearly discernible signal in these robust regions/seasons is typically 50–100 years, except in the Sahara where even longer times are required.

Land-use change is a potential contributor to climate change in the 21st century (see also Box 11.4). Taylor et al. (2002) project drying over the Sahel of 4% from 1996 to 2015 due to changing land use, but suggest that the potential exists for this contribution to grow substantially further into the century. Maynard and Royer (2004a) suggest that estimated land-use change scenarios for the mid 21st century would have only a modest compensating effect on the greenhouse-gas induced moistening in their model. In neither of these studies is there a dynamic vegetation model.

Several climate change projections based on RCM simulations are available for southern Africa but are much scarcer for other regions. Tadross et al. (2005b) examine two RCMs, PRECIS and MM5, nested for Southern Africa in a time-slice AGCM based in turn on lower resolution HADCM3 coupled simulations for SRES A2. During the early summer season, October-December, both models predict drying over the tropical western side of the continent, responding to the increase in high pressure systems entering from the west, with MM5 indicating that the drying extends further south and PRECIS further east. The drying in the west continues into late summer, but there are increases in total rainfall towards the east in January and February, a feature barely present in the consensus of the MMD models. Result obtained by downscaling one global model must be assessed in the context of the variety of responses in Southern Africa among the MMD models (Supplementary Material Figure S11.13).

Empirical downscaling has been pursued by Hewitson and Crane (2006) who provide projections for daily precipitation as a function of 6 GCM simulations. The degree of convergence in the downscaled results for SRES A2 near the end of the 21st century suggests more commonality in GCM-projected changes in daily circulation, on which the downscaling is based, than in the GCM precipitation responses. Figure 11.3 shows the response of mean June-July-August (JJA) monthly total precipitation for station locations across Africa. The consensus of these downscaling results shows increased precipitation in east Africa extending into southern Africa, especially in JJA; strong drying in the core Sahel in JJA with some coastal wetting, and moderate wetting in December-February. There is also drying along the Mediterranean coast, and, in most models, drying in the western portion of southern Africa. The downscaling also shows marked local-scale variation in the projected changes, for example, the contrasting changes on the west and east of Madagascar, and on the coastal and inland borders of the Sahel.

[INSERT FIGURE 11.3 HERE]

While this result is generally consistent with the underlying GCMs and the composite MMD projections, there is a tendency for greater Sahel drying than in the underlying GCMs, providing further rationale (alongside the large spread in model responses and poor coupled model performance in simulating droughts of the magnitude observed in the 20th century) for viewing with caution the projection for a modest increase in Sahel rainfall in the ensemble mean of the MMD models.

11.2.3.3 Extremes

Research on changes in extremes specific to Africa, in either models or observations, is limited. A general increase in the intensity of high-rainfall events, associated in part with the increase in atmospheric water vapour, is expected in Africa, as in other regions. Regional modelling and downscaling results (Tadross et al., 2005b) both support an increase in the rainfall intensity in Southern Africa. In regions of mean drying, there is generally a proportionally larger decrease in the number of rain days, indicating compensation between intensity and frequency of rain. In the downscaling results of Hewitson and Crane (2006) and

Tadross et al. (2005b), changes in the median precipitation event magnitude at the station scale do not always mirror the projected changes in seasonal totals.

There is little modelling guidance on possible changes in tropical cyclone impacting the southeast coast of Africa. Thermodynamic arguments for increases in precipitation rates and intensity of tropical storms (see Chapter 10) are applicable to these Indian Ocean storms as for other regions, but changes in frequency and spatial distribution remain uncertain. In a time-slice simulation with a 20 km AGCM, Oouchi et al. (2006) obtain a significant reduction in the frequency of tropical storms in the Indian Ocean.

Using the definition of "extreme seasons" given in Section 11.1.2, the probability of encountering extremely warm, wet and dry seasons, as estimated by the MMD models, is provided in Table 11.1. As in most tropical regions, all seasons are extremely warm by the end of the 21st century, with very high confidence under the A1B scenario. Although the mean precipitation response in West Africa is less robust than in East Africa, the increase in the number of extremely wet seasons is comparable in both, increasing to roughly 20% (so that 1/5 of the seasons are extremely wet, as compared to 1/20 in the control period in the late 20th century). In Southern Africa, the frequency of extremely dry austral winters and springs increases to roughly 20%, while the frequency of extremely wet austral summers doubles in this ensemble of models.

11.3 Europe and the Mediterranean

Assessments of projected climate change for Europe:

> Annual mean temperatures in Europe are likely to increase more than the global mean. The warming in northern Europe is likely to be largest in winter and that in the Mediterranean area largest in summer. The lowest winter temperatures are likely to increase more than average winter temperature in northern Europe, and the highest summer temperatures are likely to increase more than average summer temperature in southern and central Europe.
>
> Annual precipitation is very likely to increase in most of northern Europe and decrease in most of the Mediterranean area. In central Europe, precipitation is likely to increase in winter but decrease in summer. Extremes of daily precipitation will very likely increase in northern Europe. The annual number of precipitation days is very likely to decrease in the Mediterranean area. The risk of summer drought is likely to increase in central Europe and in the Mediterranean area.
>
> Confidence on future changes in windiness is relatively low, but it seems more likely than not that there will be an increase in average and extreme wind speeds in northern Europe.
>
> Snow season is very likely to shorten in all of Europe, and snow depth is likely to decrease in at least most of Europe.

Although many features of the simulated climate change in Europe and the Mediterranean area are qualitatively consistent between models and qualitatively well-understood in physical terms, substantial uncertainties remain. Simulated seasonal-mean temperature changes vary even on the subcontinental scale by a factor of 2–3 among the current generation of AOGCMs. Similarly, while agreeing on a large-scale increase in winter-half-year precipitation in the northern and decrease in summer-half-year precipitation in the southern parts of the area, models disagree on the magnitude and geographical details of precipitation change. These uncertainties reflect the sensitivity of the European climate change to the magnitude of the global warming and the changes in the atmospheric circulation and the Atlantic MOC. Deficiencies in modelling the processes that regulate the local water and energy cycles in Europe also introduce uncertainty, for both the changes in mean conditions and extremes. Finally, the substantial natural variability of European climate is a major uncertainty particularly for short-term climate projections in the area (e.g., Hulme et al., 1999).

11.3.1 Key Processes

In addition to global warming and its direct thermodynamic consequences such as increased water vapour transport from low to high latitudes (Section 11.1.3), several other factors may shape future climate changes in Europe and the Mediterranean area. Variations in the atmospheric circulation influence the European climate both on interannual and longer time scales. Recent examples include the central European heat wave in the summer 2003, characterized by a long period of anticyclonic weather (see Chapter 3, Box 3.5), the severe cyclone-induced flooding in central Europe in August 2002 (see Chapter 3, Box 3.6), and the strong warming of winters in northern Europe from the 1960's to 1990's that was affected by an upward trend in the NAO (Hurrell and van Loon, 1997; Räisänen and Alexandersson, 2003; Scaife et al., 2005). On fine geographical scales, the effects of atmospheric circulation are modified by topograpgy particularly in areas of complex terrain (Fernandez et al., 2003; Bojariu and Giorgi, 2005).

Europe, particularly its northwestern parts, owes its relatively mild climate partly to the northward heat transport by the Atlantic Meridional Overturning Circulation (MOC) (e.g., Stouffer et al., 2006). Most models suggest increased greenhouse-gas concentrations will lead to a weakening of the MOC (see Chapter 10, Section 10.3) which will act to reduce the warming in Europe. However, in the light of our present understanding, it is very unlikely to reverse the warming to cooling (see Section 11.3.3.1).

Local thermodynamic factors also affect the European climate and are potentially important for its future changes. In those parts of Europe that are at present snow-covered in winter, decrease of snow is likely to induce a positive feedback, further amplifying the warming. In the Mediterranean region and at times in central Europe, feedbacks associated with the drying of the soil in summer are important even in the present climate. For example, they acted to exacerbate the heat wave of 2003 (Black et al., 2004; Fink et al., 2004).

11.3.2 Skill of Models in Simulating Present Climate

AOGCMs show a range of performance in simulating the climate in Europe and the Mediterranean area. Simulated temperatures in the MMD models vary on both sides of the observational estimates in summer but are mostly lower than observed in the winter half-year, particularly in northern Europe (Supplementary material Table S11.1). Excluding one model with extremely cold winters in northern Europe, the seasonal area mean temperature biases in the Northern Europe region (NEU) vary from –5°C to 3°C and those in the Southern Europe and Mediterranean region (SEM) from –5°C to 6°C, depending on model and season. The cold bias in northern Europe tends to increase towards northeast, reaching in the ensemble mean –7°C in the northeast of European Russia in winter. This cold bias coincides with a too weak north-south gradient in the winter mean sea level pressure, which implies too weak westerly flow from the Atlantic Ocean to northern Europe in most models (Supplementary Material Figure S11.22).

Biases in simulated precipitation vary substantially with season and location. The average simulated precipitation in NEU exceeds that observed from autumn to spring (Supplementary material Table S11.1), but the interpretation of the difference is complicated by the observational uncertainty associated with the undercatch of, in particular, solid precipitation (e.g., Adam and Lettenmaier, 2003). In summer, most models simulate too little precipitation, particularly in the eastern parts of the area. In SEM, the area and ensemble mean precipitation are close to observations.

RCMs capture the geographical variation of temperature and precipitation in Europe better than global models but tend to simulate too dry and warm conditions in southeastern Europe in summer, both when driven by analysed boundary conditions (Hagemann et al., 2004) and GCM data (e.g., Jacob et al., 2006). Most but not all RCMs also overpredict the interannual variability of summer temperatures in southern and central Europe (Lenderink et al., 2006; Vidale et al., 2006; Jacob et al., 2006). The excessive temperature variability coincides with excessive interannual variability in either shortwave radiation or evaporation, or both (Lenderink et al., 2006). A need for improvement in the modelling of soil, boundary layer and cloud processes is implied. One of the key model parameters may be the depth of the hydrological soil reservoir, which appears to be too small in many RCMs (van den Hurk et al., 2005).

The ability of RCMs to simulate climate extremes in Europe has been addressed in several studies. In the PRUDENCE simulations (Box 11.2), the biases in the tails of the temperature distribution varied substantially between models but were generally larger than the biases in average temperatures (Kjellström et al., 2006). Inspection of the individual models showed similarity between the biases in daily and interannual variability, suggesting that similar mechanisms may be affecting both.

The magnitude of precipitation extremes in RCMs is model-dependent. In a comparison of the PRUDENCE RCMs, Frei et al. (2006) found the area-mean 5-year return values of one-day precipitation in the vicinity of the European Alps to vary by up to a factor of two between the models. However, except for too low extremes in the southern parts of the area in summer, the set of models as a whole showed no systematic tendency to over- or under-estimate the magnitude of the extremes when compared with gridded observations. A similar level of skill has been found in other model verification studies made for European regions (e.g., Booij, 2002; Semmler and Jacob, 2004; Fowler et al., 2005; see also Frei et al., 2003).

Evidence of model skill in simulation of wind extremes is mixed. Weisse et al. (2005) found an RCM to simulate a very realistic wind climate over the North Sea, including the number and intensity of storms, when driven by analysed boundary conditions. However, most PRUDENCE RCMs, while quite realistic over sea, severely underestimate the occurrence of very high wind speeds over land and coastal areas (Rockel and Woth, 2006). Realistic frequencies of high wind speeds were only found in the two models that used a gust parameterization to mimic the large local and temporal variability of near-surface winds over land.

Box 11.2: The PRUDENCE Project

The 'Prediction of Regional scenarios and Uncertainties for Defining European Climate changerisks and Effects – PRUDENCE' project involved over twenty European research groups. The main objectives of the project were to provide dynamically-downscaled high-resolution climate change scenarios for Europe at the end of the 21st century, and to explore the uncertainty in these projections. Four sources of uncertainty were studied: (i) *Sampling uncertainty* due to the fact that model climate is estimated as an average over a finite number (30) of years, (ii) *Regional model uncertainty* due to the fact that regional climate models use different techniques to discretize the equations and to represent sub-grid effects, (iii) *Emission uncertainty* due to choice of IPCCSRES emission scenario, and (iv) *Boundary uncertainty* due to the different boundary conditions obtained from different global climate models.

Each PRUDENCE experiment consisted of a control simulation representing the period 1961–1990 and a future scenario simulation representing 2071–2100. A large fraction of the simulations used the same boundary data (from HadAM3H for the A2 scenario) to provide a detailed understanding of the regional model uncertainty. Some simulations were also made for the B2 scenario, and by using driving data from two other GCMs and from different ensemble members from the same GCM. More details are provided in e.g., Christensen et al. (2006), Déqué et al. (2005) and http://prudence.dmi.dk.

11.3.3 Climate Projections

11.3.3.1 Mean Temperature

The observed evolution of European temperatures in the 20th century, characterised by a warming trend modulated by multidecadal variability, was well within the envelope of the MMD simulations (Figure 11.4).

[INSERT FIGURE 11.4 HERE]

In this century, the warming is projected to continue at a rate somewhat greater than its global mean, with the increase in 20-year mean temperatures becoming clearly discernible (as defined in Section 11.1.2) within a few decades. Under the A1B scenario, the simulated area and annual mean warming from 1980–1999 to 2080–2099 varies from 2.3 to 5.3°C in NEU and from 2.2 to 5.1°C in SEM. The warming in northern Europe is likely to be largest in winter and that in the Mediterranean area largest in summer (Figure 11.5). Seasonal mean temperature changes typically vary by a factor of three among the MMD models (Table 11.1); however the upper end of the range in NEU in DJF is reduced from 8.1°C to 6.7°C when one model with an extreme cold bias in present-day winter climate is excluded. Further details are given in Table 11.1 and Supplementary Material Figures S11.2-4.

[INSERT FIGURE 11.5 HERE]

Although changes in atmospheric circulation have a significant potential to affect temperature in Europe (e.g., Dorn et al., 2003), they are not the main cause of the projected warming (e.g., Rauthe and Paeth, 2004; van Ulden et al, 2006; Stephenson et al., 2006). A regression-based sudy using five of the MMD models (van Ulden and van Oldenborgh, 2006) indicated that in a region comprising mainly of Germany, circulation changes enhanced the warming in most models in winter (due to an increase in westerly flow) and late summer (due to a decrease in westerly flow), but reduced the warming slightly in May and June. However, the circulation contribution to the simulated temperature changes (typically –1°C to 1.5°C depending on model and month) was generally much smaller than the total simulated warming in the late 21st century.

Despite a decrease in the North Atlantic MOC in most models (see Chapter 10, Section 10.3), all the MMD simulations show warming in Great Britain and continental Europe, as other climatic effects of increased greenhouse gases dominate over the changes in ocean circulation. The same holds for earlier increased greenhouse gas simulations except for a very few (Russell and Rind, 1999; Schaeffer et al., 2004) with slight cooling along the northwestern coastlines of Europe but warming over the rest of the continent. The impact of MOC changes depends on the regional details of the change, being largest if ocean convection is suppressed in high latitudes where the sea-ice feedback may amplify atmospheric cooling (Schaeffer et al., 2004). AOGCM sensitivity studies with an artificial shutdown of the MOC and no changes in greenhouse gas concentrations, typically show a 2–4°C annual mean cooling in most of Europe, with larger cooling in the extreme northwestern parts (e.g., Stouffer et al., 2006).

SD studies tend to show a similar large-scale warming as dynamical models but with finer-scale regional details affected by factors such as distance from the coast and altitude (e.g., Benestad, 2005; Hanssen-Bauer et al., 2005). Comparing RCM and SD projections for Norway downscaled from the same GCM, Hanssen-Bauer et al. (2003) found the largest differences between the two approaches in winter and/or spring at locations with frequent temperature inversions in the present climate. A larger warming at these locations in the SD projections was found consistent with increased winter wind speed in the driving GCM and reduced snow cover, both of which suppress formation of ground inversions.

11.3.3.2 Mean Precipitation

AOGCMs indicate a south-north contrast in precipitation changes across Europe, with increases in the north and decreases in the south (Figure 11.5). The annual area-mean change from 1980–1999 to 2080–2099 in the MMD-A1B projections varies from 0 to 16% in NEU and from –4% to –27% in SEM (Table 11.1). The largest increases in northern and central Europe are simulated in winter. In summer, the NEU area mean changes vary in sign between models, although most models simulate increased (decreased) precipitation north (south) of about 55°N. In SEM, the most consistent and in per cent terms largest decreases occur in summer, but the area mean precipitation in the other seasons also decreases in most or all models. More detailed statistics are given in Table 11.1. Increasing evaporation makes the simulated decreases in annual precipitation minus evaporation to extend a few hundred kilometres further north in central Europe than decreasing precipitation (Supplementary Material Figure S11.1).

Both circulation changes and thermodynamic factors appear to affect the simulated seasonal cycle of precipitation changes in Europe. Applying a regression method to five of the MMD simulations, van Ulden and van Oldenborgh (2006) found that in a region comprising mainly of Germany circulation changes played a major role in all seasons. In most models increases in winter precipitation were enhanced by increased westerly winds with decrease in summer precipitation largely due to more easterly and anticyclonic flow. However, differences in the simulated circulation changes between the individual models were accompanied by large differences in precipitation change particularly in summer. The residual precipitation change varied less with season and between models, being generally positive as expected from the increased moisture transport capacity of a warmer atmosphere. In a more detailed study of one model, HadAM3P, Rowell and Jones (2006) showed that decreases in summer precipitation in continental and southeastern Europe were mainly associated with thermodynamic factors. These included reduced relative humidity resulting from larger continental warming compared to surrounding sea areas and reduced soil moisture due mainly to spring warming causing earlier snowmelt. Given the confidence in the warming patterns driving these changes, the reliability of the simulated drying was assessed as being high.

Changes in precipitation may vary substantially on relatively small horizontal scales, particularly in areas of complex topography. Details of these variations are sensitive to changes in the atmospheric circulation, as illustrated in Figure 11.6 for two PRUDENCE simulations that only differ with respect to the driving global model. In one , an increase in westerly flow from the Atlantic Ocean (caused by a large increase in the north-south pressure gradient) is accompanied by increases of up to 70% in annual precipitation over the Scandinavian mountains. In the other, with little change in the average pressure pattern, the increase is in the range of 0–20%. When compared with circulation changes in the more recent MMD simulations, these two cases fall in the opposite ends of the range. Most MMD models suggest an increased north-south pressure gradient across northern Europe, but the change is generally smaller than in the top row of Figure 11.6.

[INSERT FIGURE 11.6 HERE]

SD based projections of precipitation change in Europe tend to support the large-scale picture from dynamical models (e.g., Busuioc et al., 2001; Beckmann and Buishand, 2002; Hanssen-Bauer et al., 2003, 2005; Benestad, 2005; Busuioc et al., 2006), although variations between SD methods and the dependence on the GCM data sets used (see Section 11.10.1.3) make it difficult to draw quantitative conclusions. However, some SD studies have suggested a larger small-scale variability of precipitation changes than indicated by GCM and RCM results, particularly in areas of complex topography (Hellström et al., 2001).

The decrease in precipitation together with enhanced evaporation in spring and early summer is very likely to lead to reduced summer soil moisture in the Mediterranean region and parts of central Europe (Douville et al., 2002; G. Wang, 2005). In northern Europe, where increased precipitation competes with earlier snowmelt and increased evaporation, the MMD models disagree on whether summer soil moisture will increase or decrease (G. Wang, 2005).

11.3.3.3 Temperature Variability and Extremes

Based on both GCM (Giorgi and Bi, 2005; Rowell, 2005, Clark et al., 2006) and RCM simulations (Schär et al., 2004; Vidale et al., 2006), interannual temperature variability is likely to increase in summer in most areas. However, the magnitude of change is uncertain, even in central Europe where the evidence for increased variability is strongest. In some PRUDENCE simulations, interannual summertime temperature variability in central Europe doubled from 1961–1990 to 2071–2100 under the A2 scenario, while others showed almost no change (Vidale et al., 2006). Possible reasons for the increase in temperature variability are reduced soil moisture, which reduces the capability of evaporation to damp temperature variations, and increased land-sea contrast in average summer temperature (Rowell, 2005; Lenderink et al., 2006).

Simulated increases in summertime temperature variability also extend to daily time scales. Kjellström et al. (2006) analysed the PRUDENCE simulations and found the intermodel differences in the simulated temperature change increase towards the extreme ends of the distribution. However, a general increase in summertime daily temperature variability was evident especially in southern and central parts of Europe, with the highest maximum temperatures increasing more than the median daily maximum temperature (Supplementary material Figure S11.23). Similarly, Shkolnik et al. (2006) reported a simulated increase in summertime daily time scale temperature variability in midlatitude western Russia. These RCM results are supported by GCM studies of Hegerl et al. (2004), Meehl and Tebaldi (2004) and Clark et al. (2006).

In contrast with summer, models project reduced temperature variability in most of Europe in winter, both on interannual (Räisänen, 2001; Räisänen et al., 2003; Giorgi et al., 2004; Giorgi and Bi, 2005; Rowell, 2005) and daily time scales (Hegerl et al., 2004; Kjellström et al., 2006). In the PRUDENCE simulations, the lowest winter minimum temperatures increased more than the median minimum temperature especially in eastern, central and northern Europe, although the magnitude of this change was strongly model-dependent (Supplementary Material Figure S11.23). The geographical patterns of the change indicate a feedback from reduced snow cover, with a large warming of the cold extremes where snow retreats but a more moderate warming in the mostly snow-free southwestern Europe (Rowell, 2005; Kjellström et al., 2006).

Along with the overall warming and changes in variability, heat waves are very likely to increase in frequency, intensity and duration (Barnett et al., 2006; Clark et al., 2006; Tebaldi et al., 2006). Conversely, the number of frost days is very likely to decrease (Tebaldi et al., 2006).

11.3.3.4 Precipitation Variability and Extremes

In northern Europe and in central Europe in winter, where time mean precipitation is simulated to increase, high extremes of precipitation are very likely to increase in magnitude and frequency. In the Mediterranean area and in central Europe in summer, where reduced mean precipitation is projected, extreme short-term precipitation may either increase (due to the increased water vapour content of a warmer atmosphere) or decrease (due to a decreased number of precipitation days, which if acting alone would also make heavy precipitation less common). These conclusions are based on several GCM (e.g., Semenov and Bengtsson, 2002; Voss et al. 2002; Hegerl et al. 2004; Wehner, 2004; Kharin and Zwiers, 2005; Tebaldi et al., 2006) and RCM (e.g., Jones and Reid, 2001; Räisänen and Joelsson, 2001; Booij, 2002; Christensen and Christensen, 2003, 2004; Pal et al., 2004; Räisänen et al., 2004; Ekström et al., 2005; Beniston et al., 2006; Frei et al., 2006; Gao et al., 2006a; Shkolnik et al., 2006) studies. However, there is still a lot of quantitative uncertainty in the changes of both mean and extreme precipitation.

Time scale also matters. Although there are some indications of increased interannual variability particularly in summer precipitation (Räisänen, 2002; Giorgi and Bi, 2005; Rowell, 2005), changes in the magnitude of long-term (monthly to annual) extremes are expected to follow the changes in mean precipitation more closely than those in short-term extremes (Räisänen, 2005). On the other hand, changes in the frequency of extremes tend to increase with increasing time scale even when this is not the case for the changes in the magnitude of extremes (Barnett et al., 2006).

An illustration of the possible characteristics of precipitation change is given in Figure 11.7. The eight models in this PRUDENCE study (Frei et al., 2006) projected an increase in mean precipitation in winter both in southern Scandinavia and central Europe, due to both increased wet day frequency and increased mean precipitation for the wet days. In summer, a decrease in the number of wet days led to a decrease in mean precipitation particularly in central Europe. Changes in extreme short-term precipitation were broadly similar to the change in average wet-day precipitation in winter. In summer, extreme daily precipitation increased in most models despite the decrease in mean precipitation, although the magnitude of the change was highly model-dependent. However, this study only covered the uncertainty associated with the choice of the RCM, not those associated with the driving GCM and the emissions scenario.

[INSERT FIGURE 11.7 HERE]

Much larger changes are expected in the recurrence frequency of precipitation extremes than in the magnitude of extremes (Huntingford et al., 2003; Barnett et al., 2006; Frei et al., 2006). For example, Frei et al. (2006) estimated that, in Scandinavia under the A2 scenario, the highest 5-day winter precipitation totals occurring once in 5 years in 2071–2100 would be similar to those presently occurring once in 8–18 years (the range reflects variation between the PRUDENCE models). In the MMD simulations, large increases occur in the frequencies of both high winter precipitation in northern Europe and low summer precipitation in southern Europe and the Mediterranean area (Table 11.1).

The risk of drought is likely to increase in southern and central Europe. Several model studies have indicated a decrease in the number of precipitation days (e.g., Semenov and Bengtsson, 2002; Voss et al., 2002; Räisänen et al., 2003; 2004; Frei et al., 2006) and an increase in the length of the longest dry spells in this area (Voss et al., 2002; Pal et al., 2004; Beniston et al., 2006; Gao et al., 2006a; Tebaldi et al., 2006). By contrast, the same studies do not suggest major changes in dry spell length in northern Europe.

11.3.3.5 Wind Speed

Confidence on future changes in windiness in Europe remains relatively low. Several model studies (e.g., Zwiers and Kharin, 1998; Knippertz et al., 2000; Leckebusch and Ulbrich, 2004; Pryor et al., 2005a; van den

Hurk et al., 2006) have suggested increased average and/or extreme wind speeds in northern and/or central Europe, but some studies point toward the opposite direction (e.g., Pryor et al., 2005b). The changes in both average and extreme wind speeds may be seasonally variable, but the details of this variation appear to be model-dependent (e.g., Räisänen et al., 2004; Rockel and Woth, 2006).

A key factor is the change in the large-scale atmospheric circulation (Räisänen et al., 2004; Leckebusch et al., 2006). Simulations with increased north-south pressure gradient across northern Europe (e.g., top of Figure 11.6) tend to indicate stronger winds in northern Europe, both because of the larger time-averaged pressure gradient and a northward shift in cyclone activity. Conversely, the northward shift in cyclone activity tends to reduce windiness in the Mediterranean area. On the other hand, simulations with little change in the pressure pattern tend to show only small changes in the mean wind speed (bottom of Figure 11.6). Most of the MMD projected pressure changes fall between the two PRUDENCE simulations in Figure 11.6, which suggests that the most likely outcome for windiness might be between these two cases.

Extreme wind speeds in Europe are mostly associated with strong winter cyclones (e.g., Leckebush and Ullbrich, 2004), the occurrence of which is only indirectly related to the time mean circulation. Nevertheless, models suggest a general similarity between the changes in average and extreme wind speeds (Knippertz et al., 2000; Räisänen et al., 2004). A caveat to this conclusion is that, even in most RCMs, the extremes of wind speed over land tend to be too low (see Section 11.3.2).

11.3.3.6 Mediterranean Cyclones

Several studies have suggested a decrease in the total number of cyclones in the Mediterranean Sea (Lionello et al., 2002; Vérant, 2004; Somot 2005; Leckebusch et al., 2006; Pinto et al., 2006; Ulbrich et al., 2006), but there is no agreement on whether the number of intense cyclones will increase or decrease (Lionello et al., 2002; Pinto et al., 2006).

11.3.3.7 Snow and Sea-Ice

The overall warming is very likely to shorten the snow season in all of Europe. Snow depth will also likely be reduced at least in most areas, although increases in total winter precipitation may counteract the increased melting and decreased fraction of solid precipitation associated with the warming. The changes may be large, including potentially a one-to-three month shortening of the snow season in northern Europe (Räisänen et al., 2003) and a 50–100% decrease in snow depth in most of Europe (Räisänen et al., 2003; Rowell, 2005) by the late 21st century. However, snow conditions in the coldest parts of Europe, such as northern Scandinavia and northwestern Russia (Räisänen et al., 2003; Shkolnik et al., 2006) and the highest peaks of the Alps (Beniston et al., 2003) appear to be less sensitive to the temperature and precipitation changes projected for this century than those at lower latitudes and altitudes (see also Box 11.3).

The Baltic Sea is likely to lose a large part of its seasonal ice cover during this century. Using a regional atmosphere-Baltic Sea model (Meier et al., 2004), the average winter maximum ice extent decreased by about 70% (60%) from 1961–1990 to 2071–2100 under the A2 (B2) scenario. The length of the ice season was projected to decrease by 1–2 months in northern parts and 2–3 months in the central parts. Comparable decreases in Baltic Sea ice cover were projected in earlier studies (Haapala et al., 2001; Meier, 2002).

11.4 Asia

Assessment of projected climate change for Asia:

> All of Asia is very likely to warm during this century; the warming is likely to be well above the global mean in Central Asia, Tibetan Plateau and Northern Asia, above in East Asia and South Asia, and similar in Southeast Asia. It is very likely that summer heat waves / hot spells in East Asia will be of longer duration, more intense, and more frequent. Fewer very cold days are very likely in East Asia and South Asia.

Winter precipitation will very likely increase in Northern Asia and the Tibetan Plateau, and likely in Eastern Asia and the southern parts of Southeast Asia. Summer precipitation will likely increase in Northern Asia, East Asia, South Asia and most of Southeast Asia, but it will likely decrease in Central Asia. There will be very likely an increase in frequency of intense precipitation events in parts of South Asia, and in East Asia.

Extreme rainfall and winds associated with tropical cyclones are likely to increase in East Asia, Southeast Asia and South Asia. Monsoonal flows and the tropical large-scale circulation is likely to be weaken.

While broad aspects of Asian climate change show consistency between AOGCM simulations, there remain a number of sources of uncertainty. A lack of observational data in some areas limits model assessment. There has been little assessment of the projected changes in regional climatic means and extremes. There are substantial inter-model differences in representing monsoon processes and a lack of clarity over changes in ENSO further contributes to uncertainty about future regional monsoon and tropical cyclone behaviour. Consequently, quantitative estimates of projected precipitation change are difficult. It is likely that some local climate changes will vary significantly from regional trends due to the region's very complex topography and marine influences.

11.4.1 Key Processes

As monsoons are the dominant phenomena over much of Asia, the factors that influence the monsoonal flow and precipitation are of central importance for understanding climate change in this region. Precipitation is affected both by the strength of the monsoonal flows and the amount of water vapour transported. Monsoonal flows and the tropical large-scale circulation often weaken in global warming simulations (e.g., Knutson and Manabe, 1995). This arises out of an increase in dry static stability associated with the tropical warming in these models, and the reduction in adiabatic warming/cooling needed to balance a given amount of radiative cooling/condensational heating (e.g., Betts, 1998). But there is an emerging consensus that the effect of enhanced moisture convergence in a warmer moister atmosphere dominates over any such weakening of the circulation, resulting in increased monsoonal precipitation (e.g., Douville et al., 2000; Giorgi et al., 2001a, b; Stephenson et al., 2001; Dairaku and Emori, 2006; Ueda et al., 2006).

There is an association of the phase of ENSO with the strength of the summer monsoons (Pant and Rupa Kumar, 1997), so changes in ENSO will have an impact on these monsoons. However, such an impact can be compounded by a change in the ENSO / South Asian monsoon connection under greenhouse-gas warming (Krishna Kumar et al., 1999; Ashrit et al., 2001; Sarkar et al., 2004; see Chapter 3, Section 3.7). Moreover, there is a link between Eurasian snow cover and the strength of the monsoon (see also, Chapter 3, Section 3.7), with the monsoon strengthening if snow-cover retreats. Aerosols, particularly absorbing aerosols, further modify monsoonal precipitation (e.g., Ramanathan et al., 2005 for South Asia), as do modifications of vegetation cover (e.g., Chen et al., 2004 for East Asia). However, most emission scenarios suggest that future changes in regional climate are still likely to be dominated by increasing greenhouse gas forcing rather than changes in sulphate and absorbing aerosols, at least over the South Asian region.

For South Asia, the monsoon depressions and tropical cyclones generated over the Indian seas modulate the monsoon anomalies. For East Asia, the monsoonal circulations are strengthened by extratropical cyclones energized in the lee of the Tibetan plateau and by the strong temperature gradient along the East Coast. ENSO's influence on the position and strength of the subtropical high pressure in the North Pacific influences both typhoons and other damaging heavy rainfall events, and has been implicated in observed interdecadal variations in typhoon tracks (Ho et al., 2004). This suggests that the spatial structure of warming in the Pacific will be relevant for changes in these features. The dynamics of the Meiyu-Changma-Baiu rains in the early summer, which derive from baroclinic disturbances strongly modified by latent heat release remain poorly understood. While an increase in rainfall in the absence of circulation shifts is expected, relatively modest shifts or changes in timing can significantly affect East Chinese, Korean and Japanese climates.

Over central and Southeast Asia and the maritime continent, interannual rainfall variability is significantly affected by ENSO (e.g., McBride et al., 2003), particularly the June to November rainfall in southern and eastern parts of the Indonesian Archipelago, which is reduced in El Niño years (Aldrian and Susanto, 2003). Consequently the pattern of ocean temperature change across the Pacific is of central importance to climate change in this region.

In Central Asia, including the Tibetan Plateau, the temperature response is strongly influenced by changes in winter and spring snow cover, the isolation from maritime influences, and the spread of the larger wintertime Arctic warming into the region. With regard to precipitation, a key issue is related to the moisture transport from the northwest by westerlies and polar fronts. How far the projected drying of the neighbouring Mediterranean penetrates into these regions is likely to be strongly dependent on accurate simulation of these moisture transport processes. The dynamics of climate change in the Tibetan Plateau are further complicated by the high altitude of this region and its complex topography with large elevation differences.

11.4.2 Skill of Models in Simulating Present Climate

Regional-mean temperature and precipitation in the MMD models show biases when compared with observed climate (Supplementary material Table S11.1). The multiple model mean shows a cold and wet bias in all regions and in most seasons, and the bias of the annual-average temperature ranges from –2.5ºC over the Tibetan Plateau (TIB) to –1.4ºC over South Asia (SAS). For most regions there is a 6–7ºC range in the biases from individual models with a reduced bias range in Southeast Asia (SEA) of 3.6ºC. The median bias in precipitation is small (less than 10%) in Southeast Asia, South Asia, and Central Asia (CAS), larger in Northern Asia and East Asia (NAS and EAS, around +23%), and very large in the Tibetan Plateau (+110%). Annual biases in individual models are in the range of –50% to +60% across all regions except the Tibetan Plateau where some models show annual precipitation 2.5 times the observed and even larger seasonal biases occur in winter and spring. These global models clearly have significant problems over Tibet, due to the difficulty in simulating the effects of the dramatic topographic relief, as well as the distorted albedo feedbacks due to extensive snow cover. However, with only limited observations available, predominantly in valleys, large errors in temperature and significant underestimates of precipitation are likely.

South Asia

Over South Asia, the summer is dominated by the southwest monsoon, which spans the four months from June to September, and dominates the seasonal cycles of the climatic parameters. While most models simulate the general migration of seasonal tropical rain, the observed maximum rainfall during the monsoon season along the west coast of India, the north Bay of Bengal and adjoining northeast India is poorly simulated in many models (Lal and Harasawa, 2001; Rupa Kumar and Ashrit, 2001; Rupa Kumar et al., 2002, 2003). This is likely linked to the coarse resolution of the models as the heavy rainfall over these regions is generally associated with the steep orography. However, the simulated annual cycles in South Asian mean precipitation and surface air temperature are reasonably close to the observed (Supplementary material Figure S11.24). The MMD models capture the general regional features of the monsoon, such as the low rainfall amounts coupled with high variability over northwest India. However, there has not yet been sufficient analysis of whether finer details of regional significance are simulated more adequately in the AR4 models.

Recent work indicates that time-slice experiments using and AGCM with prescribed SSTs, as opposed to a fully coupled system, are not able to accurately capture the South Asian monsoon response (Douville, 2005). Thus, neglecting the short-term SST feedback and variability seems to have a significant impact on the projected monsoon response to global warming, complicating the regional downscaling problem. However, May (2004a) notes that the high resolution ECHAM4 GCM (T106) simulates the variability and extremes of daily rainfall (intensity as well as frequency of wet days) in good agreement with the observations (GPCP, Huffman et al., 2001).

Three-member ensembles of baseline simulations (1961–1990) from a regional climate model (PRECIS) at 50 km resolution have confirmed that significant improvements in the representation of regional processes over South Asia can be achieved (Rupa Kumar et al., 2006). For example, the steep gradients in monsoon precipitation with a maximum along the western coast of India are well represented in PRECIS.

East Asia

Simulated temperatures in most MMD models are too low in all seasons over East Asia; the mean cold bias is largest in winter and smallest in summer. Zhou and Yu (2006) show that over China, the models perform reasonably in simulating the dominant variations of the mean temperature over China, but not the spatial distributions. The annual precipitation over East Asia exceeds the observed estimates in almost all models and the rain band in mid-latitudes is shifted northward in seasons other than summer. This bias in the placement of the rains in Central China also occurred in earlier models (e.g., Zhou and Li, 2002; Gao et al., 2004). In winter, the area-mean precipitation is overestimated by over 50% on average due to strengthening of the rain band associated with extratropical systems over South China. The bias and inter-model differences in precipitation are smallest in summer but the northward shift of this rain band results in large discrepancies in summer rainfall distribution over Korea, Japan and adjacent seas.

Kusunoki et al. (2006) find that the simulation of the Meiyu-Changma-Baiu rains in the East Asian Monsoon is improved substantially with increasing horizontal resolution. Confirming the importance of resolution, RCMs simulate more realistic climatic characteristics over East Asia than AOGCMs, either driven by re-analysis or by AOGCMs (e.g., Ding et al., 2003; Oh et al., 2004; Fu et al., 2005; Ding et al., 2006; Sasaki et al., 2006b). Several studies reproduce the fine-scale climatology of small areas using a multiply nested RCM (Im et al., 2006) and a very high resolution (5 km) RCM (Yasunaga et al., 2006). Gao et al. (2006b) reported that simulated East Asia large-scale precipitation patterns are significantly affected by resolution, particularly during the mid- to late-monsoon months, when smaller scale convective processes dominate.

Southeast Asia

The broad-scale spatial distribution of temperature and precipitation in DJF and JJA averaged across the MMD models compares well with observations. Rajendran et al. (2004) examined current climate simulation in the MRI coupled model. Large-scale features were well simulated, but errors in the timing of peak rainfall over Indochina were considered a major shortcoming. Collier et al. (2004) assessed the performance of the CCM3 model in simulating tropical precipitation forced by observed SST. Simulation was good over the Maritime continent compared to the simulation for other tropical regions. B.Wang et al. (2004) assessed the ability of eleven AGCMs in the Asian-Australian monsoon region simulation forced with observed SST variations. They found that the models' ability to simulate observed interannual rainfall variations was poorest in the Southeast Asian portion of the domain. Since current AOGCMs continue to have some significant shortcomings in representing ENSO variability (see Chapter 8, Section 8.4), the difficulty of projecting changes in ENSO-related rainfall in this region are compounded.

Rainfall simulation across the region at finer scales has been examined in some studies. The CSIRO stretched-grid model CCAM at 80-km resolution show reasonable precipitation simulation in JJA, although Indochina tended to be drier than in the observations (McGregor and Nguyen, 2003). Aldrian et al. (2004a) have conducted a number of simulations with the MPI regional model for an Indonesian domain, forced by reanalyses and by ECHAM4 GCM. The model was able to represent the spatial pattern of seasonal rainfall. It was found that a resolution of at least 50 km was required to simulate rainfall seasonality correctly over Sulawesi. The formulation of a coupled regional model improves regional rainfall simulation over the oceans (Aldrian et al., 2004b). Arakawa and Kitoh (2005) have demonstrated an accurate simulation of the diurnal cycle of rainfall over Indonesia in an AGCM of 20-km horizontal resolution.

Central Asia and Tibet

Due to the complex topography and the associated mesoscale weather systems of the high altitude and arid areas, GCMs typically perform poorly over the region. Importantly, the GCMs, and to a lesser extent RCMs, tend to overestimate the precipitation over arid and semi arid areas in the north (e.g., Small et al., 1999; Gao et al., 2001; Elguindi and Giorgi, 2006).

Over Tibet, the few available RCM simulations generally exhibit improved performance in the simulation of present-day climate compared to GCMs (e.g., Gao et al., 2003a, b; Zhang et al., 2005a). For example, the GCM simulation of Gao et al. (2003a) overestimated the precipitation over the northwestern Tibetan Plateau by a factor of 5–6, while in an RCM nested in this model, the overestimate was less than a factor of 2.

11.4.3 Climate Projections

11.4.3.1 Temperature

The temperature projections for the 21st century based on MMD-A1B models (Figure 11.8 and Table 11.1) represent a strong warming over the 20th century. Warming is similar to the global mean warming in Southeast Asia (mean warming from 1980–1999 to 2080–2099, 2.5°C). Warming greater than the global mean is projected for South Asia (3.3°C) and East Asia (3.3°C), and much more than the global mean in the continental interior of Asia (3.7°C in Central Asia, 3.8°C in Tibet and 4.3°C in Northern Asia). In four out of the six regions, the largest warming occurs in DJF, but in Central Asia the maximum occurs in JJA. In Southeast Asia, the warming is nearly the same throughout the year. Model to model variation in projected warming is typically about three quarters of the mean warming (e.g., 2.0–4.7°C for annual-mean warming in South Asia). The 5–95% ranges based on Tebaldi et al. (2005) suggest a slightly smaller uncertainty than the full range of the model results (Supplementary material Table S11.2). Because the projected warming is large compared to the interannual temperature variability, a large majority, of individual years and seasons in the late 21st century are likely to be extremely warm by present standards (Table 11.1). The projections for changes in mean temperature and, where available, temperature extremes, are discussed below in more detail for individual Asian regions.

[INSERT FIGURE 11.8 HERE]

South Asia
For the A1B scenario, the MMD-A1B models show a median increase of 3.3ºC (see Table 11.1) in annual-mean temperature by the end of the 21st century. The median warming varies seasonally from 2.7ºC in JJA to 3.6ºC in DJF, and is likely to increase northward in the area, particularly in winter, and from sea to land (Figure 11.9). Studies based on earlier AOGCM simulations (Douville et al., 2000; Lal and Harasawa, 2001; Lal et al., 2001; Rupa Kumar and Ashrit, 2001; Rupa Kumar et al., 2002, 2003; Ashrit et al., 2003; May, 2004b) support this picture. The tendency of the warming to be more pronounced in winter is also a conspicuous feature of the observed temperature trends over India (Rupa Kumar et al., 2002, 2003).

[INSERT FIGURE 11.9 HERE]

Downscaled projections using the regional climate model HadRM2 indicate future increases in extreme daily maximum and minimum temperatures all over South Asia due to increase in greenhouse-gas concentrations. This projected increase is of the order of 2–4°C in the mid 21st century under the IS92a scenario both in minimum and maximum temperatures (Krishna Kumar et al., 2003). Results from a more recent regional climate model PRECIS indicate that the night temperatures increase faster than the day temperatures, with the implication that cold extremes are very likely to be less severe in the future (Rupa Kumar et al., 2006).

East Asia
The MMD-A1B models indicate a median warming of 3.3ºC (Table 11.1) by the end of the 21st century, which varies seasonally from 3.0ºC in JJA to 3.6ºC in DJF. The warming tends to be largest in winter, especially in the northern inland area (Figure 11.9) but the area-mean difference from the other seasons is not large. There is no obvious relationship between model bias and the magnitude of the warming. The spatial pattern of larger warming over northwest EAS (Figure 11.9) is very similar to the ensemble-mean of the earlier models. RCM simulations show mean temperature increases similar to that in AOGCMs (Gao et al., 2001, 2002; Kwon et al., 2003; Jiang, 2005; Kurihara et al., 2005; Y.L.Xu et al., 2005).

Daily maximum and minimum temperatures will very likely increase in East Asia, resulting in more severe warm but less severe cold extremes (Gao et al., 2002; Mizuta et al., 2005; Y.L.Xu et al., 2005; Boo et al., 2006). Mizuta et al. (2005) analyzed temperature-based extreme indices over Japan with a 20-km mesh AGCM and found the changes in the indices to be basically those expected from the mean temperature increase, with changes in the distribution around the mean playing no large role. Boo et al. (2005) reported similar results for Korea. Gao et al. (2002) and Y.L.Xu et al. (2005) found reduced diurnal temperature range in China and larger increases in daily minimum than maximum temperatures.

Southeast Asia
For MMD-A1B, the median warming for the region is 2.5ºC by the end of the 21st century, and with little seasonal variation (Table 11.1). Simulations by the DARLAM regional model (McGregor et al., 1998) and more recently by the CSIRO stretched-grid model (McGregor and Dix, 2001) centred on the Indochina Peninsula (AIACC 2004, at a resolution of 14 km) have demonstrated the potential for significant local variation in warming, particularly the tendency for warming to be significantly stronger over the interior of the landmasses than over the surrounding coastal regions. A tendency for the warming to be stronger over Indochina and the larger landmasses of the archipelago is also visible in the MMD models (Chapter 10, Figure 10.8 and Figure 11.9). As in other regions, the magnitude of the warming depends on the forcing scenario.

Central Asia and *Tibet*
For MMD-A1B Central Asia warms with a median of 3.7ºC, and Tibet by 3.8ºC (Table 11.1) by the end of the 21st century.. The seasonal variation in the simulated warming is modest. Findings from earlier multi-model studies (Zhao et al., 2002; Y.Xu et al. 2003a, b; Meleshko et al., 2004; Y. Xu et al., 2005) are consistent with the MMD models' results.

An RCM study by Gao et al. (2003b) indicated greater warming over the Plateau compared to surrounding areas, with the largest warming at highest altitudes, e.g., over the Himalayas (see also Box 11.3). The higher temperature increase over high-altitude areas can be explained by the decrease in surface albedo associated with the melting of snow and ice (Giorgi et al., 1997). This phenomenon is found to different extents in some although not all of the MMD models, and it is visible in the multi-model mean changes particularly in the winter (Figure 11.9).

11.4.3.2 Precipitation and Associated Circulation Systems

The consensus of MMD models indicates an increase in annual precipitation in most of Asia during this century, the percent increase being largest and most consistent between models in North and East Asia (Figure 11.9, Table 11.1). The main exception is Central Asia, particularly its western parts, where most models simulate reduced precipitation in the summer. Based on these simulations, sub-continental mean winter precipitation will very likely increase in Northern Asia and the Tibetan Plateau and likely increase in Central, Southeast and East Asia. Summer precipitation will likely increase in North, South, Southeast, and East Asia, but decrease in Central Asia. Probability estimates from Tebaldi et al. (2005) (Supplementary Material Table S11.2) support these judgments.

The projected decrease in mean precipitation in Central Asia is accompanied by an increase in the frequency of very dry spring, summer and autumn seasons; conversely, in winter where models project increases in the mean precipitation, very high precipitation becomes more common (Table 11.1). The projections for changes in mean precipitation and, where available, precipitation extremes, are discussed in more detail below for individual Asian regions. Where appropriate, the connection to changes in precipitation-bringing circulation systems is also discussed. Smaller (slightly larger) changes are generally projected for the B1 (A2) scenario, but the inter-scenario differences are small compared with the inter-model differences.

South Asia
Most of the MMD-A1B models project a decrease of precipitation in DJF (the dry season), and an increase during the rest of the year. The median change is 11% by the end of the 21st century, and seasonally is –5% in DJF and 11% in JJA., with a large inter-model spread (Table 11.1). The probabilistic method of Tebaldi et al. (2005) similarly shows a large spread, although only 3 of the 21 models project a decrease in annual precipitation. This qualitative agreement on increasing precipitation for most of the year is also supported by earlier AOGCM simulations (Lal and Harasawa, 2001; Lal et al., 2001; Rupa Kumar and Ashrit, 2001; Rupa Kumar et al., 2002, 2003; Ashrit et al., 2003; May, 2004b).

In a study with four GCMs, Douville et al. (2000) found a significant spread in the summer monsoon precipitation anomalies despite a general weakening of the monsoon circulation (see also May, 2004b). They concluded that the changes in atmospheric water content, precipitation and land-surface hydrology under greenhouse forcing could be more important than the increase in the land-sea thermal gradient for the future evolution of monsoon precipitation. Stephenson et al. (2001) proposed that the consequences of climate change could manifest in different ways in the physical and dynamical components of monsoon circulation. Douville et al. (2000) also argue that the weakening of ENSO-monsoon correlation could be explained by a possible increase in precipitable water as a result of global warming, rather than by an increased land-sea thermal gradient. However, model diagnostics using ECHAM4 to investigate this aspect indicate that both the above mechanisms can play a role in monsoon changes in a greenhouse-gas warming scenario. Ashrit et al. (2001) showed that the monsoon deficiency due to El Niño might not be as severe while the favourable impact of La Niña seems to remain unchanged. In a later study using the CNRM GCM, Ashrit et al. (2003) found that the simulated ENSO-monsoon teleconnection shows a strong modulation on multi-decadal time scales, but no systematic change with increasing amounts of greenhouse gases.

ECHAM4 time-slice experiments indicate a general increase in the intensity of heavy rainfall events in the future, with large increases over the Arabian Sea and the tropical Indian Ocean, in northern Pakistan and northwest India, as well as in northeast India, Bangladesh and Myanmar (May, 2004a). The regional climate model HadRM2 shows an overall decrease in the annual number of rainy days by up to 15 days over a large part of South Asia, under IS92a scenario in the 2050s, but with an increase in the precipitation intensity as well as extreme precipitation (Krishna Kumar et al., 2003). Simulations with the PRECIS RCM also projects substantial increases in extreme precipitation over a large area, particularly over the west coast of India and west central India (Rupa Kumar et al., 2006). Dairaku and Emori (2006) showed from a T106 AGCM simulation that the increased extreme precipitation over land in South Asia would arise mainly from dynamic effect, i.e., enhanced upward motion due to the northward shift of monsoon circulation.

Based on regional HadRM2 simulations, Unnikrishnan et al. (2006) reported increases in frequency as well as intensities of tropical cyclones in the 2050s under IS92a scenario in the Bay of Bengal, which will cause more heavy precipitation in the surrounding coastal regions of South Asia, during both southwest and northeast monsoon seasons.

East Asia

The MMD-A1B models project an increase in precipitation in East Asia in all seasons. The median change at the end of the 21st century is +9% in the annual mean with little seasonal difference, and a large model spread in DJF (Table 11.1). In winter this increase contrasts with a decrease in precipitation over the ocean to the southeast, where reduced precipitation corresponds well with increased mean sea level pressure. While the projections have good qualitative agreement, there remain large quantitative differences between the models which is consistent with previous studies (e.g., Giorgi et al., 2001a; Hu et al., 2003; Min et al., 2004).

Based on MMD models, Kimoto (2005) projects increased Meiyu-Changma-Baiu activity associated with the strengthening of anticyclonic cells to its south and north, and Kwon et al. (2005) shows an increased East Asia summer precipitation due to an enhanced monsoon circulation in the decaying phase of El Niño. A 20-km mesh AGCM simulation shows that Meiyu-Changma-Baiu rainfall increases over the Yangtze River valley, the East China Sea and western Japan, while rainfall decreases to the north of these areas mostly due to the lengthening of the Meiyu-Changma-Baiu (Kusunoki et al., 2006). Simulation by RCMs supports the results by AOGCMs. For example, Kurihara et al. (2005) shows increase of precipitation over western Japan in summer.

Kitoh and Uchiyama (2006) investigated the onset and withdrawal times of the Asian summer rainfall season in 15 MMD simulations (Figure 11.10). They found a delay in early summer rain withdrawal over the region extending from Taiwan to Ryukyu Islands to the south of Japan, but an earlier withdrawal over the Yangtze Basin, although the latter is not significant due to large inter-model variation. Changes in onset dates are smaller.

[INSERT FIGURE 11.10 HERE]

Yasunaga et al. (2006) used a 5-km mesh cloud-resolving RCM, to investigate summer rainfall in Japan. They found no changes in rainfall in June but increased rainfall in July in a warmer climate. The increase in July can be attributed to the more frequent large-precipitation systems.

Intense precipitation events will very likely increase in East Asia, consistent with the historical trend in this region (Fujibé et al., 2005; Zhai et al., 2005). Kanada et al. (2005) showed, using a 5-km RCM, that the confluence of disturbances from the Chinese Continent and from the East China Sea would often cause extremely heavy precipitation over Kyushu Island of Japan in July in a warmer climate. An increase in the frequency and intensity of heavy precipitation events also occurs in Korea in the long RCM simulation of Boo et al. (2006). Similarly based on RCM simulations, Y.L. Xu et al. (2005) reported more extreme precipitation events over China. Gao et al. (2002) found a simulated increase in the number of rainy days in Northwest China, and a decrease of rain days but an increase of heavy rainy days over South China. Kitoh et al. (2005) reported similar result in South China by AOGCM.

Kimoto et al. (2005) suggests that frequencies of non-precipitating and heavy ($\geq$30 mm day^{-1}) rainfall days would increase significantly at the expense of relatively weak (1–20 mm day^{-1}) rainfall days in Japan . Mizuta et al. (2005) find significantly more days with heavy precipitation and stronger average precipitation intensity in western Japan and Hokkaido Island. Hasegawa and Emori (2005) showed that daily precipitation associated with tropical cyclones over western North Pacific would increase.

The previously noted weakening of East Asian winter monsoon (e.g., Hu et al. 2000) is further confirmed by recent studies (e.g., Kimoto, 2005; Hori and Ueda, 2005).

Southeast Asia

Area-mean precipitation over Southeast Asia increases in most MMD models, with a median change of about 7% in all seasons (Table 11.1), but the projected seasonal changes vary strongly within the region. The seasonal confidence intervals based on the methods of Tebaldi et al. (2004, 2005) are similar for DJF and JJA (roughly –4% to 17%). The strongest and most consistent increases broadly follow the ITCZ, lying over northern Indonesia and Indochina in JJA, and over southern Indonesia and Papua New Guinea in DJF (Figure 11.9). Away from the ITCZ, precipitation decrease is often simulated. The pattern is broadly one of wet season rainfall increase and dry season decrease.

Earlier studies of precipitation change in the area have in some cases suggested a worse inter-model agreement than found for the MMD models. Both Giorgi et al. (2001a) and Ruosteenoja et al. (2003) found inconsistency in the simulated direction of precipitation change in the region, but a relatively narrow range of possible changes; similar results were found over an Indonesian domain by Boer and Faqih (2004). Compositing the projections from a range of earlier simulations forced by the IS92a scenario, Hulme and Sheard (1999a, b) found a pattern of rainfall increase across Northern Indonesia and the Philippines, and decrease over the southern Indonesian archipelago. More recently Boer and Faqih (2004) compared patterns of change across Indonesia from five AOGCMS and obtained highly contrasting results. They concluded that ‘no generalisation could be made on the impact of global warming on rainfall’ in the region.

The regional high-resolution simulations of McGregor et al. (1998) and McGregor and Dix (2001; AIACC, 2004) have demonstrated the potential for significant local variation in projected precipitation change. The simulations showed considerable regional detail in the simulated patterns of change, but little consistency across the three simulations. The authors related this result to significant deficiencies in the current-climate simulations of the models for this region.

Rainfall variability will be affected by changes to ENSO and its effect on monsoon variability, but this is not well understood (see Chapter 10, Sections 10.3). However, as Boer and Faqih (2004) noted, those parts of Indonesia that experience mean rainfall decrease are likely to also experience increases in drought risk. It is also likely that the region will share the general tendency for daily extreme precipitation to become more intense under enhanced greenhouse conditions, particularly where the mean precipitation is projected to increase. This has been demonstrated in a range of global and regional studies (see Chapter 10, Section 10.3), but needs explicit study for the Southeast Asian region.

The northern part of the Southeast Asian region will be affected by any change to tropical cyclone characteristics. As noted in Chapter 10, Section 10.3, there is evidence in general of likely increases in tropical cyclone intensity, but less consistency about how occurrence will change (see also Walsh, 2004). The likely increase in intensity (precipitation and winds) has been supported for the NW Pacific (and other regions) by the recent modelling study of Knutson and Tuleya (2004). The high-resolution time-slice modelling experiment of Hasegawa and Emori (2005) also demonstrated an increase in tropical cyclone precipitation in the western North Pacific, but not an increase in tropical cyclone intensity. Wu and Wang (2004) examined possible changes in tracks in the NW Pacific due to changes in steering flow in two GFDL enhanced greenhouse-gas experiments. Tracks moved more northeasterly, possibly reducing tropical cyclone frequency in the Southeast Asian region. Since most of the tropical cyclones form along the monsoon trough and are also influenced by ENSO, changes to occurrence, intensity and characteristics of tropical cyclones and their interannual variability will be affected by changes to ENSO (see Chapter 10, Section 10.3).

Central Asia and Tibet

Precipitation over Central Asia increases in most MMD-A1B projections for DJF but decreases in the other seasons. The median change by the end of the 21st century is an overall –3% in the annual mean, with +4% in DJF and –13% in JJA (the dry season) and (Table 11.1). This seasonal variation in the changes is broadly consistent with the earlier multi-model study of Meleshko et al. (2004), who however found an increase in summer precipitation in the northern part of the area.

Over the Tibetan Plateau, all MMD-A1B models project increased precipitation in DJF (median 19%). Most but not all models also simulate increased precipitation in the other seasons (Table 11.1). Earlier studies both by AOGCMs and RCMs are consistent with these findings (Y.Xu et al., 2003a,b; Gao et al., 2003b; Y. Xu et al., 2005).

Box 11.3: Climatic Change in Mountain Regions

Although mountains differ considerably from one region to another, one common feature is the complexity of their topography. Related characteristics include rapid and systematic changes in climatic parameters, in particular temperature and precipitation, over very short distances (Becker and Bugmann, 1997); greatly enhanced direct runoff and erosion; systematic variation of other climatic (e.g., radiation) and environmental factors, such as soil types. In some mountain regions, it has been shown that there is an elevation dependence on temperature trends and anomalies (Giorgi et al., 1997), a feature that is not, however, systematically observed in all upland areas (e.g., Vuille and Bradley, 2000, for the Andes).

Few model simulations have attempted to directly address issues related specifically to future climatic change in mountain regions, primarily because the current spatial resolution of general circulation models (GCM) and even regional climate models (RCM) is generally too crude to adequately represent the topographic detail of most mountain regions and other climate-relevant features such as land-cover that are important determinants in modulating climate in the mountains (Beniston et al., 2003). High-resolution RCM simulations (5-km and 1-km grid scales) are used for specific investigations of processes such as surface runoff, infiltration evaporation, and extreme events such as precipitation (Kanada et al., 2005; Yasunaga et al., 2006; Weisman et al., 1997; Walser and Schär, 2004) and damaging wind storms (Goyette et al., 2003), but these simulations are too costly to operate in a “climate mode”. Because of the highly complex terrain, empirical/statistical downscaling techniques have often been seen as a very valuable tool to generate climate change information in mountaineous regions (e.g., Benestad, 2005; Hanssen-Bauer et al., 2005).

Projections of changes in precipitation patterns in mountains are unreliable in most GCMs because the controls of topography on precipitation are not adequately represented. In addition, it is now recognized that the superimposed effects of natural modes of climatic variability such as El Niño/Southern Oscillation (ENSO) or the North Atlantic Oscillation (NAO) can perturb mean precipitation patterns on time scales ranging from seasons to decades (Beniston and Jungo, 2001). Even though there has been progress in reproducing some of these mechanisms in coupled ocean-atmosphere models (Osborn et al., 1999), deficiencies remain and prevent a good simulation of these large scale modes of variability, see also Chapter 8, Section 8.4. However, several studies indicate that the higher resolution of RCMs and GCMs can represent observed mesoscale patterns of the precipitation climate that are not resolved in caurse resolution GCMs (Kanada et al., 2005; Yasunaga et al., 2006; Frei et al., 2003; Schmidli et al., 2006).

Snow and ice are, for many mountain ranges, a key component of the hydrological cycle, and the seasonal character and amount of runoff is closely linked to cryospheric processes. In temperate mountain regions, the snow-pack is often close to its melting point, so that it may respond rapidly to minor changes in temperature. As warming increases in the future, regions where snowfall is the current norm will increasingly experience precipitation in the form of rain (e.g., Leung et al., 2004). For every °C increase in temperature, the snowline will on average rise by about 150 m. Although the concept of defining the snowline is difficult to determine in the field, it is established that at lower elevations the snowline is very likely to rise by more than this simple average estimate (e.g., Martin et al., 1994; Vincent, 2002; Gerbaux et al., 2005; see also Chapter 4, Section 4.2). Beniston et al. (2003) have shown that for a 4°C shift in mean winter temperatures in the European Alps, as projected by recent RCM simulations for climatic change in Europe under the A2 emissions scenario, snow duration is likely to be reduced by 50% at altitudes near 2000 m and 95% at levels below 1000 m. Where some models predict an increase in wintertime precipitation, this increase does not compensate for the effect of changing in temperature. Similar reductions in snow cover that will affect other mountain regions of the world will have a number of implications, in particular for early seasonal runoff (e.g., Beniston, 2003), and the triggering of the annual cycle of mountain vegetation (Cayan et al., 2001; Keller et al., 2005).

Because mountains are the source region for over 50% of the globe's rivers, the impacts of climatic change on mountain hydrology not only impacts the mountains themselves but also populated lowland regions that depend on mountain water resources for domestic, agricultural, energy and industrial supply. Water resources for populated lowland regions are influenced by mountain climates and vegetation; shifts in intra-annual precipitation regimes could lead to critical water amounts resulting in greater flood or drought episodes (e.g., Barnett et al., 2005; Graham et al., 2006).

11.5 North America

Assessment of projected climate change for North America:

> All of North America is very likely to warm during this century, and the annual mean warming is likely to exceed the global mean warming in most areas. In northern regions, warming is likely to be largest in winter, and in the South-West USA largest in summer. The lowest winter temperatures are likely to increase more than the average winter temperature in northern North America, and the highest summer temperatures are likely to increase more than the average summer temperature in South-West USA.
>
> Annual-mean precipitation is very likely to increase in Canada and North-East USA, and likely to decrease in the South-West USA.In southern Canada, precipitation is likely to increase in winter and spring, but decrease in summer.
>
> Snow season length and snow depth are very likely to decrease in most of North America, except in the northernmost part of Canada where maximum snow depth is likely to increase.

The uncertainties in regional climate changes over North America are strongly linked to the ability of AOGCMs to reproduce the dynamical features affecting the region (Chapter 10). AOGCMs exhibit large model-to-model differences in ENSO and NAO/AO responses to climate changes. Changes in the Atlantic MOC are uncertain and thus so will the magnitude of consequent reduced warming in extreme northeastern part of North America and cooling here cannot be totally excluded. The Hudson Bay and Canadian Archipelago are poorly resolved by AOGCMs, contributing to uncertainty in ocean circulation and sea-ice changes and their influence on climate of northern regions. Tropical cyclones are not resolved by MMD models and inferred changes of frequency, intensity and tracks of disturbances making landfall in southeast regions remain uncertain. At the coarse horizontal resolution of MMD models, high terrain is poorly resolved which likely results in an underestimation of warming associated with snow-albedo feedback at high elevations in western regions. Little is known about the dynamical consequences of the larger warming over land than over ocean which may impact on the northward displacement and intensification of the subtropical anticyclone off the West Coast. This could affect the subtropical North Pacific eastern boundary current, the offshore Ekman transport, the upwelling and its cooling effect on SST, the persistent marine stratus clouds and thus precipitation of the southwest USA.

The uncertainty associated with RCM projections of climate change over North America remain large despite the investments made in increasing horizontal resolution. All reported RCM projections were driven by earlier AOGCMs that exhibited larger biases than MMD models. Coordinated ensemble RCMs projections over North America are not yet available making it difficult to compare results.

11.5.1 Key Processes

Central and northern regions of North America are under the influence of midlatitude cyclones. AOGCM projections (Chapter 10) generally indicate a slight poleward shift in storm tracks, an increase in the number of strong cyclones but a reduction in medium-strength cyclones over Canada and poleward of 70N. Consequent with the projected warming, the atmospheric moisture transport and convergence is projected to increase, resulting in a widespread increase of annual precipitation over most of the continent except the south and southwestern part of the USA and over Mexico.

The southwest region is very arid, under the general influence of a subtropical ridge of high pressure associated with the thermal contrast between land and adjacent ocean. The North American Monsoon System (NAMS) develops in early July (e.g., Higgins and Mo, 1997); the prevailing winds over the Gulf of California undergo a seasonal reversal, from northerly in winter to southerly in summer, bringing a pronounced increase in rainfall over the southwest USA and ending the late spring wet period in the Great Plains (e.g., Bordoni et al., 2004). The projection of smaller warming over the Pacific Ocean than over the continent, and amplification and northward displacement of the subtropical anticyclone, is likely to induce a decrease of annual precipitation for southwestern USA and northern Mexico.

The Great Plains Low-Level Jet (LLJ) is a dynamical feature that transports considerable moisture from the Gulf of Mexico into central USA, playing a critical role in the summer precipitation there. Several factors, including the land-sea thermal contrast, contribute to the strength of the moisture convergence during the night and early morning, resulting in prominent nocturnal maximum precipitation in the plains of USA (such as Nebraska, Iowa) (e.g., Augustine and Caracena, 1994). The projections of climate changes indicate an increased land-sea thermal contrast in summer, with anticipated repercussions on the LLJ.

Interannual variability over North America is connected to two large-scale patterns of oscillation (see Chapter 3), the El Niño – Southern Oscillation (ENSO) and the North Atlantic / Arctic Oscillation (NAO/AO). The MMD model projections indicate an intensification of the polar vortex and many a decrease of the Arctic surface pressure, which contributes to an increase of the AO/NAO index; the uncertainty is large however due to the diverse responses of AOGCM in the Aleutian Low (Chapter 10). The MMD models projections indicate a shift towards mean El Niño-like conditions, with the eastern Pacific warming more than the western Pacific; there is a wide range of behaviour among the current models, with no clear indication regarding possible changes in the amplitude or period of El Niño (Chapter 10).

11.5.2 Skill of Models in Simulating Present Climate

Individual AOGCMs in the MMD vary in their ability to reproduce the observed patterns of pressure, surface air temperature and precipitation over North America (Chapter 8). The ensemble mean of MMD models reproduces very well the annual-mean mean sea level pressure distribution (Section 8.4). The maximum error is of the order of ±2 hPa, with the simulated Aleutian low pressure extending somewhat too far north, probably due to the inability of coarse-resolution models to adequately resolve the high topography of the Rocky Mountains that blocks incoming cyclones in the Gulf of Alaska. Conversely the pressure trough over the Labrador Sea is not deep enough. The depth of the thermal low pressure over the southwest region in summer is somewhat excessive.

The MMD models simulate successfully the overall pattern of surface air temperature over North America, with reduced biases compared to those reported in the TAR. Ensemble-mean region-mean bias ranges from -4.5°C to 1.9°C for the 25 to 75 percentile range, and medians vary from –2.4°C to +0.4°C depending on region and season (Supplementary Material Table S11.1). The ensemble mean of MMD models reproduces the overall distribution of annual-mean precipitation (Supplementary Material Table S11.1), but almost all models overpredict precipitation for western and northern regions. The ensemble-mean region-mean percent precipitation bias medians vary from –16% to +93% depending on region and season. The ensemble-mean precipitation is excessive on the windward side of major mountain ranges. The ensemble-mean excess reaches 1 to 2 mm/day over high terrain in the west of the continent.

RCMs are quite successful in reproducing the overall climate of North America when driven by reanalyses,. Over a 10° x 10° Southern Plains region, an ensemble of six RCMs in the North American Regional Climate Change Assessment Program (NARCCAP; Mearns et al., 2005) had 76% of all monthly temperature biases within ±2°C and 82% of all monthly precipitation biases within ±50%, based on preliminary results for a single year. RCMs' simulations over North America exhibit rather high sensitivity to parameters such as domain size (e.g., Juang and Hong, 2001; Pan et al., 2001; Vannitsem and Chomé, 2005) and the intensity of the large-scale nudging if used (e.g., von Storch et al., 2000; Miguez-Macho et al., 2004). RCMs are in general more skilful at reproducing cold-season temperature and precipitation (e.g., Pan et al., 2001; Han and Roads, 2004; Plummer et al., 2006) because the warm-season climate is more controlled by mesoscale and convective-scale precipitation events which are harder to simulate (Giorgi et al., 2001a; Leung et al., 2003; Liang et al., 2004; Jiao and Caya, 2006). On the other hand Gutowski et al. (2004) found that spatial patterns of monthly precipitation for the USA, when viewed as a whole rather than broken into individual regions, were better simulated in summer than winter. Several studies point to the large sensitivity of RCMs to parameterisation of moist convection, including the vertical transport of moisture from the boundary layer (Chaboureau et al., 2004; Jiao and Caya, 2006) and entrainment mixing between convective plumes and the local environment (Derbyshire et al., 2004). In a study of the simulation of the 1993-summer flood in the central USA by 13 RCMs, Anderson et al. (2003) found that all models produced a precipitation maximum that represented the flood, but most under-predicted it to some degree, and 10 out of 13 of the models succeeded in reproducing the observed nocturnal maxima of precipitation. Leung et al. (2003) examined 95th percentile of daily precipitation and found generally good agreement across many areas of the Western USA.

A survey of recently published RCM current-climate simulations nested with AOGCM reveals that biases in surface air temperature and precipitation are two to three times larger than the simulations nested with reanalyses. The sensitivity of simulated surface air temperature to changing lateral boundary conditions from reanalyses to AOGCMs appears high in winter and low in summer (Han and Roads, 2004; Plummer et al., 2006). Most RCM simulations to-date for North America have been made for time slices that are too short to properly sample natural variability. Some RCMs have employed less than optimal formulations, such as outdated parameterisations (e.g., bucket land-surface scheme), too few levels in the vertical (e.g., 14) or a too low uppermost computational level (e.g., 100 hPa).

11.5.3 Climate Projections

11.5.3.1 Surface Air Temperature

The ensemble mean of MMD models projects a generalised warming for the entire continent with the magnitude projected to increase almost linearly with time (Figure 11.11). On an annual-mean basis, projected surface air temperature warming varies from 2 to 3°C along the western, southern and eastern continental edges (where at least 16 out of the 21 models project a warming in excess of 2°C) up to more than 5°C in the northern region (where 16 out of the 21 AOGCMs project a warming in excess of 4°C). This warming exceeds the spread amongst models by a factor of 3 to 4 over most of the continent. The warming in the USA is projected to exceed 2°C by nearly all models, and to exceed 4°C by more than 5 AOGCMs out of 21. More regional and seasonal detail on ranges of projected warming is provided in Table 11.1 and Supplementary Table S11.2.

[INSERT FIGURE 11.11 HERE]

The largest warming is projected to occur in wintertime over northern parts of Alaska and Canada, reaching 10°C in the northernmost parts, due to the positive feedback from a reduced period of snow cover. The ensemble-mean northern warming varies from more than 7°C in winter (nearly all AOGCMs project a warming exceeding 4°C) to as little as 2°C in summer. In summertime, ensemble-mean projected warming ranges between 3 and 5°C over most of the continent, with smaller values near the coasts. In western,central and eastern regions the projected warming has less seasonal variation and is more modest, especially near the coast consistent with less warming over the oceans. The warming could be larger in winter over elevated areas as a result of snow-albedo feedback, an effect that is poorly modelled by AOGCMs due to insufficient horizontal resolution (see also Box 11.3).In winter, the northern part of the eastern region is projected to warm most while coastal areas are projected to warm by only 2 to 3°C.

The climate-change response of RCMs is sometimes different from that of the driving AOGCM. This appears to be the result of a combination of factors, including the use of different parameterisations (convection and land-surface processes are particularly important over North America in summer) and resolution (different resolution may lead to differing behaviour of a same parameterisation). For example, Chen et al. (2003) found that two RCMs projected larger temperature changes in summer than their driving AOGCM. In contrast, the projected warming of an RCM compared to its driving AOGCM was found by to be 1.5°C less in the central USA (Pan et al., 2004 and Liang et al., 2006), a region where observations have shown a cooling trend in recent decades. This resulted in an area of little warming which may have been due to a changing pattern of the low-level jet (LLJ) frequency and associated moisture convergence. It is argued that the improved simulation of the LLJ in the RCM is made possible owing to its increased horizontal and vertical resolution. However, other RCMs with similar resolution do not produce the same response.

11.5.3.2 Precipitation

As a consequence of the temperature dependence of the saturation vapour pressure in the atmosphere, the projected warming is expected to be accompanied by an increase of atmospheric moisture flux and of its convergence / divergence intensity. This results in a general increase of precipitation over most of the continent except the southwestern most part (Figure 11.12). The ensemble mean of MMD models projects a fractional increase of annual-mean precipitation in the north reaching +20%, which is twice the inter-model spread, so likely significant; the projected increase reaches as much as +30% in wintertime. Because the increased saturation vapour pressure can also yield greater evaporation, projected increases in annual precipitation are partially offset by increases in evaporation; regions in central North America may experience net surface drying as a result (see Supplementary Material Figure S11.1). Again see Table 11.1 and Supplementary Table S11.2 for more regional and seasonal details noting that regional averaging hides important north-south differences.

[INSERT FIGURE 11.12 HERE]

In keeping with the projected northward displacement of the westerlies and the intensification of the Aleutian low (Section 11.5.3.3), northerm region precipitation is projected to increase, by the largest amount in autumn and by the largest fraction in winter. Due to the increased precipitable water, the increase in precipitation amount would likely be larger on the windward slopes of the mountains in the west with orographic precipitation. In western regions, modest changes in annual-mean precipitation are projected but the majority of AOGCMs indicate an increase in winter and a decrease in summer. Models show greater consensus on winter increases (ensemble mean maximum of 15%) to the north and on summer decreases (ensemble mean maximum of –20%) to the south. These decreases are consistent with enhanced subsidence and drier airmass flowing in southwest USA and northern Mexico resulting from an amplication of the subtropical anticyclone off the west coast due to the land-sea contrast in warming (e.g., Mote and Mantua, 2002). However, this reduction is close to the inter-model spread so it contains large uncertainty, an assessment that is reinforced by the fact that some AOGCMs project an increase of precipitation.

In central and eastern regions, projections from the MMD models show the same characteristics as in the west of greater consensus of winter increases to the north and summer decreases to the south. The line of zero change is oriented more or less west-to-east and moves north from winter to summer. The line of zero change is also projected to lie further to the north under SRES scenarios with larger GHG amounts. However, uncertainty around the projected changes is large and the changes do not scale well across different SRES scenarios.

Govindasamy (2003) found that, averaged over the USA, the few existing time-slice simulations with high-resolution AGCM results do not significantly differ from those obtained with AOGCMs. Available RCM simulations provide little extra information on average changes. Some RCMs project precipitation changes of different sign, either locally (Chen et al., 2003) or over the whole continental USA (Han and Roads (2004), where in summer the AOGCM generally produced a small increase and the RCM a substantial decrease). In contrast, Plummer et al. (2006) found only small differences in precipitation responses using two sets of physical parameterisations in their RCM, despite the fact that one corrected significant summertime precipitation excess present in the other.

11.5.3.3 Temperature and Precipitation Extremes

Several RCM studies focused particularly on changes in extreme temperature events. Bell et al. (2004) examined changes in temperature extremes in their simulations centred on California. They found increases in extreme temperature events, both as distribution percentiles and threshold events, prolonged hot spells and increased diurnal temperature range. Leung et al. (2004) examined changes in extremes in their RCM simulations of the western USA; in general they found increases in diurnal temperature range in six sub-regions of their domain in summer. Diffenbaugh et al. (2005) found that the frequency and magnitude of extreme temperature events changes dramatically under SRES A2, with increases in extreme hot events and decrease in extreme cold events.

In a study of precipitation extremes over California, Bell et al. (2004) found that changes in precipitation exceeding the 95th percentile followed changes in mean precipitation, with decreases in heavy precipitation found for most areas. Leung et al. (2004) found that extremes in precipitation during the cold season increased in the northern Rockies, the Cascades, the Sierra and British Columbia by up to 10% for 2040–2060, although mean precipitation was mostly reduced, in accord with earlier studies (Giorgi et al., 2001a). In a large river basin in the Pacific Northwest, increases in rainfall over snowfall and rain-on-snow events increased extreme runoff by 11%, which would contribute to more severe flooding. In their 25-km RCM simulations covering the entire USA, Diffenbaugh et al. (2005) found widespread increases in extreme precipitation events under SRES A2, which they determined to be significant.

11.5.3.4 Atmospheric Circulation

In general the projected climate changes over North America follow the overall features of those over the Northern Hemisphere (NH) (Chapter 10). The MMD models project a northward displacement and strengthening of the mid-latitude westerly flow, most pronounced in autumn and winter. Surface pressure is projected to decrease in the north, with a northward displacement of the Aleutian low-pressure centre and a northwestward displacement of the Labrador Sea trough, and to decrease slightly in the south. The lowering surface pressure in the north is projected to be strongest in wintertime, reaching –1.5 to –3 hPa, in part as a result of the warming of the continental Arctic airmass. On an annual basis, the pressure decrease in the north exceeds the spread amongst models by a factor of 3 on an annual-mean basis and 1.5 in summer, so it is significant. The East Pacific subtropical anticyclone is projected to intensify in summer, particularly off the coast of California and Baja California, resulting in an increased airmass subsidence and drier airflow over southwestern North America. The pressure increase (less than 0.5 hPa) is small compared to the spread amongst models, so this projection is rather uncertain.

11.5.3.5 Snowpack, Snowmelt and River Flow

The ensemble-mean MMD models project a general decrease of snow depth (Chapter 10) as a result of delayed autumn snow fall and earlier spring snow melt. In some regions where winter precipitation is projected to increase, the increases snow fall can more than make up for the shorter snow season and yield increased snow accumulation. Snow depth increases are projected by some GCMs over some land around the Arctic Ocean (Chapter 10, Figure S10.1) and by some RCMs in the northern-most part of the North-West Territories (Figure 11.13). In principle a similar situation could arise at lower latitudes at high elevations over the Rocky Mountains though most models project a widespread decrease of snow depth here (Kim et al., 2002; Snyder et al., 2003; Leung et al., 2004, see also Box 11.3).

[INSERT FIGURE 11.13 HERE]

Much statistical downscaling (SD) research activity has focused on resolving future water resources in the complex terrain of the western USA. Studies typically point to a decline in winter snowpack and hastening of the onset of snowmelt caused by regional warming (Hayhoe et al., 2004; Salathé, 2005). Comparable trends towards increased mean annual river flows and earlier spring peak flows have also been projected by two SD techniques for the Saguenay watershed in northern Québec, Canada (Dibike and Coulibaly, 2005). Such changes in the flow regime also favour increased risk of winter flooding, lower summer soil moisture and river flows. However, differences in snowpack behaviour derived from AOGCMs depend critically on the realism of downscaled wintertime temperature variability and its interplay with precipitation and snowpack accumulation and melt (Salathé, 2005). Hayhoe et al. (2004) produced a standard set of statistically downscaled temperatures and precipitations scenarios for California; under both the A1F1 and B1 scenarios, they found overall declines in snowpack.

11.6 Central and South America

Assessment of projected climate change for Central and South America:

> All of Central and South America is very likely to warm during this century. The annual mean warming is likely to be similar to the global mean warming in Southern South America but larger than the global mean warming in the rest of the area.
>
> Annual precipitation is likely to decrease in most of Central America, with the relatively dry boreal spring becoming drier. Annual precipitation is likely to decrease in the southern Andes, WITH RELATIVE PRECIPITATION CHANGES BEING LARGEST IN SUMMER. A caveat on the local scale is that changes in atmospheric circulation may induce large local variability in precipitation changes in mountainous areas. PRECIPITATION IS LIKELY TO INCREASE IN TIERRA DEL FUEGO DURING WINTER AND in south-eastern South America during summer.Precipitation is likely to increase in south-eastern South America during summer.
>
> It is uncertain how annual and seasonal mean rainfall will change over northern South America, including the Amazon forest. In some regions there is qualitative consistency among the simulations (rainfall increasing in Ecuador and northern Peru, and decrease in the northern tip of the continent and in southern northeast Brazil).

The systematic errors in simulating current mean tropical climate and its variability (Chapter 8, Section 8.6) and the large inter-model differences in future changes of El Niño amplitude (Chapter 10, Section10.3) preclude a conclusive assessment of the regional changes over large areas of Central and South America. Most MMD models are poor in reproducing the regional precipitation patterns in their control experiment and have a small signal to noise ratio, in particular over most of AMZ. The high and sharp Andes Mountains are unresolved in low-resolution models, affecting the assessment over much of the continent. As with all land masses, the feedbacks from land use and land cover change are not well accommodated, and lend some degree of uncertainty. The potential for abrupt changes in biogeochemical systems in AMZ remains as a source of uncertainty (see Chapter 10, Box 10.1). Large differences in the projected climate sensitivities in the climate models incorporating these processes and a lack of understanding of processes were identified (Friedlingstein et al., 2003). Over Central America, tropical cyclones may become an additional source of uncertainty for regional scenarios of climate change, since the summer precipitation over this region may be affected by systematic changes in hurricane tracks and intensity.

11.6.1 Key Processes

Over much of Central and South America, changes in the intensity and location of tropical convection are the fundamental concern, but extratropical disturbances also play a role in Mexico’s winter climate and throughout the year in Southern South America. A continental barrier over Central America and along the Pacific coast in South America and the world's largest rainforest are unique geographical features that shape the climate in the area.

Climate over most of Mexico and Central America is characterized by a relatively dry winter and a well-defined rainy season from May through October (Magaña et al., 1999). The seasonal evolution of the rainy season is to a large extent, the result of air sea interactions over the Americas warm pools and the effects of topography over a dominant easterly flow, as well as the temporal evolution of the Inter Tropical Convergence Zone (ITCZ). During the boreal winter, the atmospheric circulation over the Gulf of Mexico and the Caribbean Sea is dominated by the seasonal fluctuation of the Subtropical North Atlantic Anticyclone, with invasions of extratropical systems that affect mainly Mexico and the western portion of the Great Antilles.

A warm season precipitation maximum, associated with the South American Monsoon System (Vera et al., 2006), dominates the mean seasonal cycle of precipitation in tropical and subtropical latitudes over South America. Amazonia has had increasing rainfall over the last 40 years, despite deforestation, due to global-scale water vapour convergence (Chen et al., 2001; see also Chapter 3, Section 3.3). The future of the rainforest is not only of vital ecological importance, but also central to the future evolution of the global carbon cycle, and as a driver of regional climate change. The monsoon system is strongly influenced by ENSO (e.g., Lau and Zhou, 2003), and thus future changes in ENSO will induce complementary changes in the region. Displacements of the South Atlantic Convergence Zone have important regional impacts such as the large positive precipitation trend over the recent decades centred over southern Brazil (Liebmann et al., 2004). There are well-defined teleconnection patterns (the Pacific-South American modes, Mo and Nogués-Paegle, 2001) whose preferential excitation could help shape regional changes. The Mediterranean climate of much of Chile makes it sensitive to drying as a consequence of poleward expansion of the South Pacific subtropical high, in close analogy to other regions downstream of oceanic subtropical highs in the Southern Hemisphere. South Eastern South America would experience an increase in precipitation from the same poleward storm track displacement.

11.6.2 Skill of Models in Simulating Present Climate

In the Central America (CAM) and Amazonia (AMZ) regions, most models in the MMD have a cold bias of 0–3°C, except in AMZ in SON (Supplementary Material Table S11.1). In Southern South America (SSA) average biases are close to zero. The biases are unevenly geographically distributed (Supplementary material Figure S11.25). The MMD mean climate shows a warm bias around 30°S (particularly in summer) and in parts of central South America (especially in SON). Over the rest of South America (central and northern Andes, eastern Brazil, Patagonia) the biases tend to be predominantly negative. The SST biases along the western coasts of South America are likely related to weakness in oceanic upwelling.

For the CAM region, the multi-model scatter in precipitation is substantial, but half of the models lie in the range of (–15%, 25%) in the annual mean. The largest biases occur during the boreal winter and spring seasons, when precipitation is meagre (Supplementary Material Table S11.1). For both AMZ and SSA, the ensemble annual mean climate exhibits drier than observed conditions, with about 60% of the models having a negative bias. Unfortunately, this choice of regions for averaging is particularly misleading for South America since it does not clearly bring out critical regional biases such as those related to rainfall underestimation in the Amazon and La Plata basins (Supplementary Material Figure S11.26). Simulation of the regional climate is seriously affected by models' deficiencies at low latitudes. In particular, the MMD ensemble tends to depict a relatively weak ITCZ, which extends southward of its observed position. The simulations have a systematic bias towards underestimated rainfall over the Amazon Basin. The simulated subtropical climate is typically also adversely affected by a dry bias over most of South Eastern South America and in the South Atlantic Convergence Zone, especially during the rainy season. In contrast, rainfall along the Andes and in NE Brazil is excessive in the ensemble mean.

Some aspects of the simulation of tropical climate with AOGCMs has improved. However, in general, the largest errors are found where the annual cycle is weakest, such as over tropical South America, see e.g., Chapter 8, Section 8.3. AGCMs approximate the spatial distribution of precipitation over the tropical Americas, but they do not correctly reproduce the temporal evolution of the annual cycle in precipitation, specifically the mid summer drought (Magaña and Caetano, 2005). Tropical cyclones are important contributers to precipitation in the region. If close to the continent, then they will produce large amounts of precipitation over land, and if far from the coast, moisture divergence over the continental region enhances drier conditions.

Zhou and Lau (2002) analyse the precipitation and circulation biases in a set of 6 AGCMs provided by the CLIVAR Asian-Australian Monsoon AGCM Intercomparison Project (Kang et al., 2002). This model ensemble captures some large-scale features of the South American Monsoon System reasonably well including the seasonal migration of monsoon rainfall and the rainfall associated with the SACZ. However, the South Atlantic subtropical high and the Amazonia low are too strong, whereas low-level flow tends to be too strong during austral summer and too weak during austral winter. The model ensemble captures the Pacific-South American pattern quite well, but its amplitude is generally underestimated.

Regional models are still being tested and developed for this region. Relatively few studies using RCMs for Central and South America exist, and those that do are constrained by short simulation length. Some studies (Chou et al., 2000; Nobre et al., 2001; Druyan et al., 2002) examine the skill of experimental dynamic downscaling of seasonal predictions over Brazil. Results suggest that both more realistic GCM forcing and improvements in the RCMs are needed. Seth and Rojas (2003) performed seasonal integrations with emphasis on tropical South America applying reanalyses boundary forcing. The model was able to simulate the different rainfall anomalies and large-scale circulations but, as a result of weak low-level moisture transport from the Atlantic, rainfall over the western Amazon was under-simulated. Vernekar et al. (2003) followed a similar approach to study the low-level jets and reported that the RCM produces better regional circulation details than does the reanalysis. However, an ensemble of four RCMs did not provide a noticeable improvement in precipitation over the driving large-scale reanalyses (Roads et al., 2003).

Other studies (Rojas and Seth, 2003; Misra et al., 2003) analyse seasonal RCM simulations driven by AGCM simulations. Relative to the AGCMs, regional models generally improve the rainfall simulation and the tropospheric circulation over both tropical and subtropical South America. However, AGCM-driven RCMs degrade compared with the reanalyses-driven integrations and they could even exacerbate the dry bias over sectors of AMZ and perpetuate the erroneous ITCZ over the neighbouring ocean basins from the AGCMs. Menéndez et al. (2001) used a RCM driven by a stretched-grid AGCM with higher resolution over the southern mid-latitudes to simulate the winter climatology of SSA. They find that both the AGCM and the regional model have similar systematic errors but the biases are reduced in the RCM. Analogously, other RCM simulations for SSA have given too little precipitation over the subtropical plains and too much over elevated terrain (e.g., Nicolini et al., 2002; Menéndez et al., 2004).

11.6.3 Climate Projections

11.6.3.1 Temperature

The warming as simulated by the MMD-A1B projections increases approximately linearly with time during this century, but the magnitude of the change and inter-model range are greater over CAM and AMZ than over SSA (Figure 11.14). The annual mean warming under the A1B scenario from 1980–1999 to 2080–2099 varies in the CAM region from 1.8 to 5.0°C, with half of the models within 2.6–3.6°C and a median of 3.2°C. The corresponding numbers for AMZ are 1.8 to 5.1°C, 2.6–3.7°C and 3.3°C, and those for SSA 1.7 to 3.9°C, 2.3–3.1°C and 2.5°C (Table 11.1). The median warming is close to the global ensemble mean in SSA but about 30% above the global mean in the other two regions. As in the rest of the tropics, the signal to noise ratio is large for temperature, and it requires only 10 years for a 20 year mean temperature, growing at the rate of the median A1B response, to be clearly discernible above the models' internal variability.

[INSERT FIGURE 11.14 HERE]

The simulated warming is generally largest in the most continental regions, such as inner Amazonia and northern Mexico (Figure 11.15). Seasonal variation in the regional area mean warming is relatively modest, except in CAM where there is a difference of 1°C in median values between DJF and MAM (Table 11.1). The warming in central Amazonia tends to be larger in JJA than in DJF, while the reverse is true over the Altiplano where, in other words, the seasonal cycle of temperature is simulated to increase (Figure 11.15). Similar results were found by Boulanger et al. (2006) who studied the regional thermal response over South

America by applying a statistical method based on neural networks and Bayesian statistics to find optimal weights for a linear combination of MMD models.

[INSERT FIGURE 11.15 HERE]

For the variation of seasonal warming between the individual models, see Table 11.1. As an alternative approach to estimating uncertainty in the magnitude of the warming, the 5% and 95% quantiles for temperature change at the end of the 21st century, assessed from the method of Tebaldi et al. (2005) are typically within ±1°C of the median value in all three of these regions (Supplementary material Table S11.2).

11.6.3.2 Precipitation

The MMD models suggest a general decrease in precipitation over most of Central America, consistent with Neelin et al. (2006), where the median annual change by the end of the 21st century is –9% under the A1B scenario, and half of the models project area mean changes from –16% to –5% although the full range of the projections extends from –48% to 9%. Median changes in area mean precipitation in Amazonia and Southern South America are small and the variation between the models is also more modest than in Central America, but the area means hide marked regional differences (Table 11.1, Figure 11.15).

Area mean precipitation in Central America decreases in most models in all seasons. It is only in some parts of North Eastern Mexico and over the eastern Pacific, where the ITCZ forms during JJA that increases in summer precipitation are projected (Figure 11.15). However, since tropical storms can contribute a significant fraction of the rainfall in the hurricane season in this region, these conclusions might be modified by the possibility of increased rainfall in storms not well captured by these global models. In particular, if the number of storms does not change, Knutson and Tuleya (2004) estimate nearly a 20% increase in average precipitation rate within 100 km of the storm centre at the time of CO_2 doubling.

For South America, the multi-model mean precipitation response (Figure 11.15) indicates marked regional variations. The annual mean precipitation is projected to decrease over northern South America near the Caribbean coasts, as well as over large parts of northern Brazil, Chile and Patagonia, while it is projected to increase in Colombia, Ecuador and Peru, around the equator and in South Eastern South America. The seasonal cycle modulates this mean change especially over the Amazon basin where monsoon precipitation increases in DJF and decreases in JJA. In other regions (e.g., Pacific coasts of northern South America, a region centred over Uruguay, Patagonia) the sign of the response is preserved throughout the seasonal cycle.

As seen in the bottom panels in Figure 11.15, most models project a wetter climate near the Rio de la Plata and drier conditions along much of the southern Andes, especially in DJF. However, when estimating the likelihood of this response, the qualitative consensus within this set of models should be weighed against the fact that most models show considerable biases in regional precipitation patterns in their control simulations.

The poleward shift of the South Pacific and South Atlantic subtropical anticyclones is a robust response across the models. Parts of Chile and Patagonia are influenced by the polar boundary of the subtropical anticyclone in the South Pacific and experience particularly strong drying because of the combination of the poleward shift of circulation and increase of moisture divergence. The strength and position of the subtropical anticyclone in the South Atlantic is known to influence the climate of South Eastern South America and the South Atlantic Convergence Zone (Robertson et al., 2003; Liebmann et al., 2004). The increase in rainfall in South Eastern South America is related with a corresponding poleward shift of the Atlantic storm track (Yin, 2005).

Some projected changes in precipitation (such as the drying over east-central Amazonia and northeast Brazil and the wetter conditions over south-eastern South America) could be a partial consequence of the El Niño-like response projected by the models (Chapter 10, Section 10.3). The accompanying shift and alterations of the Walker circulation would directly affect tropical South America (Cazes Boezio et al., 2003) and affect Southern South America through extratropical teleconnections (Mo and Nogués-Paegle, 2001).

Although feedbacks from carbon cycle and dynamic vegetation are not included in MMD models, a number of coupled carbon cycle-climate projections have been performed since the TAR (see Chapter 7, Section 7. 2 and Chapter 10, Section 10.4.1). The initial carbon-climate simulations suggest that drying of the Amazon potentially contribute to acceleration of the rate of anthropogenic global warming by increasing atmospheric carbon dioxide (Cox et al., 2000; Jones et al., 2003; Friedlingstein et al., 2001; Dufresne et al., 2002). These models display large uncertainty in climate projections and differ in the timing and sharpness of the changes (Friedlingstein et al., 2003). Changes in carbon dioxide are related to changes in precipitation in regions such as northern Amazon (Zeng et al., 2004). In a version of HadCM3 model with dynamic vegetation and an interactive global carbon cycle (Betts et al., 2004), a tendency to a more El Niño like state contributes to reduced rainfall and vegetation dieback in the Amazon (Cox et al., 2004). But the version of HadCM3 participating in the MMD projects by far the largest reduction in annual rainfall over AMZ (–21% for the A1B scenario). This stresses the necessity of being very cautious in interpreting carbon cycle impact on the regional climate and ecosystem change until there is more convergence among models on projections for rainfall in the Amazon with fixed vegetation. Box 11.4 summarises some of the major issues related to regional land use/land changes in the context of climate change.

11.6.4 Extremes

Little research is available on extremes of temperature and precipitation for this region. Table 11.1 provides estimates on how frequently the seasonal temperature and precipitation extremes as simulated in 1980–1999 are exceeded in using the A1B scenario. Essentially all seasons and regions are extremely warm by this criterion by the end of the century. In Central America, the projected time mean precipitation decrease is accompanied by more frequent dry extremes in all seasons. In South America, models anticipate extremely wet seasons in about 27% (in AMZ) and 13% (in SSA) of all DJF seasons in the period 2080–2099. The corresponding frequencies for extremely dry JJA seasons would be 16% (in AMZ) and 11% (in SSA). However, a more careful analysis is required to determine how often these wet and dry extremes are projected by the same model before concluding that both extremes are likely to increase. Austral winter (summer) seasons would not project significant changes in the frequency of extremely wet (dry) seasons.

On the daily time scale, Hegerl et al. (2004) analysed an ensemble of simulations from two AOGCMs and found that both models simulate a temperature increase in the warmest night of the year larger than the mean response over the Amazon Basin but smaller than the mean response over parts of SSA. Concerning extreme precipitation, both models project more intense wet days per year over large parts of South Eastern South America and central Amazonia and weaker precipitation extremes over the coasts of NE Brazil. Intensification of the rainfall amounts are deemed significant by a majority of MMD models over parts of South Eastern South America and most of AMZ but with longer periods between rainfall events, except in north-western South America where it rains more frequently (Meehl et al., 2005; Tebaldi et al., 2006).

Box 11.4: Land-Use/Cover Change Experiments Related to Climate Change

Land use and land cover change (LUCC) significantly affect climate at the regional and local scales (e.g., Hansen et al., 1998; Kabat et al., 2002, Bonan, 2001; Foley et al., 2005). Recent modelling studies also show that in some instances these effects can extend beyond the areas where the land cover changes occur, through climate teleconnection processes (e.g., Pielke et al., 2002; Marland et al., 2003). Changes in vegetation result in alteration of surface properties, such as albedo and roughness length, and alter the efficiency of ecosystems to exchange water, energy and carbon dioxide with the atmosphere (for more details see Chapter 7, Section 7.2). The effects differ widely based on the type of and location of the ecosystem altered. The effects of LUCC on climate can also be divided into biogeochemical and biophysical (Brovkin et al., 1999) (see Chapter 7, Section 7.2 and Chapter 2, Section 2.5 for discussion of these effects).

The net effect of human land-cover activities increases the concentration of greenhouse gases in the atmosphere, thus increasing warming (see Chapter 7, Section 7.2 and Chapter 10, Section 10.4 for further discussion); it has been suggested that these land cover emissions have been underestimated in the future climate projections used in the SRES scenarios (Sitch, 2005). Climate models assessed in this report incorporate various aspects of the effects of land cover change including representation of the biogeochemical flux, inclusion of dynamic land use where natural vegetation shifts as climate changes, and explicit human land cover forcing. In all cases these efforts should be considered at early stages of development. (See Chapters 2 and 7 and Chapter 10, Table 10. 1 for more details on many of these aspects).

One important land-cover conversion impact, generally not simulated in GCMs, is urbanization. Although small in aerial extent, conversion to urban land cover creates urban heat islands associated with considerable warming (Arnfield, 2003). Since much of the world population lives in urban environments (and this proportion may increase, thus expanding urban areas), many people will be exposed to climates that combine expanded urban heat island effects together with increased temperature from greenhouse gas forcing. See Chapter 7, Box 7.2 for more details on urban land use effects

One major shift in land use, relevant historically and in the future, is conversion of forest to agriculture and agriculture back to forest. Most areas well suited to large scale agriculture have already been converted to this land use/cover type. Yet land-cover conversion to agriculture may continue in the future, especially in parts of western North America, tropical areas of south and central America, and arable regions in Africa and south and central Asia (IPCC, 2001; RIVM, 2002). In the future, mid-latitude agricultural areal expansion (especially into forested areas) could possibly result in cooling that would offset a portion of the expected warming due to greenhouse gas effects alone. In contrast, reforestation may occur in eastern North America and the eastern portion of Europe. In these areas climate effects may include local warming associated with reforestation due to decreased albedo values (Feddema et al., 2005).

Tropical land cover change results in a very different climate response compared to mid-latitude areas. Changes in plant cover and the reduced ability of the vegetation to transpire water to the atmosphere lead to warmer temperatures by as much as 2°C in regions of deforestation (Gedney and Valdes, 2000; Costa and Foley, 2000; De Fries et al., 2002). The decrease in transpiration acts to reduce precipitation, but this effect may be modified by changes in atmospheric moisture convergence. Most model simulations of Amazonian deforestation suggest reduced moisture convergence which would amplify the decrease in precipitation (e.g., McGuffie et al., 1995; Costa and Foley, 2000; Avissar and Worth, 2005). However, increased precipitation and moisture convergence in Amazonia during the last decades contrast with this expectation, suggesting that deforestation has not been the dominant driver of the observed changes (see Section 11.6).

Tropical regions also have the potential to affect climates beyond their immediate areal extent (Chase et al., 2000; Delire et al., 2001; Voldoire and Royer, 2004; Avissar and Werth, 2005; Feddema et al., 2005, Snyder, 2006). For example, changes in convection patterns can affect the Hadley circulation and thus propagate climate perturbations into the midlatitudes. In addition, tropical deforestation in the Amazon has been found to affect sea surface temperatures in nearby ocean locations, further amplifying teleconnections (Avissar and Werth, 2005; Feddema et al., 2005; Neelin and Su, 2005; Voldoire and Royer, 2005). However, studies also indicate that there are significantly different responses to similar land use changes in other tropical regions and that responses are typically linked to dry season conditions (Voldoire and Royer, 2004a,; Feddema et al., 2005). However tropical land cover change in Africa and southeast Asia appears to have weaker local impacts in large part due to influences of the Asian and African monsoon circulation systems (Voldoire and Royer, 2005; Mabuchi et al., 2005a,b).

Several land cover change studies have explicitly assessed the potential impacts (limited to biophysical effects) associated with specific future IPCC SRES land cover change scenarios, and the interaction between land cover change and greenhouse gas forcings (De Fries et al., 2002; Maynard and Royer, 2004a; Sitch et al., 2005; Feddema et al., 2005; Voldoire, 2006). In the A2 scenario large-scale Amazon deforestation could double the expected warming in the region (De Fries et al., 2002; Feddema et al., 2005). Lesser local impacts are expected in tropical Africa and south Asia, in part because of the difference in regional circulation patterns (Delire et al., 2001; Maynard and Royer, 2004a,b; Feddema et al., 2005; Mabuchi et al., 2005a,b). In mid-latitude regions land cover induced cooling could offset some of the greenhouse gas induced warming. Feddema et al., 2005 suggest in the B1 scenario (where reforestation occurs in many areas and there are other low impact tropical land cover changes) there are few local tropical climate or teleconnection effects . However, in this scenario mid-latitude reforestation could lead to additional local warming compared to green house gas forcing scenarios alone.

These simulations suggest that the effects of future land-cover change will be a complex interaction of local land-cover change impacts combined with teleconnection effects due to land-cover change elsewhere, in particular the Amazon, and areas surrounding the Indian Ocean. However, projecting the potential outcomes of future climate effects due to land-cover change is difficult for two reasons. First, there is considerable uncertainty regarding how land cover will change in the future. In this context, the past may not be a good indicator of the types of land transformation that may occur in the future. For example, if land cover change becomes a part of climate change mitigation (e.g., carbon trading) then a number of additional factors that include carbon sequestration in soils and additional land cover change processes will need to be incorporated in scenario development schemes. Second, current land-process models cannot simulate all the potential impacts of human land-cover transformation. Such processes as adequate simulation of urban systems, agricultural systems, ecosystem disturbance regimes (e.g., fire) and soil impacts are not yet well represented.

11.7 Australia – New Zealand

Assessment of projected climate change for Australia and New Zealand:

> All of Australia and New Zealand are very likely to warm during this century, with amplitude somewhat larger than that of the surrounding oceans, but comparable overall to the global mean warming. The warming is smaller in the south, especially in winter, with the warming in the South Island of New Zealand likely to remain smaller than the global mean. Increased frequency of extreme high daily temperatures in Australia and New Zealand, and decrease in the frequency of cold extremes is very likely.
>
> Precipitation is likely to decrease in Southern Australia in winter and spring. Precipitation is very likely to decrease in Southwestern Australia in winter. Precipitation is likely to increase in the west of the South Island of New Zealand. Changes in rainfall in Northern and Central Australia are uncertain. Extremes of daily precipitation will very likely increase. The effect may be offset or reversed in areas of significant decrease in mean rainfall (southern Australian in winter and spring.). Increase in potential evaporation is likely. Increased risk of drought in southern areas of Australia is likely.
>
> Increased mean wind speed across the southern island of New Zealand, particularly in winter, is likely.

There are significant factors contributing to uncertainty in projected climate change for the region. ENSO significantly influences rainfall, drought and tropical cyclone behaviour in the region and it is uncertain how ENSO will change in the future. Monsoon rainfall simulations and projections vary substantially from model to model thus we have little confidence in model precipitation projections for Northern Australia. More broadly across the continent summer rainfall projections vary substantially from model to model reducing confidence in their reliability. Also, no detailed assessment of MMD model performance over Australia or New Zealand is available which hinders efforts to establish the reliability of projections from these models. Finally, downscaling of MMD model projections are not yet available for New Zealand but are much needed because of the strong topographical control of New Zealand rainfall.

11.7.1 Key Processes

Key climate processes affecting the Australian region include the Australian monsoon (the southern hemisphere counterpart of the Asian monsoon), the Southeast trade wind circulation, the subtropical high-pressure belt and the midlatitude westerly wind circulation with its embedded disturbances. The latter two systems also predominate over New Zealand. Climatic variability in Australia and New Zealand is also strongly affected by the El Niño-Southern Oscillation system (McBride and Nicholls, 1983; Mullan, 1995) modulated by the Interdecadal Pacific Oscillation (IPO) (Power et al., 1999; Salinger et al., 2001). Tropical cyclones occur in the region, and are a major source of extreme rainfall and wind events in northern coastal Australian, and, more rarely, in the north island of New Zealand (Sinclair, 2002). Rainfall patterns in New Zealand are also strongly influenced by the interaction of the predominantly westerly circulation with its very mountainous topography.

Apart from the general increase in temperature that the region will share with most other parts of the globe, details of anthropogenic climate change in the Australia-New Zealand region will depend on the response of the Australian monsoon, tropical cyclones, the strength and latitude of the midlatitude westerlies, and ENSO.

11.7.2 How Well is the Climate of the Region Currently Simulated?

There are relatively few studies of the quality of the MMD global models in the Australia/New Zealand area. The ensemble mean of MMD models has a systematic low pressure bias near 50°S at all longitudes in the Southern hemisphere, including the Australia/New Zealand sector, corresponding to an equatorward displacement of the midlatitude westerlies (see Chapter 8). On average, midlatitude storm track eddies are displaced equatorward (Yin, 2005) and deep winter troughs over southwest Western Australia are over-represented (Hope 2006a, b). How this bias might affect climate change simulations is unclear. One can hypothesize that by spreading the effects of midlatitude depressions too far inland, the consequences of a poleward displacement of the westerlies and the storm track might be exaggerated, but the studies needed to test this hypothesis are not yet available.

The simulated surface temperatures in the surrounding oceans are typically warmer than observed, but at most by 1°C in the composite. Despite this slight warm bias, the ensemble mean temperatures are biased cold over land, especially in winter in the Southeast and Southwest of the Australian continent, where the cold bias is larger than 2°C. On large scales, the precipitation also has some systematic biases (see Supplementary Material Table S11.1). Averaged across Northern Australia, the median model error is 20% more precipitation than observed, but the range of biases in individual models is large (–71% to +131%). This is discouraging with regard to confidence in many of the individual models. Consistent with this, Moise et al. (2005) identified simulation of Australian monsoon rainfall as a major deficiency of many of the AOGCM simulations included in CMIP2. The median annual bias in the southern Australian region is negative 6%, and the range of biases –59% to +36%. In most models the northwest is too wet and the northeast and east coast too dry, and the central arid zone is insufficiently arid.

The Australasian simulations in the AOGCMs utilized in the TAR have recently been scrutinized more closely, in part as a component of a series of national and state-based climate change projection studies (e.g., Whetton et al., 2001; McInnes et al., 2003; Hennessy et al., 2004a; McInnes et al., 2004; Hennessy et al., 2004b, Cai et al., 2003a). Some high-resolution regional simulations were also considered in this process. The general conclusion has been that large-scale features of Australian climate were quite well simulated. In winter, temperature patterns were poorer in the south where topographic variations have a stronger influence although this was alleviated in the higher resolution simulations. A set of the TAR AOGCM simulations was also assessed for the New Zealand region by Mullan et al. (2001a) with similar conclusions. The models were able to represent ENSO-related variability in the Pacific and the temperature and rainfall teleconnection patterns at the Pacific-wide scale, but there was considerable variation in model performance at finer scale (such as over the New Zealand region).

Decadal-scale variability patterns in the Australian region as simulated by the CSIRO AOGCM were considered by Walland et al. (2000) and found 'broadly consistent' with the observational studies of Power et al. (1998). On smaller scales, Suppiah et al. (2004) directly assessed rainfall-producing processes in the model in Victoria by comparing the simulated correlation between rainfall anomalies and pressure anomalies against observations. They found that this link was simulated well by most models in winter and autumn, but less well in spring and summer. As a result of this they warned that the spring and summer projected rainfall changes should be viewed as less reliable.

Pitman and McAvaney (2004) examined the sensitivity of GCM simulations of Australian climate to methods of representation of the surface energy balance. They found that the quality of the simulation of variability was strongly affected by the land surface model, but that simulation of climate means, and the changes in those means in global warming simulations, was less sensitive to the scheme employed.

Statistical downscaling methods have been employed in the Australian region and have demonstrated good performance at representing means variability and extremes of station temperature and rainfall (Timbal and McAvaney, 2001; Timbal, 2004; Charles et al., 2004) based on broad-scale observational or climate model predictor fields. The method of Charles et al. (2004) is able to represent spatial coherence at the daily timescale in station rainfall, thus enhancing its relevance to hydrological applications.

11.7.3 Projected Regional Climate Change

In addition to the MMD models, numerous studies have been conducted with earlier models. Recent regional average projections are provided in Giorgi et al. (2001b) and Ruosteenoja et al. (2003). The most recent national climate change projections of CSIRO (2001) were based on the results of eight AOGCMs and one higher resolution regional simulation. The methodology (and simulations) used in these projections is described in Whetton et al. (2005) and follows closely that described for earlier projections in Whetton et al. (1996). More detailed projections for individual states and other regions have also been prepared in recent years (Whetton et al., 2001; McInnes et al., 2003; Hennessy et al., 2004a; McInnes et al., 2004; Hennessy et al., 2004b, Cai et al., 2003a, IOCI 2005). This work has focused on temperature and precipitation, with additional variables such as potential evaporation and winds being included in the more recent assessments.

A range of dynamically downscaled projections have been undertaken for Australia using the DARLAM regional model (Whetton et al., 2001) and the CCAM stretched grid model (McGregor and Dix, 2001) at resolutions of 60 km across Australia and down to 14 km for Tasmania (McGregor, 2004). These projections use forcing from recent CSIRO AOGCM projections. Downscaled projected climate change using statistical methods has also been recently undertaken for parts of Australia (e.g., Timbal and McAvaney, 2001; Charles et al., 2004; Timbal, 2004;) and New Zealand (Mullan et al., 2001a; Ministry for the Environment, 2004).

11.7.3.1.1 Mean temperature

In both the southern and northern Australia regions, the projected MMD-A1B warming in the 21st century represents a significant acceleration on warming over that observed in the 20th Century (Figure 11.16). The warming is larger than the surrounding oceans, but only comparable to, or slightly larger than the global mean warming. Averaging over the region south of 30°S (SAU), the median 2100 warming among all of the models is 2.6 ^{o}C (with an interquartile range of 2.4 to 2.9 ^{o}C) whereas the median warming averaged over the region north of 30°S (NAU) is 3.0 ^{o}C (range of 2.8 to 3.5 ^{o}C). The seasonal cycle in the warming is weak, but with larger values (and larger spread amongst model projections) in summer (DJF). Across the MMD models, the warming is well correlated with the global mean warming, with a correlation of 0.79, so that more than half of the variance among models is controlled by global rather than local factors, as in many other regions. The range of responses is comparable but slightly smaller than the range in global mean temperature responses and warming over equivalent time periods under the B1, A1B, and A2 scenarios is close to the ratios of the global mean responses The warming varies subregionally, with less warming in coastal regions, Tasmania, and the South Island of New Zealand, and greater warming in Central and Northwest Australia (see Chapter 10, Figure 10.8).

[INSERT FIGURE 11.16 HERE]

These results are broadly (and in many details) similar to those described in earlier studies, so other aspects of these earlier studies can be assumed to remain relevant. For the CSIRO (2001) projections, pattern-scaling methods were used to provide patterns of change rescaled by the range of global warming given by IPCC (2001) for 2030 and 2070 based on the SRES scenarios. By 2030, the warming is 0.4 to 2°C over most of Australia, with slightly less warming in some coastal areas and Tasmania, and slightly more warming in the northwest. By 2070, annual average temperatures increase by 1 to 6°C over most of Australia with spatial variations similar to those for 2030. Dynamically downscaled mean temperature change typically does not differ very significantly from the picture based on AOGCMs (e.g., see Whetton et al., 2002). Projected warming over New Zealand (allowing for the IPCC (2001) range of global warming and differences in the regional results of six GCMs used for downscaling) is 0.2 to 1.3°C by the 2030s and 0.5 to 3.5°C by the 2080s (Ministry for the Environment, 2004).

11.7.3.2 Mean Precipitation

[INSERT FIGURE 11.17 HERE]

A summary of projected precipitation changes from the MMD models is presented in Figure 11.17 and Table 11.1. The most robust feature is the reduction in rainfall along the south coast in JJA (not including Tasmania) and in the annual mean and a decrease is also strongly evident in SON. The percentage JJA change in 2100 under the A1B scenario for Southern Australia has an interquartile range of –26% to –7% and by comparison the same range using the probabilistic method of Tebaldi et al (2004) is –13% to –6% (Table S11.2). There are large reductions to the south of the continent in all seasons, due to the poleward movement of the westerlies and embedded depressions (Cai et al., 2003b; Yin, 2005; Chapter 10), but this reduction extends over land during winter when the storm track is placed furthest equatorward. Due to polewards drift of the storm track as it crosses Australian longitudes, the strongest effect is in the Southwest, where the ensemble-mean drying is in the 15–20% range. Hope (2006a, b) has shown a southward or longitudinal shift in storms away form southwestern Australia in the MMD simulations. To the east of Australia and over New Zealand, the primary storm track is more equatorward, and the north/south drying/moistening pattern associated with the poleward displacement is shifted equatorward as well. The result is a robust projection of increased rainfall in the South Island (especially its southern half), possibly accompanied by a decrease in the north part of the North Island. The South Island increase is likely to be modulated by the strong topography (see Box 11.3) and to appear mainly up wind of the main range.

Other aspects of simulated precipitation change appear less robust. On the east coast of Australia, there is a tendency in the models for an increase in rain in the summer and a decrease in winter, with a slight annual decrease. However, consistency amongst the models on these features is weak.

These results are broadly consistent with results based on earlier GCM simulations. In the CSIRO (2001) projections based on a range of nine simulations, projected ranges of annual average rainfall change tend toward decrease in the south-west and south but show more mixed results elsewhere (Whetton et al., 2005). Seasonal results showed that rainfall tended to decrease in southern and eastern Australia in winter and spring, increase inland in autumn and increase along the east coast in summer. Moise et al. (2005) also found a tendency for winter rainfall decreases across southern Australia and a slight tendency for rainfall increases in eastern Australia in 18 CMIP2 simulations under 1% per year CO_2 increase.

Whetton et al. (2001) demonstrated that inclusion of high-resolution topography could reverse the simulated direction of rainfall change in parts of Victoria (see Box 11.3). In a region of strong rainfall decrease as simulated directly by the GCMs, two different downscaling methods (Charles et al., 2004; Timbal, 2004) have been applied to obtain the characteristics of rainfall change at stations (Timbal, 2004; IOCI, 2002, 2005). The downscaled results continued to show the simulated decrease, although the magnitude of the changes was moderated relative to the GCM in the Timbal (2004) study. Downscaled rainfall projections for New Zealand (incorporating differing results of some six GCMs) showed a strong variation across the Islands (Ministry for the Environment, 2004). The picture that emerges is that the pattern of precipitation changes described above in the global simulations is still present, but with the precipitation changes focused on the upwind sides of the islands, with the increase in rainfall in the South concentrated in the West, and the decrease in the North concentrated in the East.

11.7.3.3 Snow Cover

The likelihood that precipitation will fall as snow will decrease as temperature rises. Hennessy et al. (2003) modelled snowfall and snow cover in the Australian Alps under the CSIRO (2001) projected temperature and precipitation changes, and obtained very marked reductions in snow. The total alpine area with at least 30 days of snow cover decreases 14–54% by 2020, and 30–93% by 2050. Because of projected increased winter precipitation over the Southern Alps, it is less clear that mountain snow will be reduced in New Zealand (Ministry for the Environment, 2004, see also Box 11.3). However, marked decreases on average snow water over New Zealand (60% by 2040 under the A1B scenario) have been simulated by Ghan and Shippert (2006) using a high-resolution subgrid-scale orography in a global model that simulates little change in precipitation.

11.7.3.4 Potential Evaporation

Using the method of Walsh et al. (1999) changes to potential evaporation in the Australian region have been calculated for a range of enhanced greenhouse climate model simulation (Whetton et al., 2002; McInnes et al., 2003; Hennessy et al., 2004a; McInnes et al., 2004; Hennessy et al., 2004b; Cai et al., 2003a;). In all cases increases in potential evaporation were simulated, and in almost all cases the moisture balance deficit became larger. This has provided a strong indication of the Australian environment becoming drier under enhanced greenhouse conditions.

11.7.3.5 Temperature and Precipitation Extremes

Where the analysis has been done for Australia (e.g., Whetton et al., 2002) the effect on changes in extreme temperature due to simulated changes in variability is small relative to the effect of the change in the mean. Therefore, most regional assessments of changes in extreme temperatures have been based on adding a projected mean temperature change to each day of a station observed data set. Based on the CSIRO (2001) projected mean temperature change scenarios, the average number of days over 35°C each summer in Melbourne would increase from 8 at present to 9–12 by 2030 and 10–20 by 2070 (CSIRO, 2001). In Perth, such hot days would rise from 15 at present to 16–22 by 2030 and 18–39 by 2070 (CSIRO, 2001). On the other hand, cold days become much less frequent. For example, Canberra's current 44 winter days of minimum temperature below zero is projected to be 30–42 by 2030 and 6–38 by 2070 (CSIRO, 2001).

Changes in extremes in New Zealand have been assessed using a similar methodology and simulations (Mullan et al., 2001b). Decreases in the frequency of days below zero of 5–30 days per year by 2100 are projected for New Zealand, particularly for the lower North Island and the South Island. Increases in the number of days above 25°C of 10–50 days per year by 2100 are projected.

A range of GCM and regional modelling studies in recent years have identified a tendency for daily rainfall extremes to increase under enhanced greenhouse conditions in the Australian region (e.g., Hennessy et al., 1997; Whetton et al., 2002; Watterson and Dix, 2003; Suppiah et al., 2004; McInnes et al., 2003; Hennessy et al., 2004b Kharin and Zwiers, 2005). Commonly return periods of extreme rainfall events halve in late 21st century simulations. This tendency can apply even when average rainfall is simulated to decrease, but not necessarily when this decrease is marked (see Timbal, 2004). Recently Abbs (2004) dynamically downscaled to a resolution of 7 km current and enhanced greenhouse cases of extreme daily rainfall occurrence in northern NSW and southern Queensland as simulated by the CSIRO GCM. The downscaled extreme events for a range of return periods compared well with observations and the enhanced greenhouse simulations for 2040 showed increases of around 30% in magnitude, with 1 in 40 year event becoming the 1 in 15 year event. Less work has been done on projected changes to rainfall extremes in New Zealand, although the recent analysis of Ministry for the Environment (2004) based on Semenov and Bengtsson (2002) indicates the potential for extreme winter rainfall (95% percentile) to change by between –6% and +40%.

Where GCMs simulate a decrease in average rainfall it may be expected that there would be an increase in the frequency of dry extremes (droughts). Whetton and Suppiah (2003) examined simulated monthly frequencies of serious rainfall deficiency for Victoria, which showed strong average rainfall decrease in most simulations considered. There was a marked increase in the frequency of rainfall deficiencies in most simulations, with doubling in some cases by 2050. Using a slightly different approach, likely increases in the frequency of drought have also been established for the states of South Australia, NSW and Queensland (McInnes et al., 2003; Walsh et al., 2002; Hennessy et al., 2004c). Mullan et al. (2005) has shown that by the 2080s in New Zealand, there may be significant increase in drought frequency in the east of both islands.

11.7.3.6 Tropical Cyclones

There have been a number of recent regional model-based studies of changes in tropical cyclone behaviour in the Australian region (e.g., Walsh and Katzfey, 2000; Walsh and Ryan, 2000; Walsh et al., 2004), which have examined aspects of number, tracks and intensities under enhanced greenhouse conditions. There is no clear picture with respect to regional changes in frequency and movement, but increases in intensity are indicated. For example Walsh et al. (2004) obtained, under 3 × CO_2 conditions, a 56% increase in storms of maximum wind speed of greater than $30ms^{-1}$. It should also be noted that ENSO fluctuations have a strong impact on patterns of tropical cyclone occurrence in the region, and that therefore uncertainty with respect future ENSO behaviour (see Chapter 10, Section 10.3) contributes to uncertainty with respect tropical cyclone behaviour (Walsh, 2004). Also see Chapter 10, Section 10.3.6.3 for a global assessment of changes to tropical cyclone characteristics.

11.7.3.7 Winds

The ensemble mean projected change in wintertime sea level pressure may be seen in Chapter 10, Figure 10.9 based on the MMD simulations. Much of Australia lies to the north of the centre of the high-pressure anomaly. With the mean latitude of maximum pressure near 30°S at this season this corresponds to a modest strengthening of the mean wind over inland and northern areas and a slight weakening of the mean westerlies on the southern coast, consistent with Hennessy et al. (2004b). Studies of daily extreme winds in the region using high-resolution model output (McInnes et al., 2003) indicated increases of up to 10% across much of the northern half of Australia and the adjacent oceans during summer by 2030. In winter, the pressure gradient is simulated to increase over the South Island of New Zealand (see Chapter 10, Figure 10.9) implying increased windiness. This increase is present in all of the MMD-A1B projections.

11.8 Polar

Assessment of projected climate change for Polar regions:

> The Arctic is very likely to warm during this century in most areas, and the annual mean warming is very likely to exceed the global mean warming. Warming is projected to be largest in winter and, in the Arctic, smallest in summer.
>
> Annual Arctic precipitation is very likely to increase. It is very likely that the relative precipitation increase is largest in the winter and smallest in summer.
>
> Arctic sea ice is very likely to decrease in its extent and thickness. It is uncertain how the Arctic Ocean circulation will change.
>
> It is likely that the Antarctic will be warmer and that the precipitation will increase over the continent.
>
> It is uncertain to what extent the frequency of extreme temperature and precipitation events will change in the polar regions.

Polar climate involves a large natural variability on interannual, decadal and longer time scales, which is an important source of uncertainty. The projections of the trends of the underlying teleconnections (like NAM or ENSO) contain substantial uncertainty (see Chapter 10). Further, our understanding of the polar climate system is still incomplete due to its complex atmosphere-land-cryosphere-ocean-ecosystem interactions involving a variety of distinctive feedbacks. Processes that are not particularly well represented in the models are clouds, planetary boundary layer processes, and sea ice. Additionally, the resolution of global models is still not adequate to resolve important processes in the polar seas. All this contributes to a rather large range of present-day and future simulations, which may reduce our confidence in the future projections. A serious problem is the lack of observations against which to assess models, and for developing process knowledge, particularly over Antarctica.

11.8.1 Arctic

11.8.1.1 Key Processes

The Arctic climate is characterized by a distinctive complexity due to numerous nonlinear interactions between and within the atmosphere, cryosphere, ocean, land, and ecosystems. Sea ice plays a crucial role in the Arctic climate, particularly through its albedo. Reduction of ice extent leads to warming due to increased absorption of solar radiation at the surface. Substantial low-frequency variability is evident in various atmosphere and ice parameters (Polyakov et al., 2003a, b), complicating the detection and attribution of

Arctic changes. Natural multi-decadal variability has been suggested as partly responsible for the large warming in the 1920s–1940s (Johannessen et al., 2004; Bengtsson et al., 2004) followed by cooling until the 1960s. In both models and observations, the interannual variability of monthly temperatures is a maximum in high latitudes (Räisänen, 2002). Natural atmospheric patterns of variability on annual and decadal time scales play an important role in the Arctic climate. Such patterns include the NAM, NAO, PNA and the PDO, which are associated with prominent Arctic regional precipitation and temperature anomalies (see Chapter 3, Box 3.4 and Section 3.6). For instance, the positive NAM/NAO phase is associated with warmer/wetter winters in Siberia and colder/drier winters in western Greenland and north-eastern Canada. NAM/NAO showed a trend towards its positive phase over the last 3–4 decades, although it returned to near its long-term mean state in the last five years (see Chapter 3, Section 3.6). In the future, global models project a positive trend in the NAO/NAM during the 21st century (see Chapter 10, Section 10.3). There is substantial decadal-to-interdecadal atmospheric variability in the North Pacific over the 20th century, associated with fluctuations in the strength of the wintertime Aleutian Low, which co-vary with North Pacific SST in the PDO (see Chapter 3, Section 3.6). A deeper and eastward shifted Aleutian Low advects warmer and moister air into Alaska. While some studies have suggested that the Brooks Range effectively isolates Arctic Alaska from much of the variability associated with North Pacific teleconnection patterns (e.g., L'Heureux et al., 2004), other studies find relationships between the Alaskan and Beaufort-Chukchi region's climate and North Pacific variability (Stone, 1997; Curtis et al., 1998; Lynch et al., 2004). Patterns of variability in the North Pacific, and their implications for climate change, are especially difficult to sort out due to the presence of several patterns (NAM, PDO, PNA) with potentially different underlying mechanisms (see Chapter 3).

11.8.1.2 Present Climate: Regional Simulation Skill

Many processes are still poorly understood and thus continue to pose a challenge for climate models (ACIA, 2005). In addition, the evaluation of simulations in the Arctic is made difficult by the uncertainty in the observations. The few available observations are sparsely distributed in space and time and different data sets often differ considerably (Serreze and Hurst, 2000; Liu et al., 2005; Wyser and Jones, 2005; ACIA, 2005). This holds especially for precipitation measurements which are problematic in cold environments (Goodison et al., 1998; Bogdanova et al., 2002).

Few pan-Arctic atmospheric RCMs are in use. When driven by analyzed lateral and sea-ice boundary conditions, RCMs tend to show smaller temperature and precipitation biases in the Arctic compared to GCMs, indicating that sea ice simulation biases and biases originating from lower latitudes contribute substantially to the contamination of GCM results in the Arctic (e.g., Dethloff et al., 2001; Wei et al., 2002; Lynch et al., 2003; Semmler et al., 2005). However, even under a very constrained experimental design, there can be considerable across-model scatter in RCM simulations (Tjernström et al., 2005; Rinke et al., 2006). The construction of coupled atmosphere-ice-ocean RCMs for the Arctic is a recent development (Maslanik et al., 2000; Rinke et al., 2003; Debernard et al., 2003; Mikolajewicz et al., 2005).

Temperature

The simulated spatial patterns of the MMD ensemble mean temperatures agree closely with those of the observations throughout the annual cycle. Generally, the simulations are 1–2°C colder than the ERA40 reanalyses with the exception of a cold bias maximum of 6–8°C in the Barents Sea (particularly in winter/spring) caused by overestimated sea ice in this region (Chapman and Walsh, 2006a; see also Chapter 8, Section 8.3). Compared with earlier model versions, the annual temperature simulations improved in the Barents and Norwegian Seas and Sea of Okhotsk, but some deterioration is noted in the central Arctic Ocean and the high terrain areas of Alaska and northwest Canada (Chapman and Walsh, 2006a). The mean model ensemble bias is relatively small compared to the across-model scatter of temperatures. The annual mean root-mean-squared error in the individual MMD models ranges from 2°C to 7°C (Chapman and Walsh, 2006a). Compared with previous models, the MMD simulated temperatures are more consistent across the models in winter, but somewhat less so in summer. There is considerable agreement between the modelled and observed interannual variability both in magnitude and spatial pattern.

Precipitation

The AOGCM simulated monthly precipitation varies substantially among the models throughout the year but the MMD ensemble mean monthly means are within the range of different observational data sets. This is an improvement compared to earlier simulations (Walsh et al., 2002; ACIA, 2005), particularly from autumn to spring (Kattsov et al., 2006). The ensemble mean bias varies with season and remains greatest in spring and smallest in summer. The annual bias pattern (positive over most parts of the Arctic) can be partly attributed to coarse orography and to biased atmospheric storm tracks and sea ice cover (see Chapter 8). The MMD models capture the observed increase of the annual precipitation through the 20th century (see Chapter 3, Section 3.3).

Sea Ice and Ocean

Arctic sea ice biases in present-day MMD simulations are discussed in Chapter 8, Section 8.3. Arctic ocean-sea ice RCMs under realistic atmospheric forcing are increasingly capable of reproducing the known features of the Arctic Ocean circulation and observed sea ice drift patterns. The inflow of the two branches of Atlantic origin via the Fram Strait and the Barents Sea and their subsequent passage at mid-depths in several cyclonic circulation cells are present in most recent simulations (Karcher et al., 2003; Maslowski et al., 2004; Steiner et al., 2004). Most of the models are biased towards overly salty values in the Beaufort Gyre and thus too little fresh water storage in the Arctic halocline. Several potential causes have been identified, among them a biased simulation of Arctic shelf processes and wind forcing. Most hindcast simulations with these RCMs show a reduction in the Arctic ice volume over recent decades (Holloway and Sou, 2002).

11.8.1.3 Climate Projections

Temperature

A northern high-latitude maximum in the warming (“polar amplification”) is consistently found in all AOGCMs (see Chapter 10, Section 10.3). The simulated annual mean Arctic warming exceeds the global mean warming by roughly a factor of two in the MMD models, while the wintertime warming in the central Arctic is a factor of 4 larger than the global annual mean when averaged over the models. These magnitudes are comparable to those obtained in previous studies (Holland and Bitz, 2003, ACIA, 2005). The consistency between observations and the ensemble mean 20th century simulations (Figure 11.18), combined with the fact that the near future projections (2010–2029) continue the late 20th century trends in temperature, ice extent and thickness with little modification (Serreze and Francis, 2006), increases confidence in this basic polar amplified warming pattern, despite the inter-model differences in the amount of polar amplification.

[INSERT FIGURE 11.18 HERE]

At the end of the 21st century, the projected annual warming in the Arctic is 5°C, estimated by the MMD-A1B ensemble mean projection (Figure 11.21). There is a considerable across-model range of 2.8–7.8°C (Table 11.1). Larger (smaller) mean warming is found for the A2 (B1) scenario with 5.9°C (3.4°C), with a proportional across-model range. The across-model and across-scenario variability in the projected temperatures are both considerable and of comparable amplitude (Chapman and Walsh, 2006a).

Both over ocean and land, the largest (smallest) warming is projected in winter (summer) (Table 11.1, Figure 11.19). But, the seasonal amplitude of the temperature change is much larger over ocean than over land due the presence of melting sea ice in summer keeping the temperatures close to the freezing point. The surface air temperature over the Arctic Ocean region is generally warmed more than over Arctic land areas (except in summer). The range between the individual simulated changes remains large (Figure 11.19, Table 11.1). By the end of the century, the mean warming ranges from 4.3°C to 11.4°C in winter, and from 1.2°C to 5.3°C in summer under the A1B scenario. The corresponding 5th to 95th confidence intervals are given in Supplementary Material Table S11.2. In addition to the overall differences in global warming, difficulties in simulating sea ice, partly related to biases in the surface wind fields, as well as deficiencies in cloud schemes, are likely responsible for much of the inter-model scatter. Internal variability plays a secondary role when examining these late 21st century responses.

[INSERT FIGURE 11.19 HERE]

The annual mean temperature response pattern at the end of the 21st century under the A1B scenario (Supplementary Material Figures S11.27 and S11.11) is characterized by a robust and large warming over the central Arctic Ocean (5–7°C), dominated by the warming in winter/autumn associated with the reduced sea ice. The maximum warming is projected over the Barents Sea though this could result from an overestimated albedo feedback caused by removal of the present-day simulations' excessive sea ice cover. A region of reduced warming (<2°C, even slight cooling in several models) is projected over the northern North Atlantic, which is consistent among the models. This is due to weakening of the MOC (see Chapter 10, Section 10.3).

While the natural variability in Arctic temperatures is large compared to other regions, the signals are still large enough to emerge quickly from the noise (Table 11.1). Looking more locally, as described by Chapman and Walsh (2006a), Alaska is perhaps the land region with the smallest signal-to-noise ratio, and is the only Arctic region in which the 20-year-mean 2010–2029 temperature is not clearly discernible from the 1981–2000 mean in the MMD models. But even here the signal is clear by mid-century in all three scenarios.

The regional temperature responses are modified by changes in circulation patterns (Chapter 10). In winter, shifts in NAO phase can induce interdecadal temperature variations of up to 5 K in the Eastern Arctic (Dorn et al., 2003). The MMD models project winter circulation changes consistent with an increasingly positive NAM/NAO (see Chapter 10, Section 10.3), which acts to enhance the warming in Eurasia and western North America. In summer, circulation changes are projected to favour warm anomalies north of Scandinavia and extending into the eastern Arctic, with cold anomalies over much of Alaska (Cassano et al., 2006). However, deficiencies in the Arctic summertime synoptic activity in these models reduce our confidence in the detailed spatial structure. Also, these circulation-induced temperature changes are not large enough to change the relatively uniform summer warming seen in the MMD models

The patterns of temperature changes simulated by RCMs are quite similar to those simulated by GCMs. However, they show an increased warming along the sea ice margin possibly due to a better description of the mesoscale weather systems and air-sea fluxes associated with the ice edge (ACIA, 2005). The warming over most of the central Arctic and Siberia, particular in summer, tends to be lower in RCMs (by up to 2 K) probably due to more realistic present-day snow pack simulations (ACIA, 2005).The warming is modulated by the topographical height, snow cover and associated albedo feedback as shown for the region of northern Canada and Alaska (see Section 11.5.3).

Precipitation
The MMD simulations show a general increase in precipitation over the Arctic at the end of the 21st century (Table 11.1; Supplementary Material Figure S11.28). The precipitation increase is robust among the models (Table 11.1; Supplementary Material Figure S11.19) and qualitatively well understood, attributed to the projected warming and related increased moisture convergence (Chapter 10, Section 10.3). The very strong correlation between the temperature and precipitation changes (~5% precipitation increase per degree warming) across the model ensemble is worth noting (Figure 11.20). Thus, both the sign and the magnitude (per degree warming) of the percentage precipitation change are robust among the models.

[INSERT FIGURE 11.20 HERE]

The spatial pattern of the projected change (Supplementary material Figure S11.28) shows greatest percentage increase over the Arctic Ocean (30–40%) and smallest (and even slight decrease) over the northern North Atlantic (<5%). By the end of the 21st century, the projected change in the annual mean Arctic precipitation varies from 10% to 28%, with an MMD-A1B ensemble median of 18% (Table 11.1). Larger (smaller) mean precipitation increase is found for the A2 (B1) scenario with 22% (13%). The percentage precipitation increase is largest in winter and smallest in summer, consistent with the projected warming (Figure 11.19; Table 11.1). The across-model scatter of the precipitation projections is substantial (Figure 11.19; Table 11.1). The Tebaldi et al. (2005) 5th to 95th quantile confidence interval of percentage precipitation change in winter is 13–36% and in summer 5–19% (Supplementary Material Table S11.2).

Differences between the projections for different scenarios are small in the first half of the 21st century but increase later. Differences among the models increase rapidly as the spatial domain becomes smaller (ACIA, 2005). The geographical variation of precipitation changes is determined largely by changes in the synoptic circulation patterns. During winter, the MMD models project a decreased (increased) frequency of strong Arctic high (Icelandic low) pressure patterns which favour precipitation increases along the Canadian west coast, southeast Alaska and North Atlantic extending into Scandinavia (Cassano et al., 2006). Projections with RCMs support the broad-scale messages whilst adding expected local and regional detail (ACIA, 2005).

By the end of the 21st century, the MMD-A1B ensemble projected precipitation increase is significant (Table 11.1), particularly the annual and cold season (winter/autumn) precipitation. However, local precipitation changes in some regions and seasons (particularly in the Atlantic sector and generally in summer) remain difficult to discern from natural variability (ACIA, 2005).

Extremes of Temperature and Precipitation
Very little work has been done in analyzing future changes in extreme events in the Arctic. However, the MMD simulations indicate that the increase in mean temperature and precipitation will be combined with an increase in the frequency of very warm and wet winters and summers. Using the definition of extreme season in Section 11.1.2, every DJF and JJA season, in all model projections, is "extremely" warm in the period 2080–2099 (Table 11.1). The corresponding numbers for extremely wet seasons are 90% and 85% for DJF and JJA. For the other scenarios, the frequency of extremes is very similar, except that for the wet seasons under B1 which is smaller (~63%).

Cryosphere
Northern Hemisphere sea ice, snow, and permafrost projections are discussed in Chapter 10, Section 10.3; projected changes in the surface mass balance of Arctic glaciers and of the Greenland ice sheet are discussed in Chapter 10, Sections 10.3 and 10.7.

Arctic Ocean
A systematic analysis of future projections for the Arctic Ocean circulation is still lacking. Coarse resolution in global models prevents the proper representation of local processes that are of global importance (such as the convection in the Greenland Sea which impacts the deep waters in the Arctic Ocean and the intermediate waters that form overflow waters). The MMD models project a reduction in the MOC in the Atlantic Ocean (see Chapter 10, Section 10.3). Correspondingly, the northward oceanic heat transport decreases south of 60°N in the Atlantic. However, CMIP model assessment showed a projected increase of the oceanic heat transport at higher latitudes, associated with a stronger sub-Arctic gyre circulation in the models (Holland and Bitz, 2003). The Atlantic Ocean north of 60°N freshens during the 21st century, in pronounced contrast to the observed development in the late 20th century (Wu et al., 2003).

11.8.2 Antarctic

11.8.2.1 Key Processes

Over Antarctica, there is special interest in changes in accumulation of snow expected to accompany global climate change as well as the pattern of temperature change, particularly any differences in warming over the peninsula and the interior of the ice sheet. As in the Arctic, warming of the troposphere is expected to increase precipitation. However, circulation changes in both ocean and atmosphere can alter the pattern of air masses which would modify both precipitation and temperature patterns substantially over the region.

The dominant patterns controlling the atmospheric seasonal to interannual variability of the Southern Hemisphere extra-tropics are the SAM and ENSO (see Chapter 3, Section 3.6). Signatures of these patterns in the Antarctic have been revealed in many studies (reviews by Carleton, 2003 and Turner, 2004). The positive phase of the SAM is associated with cold anomalies over most of Antarctica and warm anomalies over the Antarctic Peninsula (Kwok and Comiso, 2002a). Over recent decades, a drift towards the positive phase in the SAM is evident (see Chapter 3, Section 3.6). Observational studies have presented evidence of pronounced warming over the Antarctic Peninsula, but little changes over the rest of the continent during the last half of the 20th century (see Chapter 3, Section 3.6; Chapter 4, Section 4.6). The response of the SAM in transient warming simulations is a robust positive trend but the response to the ozone hole in the late 20th century, which is also a positive perturbation to the SAM, makes any simple extrapolation of current trends into the future uncertain (see Chapter 10, Section 10.3).

Compared to the SAM, the Southern Oscillation (SO) shows weaker association with surface temperature over Antarctica but the correlation with SST and sea ice variability in the Pacific sector of the Southern Ocean is significant (e.g., Kwok and Comiso, 2002b; Renwick, 2002; Yuan, 2004; Bertler et al., 2004). Correlation between the SO index and Antarctic precipitation/accumulation has also been studied but the persistence of the signal is not clear (Bromwich et al., 2000; Genthon and Cosme, 2003; Guo et al., 2004; Bromwich et al., 2004a; Genthon et al., 2005). Recent work suggests that this intermittence is due to nonlinear interactions between ENSO and SAM that vary on decadal time scales (Fogt and Bromwich, 2006; L'Heureux and Thompson, 2006). The SO index has a negative trend over recent decades (corresponding to a tendency towards more El-Nino conditions in the Equatorial Pacific; see Chapter 3, Section 3.6) associated with sea ice cover anomalies in the Pacific sector, namely negative (positive) anomalies in the Ross and Amundsen Seas (Bellingshausen and Weddell Seas) (Kwok and Comiso, 2002a). However, a definitive assessment of ENSO amplitude and frequency changes in the 21st century cannot be made (see Chapter 10).

11.8.2.2 Present Climate: Regional Simulation Skill

Evaluating temperature and precipitation simulations over Antarctica is difficult due to sparse observations and often relies on numerical weather prediction (re)analyses. However, significant differences between those have been found, and comparisons with station observations show that the surface temperature can be subject of considerable biases (Connolley and Harangozo, 2001; Bromwich and Fogt, 2004). Marked improvement in the bias is seen after the satellite era (~1978) (Simmons et al., 2004), and parts of the bias is explained by the reanalyses' smoothing of the sharp changes in the terrain near coastal stations. Satellite-derived monthly surface temperatures agree with Antarctic station data with an accuracy of 3°C (Comiso, 2000). Precipitation evaluation is even more challenging and the different (re)analyses differ significantly (Connolley and Harangozo, 2001; Zou et al., 2004). Very few direct precipitation gauge and detailed snow accumulation data are available, and these are uncertain to varying degrees (see Chapter 4, Section 4.6).

Major challenges face the simulation of the atmospheric conditions and precipitation patterns of the polar desert in the high interior of East Antarctica (Guo et al., 2003; Bromwich et al., 2004a; Pavolonis et al., 2004, Van de Berg et al., 2005). Driven by analyzed boundary conditions, RCMs tend to show smaller temperature and precipitation biases in the Antarctic compared to the GCMs (Bailey and Lynch, 2000; Van Lipzig et al., 2002a,b; Van den Broeke and Van Lipzig, 2003; Bromwich et al., 2004b; Monaghan et al., 2006). Krinner et al. (1997) show the value of a stretched model grid with. higher horizontal resolution over the Antarctic as compared to standard GCM formulations. Despite these promising developments, since TAR there has been no coordinated comparison of the performance of GCMs, RCMs and other alternatives to global GCMs over Antarctica.

Temperature

Compared to NCEP reanalyses, the MMD ensemble annual surface temperatures are in general slightly warmer in the Southern Ocean to the north of the sea ice region. The mean bias is predominantly less than 2°C (Carril et al., 2005) which may indicate a slight improvement compared to previous models due to better simulation of the position and depth of the Antarctic trough (Carril et al., 2005; Raphael and Holland, 2006). The temperature bias over sea ice is larger. Biases over the continent are several degrees where the model topography is erroneous (Turner et al., 2006). However as emphasized above, the biases have to be seen in the context of the uncertainty in the observations. Changes in cloud and radiation parameterizations have been shown to change the temperature simulation significantly (Hines et al., 2004). A lateral nudging of a stretched-grid GCM (imposing the correct synoptic cyclones from 60°S and lower latitudes) brings the model in better agreement with observations but significant biases remain (Genthon et al., 2002).

The spread in the individual MMD simulated patterns of surface temperature trends in the past 50 years is very large, but in contrast to previous models, the multi-model composite of the MMD models qualitatively captures the observed enhanced warming trend over the Antarctic Peninsula (Chapman and Walsh, 2006b). The general improvements in resolution, sea ice models and cloud-radiation packages have evidently contributed to improved simulations. The ensemble-mean temperature trends show similarity to the observed spatial pattern of the warming, for both annual and seasonal trends. For the annual trend, this includes the warming of the peninsula and near coastal Antarctica and neutral or slight cooling over the sea ice covered regions of the Southern Ocean. While the large spread among the models is not encouraging, this level of agreement suggests that some confidence in the ensemble mean 21st century projection is appropriate.

Precipitation
The MMD models simulate the position of the storm tracks reasonably well but nearly all show some deficiency in the distribution and level of cyclone activity compares to reanalyses (see Chapter 8, Section 8.3). RCMs generally capture the cyclonic events affecting the coast and the associated synoptic variability of precipitation with more fidelity (Adams, 2004; Bromwich et al., 2004a). Over the 20th century, the MMD models simulate changes in storm track position that are generally consistent with observed changes (i.e., poleward displacement of the storm tracks; see Chapter 9, Section 9.5 and Chapter 10, Section 10.3).

The precipitation simulations contain uncertainty both in GCMs and RCMs, on all timescales (Covey et al., 2003; Bromwich et al., 2004a, b; Van de Berg et al., 2005) as a result of model physics limitations. All atmospheric models, including the models underlying the reanalyses, have incomplete parameterizations of polar cloud microphysics and ice-crystal precipitation. The simulated precipitation depends, among others things, on the simulated sea ice concentrations and is strongly affected by biases in the sea ice simulations (Weatherly, 2004). Recent RCM simulations driven by observed sea ice conditions demonstrate good precipitation skill (Monaghan et al., 2006; Van de Berg et al., 2005). However as emphasized above, the observational uncertainty contributes to uncertainty in the differences between observations and simulations.

Sea Ice
The performance biases of Southern Hemisphere sea ice conditions in present-day MMD simulations are discussed in Chapter 8, Section 8.3.

11.8.2.3 Climate Projections

Very little effort has been spent to model the future climate of Antarctica at a spatial scale finer than that of GCMs.

Temperature
At the end of the 21st century, the annual warming over the Antarctic continent is moderate but significant (Figure 11.21; Table 11.1; Chapman and Walsh, 2006b). It is estimated to be 2.6°C by the median of the MMD-A1B models with a range from 1.4 to 5.0°C across the models (Table 11.1). Larger (smaller) warming is found for the A2 (B1) scenario with mean value of 3.1°C (1.8°C). These warming magnitudes are similar to previous estimates (Covey et al., 2003). The annual mean MMD model projections show a relatively uniform warming over the entire continent (with a maximum in the Weddell Sea) (Figure 11.21; Carril et al., 2005; Chapman and Walsh, 2006b). They do not show a local maximum warming over the Antarctic Peninsula. This is a robust feature among the individual models (Supplementary Material Figure S11.12). Thus, the pattern of observed temperature trend in the last half of the 20th century (warming over the Antarctic Peninsula, little changes over the rest of the continent) is not projected to continue throughout the 21st century, despite a projected positive SAM trend (see Chapter 10, Section 10.3). It has been argued that two distinct factors have contributed to the observed SAM trend, greenhouse gas forcing and the ozone hole formation (Stone et al., 2001; Shindell and Schmidt, 2004). Their relative importance for the peninsular warming is not readily understood (see Chapter 10).

[INSERT FIGURE 11.21 HERE]

The mean Antarctic temperature change does not show a strong seasonal dependency; the MMD-A1B ensemble mean winter (summer) warming is 2.8°C (2.6°C) (Table 11.1; Supplementary Material Figure S11.29; Chapman und Walsh, 2006b). This is also illustrated by how close the Tebaldi et al. (2005) 5th to 95th confidence interval for the two seasons is: 0.1–5.7°C in summer and 1.0–4.8°C in winter (Supplementary Material Table S11.2). However over the Southern Oceans, the temperature change is larger in winter/autumn than in summer/spring, which can primarily be attributed to the sea ice retreat (see Chapter 10, Section 10.3).

Precipitation

Almost all MMD models simulate a robust precipitation increase in the 21st century (Supplementary Material Figures S11.29 and S11.30; Table 11.1). However, the scatter among the individual models is considerable. By the end of the 21st century, the projected change in the annual precipitation over the Antarctic continent varies from –2% to 35%, with a MMD-A1B ensemble median of 14% (Table 11.1). Similar (smaller) mean precipitation increase is found for the A2 (B1) scenario with values of 15% (10%). The spatial pattern of the annual change is rather uniform (Supplementary material Figure S11.30). The projected relative precipitation change shows a seasonal dependency, and is larger in winter than in summer (Supplementary material Figure S11.29). The Tebaldi et al. (2005) 5th to 95th confidence interval for winter is –1 to 34% and in summer –6 to 22% (Supplementary Material Table S11.2). The projected increase in precipitation over Antarctica and thus greater accumulation of snow, without substantial surface melting, will contribute negatively to sea level rise relative to present-day (see Chapter 10, Section 10.6). It is notable that the most recent model studies of Antarctic precipitation show no significant contemporary trends (Monaghan et al., 2006; Van de Berg et al., 2005; Van den Broeke et al., 2006) (see Chapter 4, Section 4.6).

The moisture transport to the continent by synoptic activity represents a large fraction of net precipitation (Noone and Simmonds, 2002; Massom et al., 2004). During summer and winter, a systematic shift towards strong cyclonic events is projected in the MMD models (see Chapter 10, Section 10.3). Particularly the frequency of occurrence of deep cyclones in the Ross Sea to Bellingshausen Sea sector is projected to increase by 20–40% (63%) in summer (winter) by the mid of the 21st century (Lynch et al., 2006). Related to this, the precipitation over the sub-Antarctic seas and Antarctic Peninsula are projected to increase.

Extremes of Temperature and Precipitation

Very little work has been done in analyzing future changes in extreme events in the Antarctic. However, the MMD simulations indicate that the increase in mean temperature and precipitation will be combined with an increase in the frequency of very warm and wet winters and summers. Using the definition of “extreme” seasons provided in Section 11.1.2, the MMD models predict extremely warm seasons in about 85% of all DJF and 83% of all JJA seasons in the period 2080–2099, as averaged over all models (Table 11.1). The corresponding numbers for extremely wet seasons are 34% and 59%. For the B1 scenario, the frequency of extremes is smaller, with little difference between A1B and A2.

Sea ice and Antarctic ice sheet.

Southern hemisphere sea ice projections are discussed in Chapter 10, Section 10.3. The projections of the Antarctic ice sheet surface mass balance are discussed in Chapter 10, Section 10.6.

11.9 Small Islands

Assessment of projected climate change for Small Islands regions:

> Sea levels will likely continue to rise on average during the century around the small islands of the Caribbean Sea, Indian Ocean and Northern and Southern Pacific Oceans. Models indicate that the rise will not be geographically uniform but large deviations among models make regional estimates across the Caribbean, Indian and Pacific Oceans uncertain.
>
> All Caribbean, Indian Ocean, as well as North and South Pacific islands are very likely to warm during this century. The warming is likely to be somewhat smaller than the global, annual mean warming in all seasons.
>
> Summer rainfall in the Caribbean is likely to decrease in the vicinity of the Greater Antilles but changes elsewhere and in winter are uncertain. Annual rainfall is likely to increase in the northern Indian Ocean with increases likely in the vicinity of the Seychelles in DJF, and in the vicinity of the Maldives in JJA while decreases are likely in the vicinity of Mauritius in JJA. Annual rainfall is likely to increase in the equatorial Pacific, while decreases are projected by most models just east of French Polynesia in DJF.

Since AOGCMs do not have sufficiently fine resolutions to see the islands, the projections are given over ocean surfaces rather than over land and very little work has been done in downscaling these projections to individual islands. Assessments are made difficult also because some climatic processes are still not well understood, such as the Midsummer Drought in the Caribbean and the ocean-atmosphere interaction in the Indian Ocean. Furthermore, there is insufficient information on future SST changes to determine regional distribution of cyclone changes. Large deviations among models make regional distribution of sea level rise uncertain and the number of models addressing storm surges is very limited.

11.9.1 Key Processes

Climate change scenarios for small islands of the Caribbean Sea, Indian Ocean and Pacific Ocean are included in AR4 for a number of reasons. Ocean-atmosphere interactions play a major role in determining the climate of the islands and including their climate in the projections for neighbours with larger landmasses would miss features peculiar to the islands themselves. Many small islands are sufficiently removed from large landmasses so that atmospheric circulation may be different over the smaller islands compared to their larger neighbours, e.g., in the Pacific Ocean. For the Caribbean that is close to large landmasses in Central America and northern South America, some islands partly share climate features of one, while others partly share features of the other. At the same time the Caribbean islands share many common features that are more important than those shared with the larger landmasses, such as the strong relationship of their climate to sea surface temperature.

11.9.1.1 Caribbean

The Caribbean region spans roughly the area between 10°N to 25°N and 85°W to 60°W. Its climate can be broadly characterized as dry winter/wet summer with orography and elevation being significant modifiers on the sub regional scale (Taylor and Alfero, 2005). The dominant synoptic influence is the North Atlantic subtropical high (NAH). During the winter the NAH is southernmost and the region is generally at its driest. With the onset of the spring, the NAH moves northward, the trade wind intensity decreases and the equatorial flank of the NAH becomes convergent. Concurrently easterly waves traverse the Atlantic from the coast of Africa into the Caribbean. These waves frequently mature into storms and hurricanes under warm sea surface temperatures and low vertical wind shear, generally within a 10–20ºN latitudinal band. They represent the primary rainfall source and their onset in June and demise in November roughly coincides with the mean Caribbean rainy season. In the coastal zones of Venezuela and Columbia, the wet season occurs later, from October to January (Martis et al., 2002). Inter annual variability of the rainfall is influenced mainly by ENSO events through their effect on sea surface temperatures in the Atlantic and Caribbean basins. The late rainfall season tends to be drier in El Niño years and wetter in La Niña years (Giannini et al., 2000, Martis et al., 2002, Taylor et al., 2002) and tropical cyclone activity diminishes over the Caribbean during El Niño summers (Gray, 1984). However the early rainfall season in the Central and Southern Caribbean tends to be wetter in the year after an El Niño and drier in a La Niña year (Chen and Taylor, 2002). The phase of the NAO modulates the behaviour of warm ENSO events (Giannini et al., 2001). A positive NAO implies a stronger than normal NAH and amplifies the drying during a warm ENSO. On the other hand negative NAO amplifies the precipitation in the early rainfall season in the year after an El Niño.

11.9.1.2 Indian Ocean

The Indian Ocean region refers to the area between 35°S to 17.5°N and 50°E to 100°E. The climate of the region is influenced primarily by the Asian monsoons (See section 11.4.1 for processes influencing monsoons). During January the Inter-Tropical Convergence Zone (ITCZ), is located primarily in the southern hemisphere. The region north of the ITCZ then experiences northeasterly trade winds (northeast monsoons) and that to the south, the southeasterly trades. During northern summer the ITCZ is located in the north and virtually covers the entire Bay of Bengal, the surrounding lands, and the eastern Arabian Sea. The winds in the north turn into strong southwesterlies (southwest monsoons), while the southeasterlies persist in the south. Precipitation and wind stress bring about a response that is distinctly different in the northern and southern parts of the Indian Ocean (International CLIVAR Project Office, 2006**)**. The wet (dry) season in the Maldives occurs during the southwest (northeast) monsoons. From May to October the southeast trades dominate in the Seychelles and the climate is relatively cool and dry, and December to March is the principal wet season with winds mainly from west to northwest.

While the monsoons recur each year, their irregularity at a range of time-scales from weeks to years depends on feedback from the ocean in ways that are not fully understood. Intraseasonal variability is associated with the Monsoon Intraseasonal Oscillation (MISO) and Madden–Julian Oscillation (MJO), which are long-lasting weather patterns that evolve in a systematic way for periods of four to eight weeks. On an interannual and decadal scale statistical methods have shown that while there are periods of high correlation between ENSO and monsoon variation, there are decades where there appears to be little or no association at all (International CLIVAR Project Office, 2006; this is also discussed in Chapter 10, Section 10.3.5.4). A modulating factor is the Indian Ocean Dipole or Indian Ocean Zonal Mode (IOZM), a large interannual variation in zonal SST gradient, which is discussed in Chaper 3, Section 3.6.. The magnitude of the secondary rainfall maximum from October to December in East Africa is strongly correlated with IOZM events, and the positive phase of IOZM, with higher SSTs in the west, counters the drying effect which ENSO has on monsoon rainfall (Ashok et al., 2001).

11.9.1.3 Pacific

The Pacific region refers to equatorial, tropical and subtropical region of the Pacific in which there is a high density of inhabited small islands. Broadly, it is the region between 20°N and 30°S and 120°E to 120°W. The major climatic processes which play a key role in the climate of this region are the easterly trade winds (both north and south of the equator), the southern hemisphere high pressure belt, the intertropical convergence zone (ITCZ) and the South Pacific Convergence zone (SPCZ, see Vincent, 1994), which extends from the ICTZ near the equator due north of New Zealand south-eastward to at least 21°S, 130°W. The region has a warm, highly maritime climate and rainfall is abundant. The highest rainfall follows the seasonal migration of the ITCZ and SPCZ. Year to year climatic variability in the region is very strongly affected by ENSO events. During El Niño conditions, rainfall increases in the zone northeast of the SPCZ (Vincent, 1994). Tropical cyclones are also a feature of climate of the region, except within ten degrees of the equator, and are associated with extreme rainfall, strong winds and storm surge. Many islands in the region are very low lying, but there are also many with strong topographical variations. In the case of the latter, orographic effects on rainfall amount and seasonal distribution can be strong.

11.9.2 Skill of Models in Simulating Present Climate

The ability of the MMD to simulate present climate in the Caribbean, Indian Ocean and North and South Pacific Ocean is summarized in Supplementary Material Table S11.1. In general the biases in about half of the temperature simulations are less than 1ºC in all seasons, so that the model performances were, on the whole, satisfactory. There were however large spreads in precipitation simulations. During the last decade steady progress has been made in simulating and predicting ENSO using coupled GCMs. However serious systematic errors in both the simulated mean climate and the natural variability persist (See Chapter 8, Section 8.4.7)

11.9.2.1 Caribbean

Simulations of the annual Caribbean temperature in the 20th century (1980–1999) by MMD gave an average that agreed closely with climatology, differing by less than 0.1°C. The deviations of 50% individual models from the climatology ranged from –0.3°C to +0.3°C. Thus the models have good skill in simulating annual temperature. The average of the MMD simulations of precipitation however underestimates the observed precipitation by approximately 30%. The deviations in individual models range from –64% to +20%, much greater than the deviations in temperature simulations. Recently the Parallel Climate Model (T42), a fully coupled global climate model, was found to be capable of simulating the main climate features over the Caribbean region (Angeles et al., 2006), but it also underestimated the area average precipitation across the Caribbean. Martinez-Castro et al., (2006), in a sensitivity experiment, concluded that the regional RegCM3 model, using the Anthes-Kuo cumulus parameterization scheme, can be used for long term area-averaged climatology.

11.9.2.2 Indian Ocean

For annual temperature in the Indian Ocean in the 20th century (1980–1999), the mean value of the MMD outputs overestimated the climatology by 0.6°C, with 50% of deviations ranging from 0.2°C to 1.0°C. For rainfall the model consensus was only slightly below the mean precipitation by 3%, and the model deviations ranged from –22% to +20%. There are however problems with the simulation of year-to-year variation. Many of the important climatic effects of the MJO, including its impacts on rainfall variability in the monsoons, are still poorly simulated in contemporary climate models (See Chapter 8, Section 8.4).

11.9.2.3 Pacific

Climate model simulations of current-climate means of temperature and precipitation were investigated by Jones et al. (2000, 2002) and Lal et al. (2002) for the South Pacific. AOGCMs available at the time of these studies simulated well the broad scale patterns of temperature and precipitation across the region, with the precipitation patterns more variable than for temperature in the models considered, and showing some significantly underestimating or overestimating of the intensity of rainfall in the high rainfall zones. All models simulated a broad rainfall maximum stretching across the SPCZ and ITCZ, but not all models resolved a rainfall minimum between these two regions. A problem of simulating the spatial structure of the MJO resulting in tendencies for the convective anomaly to split into double ITCZs in the Pacific is also discussed in Chapter 8, Section 8.4.8.

Analysis of the MMD simulations shows that the average model value overestimated the mean annual temperature from 1980–1999 by 0.9°C over a southern Pacific region, with 50% of deviations varying from 0.6°C to 1.2°C. Over the North Pacific, the consensus temperature simulation for same the period was only 0.7°C above the climatology, with half of model deviations from climatology ranging from 0.2°C to 1.0°C. Average precipitation was overestimated by 10%, but individual model values varied from –7% to 31% in the southern Pacific region, whereas in the northern Pacific the mean model output for precipitation almost agreed with climatology. The individual models deviated from –13% to 13%. Thus the models were better at simulating rainfall in the northern Pacific than in the southern Pacific and the quality of the simulations, both north and south, were not much different than for the Indian Ocean.

11.9.3 Temperature and Precipitation Projections

Scenarios of temperature change (ºC) and percentage precipitation change from 1980–1999 to 2080–2099 are summarized in Table 11.1, which is described in Section 11.1.3. A small value of T implies a large signal to noise ratio and it can be seen that, in general, the signal to noise ratio is greater for temperature than for precipitation change. The probability of extreme warm seasons is 100% in all cases for the small islands and the scenarios of warming are all very significant by the end of the century. Approximate results for A2 and B1 scenarios and for other future times in this century can obtained by scaling the A1B values, as described in Section 11.1.3.

[INSERT FIGURE 11.22 HERE]

The temporal evolution of temperature as simulated by MMD in the 20th and 21st centuries are also show in Figure 11.22, for oceanic regions including the Caribbean (CAR), Indian Ocean (IND), North Pacific Ocean (NPA) and South Pacific Ocean (SPA). In general it can be seen, by comparison with Box 11.1, Figure 1, that the temperature increases for the small islands are less than for the continental regions. Also seen from the figures is the almost linear nature of the evolution. The ranges for the A2 and B1 scenarios at the end of the 21st century are given by the red and blue vertical lines respectively. Temperature and precipitation projections for the small island regions will be discussed below in the context of Table 11.1.

11.9.3.1 Caribbean

The MMD simulated annual temperature increases at the end of the 21st century ranges from 1.4 to 3.2°C with a median of 2.0°C, somewhat below the global average. Fifty percent of the models give values differing from the median by only ± 0.4°C. Statistical downscaling of HadCM3 results using A2 and B2 emission scenarios gives around 2°C rise in temperature by 2080's, approximately the same as the HadCM3 model. The agreement between the AOGCMs and the downscaling analysis gives a high level of confidence in the temperature simulations. The downscaling was performed with the use of the SDSM model developed by Wilby et al. (2002) as part of an AIACC SIS06 project (http://www.aiaccproject.org). Angeles et al., (2006) also simulated 1ºC rise, approximately, in sea surface temperature up the 2050's using an IS92a scenario. There were no noticeable differences in monthly changes (See Supplementary Material Figure S11.31). Observations suggests that warming is ongoing (Peterson et al., 2002).

According to Table 11.1 most models show decreases in annual precipitation and a few increases, varying from –39% to +11%, with a median of –12%. Figure 11.23 show the mean annual decrease is spread across the entire region. In DJF some areas of increases are noted and in JJA the region wide decrease is enhanced, especially in the region of the Greater Antilles, where the model consensus is also strong. Monthly changes in the Caribbean are shown in Supplementary Material Figure S11.32, which also shows that the simulations for the Caribbean have a greater spread compared to the other oceanic regions (IND, NPA, SPA in S11.32). HadCM3 results downscaled for A2 and B2 emission scenarios using SDSM, also show a near linear decrease in summer precipitation to 2080s for a station in Jamaica. SDSM downscaled results for stations in Barbados and Trinidad however show increases rather than decreases. Thus there is consensus between MMD and the downscaled results for the Greater Antilles in JJA but not for the other islands, and also not on an annual basis. Angeles et al., (2006) also simulated decreases up to the middle of the century in the vicinity of the Greater Antilles but not in the other islands in the late rainfall season. Table 11.1 shows that the decrease in JJA has the largest signal to noise ratio. The decrease is in agreement with the expected drying in the subtropics discussed in Chapter 9, Section 9.5 and Secton 11.1 of this chapter. In multimodel analysis, most models show an increase in NAO (See Chapter 10, Section 10.3), and consensus on temperature changes in the Pacific indicate an El Niño like pattern with higher temperatures in the eastern Pacific (See Chapter 10, Section 10.3.). These conditions are associated with drying in the Caribbean. Observed trends in precipitation are unclear. While Peterson et al., (2002) found no statistically significant trends in mean precipitation amounts from 1950's to 2000, Neelin et al., (2006) note a modest but statistically significant summer drying trend over recent decades in the Caribbean in several observational data sets.

[INSERT FIGURE 11.23 HERE.]

11.9.3.2 Indian Ocean

Based on MMD consensus the annual temperature is projected to increase by about 2.1°C, somewhat below the global average with individual models ranging from 1.4 to 3.7° and at least half of the models giving values quite close to the mean. All models show temperature increases for all month with no significant seasonal variation (Supplementary Material Figure S11.31). Evidence of temperature increases from 1961–1990 in the Seychelles is provided by Easterling et al., (2003) who found that the percentage of time when the minimum temperature was below the 10th percentile is decreasing, and the percentage of time where the minimum temperature exceeded the 90th percentile is increasing. Similar results were obtained for the maximum temperatures. This is consistent with general patterns of warming elsewhere (See Chapter 3).

The annual precipitation changes for individual MMD varied from –2% to 20% with a median of 4% and 50% of the models giving changes for 3% to 5%. Thus there is some level of confidence in the precipitation results although not as high as for temperature. Figure 11.24 shows that the annual increase is restricted mainly to the north Indian Ocean, where the model consensus is greatest especially in the vicinity of the Maldives. In DJF some increases are noted in the south. Model agreement for increases is greatest for the Seychelles in DJF and for the Maldives in JJA. There is also strong consensus for decreases in the vicinity of Mauritius in JJA. Chapter 10, Section 10.3.5 and Section 11.4 discuss changes in monsoon behaviour in a warmer climate. There is an emerging consensus that the effect of enhanced moisture convergence in a warmer atmosphere will dominate over possible weaker monsoonal flows and tropical large-scale circulation in global warming simulations, resulting in increased monsoonal precipitation. Easterling et al., (2003), found evidence that extreme rainfall tended to increase from 1961–1990. (See also Section 11.4.3, Future Projections for South Asia).

[INSERT FIGURE 11.24 HERE.]

11.9.3.3 Pacific

Projected regional temperature changes in the South Pacific based on a range of AOGCMs have been prepared by Lal et al., (2002); Ruosteenoja et al., (2003) and Lal (2004). Jones et al., (2000, 2002) and Whetton and Suppiah (2003) also considered patterns of change. Broadly simulated warming in the South Pacific closely follows the global average warming rate. However there is a tendency in many models for the warming to be a little stronger in the central equatorial Pacific (North Polynesia) and a little weaker to the South (South Polynesia).

The MMD-A1B projections for the period 2080 to 2099 show an increase in annual temperature of 1.8ºC, somewhat below the global average over the South Pacific (Table 11.1). The individual model values vary respectively from 1.4°C to 3.1° and at least half of the models gave values very close to the mean. All model show increases, slightly less in the second half of the year compared to the first (Supplementary Material Figure S11.31). Over the North Pacific, the simulations give an increase in temperature of 2.3°C, slightly below the global average with values ranging from 1.5°C to 3.7°C and 50% of the models within ±0.4°C of the mean. All model show increases, more in the second half of the year compared to the first (Supplementary Material Figure S11.31).

For the same period, 2080 to 2099, annual precipitation increases over the Southern Pacific when averaged over all MMD was close to 3%, with individual models giving values from –4% to +11% and 50% of the models showing increases from 3% to 6%. The time for a discernable signal is relatively low. (Table 11.1). Most of these increases were in the first half of the year (Supplementry Material Figure S11.32) . For precipitation in the Northern Pacific an increase of 5% was found, with individual models giving values from 0% to 19% increases and at least half of the model within -2% and +5% of the median. The time for a discernable signal is relatively large. Most of these increases were in the latter half of the year (Supplementary Material Figure S11.32). Figure 11.25 illustrates the spatial distribution of annual, DJF and JJA rainfall changes and inter-model consistency. It can be seen that the tendency for precipitation increase in the Pacific is strongest in the region of the ITCZ due to increased moisture transport described in Section 11.1.3.1. Griffiths et al., (2003) found that there was in increasing trend from 1961–2000 in mean rainfall in and northeast of the SPCZ in the southern Pacific. As for the Indian Ocean, there is some level of confidence in the precipitation results for the Pacific, but not as high as for the temperature results.

Changes in rainfall variability in the South Pacific were analyzed by Jones et al. (2000) using IPCC (1996) scenarios, but more recent simulations have not been examined. These changes will be strongly driven by changes to ENSO, and this is not well understood (see Chapter 10, Section 10.3).

[INSERT FIGURE 11.25 HERE]

11.9.4 Sea Level Rise

Sea level is projected to rise between the present (1980-1999) and the end of this century (2090-2099) 0.35 m (0.23 to 0.47 m) for the A1B scenario (see Chapter 10, Section 10.6). Due to ocean density and circulation changes, the distribution will not be uniform and Chapter 10, Figure 10.32 shows a distribution in local sea level change based on ensemble mean of 14 AOGCM's. A lower than average rise in the Southern Ocean can be seen possibly due to increased wind stress. Also obvious is a narrow band of pronounced sea-level rise stretching across the southern Atlantic and Indian Oceans at about 40ºS. This is also seen in the southern Pacific at about 30ºS. However large deviations among models make estimates of distribution across the Caribbean, Indian and Pacific Oceans uncertain. Extreme sea level changes, including storm surges, are discussed in Box 11.5 in a broader context. The range of uncertainty cannot be reliably quantified due to the limiting set of models addressing the problem.

Global sea-level rise over the 20th century is discussed in Chapter 5, Section 5.5; the best estimate is 0.17 ± 0.05 m. From estimates of observed sea level rise from 1950 to 2000 by Church et al., (2004), the rise in the Caribbean appeared to be near the global mean. Church et al., (2006) also estimated the average rise in the region of the Indian and Pacific Ocean to be close to the global average. There have been large observed variations in sea-level rise in the Pacific Ocean mainly due to ocean circulations associated ENSO events. From 1993 to 2001, all the data show large rates of sea-level rise over the western Pacific and eastern Indian Ocean and sea level falls in the eastern Pacific and western Indian Ocean (Church et al., 2006). Observed sea level rise in the Pacific and Indian Oceans is discussed in Chapter 5.

11.9.5 Tropical Cyclones

There have been fewer models simulating tropical cyclones in the context of climate change than those simulating temperature and precipitations changes and sea-level rise, mainly because of the computational burden associated with the high resolution needed to capture the characteristics of tropical cyclones. Accordingly there is less certainty about the changes in frequency and intensity of tropical cyclones on a regional basis than for temperature and precipitation changes. An assessment of results for projected changes in tropical cyclones is presented in Chapter 10, Section 10.3.6.3, and a synthesis is given at the end of the section. Regional model-based studies of changes in tropical cyclone behaviour in the southwest Pacific include works by Nguyen and Walsh (2001) and Walsh (2004). Walsh concluded that in general there is no clear picture with respect to regional changes in frequency and movement, but increases in intensity are indicated. It should also be noted that ENSO fluctuations have a strong impact on patterns of tropical cyclone occurrence in the southern Pacific, and that therefore uncertainty with respect future ENSO behaviour (see Chapter 10, Section 10.3) contributes to uncertainty with respect tropical cyclone behaviour (Walsh, 2004).

Box 11.5: Coastal Zone Climate Change

Introduction

Climate change has the potential to interact with the coastal zone in a number of ways including inundation, erosion and salt water intrusion into the water table. Inundation and intrusion will clearly be affected by the relatively slow increases in mean sea level over the next century and beyond. Mean sea level is dealt with in Chapter 10 and here we concentrate on changes in extreme sea level which have the potential to significantly affect the coastal zone. There is insufficient information on changes in waves or near-coastal currents to provide an assessment of effects of climate change on erosion.

The characteristics of extreme sea level events are dependent on the atmospheric storm intensity and movement and coastal geometry. In many locations, the risk of extreme sea levels is poorly defined under current climate conditions because of sparse tide gauge networks and relatively short temporal records. This gives a poor baseline for assessing future changes and detecting changes in observed records. Using results from 141 sites worldwide for the last four decades Woodworth and Blackman (2004) found that at some locations extreme sea levels have increased and that the relative contribution from changes in mean sea level and atmospheric storminess depended on location.

Methods of simulating extreme sea levels

Climate driven changes in extreme sea level will come about because of the increases in mean sea level and changes in the track, frequency or intensity of atmospheric storms. (From the perspective of coastal flooding the vertical movement of land, for instance due to post glacial rebound, is also important when considering the contribution from mean sea level change.) To provide the large-scale context for these changes global climate models are required though their resolution (typically 150 to 300 km horizontally) is too coarse to represent the details of tropical cyclones or even the extreme winds associated with mid-latitude cyclones. However, some studies have used global climate model forcing directly to drive storm surge models to provide estimates of changes in extreme sea level (e.g., Flather and Williams, 2000). To obtain more realistic simulations from the large-scale drivers three approaches are used, dynamical and statistical downscaling and a stochastic method (see Section 11.10 for general details).

As few regional climate models currently have an ocean component, these are used to provide high resolution (typically 25 to 50 km horizontally) surface winds and pressure to drive a storm surge model (e.g., Lowe et al., 2001). This sequence of one-way coupled models is usually carried out for a present day (Debenard et al., 2003) or historic baseline (e.g., Flather et al., 1998) and a period in the future (e.g., Lowe et al., 2001 and Debenard et al., 2003). In the statistical approach, relationships between large scale synoptic conditions and local extreme sea levels are constructed. These relationships can be developed using either analyses from weather prediction models and observed extreme sea levels, or using global climate models and present day simulations of extreme water level made using the dynamic methods described above. Simulations of future extreme sea level are then derived from applying the statistical relationships to the future large-scale atmospheric synoptic conditions simulated by a global climate model (e.g., von Storch and Reichardt, 1997). The statistical and dynamical approach can be combined, using a statistical model to produce the high resolution wind fields forcing the wave and storm surge dynamical models (Lionello et al 2003). Similarly, the stochastic sampling method identifies the key characteristics of synoptic weather events responsible for extreme sea levels (intensity and movement) and represents these by frequency distributions. For each event simple models are used to generate the surface wind and pressure fields and these are applied to the storm surge model (e.g., Hubbert and McInnes, 1999). Modifications to the frequency distributions of the weather events to represent changes under enhanced greenhouse conditions are derived from global climate models and then used to infer a future storm surge climatology.

Extreme sea level changes – sample projections from three regions

1. Australia

In a study of storm surge impacts in northern Australia, a region with only a few short sea level records, McInnes et al. (2005) used stochastic sampling and dynamical modelling to investigate the implications of climate change on extreme storm surges and inundation. Cyclones occurring in the Cairns region from 1907 to 1997 were used to develop probability distribution functions governing the cyclone characteristics of speed and direction with an extreme value distribution fitted to the cyclone intensity. Cyclone intensity distribution was then modified for enhanced greenhouse conditions based on Walsh and Ryan (2000) in which cyclones off northeast Australia were found to increase in intensity by about 10%. No changes were imposed upon cyclone frequency or direction since no reliable information is available on the future behaviour of the main influences in these, respectively ENSO or mid-level winds. Analysis of the surges resulting from 1000 randomly selected cyclones with current and future intensities show that the increased intensity leads to an increase in the height of the 1 in 100 year event from 2.6 m to 2.9 m with 1 in 100 year becoming 1 in 70 years. This also results in the areal extent of inundation more than doubling (from approximately 32 km^2 to 71 km^2). Similar increases for Cairns and other coastal locations were found by Hardy et al. (2004).

2. Europe

Several dynamically-downscalted projections of climate driven changes in extreme water levels on the European shelf region have been carried out. Woth (2005) explored the effect of two different GCMs and their projected climates changes due to two different emissions scenarios (SRES A2 and B2) on storm surges along the North Sea coast. She used data from one RCM downscaling the four GCMs simulations (Woth et al., 2006) using data from four RCMs driven by one GCM produced indistinguishable results) and demonstrated significant increases in the top 1% of events of 10-20cm above average sea-level change over the continental European North Sea coast. The changes from the different experiments were statistically indistinguishable though those from the models incorporating the A2 emissions were consistently larger. When including the effects of global mean sea level rise and vertical land movements Lowe and Gregory (2005) found increases in extreme sea level are projected for the entire UK coastline using a storm surge model driven by one of the RCMs analysed by Woth et al. (2006) (Box 11.5, Figure 1). Using a Baltic Sea model driven by data from four RCM simulations Meier (2006) found the changes in storm surges to vary strongly between the simulations but with some tendency for larger increases in the 100-year surges than in the mean sea level.

[INSERT BOX 11.5, FIGURE 1 HERE]

Lionello et al. (2003) estimated the effect of CO_2 doubling on the frequency and intensity of high wind waves and storm-surge events in the Adriatic Sea. The regional surface wind fields were derived from the sea level pressure field in a 30-year long ECHAM4 T106 resolution time slice experiment by statistical downscaling and then used to force a wave and an ocean model. They found no statistically significant changes in the extreme surge level and a decrease in the extreme wave height with increased CO_2. An underestimation of the observed wave heights and surge levels calls for caution in the interpretation of these results. Using AOGCM projections, X.L. Wang et al. (2004) inferred an increase in winter and autumn seasonal mean and extreme wave heights in the northeast and southwest North Atlantic, but a decrease in the mid-latitudes of the North Atlantic. Not all changes were significant and in some regions (e.g. the North Sea) their sign was found to depend on the emissions scenario.

3. Bay of Bengal

Several dynamic simulations of storm surges have been carried out for the region but these have often involved using results from a small set of historical storms with simple adjustments (such as adding on a mean sea level or increasing wind speeds by 10%) to account for future climate change (e.g., Flather and Khandker, 1993). This technique has the disadvantage that by taking a relatively small and potentially biased set of storms it may lead to a biased distribution of water levels with an unrealistic count of extreme events. In one study using dynamical models driven by RCM simulations of current and future climates, Unnikrishnan et al. (2006) showed that despite no significant change in the frequency of cyclones there were large increases in the frequency of the highest storm surges.

Uncertainty

Changes in storm surges and wave heights have been addressed for only a limited set of models. Thus we can not reliably quantify the range of uncertainty in estimates of future coastal flooding and can only make crude estimates of the minimum values (Lowe and Gregory, 2005). There is some evidence that the dynamical downscaling step in providing data for storm surge modelling is robust, i.e. does not add to the uncertainty. However, the general low level of confidence in projected circulation changes from AOGCMs implies a substantial uncertainty in these projections.

11.10 Assessment of Regional-Climate Projection Methods

The assessment of methods recognizes the challenges posed by the complex interactions that occur at many spatial and temporal scales, involving the general circulation, cross scale feedbacks, and regional scale forcing.

11.10.1 Methods for Generating Regional-Climate Information

AOGCMs constitute the primary tool for capturing the global climate system behaviour. They are used to investigate the processes responsible for maintaining the general circulation and its natural and forced variability (Chapter 8) to assess the role of various forcing factors in observed climate change (Chapter 9) and to provide projections of the response of the system to scenarios of future external forcing (Chapter 10). As AOGCMs seek to represent the whole climate system, clearly they provide information on regional climate and climate change and relevant processes directly. For example, the skill in simulating the climate of last century when accounting for all known forcings demonstrates the causes of recent climate change (Chapter 9) and this information can be used to constrain the likelihood of future regional climate change (Stott et al., 2006; see also Section 11.10.2). AOGCM projections provide plausible future regional climate scenarios though methods to establish the reliability of the regional AOGCM scales have yet to mature. The spread within an ensemble of AOGCMs is often used to characterise the uncertainty in projected future climate changes. Some regional responses are consistent across AOGCMs simulations though for other regions the spread remains large (see Sections 11.2–11.9).

Because of their significant complexity and the need to provide multi-century integrations, horizontal resolutions of the atmospheric components of the AOGCMs in the AR4 range from 400 km to 125 km. Generating information below the grid scale of AOGCMs is referred to as downscaling. There are two main approaches, dynamical and statistical. Dynamical downscaling uses high-resolution climate models to represent global or regional sub-domains, and use either observed or lower resolution AOGCM data as their boundary conditions. Dynamical downscaling has the potential for capturing mesoscale nonlinear effects and providing coherent information between multiple climate variables. These models are formulated using physical principles and they can credibly reproduce a broad range of climates around the world, which increases confidence in their ability to downscale realistically future climates. The main drawbacks of dynamical models are their computational cost and that in future climates the parameterisation schemes they use to represent sub-grid-scale processes may be operating outside the range they were designed for.

Empirical statistical downscaling (SD) methods use cross-scale relationships that have been derived from observed data and apply these to climate model data. SD methods have the advantage of being computationally inexpensive, able to access finer scales than dynamical methods and applicable to parameters that cannot be directly obtained from the RCM outputs. They require observational data at the desired scale for a long enough period to allow the method to be well trained and validated. The main drawbacks of SD methods are that they assume the derived cross-scale relationships remain stable when the climate is perturbed, they cannot effectively accommodate regional feedbacks and, in some methods, can lack coherency between multiple climate variables.

11.10.1.1 High-Resolution Atmosphere-Only GCMs

Atmosphere-only GCMs (AGCMs) include interactive land-surface schemes as in an AOGCM but require information on sea surface temperatures and sea-ice (SSTI) as a lower boundary condition. Given the short time-scales associated with the atmosphere and land-surface components compared to those in the ocean, relatively short time slices (a few decades) can be run at high resolution. The SSTI information required can be derived from observations or AOGCMs. The use of observations can improve simulations of current climate but combing these with AOGCM-derived changes for the future climate (e.g., Rowell, 2005) increases the risk of inconsistency in the projected climate. The absence of two-way feedback between the atmosphere and ocean in AGCMs can cause a significant distortion of the climatic variability (Bretherton and Battisti, 2000), as documented over regions such as the Indian Ocean and the South Asian monsoon (Douville, 2005; Inatsu and Kimoto, 2005). The large-scale climate responses of AGCM and AOGCM appear to be similar in many regions; when and where they differ, the consistency of the oceanic surface boundary condition may be questioned (May and Roeckner, 2001; Govindasamy et al., 2003). Further research is required to determine if the similarity is sufficient for the time-slice approach with AGCMs to be considered a robust downscaling technique.

Model grids of 100 km and finer have become feasible and 50 km will likely be the norm in the near future (Bengtsson, 1996; May and Roeckner, 2001; Déqué and Gibelin, 2002; Govindaswamy, 2003). High performance computer systems now allow global computations at 20 km (e.g., May, 2004a; Mizuta et al., 2006), although for short time slices only. Evaluated on the scale typical of current AOGCMs, nearly all quantities simulated by high-resolution AGCMs agree better with observations, but the improvements vary significantly with regions (Duffy et al., 2003) and specific variables, and extensive recalibration of parameterisations is often required. Notable improvements occur in orographic precipitation and improved dynamics of mid-latitude weather systems (see Chapter 10). The highest resolution offers the prospect of credible simulations of the climatology of tropical cyclones (e.g., May, 2004a; Mizuta et al., 2006). Coordinated multi-model experiments are needed however to optimize the value of these high-resolution studies for general assessment.

An alternative to uniform high-resolution is variable-resolution AGCMs (VRGCM; e.g., Déqué and Piedelievre, 1995; Krinner et al., 1997; Fox-Rabinovitz et al., 2001; McGregor et al., 2002; Gibelin and Déqué, 2003). The VRGCM approach is attractive as it permits, within a unified modelling framework, a regional increase of resolution while retaining the interaction of all regions of the atmosphere. Numerical artefacts due to stretching have been shown to be small when using modest stretching factors (e.g., Lorant and Royer, 2001). VRGCMs results capture, over the high-resolution region, finer scale details than uniform-resolution models while retaining global skill similar to uniform-resolution simulations with the same number of grid points.

11.10.1.2 Nested Regional Climate Models

The principle behind nested modelling is that, consistent with the large-scale atmospheric circulation, realistic regional climate information can be generated by integrating a regional climate model (RCM) if the following premises are satisfied: time-varying large-scale atmospheric fields (winds, temperature and moisture) are supplied as lateral boundary conditions (LBC) and SSTI as lower boundary conditions; the control from the LBCs keeps the interior solution of the RCM consistent with the driving atmospheric circulation; and subgrid-scale physical processes are suitably parameterised, including fine-scale surface forcing such as orography, land-sea contrast and land use.

A typical RCM grid for climate-change projections is around 50 km, although some climate simulations have been performed using grids of 15 or 20 km (e.g., Leung et al., 2003, 2004; Christensen and Christensen, 2004; Kleinn et al., 2005). Recently projections of climate changes for East Asia have been completed with a 5-km non-hydrostatic RCM (Kanada et al., 2005; Yoshizaki et al., 2005; Yasunaga et al., 2006), but only for short simulations. Following the trend in global modelling, RCMs are increasingly coupled interactively with other components of the climate system, such as regional ocean and sea ice (e.g., Bailey and Lynch 2000; Döscher et al., 2002; Rinke et al., 2003; Bailey et al. 2004; Meier et al., 2004; Sasaki et al., 2006a), hydrology, and with interactive vegetation (Gao and Yu, 1998; Xue et al., 2000).

Multi-decadal RCM experiments are becoming standard (e.g., Whetton et al., 2000; Kwon et al., 2003; Leung et al., 2004; Kjellström et al., 2006; Plummer et al., 2006) including the use of ensembles (Christensen et al., 2002), enabling a more thorough validation and exploration of projected changes. In multi-year ensemble simulations driven by atmospheric reanalyses of observations, Vidale et al. (2003) have shown that RCMs have skill in reproducing interannual variability in precipitation and surface air temperature. The use of ensemble simulations has enabled quantitative estimates regarding the sources of uncertainty in projections of regional climate changes (Rowell, 2005; Déqué et al., 2005, 2006; Frei et al., 2006; Graham et al., 2006; Beniston et al., 2006). Combining information from four RCM simulations, Christensen et al. (2001) and Rummukainen et al. (2003) demonstrated that it is feasible to explore not only uncertainties related to projections in the mean climate state, but also for higher order statistics.

The difficulties associated with the implementation of LBCs in nested models are well documented (e.g., Davies, 1976; Warner et al., 1997). As time progresses in a climate simulation, the RCM solution turns from an initial-value problem gradually more into a boundary-value problem (BVP). The mathematical interpretation is that nested models represent a fundamentally ill-posed BVP (Staniforth, 1997; Laprise, 2003). The control exerted by LBCs on the internal solution generated by RCMs appears to vary with the size of the computational domain (e.g., Rinke and Dethloff, 2000), as well as location and season (e.g., Caya and Biner, 2004). In some applications, the flow developing within the RCM domain may become inconsistent with the driving LBC. This may (Jones et al., 1997) or may not (Caya and Biner, 2004) impact on climate statistics. RCMs are normally only driven by LBCs with high time resolution to capture the time variations of large-scale flow. Some RCMs also use nudging or relaxation of large scales in the interior of the domain (e.g., Kida et al., 1991; von Storch et al., 2000; Biner et al., 2000). This has proved useful to minimize the distortion of the large scales in RCMs (von Storch et al., 2000; Mabuchi et al., 2002; Miguez-Macho et al., 2004), although it can also hide model biases. One-way RCM-GCM coupling is mostly used, although recently a two-way nested RCM has been developed (Lorenz and Jacob, 2005) thus achieving interaction with the global atmosphere as with variable-resolution AGCM.

The ability of RCMs to simulate the regional climate depends strongly on the realism of the large-scale circulation that is provided at the LBCs (e.g., Pan et al., 2001). Latif et al. (2001) and Davey et al. (2002) have shown that strong biases in the tropical climatology of AOGCMs can impact negatively on downscaling studies for several regions of the world. Nonetheless, the reliability of nested models, i.e. their ability to generate meaningful fine-scale structures that are absent in the LBCs, is clear. A number of studies have shown that the climate statistics of atmospheric small scales can be recreated with the right amplitude and spatial distribution, even if these small scales are absent in the LBCs (Denis et al., 2002, 2003; Antic et al., 2005; Dimitrijevic and Laprise, 2005). This implies that RCMs can add value at small scales to climate statistics when driven by AOGCMs with accurate large scales. Overall the skill at simulating current climate has improved with the MMD AOGCMs (Chapter 8), which will lead to higher quality LBCs for RCMs.

11.10.1.3 Empirical/Statistical Downscaling Methods

A complementary technique to RCMs is to use derived relationships linking large-scale atmospheric variables (predictors) and local/regional climate variables (predictands). The local/regional scale climate-change information is then obtained by applying the relationships to equivalent predictors from AOGCM simulations. The guidance document (Wilby et al., 2004) from the IPCC Task Group on Data and Scenario

Support for Impact and Climate Analysis (TGICA) provides a comprehensive background on this approach and covers important issues in using SD applications. SD methods cover regression-type models including both linear or nonlinear relationships, unconditional or conditional weather generators for generating synthetic sequences of local variables, techniques based on weather classification which draw on the more skilful attributes of models to simulate circulation patterns, analogue methods which seek equivalent weather states from the historical record; and a combination of these techniques possibly being most appropriate. An extension to SD is the statistical-dynamical downscaling (SDD) technique (e.g., Fuentes and Heimann, 2000) which combines weather classification with RCM simulations. A further development is the application of SD to high-resolution climate models output (Lionello et al., 2003; Imbert and Benestad, 2005).

SD research has shown an extensive growth in application, and include an increased availability of generic tools for the impact community (e.g., SDSM, Wilby et al., 2002; clim.pact package, Benestad, 2004b; package pyclimate, Fernández and Sáenz, 2003); applications in new regions (e.g. Asia, Chen and Chen, 2003); the use of techniques to address exotic variables such as phenological series (Matulla et al., 2003); extreme heat-related mortality (Hayhoe et al., 2004), ski season (Scott et al., 2003), land use (Solecki and Oliveri, 2004), stream flow or aquatic ecosystems (Cannon and Whitfield, 2002; Blenckner and Chen, 2003).; the treatment of climate extremes (e.g., Katz et al., 2002; X.L. Wang et al., 2004; Seem, 2004); Caires et al., 2006); inter-comparison studies evaluating methods (e.g., STARDEX, Goodess et al., 2006; Schmidli et al., 2006; Haylock et al., 2006); application to multi-model and multi-ensemble simulations in order to express model uncertainty alongside other key uncertainties (e.g., Benestad, 2002a,b; Hewitson and Crane, 2006; Wang and Swail, 2006b); assessing non-stationarity in climate relationships (Hewitson and Crane, 2006); and spatial interpolation using geographical dependencies (Benestad, 2005). In some cases SD methods have been used to project statistical attributes instead of raw values of the predictand, for example the probability of rainfall occurrence, precipitation, wind or wave height distribution parameters and extreme event frequency (e.g. Abaurrea and Asin, 2005; Beckmann and Buishand, 2002; Buishand et al., 2004; Busuioc and von Storch, 2003; Wang and Swail, 2006a,b; Diaz-Nieto and Wilby, 2005; Pryor et al., 2005a,b).

Evaluation of SD is done most commonly through cross-validation with observational data for a period that represents an independent or different "climate regime" (e.g., Bartman et al., 2003; Busuioc, 2001a; Trigo and Palutikof, 2001; Hanssen-Bauer et al., 2003). Stationarity, i.e. whether the statistical relationships are valid under future climate regimes, remains a concern with SDMs. This is only weakly assessed through cross-validation tests because future changes in climate are likely to be substantially larger than observed historical changes. This issue was assessed in Hewitson and Crane (2006) where, within the SD method used, the non-stationarity was shown to result in an underestimation of the magnitude of the change. In general, the most effective SD methods are those that combine elements of deterministic transfer functions and stochastic components (e.g., Beersma and Buishand, 2003; Katz et al., 2003; Busuioc and von Storch, 2003; Palutikof et al., 2002; X.L. Wang et al., 2004; Lionello et al., 2003; Hewitson and Crane, 2006; Wilby et al., 2003; Hansen and Mavromatis, 2001). Regarding the predictors, the best choice appears to combine dynamical and moisture variables, especially in cases where precipitation is the predictand (e.g., Wilby et al., 2003).

Pattern scaling is a simple statistical method for projecting regional climate change, which involves normalising AOGCMs' response patterns according to the global-mean temperature. These normalised patterns are then rescaled using global-mean temperature responses estimated for different emissions scenarios from a simple climate model (see Chapter 10). Some developments were made using various versions of scaling techniques (e.g., Christensen et al., 2001; Mitchell, 2003; Ruosteenoja et al., 2006; Salathé, 2005). For example, Ruosteenoja et al. (2006) developed a pattern-scaling method using linear regression to represent the relationship between the local AOGCM-simulated temperature and precipitation response and the global-mean temperature change. Another simple statistical technique is to use the GCM output for the variable of interest (i.e. the predictand) as the predictor and then apply a simple local change factor/scaling procedure (e.g., Diaz-Nieto and Wilby, 2005; Hanssen-Bauer et al., 2003; Widmann et al., 2003; IPCC, 2001, Chapter 13).

Many studies have been performed since the TAR comparing various SD methods. In general, conclusions about one method versus another are dependent on region and the criteria used for comparison, and on the inherent attributes of each method. For example, Diaz-Nieto and Wilby (2005) downscale river flow and find that while two methods give comparable results, they differ in responses as a funciton of how the methods treat multidecadal variability.

When comparing the merits of SD methods based on daily and monthly downscaling models, in terms of their ability to predict monthly means, daily models are better (e.g., Buishand et al., 2004). In terms of nonlinearity in downscaling relationships, Trigo and Palutikof (2001) noted that complex nonlinear models may not be better than simpler linear/slightly nonlinear approaches for some applications. However, Haylock et al. (2006) found that SD methods based on nonlinear artificial neural networks are best at modelling the interannual variability of heavy precipitation but they underestimate extremes. There remains much downscaling work that goes unreported, SD activities being often implemented pragmatically for serving specific project needs, rather than for use by a broader scientific community; this is especially the case in developing nations. In some cases this work is only found within the soft literature, for example, the AIACC project (http://www.aiaccproject.org/), which supports impact studies in developing nations.

11.10.1.4 Inter-Comparison of Downscaling Methods

At the time of the TAR SD methods were viewed as a complementary technique to RCMs for downscaling regional climate, each approach having distinctive strengths and weaknesses. The conclusion of the TAR that SD and RCMs are comparable for simulating current climate still holds.

Since the TAR a few additional studies have systematically compared the SD and RCM approaches (e.g., Huth et al., 2001; Hanssen-Bauer et al., 2003, 2005; Wood et al., 2004, Busuioc et al., 2006; Schmidli et al., 2006; Haylock et al., 2006). These related mainly to the similarity of the climate-change signal (e.g., Hanssen-Bauer et al., 2003). A more complex study considered additional information about the RCM skill in simulating the current regional climate features and reproducing the connection between large-scale and regional-scale patterns used for fitting the SD method (Busuioc et al., 2006). Other studies following the STARDEX project (e.g., Schmidli et al., 2006; Haylock et al., 2006) compared the two approaches in terms of their skill in reproducing current climate features, as well as in terms of climate-change signal derived from their outputs, focusing on climate extremes and complex topography processes over Europe.

11.10.2 Quantifying Uncertainties

11.10.2.1 Sources of Regional Uncertainty

Most sources of uncertainty on regional scales are similar to those on the global scale (Chapter 10, Section 10.5 and Box 10.2), but there are both changes in emphasis and new issues that arise in the regional context. Spatial inhomogeneity of both land use/cover change (De Fries et al., 2002; Chapter 2 and Chapter 7, Section 7.2, and Box 11.4) and aerosol forcing adds to regional uncertainty. When analyzing studies involving models to add local detail, the full cascade of uncertainty through the chain of models has to be considered. The degree to which these uncertainties influence the regional projections of different climate variables is not uniform. An indication of this is, for example, that models agree more readily on the sign and magnitude of temperature changes than of precipitation changes.

The regional impact of these uncertainties in climate projections has been illustrated by several authors. For example, incorporating a model of the carbon-cycle into a coupled AOGCM gave a dramatically enhanced response to climate change over the Amazon basin (Cox et al., 2000; Jones et al., 2003) and Borneo (Kumagi et al., 2004). Further, the scale of the resolved processes in a climate model can significantly affect its simulation of climate over large regional scales (Pope and Stratton, 2002; Lorenz and Jacob, 2005). Frei et al. (2003) show that models with the same representation of resolved processes but different representations of sub-grid-scale processes can represent the climate differently. The regional impact of changes in the representation of the land-surface feedback is demonstrated by, for example, Oleson et al. (2004) and Feddema et al. (2005) (See also Box 11.4 on land use change).

Evaluation of uncertainties at regional and local scales is complicated by the smaller ratio of the signal to the internal variability, especially for precipitation, which makes the detection of a response more difficult. Also, the climate may itself be poorly known on regional scales in many data-sparse regions. Thus evaluation of model performance as a component of an analysis of uncertainty can itself be problematic.

11.10.2.2 Characterizing and Quantifying Regional Uncertainty

11.10.2.2.1 Review of regional uncertainty portrayed in the TAR

In the TAR uncertainties in regional climate projections were discussed, but methods for quantifying them were relatively primitive. For example, in the TAR chapter on regional projections (Giorgi et al., 2001a), uncertainties in regional projections of climate change from different GCMs were qualitatively portrayed (e.g., large or small increases/decreases in precipitation) based only on simple agreement heuristics (e.g., seven of the nine models showed increases). Early examples of quantitative estimates of regional uncertainty include portraying the median and inter-model range of a variable (e.g., temperature) across a series of model projections and attaching probabilities to a group of scenarios on a regional scale (New and Hulme, 2000; Jones, 2000).

11.10.2.2.2 Using multi-model ensembles

A number of studies have taken advantage of multi-model ensembles formed by GCMs that have been driven by the same forcing scenarios to generate quantitative measures of uncertainty, particularly probabilistic information on a regional scale. Table 11.3 summarizes aspects of the methods reviewed below and in Section 11.10.2.2.3. The results highlighted in Chapter 10, Section 10.5 and Box 10.2 on climate sensitivity demonstrate that multi-model ensembles explore only a limited range of the uncertainty. Also, the distribution of GCM sensitivities is not by construction a representative sample from those probability distributions and thus the regional probabilities generated using multi-model ensembles will not represent the full spread of possible regional changes.

[INSERT TABLE 11.3 HERE]

Räisänen and Palmer (2001) used 17 GCMs forced with an idealised annual increase in CO_2 of 1% to calculate the probability of exceedance of thresholds of temperature increase (e.g., >1°C) and precipitation change (e.g., <–10%). These were used to demonstrate that a probabilistic approach has advantages over conventional deterministic estimates by demonstrating the economic value of a probabilistic assessment of future climate change. Giorgi and Mearns (2002) developed measures of uncertainty for regional temperature and precipitation change by weighting model results according to biases in their simulation of present-day climate and convergence of their projections to the ensemble's consensus. Their REA (Reliability Ensemble Average) method was applied to the 9 GCMs assessed in the TAR to provide uncertainty estimates separately for the A2 and B2 SRES emission scenarios for 22 large sub-continental regions.

Tebaldi et al. (2004, 2005) used a Bayesian approach to define a formal statistical model for deriving probabilities from an ensemble of projections forced by a given SRES scenario. Using the Giorgi and Mearns (2002 and 2003) approach, model bias and convergence criteria determine the shape and width of the posterior PDFs of temperature and precipitation change signals. Expert judgement can be incorporated in the form of prior distributions that have the effect of assigning different relative weights to the two criteria (Tebaldi et al., 2004, Lopez et al. 2006). The method developed by Furrer et al. (2006) to combine GCMs' output at the grid point scale into probabilistic projections is described in detail in Chapter 10. By straightforward area averaging PDFs of climate change at the regional scale can be obtained. When this is done for the Giorgi and Francisco (2000) regions, the regional PDFs from Furrer et al. agree overall with the empirical histogram of the ensemble projections and the Tebaldi et al. (2004) PDFs, with relatively small differences in spread and generally no clear difference in location.

Greene et al. (2006) used a Bayesian framework to model an ensemble of GCM projections under individual SRES scenarios by an extension of methods used for seasonal ensemble forecasting. The set of GCM simulations of the observed period 1902–1998 are individually aggregated in area averaged annual or seasonal time series and jointly calibrated through a linear model to the corresponding observed regional trend. The calibration coefficients and their uncertainty are estimated and then applied to the future projections to provide probabilistic forecast of future trends. Two critical assumptions are responsible for this method's results being so different from the ensemble projections or the PDFs produced by Tebaldi et al. (2004,2005) (see Figure 11.26 and Supplementary Material Figures S11.33-35). Firstly, the method attributes large uncertainty to models which are unable to reproduce historical trends despite the uncertainty in the relatively weak forcings in the historical period and the large natural variability on regional scales. Second, a strong stationarity assumption is required to extrapolate the relationship derived over the historical record to future trends, which involve a different combination of and some significantly stronger forcings. The significantly smaller warming and the large width of the PDFs (at times including negative values) labelled by a "G" in Figure 11.26 are then interpretable as a result of this stationarity constraint and the large uncertainty in the fitting of the trends. They contrast starkly with the larger warming represented in the histograms of model projections and their synthesis in the Tebaldi et al. and in the Furrer et al. (not shown) PDFs. This is particularly so in the lower latitude regions of Africa, South Asia and the Southern Hemisphere, possibly as a consequence of particularly weak trends in the observations and/or relatively worse performance of the GCMs.

[INSERT FIGURE 11.26 HERE]

Dessai et al. (2005) apply the idea of simple pattern scaling (Santer et al., 1990), to a multi-model ensemble of AOGCMs. They "modulate" the normalized regional patterns of change by the global mean temperature changes generated under many SRES scenarios and climate sensitivities through MAGICC, a simple probabilistic energy balance model (Wigley and Raper, 2001). Their work is focused on measuring the changes in PDFs as a function of different sources of uncertainty. In this analysis, the impact of the SRES scenarios turns out to be the most relevant for temperature changes, particularly in the upper tail of the distributions while the GCM weighting does not produce substantial differences. This result is probably dependent on the long horizon of the projections considered (late 21st century). Arguably, the emission scenario would be less important in the short-to-mid-term. Climate sensitivity has an impact mainly in the lower tail of the distributions. For precipitation changes, all sources of uncertainty seem relevant but the results are very region-specific and thus difficult to generalize. More work to test the robustness of these conclusions is needed, especially when these are obviously not consistent with the results in Ch. 10, Figure 10.29. For example, the use of pattern scaling will likely underestimate the range of projections that would be obtained by running a larger ensemble of GCMs (Murphy et al., 2004).

The work described above has involved either large area averages of temperature and precipitation change or statistical modelling at the grid box scale. Good and Lowe (2006) show that trends for large area and grid box average projections for precipitation are often very different from the local trends within the area. This demonstrates the inadequacy of inferring the behaviour at fine-scales from that of large-area averages.

11.10.2.2.3 Using perturbed physics ensembles

Another method for exploring uncertainties in regional climate projections is the use of large perturbed physics ensembles (described in detail in Chapter 10). These allow a characterisation of the uncertainty due to poorly constrained parameters within the formulation of a model. Harris et al. (2006) have combined the results from a 17 member ensemble (Collins et al., 2006) with a larger perturbed physics ensemble investigating the equilibrium climate response to a doubling of CO_2 (Webb et al., 2006). They developed a bridge between spatial patterns of the transient and equilibrium climate response by way of simple pattern scaling (Santer et al., 1990) allowing results from the large ensemble to be translated into PDFs of time dependent regional changes. Uncertainties in surface temperature and precipitation changes are derived (Supplementary Material Figures S11.36 and S11.37), which arise from the poorly-constrained atmospheric model parameters, internal variability and pattern scaling errors. The latter are quantified by comparing the scaled equilibrium response with the transient response for 17 model versions with identical parameter settings. Errors introduced by the pattern-scaling technique are largest when the transient response varies non-linearly with global temperature, as is the case for precipitation in certain regions.

11.10.2.2.4 Other approaches to quantifying regional uncertainty

As described in Chapter 10, Stott and Kettleborough (2002) provide PDFs of future change in climate by making use of the robust observational constraints on a climate model's response to greenhouse gas and sulphate aerosol forcings that underpin the attribution of recent climate change to anthropogenic sources. The study by Stott et al. (2006a) is the first to adapt this method for continental scales. It considers two methods of constraining future continental temperature projections, one based on using observed historical changes only over the region of interest, one based on using observed changes of global temperature patterns. The first approach produces wider PDFs, since the uncertainty of detection at the regional scale is larger. The second approach incorporates more information, hence reducing the uncertainty, but assumes that the GCM represents correctly the relationship between global mean and regional temperature change. In contrast to the studies of Section 11.10.2.2.2 this work uses projections from a single GCM (HadCM3), though Stott et al. (2006b) have confirmed the results of this methodology for other models.

In general, the regional sections of this chapter assess the uncertainty in regional changes based on expert understanding of the relevant processes, rather than by formal probabilistic methods which are still in their infancy and currently do not provide definitive results. An approach to a process-based assessment of the reliability of modelled climate change responses and thus uncertainties in its future projections has been proposed by Rowell and Jones (2006). They perform an assessment of the physical and dynamical mechanisms responsible for a specific future outcome, in their case European Summer drying. Their analysis isolates the contribution of the four major mechanisms analysed, spatial pattern of warming, other large-scale changes, reduced spring soil moisture and summer soil moisture feedbacks. In certain regions the second process makes a minor contribution with the first and third dominating. This leads to the conclusion that the sign of the change is robust as confidence in the processes underlying these mechanisms is high.

11.10.2.2.5 Combined uncertainties: GCMs, emissions, and downscaling techniques

It is important to quantify the relative importance of the uncertainty arising from the downscaling step (from the RCM formulation or the assumptions underlying an ESD method) against the other sources of uncertainty. For example, in the application of statistical downscaling methods to probabilistic scenarios, Benestad (2002b, 2004a) used a multi-model ensemble coupled to statistical downscaling to derive tentative probabilistic scenarios at a regional scale for Northern Europe.

The PRUDENCE project (Box 11.2) provided the first opportunity to weigh these various sources of uncertainty for simulations over Europe. Rowell (2005) evaluated a 4 dimensional matrix of climate modelling experiments that included two different emissions scenarios, 4 different GCM experiments, multiple ensemble members within the latter to assess internal variability, and 9 different RCMs, for the area of the British Isles. He found that the dynamical downscaling added a small amount of uncertainty compared to the other sources for temperature evaluated as monthly/seasonal averages. For precipitation the relative contributions of the four sources of uncertainty are more balanced. Déqué et al. (2005, 2006) show similar results for the whole of Europe, as do Ruosteenoja et al. (2006) for subsections of Europe. Kjellström et al. (2006) found that the differences among different RCMs driven by the same GCM become comparable to those among the same RCM driven by different GCMs when evaluating daily maximum and minimum temperatures. However, mean responses in the PRUDENCE RCMs were often quite different to that of the driving GCM. This suggests that some of the spread in RCM responses may be unrealistic due to model inconstistency (Jones et al., 1997). However, it should be noted that only a few of the RCMs in PRUDENCE were driven by more than one GCM which adds further uncertainty regarding these conclusions. Other programs similar to PRUDENCE have begun for other regions of the world, such as NARCCAP (North American Regional Climate Change Assessment) over North America (Mearns et al., 2005), CREAS (Regional Climate Change Scenarios for South America, Marengo and Ambrizzi (2006)), and CLARIS (A Europe-South America Network for Climate Change Assessment and Impact Studies, http://www.claris-eu.org) over South America.

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Frequently Asked Question 11.1: Do Projected Changes in Climate Vary from Region to Region?

Climate varies from region to region. This variation is driven by the uneven distribution of solar heating, the individual responses of the atmosphere, oceans and land surface, the interactions between these, and the physical characteristics of the regions. The perturbations of the atmospheric constituents that lead to global changes affect certain aspects of these complex interactions. Some human-induced factors that affect climate ("forcings") are global in nature, while others differ from one region to another. For example, carbon dioxide, which causes warming, is distributed evenly around the globe, regardless of where the emissions originate, whereas sulphate aerosols (small particles) that offset some of the warming, tend to be regional in their distribution. Furthermore, the response to forcings is, in part, governed by feedback processes that may operate in different regions from those in which the forcing is greatest. Thus the projected changes in climate will also vary from region to region.

Latitude is a good starting point for considering how changes in climate will affect a region. For example, while warming is expected everywhere on Earth, the amount of projected warming generally increases from the tropics to the poles in the Northern Hemisphere. Precipitation is more complex, but also has some features that are latitude-dependent. In latitudes adjacent to the polar regions, precipitation is projected to increase, while decreases are projected in many regions adjacent to the tropics (see FAQ 11.1, Figure 1). Increases are projected in tropical precipitation during rainy seasons, e.g. monsoons, and over the tropical Pacific in particular.

[INSERT FAQ 11.1, FIGURE 1 HERE]

The location with respect to the oceans and mountain ranges is also an important factor. Generally, the interiors of continents are projected to warm more than the coastal areas. Precipitation responses are especially sensitive not only to the continental geometry, but the shape of nearby mountain ranges and wind flow direction. Monsoons, extratropical cyclones and hurricanes/typhoons are all influenced in different ways by these region-specific features.

Some of the most difficult aspects of understanding and projecting changes to regional climate relate to possible changes in the circulation of the atmosphere and oceans, and their patterns of variability. Although general statements covering a variety of regions with qualitatively similar climates can be made in some cases, nearly every region is idiosyncratic in some ways. This is true whether it is the coastal zones surrounding the distinctive subtropical Mediterranean Sea, or the distinctive extreme weather in the North American interior that depends on moisture transport from the Gulf of Mexico, or the interactions between vegetation distribution, oceanic temperatures and atmospheric circulation that help control the southern limit of the Sahara Desert.

While developing an understanding of the correct balance of global and regional factors remains a challenge, the understanding of these factors is steadily growing, increasing our confidence in regional projections.

Tables

Table 11.1. Temperature and precipitation projections by the AR4 global models

Averages over a number regions of the projections by a set of 21 global models for the A1B scenario. The mean temperature and precipitation responses are first averaged for each model over all available realizations of the 1980–1999 period from the 20C3M simulations and the 2080–2099 period of A1B. Computing the difference between these two periods, the table shows the minimum, maximum, median (50%), and 25% and 75% quartile values among the 21 models, for temperature in degrees Celsius and precipitation as a per cent change. Regions in which the middle half (25–75%) of this distribution is all of the same sign in the precipitation response are colored light brown for decreasing and light blue for increasing precipitation. Signal-to-noise ratio for these 20 year mean responses is indicated by first computing a consensus standard deviation of 20 year means, using those models that have at least 3 realizations of the 20C3M simulations, and using all twenty year periods in the 20th century. The signal is assumed to increase linearly in time, and the time required for the median signal to reach 2.83 (2 × √2) times the standard deviation is displayed as an estimate of when this signal is significant at the 95% level. These estimates of the times for emergence of a clearly discernable signal are only shown for precipitation when the models are in general agreement on the sign of the response, as indicated by the coloring. The frequency of extremely warm, wet, and dry seasons, averaged over the models, is also presented, as described in Section 11.2.1.The values are given in per cent and are only shown when at least 14 out of the 21 models agree on an increase (bold) or a decrease in the extremes. A value of 5% indicates no change, as this is the nominal value for the control period by construction. The regions are defined by rectangular latitude/longitude boxes and the coordinates of the bottom left hand and top right hand corners of these are given in degrees in the first column under the region acronym. Information is provided for land areas contained in the boxes except for the Small Islands regions where sea areas are used and for Antarctica where both land and sea areas are used.

		Temperature Response						% Precipitation Response						Extreme Seasons		
REGION	SEASON	MIN	25	50	75	MAX	T YRS	MIN	25	50	75	MAX	T YRS	WARM	WET	DRY
AFRICA																
WAF	DJF	2.3	2.7	3.0	3.5	4.6	10	-16	-2	6	13	23		**100**	**21**	4
	MAM	1.7	2.8	3.5	3.6	4.8	10	-11	-7	-3	5	11		**100**		
12S,20W to 22N,18E	JJA	1.5	2.7	3.2	3.7	4.7	10	-18	-2	2	7	16		**100**	**19**	
	SON	1.9	2.5	3.3	3.7	4.7	10	-12	0	1	10	15		**100**	**15**	
	ANN	1.8	2.7	3.3	3.6	4.7	10	-9	-2	2	7	13		**100**	**22**	
EAF	DJF	2.0	2.6	3.1	3.4	4.2	10	-3	6	13	16	33	55	**100**	**25**	1
	MAM	1.7	2.7	3.2	3.5	4.5	10	-9	2	6	9	20	>100	**100**	**15**	4
12S,22E to 18N,52E	JJA	1.6	2.7	3.4	3.6	4.7	10	-18	-2	4	7	16		**100**		
	SON	1.9	2.6	3.1	3.6	4.3	10	-10	3	7	13	38	95	**100**	**21**	3
	ANN	1.8	2.5	3.2	3.4	4.3	10	-3	2	7	11	25	60	**100**	**30**	1
SAF	DJF	1.8	2.7	3.1	3.4	4.7	10	-6	-3	0	5	10		**100**	**11**	
35S,10E to 12S,52E	MAM	1.7	2.9	3.1	3.8	4.7	10	-25	-8	0	4	12		**98**		
	JJA	1.9	3.0	3.4	3.6	4.8	10	-43	-27	-23	-7	-3	70	**100**	1	**23**
	SON	2.1	3.0	3.7	4.0	5.0	10	-43	-20	-13	-8	3	90	**100**	1	**20**
	ANN	1.9	2.9	3.4	3.7	4.8	10	-12	-9	-4	2	6		**100**	4	**13**
SAH	DJF	2.4	2.9	3.2	3.5	5.0	15	-47	-31	-18	-12	31	>100	**97**		**12**
18N,20 to 30N,65E	MAM	2.3	3.3	3.6	3.8	5.2	10	-42	-37	-18	-10	13	>100	**100**	2	**21**
	JJA	2.6	3.6	4.1	4.4	5.8	10	-53	-28	-4	16	74		**100**		
	SON	2.8	3.4	3.7	4.3	5.4	10	-52	-15	6	23	64		**100**		
	ANN	2.6	3.2	3.6	4.0	5.4	10	-44	-24	-6	3	57		**100**		
EUROPE																
NEU	DJF	2.6	3.6	4.3	5.5	8.2	40	9	13	15	22	25	50	**82**	**43**	0
	MAM	2.1	2.4	3.1	4.3	5.3	35	0	8	12	15	21	60	**79**	**28**	2

		Temperature Response						% Precipitation Response						Extreme Seasons		
	JJA	1.4	1.9	2.7	3.3	5.0	25	-21	-5	2	7	16		88	11	
	SON	1.9	2.6	2.9	4.2	5.4	30	-5	4	8	11	13	80	87	20	2
	ANN	2.3	2.7	3.2	4.5	5.3	25	0	6	9	11	16	45	96	48	2
SEM	DJF	1.7	2.5	2.6	3.3	4.6	25	-16	-10	-6	-1	6	>100	93	3	12
30N,10W to 48N,40E	MAM	2.0	3.0	3.2	3.5	4.5	20	-24	-17	-16	-8	-2	60	98	1	31
	JJA	2.7	3.7	4.1	5.0	6.5	15	-53	-35	-24	-14	-3	55	100	1	42
	SON	2.3	2.8	3.3	4.0	5.2	15	-29	-15	-12	-9	-2	90	100	1	21
	ANN	2.2	3.0	3.5	4.0	5.1	15	-27	-16	-12	-9	-4	45	100	0	46
ASIA																
NAS	DJF	2.9	4.8	6.0	6.6	8.7	20	12	20	26	37	55	30	93	68	0
50N,40E to 70N,180E	MAM	2.0	2.9	3.7	5.0	6.8	25	2	16	18	24	26	30	89	66	1
	JJA	2.0	2.7	3.0	4.9	5.6	15	-1	6	9	12	16	40	100	51	2
	SON	2.8	3.6	4.8	5.8	6.9	15	7	15	17	19	29	30	99	65	0
	ANN	2.7	3.4	4.3	5.3	6.4	15	10	12	15	19	25	20	100	92	0
CAS	DJF	2.2	2.6	3.2	3.9	5.2	25	-11	0	4	9	22		84	8	
30N,40E to 50N,75E	MAM	2.3	3.1	3.9	4.5	4.9	20	-26	-14	-9	-4	3	>100	94		16
	JJA	2.7	3.7	4.1	4.9	5.7	10	-58	-28	-13	-4	21	>100	100	3	20
	SON	2.5	3.2	3.8	4.1	4.9	15	-18	-4	3	9	24		99		
	ANN	2.6	3.2	3.7	4.4	5.2	10	-18	-6	-3	2	6		100		12
TIB	DJF	2.8	3.7	4.1	4.9	6.9	20	1	12	19	26	36	45	95	40	0
30N,50E to 75N,100E	MAM	2.5	2.9	3.6	4.3	6.3	15	-3	4	10	14	34	70	96	34	2
	JJA	2.7	3.2	4.0	4.7	5.4	10	-11	0	4	10	28		100	24	
	SON	2.7	3.3	3.8	4.6	6.2	15	-8	-4	8	14	21		100	20	
	ANN	2.8	3.2	3.8	4.5	6.1	10	-1	2	10	13	28	45	100	46	1
EAS	DJF	2.1	3.1	3.6	4.4	5.4	20	-4	6	10	17	42	>100	96	18	2
20N,100E to 50N,145E	MAM	2.1	2.6	3.3	3.8	4.6	15	0	7	11	14	20	55	98	35	2
	JJA	1.9	2.5	3.0	3.9	5.0	10	-2	5	9	11	17	45	100	32	1
	SON	2.2	2.7	3.3	4.2	5.0	15	-13	-1	9	15	29		100	20	3
	ANN	2.3	2.8	3.3	4.1	4.9	10	2	4	9	14	20	40	100	47	1
SAS	DJF	2.7	3.2	3.6	3.9	4.8	10	-35	-9	-5	1	15		99		
5N,64E to 50N,100E	MAM	2.1	3.0	3.5	3.8	5.3	10	-30	-2	9	18	26		100	14	
	JJA	1.2	2.2	2.7	3.2	4.4	15	-3	4	11	16	23	45	96	32	1
	SON	2.0	2.5	3.1	3.5	4.4	10	-12	8	15	20	26	50	100	29	3
	ANN	2.0	2.7	3.3	3.6	4.7	10	-15	4	11	15	20	40	100	39	3
SEA	DJF	1.6	2.1	2.5	2.9	3.6	10	-4	3	6	10	12	80	99	23	2
11S,95E to 20N,115E	MAM	1.5	2.2	2.7	3.1	3.9	10	-4	2	7	9	17	75	100	27	1
	JJA	1.5	2.2	2.4	2.9	3.8	10	-3	3	7	9	17	70	100	24	2
	SON	1.6	2.2	2.4	2.9	3.6	10	-2	2	6	10	21	85	99	26	3
	ANN	1.5	2.2	2.5	3.0	3.7	10	-2	3	7	8	15	40	100	44	1
NORTH AMERICA																
ALA	DJF	4.4	5.6	6.3	7.5	11.0	30	6	20	28	34	56	40	80	39	0
60N,170W to 72N,103W	MAM	2.3	3.2	3.5	4.7	7.7	35	2	13	17	23	38	40	69	45	0
	JJA	1.3	1.8	2.4	3.8	5.7	25	1	8	14	20	30	45	86	51	1
	SON	2.3	3.6	4.5	5.3	7.4	25	6	14	19	31	36	40	86	51	0
	ANN	3.0	3.7	4.5	5.2	7.4	20	6	13	21	24	32	25	97	80	0
CGI	DJF	3.3	5.2	5.9	7.2	8.5	20	6	15	26	32	42	30	95	58	0
50N,103W to	MAM	2.4	3.2	3.8	4.6	7.2	20	4	13	17	20	34	35	94	49	1
	JJA	1.5	2.1	2.8	3.7	5.6	15	0	8	11	12	19	35	99	46	1
	SON	2.7	3.4	4.0	5.7	7.3	20	7	14	16	22	37	35	99	62	0

		Temperature Response						% Precipitation Response						Extreme Seasons		
85N,10W	ANN	2.8	3.5	4.3	5.0	7.1	15	8	12	15	20	31	25	**100**	**90**	0
WNA	DJF	1.6	3.1	3.6	4.4	5.8	25	-4	2	7	11	36	>100	**80**	**18**	3
30N,50E	MAM	1.5	2.4	3.1	3.4	6.0	20	-7	2	5	8	14	>100	**87**	**14**	
to	JJA	2.3	3.2	3.8	4.7	5.7	10	-18	-10	-1	2	10		**100**	3	
75N,100E	SON	2.0	2.8	3.1	4.5	5.3	20	-3	3	6	12	18	>100	**95**	**17**	2
	ANN	2.1	2.9	3.4	4.1	5.7	15	-3	0	5	9	14	70	**100**	**21**	2
CNA	DJF	2.0	2.9	3.5	4.2	6.1	30	-18	0	5	8	14		**71**	**7**	
30N,103W	MAM	1.9	2.8	3.3	3.9	5.7	25	-17	2	7	12	17	>100	**81**	**19**	4
to	JJA	2.4	3.1	4.1	5.1	6.4	20	-31	-15	-3	4	20	>100	**93**		**15**
50N,85W	SON	2.4	3.0	3.5	4.6	5.8	20	-17	-4	4	11	24		**91**	**11**	
	ANN	2.3	3.0	3.5	4.4	5.8	15	-16	-3	3	7	15		**98**		
ENA	DJF	2.1	3.1	3.8	4.6	6.0	25	2	9	11	19	28	85	**78**	**24**	
25N,85W	MAM	2.3	2.7	3.5	3.9	5.9	20	-4	7	12	16	23	60	**86**	**23**	2
to	JJA	2.1	2.6	3.3	4.3	5.4	15	-17	-3	1	6	13		**98**		
50N,50W	SON	2.2	2.8	3.5	4.4	5.7	20	-7	4	7	11	17	>100	**97**	**19**	
	ANN	2.3	2.8	3.6	4.3	5.6	15	-3	5	7	10	15	55	**100**	**29**	1
CENTRAL AND SOUTH AMERICA																
CAM	DJF	1.4	2.2	2.6	3.5	4.6	15	-57	-18	-14	-9	0	>100	**96**	2	**25**
10N,116W	MAM	1.9	2.7	3.6	3.8	5.2	10	-46	-25	-16	-10	15	75	**100**	2	**18**
to	JJA	1.8	2.7	3.4	3.6	5.5	10	-44	-25	-9	-4	12	90	**100**		**24**
30N,83W	SON	2.0	2.7	3.2	3.7	4.6	10	-45	-10	-4	7	24		**100**		**15**
	ANN	1.8	2.6	3.2	3.6	5.0	10	-48	-16	-9	-5	9	65	**100**	2	**33**
AMZ	DJF	1.7	2.4	3.0	3.7	4.6	10	-13	0	4	11	17	>100	**93**	**27**	4
20S,82W	MAM	1.7	2.5	3.0	3.7	4.6	10	-13	-1	1	4	14		**100**	**18**	
to	JJA	2.0	2.7	3.5	3.9	5.6	10	-38	-10	-3	2	13		**100**		
12N,.34W	SON	1.8	2.8	3.5	4.1	5.4	10	-35	-12	-2	8	21		**100**		
	ANN	1.8	2.6	3.3	3.7	5.1	10	-21	-3	0	6	14		**100**		
SSA	DJF	1.5	2.5	2.7	3.3	4.3	10	-16	-2	1	7	10		**100**		
56S-76W	MAM	1.8	2.3	2.6	3.0	4.2	15	-11	-2	1	5	7		**98**	8	
to	JJA	1.7	2.1	2.4	2.8	3.6	15	-20	-7	0	3	17		**95**		
20S,40W	SON	1.8	2.2	2.7	3.2	4.0	15	-20	-12	1	6	11		**99**		
	ANN	1.7	2.3	2.5	3.1	3.9	10	-12	-1	3	5	7		**100**		
AUSTRALIA AND NEW ZEALAND																
NAU	DJF	2.2	2.6	3.1	3.7	4.6	20	-20	-8	1	8	27		**89**		
30S,110E	MAM	2.1	2.7	3.1	3.3	4.3	20	-24	-12	1	15	40		**92**		3
to	JJA	2.0	2.7	3.0	3.3	4.3	25	-54	-20	-14	3	26		**94**	3	
11S,155E	SON	2.5	3.0	3.2	3.8	5.0	20	-58	-32	-12	2	20		**98**		
	ANN	2.2	2.8	3.0	3.5	4.5	15	-25	-8	-4	8	23		**99**		
SAU	DJF	2.0	2.4	2.7	3.2	4.2	20	-23	-12	-2	12	30		**95**		
45S,110E	MAM	2.0	2.2	2.5	2.8	3.9	20	-31	-9	-5	13	32		**90**		**6**
to	JJA	1.7	2.0	2.3	2.5	3.5	15	-37	-20	-11	-4	9	>100	**95**		**17**
30S,155E	SON	2.0	2.6	2.8	3.0	4.1	20	-42	-27	-14	-5	4	>100	**95**		**15**
	ANN	1.9	2.4	2.6	2.8	3.9	15	-27	-13	-4	3	12		**100**		
POLAR REGIONS																
ARC*	DJF	4.3	6.0	6.9	8.4	11.4	15	11	19	26	29	39	25	**100**	**90**	0
60N,180E	MAM	2.4	3.7	4.4	4.9	7.3	15	9	14	16	21	32	25	**100**	**79**	0
to	JJA	1.2	1.6	2.1	3.0	5.3	15	4	10	14	17	20	25	**100**	**85**	0
	SON	2.9	4.8	6.0	7.2	8.9	15	9	17	21	26	35	20	**100**	**96**	0

		Temperature Response						% Precipitation Response						Extreme Seasons		
90N,180W	ANN	2.8	4.0	4.9	5.6	7.8	15	10	15	18	22	28	20	**100**	**100**	0
ANT*	DJF	0.8	2.2	2.6	2.8	4.6	20	-11	5	9	14	31	50	**85**	**34**	3
90S,180E	MAM	1.3	2.2	2.6	3.3	5.3	20	1	8	12	19	40	40	**88**	**54**	0
to	JJA	1.4	2.3	2.8	3.3	5.2	25	5	14	19	24	41	30	**83**	**59**	0
60S,180W	SON	1.3	2.1	2.3	3.2	4.8	25	-2	9	12	18	36	45	**79**	**42**	1
	ANN	1.4	2.3	2.6	3.0	5.0	15	-2	9	14	17	35	25	**99**	**81**	1
SMALL ISLANDS																
CAR	DJF	1.4	1.8	2.1	2.4	3.2	10	-21	-11	-6	0	10		**100**	2	
10N,85W	MAM	1.3	1.8	2.2	2.4	3.2	10	-28	-20	-13	-6	6	>100	**100**	3	**18**
to	JJA	1.3	1.8	2.0	2.4	3.2	10	-57	-35	-20	-6	8	60	**100**	2	**40**
25N,60W	SON	1.6	1.9	2.0	2.5	3.4	10	-38	-18	-6	1	19		**100**		**22**
	ANN	1.4	1.8	2.0	2.4	3.2	10	-39	-19	-12	-3	11	60	**100**	3	**39**
IND	DJF	1.4	2.0	2.1	2.4	3.8	10	-4	2	4	9	20	>100	**100**	**19**	1
35S,50E	MAM	1.5	2.0	2.2	2.5	3.8	10	0	3	5	6	20	80	**100**	**22**	1
to	JJA	1.4	1.9	2.1	2.4	3.7	10	-3	-1	3	5	20		**100**	**17**	
17.5N,100E	SON	1.4	1.9	2.0	2.3	3.6	10	-5	2	4	7	21	>100	**100**	**17**	2
	ANN	1.4	1.9	2.1	2.4	3.7	10	-2	3	4	5	20	65	**100**	**30**	2
MED	DJF	1.5	2.0	2.3	2.7	4.2	25	-25	-16	-14	-10	-2	85	**96**	1	**18**
30N,5W	MAM	1.5	2.1	2.4	2.7	3.7	20	-32	-23	-19	-16	-6	65	**99**	0	**32**
to	JJA	2.0	2.6	3.1	3.7	4.7	15	-64	-34	-29	-20	-3	60	**100**	1	**36**
45N,35E	SON	1.9	2.3	2.7	3.2	4.4	20	-33	-16	-10	-5	9	>100	**99**	2	**21**
	ANN	1.7	2.2	2.7	3.0	4.2	15	-30	-16	-15	-10	-6	45	**100**	0	**50**
TNE	DJF	1.4	1.9	2.1	2.3	3.3	10	-35	-8	-6	3	10	>100	**100**		
0,30W	MAM	1.5	1.9	2.0	2.2	3.1	15	-16	-7	-2	6	39	>100	**100**		
to	JJA	1.4	1.9	2.1	2.4	3.6	15	-8	-2	2	7	13	>100	**100**		
40N,10W	SON	1.5	2.0	2.2	2.6	3.7	15	-16	-5	-1	3	9	>100	**100**		
	ANN	1.4	1.9	2.1	2.4	3.5	15	-7	-3	1	3	7	>100	**100**		
NPA	DJF	1.5	1.9	2.4	2.5	3.6	10	-5	1	3	6	17	>100	**100**	**20**	2
0,150E	MAM	1.4	1.9	2.3	2.5	3.5	10	-17	-1	1	3	17		**100**	**14**	
to	JJA	1.4	1.9	2.3	2.7	3.9	10	1	5	8	14	25	55	**100**	**43**	1
40N,120W	SON	1.6	1.9	2.4	2.9	3.9	10	1	5	6	13	22	50	**100**	**31**	1
	ANN	1.5	1.9	2.3	2.6	3.7	10	0	3	5	10	19	60	**100**	**35**	1
SPA	DJF	1.4	1.7	1.8	2.1	3.2	10	-6	1	4	7	15	80	**100**	**19**	4
55S,150E	MAM	1.4	1.8	1.9	2.1	3.2	10	-3	3	6	8	17	35	**100**	**35**	1
to	JJA	1.4	1.7	1.8	2.0	3.1	10	-2	1	3	5	12	70	**100**	**27**	3
0,80W	SON	1.4	1.6	1.8	2.0	3.0	10	-8	-2	2	4	5		**100**		
	ANN	1.4	1.7	1.8	2.0	3.1	10	-4	3	3	6	11	40	**100**	**40**	3

*Notes:
ARC = land + ocean
ANT = land only

Table 11.2. Projected changes in climate extremes. This table summarizes key phemomena for which there is confidence in the direction of projected change based on the current scientific evidence. The included phenomena are those where confidence ranges between medium and very likely, and are listed with the notation of: VL (Very Likely); L (Likely); and M (Medium confidence).

Special notes: In addition to changes listed in the table, there are two phenomena of note for which there is little confidence. The issue of drying and associated risk of drought in the Sahel remains uncertain as discussed in Section 11.2.4.2. The change in mean duration of tropical cyclones can not be assessed with confidence at this stage due to insufficient studies.

Temperature-Related Phenomena	
Change in phenomenon	**Projected changes**
Higher maxTmax, more hot / warm summer days	VL (consistent across model projections) maxTmax increases at same rate as the mean or median[i] over northern Europe[ii], Australia and New Zealand[iii] L (fairly consistent across models, but sensitivity to land-surface treatment) maxTmax increases more than the median over southern and central Europe[iv], and southwest USA[v] L (consistent with projected large increase in mean temperature) Large increase in probability of extreme warm seasons over most part of the world[vi]
Longer duration, more intense, more frequent heat waves / hot spells in summer	VL (consistent across model projections) Over almost all continents[vii], but particularly central Europe[viii], western USA[ix], East Asia[x] and Korea[xi]
Higher maxTmin; more warm and fewer cold nights	VL (consistent with higher mean temperatures) Over most continents[xii]
Higher minTmin	VL (consistent across model projections) minTmin increases more than the mean in many mid-and high-latitude locations[xiii], particularly in winter over most of Europe except the southwest[xiv]
Higher minTmax, fewer cold days	L (consistent with warmer mean temperatures) minTmin increases more than the mean in some areas[xv]
Fewer frost days	VL (consistent across model projections) Decrease in number of days with below freezing temperatures everywhere[xvi]
Fewer cold outbreaks; fewer, shorter, less intense cold spells / cold extremes in winter	VL (consistent across model projections) Northern Europe, South Asia, East Asia[xvii] L (consistent with warmer mean temperatures) Most other regions[xviii]
Reduced diurnal temperature range (DTR)	L (consistent across model projections) Over most continental regions, night temperatures increase faster than the day temperatures[xix]
Temperature variability on interannual and daily time scales	L (general consensus across model projections) Reduced in winter over most of Europe[xx] Increase in central Europe in summer[xxi]
Moisture-Related Phenomena	
Change in phenomenon	**Projected changes**
Intense precipitation events	VL (consistent across model projections; empirical evidence, generally higher precipitation extremes in warmer climates) Much larger increase in the frequency than in the magnitude of precipitation extremes over most land areas in middle latitudes[xxii], particularly over northern Europe[xxiii], Australia and New Zealand[xxiv] Large increase during the Indian summer monsoon season over Arabian Sea, tropical Indian Ocean, South Asia[xxv] Increase in summer over south China, Korea, and Japan[xxvi] L (some inconsistencies across model projections)

	Increase over central Europe in winter[xxvii] Increase associated with tropical cyclones over Southeast Asia, Japan[xxviii] Uncertain Changes in summer over Mediterranean and central Europe [xxix] L decrease (consistent across model projections) Iberian Peninsula[xxx],
Wet days	L (consistent across model projections) Increase in number of days at high latitudes in winter, and over northwest China[xxxi] Increase over the ITCZ[xxxii] Decrease in South Asia[xxxiii] and the Mediterranean area[xxxiv]
Dry spells (periods of consecutive dry days)	VL (consistent across model projections) Increase in length and frequency over the Mediterranean area[xxxv], southern areas of Australia, New Zealand[xxxvi] L (consistent across model projections) Increase in most subtropical areas[xxxvii] Little change over northern Europe[xxxviii]
Increased continental drying and associated risk of drought	L (consistent across model projections; consistent change in P-E, but sensitivity to formulation of land-surface processes) In summer over many mid-latitude continental interiors, e.g. central[xxxix] and southern Europe, Mediterranean area[xl], in boreal spring and dry periods of the annual cycle over Central America[xli]
Tropical Cyclones (typhoons and hurricanes)	
Change in phenomenon	**Projected changes**
Increase in peak wind intensities	L (high-resolution AGCM and embedded hurricane-model projections) Over most tropical cyclone areas[xlii]
Increase in mean and peak precipitation intensities	L (high-resolution AGCM projections and embedded hurricane-model projections) Over most tropical cyclone areas[xliii], South[xliv], East[xlv] and southeast Asia[xlvi]
Changes in frequency of occurrence	M (some high-resolution AGCM projections) Decrease in number of weak storms, increase in number of strong storms[xlvii] M (several climate model projections) Globally averaged decrease in number, but specific regional changes dependent on SST change[xlviii] Possible increase over the North Atlantic[xlix]
Extratropical Cyclones	
Change in phenomenon	**Projected changes**
Changes in frequency and position	L (consistent in AOGCM projections) Decrease in the total number of extratropical cyclones[l] Slight poleward shift of storm track and associated precipitation, particularly in winter[li]
Change in storm intensity and winds	L (consistent in most AOGCM projections, but not explicitly analysed for all models) Increased number of intense cyclones[lii] and associated strong winds, particularly in winter over the North Atlantic[liii], central Europe[liv] and Southern Island of New Zealand[lv] More likely than not Increased windiness in northern Europe and reduced windiness in Mediterranean Europe[lvi]
Increased wave height	L (based on projected changes in extratropical storms) Increased occurrence of high waves in most midlatitude areas analyzed, particularly the North Sea[lvii]

Table 11.3. Methods for generating probabilistic information from future climate simulations at continental and sub-continental scales, SRES – scenario specific. Results from the methods of Greene et al. (2006) and Tebaldi et al. (2004, 2005) are displayed in Figure 11.26.

	Input Type			**Methodological Assumptions**	
Reference	**Experiment**	**Spatial Scale**	**Time Resolution**	**Synthesis Method and Results**	**Model Performance Evaluation**
Furrer et al. (2006)	Multimodel Ensemble	Grid points (after interpolation to common grid)	Seasonal multidecadal averages	Bayesian approach. AOGCMs are assumed independent. Large scale patterns projected on basis functions, small scale modeled as an isotropic Gaussian process. Spatial dependence fully accounted for by spatial model. PDFs at grid point level, jointly derived accounting for spatial dependence	Model performance not explicitly brought to bear.
Giorgi and Mearns (2003)	Multimodel Ensemble	Regional averages (Giorgi and Francisco, 2000)	Seasonal multidecadal averages	Cumulative Distribution Functions derived by counting threshold exceedances among members, and weighing the counts by the REA-method. Stepwise CDFs at the regional levels	Model performance (Bias and Convergence) explicitly quantified in each AOGCMs' weight. Observable at same spatial scale and time resolution, for period 1961-1990.
Greene et al. (2006)	Multimodel Ensemble	Regional averages (Giorgi and Francisco, 2000)	Seasonal and annual averages	Bayesian approach. AOGCMs dependence is modeled. Linear regression of observed values on model's values (similar to Model-Output-Statistics approach used in weather forecasting and seasonal forecasting) with coefficients estimates applied to future simulations. PDFs at regional level	Model performance measured on 1902-1998 historical trend reproduction at same spatial scale and time resolution.
Harris et al. (2006)	Perturbed Physics Ensemble (PPE)	Grid points	Seasonal multiannual averages	Scaled equilibrium response patterns from a large slab-model PPE, using transient responses of an EBM driven by PPE climate feedbacks. Quantifying scaling error, against a smaller PPE of transient simulations, to include in PDFs. PDFs at arbitrary level of aggregation	All model versions assumed equally likely
Stott et al. (2006a)	Single Model (HADCM3)	Continental averages	Annual decadal averages	Linear scaling factor estimated through optimal fingerprinting approach at continental scales or at global scale and applied to future projections, with estimated uncertainty. Natural variability estimated from control run added as additional uncertainty component. PDFs at the continental scale level	Not applicable
Tebaldi et al. (2004, 2005)	Multimodel Ensemble	Regional averages (Giorgi and Francisco, 2000)	Seasonal multidecadal averages	Bayesian approach. AOGCMs are assumed independent. Normal likelihood for their projections, with AOGCM-specific variability. PDFs at the regional level	Model performance (Bias and Convergence) implicitly brought to bear through likelihood assumptions. Observable at same spatial scale and time resolution, for period 1961-1990 in original papers, for period 1980-1999 for results displayed in this report.

[i] Kharin and Zwiers (2005)
[ii] §11.3.3.3, Supplementary material Figure S11.23, PRUDENCE, Kjellström et al. (2006)
[iii] §11.7.3.5, CSIRO (2001)
[iv] §11.3.3.3, PRUDENCE, Kjellström et al. (2006)
[v] §11.5.3.3, Bell et al. (2004),
[vi] Table 11.1
[vii] §11.3.3.3, Tebaldi et al. (2006), Meehl and Tebaldi (2004)
[viii] §11.5.3.3, Barnett et al. (2006), Clark et al. (2006), Tebaldi et al. (2006), Gregory and Mitchell (1995), Zwiers and Kharin (1998), Hegerl et al. (2004), Meehl and Tebaldi (2004)
[ix] §11.5.3.3, Bell et al. (2004), Leung et al. (2004)
[x] §11.4.3.2, Gao et al. (2002)
[xi] §11.4.3.2, Kwon et al. (2005), Boo et al. (2006)
[xii] §11.3.3.2, §11.4.3.1
[xiii] Kharin and Zwiers (2005)
[xiv] §11.3.3.2, Fig. 11.3.3.3, PRUDENCE
[xv] §11.7.3.5, Whetton et al. (2002)
[xvi] Tebaldi et al. (2006), Meehl and Tebaldi (2004), §11.3.3.2, PRUDENCE, §11.7.3.1, CSIRO (2001), Mullan et al. (2001b)
[xvii] §11.3.3.2, PRUDENCE, Kjellström et al. (2006), §11.4.3.2, Gao et al. (2002), Rupa Kumar et al. (2006)
[xviii] §11.1.3
[xix] §11.5.3.3, Bell et al. (2004), Leung et al. (2004), §11.4.3.2, Rupa Kumar et al. (2006), Mizuta et al. (2005)
[xx] §11.3.3.2, Räisänen (2001), Räisänen and Alexandersson (2003), Giorgi and Bi (2005), Zwiers and Kharin (1998), Hegerl et al. (2004), Kjellström et al. (2006)
[xxi] §11.3.3.2, PRUDENCE, Schär et al. (2004), Vidale et al. (2006)
[xxii] §11.3.3.4, Groisman et al. (2005), Kharin and Zwiers (2005), Hegerl et al. (2004), Semenov and Bengtsson (2002), Meehl et al. (2006)
[xxiii] §11.3.3.4, Räisänen (2002), Giorgi and Bi (2005), Räisänen (2005)
[xxiv] §11.1.3, §11.7.3.2, §11.3.3.4, Huntingford et al. (2003), Barnett et al. (2006), Frei et al. (2006), Hennessy et al. (1997), Whetton et al. (2002), Watterson and Dix (2003), Suppiah et al. (2004), McInnes et al. (2003), Hennessy et al. (2004b), Abbs (2004), Semenov and Bengtsson (2002)
[xxv] §11.4.3.2, May (2004a), Rupa Kumar et al. (2006)
[xxvi] §11.4.3.2, Gao et al. (2002), Boo et al. (2006), Kimoto et al. (2005), Kitoh et al. (2005), Mizuta et al. (2005)
[xxvii] §11.3.3.4, PRUDENCE, Frei et al. (2006), Christensen and Christensen (2003, 2004)
[xxviii] §11.1.3, §11.4.3.2, Kimoto et al. (2005), Mizuta et al. (2005), Hasegawa and Emori (2005), Kanada et al. (2005)
[xxix] §11.3.3.4, PRUDENCE, Frei et al. (2006), Christensen and Christensen (2004), Tebaldi et al. (2006)
[xxx] §11.3.3.4, PRUDENCE, Frei et al. (2006)
[xxxi] §11.4.3.2, Gao et al. (2002), Hasegawa and Emori (2005)
[xxxii] Semenov and Bengtsson (2002)
[xxxiii] §11.4.3.2 Krishna Kumar et al. (2003)
[xxxiv] §11.3.3.4, Semenov and Bengtsson (2002), Voss et al. (2002); Räisänen et al. (2004); Frei et al. (2006)
[xxxv] §11.3.3.4, Semenov and Bengtsson, 2002; Voss et al. 2002; Hegerl et al. 2004; Wehner, 2004; Kharin and Zwiers, 2005; Tebaldi et al., 2006
[xxxvi] §11.1.3, §11.7.3.2, §11.7.3.4, Whetton and Suppiah (2003), McInnes et al. (2003), Walsh et al. (2002), Hennessy et al. (2004c), Mullan et al. (2005)
[xxxvii] §11.1.3
[xxxviii] §11.3.3.4, Beniston et al. (2006), Tebaldi et al. (2006), Voss et al. (2002)

[xxxix] §11.3.3.2, Rowell and Jones (2006)
[xl] §11.1.3, §11.3.3.4, Voss et al. (2002)
[xli] §11.1.3
[xlii] Knutson and Tuleya (2004)
[xliii] Knutson and Tuleya (2004)
[xliv] §11.4.3.2, Unnikrishnan et al. (2006)
[xlv] §11.3.4, Hasegawa and Emori (2005)
[xlvi] §11.3.4, Hasegawa and Emori (2005), Knutson and Tuleya (2004)
[xlvii] Oouchi et al. (2006)
[xlviii] Hasegawa and Emori (2005)
[xlix] Sugi et al. (2002), Oouchi et al. (2006)
[l] §11.3.3.6, Yin (2005), Lambert and Fyfe (2006), §11.3.3.5, Lionello et al. (2002), Leckebusch et al. (2006), Vérant (2004), Somot (2005)
[li] §11.1.3, Yin (2005), Lambert and Fyfe (2006)
[lii] §11.1.2, §11.3.3.5, Yin (2005), Lambert and Fyfe (2006)
[liii] §11.3.3.5, Leckebusch and Ulbrich (2004)
[liv] §11.3.3.5, Zwiers and Kharin (1998), Knippertz et al. (2000), Leckebusch and Ulbrich (2004), Pryor et al. (2005a), Lionello et al. (2002), Leckebusch et al. (2006), Vérant (2004), Somot (2005)
[lv] §11.1.3, §11.7.3.7
[lvi] §11.3.3.5, Lionello et al. (2002), Leckebusch et al. (2006), Vérant (2004), Somot (2005)
[lvii] X.L. Wang et al. (2004)