Chapter 11: Regional Climate Projections

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Executive Summary

Introduction
This chapter assess regional climate information from all sources, including Atmosphere-Ocean General Circulation Models (AOGCMs) and various downscaling techniques used to enhance regional detail. These methods have substantially matured since the IPCC WGI Third Assessment Report (IPCC, 2001) (hereafter TAR) and have become widely applied. In several cases large-scale coordination of efforts has been undertaken to conduct multi-ensemble climate change simulations.

The advances in methods have also allowed the important advancement in the understanding and quantification of uncertainty surrounding projections (see Section 11.2.2). In particular the evolving work around developing probability distribution functions (PDFs) based on multi-model ensembles allow for defensible probabilistic interpretations of climate projections and assessment of risk factors. Systematic use of ensemble simulations opens for a possibility to the span the probability envelope of possible climate evolution pathways, while other methods seek to put constraints to possible future change. Collectively these techniques are beginning to provide insight into the combined uncertainties from different sources.

Building on the advances in models and analysis, and within the growing understanding of uncertainty and climate constraints, this chapter is, in contrast to the TAR, in a position to make clear assessments of regional change.

Simulations of present day
The current generation AOGCMs simulate many aspects of the atmospheric and oceanic general circulation well, and have continued to improve since the TAR. The ensemble mean of the global models in the PCMDI/AR4 archive provides a simulation of the present day climate superior in most continental and sub-continental regions to that of any individual model in the archive. Several AGCMs have been applied at high resolution illustrating a general improved performance as the dynamics and large scale flow improve with increases in resolution. The direct consequence has been improved simulation of regional climates in the GCMs, and in the simulated climates using regional climate models (RCMs) nested in GCMs. Similarly, empirical downscaling techniques consistently show skill in deriving accurate local climate representation from the GCM-scale atmospheric forcing.

For RCM based downscaling, multi-model ensemble simulations have demonstrated that ensemble mean biases can be very small, generally temperature biases remain within 1ºC and precipitation biases are less than 30%. While individual models generally are of similar quality as when assessed for TAR, the multi-model ensemble mean performs very well.

Simulations of future climate change
For many of the regions of the world it is now possible to make robust statements as to some of the attributes of the projected change, either based on the direct regional climates from AOGCM and downscaling methods, or on examination of the GCM simulation of the governing large scale processes for a region. The strength and specificity of these statements is region dependant and summarized in Box 11.1. This represents a significant advance over the TAR.

Climate means
- *Temperature projections:* These are comparable in magnitude to those of the TAR, however the confidence in the regional projections is higher than in the TAR due to better statistics (more simulations available), improved models, a better understanding of the role of model deficits, and generally more advanced analyses of the results. As in the TAR, significant warming (in most cases greater than the global mean) is very likely over nearly all landmasses.
- *Precipitation projections:* These are comparable in magnitude to those of TAR, with greater confidence in the projections for some regions. There are indications of convergence between AOGCM models in their regional projections, and in the downscaled projections for some regions. For some regions there are grounds for stating the projected precipitation changes as likely or very likely. To differing degrees, there remains uncertainty in the regional projections depending on the

region. For some regions confidence in the projected change is weak, even in terms of the direction of precipitation change.

Climate variability and extremes

There is a large increase in the available analyses on changes in extremes. This allows for a more comprehensive assessment for most regions in the world (see Chapter 9 on detection issues). The general findings are in line with the assessment made in TAR. However, the increasing number of specialised analyses supply a higher level of confidence compared to the TAR, especially with regard to historical change; notable improvements in confidence relate to the regional statements concerning heat waves, heavy precipitation, and droughts, while changes in wind storms seem highly dependent on detailed regional changes in atmospheric circulation, where a significant convergence between AOGCMs is still lacking.

- *Africa*: All of Africa is very likely to warm during this century; Annual rainfall is very likely to decrease in much of North Africa and Northern Sahara; Winter rainfall will very likely decrease in much of Southern Africa
- *Mediterranean and Europe:* All of Europe is very likely to warm during this century; The lowest winter temperatures are very likely to increase more than the average winter temperature in northern Europe; Annual precipitation is very likely to increase in most of northern Europe and decrease in most of the Mediterranean area; 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;
- *Asia:* All of Asia is very likely to warm during this century;
- *North America:* All of North America is very likely to warm during this century; The lowest winter temperatures are very likely to increase more than the average winter temperature in northern North America; Annual precipitation is very likely to increase in northern part of North America
- *Central and South America:* All of Central and South America is very likely to warm during this century; Annual precipitation is very likely to increase in south eastern South America
- *Australia – New Zealand:* All of Australia and New Zealand are very likely to warm during this century; There will very likely be an increase in rainfall in the South Island of New Zealand; Increased frequency of extreme high daily temperatures, and decrease in the frequency of cold extremes is very likely; Increased risk of drought in southern areas of Australia is very likely
- *Polar:* 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; Annual Arctic precipitation is very likely to increase; Arctic sea ice is very likely to decrease in its extent and thickness
- *Small Islands:* Changes are less well understood than elsewhere

There remains a need for large coordinated efforts to provide better and more comprehensive analysis of climate change in and for many regions. The apparent convergence of projected change over large portions of the World by AOGCMs seem to justify such endeavours. However, in regions where there is a lack of convergence, further insight into the understanding of model deficits is clearly needed. Developing nations are still disadvantaged in the sophistication, clarity, and breadth of climate change projections.

11.1 Introduction

11.1.1 The Need for a Regional Focus and Regional Projections

Scientific understanding of anthropogenic global climate change has advanced notably in recent years, and led to commensurate developments of mitigation strategies. International discussions on mitigation are primarily founded on our present understanding of global-scale change. Opposed to mitigation, adaptation is inherently a local and regional scale issue, and limited by the measure of confidence in the projected changes at these scales. It is at regional scales that credible information of probable climate change and the associated uncertainties is mostly needed. The possible consequences of climate change within some regions may even motivate some countries to commit to and argue for further mitigation practises.

Ideally Global Climate Models (GCMs) should be able to provide information at the regional scale they are able to resolve, but the majority of efforts in model development have been concentrated on improving the ability to describe specific geophysical phenomenons, e.g., El Niño, monsoon systems, sea-ice, etc. thereby at the same time obviously lacking specific attention to certain aspects of model performance in many other regions of the World. Therefore, alternative methods have been developed to derive detailed regional information in response to geophysical processes at finer scales than that resolved by GCMs. Through nested Regional Climate Models (RCMs) or empirical downscaling, these developments in turn have generated new and alternative ways to assess important regional processes central to climate change. This further allows development and validation of models to simulate the key dynamical and physical processes of the climate system.

Within the impacts and adaptation community there is a growing move toward integrated assessment, wherein regional climate change projections form a principal factor for decision support systems aimed at reducing vulnerability (Bales et al., 2004). At present the regional projections are perhaps the weakest link in this process, and the bulk of information readily available for policy and resource managers (such as via the IPCC DDC) is largely derivatives of GCMs, the data of which have limited skill in accurately simulating local scale climates, especially as regards the key parameter of precipitation. GCM data are commonly mapped as continuous fields (as in IPCC, 2001, Chapter 9), which do not convey the low skill of the model for many regions, or are area aggregated (as in IPCC 2001, Chapter 10) which renders the results of little value for local application.

In view of the pressing need for regional projections, much effort has been expended in recent years on developing regional projections through the above mentioned methodologies, and significant advances made to downscale the GCM skilful scale to the regional and local scales, either through high resolution dynamical modelling, or via empirical cross scale functions. However, to date, much of the work remains at the level of methodological development. Climate change projections that are tailored to the needs of the impacts community, and which demonstrate convergence of the projections across different forcing GCMs, are only now beginning to become more available. An additional challenge is to be able to anchor the regional climate projections reasonably well within a given set of emission scenarios, otherwise the notion that climate sensitivity might be more uncertain than previously believed (see e.g., Chapter 10.5) would indicate that regional results would not be important at all, given the large-scale uncertainties.

11.1.2 Summary of TAR

The analysis of regional climate projections in the TAR (IPCC, 2001; Chapter 10) was based upon a thorough discussion of various regionalisation methods. Since the chapter was an entirely new effort compared to the two previous assessment reports; the SAR (IPCC, 1995) and the FAR (IPCC, 1990), most of the effort within the chapter was spent on assessing the strength and weaknesses of these methods, building to a large extent on illustrative examples chosen from various geographical locations. Since at the time only limited efforts had been made to analyse regional climate change projections in a coordinated fashion, the actual projections assessed were also limited. The central results regarding projected changes in seasonal temperature and precipitation were almost entirely based on analysis of the 9 coarse resolution AOGCMs which had performed a transient experiments representing at least the period 1960–2100 with the specifications for the A2 and B2 emission scenarios. In contrast to both the SAR and the FAR where only results for 7 (5) broad continental-scale regions were assessed, 23 sub-continental regions were considered within the TAR. The analysis was restricted to two seasons boreal summer; June-July-August (JJA) and boreal winter; December-January-February (DJF)

11.1.2.1 Simulations of present day climate

The basic findings of the TAR were that the analysed coarse resolution AOGCMs were able to simulate atmospheric general circulation features well in general, but that at the regional scale the models showed highly variable region-to-region area-averaged biases for both temperature – typically within 4ºC of that observed, and precipitation – mostly between –40 and +80% of the observed values. In most cases these biases were improvements when compared to the models assessed within the SAR.

Results from a few high resolution AGCMs that were available at the time strongly suggested that increasing resolution would further improve models' dynamics and large-scale flow, leading to better regional details in the climate simulations. This was supported by the finding that RCMs also operating at substantially higher resolution than AOGCMs consistently improve the spatial details of the simulated climate, and when driven by observed boundary conditions biases are mostly much lower than those of AOGCMs. Likewise statistical downscaling of AOGCM simulations in most cases was assessed to provide enhanced performance for most applications.

11.1.2.2 Simulations of climate change

Based on the available AOGCM information for the period 2071–2100, it was found with some confidence that it is very likely that with a few exceptions (Southeast Asia and South America in JJA) all land areas will warm more than the global average, particularly at high latitudes. The following changes in precipitation were found to be likely: precipitation will increase over northern mid-latitude regions in winter and over high latitude regions in both winter and summer; in DJF, rainfall will increase in tropical Africa, show little change in Southeast Asia and decrease in Central America; there will be increase or little change in JJA over South Asia; and precipitation will decrease over Australia and the Mediterranean region in JJA. Studies with regional models indicate that at finer scales changes may be substantially different in magnitude from these large sub-continental findings.

At the time of the TAR the amount of 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 at the regional level. However, some statements of a more generic nature could be made, but with somewhat lower confidence than for the changes in the mean. For example it was stated that daily to interannual temperatures are likely to decrease in winter and increase in summer for mid-latitude Northern Hemisphere land areas. Daily high temperature extremes will likely increase in frequency. Future increase in mean precipitation will very likely lead to an increase in variability. Extreme precipitation may increase in some regions, but only specially analysed regions were considered. Furthermore, there were indications from simulations that droughts or dry spells may increase in occurrence in some regions (Europe, North America and Australia).

11.1.3 Developments Since the TAR

It is evident that the climate of a given region is determined by the interaction between external forcings and atmospheric and oceanic circulations that occur at many spatial scales, for a wide range of temporal scales. Examples of regional and local scale forcings are those due to complex topography, land-use characteristics, inland bodies of water, land ocean contrasts, atmospheric aerosols, radiatively active gases, snow, sea ice, and ocean current distribution. Moreover, teleconnection patterns such as ENSO and NAO can strongly influence the climate variability of a region. The difficulties related to the simulation of regional climate and climate change are therefore quite apparent. Many of these difficulties troubled a quantitative assessment of projected regional climate changes for both the regional mean state and particularly regarding extreme events and forced TAR to put relatively low confidence in many of the specific regional statements. In the TAR a number of key priorities to address this problem were therefore listed, and progress has been made within most of these priorities.

11.1.3.1 GCMs

GCMs have steadily improved their general performance (compare with Chapter 10) although not necessarily in all regions for all variables analysed, many of the state-of-the-art GCMs has been run for a great range of forcing scenarios (e.g., Chapter 10) and much more attention to both the general performance and aspects of climate change response of these models at the regional scale has taken place since the TAR. Likewise a considerable effort has gone into the analysis of these model simulations in the evaluation of simulated climate variability and extreme events (e.g., Chapter 10.4.3). The 20-model ensemble of global models assembled in the PCMDI/AR4 archive has provided the clearest view to date of which aspects of continental and sub-continental climate changes are robust across models and which are not. Perturbed physics model ensembles (e.g., Murphy et al., 2004; Stainforth et al., 2005) are beginning to add to this information as well. There are more high resolution time-slice studies with uncoupled atmospheric models, ranging up to the 20 km resolution (e.g., Mizuta et al., 2005) but coordinated multi-model time-slice experiments will be needed to optimize the value of these studies for assessments.

11.1.3.2 RCMs

While most of the RCM work on climate change issues dealt with in the TAR only considered simulations of limited duration (months to a decade), with hardly any study exploring time scale beyond a decade (see IPCC, 2001, Appendix 10.3), experiments with RCMs of 20–30 year duration have become standard by many groups around the world (e.g., Christensen et al., 2002; Leung et al., 2004). This has enabled a more stringent validation of their performance in climate mode, and the general quality and understanding of RCM performance for many regions have greatly improved since the TAR (see Section 11.2.1.3). The need for comparative studies using different RCMs to downscale climate change information from GCMs has also been confirmed by the scientific community. Christensen et al. (2001) with later updates by Rummukainen et al. (2003) combined the information from four different RCM climate change experiments for Scandinavia. They showed that by adding information from different runs and applying a simple pattern scaling argument, it became possible to quantify the uncertainty related to projections in the mean climate state, but also for higher order statistics.

In the European initiative PRUDENCE (Christensen et al., 2002; 2005) as many as 10 RCMs were applied to explore the uncertainties in regional climate change projections due to RCM formulation as well as GCM formulation, and scenario specification, as combinations of downscaling experiments from 3 different GCMs and two SRES scenarios were combined. This enabled some first rough quantitative estimates of the uncertainty in climate change projections due to these sources of uncertainty to be made (Deque et al., 2005ab; Frei et al. 2005; Graham et al. 2005; Beniston et al., 2005).

With more studies focusing on the 30 year time scale much more emphasis has been devoted to the analysis of extremes compared to what was available for TAR. As some modelling centres have conducted ensemble simulations with their RCM the data backing for the statistical analysis of extreme event has also improved. Within the PRUDENCE project two groups downscaled three completely independent members from an Hadley Centre ensemble simulation of an A2 scenario (Christensen et al., 2005; Deque et al., 2005). Thereby enabling an analysis based on 90 year of control and scenario instead of two times 30 years.

Another significant change compared to the situation in preparing TAR is that many RCMs have been adjusted to operate at the 20km scale and even finer scales (e.g., Leung et al., 2003ab; Christensen & Christensen, 2004, 2005; Grell et al., 2000). For TAR only one group had efforts at this resolution (e.g., Christensen et al., 1998) representing a period long enough to give climate information. It appears that it is still possible to obtain improved patterns of precipitation for example by increasing the resolution. Figure 11.1.1 demonstrates that in order to depict essential geographical details in the precipitation patterns in the Alps, inter grid distances below 20km may even be required.

[INSERT FIGURE 11.1.1 HERE]

Coupled modelling is the norm in global climate modelling. Steps towards coupled modelling have been taken also in regional climate modelling since TAR (Döscher et al., 2002; Rummukainen et al., 2004; Schrum et al., 2003). In addition to providing a more realistic simulation of climate in regions where water bodies are characterised by sub-GCM detail, it is very useful for studies focusing on coastal regions, the marginal sea ice zone and regional oceans as such (e.g., Döscher and Meier, 2004; Meier et al., 2004).

As mentioned above, many RCMs have since TAR been run for periods of 30 years per time-slice. Few RCMs have even attempted transient experiments, run from some present-day climate through the whole 21st Century (Kwon et al., 2003; Kjellström et al., 2005). Transient RCM-runs improve the means for evaluating pattern-scaling techniques for regional studies, provide coherent regional climate projections for different time horizons and also facilitate regional-scale impact studies dealing with topics that are affected by the transience (e.g., ecosystems and forestry).

11.1.3.3 Empirical/statistical[1] downscaling

At the time of the TAR empirical downscaling was viewed as a complementary technique to RCMs for downscaling regional climate, each approach having respective strengths and weaknesses. This situation, with some caveats, remains largely unchanged, although the plethora of empirical and statistical techniques in use at the time of the TAR (IPCC, 2001, Appendix 10.4) has greatly expanded in the subsequent years. This situation is indicative of the urgent need for scenarios by the impacts community. Empirical techniques are additionally attractive due to computational efficiencies and because of the ability to downscale directly to attributes that are not readily available from an RCM (e.g., streamflow; Cannon and Whitfield, 2002). However, unlike the RCM community, there has been little development of coherent multi-technique research programmes assessing the relative merits of different empirical techniques.

Development of understanding of the relative strengths and weaknesses of empirical downscaling has to some degree advanced with a number of studies assessing the utility for different applications (for example, Wilby et al., 2002; Salathe, 2003, or Mehrotra et al., 2004). There remains, however, much downscaling work that goes unreported, where downscaling is implemented for the pragmatic purpose of serving a project need, rather than explicitly for use in a broader scientific community, this is especially the case in developing nations. In some cases this work is only found within the project literature, for example, the AIACC project (http://www.aiaccproject.org/), which supports impact studies in developing nations.

11.2 Assessment of Regional-Climate Projection Methods

11.2.1 Generating Regional Information

This section describes the main approaches to generate regional-scale climate-change projections. These can roughly be divided into two classes: dynamical and empirical. The dynamical approach employs physically based numerical climate models, either fully coupled atmosphere-ocean global models (CGCM) or, in dymanical-downscaling mode, atmosphere-only global models (AGCM), of uniform or variable resolution, and nested regional climate models (RCM). The empirical-downscaling approach employs statistical downscaling and pattern scaling of climate projections from climate models.

11.2.1.1 CGCM results

Global General Circulation Models of the atmosphere and land-surface, coupled with ocean and sea-ice components (CGCMs), represent the corner stone of efforts at simulating the global climate system. As is evident from the prominent role that they occupy in this report, CGCMs are the primary tool in studies of the maintenance and evolution of the climate, its natural variability and its response to external forcing. Because of the computational expense of integrating these models for several simulated centuries, a cost that increases rapidly with increasing horizontal resolution, CGCMs employ rather coarse computational meshes: horizontal resolutions of the atmospheric components of the CGCMs in the AR4 range roughly from 400 km down to 125 km.

[1] Within the literature the terms empirical and statistical downscaling are often used interchangeably. Although there are distinctions that may be drawn between the terms, pragmatically they both refer to the dependency on historical data for formulating the cross-scale relationships (in contrast to dynamical models which use a core base on explicit formulation of atmospheric physics and dynamics).

The process of regional-scale climate-change assessment begins of necessity with an evaluation of the ability of models to simulate changes in climate. Weighting of different models according to their strength in simulating present climate is only part of the issue; robustness of climate-change response and responsible mechanisms across models is also important (Giorgi and Mearns, 2002, 2003). While some physical processes are robust in CGCMs simulations, for others the spread is large, particularly at regional scales. While small spread does not necessarily imply small uncertainty, a large spread makes attempts at regional downscaling quixotic. An attempt is made in this chapter to provide information about the spread in CGCMs' projections for each of the regions in Section 11.3.

Studies of environmental, societal and economic impacts associated with anticipated climate changes would benefit from spatially detailed information at scales finer than is currently feasible with CGCMs. A variety of methods are used to "downscale" the climate-change scenarios generated by CGCMs: "time-slice" simulations of AGCMs and RCMs, and empirical/statistical techniques applied upon projections from CGCMs, AGCMs or RCMs. The main advantage of dynamical downscaling approach (AGCM, RCM) is that it is physically based, and hence has the potential for providing added value, particularly for situations in which local changes are produced by processes with spatial scales that are not captured by CGCMs (such as sharp land-sea or land-use contrasts), and for capturing nonlinear effects (such as mesoscale circulations) under perturbed forcing conditions; their main drawback is computational cost. Empirical methods on the other hand require limited computational resources; they rely however on the assumption that statistical relationships that prevail under current climate will remain under perturbed climate. A practical drawback of statistical methods is that they need long time series of reliable, homogeneous station data to develop the statistics; for many regions of the world, such data does not exist. The geographical distribution of stations may be far from optimal for coverage (e.g., along shore), making them non-representative of surrounding conditions.

11.2.1.2 High-resolution AGCMs

Atmosphere-only climate models (AGCMs) can increase their horizontal resolution beyond that utilised in current CGCMs. The lower boundary conditions (BC) required by AGCMs over oceans (temperature and sea ice) are prescribed from observations or CGCMs' simulations. With AGCMs multiple simulations are not required to fine-tune the atmosphere-ocean fluxes as with CGCMs, and only decades rather than centuries are required to obtain satisfactory climate statistics. As a result, AGCMs' resolutions of 100 km and finer have become feasible at many facilities; a resolution of 50 km will likely be the norm for AGCMs in the near future (Cubasch, 1995; Bengtsson, 1996; Brancovic and Gregory, 2001; May 2001; Déqué and Gibelin, 2002; Govindaswamy, 2003). The largest existing computational resources now allow global time slice computations at 20 km resolution (ref).

In high-resolution simulations, the most dramatic improvements occur because of the better simulation of orographic forcing on variables such as precipitation, but there are also improvements in polar climate, monsoonal circulations and mid-latitude weather systems (Boyle, 1993; Déqué and Piedelievre, 1995; Lal, 1997; Stendel and Roeckner, 1998; Stratton, 1999; Duffy et al., 2003; Geng and Sugi, 2003; Iorio, 2004). On the scale typical of current CGCMs, nearly all quantities simulated by higher resolution models agree better with observations (Duffy et al., 2003). Because tropical waves, as well as hurricanes and typhoons, are of smaller scale than typical midlatitude weather systems, tropical meteorology is often an important focus of higher resolution climate simulations (e.g., Bengtsson, 1995, earth simulator ref).

As a result of the absence of two-way feedback between the atmosphere and ocean in AGCMs, climatic variability could be distorted, due to the increased thermal damping of low-frequency internal atmospheric variability (Bretherton and Battisti, 2000). There is also growing evidence that the decoupling can cause significant distortion of the climate over the Indian ocean and the South Asian monsoon. Due to the difference in the resolution of AGCMs and CGCMs, their large-scale climate responses also run the risk of being different, leading one to question the consistency of the oceanic lower BC. In practice, however, the large-scale responses appear to be similar in many regions, lending confidence that the time-slice approach with AGCMs can be considered a valid downscaling technique.

An alternative to uniform high-resolution AGCMs is that of variable-resolution (including stretched-grid) AGCMs (VRGCM; e.g., Déqué and Piedelievre, 1995; Fox-Rabinovitz et al., 2001, 2005; McGregor et al., 2002; Gibelin and Déqué, 2003). The VRGCM approach is attractive as it permits to achieve, within a unified modelling framework, a regional increase of resolution while retaining the full interaction of all regions of the globe. Constraints must be satisfied for accurate results, including using a conservative stretching factor between adjacent grid points and keeping resolution outside the region of interest to no less than that typical of CGCMs. When respecting these constraints, VRGCMs results display some ability at capturing, over the high-resolution region, finer scale details that are out of reach for the coarser uniform-resolution models, while retaining global skill similar to uniform-resolution simulations with the same number of grid points. Numerical artefacts due to stretching have been shown to be small when using modest stretching factors, at least in simplified context (e.g., Lorant and Royer, 2002). Appropriately adjusting subgrid-scale parameterizations represents a challenge within the VRGCM framework; some groups promote the use of an accessory intermediate uniform-resolution grid to compute diabatic tendencies (e.g., Fox-Rabinovitz et al., 2005) to minimise the potential negative interaction of parameterisations with non-uniform resolution.

The modest improvements with increased resolution in some aspects such as convection-dominated continental summer precipitation, suggest that improvements in the physical parameterisations are also required beyond simple increase in resolution. Land-surface processes, through their interaction with the overlaying atmosphere, play an important role in determining continental climate. Soil freezing processes can have significant effects on regional boreal climate (e.g., Poutou et al., 2004). Wetlands and lakes occupy a large fraction of mid- and high-latitude continents, and yet are not accounted for in most CGCMs. Wetlands seem to play a more important role than lakes in cooling the boreal regions in summer and in humidifying the atmosphere (e.g., Krinner, 2003). Overly coarse vertical resolution is a remaining problem in most AGCMs, potentially masking some of the anticipated benefits from increased horizontal resolution.

11.2.1.3 Nested RCMs

The development of nested, limited-area, regional climate models (RCMs) has been motivated by the desire to perform high-resolution climate simulations at the most affordable computational cost. The principle behind regional climate modelling is that an RCM can generate realistic regional climate information that is consistent with the driving large-scale atmospheric circulation if the following premises are satisfied: (1) time-varying atmospheric fields (winds, temperature and moisture data), supplied either by analyses or GCMs, are provided as lateral BC, and sea surface temperature and sea ice are provided as lower BC, (2) subgrid-scale physical processes are suitably parameterised, and (3) fine-scale surface forcings (such as orography, land-sea contrast, land use) are suitably resolved at the high resolution. The first successful demonstration was realised by Dickinson et al. (1989) and Giorgi and Bates (1989). A feature distinguishing AGCMs and RCMs is that the former are only constrained via oceanic lower BC from CGCMs, while the latter are additionally constrained by atmospheric lateral BC from CGCMs or AGCMs; in both cases the interactions are one-way only, without feedback from the high-resolution atmosphere back onto the driving model state.

Nested models have been used extensively for short-range numerical weather prediction. Unlike global models RCMs, owing to their finite domain size, require closure at their largest resolved scale, an issue that has traditionally been addressed as a physical-space, boundary-value problem (e.g., Davies, 1976). The difficulties associated with the implementation of lateral BC are well documented (e.g., Warner et al., 1997). The traditional mathematical interpretation is that nested models represent a fundamentally ill-posed boundary-value problem. These difficulties can be compounded in RCMs owing to the length of the simulations. The control exerted by lateral BC 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 weather regime, mid-tropospheric flow through the domain, location and season; for example, the control is weak in mid-latitudes summer, particularly in absence of topographic forcing, for fields such as precipitation. In some applications, the flow developing within the RCM domain may become incoherent with the nesting BC; the phenomenon is referred to as "intermittent divergence in phase space", and is analogous to the classical predictability limits of initial-value problems with global models. Following earlier work with spectral RCMs (Kida et al., 1991; Waldron et al., 1996), von Storch et al. (2000) and Biner et al. (2000) have published results of RCM simulations in which the large scales of the RCM are forced in the interior to satisfy the nesting fields throughout the RCM's domain. The resulting so-called "large-scale nudging" has the advantage of ensuring consistency of large-scale features in RCM and nesting GCM; in practice it has the additional benefit of reducing the numerical noise near the lateral boundaries. Large-scale nudging has also been used as a kind of poor-man assimilation system, to reconstruct historical weather analyses from low-resolution objective analyses.

Several fundamental issues of RCMs have been reviewed in Wang et al. (2004). One concerns the predictability of nested models: whether RCMs can generate meaningful small-scale features that are absent in the lateral BC. With a simplified approach nicknamed the Big-Brother Experiment (BBE; de Elía et al., 2002), RCMs have been found to be able to recreate the right amplitude of small-scale features that are absent in lateral BC, but is incapable of reproducing it with a root-mean-square measure of error. This implies that RCMs could add value to climate statistics rather than to daily weather events; this has since been confirmed for several seasons and regions (Denis et al., 2002, 2003; Antic et al., 2004; Dimitrijevic and Laprise, 2005). In multi-year ensemble simulations, RCMs have been shown to have skill in reproducing interannual variability in precipitation and surface air temperature, although the skill varies strongly with regions and seasons, being weakest in summer over continents (Vidale et al., 2003). The ultimate proof of the validity of the nested approach rests in RCMs' skill to simulate climate with fidelity. Over the past decade, RCMs have been applied successfully to several regions around the world, to simulate recent past climate as well as climate-change projections. Typical RCM grid mesh for climate-change projections is around 50 km, although some climate simulations have been performed at higher resolutions, with meshes such as 20 km. The aforementioned BBE studies have revealed that criteria for the spatial and temporal resolution of nesting information and RCMs' resolution are intricately related (at least in the case without large-scale nudging): for example, a 45-km mesh RCM requires nesting data to satisfy a minimum resolution equivalent to T30 and a maximum time interval of 12 hours.

Since the ability of RCMs to simulate the regional climate depends strongly on the realism of the large-scale circulation that is provided at the lateral BC (e.g., Pan et al., 2001; de Elía et al., 2006), reduction of errors in GCMs remain a priority for the climate modelling community. For example, Latif et al. (2001) and Davey et al. (2002) have shown strong biases in the tropical climatologies of CGCMs, which would impact negatively on downscaling studies for several regions of the world. Overall the skill at simulating current climate has improved with AR4 CGCMs, which will lead to higher quality of BC for RCMs; it is important to note however that, unless otherwise indicated, RCMs results reported in this AR4 are mostly based on simulations driven by TAR-generation CGCMs. Continued efforts are required to further improve parameterisations in regional and global models. RCMs are increasingly coupled interactively with other components of the climate system, such as regional ocean and sea ice, hydrology, and some work has been initiated with interactive vegetation. Coarse vertical resolution is a remaining problem in several RCMs, potentially masking some of the benefits from increased horizontal resolution.

11.2.1.4 Physically based off-line downscaling

Another downscaling technique has been applied to represent the effect of fine-scale variability in land surface and terrain height. The physically based off-line downscaling (PBOLD) technique consists in off-line running a detailed set of physical parameterisations fed by atmospheric fields from a prior CGCM or RCM simulation. Higher resolution details are achieved in two possible ways, either by using a set of multiple terrain-elevation and land-surface classes within each climate model grid cell (e.g., Ghan et al., 2002, 2006; Leung and Ghan, 2005), or by using a spatially distributed finer resolution grid for applying the physical parameterisations (e.g., Goyette and Laprise, 1996). In either variant, the full column atmospheric and land-surface physical parameterisations are applied, with the orographic forcing on temperature and water vapour determined from an estimated vertical displacement of air parcels given the atmospheric stability and detailed terrain elevation. The PBOLD approach permits the physically based representation of fine-scale surface heterogeneities that would be computationally prohibitive to resolve with a fully coupled high-resolution climate model.

One disadvantage of the PBOLD technique is its limited ability to represent rain shadows, the maximum simulated precipitation occurring at higher elevation with similar amounts at the same elevation on the windward and leeward sides of mountain ranges; Goyette and Laprise (1996) proposed an ad hoc orographic-lift term based on the projection of the low-level wind velocity on the local slope to alleviate this problem. Although the off-line strategy does not permit feedback of the downscaled variables on the driving climate model, Ghan et al. (2002) have shown that the neglected effects are generally smaller than model biases. The same study showed that simulations of CAM2 with PDOLD are clearly superior for surface air temperature and precipitation, and particularly significant for snow because of its extreme sensitivity to temperature, and hence surface elevation, around the freezing point.

11.2.1.5 Empirical/statistical downscaling

A complementary technique to RCMs is the use of empirically derived relationships linking large-scale atmospheric variables (predictors) and local/regional climate variables (predictands). This technique, commonly referred as empirical or statistical downscaling (SD), is analogous to the "perfect prog" approaches and "model output statistics" (MOS) used for short-range numerical weather prediction (Wilby et al., 2000). The local/regional climate-change information is obtained by applying the derived relationships to equivalent variables from GCM simulations.

The main advantages of SD techniques is that they are computationally inexpensive,can be used to derive variables not available from RCMs, and allow downscaling to the point scale. As with RCMs, care is required in application, and key assumptions and limitations need to be recognized. The IPCC TGICA guidance document (Wilby et al., 2004) provides a comprehensive background to using this approach and covers important issues to be addressed in any robust downscaling. Important elements to be highlighted include; The predictors relevant to the local predictand should be realistically modelled by the GCM; the statistical relationship between predictands and predictors has to remain valid for future altered climate or non-stationarity appropriately accommodated; and the predictors should sufficiently incorporate the future climate-change signal. As SD techniques, on the face of it, are easily implemented, a concern remains that not all SD applications fully address all aspects for a robust solution.

Methodological issues aside, the main pragmatic limitation is the need for historical observational data that comprehensively spans the natural variability of the climate. Such data are not available for some regions. Important developments in SD research have been done since the TAR reflecting a maturing of the approach and implementation in climate impact studies. Developments include: increased availability of downscaling tools for the impacts community (e.g., SDSM, Wilby et al., 2002), use of generic downscaling techniques in novel ways (exotic variables such as phenological series and plant disease: Matulla et al., 2003; Seem, 2004); extreme events (e.g., Katz et al., 2002; Wang et al., 2003; Seem, 2004), inter-comparison studies evaluating statistical methods (e.g., STARDEX), downscaling from multi-model and multi-ensemble simulations in order to express climate-model uncertainty alongside other key uncertainties (e.g. Benestad, 2002a,b; Hewitson and Crane, 2005), and accommodation of non-stationarity in climate relationships with conservative methodologies (Hewitson and Crane, 2005).

The SD models can be grouped in three categories: regression models, weather classification, and weather generators. Each of these approaches has relative strengths and weaknesses as fully outlined in the TGICA guidance document (Wilby et al., 2004).

11.2.1.5.1 Methodological approaches

Regression models represent linear or nonlinear relationships between predictands and large-scale predictors. Linear techniques include; multiple regression (Benestad, 2002a,b; Hansen-Bauer et al., 2003; Matulla et al., 2003; Palutikof et al., 2002; Bartman et al., 2003; Huth et al., 2001; 2003, 2005), canonical correlation analysis (CCA) (Bartman et al., 2003; Benestad, 2001; Busuioc et al., 2001; 2003; Chen and Chen, 2003; Penlap et al., 2004, Lionello et al., 2003) and singular value decomposition analysis (SVD) (Widmann et al., 2003; Huth, 2002). Non-linear regression models based on artificial neural networks (ANNs) allow fitting a more general class of statistical models (e.g., Schoof and Pryor, 2001; Cavazos et al., 2002; Hewitson and Crane, 2002; Trigo and Palutikof, 2001). Regression models have been used to derive statistics of a range of local variables such as probability of rainfall occurrence, precipitation / wind distribution parameters, frequency of extreme events, percentiles of rainfall /wave height (e.g., Abaurrea and Asin, 2005, Beckmann and Buishand, 2002, Buishand et al., 2004, Busuioc and von Storch, 2003, Diaz-Nieto and Wilby, 2005, Wang et al., 2004, Wang and Swail, 2004, Pryor et al., 2005). The main weaknesses of the regression methods are poor representation of the high frequency component of variance.

Weather generators (WGs) are a mature approach for generating synthetic sequences of local variables that replicate their observed statistical attributes (such as the mean and variance) but not necessarily the observed sequences of events (e.g., Abaurrea and Asin, 2003; Buishand et al., 2004; Huth et al., 2001; Busuioc and von Storch, 2003; Katz et al., 2003; Palutikof et al., 2002; Wilby et al., 2002c, 2003; Diaz-Nieto and Wilby, 2005; Pryor et al., 2005). Generally these models focus on the daily time scale, as required in many impact studies and are commonly WGs are adapted for statistical downscaling by conditioning their parameters on large-scale atmospheric predictors. In many cases weather generators continue to be the method of choice for agricultural applications and have been the subject of several comparisons (Mavromatis and Hansen, 2001; Qian et al., 2004). These studies show that crop responses based on weather generators can be sensitive to assumptions about the extent and nature of variability of the derived weather sequences under climate change.

The statistical downscaling methods based on the occurrence of generalized weather states relate local or regional climate variables to weather patterns. Methods range from analogues (e.g., Beersma and Buishand, 2003) to objective methods (e.g., Cavazos et al., 2002; Hewitson and Crane, 2002, 2005) or subjective classification (e.g., Palutikof et al., 2002; Risbey et al., 2002). The relationships to weather patterns can be either in terms of the mean response, or explicitly accommodate the stochastic component by sampling the PDF of the local response to the weather mode. Advantages of this approach include the fact that climate change is estimated as a direct function of the frequency of circulation patterns—a more skilful attribute of GCMs. Hewitson and Crane (2005) show how this can achieve significant convergence between the downscaled regional change projections of different GCMs. In addition, this method can reproduce both the low and high frequency components of the variance, including extreme events, and has been extended to both multi-site and multi-variate series (e.g., Palutikof et al., 2002; Hewitson and Crane, 2005)An extreme form of weather typing is the analogue method (see 11.2.1.6). Beersma and Buishand (2003) presented an extension of the analogue method by using a non-parametric nearest-neighbour re-sampling technique to generate multi-site sequence of daily temperature and precipitation.

Weather classification is also used in statistical-dynamical downscaling (SDD) (e.g., Fuentes and Heimann, 2000) that combines the two approaches (statistical and dynamical). In this approach a RCM is used to simulate local climate from similar episodes of different weather classes, and the results then statistically evaluated using the frequency of occurrence. An advantage of the SDD technique over other Sds is that it specifies a complete, dynamically coherent, three-dimensional climate state.

11.2.1.5.2 Issues in statistical downscaling

Since the TAR a growing number of studies analysed the sensitivity of local/regional climate-change scenarios to the selection of downscaling models and predictors (e.g., Beckmann and Buishand, 2002; Benestad, 2002; Cavazos and Hewiston, 2005; Diaz-Nieto and Wilby, 2005; Hansen-Bauer et al., 2004; Huth, 2003; Trigo and Palutikof, 2001). These studies have highlighted the need for care in implementation in the same manner care in needed in RCM applications with the choice of paramaterizations and tuning. Notable is the necessity for care in the choice of predictors in relation to the nature of the local predictand. At a minimum the predictors should be reasonably represented in the GCM, and have a relevant, physical, and interpretable relationship to the predictand, and reflect the climate change signal. In most cases this will require dynamical and moisture variables. The position and size of the predictor domain is also important (e.g., Benestad, 2001; Brinkmann, 2002). The best choice of predictors is to combine dynamical and moisture variables. Other studies have shown that using the GCM-simulated precipitation as a predictor for can improve skill (Salathé, 2003; Widmann et al., 2003), although subject to the skill of the GCM precipitation.

As with RCMs, evaluation of the SD technique is crucial for obtaining a reliable climate-change scenario. Most commonly this is through cross-validation of the SD relationships with observational data from an independent data set for a period that could represent an independent or different “climate regime” (e.g., Busuioc et al., 2001; Trigo and Palutikof, 2001; Hansen Bauer et al., 2003).

Stationarity remains a concern with SD, as to whether the relationships are valid under future climate regimes, and is only weakly assessed through cross-validation tests. A convergence of the climate-change signals across CGCMs, RCMs and Sds can further strengthen the results (e.g., Hewitson and Crane, 2005). More recently, the degree of non-stationarity in a projected climate change has been assessed as part of a SD application (Hewitson and Crane, 2005).

The choice of SD technique will determine the degree to which different aspects of temporal variance (especially extremes) can be derived. Most appropriate are methods that I both low and high frequency components of the variance (e.g., Beersma and Buishand, 2003; Katz et al., 2003; Busuioc and von Storch, 2003; Palutikof et al., 2002; Wang et al., 2004; Lionello et al., 2003; Hewitson and Crane, 2005; Wilby et al., 2003; Hansen and Mavromatis, 2001; Katz et al., 2003).

Most importantly it needs to be recognized that feedbacks are not accommodated in SD downscaling, other than to the degree that feedbacks may be addressed through any stochastic component of the SD method. For example, under weak synoptic forcing feedbacks from vegetation, may play an important role. As such SD techniques reflect first order response of the regional climate to the GCM simulated large scale forcing.

11.2.1.6 Pattern scaling of climate model simulations

Pattern-scaling methods allow obtaining regional climate-change scenarios for a large number of forcing scenarios for which CGCM simulations are not available, by combining CGCM-simulated patterns with simple climate models (SCM) results. The approach involves normalising CGCMs' response patterns according to the global mean temperature. These normalised patterns are then rescaled using a scalar derived from SCM under all forcing scenarios of interest.

This pattern-scaling method, first suggested by Santer et al. (1990), was then developed using various versions of scaling techniques (e.g., Christensen et al., 2001; Mitchell, 2003; Ruosteenoja et al., 2005; Salathé , 2005). For example, Ruosteenoja et al. (2005) developed a super-ensemble pattern-scaling method using linear regression to represent the relationship between the local CGCM-simulated temperature and precipitation response and the global mean temperature change simulated by the SCM MAGICC (IPCC, 2001, Appendix 9.1). In order to reduce the noise induced by the GCM internal variability (common problem to all scaling methods), the scaling was carried out using an ensemble mean instead of an individual GCM response. The method was applied for 6 CGCMs and PRUDENCE RCMs.

11.2.1.7 Other methods

There are alternative techniques for generating high-resolution climate-change scenarios, other than the application of RCM and SD schemes presented above. These approaches include the spatial interpolation of grid-point data to the required local-scale, construction of spatial/emporal analogues using historic climate data (Gangopadhyay et al., 2005), and the use of simple change factors/simple scaling procedure (e.g., Diaz-Nieto and Wilby, 2005; Hansen Bauer et al., 2003; Widmann et al., 2003).

Climate-change analogues are developed from climate records that may be similar to the future climate for a given region. The analogue can originate from either past climate data (temporal analogue) or from another region (spatial analogue). A major advantage of the analogue approach is that the future climate scenario and associated impacts may be described at greater temporal and spatial resolutions than might otherwise be possible. A disadvantage is that the analogue model cannot make any projections outside the range of already measured values.

One of the most popular procedures for rapid impact assessment involves the use of a "change factor". This technique consists in adding the change (against the reference climatology) of the equivalent climate variable for the CGCM grid-box closest to the target site, to each day in the reference period. A disadvantage of this method is that the scaled and baseline scenarios only differ in terms of their respective means; all other parameters such as temporal variability remain unchanged in the future. The procedure also assumes that the spatial pattern of the present climate remains unchanged. The method does not easily apply to precipitation record because the addition (or multiplication) of observed precipitation by CGCM precipitation changes can affect the number of rain days, the size of extreme events, and even result in negative precipitation amounts.

11.2.1.8 Inter-comparison of SD downscaling methods

Many studies comparing several SD techniques (Buishand et al., 2004; Diaz-Nieto and Wilby, 2005; Matulla et al., 2003; Huth, 2002; Widmann et al., 2003; Wilby et al., 2002, 2003; Wood et al., 2004) as well as SD with CGCMs/dynamical downscaling (e.g., Huth et al., 2001; Hansen Bauer et al., 2003; Wilby et al., 2000; Wood et al., 2004) have been performed since the TAR.

In general, conclusions from comparing different SD techniques are dependent on region and criteria used for comparison, and on the inherent attributes of each SD methodology As regards temporal resolution, it is apparent that when comparing the merits of daily and monthly downscaling, daily models are preferable (e.g., Buishand et al., 2004). In terms of non-linearity in downscaling relationships, Trigo and Palutikof (2001) noted complex non-linear models are not necessarily any better than more simple linear / slightly non-linear approaches.

Since the TAR only a few studies have systematically compared the two approaches. A comparison by Wilby et al. (2000) noted the sensitivity to the choice of downscaling technique, although the SD and RCM approaches have comparable skill in reproducing the current climate. Similarly Hanssen-Bauer et al. (2003) found the SD and RCM climate signal to be quite similar. However, a major question over the findings of most inter-comparison studies is to what extent are findings transferable between locations and time periods? Nonetheless, at present the conclusion of the TAR that SD and RCM downscaling techniques are comparable would appear to still hold, even while both methodological approaches have matured and become more skilful.

11.2.2 Quantifying Uncertainties

11.2.2.1 Sources of regional uncertainty

There are numerous sources of uncertainty in projections of regional climate change. Most are the same as those on the global scale (discussed in Chapter 10, Section 5), so we give only a brief overview of these here.

The three major sources include the trajectories of future emissions and other sources of anthropogenic changes, such as land use and cover, the response of the climate system (as represented in climate models and in their components representing atmospheric chemistry and the carbon cycle) to the radiative forcing of the atmospheric concentrations of gases and aerosols derived from these emissions, and the effects of natural variability on multiple timescales. Regarding emissions, for the most part these result in well-mixed gases that have no strong regional distribution in and of themselves. However, the short lifetimes of aerosols in the atmosphere coupled with the uneven geographical distribution of the emission of their precursor chemicals results in them having a strong regional component, and thus may count as an uncertainty in regional forcings per se (see Chapter 2, Section 2.4). Land use/cover change is another important forcing that is inherently regional in scope (De Fries et al., 2002).

The second major component of uncertainty is the response of the climate system to these emissions as represented in climate models. These include uncertainties in the conversion of the emissions into concentrations of radiatively active species (i.e., via atmospheric chemistry and carbon-cycle models) uncertainty in the radiative forcing for known concentrations (particularly large for aerosols) and the uncertainties in the response of the physical climate system to these forcings resulting from incomplete representation of resolved processes (e.g., moisture advection), in the parameterizations of sub-grid-scale processes (e.g., clouds, precipitation, planetary boundary layer), in the feedback mechanisms on the global and regional scale (e.g., changes in land-use/cover affecting the atmosphere) and so on.

The regional impact of these uncertainties in the response of the climate system can be well illustrated with a few examples. Cox et al. (2000) showed that incorporating a model of the carbon-cycle into a coupled AOGCM gave a dramatically enhanced response to climate change over the Amazon basin. Kumagi et al. (2004) demonstrated similar results for the tropical rainforest in Borneo. Pope and Stratton (2002) show that the scale of the resolved processes in a climate model can significantly affect its simulation of climate over large regional scales. Similarly, 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. (2003).

One specific aspect of modelling uncertainty, which is important at regional scales, is the increasing sequence of models used to provide spatially and/or temporally detailed information. The techniques of these models' use are detailed in Section 11.2.1. Clearly uncertainties derive from both the choice of technique and the specific model(s) applied.

Uncertainty in observations, particularly as we consider higher and higher resolution simulations, is also an issue. Whether the model is reproducing correctly the climate becomes a difficult question when there are insufficient or differing observational datasets. Thus methods that include an assessment of the reliability of models when constructing future climate projections need to account for uncertainties in or lack of observations.

Finally, the inherent variability of the climate system should be included in any characterisation of the climate of a particular region over a given period. As a result, in the assessment of the uncertainties in the projections of future climate, the resulting spread in these should be compared to natural climate variability. In climate model experiments the natural internal variability is often explored by creating ensembles of simulations by varying the initial conditions of each run (see Chapter 10.5 for a more complete discussion). When assessing the likelihood that the climate in a particular period in the future will have certain consequences, the uncertainty in the projection of how this climate might change should be assessed alongside the natural variability in the climate over this period.

11.2.2.2 Quantifying regional uncertainty

11.2.2.2.1 Review of regional uncertainty portrayed in the TAR

In the Third Assessment Report (IPCC, 2001) uncertainties in regional climate projections were discussed, but methods for quantifying them were relatively primitive. For example in the chapter on regional projections (Giorgi et al., 2001), uncertainties in regional projections of climate change from different AOGCMs were qualitatively portrayed (e.g., large or small increases/decreases in precipitation) based only on simple agreement heuristics (e.g., 7 of the nine models showed increases). Other 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 (Hulme and Carter, 2000). Some early work in providing probabilistic estimates of regional climate change was portrayed in the chapter on climate scenario development (Mearns et al., 2001).

New and Hulme (2000) and Jones (2000) provide examples of advancing beyond the scenario approach to uncertainty to a probability-based approach by attaching probabilities to a group of scenarios (on a regional scale). Regarding uncertainty in results from regional models, Pan et al. (2001), evaluated the future climate produced by two different regional models, nested in the same AOGCM, in a three-way comparison. Since then, however, much more work has been accomplished in the area of quantifying uncertainties in regional climate change.

There is still much less work on regional scales compared to that produced on the global scale (see Chapter 10, Section 5). In general, large ensembles of projections from full AOGCMs are necessary to produce probabilistic estimates of sub- continental scale regions; and until very recently, sufficient computer resources have not been available for generating large ensembles.

11.2.2.2.2 Using multi-model ensembles

A number of studies have taken advantage of the growing number of AOGCMs that have run the same climate experiments, resulting in multi-model ensembles, to generate probabilistic information on a regional scale. Table 11.2.1 summarizes aspects of the methods reviewed below, together with methods described in section 11.2.2.2.3, and Figure 11.2.1 compares probability density functions (PDFs) from some of these methods for selected regions.

It is important to note that multi-model ensembles do not necessarily explore completely the uncertainty that may exist, for example, based on the full range of climate sensitivity (see Chapter 10, Box 10.2 on uncertainty in climate sensitivity). They explore only the range of climate sensitivity represented by the particular set of models making up the ensemble. For example, in the methods described below that use the results of climate model simulations developed for the IPCC AR4 and available on the PCMDI web site (www.pcmdi-llnl.gov/ipcc/about_ipcc.php)), the range of sensitivity of the models that produced simulations based on the three SRES scenarios chosen for the AR4, does not cover the extremes of the various distributions of sensitivity discussed in Chapter 10. The 5–95% confidence interval for climate sensitivity of the models ranges from about 2 to 4.4°C (Räisänen, 2005) whereas the tails of the distributions (i.e., 95th quantile) of climate sensitivity determined by other means can exceed 6°C (see Box 10.2 on Climate Sensitivity, Chapter 10). Also, the distribution of AOGCM sensitivities is arbitrary and not intended to be consistent with probabilities derived for climate sensitivity. Thus, regional probabilities generated using multi-model ensembles should be viewed as relatively conservative quantities that do not represent well, particularly, the right tail of the future regional PDFs of climate.

Räisänen and Palmer (2001) used 17 members of the CMIP2 experiments (forced with 1% annual increase in CO2) and calculated the probability of exceedance of certain values of temperature increase (> 1 °C) and relative change in precipitation exceeding some threshold (e.g., < –10%), on a model grid point level. While their goal was not to produce regional probabilities of climate change per se, their paper demonstrated that a probabilistic interpretation of climate change has advantages over conventional deterministic interpretations.

Räisänen (2005) sets the goal of producing continuous Probability Density Functions (PDFs) at the grid point level. (See Chapter 10, Section 5 for additional discussion of this work). This method is applied by assuming equal weighting among AOGCMs, but examples are also shown of how it could be generalized to the case where "bad models" are eliminated from the ensemble. Even though the method is designed to produce PDFs at the high resolution level of the model grid points, it can be adapted to derive PDFs of regionally aggregated values. The red curves in Figure 11.2.1 show examples of PDFs derived by this method.

[INSERT FIGURE 11.2.1 HERE]

Criteria have also been developed to provide differential weighting of the individual model members within a multi-model ensemble. Giorgi and Mearns (2002) took the results of the 9 AOGCMs that had appeared in the TAR and developed two criteria for weighting the individual AOGCM contribution to the final estimates and measures of uncertainty for regional temperature and precipitation change. These criteria are bias (i.e., a particular instance of constraining forecasts with observations) and convergence (how close the model projection of change is to the central tendency of the aggregated model projections). They developed estimates separately for the A2 and B2 SRES emission scenarios for 22 large sub-continental regions. While their reliability ensemble averaging (REA) method does quantify uncertainty on a regional scale, the method is not a probabilistic one. Giorgi and Mearns (2003) went on to produce cumulative probability distributions (CDFs) using the REA method by adapting a probability of incremental threshold exceedance approach similar to that adopted by Raisanen and Palmer (2001).

Tebaldi et al. (2004, 2005) approached probabilistic projections at regional scales by stating formal statistical models for an ensemble of projections, for a given season and SRES scenario. A Bayesian approach was adopted, by which current and future regional climate signal (in the form of multi-decadal averages) and model reliabilities are treated as uncertain quantities, starting with uninformative (i.e., flat) prior distributions, which are updated using data (model projections and observations) via Bayes' theorem. Posterior PDFs of temperature and precipitation change signals are thus obtained, and the relative contribution of the individual models to this final result is a function of the models' biases with respect to current climate observations and models' convergence (based on Giorgi and Mearns, 2002, 2003). Under the assumption that the natural variability remains constant at the value estimated from the observed regional records, PDFs that include this additional component of uncertainty can be derived from the posterior of the climate change signal. The blue curves in Figure 11.2.1 show examples of the PDFs derived by this method, with the addition of climate variability, whose interpretation and relevance for impacts considerations are more immediate than the posterior PDFs shown in their original paper.

Furrer et al. (2005) extended the framework used in Tebaldi et al. (2004, 2005) by modelling the high resolution fields as produced by the AOGCMs (after interpolation to a common grid), also in a Bayesian framework. (See Chapter 10, Section 5.5 for details.) The final product is a high-dimensional joint probability distribution of the field of seasonal temperature and precipitation change, and a straightforward aggregation in area-averages can produce regional PDFs.

Greene et al. (2005) used a Bayesian framework to model an ensemble of AOGCM projections under individual SRES scenarios by an extension of the – by now traditional – methods for seasonal forecasting. AOGCMs projections are extracted as seasonal annual values, and smoothed to extract low frequency trends. After a selection step, only a few AOGCMs are retained, on the basis of their performance over all regions, and across seasons. This subset of AOGCMs is used in a regression framework with observed data, and coefficient estimates and their uncertainty are derived and then applied to the smoothed future projections. The green curves in Figure 11.2.1 show examples of PDFs derived by this method, representing posterior distributions of the climate change signal, not accounting for natural variability.

Dessai et al. (2005) apply the idea of simple pattern scaling (Santer et al., 1990), to a super 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). Thus they can estimate PDFs of regional change on the basis of a high number of samples, and explore their sensitivities to SRES scenarios, climate sensitivity and AOGCM weighting through a skill score that they turn on or off in the sampling stage. Their work is focused on measuring the changes in PDFs as a function of the different sources of uncertainty. The impact of the SRES scenarios turns out to be the most relevant for temperature changes, while the AOGCM weighting does not produce substantial differences. Climate sensitivity has an impact mainly in the tails of the distributions. For precipitation changes, all sources of uncertainty seem relevant but the results are very region-specific and thus difficult to generalize.

Other studies have taken advantage of ensembles of opportunity, and derived estimates of climate change through statistical methods, but they fail to provide formal PDFs. Thus Boulanger et al. (2005) demonstrate how a calibration of current simulations from different AOGCMs, at the gridpoint level can be performed through neural network processing. They apply their method to the entire South American region, fitting full spatial patterns of temperature and precipitation to observed fields. The resulting coefficients are then applied to future projections, but without producing an estimate of their uncertainty. The same method can be applied to arbitrary regions.

Laurent and Cai (2005) use the Maximum Entropy method in order to explore optimal combinations of AOGCMs. Their study aims at demonstrating the range of possible solutions to the problem of “optimal fit”. They argue that a number of parameterizations are equally supported by the data, if the specific goals of future climate projections is taken into account. By using the Central US region as an example they show how focusing on intra-annual variability of climate variables – especially relevant for agricultural impact assessment – leads them to a specific range of parameterization for the statistical model. The uncertainty of the final estimate, however, is not addressed

Figure 11.2.1 compares PDFs estimated by the methods in Tebaldi et al. (2005), Räisänen (2005), Greene et al. (2005), as well as the empirical distribution of the AR4 AOGCM responses in the form of a histogram, with 0.5 degree bin size. Three regions are chosen as representatives of high latitudes (Northern Asia, NAS), mid latitudes (Central North America, I) and low latitudes (West Africa, WAF) . PDFs of temperature change are shown for the A2 scenario, for December through February (DJF) and June through August (JJA). The shapes and relative positions of the curves in each panel are similar under A1B and B1 (not shown), only the absolute values of change are modulated, as is expected, with smaller changes being projected under the lower-emission scenarios.

A comparison across regions and seasons produces well established results: larger warming in the higher latitudes and, for those, significantly larger warming in winter than in summer. For lower latitudes the difference in warming is not significant across the two seasons. Higher variability/uncertainty in the mid-high latitudes is associated with larger ensemble spread (and PDF width) compared to the lower latitudes. A comparison of the curves within each region highlights significant differences in the three methods’ results.

Räisänen (2005), and Tebaldi et al. (2005) when adding the natural variability component, more closely fit the empirical distribution of the AOGCMs. For the first method, this result is consistent with the choice of giving equal weight to every AOGCM and of constructing the estimate in a way that is robust to (accommodating of) outliers. For Tebaldi et al. (2005) the close fit to the histogram is the effect of the method's rewarding the models' agreement, producing PDFs whose central location is consistent with the ensemble average location (and therefore with Räisänen's) . The width being so close to the histogram's is in part the result of the order of magnitude of the inter-model variability being the same as the natural variability estimated, for most of the regions. Thus, even if the method tends to discount AOGCMs at the "fringe" of the ensemble the added uncertainty from the natural variability includes those all the same in most cases.

The Greene et al. (2005) method appears to produce PDFs of narrower width with respect to the other two methods, as an indirect consequence of performing a selection of the AOGCMs based on their current climate performance, but more importantly because the method does not model the uncertainty due to natural variability, quantifying the range of uncertainty around the climate signal only. The striking shift of the green curves for 5 out of 6 panels can be traced back to a fundamental difference in the fitting method. Greene et al. choose to calibrate AOGCMs' trend over the observed period through a linear regression framework, and apply the estimated coefficients forward, to the trends of the 21st Century. The "average warming" for the end of the century is then derived as a by-product, so that nothing ensures that the range of the point-wise estimates of such warming by the AOGCMs coincide with the range of the PDFs. For example, if models produce a stronger trend than observed in the 20th century over a region the calibration will tend to deflate the 21st century simulated trends. When combined to produce PDFs of temperature change this would likely produce conservative warming projections when compared to the actual AOGCMs' projections. Given how new most of these methods are, it is not surprising that there is no clear consensus on the 'best' method for developing probabilities.

Figure 11.2.2 extends the comparison of methods for deriving PDFs to the set of 23 regions that encompass most land areas of the world. For each region three PDFs obtained through alternative methods are represented by color bars. Each bar represents the range of values covering 90% of the probability of temperature change under the A2 SRES scenario, in DJF. The upper bar is derived by the method described in Greene et al. (2005), the middle bar is derived by computing the empirical frequency of the AOGCM responses into half degree bins, and the lower bar is derived by the method described in Tebaldi et al. (2005). The color range and gradient in each bar represents the location and width of the distribution between its 5th and 95th percentiles. The color scale is common to all bars, in order to allow comparison across methods and regions.

[INSERT FIGURE 11.2.2 HERE]

The simplest interpretation of each bar is that with 90% chance the temperature change in the region will be between the two values that can be "read" on the reference color scale as corresponding to the extremes of the bar. Similar probability statements can be construed by translating into numerical values the colors corresponding to the three other quantiles marked within the bar by white lines (25%, median and 75%).

From a more qualitative standpoint, the dominant hues in the bar are indicative of the absolute values of change within the region, and allow a comparison between regions warming more or less in absolute value, and between methods projecting more or less conservative changes. Additionally, the full spectrum of colors in a bar is suggestive of the width of the distribution, with relatively wider color ranges indicative of wider distributions, reflecting relatively larger uncertainty in the projections.

The results already highlighted by Figure 11.2.2 for three indicative regions can be gathered from this full representation as well: the method by Tebaldi et al. is in general similar in location (dominant color) and width (color range) to the empirical distribution represented in the middle bar. The PDFs for Greene et al. are in general narrower (as indicated by more homogeneous color ranges within the upper bars compared to the two other bars), not representing – by design – the natural variability component of the uncertainty. Also, for most of the regions Greene et al. projections tend to be more conservative in absolute values, the shades of color in the upper bars being generally cooler than the shades in the other two bars. While this figure portrays results only for the A2 emissions scenarios, the results of the other scenarios behave in a similar fashion (i.e., show the same patterns across the methods), but the values are shifted to the left, with lower mean values, since the emissions are lower in these other two scenarios.

The former studies have been developed either for large area averages of temperature and precipitation change, or for applying statistical modelling directly at the grid point scale. Good and Lowe ( ) tackle the link between the two levels of AOGCM output. They examine the relation between ensemble statistics derived by aggregating precipitation fields into area averages, and statistics derived at the highest level of resolution, i.e., grid point, trying to answer the question of representativeness of the former when aiming at the characterization of changes in precipitation pattern. The answer is in general that area-averages of precipitation produce trends that are often very different from what is produced at the finer scales, but the study finds a stable relation between sub-regional scale variability of the trends and inter-model variability, in a framework similar to pattern-scaling. PDFs of precipitation change are not derived, but the study claims to serve the impact research community nonetheless, by characterizing the uncertainty – in terms of non-representativeness – of the commonly produced statistics and exploring fine scale behavior and its linkages to trends at aggregated scales.

11.2.2.2.3 Using perturbed physics ensembles

Another recent development in designing ensembles of GCMs or AOGCMs is to vary the parameters of one AOGCM and generate multiple runs from the various parameter combinations. In the first systematic application of this idea, Murphy et al. (2004) perturbed 26 parameters in the representations of key atmospheric and surface processes in a version of the Hadley Centre atmospheric model coupled to a mixed layer ocean model and constrained the resulting probability density function (pdf) for climate feedback parameters like climate sensitivity with estimates of the relative reliability of the models derived from a Climate Prediction Index (CPI). They calculated PDFs for climate sensitivity both weighting with the CPI and assuming equal weight for each ensemble member (see chapter 10 for a more complete discussion).

Recent work by Harris et al. ( ) has developed a bridge between spatially complex regional projections and the equilibrium response derived from large slab-model ensembles obtained through Perturbed Physics Experiments, by way of simple pattern scaling (Santer et al. 1990). See Chapter 10 for details. Regional probability distributions of climate change are derived, under assumptions on the distribution of the net scaling error for each slab-model projection. In particular, PDFs of annual temperature change at the regional scale are derived under a 1% CO_2 increase scenario. Precipitation pattern scaling is in principle possible but the same study highlights non-linearities and thus less accuracy in the scaled projections, advocating a more complex, possibly non-linear approach to pattern scaling for this variable.

11.2.2.2.4 Other approaches to quantifying regional uncertainty

As described in Chapter 10, Stott and Kettleborough (2002) combined observational constraints on a climate model's response to greenhouse gas and sulphate aerosol forcings with its predictions of the future evolution of the climate under various emissions scenarios to provide constraints on future global temperature change. The study by Stott et al. (2005) is the first to adapt this method for the regional (or continental in this case) scales. This method applies linear scaling factors to AOGCM projections, after deriving them for current climate simulations through observational constraints, evaluated at the regional level. Differently from the studies described in Section 11.2.2.2, this strain of work uses projections from a single AOGCM, specifically HADCM3. The regional projections so derived are compared to scaled projections using factors computed at the global scale. The first approach produces wider PDFs, since the uncertainty of detection at the regional scale, forming the basis of the estimate of the scaling factors, is larger. A factor of additional variability is added in both cases representing the natural variability estimated from a control run of the same climate model.

11.2.2.2.5 Combined uncertainties: AOGCMs, emissions, and downscaling techniques

The combined uncertainty of regional projections across different forcing scenarios and different AOGCMs has been the focus of most regional uncertainty analysis. However, as mentioned in the introduction to this section, there is an additional uncertainty presented when AOGCMs are used as the starting point for statistical or dynamical downscaling. There is abundant evidence indicating the contrasts in projections from the regional results of an AOGCM and the results from a regional model which took its boundary and initial conditions from that AOGCM future experiment (see Section 11.x above). However, there has been little done so far quantifying the relative importance of the uncertainty from the downscaling step against the other sources of uncertainty (AOGCM, emissions pathway, and internal variability of the climate system). The PRUDENCE project provided the first opportunity to weigh these various sources of uncertainty for simulations over Europe. Rowell (2004) evaluated a 4 dimensional matrix of climate modelling experiments that included two different emissions scenarios, 4 different GCM or AOGCM experiments, 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. For precipitation the relative contributions of the four sources of uncertainty are more balanced. Deque et al. (2005) show similar results for the whole of Europe, as do Ruosteenja et al. (2005) for subsections of Europe. However, it should be noted that few of the RCMs in PRUDENCE were driven by more than one AOGCM, leaving some uncertainty regarding these conclusions. Other programs similar to PRUDENCE have begun for other regions of the world, such as NARCCAP over North America (Mearns et al., 2004), and CREAS over South America.

11.3 Regional Projections

11.3.1 Introduction to Regions and Relationship to WGII Regions

This section considers climate change projections on a region by region basis, which includes key regional processes, skill of models in simulating current regional climate and projections of future regional climate change. The discussion is organised according to the same regions used for discussion of impacts in WG II in the AR4 and earlier assessments: Africa, Europe and Mediterranean, Asia, North America, Central and South America, Australia-New Zealand, Polar Regions, and Small Islands. These regions are continental-scale (or based on large oceanic regions with a high density of inhabited islands) and may have a broad range of climates and be affected by a large range of climate processes. As they are generally too large to be used as a basis for conveying quantitative regional climate change information, this section also utilises the subdivision of continental and key oceanic regions indicated in Figure 11.3.1.1. Thus the continental regions introduced above are subdivided into a number of sub-continental regions, e.g., Africa is comprised of the Saharan, East African, West African and South African regions. In particular, this regionalisation is used for presenting sub-regional area-averaged precipitation and temperature change information from the new AR4 AOGCM simulations.

This regionalisation is very close to that initially devised by Giorgi and Francesco (2000) and in the TAR but includes additional oceanic regions and some other minor modifications similar to those of Carter et al. (2000) and Ruosteenoja et al. (2003). The objectives behind the original Giorgi and Francesco (2000) regions were that they should have simple shape, be no smaller than the horizontal wave length typically resolved by GCMs (judged to be a few thousand kilometres), and should recognise where possible distinct climatic regimes. Although these objectives may be met with alternative regional configurations, as yet there are no well developed options in the regional climate change literature.

[INSERT FIGURE 11.3.1.1 HERE]

Several common processes underlie climate change in a number of regions. Before proceeding to discuss regions individually, we briefly summarize some of these.

The first is a fundamental consequence of warmer temperatures and the increase in water vapor in the atmosphere (Chapter 3). Water is continually 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, and regions of convergence will get wetter and regions of divergence drier. We see the consequences of this increased moisture transport in plots of the global response of precipitation (Chapter 10), where, on average, precipitation increases in the intertropical convergence zones, decreases in the subtropics, and increases in sub-polar and polar regions. Regions of large uncertainty often lie near the boundaries between robust moistening and drying regions, with different models placing these boundaries differently.

Another important theme in the extratropics is the poleward expansion of the subtropical highs, and the poleward displacement of the midlatitude westerlies and the associated storm tracks. This circulation response is often referred to as the excitation of the positive phase of the Northern or Southern Annular Mode, or when focusing on the North Atlantic, as the positive phase of the North Atlantic Oscillation. Superposition of the tendency towards subtropical drying and poleward expansion of the subtropical highs creates especially robust drying responses on the equatorward boundaries of the 5 subtropical oceanic high centers in the South Indian, South Atlantic, South Pacific, North Atlantic and (less robustly) the North Pacific. Most of the regional projections of strong drying tendencies in the 21st century are associated with the land areas immediately downstream of these centers (Southwestern Australia, the Western Cape Provinces of South Africa, the central Andes, the Mediterranean, and (less robustly) Mexico. For discussions of our level of confidence that these circulation shifts will occur, see the discussion in Chapter 10.

A familiar theme wherever snow and ice are present is the implications for local climates of the retreat of snow and ice cover. The difficulty of quantifying these effects in regions of substantial topographic relief is a significant limitation of global models and an aspect that one hopes to improve with dynamical and statistical downscaling. The drying effect of earlier timing of the 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 well-known control that sea surface temperature anomalies exerts on tropical rainfall variability provides an important unifying theme for tropical climates,. At several points below reference is made to the "ENSO-like" change in Pacific Ocean temperatures predicted in the majority of the CGCMs, and the implications that a shift of this character has on regional precipitation projections for the 21st century.

[Placeholder for additional Table.]

[In order to assess projections of changes in extremes, regional as well as larger-scale issues require consideration. It is at the regional scale that the consequences of such changes will be felt. An additional table will address the current assessment of different forms of extreme events. For each phenomenon (e.g. heat waves), the table will summarize observed changes in the past century, model simulations of change in the phenomenon for the 20th century, and corresponding projections for change during the 21st century. Regional specificity will be provided where supporting evidence allows. The table will incorporate and synthesize material from this and other chapters as appropriate, taking into account reviews of relevant material in the first draft of this report.]

[START OF BOX 11.1]

Box 11.1:

For each of the 22 land regions (Giorgi et al., 2001) extended with an Artic region as well as regions representing small islands located in the major oceans (as delineated on Box 11.1, Figure1; see also 11.3.1), an evaluation of the quality of the PCMDI simulations for the 20C3M have been made and the projected climate change for the A1B scenario based on the same set of models have been used along with additional material to generate regional statements about the probable projection of climate change by the end of the 21st Century. Box 11.1, Figure 1 summarizes and highlights changes in regions, where general model quality, general model projection agreement, physical understanding and additional material suggests that changes are very likely or likely to occur. In the following extracts from the individual analyses for each of the regions are provided, and in most cases, statements are based on regionally averaged values from the individual continental scale regions with adequate sub divisions (see 11.3.2–11.3.9), further informed by by physically based arguments and supporting analyses from the literature that goes beyond the PCMDI data set. There are many robust regional changes that are comparable across the regions, but there are also clear differences. Therefore, each of the continental scale regions is treated separately here.

[INSERT BOX 11.1, FIGURE 1HERE]

Key processes

A fundamental consequence of warmer temperatures and the increase in water vapor in the atmosphere, water is continually transported horizontally by the atmosphere from regions of moisture divergence (particularly in the subtropics) to regions of convergence. The consequences of this increased moisture transport is, on average, precipitation increases in the intertropical convergence zones, decreases in the subtropics, and increases in sub-polar and polar regions.

In the extratropics, a poleward expansion of the subtropical highs, and the poleward displacement of the midlatitude westerlies and the associated storm tracks results in strong drying tendencies associated with the land areas immediately downstream of these centers.

Wherever snow and ice are present it has implications for local climates of the retreat of snow and ice cover. The drying effect of earlier timing of the spring snowmelt, and, more generally, the earlier reduction in soil moisture furthermore contributes to changes summertime continental climates.

Sources of information

The main source of information for this regional assessment stems from studies analysing AOGCMs (1), RCMs (2), and statistical downscaling (3). Likewise, physically plausible mechanisms such as mentioned above (4) add to the confidence in the statements.

Africa:

1. **All of Africa is very likely to warm during this century. The warming is likely to be somewhat larger than the global, annual mean warming throughout the continent and in all seasons, with drier subtropical regions (especially arid zones) warming more than the moister tropics.** Based on: 1 and 4.
2. **Annual rainfall is very likely to decrease in much of North Africa and Northern Sahara.** Based on: 1 and 4.
3. **Winter rainfall will very likely decrease in much of Southern Africa.** Based on: 1, 2, 3, and 4.
4. **There will likely be an increase in annual mean rainfall in tropical and East Africa** Based on: 1, 2, 3, and 4.
5. **It is uncertain how rainfall in the Sahel and the Southern Sahara will evolve in this century.** Based on: 1, 3, and 4.

Mediterranean and Europe:

1. **All of Europe 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 Europe, warming is likely to be largest in winter, and in the Mediterranean area in summer.** Based on: 1, 2, 3, and 4. However, the uncertainty in the Atlantic THC indicates a small possibility of cooling in northwestern Europe.
2. **The lowest winter temperatures are very likely to increase more than the average winter temperature in northern Europe, and the highest summer temperatures are likely to increase more than the average summer temperature in southern and central Europe.** Based on: 1, 2, and 4.
3. **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.** Based on: 1, 2, 3, and 4. Furthermore, process studies suggest that changes in atmospheric circulation and drying of soil in summer both contribute to the seasonal cycle in central Europe.
4. **Extremes of daily precipitation will very likely increase in northern Europe.** Based on: 1, 2, 3, 4, and empirical evidence (generally higher precipitation extremes in warmer climates).
5. **The annual number of precipitation days is very likely to decrease in the Mediterranean area** Based on: 1, 2, 3, and 4.
6. **Risk of summer drought is likely to increase in central Europe and in the Mediterranean area, because of reduced summer precipitation and increased potential evaporation.** Based on: 1, 2, 4, and process studies (evaporation efficiency increases with increasing temperature).
7. **It is uncertain whether and how wind storm frequency and/or intensity will change.** Based on: 1.
8. **Snow season length and snow depth are very likely to decrease in most of Europe.** Based on: 1, 2, and 4.

Asia (Presently only Southeast)

1. **All of Southeast Asia is very likely to warm during this century, but the annual mean warming is likely to be slightly less than the global mean warming in most areas.** Based on: 1, 2, and 4.

2. **On average, annual precipitation and wet season precipitation is likely to increase across Southeast Asia.** Based on: 1. Due to strong interactions between atmospheric circulation and topography some local deviations are expected from general trends
3. **Extreme rainfall and winds associated with tropical cyclones are likely to increase in Southeast Asia.** Based on: 1 and 2. Result may be affected or offset by changes in tropical cyclone numbers.

North America

1. **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 North America, warming is likely to be largest in winter, in the South-West USA in summer.** Based on: 1, 2, and 4. However, uncertainty associated with the Atlantic THC implies a small possibility of cooling in extreme northeastern part of North America.
2. **The lowest winter temperatures are very 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.** Based on: 1, 2, and 4.
3. **Annual precipitation is very likely to increase in northern part of North America, and likely to decrease in the South-West USA.** Based on: 1, 2, and 4.
4. **From southern British Columbia south-eastward along the USA-Canada border, precipitation is likely to increase in winter but decrease in summer.** Based on: 1, 2, and 4.
5. **Snow season length and snow depth are very likely to decrease in most of North America.** Based on: 1, 2, and 4.

Central and South America

1. **All of Central and South 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. Increases in temperature in Central America will likely be more evident during dry periods. In southern South America and Amazonia warming is likely to be largest in austral summer.** Based on: 1 and 4.
2. **In Central America it is likely that relatively dry periods of the annual cycle will become drier. It is likely that boreal spring will correspond to drier conditions and the decrease in precipitation during the Mid Summer Drought will be more intense.** Based on: 1 and 4.
3. **Annual precipitation is likely to decrease in Southern Andes.** Based on: 1 and 4. A caveat on the local scale is that changes in atmospheric circulation may induce large local variability in precipitation changes in mountainous areas. Tierra del Fuego exhibits an opposite response (precipitation likely increases).
4. **Annual precipitation is very likely to increase in south eastern South America, with a relative increase in precipitation during austral summer.** Based on: 1 and 4.
5. **It is uncertain how annual and seasonal mean rainfall will change over northern South America.** Based on: 1 and lack of understanding of processes (e.g., biogeochemical feedbacks). However, in some regions the majority of simulations suggest consistent results (rainfall would increase in Ecuador and northern Peru, and would decrease in the northern tip of the continent and in southern northeast Brazil).

Australia and New Zealand

1. **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.** Based on: 1 and 4.

2. **Annual rainfall is likely to decrease in Southern Australia in winter and spring.** Based on: 1 and 4.
3. **There will very likely be an increase in rainfall in the South Island of New Zealand.** Based on: 1 and 4.
4. **Changes in rainfall in Northern and Central Australia are uncertain.** Based on: lack of consensus in AOGCM simulations, the often inadequate simulations of the climatology of the monsoonal rains in this region, and the dependence of the rainfall trends in this region on the uncertain changes in the tropical Pacific Ocean SSTs.
5. **Increased mean windspeed across the southern island of New Zealand, particularly in winter, is likely.** Based on: 1.
6. **Increased frequency of extreme high daily temperatures, and decrease in the frequency of cold extremes is very likely.** Based on: 1, 2, and 4.
7. **Extremes of daily precipitation will very likely increase.** Based on: 1, 2, and 4. The effect may be offset or reversed in areas of significant decrease in mean rainfall (southern Australian in winter and spring.)
8. **Increase in potential evaporation is likely.** Based on: 1. The effect is primarily due to increased temperature.
9. **Increased risk of drought in southern areas of Australia is very likely.** Based on: 1, 2, and 4.

Polar

1. **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 likely to be largest in winter.** Based on: 1, 2, and 4. Support also by climate observations and paleo-climate reconstructions.
2. **Annual Arctic precipitation is very likely to increase. It is very likely that the precipitation increase is largest in the cold seasons.** Based on: 1 and 4. Support by positive AO trend over the past century, and most particularly in the last decade or two.
3. **It is likely that the Antarctic will be warmer and wetter although the magnitude is uncertain.** Based on AOGCM 1. Important uncertainties remain: natural variability; present-day simulations are hard to compare with observational data; recent observed warming (cooling) trend over Peninsula (rest of Antarctic)
4. **Arctic sea ice is very likely to decrease in its extent and thickness; see Chapter 10.** Based on: 1 and 4. Important uncertainties remain: Large present-day sea ice simulations scatter and limited ice thickness observations.
5. **It is uncertain how the Arctic Ocean will change.** Based on: Lack of systematic analysis of future projections of the Arctic Ocean. Present-day simulations are still unsatisfactory. The resolution of AOGCMs are still not adequate to resolve some important processes in the Arctic Ocean.
6. **It is uncertain to what extent the frequency of extreme temperature and precipitation events will change in the Arctic.** Based on: a small amount of material. Difficulty of validation of present-day temperature & precipitation, and therefore uncertain present-day PDF.

Small Islands

1. **Temperatures in the Caribbean Islands are likely to increase with a mean of about 2ºC by the end of the century.** Based on: 1, 3 and 4.
2. **Temperatures in the small islands of the Pacific are likely to increase.** Based on: 1
3. **Temperatures in the small islands of the Indian Ocean are likely to increase**. Based on: 1
4. **Change in precipitation in the small islands of the Pacific is uncertain.** Based on: 1
5. **Change in precipitation in the small islands of the Indian Ocean is uncertain.** Based on: 1
6. **Change in precipitation on the Caribbean islands is highly uncertain.** Based on: 1 and 3

[END OF BOX 11.1]

11.3.2 Africa

11.3.2.1 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. Even moderate variations in these rain belts, given agrarian societies and population pressures, can have profound impacts (Maynard et al., 2002). There are also regions on the northern and southern boundaries of the continent with winter rainfall regimes governed by the passage of mid-latitude depressions, that are therefore sensitive to the poleward displacement of the mid-latitude storm tracks, as is evident from the strong correlation between South African rainfall and the Southern Annular Mode (Reason, 2005) and between North African rainfall and the North Atlantic Oscillation (Lamb and Peppler, 1987) Troughs penetrating into the tropics from mid-latitude depressions also influence warm season rainfall, especially in Southern Africa, and can contribute to a sensitivity of warm season rains to the poleward displacement of the circulation as well (Todd et al., 2004; Todd and Washington, 1999). The southeast coastal regions, including the island of Madagascar, are vulnerable to topical cyclones which, given the subsistence nature of society in much of this region, have proven catastrophic in the past (Reason and Keible, 2004).

There are many pathways through which changes in the surrounding oceans can alter African climates. Southern Africa is bordered on the west by the cool Benguela current with regionally strong upwelling, contributing to the aridity in the southwest (Washington et al., 2003; Reason, 2002), and on the east by the energetic and warm Agulhas current. The warm Indian Ocean supplies most of the water for rainfall in Southern Africa, and affects the East African rains as well (Black, 2003). The seasonal Arabian Sea upwelling and Somali current, which are sensitive to the strength of the Indian monsoon, help shape the climate of the Horn of Africa. The North Atlantic, with its distinctively variable, and potentially sensitive overturning circulation, together with the waters of the Gulf of Guinea, controls the location of the Atlantic Intertropical Convergence Zone and influences rainfall in West Africa and the Sahel. Moisture supply from the Mediterranean affects not only local climates but has been shown to be important for Sahel rainfall, despite the intervening Sahara (Rowell, 2003). The correlations between ENSO and seasonal rainfall in Southern Africa (Rautenbach and Smith, 2001) and the Sahel (Janicot et al., 2001) remind us of the interconnectedness of tropical climates and the potential role of the Indo-Pacific oceans in the maintenance of African rainfall patterns.

The factors that determine the Southern boundary of the Sahara and rainfall in the Sahel have attracted special interest because of the profound drought experienced by this region in the 1970’s and 80’s. As discussed in Chapters. 8 and 9, the field has moved steadily away from explanations for variations in this region as due primarily to land use changes. A recent and thorough attempt to estimate land use changes over the latter part of 20th century and to simulate the response in a GCM shows discernible reduction in precipitation, but only of 5%, from 1960 to the 1990’s (Taylor et al., 2002), small compared to the observed drying in this period. It has become steadily more plausible that Sahel precipitation variations and trends can instead be understood to a first approximation as forced by changes in sea surface temperatures (SSTs), as early SST perturbation AGCM experiments (Folland et. Al., 1986) are continually being updated with impressive results from the most recent models (Giannini, et. Al., 2003; Hoerling, et al, 2005; Lu and Delworth, 2005), This does not imply that land surface changes play no role, but that they primarily act as feedbacks generated by the underlying response to SST anomalies. The key feature of the SST changes thought to be important for the Sahel is the north-south inter-hemispheric gradient, with a colder North Atlantic, and warmer Indian, South Atlantic and Gulf of Guinea conducive to an equatorward shift and/or a reduction in Sahel rainfall.

The focus on changes in the inter-hemispheric SST gradient has created interest in the possibility that aerosol cooling localized in the Northern hemisphere could enhance drying in this region. The work of Rotstayn and Lohmann (2002), supports this picture, as does Held, et al. (2005). A mix of internal interdecadal variability and aerosol forcing is a plausible hypothesis for the cause of the changes in interhemispheric gradients in the 20th century that are relevant to the observed Sahel rainfall variations. Quantitative estimates of the relative importance of these two factors must await more definitive estimates of the full aerosol cooling effect.

In Southern Africa as well, changing SSTs rather than changing land use patterns are considered to be the dominant factor controlling warm season rainfall trends. Strong links have been confirmed with Indian Ocean temperatures (Hoerling et al., 2005). Since recent work suggests that land-surface feedbacks may play an important role in governing both intra-seasonal variability and rainy season onset (New et al., 2004; Tadross et al., 2005a; Tadross et al., 2005b; Anyah and Semazzi, 2004), it is plausible that these land-surface feedbacks are important for climate change simulations in Southern Africa, just as in the Sahel.

Changing SSTs can affect African rainfall not only by altering moisture supply, but also by stabilizing the atmosphere to convection by warming the troposphere. ENSO may affect Africa primarily through this mechanism (Chiang and Sobel, 2002) and the increase in days with stable inversion layers over southern Africa (Freiman and Tyson, 2000; Tadross et al., 2005b) in the late-20th century suggests that the same process (possibly linked to increases in Indian ocean SSTs) plays a role in this trend as well. These observations of trends in the atmospheric circulation are consistent with observed increases in daytime temperatures and consecutive dry days (New et al., 2005). They are also consistent with projected changes from several GCMs and may promote the role of the land surface in determining local climates. The stability of the teleconnection processes is, however, uncertain.

There is little doubt that vegetation patterns help shape the climatic zones throughout much of Africa (Zeng and Neelin, 2000; Wang and Eltahir, 2000; Xue et al., 2004; Paeth, 2004, Maynard and Royer, 2004a). Vegetation changes are generally thought of as providing a positive feedback with climate change. The models in the AR4/PCMDI archive do not contain dynamic vegetation models and would likely respond more strongly to large-scale forcing, especially in semi-arid areas, if they did. But given the spread of model predictions in key areas such as the Sahel, it is not clear that adding vegetation models, with the associated additional uncertainties, would add materially to our ability to simulate African climate change at this point in time.

The possibility of the 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 Kyr BP (Foley et al., 2003; Claussen et al., 1999). The implication is that there may be the possibility of abrupt shifts from one climate/vegetation pattern to another, as climate changes.

11.3.2.2 Skill of models in simulating present and past climates

The precipitation generated by the ensemble mean of the 20 models in the PCMDI/AR4 database, averaged over the years 1979–1999 from the 20C3m integrations are displayed in Figure 11.3.2.1 for JJA and DJF. While there is substantial spread among the individual global model simulations, the ensemble mean model is generally of higher quality than any individual model. There are biases that are systematic across the ensemble, a notable example being an overestimate of rainfall in Southern Africa. Of the models in the PCMDI Archive, 90% overestimate the rainfall in this region, on average by over 20% and in some cases by as much as 80% over a wide area extending, in many cases, well into equatorial Africa (an underestimate of Amazon rainfall is just as prevalent in these models; see Section 11.3.x), and it is conceivable that these two deficiencies are related). This bias raises a concern that the sensitivity to drying in Southern Africa could be underestimated, as it is plausible that land surface feedbacks which can accentuate a drying tendency would not act as strongly if the soil is too wet to begin with.

[INSERT FIGURE 11.3.2.1 HERE]

The intertropical convergence zone in the Atlantic is displaced equatorward in nearly all models, and ocean temperatures are too warm by an average of 1–2K in the Gulf of Guinea, and typically by 3K in the intense upwelling region off the southwest coast. Clearly, the oceanic upwelling is too weak in the bulk of the AR4 models. These distortions in the Atlantic make it difficult for many of the models to simulate West African and Sahel rainfall with any precision, but a composite climate averaged over all models produces a credible pattern of rainfall nevertheless. As analyzed by Vizy and Cook (2005), in a few models the summer rains in West Africa fail to move from the Gulf onto land, so there is effectively no West African Monsoon, but most of the models do have a reasonable monsoonal climate. These authors also examine the interannual variability of SSTs in the Gulf of Guinea and the associated dipolar rainfall variations in the Sahel and the Guinean Coast, concluding by their criteria that only 4 of the models of the subset examined produce realistic co-variability of SSTs and rainfall in this region.

Simulated surface temperatures across Africa in the PCMDI archive models are too cold on average, by about 1K, with larger cold biases in drier areas. This cold bias may be reasonably linked to the positive bias in rainfall, and especially the frequency of rainfall events, in the models. Nonetheless, these temperature biases in themselves do not seem large enough to affect the credibility of the model projections, although may indicate a reduced sensitivity to land surface feedbacks in the models due to the wet bias.

Tennant (2003) examines three GCMs from the TAR in terms of their simulation of southern Africa regional inter-annual variability, and notes that the models, at least in terms of synoptic scale processes, represent the major atmospheric processes and related regional climates with credibility, albeit with identifiable systematic biases. The spatial positioning of key large-scale dynamical features is most problematic, especially with respect to the mid-latitude flow, leading to commensurate problems in precipitation fields over the continent.

The multi-model analysis of Hoerling, et al. (2005) using several of the models that contributed to the TAR, provides important evidence that atmospheric/land models 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. This work supplements a large and growing literature (important recent examples are Sutton et al., 2000; Bader and Latif, 2003, 2004; Giannini et al., 2003; Yu and Delworth, 2005) using simulations of this type to study interannual variability; a body of work that is encouraging with regard to the ability of current AGCMs to simulate responses to SST anomalies. However, there is less confidence in the ability of the coupled AOGCMS to generate appropriate interannual variability in the SSTs of the type known to affect African rainfall , as evidenced by the fact that very few of the models in the Archive produce droughts comparable in magnitude to the Sahel drought of the 70's and 80's. There are exceptions, as discussed in Chapter 8, but what distinguished these from the bulk of the models in the Archive is not understood.

Simulations of the mid-Holocene wet period in the Sahara as a response to the changes in insolation over land resulting from changes in the Earth's orbit provide background information on the quality of a model's African monsoon and biome dynamics, but the processes controlling the response to changing insolation may be rather different from those controlling the response to changing SSTs. While these modelling studies cannot be easily used as a means of quality control, the fact that GCMs continue to have difficulty in simulating the full magnitude of the mid-Holocene wet period may indicate a lack of sensitivity to other kinds of forcing. (Jolly et al., 1996; Kutzbach et al., 1997; Braconnot et al., 2000; Kukla and Gavin, 2004)

11.3.2.3 Regional simulation skill

Climate simulations, using dynamical models with a specific focus on southern Africa, are limited, and only in recent years has this issue begun to be more rigorously evaluated. As climate change occurs predominantly through relatively small changes in the balance of large-scale dynamics, it is important that GCMs used to drive regional models are as realistic as possible if to be used as tools for future climate projections. In view of the biases noted in 11.3.2.2, this suggests potential problems for embedding RCMs in the GCM fields for the purpose of downscaling from the GCM projections, because the RCM result would be strongly influenced by the position of circulation features on the lateral boundary.

There are few studies assessing how well RCMs simulate African climate. Over the east Africa regions centred on Lake Victoria Anyah and Semazzi (2004), and Song et al. (2004), following earlier work of Indeje (2001), use the RegCM2 RCM to investigate how regional climate dynamics are influenced by lake surface temperatures. However, the model simulation was for a short period, limiting the conclusions that may be drawn about the climate mode performance of the RCM. On a broader scale, Engelbrecht et al. (2002) evaluate the DARLAM RCM over southern Africa in perpetual January and July modes, with a prominent conclusion being that the model simulates excessive precipitation over the east coast escarpment of the central plateau. Arnell et al. (2003) use the HadRM3H RCM, forced by the HadAM3H GCM in a climate change experiment. Evaluation of the RCM control climate showed that the model suffered from excess precipitation over most of the southern region, which raises important questions on the degree to which soil moisture feedbacks may impact the simulated regional climate change signal. Both Hewitson et al. (2004) and Tadross et al. (2005c) evaluated the MM5 RCM under different physics and parameterization options for a domain spanning Africa south of the equator. With appropriate configuration, the MM5 simulated credible regional climates, with the seasonal mean precipitation field within 30% of observed climatology. However, both the frequency and diurnal cycle of rainfall, and hence the hydrological cycle, was dependent on the choice of convective parameterisation. Extending the application of MM5, Tadross et al. (2005b) and New et al. (2004) explore the sensitivity of the model to surface feedback processes (changes in soil moisture and vegetative cover), which suggest a positive feedback that may exacerbate regional climate change, particularly over arid and semi-arid regions. Furthermore, Tadross and Hewitson ( ) demonstrate that uncertainty in characterising the land surface leads to large uncertainties in the simulated surface climate (±2°C). Uncertainty in the simulated precipitation and regional atmosphere is promoted under synoptically forced high pressures, the frequency of which have and will increase under climate change (Tadross et al., 2005b). This suggests an increasing role for land-surface processes in the future climate of the region.

Over West Africa the number of RCM investigations is significantly fewer than their GCM counterpart. As Jenkins et al. (2002) note, this is partly because of the difficulty in setting up modelling facilities within the region. For the large part RCM studies have focused on simulating important processes of the regional climate; these include African Easterly Waves (Druyan et al., 2001), SSTs influences within the Gulf of Guinea (Vizy and Cook, 2002), and the African easterly Jet (Hsieh and Cook, 2005). Vizy and Cook (2002) went on to demonstrate that warm SSTs in the Gulf of Guinea promoted a southward shift of the ITCZ, resulting in positive rainfall anomalies along the coast and a drying over the Sahel. Gallee et al. (2004) use an RCM to simulate the 1992 rainy season, and Hsieh and Cook (2005) demonstrate that African easterly waves are closely linked to convection within the ITCZ. A 25-year simulation was undertaken by Paeth et al. (2005), which highlighted most of the above large-scale controls of West African climate, and found that the RCM (REMO) simulated the regional climate well, with the regional sea surface temperatures to be most important in forcing the regional climate. In addition, it was noted that change in land cover is directly linked to a local anomalies of the hydrological cycle.

Given the potential importance of orography (Semazzi and Sun, 1997) and land-surface-atmosphere interactions (Wang and Eltahir, 2000) on the modelled climate, the potential for RCMs to elucidate the controls of the regional climate is high. Vizy and Cook (2002), following earlier by Jenkins (1997) found that in simulating the region the RCMs are sensitive to tuning and parameterisation choices. As for southern Africa these studies suggest that RCMs need to be carefully evaluated for the West African domain before being used to investigate the local climate.

Empirical downscaling has been applied over the southern Africa region for a number of different applications. For example, Landman (2000; 2001; 2002) used empirical techniques to enhance seasonal forecasting products. For longer simulation periods Hewitson (1996) assessed the sensitivity to the assumptions underlying the method, and demonstrated that with appropriate predictor selection a robust downscaling of contemporary climate can be derived. Building on this work 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 NCEP reanalysis data results provide a close match to the historical climate record, especially over the eastern escarpment of the sub-continent – a problematic region for RCMs.

Both RCM and empirical downscaling approaches show valuable skill; each having relative strengths and weaknesses. For Africa the RCM downscaling lags the work with empirical downscaling. At present it is difficult to assess the degree of convergence in RCM-based projections from multiple GCMs, although empirical downscaling from multiple GCMs shows notable convergence (Hewitson and Crane, 2005).

11.3.2.4 Climate projections

11.3.2.4.1 Mean temperature

Global models predict a relatively uniform warming over the continent. Figure 11.3.2.2 shows the mean difference across all AR4/PCMDI models in annual mean near surface air temperature between years 2079–2099 in the A1B scenario and the years 1979–1999 in the 20C3M 20th century simulations. In most regions this ensemble mean response is between 3 and 4K, with smaller values in equatorial and coastal areas and larger values in the Western Sahara. The global mean of this response in near-surface air temperature is 2.5K, so the temperature response is about 50% larger on average than the global mean response. Indeed, every model in the archive predicts temperature responses larger than its own global mean response in Northern and Southern Africa, while 13–17 of the models predict larger than global mean responses in different areas within Western and Eastern Africa. Similar results hold for other scenarios. The largest temperature responses in North Africa are projected to occur in June-July-Aug, while the largest responses in Southern Africa occur in Sept-Oct-Nov. But the seasonal structure in the temperature response over Africa is modest as compared to many other regions. The pattern of warming is very similar to that described by Hulme et al. (2001) for a composite of models used in the TAR. The observed rate of warming over the African continent has been estimated as being comparable to the global mean warming (Hulme et al., 2001). See Table 11.3.2.1 for the more information on the range of temperature responses among the different models, which is typically a factor of 2–2.5.

[INSERT FIGURE 11.3.2.2 HERE]

There is a strong correlation across the AR4/PCMDI models between the global mean temperature response and the response in Africa. For example, regressing the SAH annual mean temperature response in A1B against the global mean temperature response, one finds that the latter explains 61% of the variance in SAH. Thus, a significant fraction of the spread in the temperature response among models has little to do with local African processes, but rather with the sum total of the global feedbacks that control (transient) climate sensitivity. This conclusion is also consistent with the observed rate of warming over the African continent, which is comparable to the global mean warming (Hulme et al., 2001).

Inspection of the AR4 Archive shows that one can predict rather well the ensemble mean temperature response in other time periods, and for the A2 and B1 scenarios, from these temperature responses for A1B in the 20802100 time frame, by rescaling linearly according to the ensemble mean global mean responses. For example, in SAH the ensemble mean annual mean temperature responses in the scenarios (B1, A1B, A2) in the 20792099 time frame are in the ratio (0.68, 1.0, 1.22) as compared to the corresponding values for the global mean responses of (0.69, 1.0, 1.17).

Over southern Africa, Tadross et al ( ) used two RCMs forced by the same GCM (HadAM3H, SRES A2), and project average temperature changes in excess of 1°C with highest temperature changes in excess of 4°C during OND, at the lower end of the spread of AR4 models. During this period some of the regions experiencing the highest temperature increases showed commensurate decreases in precipitation, suggesting that some of the increases in temperature are associated with either a reduction in latent cooling or increase in incident shortwave radiation (due to decreased cloud cover) at the surface.

11.3.2.4.2 Mean precipitation

Figure 11.3.2.3 illustrates some of the robust aspects of the precipitation response over Africa in the AR4/PCMDI models. The upper panels show the % change in precipitation averaged over the ensemble of models, between years 2079–2099 of the A1B scenario and the years 1979–1990 of the 20C3M historical integrations, for DJF, JJA and the annual mean)The lower panels show the number of models (out of 20) that predict moistening at a particular location.

[INSERT FIGURE 11.3.2.3 HERE]

The corresponding plots for the A1 and B2 scenarios are very similar once rescaled by the global mean temperature response. The ensemble mean responses also vary smoothly in time. 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 a qualitative agreement with the results in Hulme (2001) and Ruosteenoa et al. (2003) summarizing results from the TAR models.) A tendency towards moistening on the Guinean coast evident in these TAR summaries does not appear as clearly in the ensemble mean of the AR4 archive, although it is present in individual models.

The large-scale picture is one of drying in the subtropics and an increase (or little change) in rainfall in the tropics, increasing the rainfall gradients. This is an anticipated and fundamental aspect of the 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 11.3.2.3.1).

The drying along Africa's Mediterranean coast is a component of a larger scale drying pattern surrounding the Mediterranean on all sides, 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. The sign is consistent throughout the year and is generated by nearly every model in the archive. The drying signal in this composite extends into the Northern Sahara, and along 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.

In Southern Africa a roughly analogous set of processes produces drying as well. This drying is especially robust and severe in the extreme southwest in austral winter, which is a manifestation of a much broader scale poleward shift in the storm tracks across the South Atlantic and Indian oceans. The very robust drying in percentage terms in JJA 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 (of the order of 5–10% throughout the subcontinent) according to this global model consensus, occurs in the spring (Sept-Oct-Nov) and is mirrored in some RCM simulations for this region (see below), and suggests a delay in the onset of the rainy season. This springtime drying contributes to the springtime maximum in the temperature response in this region mentioned above, as evaporation is suppressed.

The increase in rainfall in East Africa, extending into the Horn of Africa is also robust across the ensemble of models, with 18 of 20 models projecting an increase in rainfall in the core of this region, east of the Great Lakes. An increase in tropical rains is a conservative expectation, assuming little or no increase in the circulation, based on an increase in atmospheric water vapour and an increased convergence of vapour into pre-existent convergent regions. This East African increase was also evident in the TAR models. What is more difficult to explain is the lack of a clear increase in the Guinean coastal rain belts and in the Sahel. (The increase at 20°N in the East Sahara is generated a large response in a few models and is not robust across the model ensemble.) A straight average across the ensemble results in modest moistening in the Sahel and with little change on the Guinean coast. But individual models generate large, but disparate, responses in this region. GFDL/CM2.1 projects strong drying in the Sahel and throughout the Sahara. MIROC3.2 (medres) model shows a strong trend with the opposite sign. These two models are near the extremes of the ensemble of responses, but they are both among the four models that Vizy and Cook (2005) find generate realistic interannual variability in the Gulf of Guinea and Sahel, and their climatologies are similar. While the drying the GFDL model is extreme within the ensemble, its 20th century simulation is not inconsistent with observations (Held et al., 2005). As one moves northwards in the Sahara, one eventually enters the latitudes to which the Mediterranean drying penetrates robustly (see Figure 11.3.2.3). In models that dry the Sahel, the entire Sahara typically dries; in others, the moistening in the Sahel transitions into the Mediterranean drying at a latitude that varies considerably from model to model.

Inspection of the AR4 Archive indicates that summer sea level pressure in projected to be reduced in the Sahara in nearly all models. Haarsma et al. (2005), argue that the moistening of the Sahel and Sahara generated in their model is a consequence of this pressure drop. That the precipitation response is far less robust than the pressure response across the AR4 models suggests a more complex picture. Maynard et al. (2002) provide a very detailed analysis of the changes in the hydrological cycle a model that projects a significant moistening in the Sahel as the climate warms, but it is difficult in analyses of tropical climates to move beyond statements of consistency towards causal mechanisms. Progress is being made in developing new methodologies for this purpose (e.g., Chou and Neelin, 2004; Lintner and Chiang, 2005) but these have not yet fully matured.

It has been argued (e.g., Paethe and Hense, 2004) that the amelioration of the Sahel drought since the 80's may be a sign of the greenhouse-gas driven increase in rainfall, providing support for those models that moisten the Sahel into the 21st century. Our view is that it is premature to take this partial amelioration as evidence of a global warming signature, and that it at least equally plausible to consider an explanation based on inter-decadal variability in inter-hemispheric SST gradients.

In any downscaling from GCM fields, land use changes presents a possibly important feedback process not captured in the global model, and cannot be ignored as a potential contributor to drying in the 21st century. However, there is general agreement that it is not the dominant factor to be considered. Taylor et al. (2002)

estimate drying over the Sahel of 4% between 2015 and 1996, but do suggest that the magnitude could grow substantially further into the next century. Maynard and Royer (2004a) indicate 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. While a variety of lines of research, including mid-Holocene modelling, supports the intuition that interactive vegetation is important in this region, the spread in prediction by models with prescribed vegetation will have to be better understood before we can learn very much from the more complex interactive-vegetation models.

Regional climate change projections based on RCM simulations are limited for the southern Africa region and even scarcer in other regions. Tadross et al. ( ) examine two RCMs (PRECIS and MM5) nested for Southern Africa in the HadAM3H (SRES A2) GCM.. The projected change from the two RCMs differed between the early summer season (Oct-Dec, OND) and late summer season (Jan-Mar, JFM). During OND both models predict drying over the tropical western side of the continent with MM5 indicating that the drying extends further south and PRECIS further east. Again there is an indication of drying in the west towards the tropics during JFM but with increases in total rainfall towards the east. These increases cover a larger statistically significant area in the PRECIS data but are of greater magnitude in the MM5 data. Examination of the monthly data indicated that these increases in rainfall in the east were confined to January and February in both models.

Generally the change in total rainfall reflected changes in the number of rain days. This reflects projected increases in the frequency of high pressures towards the west in the GCM forcing data indicating that the lateral boundaries in this region dominate the response of both models. This also serves to highlight that stronger responses may be detected in statistics of daily precipitation, as in this case it appears that total rainfall changes little, perhaps due to increases in intensity, which act in an opposite manner to the reduction in rain days.

Arnell et al. (2003) make use of the HadRM3H RCM with a macro-scale runoff model to explore the effects of Southern African climate change. Boundary fields for the RCM were from the HadAM3H GCM forced with sea surface temperatures from the HadCM3 coupled ocean-atmosphere GCM. Using 16 different ways of constructing scenarios from the model simulation output, they noted a positive runoff change of between 10% and 20%, with the regional model showing a clear difference in the large-scale runoff pattern in comparison to the GCM. While this suggests the RCM has added value, there remains notable uncertainty in light of the significant precipitation bias in the model.

Projections based on empirical downscaling have been developed over a number of years, and for Africa are more widely available than from RCM based approaches. Building on earlier work, Hewitson and Crane (2005) provide projections for daily precipitation as a function of 6 GCM (3 from the TAR, 3 from the AR4/PCMDI archive) simulations of climate change. The empirical method explicitly depends on the synoptic scale atmospheric features, is inherently conservative, and as such likely to under-estimate rather than over-estimate the climate change signal. By using the more robust circulation fields of the GCM the downscaling bypasses the native parameterized precipitation of the GCM. The downscaling of the GCM control climates (30 year period) show some small wet bias but captures the spatial detail of the regional precipitation gradients well. The downscaled results for the GCM output between the control period and the 2070–2099 (TAR models) period, and the 2080–2099 period (AR4/PCMDI archive models), for the SRES A2 emissions scenario, show notable convergence in the projected change with fine spatial detail. The convergence in the projected anomaly pattern at this resolution suggests that the GCMs have significant commonality in the projected changes of daily synoptic circulation, on which the downscaling is based. Figure 11.3.2.4 shows the Africa climate change anomaly of mean June-July-August monthly total precipitation (aggregated from the downscaled daily data) for station locations across Africa. The downscaled results largely agree between the 6 GCMs as to the broad spatial detail of the pattern change, although showing some difference in magnitude. Most notably of the seasonally dependant consensus changes are:

- the increased precipitation in east Africa and extending into southern Africa, especially in June-August,

- strong drying in the core Sahel in June-July-August with some coastal wetting, and moderate wetting in December-February,
- most downscaled models showing drying to the west in southern Africa, and on the Mediterranean coast.

[INSERT FIGURE 11.3.2.4 HERE]

However, 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 boarders of the Sahel.

As noted by Tadross et al. ( ), some of the more relevant changes are found in the statistics of the daily rainfall, and the empirical downscaling show a moderate increase in heavy rainfall events for much of the southern rainfall region, and changes in the median precipitation event magnitude that, at the station scale, does not always mirror the projected changes in seasonal totals. Qualitatively, the downscaled anomalies are consistent with the native GCM fields at GCM resolutions.

There is a consistent tendency for greater Sahel drying than in the underlying GCM in these empirical downscaling results, providing further rationale (alongside the large AR4 global model spread and poor coupled model performance in simulating droughts of the magnitude observed in the 20th century) for resisting a projection of ameliorating conditions in the Sahel in the 21st century common to much of the recent literature

11.3.2.5 Uncertainties

- Systematic errors across the ensemble of global models (excessive rainfall in the south, southward displacement of Atlantic ITCZ, insufficient upwelling of the West Coast) emphasize that the robustness of the large-scale response is only a necessary but not a sufficient condition for its reliability.
- The potential significance of land surface feedbacks and the accurate characterisation of the land surface, especially in semi-arid regions, adds a layer of uncertainty to the climate projections for these areas. Vegetation feedbacks and feedbacks from dust aerosol production are not included in any of the global models.
- RCMs are still being developed for different African regions; experience as to the extent they can successfully downscale precipitation is limited. The intensity with which they simulate the local hydrological cycle may affect their ability to respond accurately to changes in regional forcings (e.g., synoptic, land surface, SST).
- Empirical downscaling, while subject to assumptions of predictor choice and issues of stationarity, does appear to reach relatively robust results and indicate a convergence when trained with different GCMs. However, empirical techniques cannot capture changes in local feedback mechanisms.
- Absence of realistic variability in Sahel in most 20th century simulations casts some doubt on reliability of coupled models in this region.
- There is insufficient information on which to assess possible changes in tropical storm distribution

11.3.3 Europe and the Mediterranean

11.3.3.1 Key processes

In addition to global warming and its direct thermodynamic consequences, such as the ability of a warmer atmosphere to transport more water vapour from low to high latitudes (e.g., Manabe and Wetherald, 1987), future climate changes in Europe and the Mediterranean area may be affected by several other factors. Variations in the atmospheric circulation induce substantial variations in 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 (e.g., Fink et al., 2004), and the strong warming of winters in northern Europe from the 1960’s to 1990’s, attributed mainly to a shift towards the positive phase of the NAO (e.g., Hurrell and van Loon, 1997; Räisänen and Alexandersson, 2003).

Although the NAO has the highest influence upon the northwestern winter European climate (e.g., Busuioc et al., 2001b; Wilby et al., 2002; Hurell et al., 2003; Uvo, 2003; Haylock and Goodess, 2004), it is also responsible for the interdecadal variability of the Mediterranean precipitation (Quadrelli et al., 2001; Goodess and Jones, 2002; Xoplaki et al., 2004) and southeastern European climate (e.g., Bojariu and Paliu, 2001) and controls the snow cover and surface-atmosphere temperature feedback in the alpine region (Beniston, 2005). Additional processes such as Mediterranean cyclogenesis (discussed below), Euro-Atlantic blocking (e.g., Quadrelli et al., 2001; Xoplaki et al., 2003b; Valero et al., 2004; Tomozeiu et al., 2005), the Eastern Atlantic/ Scandinavian patterns (e.g., Quadrelli et al., 2001; Domonkos et al., 2003; Tomozeiu et al., 2002) and North Atlantic/Mediterranean SST patterns (Wilby et al., 2002; Benestad and Melsom, 2002; Xoplaki et al., 2003a) also play important roles in the European climate variability. On fine geographic scales, the effects of atmospheric circulation are modified by topography particularly in mountainous areas (Bojariu and Giorgi, 2005).

For the southern part of the area, Mediterranean cyclogenesis and Mediterranean subsynoptic cyclones strongly influence the local climate and particularly precipitation (Alpert et al., 1990; Trigo et al., 2000). Most of the floods both in the Northern Mediterranean and in the Middle East are associated with these lows. Mediterranean cyclogenesis is mainly due to the phasing of high level vorticity anomalies and low level orography and thermal forcings (Alpert et al., 1990; Trigo et al., 2002). Trigo et al. (2000) also show a correlation between the decline in Mediterranean rainfall and the weakening of Mediterranean cyclones over the last decades. The Mediterranean area is also one of the areas in the world where an increase in extreme daily rainfall has been observed in spite of a decrease in total precipitation (Alpert et al., 2002).

Europe, particularly its northwestern parts, owes some portion of its relatively mild winter climate to the northward heat transport by the North Atlantic Thermohaline Circulation (THC) (e.g., Vellinga and Wood, 2002). If increased greenhouse gas concentrations lead to a weakening of the THC, as suggested by most AOGCMs (Chapter 10), this will act to reduce the warming in Europe. However, models do not support a reversal of the warming to cooling (Section 11.3.3.3.1; Chapter 10).

Local thermodynamic factors also affect the European climate and are potentially important for its future changes. In the northern and eastern parts of the continent that are at present snow-covered in winter, reductions of snow are likely to induce a positive feedback, further amplifying the warming. The decrease in snow cover may have a particularly large impact on the lowest winter temperatures (Section 11.3.3.3.2). In the Mediterranean region and occasionally in central Europe, feedbacks associated with the drying of the soil in summer are important even in the present climate. For example, they appeared to exacerbate the heat wave of 2003 (Fink et al., 2004).

11.3.3.2 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 AR4 models vary on both sides of the observational estimates in summer but are mostly lower than observed in the winter half-year, particularly in NEU (Table 11.3.3.1). Excluding one model with extremely cold winters in northern Europe, the seasonal area mean temperature biases in NEU vary from –6°C to 3°C, and those in SEU from –5°C to 4°C, depending on model and season. The biases vary geographically within both regions. In particular, 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.

There is a wide range of geographic variation and model-to-model variation in the precipitation biases within Europe and the Mediterranean area. The, average simulated precipitation in NEU exceeds the observational estimate from autumn to spring (Table 11.3.3.1), but the interpretation of the difference is complicated by the observational uncertainty associated with the undercatch of, in particular, solid precipitation (e.g., Legates and Willmott, 1990; Rubel and Hantel, 2001). In summer, most models simulate too little precipitation, particularly in the eastern parts of the area. In SEU, the area and ensemble mean precipitation is close to observations.

The distribution of time-mean sea-level pressure over Europe and surrounding areas is simulated realistically in many but not all of the current AOGCMs (e.g., van Ulden and van Oldenborgh, 2005). However, most models simulate too high pressure over the European sector of the Arctic Ocean and too low pressure in the latitude band 50°-55°N, particularly in winter and spring. As regards the origin of the temperature and precipitation biases, the biases in the pressure distribution and the resulting biases in the near-surface atmospheric flow may be equally important as other sources of error (van Ulden and van Oldenborgh, 2005).

Notwithstanding their dependence on the boundary data used, RCMs capture the geographical variation of temperature and precipitation in Europe more realistically than global models. However, RCMs 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., Räisänen et al., 2003; Jacob et al., 2005; Figure 11.3.3.1). Most but not all RCMs also overpredict the interannual variability of summer temperatures in central and southern Europe (Lenderink et al., 2005; Vidale et al., 2005; Jacob et al., 2005). Lenderink et al. (2005) show that, depending on the RCM, the overestimate in temperature variability is forced by excessive interannual variability in either shortwave radiation or evaporation, or both. A need for improvement in the modelling of soil, boundary layer and cloud processes is implied.

[INSERT FIGURE 11.3.3.1 HERE]

The ability of RCMs to simulate climate extremes in Europe has been addressed in several studies. In the PRUDENCE simulations (see Box 11.2), the biases in the tails of the temperature distribution were generally larger than the biases in average temperatures (Kjellström et al., 2005). Most models underestimated the 95th percentile of summer maximum temperatures in Scandinavia and in the British Isles, but overestimated it in eastern Europe. The 5th percentile of winter minimum temperatures was generally too high in western, central and northern parts of Europe, but too low in eastern Europe. However, these biases varied substantially between the RCMs, not only in magnitude but in most areas also in sign.

Frei et al. (2005) compared extremes of daily precipitation in the vicinity of the European Alps between observations and seven RCMs driven by boundary data from Hadley Centre global models. The 5-year return values of maximum one-day precipitation varied by up to a factor of two among the RCMs, differing frequently by several tens of percent from the observed values. Nevertheless, the biases in the extremes were not larger than those in mean precipitation and average wet-day precipitation intensity. Moreover, except for generally too low extremes in the southern parts of the Alpine area in summer, the set of models as a whole showed no systematic tendency to over- or underestimate the magnitude of the extremes. The models also showed skill in simulating the mesoscale patterns of extreme precipitation associated with the complicated topography of the Alpine area. Buonomo et al. (2005) show similar results for two RCMs compared with high resolution observations over the UK for one and 30-day precipitation extremes in the range of 2 to 20 year return periods. Other model verification studies made for European regions (e.g., Booji, 2002; Semmler and Jacob, 2004; Fowler et al., 2005, see also Frei et al. 2003) support these findings.

Weisse et al. (2005) found the REMO 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 (17.2 m/s or more) over land and coastal areas (Rockel and Woth, 2005). Although this might also be affected by the boundary data set used, the main explanation appears to be the lack of gust parameterizations which would be needed to mimic the large local and temporal variability of near-surface winds over land. Realistic frequencies of high wind speeds were only found in those two PRUDENCE RCMs that applied a gust parameterization.

[START OF BOX 11.2]

Box 11.2: The PRUDENCE Project

The 'Prediction of Regional scenarios and Uncertainties for Defining European Climate change risks and Effects – PRUDENCE' project involved over twenty European research groups. The main objectives of the project were to provide high resolution climate change scenarios for Europe at the end of the 21st century using dynamical downscaling methods with regional climate models, and to explore the sources of 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) *Radiative uncertainty* due to choice of IPCC-SRES emission scenario, and (iv) *Boundary uncertainty* due to the fact that the regional models have been run with boundary conditions from different global climate models. A large fraction of the PRUDENCE simulations (Box 11.2, Table 1) used the same boundary data (from HadAM3H for the A2 scenario) to provide a detailed understanding of the regional model uncertainty; the other uncertainties were covered in a less complete manner.

Each PRUDENCE experiment consisted of a control simulation representing the period 1961–1990 and a future scenario simulation representing 2071–2100. Box 11.2, Figure 1 illustrates the geographical region that was investigated within the project. More details are provided in e.g., Christensen et al. (2005), Déqué et al. (2005a) and http://prudence.dmi.dk.

Box 11.2, Table 1: A summary of the PRUDENCE simulations. “1” indicates that one experiment was conducted for a given GCM / emissions scenario / RCM combination, and “3” that an ensemble of three experiments with varying GCM initial values were made to study sampling uncertainty.

GCM boundaries RCM	No.1	No.2	No.3	No.4	No.5	No.6	No.7	No.8	No.9	No.10
HadAM3H +A2[a]		3	3	1	1	1	1	1	1	1
HadAM3H +B2		1		1	1	1				
ECHAM4 +A2			1	1						
ECHAM4 +B2			1	1						
ARPEGE +A2[a]	1									
ARPEGE +B2	3									

Notes:
(a) Using the same sea surface temperatures based on HadCM3 AOGCM simulations.

[END OF BOX 11.2]

Lionello et al. (2002) showed that the ECHAM4 model at T106 resolution simulates well Mediterranean cyclone characteristics such as the main cyclogenesis areas and the cyclone track number density. Vérant (2004) and Somot (2005) found that the stretched version of ARPEGE-Climate (resolution of 50 km over the Mediterranean region) also reproduces the main characteristics (cyclogenesis area, track number, seasonal cycle, interannual variability, life-time, velocity) of Mediterranean cyclones in spite of a somewhat too strong cyclone intensity.

11.3.3.3 Climate projections

11.3.3.3.1 Mean temperature

The area and annual mean warming from 1979–1998 to 2079–2098 in the AR4 SRES A1B simulations varies from 2.3 to 5.2°C in NEU and from 2.0 to 5.0°C in SEU, with an ensemble mean of 3.6°C (40% above the global ensemble mean warming) in NEU and 3.4°C (30% above the global mean) in SEU. Ensemble mean temperature changes for other periods and emissions scenarios scale approximately linearly with the global mean warming (e.g., Jylhä et al., 2004). In northern Europe, particularly in its northeastern parts, the warming is likely to be largest in winter, in the Mediterranean area in summer (Table 11.3.3.2.; Figure 11.3.3.2). The uncertainty ranges for local changes are wider than those for the subcontinental means.

For the A1B scenario in the years 2079–2098, the inter-model correlation between the global warming and the annual warming in NEU (SEU) is 0.8 (0.9). Thus, models with large (small) global warming also tend to simulate large (small) warming in Europe.

[INSERT FIGURE 11.3.3.2 HERE]

In addition to the overall global warming, changes in atmospheric circulation also have the potential to affect temperature changes in Europe. Van Ulden and van Oldenborgh (2005) estimated the contribution of circulation changes for a western part of central Europe, using a regression method and seven AOGCM simulations for the SRES A2 scenario. In most models, circulation changes enhanced the warming from 1971–2000 to 2071–21000 in winter (due to an increase in westerly flow) and late summer (due to a decrease in westerly flow), but they reduced the warming slightly in May and June. The magnitude of the circulation contribution typically ranged from –1°C to 1.5°C, but slightly larger values were found for some individual models in some months. The residual warming, unexplained by changes in circulation, was 1–5°C depending on model and season. Other studies (Rauthe and Paeth, 2004; van Ulden et al, 2005) also support the idea that circulation changes may have a significant, but not generally dominating, impact on future long-term temperature changes in Europe. Besides the circulation changes associated with anthropogenic forcing, natural variations of the circulation may cause pronounced interdecadal temperature variations even in the future (e.g., Dorn et al., 2003).

Most AOGCMs simulate a decrease in the North Atlantic Thermohaline Circulation (THC) with increasing greenhouse gas concentrations (Chapter 10). In spite of this, nearly all reported AOGCM greenhouse gas simulations indicate warming in all of Europe, as the direct atmospheric effects of increased greenhouse gases, the positive feedbacks associated with the warming and the tendency for the land to warm faster than the oceans dominate over the changes in ocean circulation. Rarely, slight cooling has been simulated along the northwestern or northern coastlines of Europe (Russell and Rind, 1999; Schaeffer et al., 2004), but even in these simulations most of Europe has experienced warming. Schaeffer et al. (2004) point out that the impact of THC changes on the atmosphere depends on the regional details of the THC change, being largest if ocean convection is suppressed in high latitudes where the sea-ice feedback may amplify atmospheric cooling. AOGCM sensitivity studies with an artificial shutdown of the THC, with no changes in greenhouse gas concentrations, indicate a 1–3°C annual mean cooling in Europe, with the largest effect in the northwestern parts of the continent in winter (e.g., Manabe and Stouffer, 1997; Vellinga and Wood, 2002).

In PRUDENCE, different RCMs simulated different temperature changes even when driven by the same GCM. In summer, these differences amounted up to about 3°C in eastern Europe. Nevertheless, the differences between the RCMs were generally smaller than the differences in warming between various GCMs (Déqué et al., 2005b; Ruosteenoja et al., 2005; Section 11.2.2.2.5). Fronzek and Carter (2005) and Jacob et al. (2005) found the HadAM3H-driven PRUDENCE RCMs to simulate generally smaller warming than HadAM3H, but it is not known if this would also hold for other driving GCMs.

More detailed local projections of temperature change have been derived by using various statistical downscaling models (SDMs). SDMs have been applied to several AOGCMs including the IPCC AR4 model ensembles, especially for northern Europe (e.g., Benestad, 2002a, 2002b, 2004; Hanssen-Bauer et al., 2003, 2005). While showing a similar large-scale signal as dynamical models, SDMs have added some regional detail to the projections that are not captured even by RCMs. For example, Hanssen-Bauer et al. (2005) found that, in most of Scandinavia, the projected warming rates during the 21st century increased with distance from the coast and with latitude. Hanssen-Bauer et al. (2003), comparing dynamical and empirical downscaled changes from the ECHAM4/OPYC3 global model found that the differences between the two approaches were largest during winter and/or spring at localities exposed to temperature inversions. It was argued that less favourable conditions for ground inversions are consistent with the future projection of increased winter wind speed in ECHAM4/OPYC3 and reduced snow cover. For other European regions, a similar signal of warming was also identified with some regional differences (Huth, 2003).

11.3.3.3.2 Temperature variability and extremes

Several studies have indicated increased temperature variability in Europe in summer, both on interannual and daily time scales. However, the magnitude of the increase is model-dependent. In some of the PRUDENCE RCM simulations, the interannual summertime temperature variability in central Europe doubled from 19611990 to 2071–2100 under the A2 scenario, while others showed almost no change (Schär et al., 2004; Vidale et al., 2005). A weaker tendency to increased variability was found in Scandinavia and in Mediterranean Europe. Lenderink et al. (2005) related the increase in variability to reduced soil moisture, which reduces the capability of evaporation to damp temperature variations, and to increased land-sea contrast in average summer temperature. In qualitative agreement with these RCM results, Giorgi and Bi (2005) found the interannual standard deviation of JJA mean temperature to increase in both northern Europe and the Mediterranean region in most of 18 recent AOGCM simulations. However, the average increase for the A2 scenario was only about 20% in 100 years. Schär et al. (2004) speculated that increased variability may have played a role in producing the European heatwave in summer 2003, but Stott et al. (2005) found no support for this conlusion in their model.

Kjellström et al. (2005) analysed the variability of daily maximum and minimum temperatures in the PRUDENCE simulations in several European regions. As was the case with the present-day biases, the intermodel differences in simulated change from 1961–1990 to 2071–2100 also increased towards the extreme ends of the temperature distribution. However, a common signal of increased summertime variability was evident especially in southern and central Europe, with the highest maximum temperatures increasing more than the median daily maximum temperature (Figure 11.3.3.3). Increased summertime temperature variability was also found in midlatitude western Russia in the RCM simulations of Shkolnik et al. (2005). Similar results were obtained in GCM studies by Gregory and Mitchell (1995), Zwiers and Kharin (1998) and Hegerl et al. (2004), and by Meehl and Tebaldi (2004) who found the simulated intensity of central European heat waves to increase more than explained by changes in average conditions alone.

[INSERT FIGURE 11.3.3.3 HERE]

In contrast with summer, models indicate reduced temperature variability in most of Europe in winter, both on interannual (Räisänen, 2001; Räisänen et al., 2003; Giorgi and Bi, 2005) and daily time scales (Zwiers and Kharin, 1998; Hegerl et al., 2004; Kjellström et al., 2005). 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 change in them was more strongly model-dependent than that in the median (Figure 11.3.3.3). The largest warming of the cold extremes occurred in those areas that had a substantial simulated snow cover (mean DJF snow fraction over 50%) in the years 1961–1990 but much less (under 25%) snow in the years 2071–2100 (Kjellström et al., 2005). In another study, Vavrus et al. (2005) analysed simulated changes in cold-air outbreaks, defined as at least two consecutive days with temperature two standard deviations below the local present-day winter mean. The seven AOGCMs in this study indicated a large decrease (generally 80–100%) in cold-air outbreaks in northern Europe by the end of the 21st century, but some of them indicated substantially smaller decreases in southern Europe.

Along with the increase in average temperatures, the annual number of frost days is very likely to decrease. In the PRUDENCE simulations under the A2 forcing scenario, the largest absolute decreases of about 60 days per year occurred in northern and eastern Europe and in the Alps (Jylhä et al., 2005), whereas larger relative decreases occurred in southern and western Europe. For the B2 scenario, the decreases were smaller. Jylhä et al. (2005) also found a general decrease in the number of days with temperature intersecting 0°C in the PRUDENCE simulations. However, the change in northernmost Europe was seasonally variable, with fewer such days in autumn and spring but more of them in winter.

11.3.3.3.3 Mean precipitation

AOGCMs indicate a south-north contrast in precipitation changes across Europe, with increases in the north and decreases in the south (Figure 11.3.3.2). The annual area mean change from 1979–1998 to 2079–2098 in the AR4 A1B simulations varies from 0 to 17% in NEU and from –3% to –26% in SEU (Table 11.3.3.2). The largest increases in northern Europe are simulated in winter, when models also tend to simulate increases in central Europe. In summer, the sign of the NEU area mean change varies between models, although most models simulate increased (decreased) precipitation north (south) of about 55°N. In SEU, the largest per cent decreases are generally simulated in summer, but the area mean winter precipitation also decreases in most models.

Changes in precipitation may vary substantially on relatively small horizontal scales, particularly in areas of complex physiography. However, the details of this variation are very sensitive to changes in the atmospheric circulation, as illustrated in Figure 11.3.3.4 by a comparison of two PRUDENCE simulations that only differ with respect to the driving global model. In the ECHAM4/OPYC3-driven simulation, an increase in westerly flow from the Atlantic Ocean (caused by a substantial increase in the north-south pressure gradient) leads to a 60–70% increase in annual precipitation at the western flank of the Scandinavian mountains. In the HadAM3H-driven simulation, with little change in the annual pressure pattern, the increase is only 0–10%. The different changes in circulation also have a larger-scale signature, with a larger contrast between increasing precipitation in northern Europe and decreasing precipitation in southern Europe in the top than the bottom row experiment. Räisänen et al. (2004) attribute this to a northward shift in cyclone activity present in ECHAM4/OPYC3 but not in HADAM3H.

[INSERT FIGURE 11.3.3.4 HERE]

The importance of circulation changes for precipitation was also demonstrated by van Ulden and van Oldenborgh (2005). They found that, in the western parts of central Europe, simulated increases in winter precipitation were in most models enhanced by increased westerly winds, whereas the general decrease in summer precipitation was largely explained by a more easterly and anticyclonic flow type (Figure 11.3.3.5). The residual precipitation change that was unexplained by changes in circulation varied much less with season, and (with the exception of summer) between the seven AOGCMs in their study, than the actual precipitation change. For most months and models, the residual change from 1971–2000 to 2071–2100 was a modest increase (0–15%). This is consistent with the idea that, under unchanged atmospheric circulation, the increased absolute humidity of a warmer atmosphere should increase the moisture transport from oceans to continents and from low to higher latitudes (e.g., Manabe and Wetherald, 1987)

[INSERT FIGURE 11.3.3.5 HERE]

The causes of reduced simulated summer precipitation in southern and central Europe were also studied by Rowell and Jones (2005), who made a series of experiments with a regional version of the HadAM3P atmospheric model to isolate the mechanisms that led to reduced precipitation in the global version of the same model. Although they found changes in the large-scale atmospheric circulation to be important in Great Britain and southern Scandinavia, other factors were dominant in continental and southeastern Europe. These included reduced relative humidity resulting from larger warming over the European continent than over the surrounding sea areas, and reduced soil moisture, affected by both earlier snowmelt and by a feedback from reduced summer precipitation. Their study also indicated that reduced soil moisture enhanced the simulated summertime warming in central and southeastern Europe by several tens of percent. Based on their results and the fact that changes in large-scale atmospheric circulation remain a relatively uncertain aspect of model results, they had higher confidence in reduced summer precipitation in continental and southeastern Europe than in Great Britain and southern Scandinavia.

In RCM simulations, changes in precipitation are less strongly governed by the driving global model than the changes in temperature. Differences in precipitation change between different RCMs, when driven by the same GCM, may be comparable to the differences between various GCMs, particularly in summer and autumn (Dequé et al., 2005b; Ruosteenoja et al., 2005; Section 11.2.2.2.5).

Various SDMs to obtain detailed information about precipitation have been developed, mostly having in mind the limited performance of AOGCMs and RCMs in simulating local precipitation especially in areas of very complex topography. Although the climate change signal derived through these techniques is dependent on the method and large-scale predictor used for their calibration (see Section 11.2.1.4), some results are generally in agreement with those obtained from AOGCMs and RCMs giving them more robustness, for example precipitation increase over almost the whole year in northern Europe (Busuioc et al., 2001a; Beckmann and Buishand, 2002; Benestad, 2002b; Hanssen-Bauer et al., 2003, 2005) and increase of winter precipitation over northwestern Romania. Other case studies showed more or less agreement. For example, Diaz-Nieto and Wilby (2005), using the GCM outputs from UKCIP02 A2 and B2 scenario simulations found precipitation increase in winter and decrease in summer over the Thames area for the periods centered on the 2020s, 2050s and 2080s. Trigo and Palutikof (2001) revealed an increase of Iberian precipitation in winter and small decreases in spring and autumn for the period 2041–2090 against 1941–1990, using the SLP predictor simulated by HadCM2.

11.3.3.3.4 Precipitation variability and extremes

In northern Europe and in central Europe in winter, where time mean precipitation is simulated to increase, both GCMs (e.g., Semenov and Bengtsson, 2002; Voss et al., 2002; Hegerl et al., 2004; Wehner, 2004; Tebaldi et al., 2005) and RCMs (e.g., Jones and Reid, 2001; Räisänen and Joelsson, 2001; Booji, 2002; Huntingford et al., 2003; Christensen and Christensen, 2004; Räisänen et al., 2004; Ekström et al., 2005; Beniston et al., 2005; Buonomo et al., 2005; Frei et al., 2005; Shkolnik et al., 2005) also indicate a general increase in precipitation extremes on the daily time scale. In an analysis of seven PRUDENCE simulations, all driven by HadAM3H or HadAM3P boundary data for the A2 scenario, Frei et al. (2005) found the average 5-year return value of winter 5-day maximum precipitation to increase in southern Scandinavia (5–20°E, 55–62°N) by 10–25% from 1961–1990 to 2071–2100, depending on the RCM. In a central European region (5–15°E, 48–54°N), the changes in winter varied from a decrease of 2% to an increase of 11%, but larger increases were found in autumn and spring. In both regions, the changes in wintertime precipitation extremes were similar to the change in precipitation intensity as averaged over all wet days but smaller than the increase in winter mean precipitation, which was also affected by an increase in the number of wet days (Figure 11.3.3.6).

[INSERT FIGURE 11.3.3.6 HERE]

Frei et al. (2005) only investigated the uncertainty associated with the choice of the RCM, not the uncertainties associated with the driving GCM and the forcing scenario. Over the British Isles, an older version of the Hadley Centre GCM-RCM system (HadCM2-HadRM2) simulated much larger increases in extreme precipitation than a more recent version (HadAM3H-HadRM3H) (Ekström et al., 2005). Driving both of these RCMs (HadRM2/3H) with HadCM2, Buonomo et al. (2005) showed that the UK and European patterns of extreme precipitation change were relatively insensitive to the change in RCM formulation. They showed large areas of significant change in daily to 30-day annual maximum precipitation. In particular, there were European average increases of 13–18% in 2–20 year return period daily precipitation with increases greatest for those extremes which are the rarest and shortest duration (i.e., the most intense), both in relative and thus absolute terms.

In the Mediterranean area and in central Europe in summer, where reduced mean precipitation is projected, short-term precipitation extremes may either increase or decrease. In an analysis of several indices of heavy precipitation in eight recent GCM simulations, Tebaldi et al. (2005) found insignificant changes of varying sign in the Mediterranean area. Frei et al. (2005) found, for the PRUDENCE simulations, a general decrease in extreme precipitation in the Iberian Peninsula throughout the year, whereas changes of varying sign were found elsewhere in southern Europe (see also Christensen and Christensen, 2003; 2004). In central Europe in summer, the change in the 5-year return value of one-day precipitation varied from –13% to 21%, with larger differences between the RCMs than in winter. However, the models consistently indicated a larger increase, or a smaller decrease, in extreme precipitation than would have been expected from the changes in the average intensity and frequency of precipitation events (Figure 11.3.3.6), a result also supported by Buonomo et al. (2005).

Simulated changes in extremely high or low precipitation accumulation on monthly and longer time scales are to a first approximation similar to the changes in mean precipitation (Räisänen, 2005). However, there are indications of increased interannual variability particularly in the Mediterranean region (Räisänen, 2002; Giorgi and Bi, 2005), which tends to make the extremes slightly more severe than expected from the changes in the mean. Both on daily and longer time scales, much larger changes are expected in the recurrence frequency of precipitation extremes than in the magnitude of extremes. For example, Frei et al. (2005) estimated that, in Scandinavia under the A2 scenario, 5-day winter precipitation totals that in the present climate occur once in 8–18 years would occur once in 5 years in 2071–2100. Similarly, using the idealized CMIP2 simulations with a gradual doubling of CO_2, Palmer and Räisänen (2002) found up to a five-fold increase in the frequency of very high DJF seasonal precipitation in northwestern Europe. Analysing a HadRM2 regional simulation driven by the HadCM2 AOGCM, Huntingford et al. (2003) found an even larger increase in the recurrency of 30-day precipitation extremes in Britain, with 40-year present-day extremes occurring once in 3–4 years in the years 2081–2100 when the HadCM2-simulated global mean temperature was 3.7°C higher.

Changes in drought in Europe have been studied using a variety of measures. Voss et al. (2002) found an increase in the length of the longest dry spells in central and southern Europe in a high-resolution GCM, consistent with a decrease in the number of precipitation days also found in this area in many other studies (e.g., Semenov and Bengtsson, 2002; Räisänen et al., 2003; 2004; Frei et al., 2005). Little change in dry spell length was found in northern Europe. Tebaldi et al. (2005) got similar results from eight recent AOGCM simulations and Beniston et al. (2005) from the PRUDENCE simulations. Räisänen (2005) found the mean of 20 CMIP2 simulations to indicate a 10–30% decrease in the 20-year minimum of JJA seasonal precipitation in southern and central Europe at doubling of CO_2, which was similar to or slightly larger than the decrease in mean JJA precipitation in these simulations. In northern Europe, no consistent signal was found among the models.

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

11.3.3.3.5 Wind speed

Although many studies have suggested increased wind speeds in northern and/or central Europe (e.g., Zwiers and Kharin, 1998; Knippertz et al., 2000; Leckebusch and Ulbrich, 2004; Pryor et al., 2005a) in the future, the results remain model- and possibly method-dependent. Slight decreases in wind speeds have also been reported, for example in a statistical downscaling study by Pryor et al. (2005b) for the Baltic Sea area.

A key uncertainty are the changes in the large-scale atmospheric circulation. Simulations that show a decrease in average sea level pressure in northern Europe and/or the northernmost Atlantic Ocean and the Barents Sea, a pattern reminiscent of the positive phase of the NAO, tend to indicate increased wind speeds in northern Europe (e.g., top row of Figure 11.3.3.4). Such a change in the pressure pattern indicates an increase in both the time-averaged pressure gradient across Europe and increased cyclone activity in northern Europe, both of which promote stronger winds. Conversely, the northward shift in cyclone activity tends to reduce windiness in the Mediterranean area. However, although most current AOGCMs indicate at least a slight shift toward the positive phase of the NAO (Chapter 10), the details of the circulation change are model-dependent. The HadAM3H experiments used to drive most of the PRUDENCE RCM simulations (e.g., the one in the bottom row of Figure 11.3.3.4) did not show a characteristic NAO-like circulation change. Thus, these simulations only showed relatively small changes in windiness, although the changes varied seasonally and there was a tendency towards increased average and extreme wind speeds in western and central Europe in winter (Räisänen et al., 2004; Beniston et al., 2005; Rockel and Woth, 2005).

In addition to the atmospheric circulation, changes in surface layer stability may also affect low-level wind speeds, especially over local water bodies (Knippertz et al., 2000; Räisänen et al., 2004). Räisänen et al. (2004) found larger increases in wintertime wind speeds over the Baltic Sea than in Sweden and attributed this to the reduced surface layer stability associated with reduced ice cover.

Extremes of wind speed in Europe are generally associated with strong winter cyclones (e.g., Leckebush and Ullbrich, 2004), the occurrence of which is only indirectly related to the mean atmospheric circulation. Nevertheless, models suggest a general similarity between the changes in average and extreme wind speeds (Knippertz et al., 2000; Räisänen et al., 2004; Figure 11.3.3.7). A caveat to this conclusion is that, even in RCMs, the extremes of wind speed over land tend to be too low, excluding a few models that use explicit gust parameterizations (Rockel and Woth, 2005).

[INSERT FIGURE 11.3.3.7 HERE]

11.3.3.3.6 Atlantic storm track and Mediterranean cyclones

Ulbrich et al. (2005a) analyzed the climate change signals in winter storm activity (computed from 2–6 day band-pass filtered sea level pressure data) from five AR4 IPCC GCMs (ECHAM5/OM1, GFD, GISS-AOM, GISS E-R and MRI) under the SRES A1b scenario. They found increasing storm track activity for the period 2081–2100 compared to 1960–1990 over the North Atlantic between Newfoundland and the British Isles. The agreement between the signals was high with a correlation ranging between 0.46 and 0.81 between the signals from the individual models and the ensemble mean, the ECHAM5/OM1 model being closest to the ensemble mean signal.

In a doubled CO_2 simulation, Lionello et al. (2002) found a small but significant decrease in the number of Mediterranean cyclones in ECHAM4 (T106), but an increase in the number of intense cyclones. A study based on several AOGCMs shows a consistent signal (Leckebusch et al., 2005). Decreases in Mediterranean cyclone number are also supported by model studies by Vérant (2004) and Somot (2005). This decrease is most emphasized in winter. The reduction of the number of cyclones may be attributed to alterations in the average sea-level pressure pattern and in the upper-tropospheric baroclinicity, showing less favourable conditions for the development of Mediterranean cyclones (Ulbrich et al., 2005b). Other charactreristics of Mediterranean cyclones, such as cyclogenesis areas and cyclone life-time, velocity and intensity, as well as the interannual variability of the cyclone track number, appear to remain unchanged (Somot, 2005).

11.3.3.3.7 Ocean wave heights and storm surges

Some studies have addressed changes in the North Atlantic Ocean wave heights. Wang et al. (2004) used the projections of a coupled climate model for three emission scenarios. They found the winter and autumn seasonal means and extremes of significant wave heights to increase in the twenty-first century in the northeast Atlantic and southwest North Atlantic, but decrease in the midlatitudes of North Atlantic. However, the changes showed decadal fluctuations and in some regions such as the North Sea even their sign was found to depend on the emission scenario.

Woth et al. (2005) analysed changes in storm surges along the North Sea coasts, forcing a hydrodynamic storm surge model with pressure and wind data from four of the HadAM3H A2 scenario driven PRUDENCE simulations. They found up to a 20–30 cm increase in the 99.5th percentile of sea surface height (above the average sea level change) from 1961–1990 to 2071–2100 along the eastern coasts of the North Sea, but no change at the east coast of the UK. Meier (2005) used a Baltic Sea ocean model driven by data from four RCM simulations to study storm surges in the Baltic Sea. The simulations gave varying results but suggested a possibility of large changes, one of them indicating the 100-year surge in the Gulf of Riga to increase 41 cm more than the average sea level.

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 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.

Changes in both wave heights and storm surges have been addressed for only a limited set of models. The connection between these phenomena and high wind speeds implies a substantial uncertainty in these projections.

11.3.3.3.8 Cryosphere

Increased melting and decreased fraction of solid precipitation due to warmer temperatures will very likely reduce the amount of snow and the length of the snow season in Europe. Increases in total winter precipitation, as projected by models, will counteract the effects of the warming but are unlikely to balance them. In an analysis of the HadAM3H-driven PRUDENCE simulations, Jylhä et al. (2005) found the average annual number of days with snow cover to decrease by 43–60 in northern Europe (55–75°N, 4–35°E) from 1961–1990 to 2071–2100 under the A2 scenario. The average DJF mean snow water equivalent decreased by 45–60%. Slightly smaller changes were found for the B2 scenario, but RCM simulations driven by ECHAM4/OPYC3 indicated larger changes for both scenarios. Further south in Europe, absolute decreases in snow season length and snow water equivalent were smaller but relative decreases larger. Results from other studies are qualitatively similar; however in their off-line snow model calculations Beniston et al. (2003) found a 4°C winter warming (as projected for the period 2071–2100) to lead to a 110–130-day decrease in snow season length at 1000 m altitude in the Swiss Alps. Snow conditions in the coldest parts of Europe, such as northern Scandinavia and northwestern Russia (Räisänen et al., 2003; Shkolnik et al., 2005) 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).

In the present climate, about a half of the Baltic Sea is ice-covered at the height of an average winter (Tinz, 1996). Jylhä et al. (2005) estimated future changes in the winter maximum ice extent from temperature changes simulated by six AOGCMs. They found that, under the A2 (B2) emission scenario, 70–100% (30–70%) of the winters in 2071–2100 would have less ice than ever observed since 1720. Simulations with a coupled regional atmosphere-Baltic Sea model (Meier et al., 2004) suggest a slightly lower sensitivity of the ice cover to temperature changes. Nevertheless, even in these simulations the average 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 simulated to decrease by 1–2 months in the northern and 2–3 months in the central parts of the Baltic Sea. Comparable reductions in Baltic Sea ice cover were found in earlier studies (Tinz, 1996; Haapala et al., 2001; Meier, 2002).

11.3.3.3.9 Mediterranean Sea oceanography

Li et al. (2005) compared A2 scenario simulations from two stretched-grid AGCMs (ARPEGE-Climate and LMDZ) focused on the Mediterranean area. Over the Mediterranean Sea, the simulations indicated a decrease in precipitation and an increase in evaporation in the end of the 21st century, and a decrease in the heat loss by the sea surface. Following the precipitation decrease over the south of Europe, river runoff fluxes of the Mediterranean Sea catchment basin also decrease under the A2 scenario (Somot et al., 2005). Using one of these simulations, Somot et al. (2005) carried out a transient simulation (1960–2099) of the Mediterranean Sea with the Mediterranean version of the OPA ocean model at 1/8° resolution. They noted a warming (3°C) and salting (0.43 psu) of the surface waters by the end of the simulation, of the whole water column (1.2°C, 0.24 psu) and of the Mediterranean Outflow Waters (MOW, 1.9°C, 0.40 psu) in agreement with observed trends over the last decades of the 20th century. Somot et al. (2005) also found a strong weakening of the Mediterranean THC (MTHC): 20% for the intermediate circulation and 60% for the deep circulation. These results were confirmed by Li et al. (2005) with the same Mediterranean Sea model but another atmospheric simulation. Changes in the MTHC could have strong impacts (Somot et al., 2005) on the Mediterranean SST, Mediterranean climate, Mediterranean Sea ecosystems and also on the Atlantic THC through the salty MOW. However, due to uncertainties (scenario, RCM, Mediterranean model), more work is needed in order to assess the response of the Mediterranean Sea to climate change.

11.3.3.4 Uncertainties

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 to 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. Agreement on changes in windiness is still rather limited. 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 thermohaline circulation. As highlighted by the PRUDENCE studies, deficiencies in the modelling of the processes that regulate the local water and energy cycles in Europe are also an important source of uncertainty, for both the changes in mean conditions and extremes. Finally, the substantial natural variability of European climate (e.g., Hulme et al., 1999; Jylhä et al., 2004) is a major uncertainty particularly for short-term climate projections in the area.

11.3.4 Asia

11.3.4.1 Key processes

The processes of central importance to Asian climate change range from factors that control the temperature response in the center of the continent, to the various effects of a warmer atmosphere on the South Asian summer monsoon, to the distinctive dynamics that control the Meiyu-Baiyu early-summer rains, to the effects of an El-Niño-like shift in the Pacific on the Maritime continent.

In *Central Asia*, large temperature responses are favored by retreat of winter and spring snowcover, the isolation from maritime influences, and diffusion of the larger wintertime Arctic warming into the region by midlatitude eddies. With regard to precipitation, a key issue is the extent to which the processes that generate drying in the Mediterranean and Middle East penetrate eastward through the southern rim of Central Asia (from Iran to Pakistan). Poleward movement of the westerly winds is expected to produce drying in the rainy season in this region, since these winds and the associated disturbances bring water vapor inland from more humid regions.

Passing across the major mountain barriers into the domain of the powerful monsoonal flows of *South Asia*, the focus shifts to the factors that control this monsoonal precipitation in both summer and winter. There are competing effects: the increase in moisture convergence even if the monsoonal flow itself is unchanged, and a possible decrease in the strength of monsoonal circulations. The latter is to be expected (Knutson and Manabe, 1998) because much of the tropics is dominated by subsidence driven by radiative cooling of the atmosphere, with the adiabatic warming due to this subsidence balancing the cooling. If the tropics stays close to a moist adiabat as it warms (see Chapter 3), the lapse rate decreases and weaker subsidence is required to balance the same radiative cooling; one anticipates weaker upward convective mass fluxes since these must balance the downward movement of mass in the subsiding regions. An emerging consensus that the effect of increasing water vapor dominates over any such weakening of the circulation (Douville et. al., 2000; Giorgi, et. Al., 2001) needs to be reassessed with improving models. The association of ENSO with weak summer monsoons (Pant and Rupa Kumar, 1997) and the evidence of secular variation in this connection (Krishna Kumar et al., 1999; Sarkar et al., 2004) provides another focus. The ability of aerosols, particularly absorbing aerosols, to modify monsoonal precipitation (Ramanathan et al., 2005), and the ability of sustained modifications of vegetation cover to do likewise (Wei and Fu, 1998), are additional issues. The Tibetan plateau provides a distinctive set of problem for climate change projections, not the least of which is the difficulty that global models have in dealing with the dramatic relief.

Moving towards *East Asia*, the monsoonal circulations are supplemented 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 monsoonal circulations remains of potential importance for climate change, and, somewhat more generally, 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), see also Figure 11.3.4.1. The Meiyu-Baiyu rains in the early summer, which derive from disturbances of baroclinic character but strongly modified by latent heat release, provide a challenge to our dynamical intuition. While one expects increases in rainfall in the absence of circulation shifts, relatively modest shifts or changes in timing that are difficult to anticipate in the absence of detailed modelling, can significantly affect East Chinese, Korean, and Japanese climates.

[INSERT FIGURE 11.3.4.1 HERE]

Issues related to monsoonal controls continue to dominate the discussion for *Southeast Asia* and the maritime continent. The difficulty in modelling the distribution of rainfall in this region, especially in the Indonesian archipelago and the importance of model deficiencies is this region for the tropic as a whole are well appreciated (e.g., Neale and Slingo, 2003). Interannual rainfall variability is significantly affected by ENSO (Hastenrath, 1987; Ropelewski and Halpert, 1989; McBride et al., 2003), particularly June to November rainfall in southern and eastern parts of the Indonesian Archipelago, which is lowered in El Niño years (Aldrian and Susanto, 2003) and also the Sumatra-Malay Peninsula-western Borneo region and regions to its east and west. A possibility of a shift towards a more El-Niño-like mean state in the Pacific has significant implications for rainfall reduction in these regions.

11.3.4.2 Skill of models in simulating present and past climates

There is substantial variation across the region in the number of studies carried out to analyze the regional skill of GCMs.. While little work has been done with a focus on Central and Southeast Asia, a considerable amount of work deals with South and East Asia.

Central Asia

Due to the complex topography and the associated meso-scale weather systems of the high altitude and arid areas, GCMs usually do not usually perform well over the region. For example, they tend to overestimate the precipitation over arid and semi arid areas in the north (e.g., Gao et al., 2001). For the PCMDI simulations of present day climate (1980–1999), the annual mean temperature bias over Central Asia ranges from –3.9 to 2.1°C across the models, with the mean of –1°C (Table 11.3.4.1). A similar cold bias is present in DJF, MAM and SON while in JJA there is a slight warm bias. Most of the PCMDI models overestimate precipitation over the region in DJF and MAM by a few percent to 40% with the average being about 20%. The majority underestimate precipitation in JJA with an average bias of about 20%. The annual mean precipitation bias is 10% when averaged across models.

Over the Tibetan part, the PCMDI models generally perform poorly. For the annual mean temperature simulation, there is a cold bias ranging from –0.4 to –6.0ºC across models, with the mean being around –3ºC (Table 11.3.4.1). All the models greatly overestimate the annual mean precipitation (50–240%), with the average bias being 120%. Similar biases are found for each of the 4 seasons, with the greatest in MAM and DJF for both temperature and precipitation. However, due to the complex topography, and a large portion of solid precipitation, observations could well be substantially underestimating the true precipitation.

The few available RCM simulations generally exhibit improved performance in the simulation of present day climate compared to the GCMs (e.g., Gao et al., 2003a, b). The GCM simulation from Gao et al. (2003a) did not accurately simulate the distribution of precipitation and overestimated the precipitation over the northwestern portion of the Tibetan Plateau by 5–6 times. However, despite this poor performance, an RCM nested in the same GCM greatly improved the simulation of precipitation distribution, although the amounts were still 1–2 times greater than the observations.

South Asia

For the PCMDI simulations, the annual mean temperature bias ranges from –4.2 to 3.2°C across the models, with a mean of –0.5°C (Table 11.3.4.1). A average cold bias of ~1°C is found in DJF and SON while a slightly warm bias is found in MAM and JJA. The annual precipitation bias is in the range of –49% to 33% with the mean of –4%. The models usually overestimate the precipitation in DJF (model mean of 33%) and underestimate it in JJA (mean bias of –11%). The average bias is small in MAM and SON.

There are a number of assessments of the skills of AOGCMs at simulating the observed broad surface climatological features of South Asia. Large-scale tropical precipitation patterns in the winter (DJF) and summer (JJAS in this case) seasons, as simulated by several AOGCMs models have been examined by Lal and Harasawa (2000), Rupa Kumar and Ashrit (2001), and Rupa Kumar et al. (2003). Over South Asia, the summer season is dominated by the southwest monsoon, which spans the four months June through September, and distinctly characterizes the seasonal cycles of precipitation, temperature, wind and a host of other climatic parameters. The season JJAS is therefore widely used to represent this unique feature of climate over South Asia. While most models simulate the general migration of tropical rain belts from the austral summer to the boreal summer, in the Indian monsoon context, the observed maximum rainfall during the monsoon season along the west coast of India and the north Bay of Bengal and adjoining northeast India is not very realistically simulated in many models (with the exception of HadCM3 and CSIRO and to some extent in DKRZ). This may possibly be linked to the coarse resolution of the models as the heavy rainfall over these regions is generally in association with the steep orography. However, the annual cycle in the simulated precipitation averaged over the South Asian region (land and sea) showed a remarkably similar pattern to the observed (Figure 11.3.4.2), though there are substantial quantitative biases (e.g., NCAR). The annual surface air temperature patterns over the South Asian region also show a general match of gross features with the observed (Figure 11.3.4.2). The models capture the gross features of the monsoon such as low rainfall amounts coupled with high variability over northwest India. However, some of the finer details of regional significance are not represented in some of the models; for instance, ECHAM4 fails to reproduce the rainfall minimum in the rain shadow region over eastern peninsula, while HadCM2 underestimates the rainfall over the Indo-Gangetic plains (Rupa Kumar et al., 2002). The simulated monsoon rainfall patterns in these models will be affected by the coarse resolution of the AOGCMs. Horizontal as well as vertical resolutions of the atmosphere in the AOGCMs appear to be strongly related to the skill of the models on regional scale. For example, both the NCAR and the GFDL models have relatively coarse horizontal resolutions. Apart from the resolution issues, recent experiments with coupled and forced GCMs indicate that time slice experiments with forced GCMs are not able to accurately capture the South Asian monsoon response simulated in a coupled system. This suggests that the ocean-atmosphere coupling is a fundamental feature of the climate system, not only at the decadal to century time scales, but also at shorter intervals. Thus, neglecting the high-frequency SST feedback and variability seems to have a significant impact on the projected monsoon response to global warming. Douville ( ) suggests that coupling an AGCM with either a regional ocean model or a slab ocean model may possibly be a compromise between computationally expensive coupled model experiments and the affordable time-slice experiments. Further, simulated changes in the Indian summer monsoon climate are sensitive to biases in the regional SST anomalies in the southern Ocean and equatorial Pacific.

[INSERT FIGURE 11.3.4.2 HERE]

Downscaling by regional climate models has been demonstrated to provide a more realistic representation of the South Asian climate, particularly the aspects of regional topographic influences (Hassell and Jones, 1999). The Hadley Centre's regional climate model PRECIS (Providing Regional Climates for Impact Studies) has recently been used in India to simulate the South Asian climate with a horizontal resolution of 50 × 50 km. Three-member ensembles of baseline simulations (1961–1990) have been performed, with and without including the sulphur cycle. These experiments confirmed that significant improvements in the representation of regional processes over South Asia can be achieved (Rupa Kumar et al., 2005). For example, the steep gradients in monsoon precipitation with a maximum along the western coast of India are remarkably well-represented in the RCM. Such details are essential to make reliable impact assessments in sectors like water resources, as most peninsular rivers are fed by topographically induced precipitation maxima. However, PRECIS does inherit some of the inherent biases of the driving GCM (HadCM3/HadAM3); for example, the simulated annual cycle indicates a stronger-than observed onset phase of the summer monsoon and the precipitation is substantially overestimated over east central India, which are very similar to the biases present in the driving GCM (Rupa Kumar et al., 2005).

East Asia

The PCMDI models show different levels of performance in simulating the mean climate over this area (Table 11.3.4.1). The simulated temperature patterns show close similarity with observations but the annual area-mean temperatures are lower than observation (except two high-resolution models) with an ensemble mean of –2.1°C, ranging from –5.3 to 0.3°C. Simulated temperature over land area are distinctively lower than observations for all seasons but over the ocean, large warm biases are present in winter and cold biases in the warm seasons. The seasonal area-mean temperature biases vary from –6.6°C to 1.6°C. Temperature bias and inter-model differences are smallest in summer (JJA) and largest in winter (DJF).

The PCMDI models reproduce the large-scale precipitation patterns but the rain band in mid-latitudes is shifted northward in seasons other than summer. Except for one model, simulated area mean precipitation exceeds the observed precipitation on an annual basis. In winter, model biases of precipitation vary from –23% to 138% and the area mean precipitation is overestimated by 56% due to strengthening of the mid-latitude rain band over the ocean. The bias and inter-model differences are smallest in summer but the mid-latitude rain band is shifted northward, resulting large discrepancies in rainfall distribution over Korea, Japan and adjacent seas. Model bias of surface pressure in East Asia is generally negative but in summer, the Northwest Pacific High is stronger than observed and this could lead to the premature northward shift of the rain band, resulting much less precipitation in this area. The models with larger cold biases tend to produce less precipitation (correlation of 0.4).

The overall performance of participating models generally show some improvements compared to the performance of earlier AOGCMs but model bias is not improved significantly (compared to Min et al., 2004)

Simulation of the major characteristics of the summer monsoon climate over South Asia, East Asia, and the western North Pacific by the new version of the Meteorological Research Institute coupled GCM (MRI-CGCM2) was analyzed by Rajendran et al. (2004). They evaluated the model performance for mean conditions and the evolution of summer monsoon rainfall and its association with SST and basic circulation parameters, and found that the model captures the basic features but with significant discrepancies in some regions.

Traditionally GCMs have shown a poor performance in simulating the East Asia monsoon precipitation patterns. The precipitation center simulated by GCMs is usually located too far north over central China (e.g., Gao et al., 2001; Gao et al., 2004), as well as in many of the PCMDI models.

However, in the work of Gao et al. (2001), where a much higher resolution regional climate model (RegCM2) was nested in the above mentioned CSIRO model results, the simulation of the precipitation was highly improved. Not only regional details but also the spatial distribution became closer to reality. This improvement can be largely attributed to the increased horizontal resolution as discussed by Gao et al. ( ). They found that simulated East Asia large-scale precipitation patterns are significantly affected by resolution. The effect of resolution is most important during the mid to late monsoon months, when smaller scale convective processes dominate. Figure 11.3.4.3 shows the spatial correlation coefficient between the simulated and observed annual mean precipitation from their different simulations. In general, it can be seen that the coefficient increases as the model resolution increases and the topography becomes more realistic Moreover, it shows that the high-resolution simulations with the coarse CSIRO topography also perform surprisingly well which suggests that the impact of resolution may be more important than the impact of topography.

[INSERT FIGURE 11.3.4.3 HERE]

There are many studies evaluating the capability of RCMs at reproducing realistic climate features in East Asia (Ding et al., 2003; Oh et al., 2004; Sasaki et al., 2005; Kadokura and Kato, 2005; Fu et al., 2005). Ding et al. (2003) showed RegCM_NCC has improved anomaly correlation coefficient (ACC) over the Yangtze River valley where the AOGCM shows a very low ACC, This is likely to be related to the realistic representation of terrains in the regional model. There has also been several simulation studies reproducing the fine-scale climatology of small areas using a nested RCM and a very high resolution RCM (Takayabu et al., 2005) and these studies show some improvement in features associated with terrain, e.g., snow area and temperature fields. However, one of the limitations of RCMs is that the RCM performances are subjected to the lateral boundary forcings (Ding et al., 2003; Takayabu et al., 2005) and they are very limited in reproducing the strong meso-scale features, such as Typhoons. It has been pointed out that the land surface, convection, and radiation processes should be improved to decrease uncertainties in RCM experiments (Ding et al., 2003; Fu et al., 2005).

Southeast Asia

Table 11.3.4.1 summarizes the PCMDI results over this region; a cold bias of 1.6°C (range of 0.2 to –3.1) is seen, while precipitation averages 8% greater than observed, with a range amongst the models of –27% to +45%. Both the precipitation and temperature biases are distributed evenly across the seasons. The broadscale spatial distribution of rainfall in DJF and JJA averaged across the AR4 runs compares well with observations.

Rajendran et al. (2004) examined current climate simulation in the MRI coupled model over an Asian domain that included Southeast Asia. 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 CCM3 in simulating tropical precipitation, with the model forced by observed sea surface temperature. Simulation was good over the Maritime continent compared to the simulation for other tropical regions. Wang et al. (2004) assessed the ability of eleven atmosphere-only GCMs to simulate climatic means and variability in the Asian-Australian monsoon region when forced with observed sea surface temperature variations. They found that the models' ability to simulate observed interannual rainfall variations were poorest in the Southeast Asian portion of the domain, where observed SST- rainfall links were reversed in the model. This represented a shortcoming in model processes that is likely to be relevant to the reliability of enhanced greenhouse simulations.

Rainfall simulation across the region at finer scale has been examined in some studies. McGregor et al. (1998) reported that a ten-year regional simulation with DARLAM at 44km resolution nested in the CSIRO Mk 2 AOGCM was generally acceptable at simulating the spatial distribution, magnitude and seasonality of the simulated precipitation. McGregor and Nguyen (2003) conducted a ten-year current climate simulation at 80km resolution centred over Indochina using the CSIRO stretched grid model CCAM nested in CSIRO Mk 3. Summer (JJA) precipitation simulation was reasonable, although Indochina tended to be drier than in the observations. Aldrian et al. (2004a,b) have conducted a number of simulations with the MPI regional model for an Indonesian domain, forced by broadscale observed conditions and by the output of the ECHAM4 GCM. Aldrian et al. (2004) found that the model was able to represent the spatial pattern of seasonal rainfall, although the monsoonal contrast over Java was poor in the simulation nested in ECHAM4. The effect of varying resolution was also examined, and it was found that a resolution of at least 50 km was required to simulate rainfall seasonality correctly over Sulawesi. A coupled regional model was used by Aldrian et al (2004b) and this formulation was found to improve regional rainfall simulation over the oceans. 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.

Finally in considering current climate simulation for Southeast Asia, it should be noted that current AOGCMs continue to have some significant shortcomings in representing ENSO variability (see Section 8.4.1.2.1).

11.3.4.3 Climate projections

11.3.4.3.1 Mean temperature

Central Asia

Only a few publications focus on climate projections over Central Asia and Southeast Asia. Application of regional climate models to simulate present climate and the future changes over Central Asian regions has only started recently (Gao et al., 2003a,b)

For the A1B emission scenario for the period 2079–2098 (compared to 1979–1998), the PCMDI models simulate an annual mean temperature increase of 2.6°C to 5.1°C with the multi-model average being 3.8°C (Table 11.3.4.2). The models agree on the warming in all seasons, with a spread across the individual models of ~3°C. The warming does not show a distinct seasonal dependency. Across the 4 seasons, JJA shows the greatest warming, 4.3°C, while DJF shows the lowest, 3.4°C. Meleshko et al. (2004) analyzed results from a multi-model ensemble of 21st century projections for Northern Eurasia under the B2 and A2 emission scenarios. They too found an increase in temperature throughout the year.

For the Tibetan plateau, under the A1B scenario for the years 2079–2098 (compared to 1979–1998), the annual temperature of the region shows an increase of 2.8–6.1°C in the PCMDI models. The multi model mean change is 4.0°C (Table 11.3.4.2). The models agree on the warming in all seasons. The warming is in all seasons in the mean ~4°C and the across-model spread is ~3°C. For the Tibetan Plateu and Northwest China, Xu et al. (2003a, 2003b) analysed the climate changes induced by greenhouse gases and aerosol based on AOGCMs' simulations and found consistent results.

Greater warming over the Plateau compared to the surrounding areas is simulated by a regional model (Gao et al., 2003), with the warming being most significant in high altitude areas, e.g., over the Himalayas. The higher temperature increase over high altitude areas can be explained by the decrease in ice-albedo feedback due to snow and ice melting (Giorgi et al., 1997). This phenomenon is found to different extents in some of the PCMDI models (e.g., medium and high resolution versions of MIROC3.2) while not in others (e.g., ECHAM5/MPI-OM). However the multi PCMDI model average change shows the largest warming over the Plateau, especially in DJF, MAM and SON.

South Asia

For the A1B scenario in the year 2079–2098 (compared to 1979–1998), the PCMDI models show an increase of 2.0–4.7°C in annual temperature in the region. The multi model mean change is 3.2°C (Table 11.3.4.2). By season, the warming ranges from 2.8°C (in JJA) to 3.5°C (in DJF). Other studies using coupled atmosphere-ocean general circulation models indicate general warming in a greenhouse gas increase scenario, the changes becoming particularly conspicuous after the 2040s (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). There is considerable consensus in temperature projections.

Considering all the land-points in India according to the resolution of each AOGCM, the average (all-India) temperature is calculated for the entire duration of model simulations and for different experiments (Figure 11.3.4.4). GHG simulations with IS92a scenarios show marked increase in temperature by the end of 21st century relative to the baseline. There is a considerable spread among the models in the magnitudes of temperature projections. In case of mean annual temperature, the increase is of the order of 3 to 6°C. The temperature however shows comparable increasing trends in IS92a and A2 sceanarios but B2 shows slightly lower trends.

[INSERT FIGURE 11.3.4.4 HERE]

All the models show positive trends indicating widespread warming into the future. Examination of the spatial patterns of annual temperature changes in the two future time slices for different models indicates that the warming is more pronounced over the northern parts of India. The different models/experiments generally indicate the increase of temperature to be of the order of 2–5°C across the region. The warming is generally higher in IS92a scenario runs compared to A2 and B2 simulations. Also, the warming is more pronounced during winter and post-monsoon months compared to the rest of the year. Interestingly, this is a conspicuous feature of the observed temperature trends from the instrumental data analyses over India (Rupa Kumar et al., 2002, 2003). Douville et al. (2000) found from GCM diagnostics that all models simulated a stronger warming over land than over sea.

East Asia

The annual mean temperature for the period of 2079–2098 (compared to 1979–1998) is projected to increase from 2.4 to 3.4°C by the PCMDI models, with an ensemble mean of 3.4°C (Table 11.3.4.2). In EAS, the warming is largest in winter, especially in the northern inland area but the area mean difference is not significant compared to the other seasons. There seems no relationship between model bias and size of warming. The uncertainty range is not larger than the other regions of the Asian continent. The ensemble mean changes of annual temperature based on SRES A2 scenario is 4.1°C, similar to the earlier model result (Min et al., 2004). The spatial pattern of larger warming over northwest EAS is closely matched with the ensemble mean of the earlier models.

Future climate changes over East Asia are projected from multi-model ensembles (MMEs) of selected coupled atmosphere-ocean general circulation model (AOGCM) simulations based on IPCC SRES A2 and B2 scenarios (Min et al., 2004). The overall projection results from four MMEs show that East Asia will experience a warmer climate in the 21st century. The projection results are not sensitive to the MME method. Area-averaged temperature changes for three 30-year periods of 2020s, 2050s, and 2080s simulated by MME7 A2 (B2) scenario ensembles are 1.2 (1.4), 2.5 (2.4), and 4.1°C (3.2°C) increase, respectively (Figure 11.3.4.5).

[INSERT FIGURE 11.3.4.5 HERE]

Spatial patterns indicate that temperature increases are larger over the continental areas than oceanic areas and that the areas of larger inter-model variability are in accord with those of stronger climate change. The inter-model variability in temperature changes is much smaller than the signal in the projection of temperature changes. A significant difference in projected patterns between A2 and B2 scenario ensembles (defined as a potential impact of greenhouse-gas mitigation) appears in the 2080s temperature field over the southwestern part of East Asia (Figure 11.3.4.6).

[INSERT FIGURE 11.3.4.6 HERE]

There has been a time-slice experiment with high resolution MRI-GSM to examine the effect of horizontal resolution on small-scale phenomena and short-term variability (Mizuta et al., 2005). Two 10-year periods are integrated: present and late 21st century. Global distributions of mean precipitation, temperature, and wind fields agree well with observation in general and it is beneficial to see improved regional-scale phenomena due to more realistic topography. However, the experiment is limited to the simulation of mean state, and did not include interannual and decadal variability.

There are several studies which downscale CGCM simulations using RCMs (Gao et al., 2001; 2002; Kwon et al., 2003; Choi et al., 2004; Kurihara et al., 2005; Kanada et al., 2005). Kwon et al. (2003) reports a 150 year simulation (1951–2100) over East Asia with 27-km resolution using MM5. The initial and boundary conditions of MM5 are provided from the simulation of ECHAM4/HOPE based on SRES A2 greenhouse gas-only scenario. Regional projections of Kwon et al. (2004) show more realistic characteristics of regional climate than other previous studies. It is projected that the area mean temperature is increased about 5°C over East Asia by 2100, which is slightly warmer than those of coupled model simulations (Boo et al., 2005). Future climate for 2081–2100 over Japan was projected by RCM20 with 20-km resolution driven by the lateral boundary provided from MRI-CGCM2 following SRES A2 scenario (Kurihara et al., 2005). Temperature increased more than 2°C during cold months, exceeding 4°C around the Okhotsk Sea and the difference of increase is about 1°C between summer and winter.

Southeast Asia

The temperature projection of the AR4 global models for the Southeast Asian region varies between 1.5 and 3.7°C with little seasonal variation (Table 11.3.4.2), There is some tendency for the warming to be stronger over Indochina and the larger landmasses of the archipelago (Figure 11.3.4.7) The range of warming in the region is slightly less than the global average warming for this set of models (1.8 to 4.1°C).

Projected regional temperature changes in the region based on a range of recent AOGCMs have been prepared by Giorgi et al. (2001) and Ruosteenoja et al (2003), and over Indonesia by Boer and Faqih (2004). Giorgi et al. (2001) found that in regional average terms, AOGCMs simulated the warming rate in the region as less than the global average rate. In Ruosteenoja et al (2003), the projected regional warming in 2070–2099 scaled to the full range of SRES scenarios was 1 to 4.5°C. The results of Boer and Faqih (2004) were broadly consistent (regional warming in 2080 of 2.5 to 3.5°C under the A2 and B2 emission scenarios).

[INSERT FIGURE 11.3.4.7 HERE]

The DARLAM regional model was used in a simulation across the region by McGregor et al. (1998) and more recently the CSIRO stretched grid model (McGregor and Dix 2001) was used in a climate change simulation centred on the Indochina Peninsula (AIACC 2004, at a resolution of 14 km). These simulations 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.

11.3.4.3.2 Mean precipitation

Central Asia

For the traditional Central Asia, and the A1B emission scenario for the period 2079–2098 (compared to 1979–1998), the PCMDI models simulate a slight decrease annual mean precipitation (average of –4%). A more pronounced decrease up to 10% is found in the southwestern part of the region, while in the northern part precipitation slightly increases. However, the individual models simulate quite different magnitudes and do not agree on the sign of the change (Table 11.3.4.2). For the annual mean, 7 models simulate an increase (of 1–6%) while the other 13 models simulate a decrease (of 1–19%) of precipitation. For the different seasons, the agreement among the models fluctuates. In JJA and MAM, most of the models project a precipitation decrease (with a very large across-model spread in JJA of 4–59%), and in DJF, most of the models agree on a precipitation increase. In contrast, in SON, the model projections are not consistent; half of the models project increase, while the others simulate a decrease.

Meleshko et al. (2004) analyzed results from a multi-model ensemble of 21st century projections for Northern Eurasia under the B2 and A2 emission scenarios. They showed an increase of precipitation in winter for the entire region. In summer, precipitation was projected to increase in the northern part of the region and decrease in the south.

For the Tibetan Plateu and Northwest China, Xu et al. (2003a, 2003b) showed a general increase of precipitation in the future. The AR4 models simulate a consistent increase of annual mean precipitation. The multi model mean increase of annual precipitation is 9%, and the individual projections range from very slight to 30% increase (Table 11.3.4.2). The precipitation increase is consistent among the models in all seasons, however the agreement is lower in JJA and SON, where 5 and 7 models project a precipitation decrease, respectively. The highest mean precipitation increase is projected for DJF with it being 19%.

South Asia

Most of the AR4 models project a decrease of precipitation in DJF, and an increase during the rest of the year by the end of the 21st century. The precipitation increase by the model ensemble mean under A1B scenario is about 10% in JJA and SON, and only 5% in MAM, while in DJF, the mean precipitation decrease is 6%. However, the spread among the individual models is considerable (Table 11.3.4.2). The precipitation increase (decrease) in JJA (DJF) ranges from 2–23% (3–36%). The ensemble mean annual precipitation is projected to increase by 8% at the end of the 21st century under A1B scenario, while the individual model simulations range from a decrease by 16% to an increase by 20%. However, only 3 of 20 models project an annual precipitation decrease (see also Kripalani et al., 2005).

Over South Asia, coupled atmosphere-ocean general circulation models indicate enhanced rainfall in a greenhouse gas increase scenario, the changes becoming particularly conspicuous after the 2040s (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). There is some disagreement among the models on rainfall changes. Rupa Kumar and Ashrit (2001) found significant differences in the projections of two state-of-art atmosphere-ocean coupled climate models, for the Asian summer monsoon rainfall. In a study with four different GCMs, Douville et al. (2000) found a significant spread in the summer monsoon precipitation anomalies despite a general weakening of the monsoon circulation (also see May, 2004b). They concluded that, for decades to come, the increase in the atmospheric water content could be more important than the increase in the land-sea thermal gradient for understanding the evolution of the monsoon precipitation. They found that the monsoon sensitivity to CO_2 doubling is not only related to changes in the horizontal transport of water vapour, but also to changes in the precipitation efficiency, which depends on soil moisture. Therefore, the treatment of land surface hydrology in the GCMs is a critical factor in determining monsoon sensitivity. Stephenson et al. (2001) argue that the consequences of climate change may be manifested in different ways in the physical and dynamical components of monsoon circulation.

Considering all the land-points in India according to the resolution of each AOGCM, the country-level (all-India) averages of rainfall are calculated for the entire duration of model simulations and for different experiments (Figure 11.3.7.4). GHG simulations with IS92a scenarios show marked increase in rainfall by the end of 21st century relative to the baseline. There is a considerable spread among the models in the magnitudes of precipitation projections, but more conspicuously in the case of summer monsoon rainfall. The increase in rainfall from the baseline period (1961–1990) to the end of 21st century ranges between 15 and 40% among the models. At a glance one can realize that the change in rainfall in A2 and B2 scenarios is not as high as that noted earlier in IS92a scenarios. Compared to A2 scenario, the B2 simulations show much subdued trends into the future. Most models project enhanced precipitation during the monsoon season, particularly over the northwestern parts of India. There is very little or no change noted in the monsoon rainfall over a major part of peninsular India.

Douville et al. (2000) found from GCM diagnostics that not all models simulate a stronger monsoon. They argue that the weakening of ENSO-monsoon correlation could also 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, recent model diagnostics using ECHAM4 to investigate this aspect indicate that both the above mechanisms can play a role in monsoon changes in a greenhouse warming scenario (Ashrit et al., 2001). This study also indicates that, while the monsoon deficiency due to El Niño may not be as severe as present in a greenhouse warming scenario, the favourable impact of La Niña seems to remain unchanged. Later, 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.

East Asia

Precipitation is projected to increase for all seasons by the PCMDI models for the period of 2079–2098, with the change being largest in winter. Some models simulated drier condition for this period but most models simulated wetter condition over the continental area. The spatial patterns show wetter continental areas and drier oceanic areas, in summer, but in other seasons precipitation increases both continental and oceanic area in the 30–40N latitude-band.

Projections from multi-model ensembles (MMEs) of selected coupled atmosphere-ocean general circulation model (AOGCM) simulations based on IPCC SRES A2 and B2 scenarios (Min et al., 2004) indicate East Asia will experience wetter climate in the 21st century and the increase is larger for greater warmings. Spatial patterns indicate that precipitation increases are larger over the continental area than the oceanic area and that the areas of larger inter-model variability are in accord with those of stronger climate change. The inter-model variability (noise) in precipitation changes is as large as that of ensemble mean (signal). No significant differences can be found between precipitation patterns of A2 and B2 scenario ensembles because of the dominant inter-model variability.

The 150-year East Asia regional projections (Kwon et al., 2003, 2004) show that the area mean precipitation is enhanced by 6% over East Asia by 2100, which is wetter than those of coupled model simulations (Boo et al., 2005). The precipitation is increased during the warm season but not in the cold season, consistent with the previous studies and AOGCM results. However, large multi-decadal variations are present in the long-term projection in accordance with observation. Precipitation projection using RCM20 (Kurihara et al., 2005) indicated that daily precipitation will increase during the warm season, June to September, with a increase rate of 10–20%, especially over western Japan.

Southeast Asia

Regional precipitation change has shown a mixed pattern in AOGCM intercomparison studies. The analysis of Giorgi et al. (2001) showed SE Asia as a region where models consistently showed little change in precipitation. In the analysis of Ruosteenoja et al. (2003) we see both simulated rainfall increase and decrease amongst the models, but with a slight bias to increase, and, consistent with Giorgi et al. (2001), a relatively narrow range of projected changes for 2070–2099 (mostly –10 to +15%). The results were very similar when analysed over an Indonesian domain by Boer and Faqih (2004). Hulme and Sheard (1999a,b) prepared patterns of rainfall change across Indonesia and the Philippines composited from a range of earlier AOGCM simulations forced by IS92a scenarios. They 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. Indeed, their conclusion was that ‘no generalisation could be made on the impact of global warming on rainfall’ in the region.

However, the set of AR4 simulations present a more consistent picture of regional precipitation increase than obtained in these earlier studies. Annual precipitation change for SEA region (comparing the period 2070–2099 in the A1B scenario to 1979–1999) averages 6% with a range of –3 to +15% (see Table 11.3.4.2). The results are very similar when broken down by season. Regional averaged, annual rainfall increases in 18 of the 20 simulations (Table 11.3.4.2). Figure 11.3.4.8 illustrates the spatial distribution of average DJF and JJA rainfall change and inter-model consistency. The region of strongest increase (at least 15 out of 20 models showing increase) broadly follows the ITCZ, lying over northern Indonesia and Indochina in JJA, and over southern Indonesia and Papua New Guinea in DJF. Away from the ITCZ precipitation decrease is often simulated. The pattern is broadly one of wet season rainfall increase and dry season decrease.

[INSERT FIGURE 11.3.4.8 HERE]

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. For example, Figure 11.3.4.9 indicates that due to topographical effects, the magnitude of simulated rainfall change can vary significantly across Indochina. More recently, Takayabu et al. (2005) compared three regional climate model simulations over Indochina (as well as other Asian domains). 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.

[INSERT FIGURE 11.3.4.9 HERE]

11.3.4.3.3 Changes in extremes

High-resolution GCMs are beginning to provide a more realistic representation of the extremes in daily precipitation during the Indian summer monsoon season, allowing the development of more reliable projections of short-duration precipitation characteristics. May (2004a) notes that the ECHAM4 GCM at a horizontal resolution of T106 simulates the variability and extremes of daily rainfall in good agreement with the observations, even better than the reanalysis ERA-40. ECHAM4 time slice experiments indicate that the intensity of heavy rainfall events is generally increased in the future (2070–2100), 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).

Keeping in view the need to analyse the changes on a smaller space-time scale to derive information related to the extremes, regional climate models provide a better handle for examining the projections of extremes. In the IS92a scenario, HadRM2 shows an overall decrease in the number of rainy days over a major part of the country. This decrease is more in western and central parts of *South Asia* (by more than 15 days) while near foothills of Himalayas and in northeast India the number of rainy days is found to increase by 5–10 days. Increase in GHG concentrations may lead to overall increase in the rainy day intensity by 1–4 mm/day except for small areas in northwest India where the rainfall intensities decrease by 1 mm/day. The model results also indicate that there will be an overall increase in the highest 1-day rainfall over a major part of *South Asia*. This increase may be up to 20 cm/day. However, in some parts of northwest India, decrease in extreme rainfall has been noticed in the GHG experiment, up to about 10 cm/day. The model also shows that there will be increase in extreme maximum and minimum temperatures all over *South Asia* due to increase in greenhouse gas concentrations. This increase will be of the order of 2–4°C both in minimum and maximum temperatures (Krishna Kumar et al., 2003). Results from the regional climate model PRECIS indicate that the night temperatures increase faster than the day temperatures in both A2 and B2 scenarios, with the possibility that the occurrence of cold extremes is likely to be less severe into the future. PRECIS 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., 2005).

Gao et al. (2002) analyzed the change of extreme events in *East Asia* focused on China using an RCM (Gao et al., 2002). They show that both daily maximum and daily minimum temperature are increased but that the diurnal temperature range is decreased due to the higher increase of minimum temperature. The number of hot spell days in summer significantly increases while the number of cold spell days in winter significantly decreases. The number of rainy days increases most noticeably in Northwest China and parts of inner Mongolia. Heavy rain days increase over some sub-regions in Southeast and South west China. Tropical storms tend to increase and the dominant path of tropical storms landing is also found in the simulation.

Kimoto et al. (2005) showed that the high-resolution version (T106 atmosphere) of their AOGCM, MIROC3.2, successfully represents the frequency distribution of daily precipitation intensity over Japan. For the 21st century projection, their result 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. The increase in non-precipitating days would occur in winter, while that in heavy rainfall days would occur mainly in warm seasons. This is consistent with the historical trend reported by Fujibe et al. (2005) from four-hourly data for hundred years over Japan, in which increased frequency of intense precipitation is found for all the seasons and regions of Japan. Mizuta et al. (2005) examined various extremes indices (Frich et al., 2002) from the results of a time-slice climate change experiment with the 20 km-mesh AGCM of MRI/JMA. They found statistically significant increases in R10 (The number of days with precipitation over 10 mm) and SDII (simple daily intensity defined as the total precipitation divided by number of wet days) in western part of Japan and Hokkaido Island. Overall evidence seems to indicate the historical and expected future increases in extremely heavy precipitation in Japan.

Some high-resolution modelling studies also investigated specific kinds of disturbances that give extremely heavy precipitation. Hasegawa and Emori (2005) showed from a time-slice climate change experiment with the T106 CCSR/NIES/ FRCGC AGCM that daily precipitation associated with tropical cyclones over western North Pacific would increase due to increased water vapor in the warmed climate. Kanada et al. (2005) showed from a time-slice climate change experiment with the 5-km mesh non-hydrostatic limited area model of MRI/JMA 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 of the warmed climate.

Mizuta et al. (2005) also examined temperature-based extremes indices (Frich et al., 2002) over Japan from the results of the 20 km-mesh AGCM of MRI/JMA and found that the changes in the indices are basically those expected from the mean temperature increase.

There are a few studies (Kwon et al., 2004; Boo et al., 2005) aimed at understanding changes in the extreme climate over the Korean Peninsula based on the long-term simulations. Kwon et al. (2004) analyzed ten indicators suggested by Frich et al. (2002) using an AOGCM simulation based on the SRES A2 scenario. They found the indicators related to minimum temperature change showed a decreasing trend but indicators such as heat wave duration index showed a distinctive upward trend, consistent with Mizura et al. (2005). Boo et al. (2005) investigated changes in regional climate arising from global warming with a high-resolution downscaling simulation for the period 1971–2100. The main focus was on temperature and precipitation extremes over Korea. Frequency distribution of daily temperature shows an increase in the mean by about 5.5°C from 1971–2000 to 2071–2100 with little change in the variance. Under the climate change scenario, hot events are expected to be more frequent and severe, while cold events occur less often and are warmer. The increasing trend of temperature is associated with an increasing trend in precipitation. The long-term increase produces an increase in the number of the days of heavy precipitation and in their corresponding amount. The increasing rate is marked in the northern region compared to the southern region, since the regional projection has large changes in local precipitation over Korea.

Lee et al. (2005) analyzed the multi-model ensemble of eight AOGCMs in the historical (20C3M) and the scenarios (A2, A1B, B1) runs to evaluate the model performance in simulating the *East Asian* summer climate and to investigate the effect of global warming on the summer climate over the *East Asia*. From comparison of the observation and the 20C3M experiment, it is found that the multi-model ensemble quite well simulates the *Northeast Asian* summer precipitation and circulation, especially in the first two EOF modes and the associated regressed field. The first EOF mode represents the decaying phase of ENSO, which contributes to the development of the Philippines Sea anticyclone. The second EOF mode is associated with the fast transition of ENSO. The circulation pattern related to the first two EOF modes in observation and the model correspond well with the patterns in the decaying and developing phases of ENSO respectively in Wu et al. (2003). In future climate, the increase of the precipitation to 2099 in the A2 and A1B simulations reaches 10% over the *Northeast Asian* region. From EOF analysis, it seems that the increased *Northeast Asian* summer precipitation due to global warming is contributed by the effect of the enhanced monsoon circulation in the decaying phase of El Niño rather than the mean linear increase of global climate or the circulation in the fast transition period of ENSO. The reason why the second mode associated with the decaying phase of ENSO becomes important in the increase of precipitation over the Northeast Asia due to the global warming is not understood.

Concerning the *East Asian* Monsoon, reproducibility of the Baiu depends on the model's horizontal resolution. A time-slice experiment with super-high-resolution global model and cloud-resolving regional climate models (20-km mesh MRI/JMA AGCM and 5-km mesh NHM) is performed (Kusunoki et al., 2005; Yasunaga et al., 2005). Results with an AGCM with 20-km grid size show that the Meiyu-Baiu rainfall increases over the Yangtze River valley, the East China Sea, and western Japan, while rainfall decreases over the Korean peninsula and northern Japan. A northward shift of the Baiu front is not clear in the warming climate, and the termination of the Baiu tends to be delayed until August. A 5-km mesh cloud resolving regional climate model is forced by 20-km mesh AGCM to investigate the small-scale response to large-scale conditions simulated by the 20-km mesh AGCM. While the rainfall does not vary in June between the present and warmed climates, there is more rainfall in July in the warmed climate. Moreover, the frequency of the precipitation greatly increases with the intensity of the precipitation in July in the warming climate simulation. Classification of the area larger than 900,000 km^2 are more frequently seen in July in the warming climate than in the present climate, resulting in more rainfall. The increase of the large system in July is the most remarkable in the vicinity of Kyushu Island, and the baroclinicity in that area is stronger in the warming climate.

Kitoh and Uchiyama (2005) investigated onset and withdrawal pentad dates in Asia summer rainfall season based on daily precipitation data of IPCC AOGCM simulations. Figure 11.3.4.10 shows the horizontal distribution of the withdrawal dates of the summer rainy season based on the climatological pentad mean precipitation for the CMAP observations, seven AOGCM ensembles for the present day (1981–2000 in 20C3M), for the 2081–2100 of the SRES A1B experiments, and its changes. At the end of the twenty-first century, changes in the withdrawal dates differ from region to region. It clearly delays near Japan and *South Asia* from Indian through Indochina peninsular, while it becomes earlier in South China. Over India and Indochina peninsula, there is about two pentads delay, while there is about four pentads delay over the Arabian Sea and the Bay of Bengal. A large delay can be found over Baiu region to the south of Japan, where one or two months' delay of rainy season withdrawal is seen. These regions experience large increases in precipitation throughout the summer season by extending the rainy season from only early summer in the present-day case to the whole summer season in the warming climate. On the contrast, rainy season ends earlier over an extensive region in South China where some regions experience more than a one month early retreat of summer rainy season. In summary, ensemble mean of AOGCM simulations with ordinary resolutions reveal, at the end of the twenty-first century under the SRES A1B scenario, a delay in Baiu rain withdrawal around Japan and an earlier withdrawal in Meiyu rain over southern China, although the change in onset dates is relatively less.

[INSERT FIGURE 11.3.4.10 HERE]

Weakening of *East Asian* winter monsoon is already noted (e.g., Hu et al., 2000). 17 AOGCM results with 1% CO_2 experiment at years 61–80 relative to years 1–21 reveals weakened winter monsoon associated with the shallower planetary wave trough over the east coast of the Eurasian Continent (Kimoto, 2005). Hori et al. (2005) defined the *East Asian* winter monsoon (EAWM) index as –v at 850 hPa averaged over 20–40N, 120–150E using 9 AOGCM output. Most models show a weakening of the *East Asian* winter monsoon accompanied by an anticyclone of 850 hPa circulation anomaly over the North Pacific, which corresponds to a weakened and/or northward pressure gradient along the eastern coast of the Eurasian continent in a weakened EAWM.

For *Southeast Asia*, few studies have been made at the regional level as to how temperature and precipitation variability and extremes may change, but it can be expected that the region would share in the global tendency for increased daily extreme high temperatures as the climate warms (see Section 10.3.6.2). Weisheimer and Palmer (2005) demonstrated that extreme seasonally averaged temperatures that currently occur in 5% of years over *Southeast Asia*, could occur in over 50% of years by the late 21st century. Rainfall variability will be affected by changes to ENSO and its effect on monsoon variability, but this is not well understood (see Sections 10.3.5.1 and 10.3.5.2). 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 should also be said that the region is likely to share in the tendency for daily extreme precipitation to become more intense under enhanced greenhouse conditions. This has been demonstrated in a range of global and regional studies (see Section 10.3.6.1), 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 Section 10.3.6.3 there is evidence in general of likely increases in tropical cyclone intensity, but less consistency about how occurrence will change (also see 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 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 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 Section 10.3.5.1).

11.3.4.4.4 Regional sea level rise

Choi et al. (2002) examined the regional sea-level rise over the Northwestern Pacific Ocean using the NCAR-CSM coupled climate model with enhanced oceanic horizontal resolution over the region. They found that the sea-level rise over that region was enhanced compared with the global average mainly due to exceptionally large warming and sea-level change near the entrance of the Kuroshio extension. Unnikrishnan et al. (2005), using HadRM2 simulations for South Asia, report an increase in the frequency of cyclonic storms in the Bay of Bengal towards 2050s in an IS92a scenario. Using the HadRM2 results to drive a storm surge model for the region, they report greater number of high surges in the IS92a scenario.

11.3.4.4.5 Uncertainties

Major uncertainties concerning projected climate change for this region are:

- Very limited assessment of simulated changes to regional climatic means and extremes by current climate models. A range of regional studies are required.
- Uncertainty regarding the future behaviour ENSO contributes significantly to uncertainty about monsoon behaviour in the region and tropical cyclone behaviour in northern parts of the region.
- High potential for local climate changes to vary significantly from regional trends due to the regions very complex topography (multiple islands and very mountainous).

11.3.5 North America

11.3.5.1 Key processes

The North American continent spans several climatic zones, from subtropical to arctic, through the mid-latitudes. The region from roughly 40° to 60° N lies in the westerlies, with an upper-level ridge over the Rocky Mountains and a trough over the Hudson Bay, particularly strong in winter. The North Pacific storm track terminates on the West Coast, and the Rocky Mountain cordillera acts as a moisture barrier for the entire continent (Figure 11.3.5.1). Under the permanent influence of the Aleutian low, the coastal regions from Alaska to Oregon receive the largest annual precipitation amounts. The thermal contrast between the cold continent in winter and the warm waters of the Gulf Stream favours the development of the North Atlantic storm track along the East Coast, from Florida to Nova Scotia. The regions northeast of the Gulf of Mexico up to Labrador also receive substantial annual precipitation amounts. Most of North America, with the exception of the southwest USA and northern Mexico, is under the influence of atmospheric moisture convergence transported by travelling weather systems; the southwest USA and northern Mexico region is very arid under the overall influence of a subtropical ridge of high pressure.

[INSERT FIGURE 11.3.5.1 HERE]

North America is affected by the two important patterns of oscillations in the Northern Hemisphere: the El Niño – Southern Oscillation (ENSO) and the North Atlantic/Arctic Oscillation (NAO/AO). The positive phase of ENSO produces above-normal rainfall over large regions of USA, from southern California, the central and Gulf Coast states, and even Florida (Hagemeyer and Almeida, 2004). The positive phase of NAO/AO, characterised by strong westerly flow, induces a cooling and drying over eastern Canada, due to the strengthened advection of cold Arctic air masses in winter.

The North America monsoon system (NAMS; e.g., Higgins et al., 1997) is a circulation that develops in early July over north-western Mexico and the south-western USA (Arizona, New Mexico, Utah, Colorado, Nevada, California). Similar to but of smaller scale than the Asian monsoon, the NAMS has associated low-level winds over the Gulf of California undergoing a seasonal reversal, from northerly prevailing winds during the winter to southerly prevailing winds during the summer. The shift of wind patterns associated with the NAMS brings a pronounced increase in rainfall over the otherwise very arid region of the southwest USA, and ends the late spring wet period in the Great Plains (e.g., Bordoni et al., 2004).

The Great Plains low-level jet (LLJ) transports considerable moisture from the Gulf of Mexico into the central USA, playing a critical role in the summer precipitation there. Several factors appear to be contributing to the strength of the moisture convergence into the Mississippi River Basin during the night and early morning, resulting in prominent nocturnal maximum in the northern plains of USA (such as Nebraska, Iowa) (e.g., Augustine and Caracena, 1994).

11.3.5.2 Simulation skill at regional scale

11.3.5.2.1 Global coupled models (CGCMs)

Coordinated experiments such as the Coupled Model Intercomparison Project (CMIP; Meehl et al., 2000) have established the skill of CGCMs in reproducing the overall general circulation of the atmosphere (e.g., Wallace and Osborn, 2002; Covey et al., 2003; forthcoming AR4 CGCMs analysis papers, 2006), as well as several features of the North American climate and its variability (e.g., Coquard et al., 2004). Models vary in their ability to reproduce the observed patterns of pressure, surface air temperature and precipitation over North America, but there are also several systematic aspects to their performance. For example, simulated mean sea level pressure is generally too low over Northern Alaska and the western part of the Canadian North-West Territories, probably due the inability of coarse-resolution models to properly block incoming cyclones in the Gulf of Alaska.

All models simulate successfully the overall pattern of surface air temperature over North America, but in models used for TAR, the model-mean surface air temperature is more than 2°C too warm over the Hudson Bay and the Canadian Prairies to its west. By contrast the model-mean surface air temperature is too cold over high elevation, despite the fact that terrain elevations are underestimated due to coarse resolution. In winter, models tend to underestimate the meridional temperature gradient and, in parts of western USA, the errors exceed the interannual temperature variability. In summer, the model-mean surface air temperature is too warm over most of North America and, in western USA, the average error greatly exceeds the interannual variability. Several models overestimate the surface air temperature in summer by as much as 3 to 6°C; other models with weaker warm bias in summer underestimate the temperature in winter and spring. Overall the normalised error (i.e., the ratio of average model errors to observed interannual variability) is smaller in winter than in summer. The model average is close to observations for some regions (e.g., over south-eastern Canada and north-eastern USA in summer), but large inter-model differences exist, indicating compensation of errors between models. A link has been noted between individual model temperature bias and variability (e.g., Räisänen, 2002): in winter the correlation is negative over most of the region while in summer it is positive mostly over the northern part of the region.

Over the western USA where the seasonal cycles are strong, some models produce a seasonal cycle for spatially averaged surface air temperature and precipitation in good agreement with observations, while others tend to over-predict precipitation in the winter or exaggerate the amplitude of the annual cycle of surface air temperature. The model-mean simulated precipitation is excessive over an elongated region from Alaska to Mexico, on the windward side of major mountain ranges, probably as an artefact of overly spatially broad and underestimated terrain height in coarse-resolution CGCMs. All models over-predict winter precipitation over the Vancouver Island area and western USA (eastern Washington, eastern Oregon, Montana, Wyoming, Utah and Nevada), with precipitation amounts more than 50% above the observations. This error appears as a failure to properly simulate the rain-shadow of mountain ranges with coarse-resolution models. In some models, this over-prediction of precipitation extends throughout the year except in July, August and September. The mean of all models fails to represent the region of high precipitation over south-eastern USA, while the north-eastern states are too wet in summer. The wet region in the Midwest is displaced westward, and summer precipitation is incorrectly represented over Mexico and the Gulf of Mexico. There is a suggestion that there may be some relationship between horizontal resolution of the atmospheric model and the ability to simulate surface air temperature throughout the year and precipitation in winter, in agreement with the results of Duffy et al. (2003). The reason appears to be that winter-time precipitation is dominated by resolved large-scale processes and interaction with topographic features, while summer-time precipitation is dominated by parameterised convection hence the weaker resolution dependence.

Several interacting factors are responsible for the simulation weaknesses of CGCMs over North America; some errors are model specifics, dependent on details of model formulation. An overly frequent occurrence of light precipitation, referred to as the drizzle problem, is noted in most models. Subgrid-scale parameterised processes such as convection appear to control precipitation in summer over North America, and most models appear rather weak in this respect, with resulting systematic excessive precipitation in summer (Coquard et al., 2004). As noted by Huth et al. (2001) and Ruosteenoja et al. (2003) some CGCMs have a strong tendency to favour surface temperatures close to 0°C, due to simplistic soil thermodynamic parameterisation that overestimates the latent heat during phase transition of soil water; this can result in an underestimation of variability in northern regions during soil melting/thawing seasons. Land surface processes, through their interaction with the overlaying atmosphere, also play an important role in determining the North American climate. Poutou et al. (2004) showed that the soil freezing processes have significant effects on regional boreal climate. Lakes and wetlands occupy a large fraction of Canada and these open water surfaces are often not accounted for in CGCMs. The results of Krinner (2003) show that wetlands seem to play a more important role than lakes in cooling the boreal regions in summer and in humidifying the atmosphere. SSTs contribute importantly to the distribution and intensity of precipitation in winter over western North America. Models using "flux adjustments" to constrain the sea surface temperature (SST) tend to exhibit smaller precipitation errors, which points to a link between SST and western continental precipitation. To remove ad hoc flux-adjustment schemes, higher spatial resolution for the ocean component is required to permit ocean eddies to form. There are indications however that higher atmospheric resolution is also required to derive the full benefits of increased ocean resolution (e.g., Roberts et al., 2004). Analysis of surface temperature indicates that warm spells over North America tend to be associated with a characteristic pattern of cool SST in eastern north and central Pacific Ocean. Gerhunov and Douville (2004) showed that this association is well reproduced in their simulated data, showing that CGCMs are able to capture the spatial signature of large-scale anomalous circulations associated with warm spell over North America. GCMs also appear to reproduce with reasonable skill the NAMS (e.g., Arritt et al., 2000), lending some confidence in their ability to represent the effects of climate change on the NAMS.

Overall the skill at simulating current climate over North America has improved with AR4 CGCMs. Current-climate simulations of AR4-generation CGCMs indicate the following characteristics over North America. The ensemble-mean of CGCMs reproduces very well the annual-mean mean sea level pressure distribution. The maximum error is of the order of ±2 hPa, with the simulated Aleutian low pressure extending somewhat too far to the North of Alaska and the pressure trough over the Labrador Sea not being deep enough; this annual-mean error pattern arises mostly from the winter biases where the errors are about twice as large (±4 hPa). In summer the depth of the simulated thermal low pressure over the southwest states is somewhat excessive. The ensemble-mean of CGCMs reproduces well the annual-mean temperature distribution. Over the Rocky Mountains simulated temperatures are too cold by more than 2°C; this cold bias is smallest in winter months over Alaska and in summer months over the southwest states. The simulated temperatures over the eastern part of the continent are too cold by more than 1°C throughout the year. The simulated temperatures over the Canadian Prairies are somewhat too warm, by more than 1°C in the annual mean and by more than 2°C in winter. The ensemble-mean of CGCMs reproduces the overall distribution of annual-mean precipitation (Figure 11.3.5.2). There is however a generalised tendency for excessive precipitation, the excess reaching 1 to 2 mm/day over high terrain in the West of the continent; over the central states North of the Gulf of Mexico, there is a precipitation deficit of 1 to 2 mm/day. The precipitation bias pattern varies little with season; an exception is the region bordering the Gulf of California – the NAMS region – where there is a deficit in summer.

[INSERT FIGURE 11.3.5.2 HERE]

11.3.5.2.2 Regional climate models

Since the TAR there have been a number of regional modelling experiments driven either by reanalyses or control runs (i.e., current-climate simulations) of CGCMS and AGCMs, or both (e.g., Pan et al., 2001; Han and Roads, 2004; Kim et al., 2002; de Elía et al., 2006).

RCM simulations driven by reanalyses

Some of the fundamental assessments from the TAR still hold true. RCMs succeed in reproducing the overall climate, including fine-scale features forced by resolved topography and land-sea contrast. RCMs simulations over North America exhibit somewhat disconcerting sensitivity to parameters such as the domain size (e.g., Juang and Hong, 2001; Pan et al., 2001; Rojas and Seth, 2003; Miguez-Macho, 2004; Vannitsem and Chomé, 2005; de Elía et al., 2006) and the intensity of the large-scale nudging (e.g., von Storch et al., 2000; Miguez-Macho et al., 2004; de Elía et al., 2006). RCMs' simulations results from the coordinated North American Regional Climate Change Assessment Program (NARCCAP) show that typically 76% of the individual models temperature biases are within the range ±2°C and 82% of the precipitation biases are within the range ±50% (Figure 11.3.5.3)

[INSERT FIGURE 11.3.5.3 HERE]

Simulations driven by reanalyses have become more specific in their goals compared to those in the TAR. While general validation based on seasonal mean values is still a focus, more research now concentrates on particular phenomena, such as daily extremes (Kunkel et al., 2002; Leung et al., 2003a), extreme floods and droughts (Anderson et al., 2003; Sushama et al., 2006), diurnal cycle of precipitation (Liang et al., 2004b), and particular regional atmospheric features such as the LLJ in the central USA and the precipitation maximum in the south central USA (Gutowski et al., 2003 and 2004) and the NAMS in the southwest of USA (Anderson et al., 2000a, 2000b; Anderson and Roads, 2002; Xu and Small, 2002).

RCMs are in general more successful at reproducing North American cold-season temperature and precipitation (e.g., Han and Roads, 2004; Pan et al., 2001), since the warm-season climate is more controlled by fine-scale, mesoscale and convective, precipitation events (Giorgi et al., 2001). This remains generally true despite the wide variety of convective parameterisation schemes (e.g., Liang et al., 2004b; Leung et al., 2003a); Gutowski et al. (2004) found however that spatial patterns of monthly precipitation for the USA were better simulated in summer than winter in their results. Strong regional topographic forcing improves the skill of regional model simulations (e.g., Wang et al., 2004).

In one RCM, Kunkel et al. (2002) found that simulated extreme precipitation events were in good agreement with observations regarding magnitudes of 1-day heavy precipitation thresholds, but for 7-day events, skill is variable across regions, being good in the east and the Great Plains, but poor in the Mississippi Valley. Gutowski et al. (2003) show that a 50-km RCM has some skill at simulating central USA precipitation extremes on daily or longer time scales, but none on shorter time scales; also resolutions of several tens of kilometres are insufficient to simulate well the diurnal cycle of precipitation in the central USA. Leung et al. (2003a) examined 95th percentile of daily precipitation and found generally good agreement across many areas of the Western USA, although it should be noted that there remain important methodological issues regarding how to appropriately compare station observations with model grid-point precipitation extremes. 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 and convergence.

Studies targeted at the representation of convection, such as the EUROCS project, indicate that all convection parameterizations tested failed to represent the gradual diurnal transition over continental North America, with moistening of the top of the planetary boundary, then the lower to mid-troposphere, after which deep precipitating convection can begin (Chaboureau et al., 2004). A large part of the error in the parameterizations arises from an incorrect sensitivity of the convection schemes to environmental humidity and the representation of entrainment mixing between convective plumes and the local environment (Derbyshire et al., 2004), processes that appear essential for the correct representation of moist convection in summer over North America.

RCM simulations of present-day climate using GCM boundary conditions

Current-day simulations of RCMs driven by control runs of GCMs are generally inferior to those driven by reanalyses, due to the errors introduced at the boundaries from the global models. Comparisons are usually made between the quality of the RCMs simulations and those of the driving GCMs. The RCMs simulations generally inherit several biases of the nesting GCMs. The sensitivity of simulated precipitations to changing lateral boundary conditions (BC) from reanalyses to GCMs appears low in winter and high in summer; for surface air temperature, however, the sensitivity appears to be much higher in winter than in summer (e.g., Han and Roads, 2004; Plummer et al., 2006). Improvements and increased resolution of the driving GCMs compared to those used to drive RCMs in the TAR have led to higher quality of BC for RCMs. It is important to note however that, unless otherwise indicated, RCMs results reported in this AR4 are mostly based on simulations driven by TAR-generation CGCMs.

11.3.5.2.3 Statistical downscaling

Since the TAR there have been numerous statistical downscaling (SD) studies but several important challenges remain largely unresolved (Leung et al., 2003) as discussed in Section 11.2.1 (and a key resource is the TGICA guidance document on statistical downscaling; Wilby et al., 2004). A significant fraction of studies were devoted to model inter-comparison; others highlighted the synergy between techniques used for statistical downscaling and those used for seasonal prediction. Although a few novel applications have emerged, regional climate-change projections by SD methods continue to be most widely applied to the water resource, agricultural and conservation sectors. However, a handful of integrated assessments have begun to appear.

11.3.5.3 Climate-change projections

11.3.5.3.1 CGCMs projections

Based on CGCMs projections under a specific scenario of GHG and aerosols evolution, the climate-change "response" to CO_2 doubling is defined as the difference between mean results for a selected time window centred on the time of doubling of CO_2 concentration and corresponding time window in a simulation with constant CO_2 concentration at the current value.

The temperature response of all TAR-generation CGCMs is positive everywhere within the region and for all months. The model-mean temperature response exceeds the inter-model standard deviation (IMSD) everywhere over the domain for all seasons, indicating that models are consistent in predicting a warming over the North American region. For most of the region, the temperature response is larger in winter and increases toward the north due too the well-known snow-albedo feedback. Over western USA, however, the model-averaged response is larger during the warm months than during the cold season; this may be an artefact of the coarse resolution of these models that underestimate the elevation of mountain ranges, and hence underestimate the snow-albedo feedback process.

There is generally poor agreement of TAR-generation CGCMs on the amplitude and even sign of the regional precipitation response over North America (Giorgi et al., 2001; Coquard et al., 2004). In summer the precipitation response is less than the spread between models over most of the region, hence the precipitation response can be said to be everywhere consistent with the null hypothesis. Over western USA, the model-average response indicates a small decrease of precipitation during the summer and fall when precipitation is weak, and a larger increase of precipitation in winter when the precipitation is stronger; but the IMSD is large, the response of some models projections disagreeing in sign. In winter the precipitation response exceeds the spread between models only over Canada, northern USA and in some places over Mexico and the extreme southern USA; precipitation response in winter indicates a consistent increase in northern high latitudes and eastern North America.

Given the wide range of response of CGCMs, it is interesting to investigate whether there is a relationship between the strength of the climate-change response of a particular model and its ability to simulate the current climate conditions; if such a relationship existed, it could be used as a confidence factor to be attributed to each projection in forming the ensemble mean. The study of Coquard et al. (2004) revealed that the existence of a relationship between the temperature response over western USA and simulation error over the northeastern Pacific region; the average the models with the smallest error predicted a modest but significantly larger warming (2.35°C) than the models with largest error (2.04°C). The same study for precipitation, however, did not show any obvious relationship, i.e., the range of precipitation responses of models with the smallest errors did not differ appreciably from that of models with larger errors.

The latest AR4-generation CGCMs climate-change projections under the SRES A1B scenario, for 20-year projections for the period 2079–2098, using the 20-year simulation period 1979-1998 as reference, give the following results for the ensemble mean over North America. The ensemble-mean of CGCMs projects an increased low-level zonal flow, with decreasing mean sea level pressure in the northern region (reaching –1.5 hPa) and a slight increase in the south (less than 0.5 hPa); this tendency is most pronounced in autumn and winter. On an annual basis, the pressure decrease in the north exceeds the IMSD by a factor 3 on an annual-mean basis and 1.5 in summer, so it is significant; the pressure increase in the south, on the other hand, is small compared to IMSD.

The ensemble-mean of AR4 CGCMs projects warming of the annual-mean surface air temperatures varying from 2 to 3°C along the western, southern and eastern continental edges (there at least 13 out of the 18 models projecting a warming in excess of 2°C), up to more than 5°C in the northern region (where 15 out of the 18 CGCMs project a warming in excess of 4°C). This warming is highly significant, exceeding the IMSD by a factor of 3 to 4 over most of the continent. The northern warming varies from more than 7°C in winter (in this season nearly all CGCMs project a warming exceeding 4°C) to as little as 2°C in summer (Figure 11.3.5.4). The warming in the USA is projected to exceed 2°C by nearly all models, and to exceed 4°C by some 7 CGCMs.

[INSERT FIGURE 11.3.5.4 HERE]

The ensemble-mean of AR4 CGCMs projects an increase of annual-mean precipitation in the North, reaching +20%, which is twice the IMSD, so significant. These precipitation changes are projected to prevail in all seasons. The winter is characterised by a more extensive increase of precipitation (exceeding +30%) while models are divided on the sign of precipitation changes in summer (Figure 11.3.5.5). The ensemble-mean of AR4 CGCMs projects a decrease of annual-mean precipitation in the South, exceeding 20 % reduction in the Southwest. This reduction is close to the IMSD, so only marginally significant; it is noteworthy however that 4 out of the 18 CGCMs do project an increase of precipitation there. In spring and summer there is a widespread projected decrease of precipitations in the South and Southwest part of the continent, with only 2 CGCMs projecting an increase of precipitation in spring there.

[INSERT FIGURE 11.3.5.5 HERE]

11.3.5.3.2 High-resolution AGCM projections

Time-slice projections with AGCMs can provide useful indications on the sensitivity of global models to resolution. Similar large-scale patterns of response are generally found in AGCMs and CGCMs, but some important regional-scale differences due to better representation of topography and other factors at high resolution. Temperature responses can vary between the AGCM and CGCM by as much as ±1 to 2°C depending on regions. Averaged over the USA, Govindasamy (2003) found that an AGCM projected a larger (smaller) increase in precipitation than the CGCM in winter (summer), resulting in insignificant differences in the annual-mean precipitation responses.

Higher-resolution AGCMs are quite skilful at reproducing cyclone tracks and intensities. In a CO_2-doubling projection, Geng and Sugi (2003) found a decrease of cyclones in the Northern Hemisphere (NH) mid-latitudes in all seasons, due to a reduction in the number of weak- and medium-strength cyclones, while strong cyclones tend to increase: 20% increase in NH summer, including over the East Coast of North America.

11.3.5.3.3 RCMs projections

Since the TAR there have been a number of RCM climate-change projections over various sub-regions of North America, using a variety of nesting CGCMs. These include projections over the western USA which has been an area of intense attention given the dominance of complex topography and high concern regarding climate change in this region of limited water resources (Kim et al., 2002; Snyder et al., 2002; Bell et al., 2004; Leung et al., 2004), the north-eastern USA (Horgrefe et al., 2004; Lynn et al., 2005), the south-eastern USA (Mearns et al., 2003), the continental USA (Pan et al., 2001; Chen et al., 2003; Han and Roads, 2002; Liang et al., 2004), western Canada (Laprise et al., 2003), and the entire North America (Plummer et al., 2006; see Figure 11.3.5.6 and 11.3.5.7).

[INSERT FIGURES 11.3.5.6 HERE]

[INSERT FIGURES 11.3.5.7 HERE]

The enhanced resolution of RCMs allows for a better representation of certain processes and their response under climate change. For example, it is found that more spatial structure of precipitation change was found in the RCM simulations that employed the higher resolution (Han and Roads, 2004). In simulations of the western USA, several studies relate to projected changes in snow amount, particularly as a function of elevation. Results confirm earlier ones presented in the TAR (Giorgi et al., 2001), that the warming in the simulations resulted in increased rainfall at the expense of snowfall, reduced accumulation or earlier snow melt (Kim et al., 2001 and 2002; Snyder et al., 2002; Leung et al., 2004), although the extent of this depended on the degree of warming and elevation. Sushama et al. (2006) studied extreme flows of six North American river basins (Fraser, Mackenzie, Yukon, Nelson, Churchill and Mississippi) and found significant decrease in the number of days with flows below the 10th percentile threshold for the high-latitude basins and significant decrease in the number of days with flows above the 90th percentile threshold for Nelson and Mississippi.

Several experiments confirm the now well-established contrast in the responses of RCMs and driving CGCM (Kim et al., 2002; Snyder et al., 2002; Mearns et al., 2003; Liang et al., 2004). A particularly interesting contrast in this regard was found by Pan et al. (2004) regarding a distinct "warming hole" in the central USA where observations have shown a cooling trend in recent decades; this area of very little warming in the climate-change experiment, which was not at all evident in the driving model, is attributed to changing pattern of the low-level jet frequency and moisture convergence. Han and Roads (2004) also found in their results that precipitation response differed significantly in summer, even averaged over the entire domain of the continental USA, with the CGCM generally producing a small precipitation increase and the RCM a substantial precipitation decrease. Han and Roads attributed the differing climate-change response to differences in the physical parameterisations used in the CGCM and RCM. Plummer et al. (2006) also found differing summertime surface air temperature climate-change responses in a RCM when two different sets of parameterisations were used; differences in precipitation responses however were generally small, despite the fact that one set of parameterisations corrected a significant summertime precipitation excess.

Multi-member ensembles of RCM climate-change projections allow exploring the uncertainty related to internal variability (e.g., Pan et al., 2001a; Yang and Arritt, 2002). In a three-member ensemble of an RCM integrated from different initial conditions, Snyder et al. (2002) found the variability among members to be low compared to the interannual variability, and recommended longer runs rather than ensembles. Leung et al. (2004) analysed a three-member ensemble of an RCM integrated over the western USA, nested by different realisations of a global model, and found that, for several river-basin areas of the domain, the variability among ensemble members for both monthly temperature and precipitation was within the variability captured by 20 years of a single simulation. In several cases, RCMs responses differ significantly from one another, even when nested by the same CGCM. For example, Chen et al. (2003) found that the RCMs disagreed, particularly in summer, regarding climate-change response: two RCMs projected larger temperature changes than did the CGCM in summer. In areas downwind of the Great Lakes, these RCMs projected precipitation increases whereas the CGCM projected precipitation decreases.

Several studies focused particularly on changes in extreme climate events. Bell et al. (2004) examined changes in temperature and precipitation 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. Changes in extreme precipitation (exceeding of 95th percentile) followed changes in mean precipitation, with decreases in heavy precipitation found for most areas, except for two hydrologic basins that experienced increases in mean precipitation. Leung et al. (2003a) examined changes in extremes in their simulations of the western USA. In general they found increases in diurnal temperature range in six sub-regions of their domain in summer. Extremes in precipitation increased in the northern Rockies, the Cascades, the Sierra and British Columbia, along with increases in mean precipitation. In two river basins, decreases in mean precipitation still resulted in increases in extreme events, a result that was reported earlier for other climate-change projections (Giorgi et al., 2001). They also noted increases in rain-on-snow events that could contribute to more severe flooding.

11.3.5.3.4 Statistical downscaling

Since the TAR there have been a large number of SD climate-change projections applied to various impact sectors and sub-regions across North America. As with RCMs, much 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 (Dettinger et al., 2004; 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 HadCM3, ECHAM4 and NCAR-PCM depend critically on the realism of GCM-downscaled wintertime temperature variability and its interplay with precipitation and snowpack accumulation and melt (Salathé, 2005).

Several articles focus on the effect of downscaled precipitation and temperature changes on agricultural potential and land quality. Bootsma et al. (2005) interpolated climate-change projections for the Atlantic region of Canada from CGCM1 to a 10–15 km grid and computed a range of agroclimatic indices (e.g., crop heat units, effective growing degree-days, water deficits) for 2010–2039 and 2040–2069. The interpolation procedure yielded smaller winter and summer temperature increases, and smaller summer and autumn precipitation increases than the SD tool (Wilby et al., 2002). Uncertainty due to multiple GCMs also increased the range of the indices. Work by Georgakakos and Smith (2001) further highlights the risks of drier than present soil moisture conditions in the south-eastern US, whereas Zhang et al. (2004) project increased soil loss and reduced wheat yield for the Oklahoma region. However, the latter study also showed that adoption of conservation tillage and no-till measures would be effective in controlling soil erosion under the climate-change scenario downscaled from HadCM3.

A key advantage of SD techniques is their potential for generating site-specific and/or exotic scenarios for specific impact sectors. For example, local wind speeds are notoriously difficult to downscale using RCMs because of highly localised controls on vertical and horizontal airflows. Nonetheless, Sailor et al. (2000) applied a neural network approach to estimate wind power from GCM output. Other challenging applications of downscaling include projections of changes in average ski seasons for southern Ontario (Scott et al., 2003), and estimates of extreme heat-related mortality in California (Hayhoe et al., 2004). Construction of land-use change scenarios for the New York Metropolitan Region involved downscaling the SRES A2 and B2 scenarios into a local narrative of alternative rural-to-urban land conversions (Solecki and Oliveri, 2004).

There have been a small, but growing number of downscaling studies that seek to integrate regional climate-change impacts and/or explore adaptation options. For example, Vanrheenen et al. (2004) showed that projected reduction in winter, spring and summer streamflow in the Sacramento-San Joaquin River basin can not be fully mitigated without demand modification and investment in water infrastructure improvements. Similarly, Payne et al. (2004) found that changes in the regime of the Columbia River could be accommodated by earlier reservoir refill and greater storage allocated for compensation flows, but at the expense of less reliable hydropower production. Quinn et al. (2001) adopted a broader perspective to assess vulnerability of other water dependent activities such as water quality, ecosystem health and socioeconomic welfare within the San Joaquin River basin. Finally, Hayhoe et al. (2004) produced a standard set of downscaled temperature and precipitation scenarios to underpin a multi-sector impact assessment for California. Large increases in temperature and extreme heat were found to drive significant impacts on temperature-sensitive sectors. For example, under both the A1F1 and B1 SRES scenarios there are overall declines in snowpack and loss of alpine and subalpine forests, as well as reduced dairy production and degraded wine quality.

11.3.5.3.5 Land-use change experiments related to climate change

North America may see significant climate impacts from the effects of land use and cover changes (LUCC) both from changes within the region and from effects taking place outside the region. The effects of LUCC may be divided based on their source or origin and by the processes responsible for the transformation (Kabat et al., 2002; Pielke et al., 2002; Marland et al., 2003). LUCC-related climate impacts can be divided into those related to biogeochemical impacts and those related to biophysical impacts (Brovkin et al., 1999).

Biogeochemical impacts affect the rate of biogeochemical processes, such as the carbon and nitrogen cycles. Human activities affect the rate of release and uptake of carbon into and from the atmosphere (Kabat et al., 2002). The net effect of human land-cover activities increases the concentration of greenhouse gases (GHG) in the atmosphere; it has been suggested that these effects have been significantly underestimated in the future climate projections used in the SRES scenarios (Sitch, 2005). Biophysical impacts include those resulting from changes in albedo, vegetation height, transpiration rates, and leaf area. Details of how these changes translate into different forcings are found in Chapter 2, Section 2.5.

Deforestation of boreal forests and conversion of mid-latitude forests and grasslands to agriculture have been simulated to cause cooling (Bonan et al., 1992). These processes tend to lead to cooling, in part by lowering average daily maximum temperatures, while daily minimum temperatures are not much affected. Because of this, the mean diurnal temperature range also decreases. If these effects are combined with the observed temperature increases in the observed record, this means that maximum temperatures remain relatively constant; i.e. the warming is offset by cooling from land cover, and the minimum temperatures are increased by the warming trend as has been observed in the recent continental temperature records (Bonan, 2001).

These simulations of anthropogenic land-cover change effects up to the present indicate that these changes could be responsible for a 2°C cooling for many of the areas that have experienced agricultural conversion (Chase et al., 2000; Betts, 2001; Bounoua et al., 2002). Over agricultural areas this cooling effect would offset a portion of the expected warming due to GHG effects in the future. One significant land-cover conversion impact, not yet simulated in GCMs, is urbanization. Although small in aerial extent, conversion to urban land cover has been shown to create urban heat islands associated with considerable warming (Arnfield, 2003). Since much of the population of North America is located in urban environments, this means that many people will be exposed to warmer climates, especially increases in mean daily minimum temperatures, a variable known to have health consequences (Karl and Knight, 1997; Meehl et al., 2005).

Much of the North American continent has already been affected by land-cover change, and land-cover conversion to agriculture may continue in the future, especially in parts of the western USA and Canada and portions of Mexico (RIVM, 2002). Countering this trend is the extensive reforestation occurring on the eastern portion of the continent, which is likely to continue in the future. In these areas climate impacts may include local warming associated with reforestation and decreased albedo values. In addition, high rates of urbanization may begin to play a role in the climate of these locations. Although urbanization is generally associated with warming, there is also a suggested link to increased precipitation rates and cloud cover over urban areas that could influence local climates in these areas (Jin et al., 2005). Depending on large-scale precipitation and moisture fluxes into the region, this could lead to different future climate outcomes.

Tropical forest conversion to agriculture has been shown to lead to significant local warming, an impact that is likely to have future implications for North American climate conditions (De Fries et al., 2002). 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. These effects dominate over that of increased albedo. On the North American continent, this could directly affect regions of Mexico and the Caribbean. Future SRES B1 and A2 scenarios differ in their projected land-cover change impacts on temperatures in this region. Although the local-scale processes should lead to a warming in many of these forested areas, in the SRES simulations these local effects are overridden by large-scale circulation impacts of land-cover change in other regions, specifically in the Amazon in this case.

Large-scale deforestation in the Amazon (as is seen in the SRES A2 scenario) is projected to lead to about 2°C warmer temperatures over the region (McGuffie et al., 1995; Gedney and Valdes, 2000; Costa and Foley, 2000). The larger scale impacts of this deforestation are not yet resolved. Avissar and Worth (2005) suggest that through teleconnection processes the entire region from northern Mexico through the USA experiences drying for at least a portion of the year. Feddema et al. ( ) find contrasting results, i.e. the warming over the Amazon is accompanied by a large reduction in the water vapour flux to the atmosphere. This slows the Hadley circulation over Middle and North America allowing the ITCZ to migrate further north, which in turn allows further northward entrainment of moist air into the region. Hence, in the A2 SRES scenario for 2100, with a near complete Amazon deforestation, Middle America will be wetter, overriding the warming and drying that might occur due to local deforestation. This same moisture source also leads to a significant increase in regions affected by the southwest monsoon in the southeastern USA. However, if there is local deforestation without accompanying deforestation of the Amazon, then the local effects will manifest themselves to lead to local warming and drying, an effect shown in the future B1 SRES scenario.

These simulations suggest that the effects of future land-cover change over the North American continent 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. 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. The past may not be a good indicator of the types of land transformation that may occur in the future. Second, current land-process models are not completely up to the task of simulating all the potential impacts of human land-cover transformation. Such processes as adequate simulation of urban systems, agricultural systems, ecosystem disturbance regimes and soil impacts are not yet represented, and if they are need they still need significant improvement before they can give a complete estimate of the climate effects from anthropogenic land transformations.

11.3.5.4 Aspects of North American climate and climate change

Until recently climate-change projections over North America using RCMs or high-resolution AGCMs have been undertaken without a coordinated effort to produce ensembles under controlled experimental conditions. As a result the present assessment is strongly based on the results of AR4 CGCMs. Unless otherwise stated, the quoted range of values that are cited corresponds essentially to those projected for the end of the century (2080–2100) under SRES A1B – a middle-range scenario comprised between SRES A2 (high) and B1 (low) – by the participating AR4 CGCMs (after eliminating some clear outliers). The range of values in parenthesis correspond to those obtained with the probabilistic scheme of Tebaldi et al. (2005) that weights both model biases and spread amongst CGCMs, for 5 and 95th percentiles of the distribution; clearly this range of values is always narrower than the first one. For all regions of North America, the magnitude of the climate changes are projected to increase almost linearly with time.

In general the projected climate changes over North America follow the overall features of those over the Northern Hemisphere. There will be a northward displacement of the mid-latitude westerly flow and its associated storm tracks, with lowering surface pressure over the northern portion of North America and a weak rise of surface pressure over the southern part. The lowering surface pressure in the North will be strongest in wintertime, reaching –1.5 to –3 hPa, in part as a result of the warming of the continental Arctic airmass. This will also be associated with a northward displacement of the Aleutian low-pressure centre and a north-westward displacement of the Labrador Sea trough. In summer, the East Pacific subtropical anticyclone is projected to broaden, strengthening particularly off the coast of California and Baja California, resulting in an increased airmass subsidence and drying over south-western North America. A generalised warming trend is projected for the entire continent, with the largest warming occurring in wintertime over northern parts of Alaska and Canada, reaching 10°C in the northernmost parts. In summertime, warming should range between 3 and 5°C over most of the continent, with weaker values near the coasts.

The magnitude of precipitation changes in climate projections appears to scale directly with the precipitation amounts in simulations of current climate. Hence it appears natural to describe precipitation projections in term of relative changes, as fraction of current precipitation amounts, rather than absolute amounts. The area-average fractional changes can be used to scale local precipitation amounts to obtain local changes in precipitation amount, which is particularly relevant in mountainous regions with important orographic 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 moisture flux and of the intensity of its convergence and divergence, resulting in a general increase of precipitation over most of the continent but the southwest most part. Precipitation is projected to increase in the northern part of the Continent, by as much as +30% in the northernmost parts in wintertime. Warming is expected to be small over the Pacific Ocean, +1 to +2°C, and larger over the continent, about +3°C over the western portion. The contrast between land and ocean projected warming is expected to contribute to the amplification of the subtropical anticyclone off the West Coast of USA (e.g., Mote and Mantua, 2002). As a consequence of the broadening Pacific subtropical anticyclone and its associated subsidence, a decrease of annual precipitation is projected for the southwest USA and northern Mexico. In summertime there should be a decrease of precipitation reaching –20% over the some West Coast states of the conterminous USA, and a weak increase of precipitation over Alaska and northern Canada.

Based upon AR4-CGCMs projections, surface air temperature changes appear to scale rather systematically between the various SRES scenarios for all regions of North America. For example, the climate-change warming for the period 2980–2099 under SRES B1 is smaller than that under SRES A1B by a factor varying between 0.65 and 0.73, and that under SRES A2 is larger by a factor between 1.07 and 1.30, for all regions and seasons. Precipitations are projected to increase for regions ALA and ENA in winter and for region GRL for winter and summer, and the fractional increase scales rather systematically: the projected increase under SRES B1 is smaller than that under A1B by a factor between 0.73 and 0.82, and that under SRES A2 is larger by a factor 1.15 and 1.29. Projected summertime precipitation changes under various SRES scenarios do not scale well with GHG amounts for regions of conterminous USA. The reason appears to be related to the fact that projections of climate changes over North America indicate both an amplification of changes (including the hydrological cycle) and a northward displacement of the mid-latitude westerly flow (and associated storm tracks) with enhanced GHG. For regions well to the North or South of the separating line between the projected precipitation increase and decrease, the amplification aspect dominates and projected climate changes scale with GHG amounts; for regions near the separating line however, the latitudinal displacement of the climate-change pattern prevents a scaling of the projected changes with GHG amounts. Déqué et al. (2006) noted a similar behaviour for projected changes over southern Europe.

The following subsections make statements specific to individual regions of North America. Unless otherwise indicated, the statements pertain to the spatial average for the region.

***ALA**, land part of region (60–72°N and 170–103°W), i.e., Alaska, Yukon and most of Canadian North-West Territories.*

Consequent with the general poleward amplification of climate-change warming, this region (as well as CGI) is expected to undergo the largest warming in North America. The warming should be larger in winter as a result of reduced period with snow cover, with temperature changes between +5 and +8°C (+6.2 and +7.6°C), and smaller in summer, with temperature changes between +2 and +4°C (+1.9 and +3.1°C).

In keeping with the northward displacement of the westerlies and the intensification of the Aleutian low, the region should undergo an increase of precipitation, particularly in winter with an increase between +15 and +40% (+21 and +32%); in summer, the increase should be between +8 and +23% (+12 and +21%). The increase in precipitation could be larger on the windward slopes of the mountains as a result of increased orographic precipitation.

***CGI**, land part of region (50–85°N and 103–10°W), i.e. Greenland, easternmost part of Canadian North-West Territories, northern part of Manitoba, Ontario and Québec, and Labrador.*

Consequent with the general poleward amplification of climate-change warming, this region (with ALA) is projected to undergo the largest warming in North America. The warming is projected to be largest in winter as a result of reduced period with snow cover, with temperature changes between +4 and +8°C (+5.3 and +6.8°C), and smaller in summer, with temperature changes between +2 and +4°C (+2.0 and +3.2°C).

In keeping with the northward displacement of the westerlies and the northwestward displacement of the Labrador Sea trough, the region is projected to undergo an increase of precipitation, particularly in winter with an increase between +10 and +30% (+14 and +25%). In summer, the increase is projected to be between +5 and +15% (+8 and +12%), August being the month with the smallest precipitation increase.

***WNA**, land part of Western North America (30–60°N and 130–103°W), i.e., BC, Alberta, Saskatchewan, Washington, Idaho, Montana, western part of Dakotas, Oregon, Wyoming, California, Nevada, Colorado, Arizona, New Mexico, West of Texas, and northernmost part of Mexico.*

A general warming is projected for the region, with modest seasonal variations of warming. The largest warming is projected to occur in July-August-September and January, from +3 to +5°C, and smaller warming in March-April-May and November, +2 to +4°C (DJF and JJA: +3.2 to +4.1°C). Warming is projected to be smallest near the West Coast, +2 to +3°C, and larger inland. In fact the warming over the Pacific Ocean is projected to be limited to 1 to 2°C. The contrast between land and ocean warming is expected to contribute to the amplification of the subtropical anticyclone off the West Coast of USA, which could have important consequences on coastal upwelling and marine stratus clouds. The warming could be larger in winter over elevated areas as a result of snow-albedo feedback, an effect that is poorly modelled by CGCMs due to insufficient horizontal resolution.

Averaged over the region, modest annual-mean precipitation changes are projected, with an increase in winter, 0 to +20% (0 and +15%), and a decrease in summer, –15% to 0% (–7 and 0%). The uncertainty around the projected changes is large however, as projections from different CGCMs and different SRES scenarios produce a wide range of values, and the changes do not scale well with variations in GHGs. The averages for the entire region hide important north-south differences: the north is projected to experience an increase of precipitations while the south should experience a decrease. The line of zero change is oriented more or less west-to-east, and it is expected move north and south with seasons, being at its southern most position in winter, through California, south Nevada and north Arizona, and should almost reach the northern limit of the region in summer. North of the line of zero change, increases could reach up to +15% at the extreme north in winter, while south of the line decreases should reach –20% in summer. The line of zero change is also projected to lie further to the North under SRES scenarios with larger GHG amounts.

***I**, land part of Central North America (30–50°N and 103–85°W), i.e., eastern part of the Dakotas, Minnesota, Wisconsin, Michigan, Iowa, Kansas, Missouri, Indiana, western part of Kentucky and Tennessee, Oklahoma, Arkansas, eastern Texas, Mississippi, and Alabama.*
A general warming is projected for this region, with modest seasonal variations of warming. The largest expected warming is projected to occur in July-August-September, from +3 to +5°C, and smaller warming in March-April-May, +2 to +5°C (DJF: +2.8 to +3.7°C; JJA: +3.3 to +4.6°C). Warming should be smallest near the Gulf Coast in winter, +2 to +3°C, and larger northward inland.

Averaged over the region, precipitation changes are projected to be modest with little seasonal variation, from –5 to +15% in February-March-April, (DJF: +2 and +9%), and –20% to +10% in July-August-September (JJA: –12 and +3%). The uncertainty around the projected changes is large, particularly for the summer season, as projections from different CGCMs and different SRES scenarios produce a wide range of values, and the changes do not scale well with variations in GHGs. The averages for the entire region hide important north-south differences: the north is projected to generally experience an increase of precipitations while the south is projected to experience a decrease. The line of zero change is oriented more or less west-to-east, and it is projected to move north and south with seasons, being at its southern most position in winter, around 35° North, and will almost reach the Canadian border in summer. North of the line of zero change, increases could reach up to +15% near the Great Lakes in winter, while south of the line decreases should reach –10% in the southern states in summer. The line of zero change is also projected to lie further to the North under SRES scenarios with larger GHG amounts.

***ENA**, land part of Eastern North America (25–50°N and 85–50°W), i.e., Ohio, eastern part of Kentucky and Tennessee, southern parts of Ontario and Québec, Canadian Maritimes, Island of Newfoundland, New England states southward to Florida.*
A general warming is projected for the region with little seasonal variations of warming, from +2.5 to +5°C (DJF: +3.0 and +3.8°C; JJA: +2.8 and +3.6°C). In winter, the northern part of the region is projected to warm most, up to +6°C in the central part of Ontario and Québec, while coastal areas are projected to warm by only +2 to +3°C.

Average over the region, precipitation changes are projected to vary from an increase in February-March-April, from +5 to +20% (DJF: +8 and +13%), to modest changes in July-August-September, from –5 to +10% (JJA: –2 and +5%). The uncertainty around the projected changes is large, particularly for the summer season, as projections from different CGCMs and different SRES scenarios produce a wide range of values, and the sign of the changes varies with different SRES scenarios. In winter the northern parts is expected to experience an increase of precipitation, reaching +25%, while the south should experience negligible changes. Summertime precipitations are projected to decrease under SRES scenarios with larger GHG amounts, except for the Appalachian region where a small increase is projected.

11.3.5.5 Uncertainties

The uncertainties of climate changes over North America have their roots in climate-change projections from CGCMs that need to faithfully simulate several dynamical features that control or affect the North American climate:

- The skill of AR4 CGCMs at simulating ENSO and NAO/AO, their projection under altered forcing, and their influence on North American climate, is largely unknown.
- The ocean circulation in the Hudson Bay and Canadian Archipelago is under resolved by CGCMs, and hence changes in sea-ice under altered forcing are poorly known, as is its influence on climate of surrounding regions.
- Large uncertainty remains in the decrease of the North Atlantic Thermohaline Circulation (THC) under altered forcing, and its influence on reduced warming of the northeast Canadian regions.
- Little is known on the changes in frequency and intensity of middle-latitude cyclones, although a general northward displacement of tracks is quite probable.

- Tropical cyclones are still under resolved by CGCMs, and hence changes under altered forcing with respect to the frequency, intensity and tracks of tropical disturbances making landfall in regions of southeast USA and Northern Mexico are mainly unknown.
- Owing to the coarse horizontal resolution of CGCMs, high terrain remains unresolved, which likely results in an underestimation of snow-albedo feedback in warming high elevations over western North America.
- Little is known on the dynamical consequences of the larger climate-change warming over land than over ocean, in particular for the northward displacement and intensification of the subtropical anticyclone off the West Coast of USA, and the potential consequences on 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 how all these elements can affect a substantial precipitation reduction of the southwest USA (e.g., Mote and Mantua, 2002).

Some uncertainties listed above relate not so much to documented weaknesses of AR4-generation CGCMs but rather to our current lack of knowledge of their skill at simulating these features. As the analysis of the recently completed simulations progresses, these identified uncertainties will either be lifted or confirmed.

The uncertainty associated with climate-change projections made with RCMs is much larger than desirable, despite the investments made with increasing horizontal resolution; typically grid meshes range from 36 to 55 km. A survey of recently published RCMs' current-climate simulations nested with CGCMs reveals biases in surface air temperature and precipitation that are two to three times larger than the recent simulations nested with reanalyses by several RCMs within NARCCAP (see Figure 11.3.5.3). This situation stems from a combination of several factors:

- All reported RCMs' projections were nested with TAR-generation CGCMs that exhibited larger biases than AR4-generation CGCMs.
- Several RCMs still employ physical parameterisation packages with poor performance, either because of their outdated design (e.g., "bucket" land-surface scheme) or because of their unacceptable sensitivity (e.g., deep convection in summertime).
- Often too few levels are used in the vertical (e.g., 14), sometimes with a too low uppermost computational level (e.g., 100 hPa).
- Most RCMs' projections were for short time slices, varying between 5 and 20 years in length.
- Ensemble runs are seldom performed, occasionally few (e.g., 3) runs are made with one, sometimes two, RCMs.
- RCM's projections were performed for a wide diversity of domains, periods and SRES scenarios, making difficult or impossible to compare results.

The North American Regional Climate Change Assessment Program (NARCCAP[2]) will permit to reduce some of these uncertainties, through the coordination of an ensemble of RCMs' simulations, nested with various AR4-generation CGCMs and performed under controlled experimental conditions, for a domain covering the continental USA, the southern part of Canada and the northern part of Mexico.

11.3.6 Central and South America

11.3.6.1 Key processes

A mix of tropical and extratropical processes are of importance in Central and South America. Over much of the continent, 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 Patagonia. ENSO plays a key role throughout much of the region, so a shift towards a more El-Niño-like state in the Pacific will have effects that will overlay and interact with tendencies from other sources, such as the poleward shift of the westerlies and the drier subtropics associated with increased moisture fluxes in the atmosphere.

Climate over most of Centralamerica (central-southern Mexico and Central America) is characterized by a relatively dry winter (November through April), and a well defined rainy season from May through October

[2] http://www.narccap.ucar.edu/

(Taylor and Alfaro, 2004), with a mid-summer minimum in late July and early August that has been attributed to air-sea interactions and teleconnections between the IAS and the eastern Pacific warm pool (Magaña et al., 1999; Magaña and Caetano, 2005). Easterly waves and tropical cyclones contribute a large percentage of the precipitation, particularly over northern Mexico (Douglass and Englehart, 1999). The Intra Americas Seas (IAS, i.e., Gulf of Mexico and Caribbean Sea) and the north eastern tropical Pacific are among the most active in the world for tropical cyclones. Interannual variability in precipitation depends in part on the proximity of tropical cyclones to the coast, implying that projections of changes in mean precipitation in this region will be partly dependent on difficult to model changes in tropical cyclone climatology. During the boreal winter, the atmospheric circulation over the IAS is dominated by the seasonal fluctuation of the Subtropical North Atlantic Anticyclone, with invasions of extratropical systems that affect Mexico and the western portion of the Great Antilles (Schultz et al., 1998; Romero Centeno et al., 2003). An El-Niño-like shift in the Pacific would displace these storms equatorward, tending to offset the effects of a planetary-wide polewards shift of the midlatitude sotrm tracks.

A warm season precipitation maximum, associated with the South American Monsoon System (SAMS), dominates the mean seasonal cycle of precipitation in tropical and subtropical latitudes. The pattern of Amazonian rainfall is determined by the interplay of land-surface feedbacks, topography, and incursions of drier and cooler air from midlatitudes (Garreaud, 2000; Vera and Vigliarolo, 2000). The future of the rainforest is, of course, of vital ecological importance, as well as being central to the future evolution of the global carbon cycle. The SAMS is strongly influenced by ENSO (e.g., Lau and Zhou, 2003), and thus future changes in ENSO will induce complementary changes in the region. The South Atlantic Convergence Zone (SACZ) plays an important role in precipitation over the southern Amazon towards southeast Brazil; displacements of the SACZ would have important regional impacts. There are well-defined teleconnection patterns (such as the Pacific-South American (PSA) modes (e.g., 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, such as Southern Australia and the Western Cape provinces of South Africa. Patagonia would experience an increase in precipitation from the same poleward storm track displacement.

11.3.6.2 Skill of models in simulating present climate

GCM climate simulations for the tropical regions have improved in some aspects since the TAR.. The interannual variability in precipitation is well simulated by numerical models in most of Centralamerica, except over northwestern Mexico (Koster et al 2000). Other mesoscale elements important for central America (such as tropical cyclones) have received only minimal examination in GCMs (Camargo and Sobel, 2004).

In general, simulations from models in the AR4/PCMDI archive tend to produce excessive precipitation during the Centralamerican winter (Dec-Jan-Feb) (Figure 11.3.6.1a), but tend to slightly underestimate it during part of the rainy season (Jun-Jul-Aug) (Figure 11.3.6.1b). The excess in simulated precipitation during the Centralamerican dry period may be as large as 40%, but it does not substantially affect total annual rainfall (not shown) since most precipitation in the region concentrates in the summer months (Higgins et al., 2004).

[INSERT FIGURE 11.3.6.1 HERE]

Over Centralamerica, annual mean temperatures increased by about 1°C during the 20th Century (Hulme and Sheard, 1999), and can be approximately reproduced by AOGCMs (Figure 11.3.6.2). The years 1994, 1995, 1997, 1998 and 2000 were among the warmest of the last century and the tendency for warming in the region continues but at a rate slower than the global average. Modest increases in precipitation of a few per cent have been recorded over the region in the 20th century (Hulme and Sheard, 1999), with most of the increase occurring in the summer rainy season (not shown). Such trends are not captured in AOGCM simulations.

[INSERT FIGURE 11.3.6.2 HERE]

The performance of the AR4 AOGCMs over southern America is summarized in Table 11.3.6.1. The seasonal area-averaged temperature biases range from about –3°C to 3°C in AMZ and from about –4°C to 5°C in SSA. The ensemble annual mean temperature is somewhat colder than the observations for both regions. In general most individual models exhibit a cold bias throughout the year, except in SON in AMZ and in DJF in SSA. The biases are unevenly geographically distributed within both regions (Figure 11.3.6.3). The AR4 models ensemble mean present climate simulations show a warm bias around 30°S (particularly strong 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 weak oceanic upwelling.

[INSERT FIGURE 11.3.6.3 HERE]

The multi-model scatter is considerable in the AR4 AOGCM precipitation as simulated for the current climate, ranging between –57% and 43% in AMZ and between –50% and 65% in SSA (Table 11.3.6.1). For both regions, the ensemble annual mean climate exhibits drier than observed conditions, with about 60% of the models having a negative biase. The geographical distribution of the bias (Figure 11.3.6.4) displays strong contrasts between the western coasts and the rest of the continent. Simulation of the regional climate is seriously affected by models' deficiencies at low latitudes. Both annual and seasonal mean rainfall simulations have similar systematic bias towards underestimated rainfall over the Amazon Basin. Over the adjacent oceans, the AR4 ensemble tends to depict a relatively weak ITCZ which extends southward of its observed position. Simulated subtropical climate is also adversely affected by a dry bias over most of south-eastern South America and in the SACZ region. On the contrary, rainfall along the Andes and in NE Brazil tends to be excessive in the ensemble mean.

[INSERT FIGURE 11.3.6.4 HERE]

Cavalcanti et al., 2002, Marengo et al., 2003, and Zhou and Lau, 2002 provide a coordinated and detailed analyses of precipitation and circulation biases in a set of 6 AGCMs in this region. The precipitation biases are qualitatively similar to those described above. However, the North and South Atlantic subtropical highs and the Amazonia low are too strong, while low level flow tends to be too strong during austral summer and too weak during austral winter. In DJF model simulations over-estimate the Chaco low, which is also displaces toward the central Andes, with distortions resulting both wqest and east of the Andes. The model ensemble simulations capture the PSA patterns quite well, but the Rossby wave pattern smoother than observed.

Relatively few studies using RCMs for 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) and Rojas and Seth (2003) performed seasonal integrations with emphasis on tropical South America to study two January-May periods with extreme rainfall anomalies, applying reanalyses and GCM boundary forcing. The model (RegCM) driven by reanalyses was able to simulate the different rainfall anomalies and large-scale circulations in the two periods, but it shows reduced rainfall in the western Amazon compared with observed estimates that are associated with weak low-level moisture transport from the Atlantic. The GCM-driven RegCM improves upon the monthly evolution of rainfall compared with that from the GCM, but degrades compared with the reanalyses-driven integrations. Misra et al. (2003) also performed austral summer simulations with a regional spectral model (RSM) driven by an ensemble of AGCM simulations. Relative to the AGCM, the RSM improves the ensemble mean simulation of precipitation and the lower- and upper-level tropospheric circulation over both tropical and subtropical South America and the neighbouring ocean basins. But the RSM exacerbates the dry bias over sectors of AMZ, and perpetuates the erroneous split ITCZ over both the Pacific and Atlantic Ocean basins from the AGCM. Menéndez et al. (2001) used a RCM (LAHM) driven by a stretched-grid AGCM (LMDZ) 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, in other regional modelling studies for SSA it was found that rainfall tends to be underestimated over the subtropical plains and overestimated over elevated terrain (e.g., Saulo et al., 2000; Menéndez et al., 2004; using Eta and MM5 regional models, respectively).

11.3.6.3 Climate projections

11.3.6.3.1 Temperature

Climate change projections for Central America of annual surface air temperature for the 2000–2099 period are based on AOGCMs. For the Central American region, most subregional analyses indicate an increase in surface temperature at an average rate of about 3°C by 2100 under SRES A1B scenario, (Figure11.3.6.2) 4°C under SRES A2, and 2° under SRES B1The magnitude of the positive trend in regional warming could be slightly lower over the Central American countries than over Mexico., (Figure 11.3.6.5).

[INSERT FIGURE 11.3.6.5 HERE]

Seasonally, increases in surface temperature tend to be larger (~4°C) during March, April, May, and in July and August (Figure 11.3.6.6) than during the rest of the year. Such tendency for warmer temperatures is coincident with present highest temperatures during spring and during the midsummer drought period (Magaña et al., 1999).

[INSERT FIGURE 11.3.6.6 HERE]

For South America, the mean for the AR4/PCMDI (SRES A1B) multi-model ensemble for surface temperature in AMZ and SSA is given in Table 11.3.6.2. AMZ would warm up by nearly 3.3°C on annual average, while SSA would also undergo a warming of about 2.7°C. Seasonal mean responses for individual models range between 1.7°C (in DJF and MAM) and 5.7°C (in JJA) over AMZ, and between 1.5°C (in DJF) and 4.3°C (in DJF) over SSA. Over AMZ, most (if not all) models experience an annual mean warming of at least 2°C, while only 20% of the models gives a response greater than 4°C. In SSA, 18 out of the 20 models considered project an annual mean warming between 2°C and 4°C.

The geographical distribution of the annual mean response over South America by the end of the 21st century (A1B scenario, AR4 simulations) and different measures of the confidence of such changes to occur are given in Figure 11.3.6.7. This annual mean temperature change is modulated by the seasonal cycle (not shown) differently in different regions. In SSA, the response is greater during DJF with a maximum centred over Bolivia and NW Argentina (i.e., the amplitude of the seasonal cycle tends to increase), while in AMZ the response is greater during JJA. The signal to noise ratio compares the strength of the climate change signal to the spread between models' responses. In general, signal exceeds noise in the entire region, but especially over Patagonia during SON and DJF. The minimum ratio is reached over Amazonia (in particular during SON). The region along 20°S in JJA and many coastal areas feature relatively large signal to noise ratio. About 90% (30%) of the models show a response larger than 2°C (4°C) over large areas of tropical and subtropical South America. As the mean response reduces southward, in the southern tip of the continent all but one or two models give a response below 2°C.

[INSERT FIGURE 11.3.6.7 HERE]

Hegerl et al. (2004) use daily minimum and maximum temperature in climate change simulations with two coupled climate models (CCCma and HadCM3) and discuss the change in the warmest night of the year at the time of doubling of CO_2. relative to the change in the warm season mean. Both models simulate an increase in this "warmest night temperature" larger than the mean response over the Amazon Basin but smaller than the mean response over SSA.

11.3.6.3.2 Precipitation

For central America there is a large dispersion in the AR4/PCMDI models projections , ranging from +10% to –30% for A1B by the end of the century (Figure 11.3.6.8). Most models indicate a negative trend in precipitation particularly during spring and mid summer (Figure 11.3.6.9). This tendency for larger changes (in percentage) in precipitation suggests that dry or at least relatively dry conditions will tend to be enhanced. This drying is associated with less cloudiness and reduced evapotranspiration in those months, leading to a seasonal maximum in the temperature response.

[INSERT FIGURE 11.3.6.8 HERE]

[INSERT FIGURE 11.3.6.9 HERE]

The composite pattern of precipitation change indicates a southward displacement of the eastern tropical Pacific ITCZ activity, closer to the equator than present (Figure 11.3.6.10). This pattern is reminiscent of the El Niño precipitation anomaly over the tropical eastern Pacific (Walliser and Gautier, 1992), which results in negative precipitation anomalies over most of Centralamerica (Ropelewsky and Halpert, 1989), and positive precipitation anomalies along the Caribbean coast of Central America (Magaña et al., 2003).

[INSERT FIGURE 11.3.6.10 HERE]

Of importance to the central American region are changes in tropical cyclone activity. Knutson and Tuleya (2004) showed a CO_2 induced increase in both storm intensity and near-storm precipitation rates using the output from 9 different CMIP climate models to drive a higher resolution version of the GFDL hurricane prediction system. The CO_2-Sea Surface Temperature changes, based on 80 year linear trends, ranged from about 0.8°C to 2.4°C. The aggregate results, averaged across all experiments, indicate a 14% increase in central pressure fall, a 6% increase in maximum surface wind speed and an 18% increase in average precipitation rate within 100 km of the storm centre. However, a human-forced signal in the tropical cyclone record will be extremely difficult to detect because of both the relatively modest size of the predicted changes in maximum potential intensity and the rather large natural multidecadal variability of these phenomena (Landsea et al., 1999). Therefore, the projected negative precipitation anomaly over the Americas warm pools could be affected, at least during summer, in relation to tropical cyclone activity.

Additional uncertainty also exists over northern Mexico in relation to the North American Monsoon System (NAMS). According to Arrit (2005), under the SRES A1B climate change simulations for 2070–2099 indicate little change in precipitation over the monsoon core region. However, the large ensemble spread and the inconsistent performance of the models in replicating the observed teleconnections from the NAMS limit confidence in the models' projections of climate change.

Over the Centralamerican region, projected changes in temperature consistently indicate an increase, while projected precipitation changes vary from model to model, with most of them indicating a negative trend in rainfall for the 21st century. An analysis of the projected changes in a month by month basis indicates that the largest expected changes in temperature and precipitation tend to coincide. During the boreal spring season, i.e., the driest period over most of Centralamerica, warming may be larger than during the other months. Dry conditions and high temperature in Centralamerica affect soil moisture. On the other hand, the other maxima in projected changes in temperature and precipitation that appear during July and August imply a more intense Mid Summer Drought or *Canicula.*

For South America, the areal mean annual response for the AR4 ensemble of A1B scenario simulations brings about near-zero values (lower than 1%)over both AMZ and SSA , but models responses range between –20.9% and 13.7% for AMZ, and between –11.7% and 7.0% for SSA (Table 11.3.6.2). Seasonal mean responses for individual models range between –36.9% (in JJA) and 21.3% (in SON) over AMZ, and between –20.6% and 17.8% (both in JJA) over SSA. About 70% (40%) of the models project a wetter climate in austral summer and autumn (winter and spring) in AMZ, while about 50–60% of all the models project a wetter climate in SSA all over the year.

The geographical distribution of the annual mean response (Figure 11.3.6.11) suggests that the large-area averaged response discussed in the previous paragraph hide marked regional differences. The annual mean precipitation would decrease over northern South America near the Caribbean coasts, as well as over large parts of northern Brazil, Chile and Patagonia, while it would increase in Colombia, Ecuador and Peru, around the equator and in south eastern South America. The seasonal cycle (not shown) 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.

[INSERT FIGURE 11.3.6.11 HERE]

The poleward shift of the South Pacific and South Atlantic subtropical anticyclones is a very firm 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 influence the climate of eastern South America (Robertson et al., 2003), including the SACZ and La Plata Basin regions, although the mechanisms are not so straightforward. The increase in rainfall in south eastern South America is likely related with a corresponding poleward shift of the Atlantic storm track. It was also speculated that the observed southward displacement of the subtropical Atlantic high would be related with a southward shift of the SACZ (Liebmann et al., 2004).

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 this El Niño-like response. The accompanying shift and alterations of the Walker circulation would directly affect tropical South America since the region is associated with ENSO through a pronounced Walker cell component in all seasons (Cazes Boezio et al., 2002). Moreover, any change in the tropical Pacific would affect SSA through extratropical teleconnections (Mo and Nogués-Paegle, 2001).

In general, the signal to noise ratio (Figure 11.3.6.11) is lower for precipitation than for surface temperature. The signal stands out against the noise only in relatively few regions: a few coasts of Ecuador and northern Peru, parts of south eastern South America, parts of southern Andes and Tierra del Fuego. These areas agree with the areas of models response coincidence (also in Figure11.3.6.11). The response will be better represented when and where this quantity is either large (i.e. most models project more precipitation) or small (i.e. most models project a decreasing precipitation trend). For example, about 90% of the models foresee a wetter climate near the Rio de la Plata (especially in DJF, not shown). The uncertainty is larger over parts of Bolivia and Brazil where the number of models projecting a wetter climate is similar to the number of models projecting a drier climate. However, even in the regions where relatively large consensus is reached for the response, the fact that most models are not able to reproduce the regional precipitation patterns in their control experiment with sufficient accuracy contributes to enhancement of the uncertainty.

Boulanger et al. (2005b) evaluates the AR4 models' skill in simulating the large-scale structure of late 20th century precipitation over South America. The method leads to an "optimal model combination" for 21st century climate change projections. . The precipitation responses for scenarios A1B and B1 strongly resemble the A2 trends but with weaker amplitudes.

11.3.6.4 Uncertainties

Most climate variability in the Centralamerican region is associated with ENSO (Amador et al., 2003). Current simulations of precipitation under SRES scenarios suggest a more El Niño like type of change pattern over the tropical eastern Pacific, that appear to lead to the negative anomalies in precipitation over the Americas warm pools. However, the contrast in precipitation anomalies between the Caribbean and Pacific coast of Central America may not be well captured under present simulations. Tropical cyclone activity is a key process of concern for Central America, but future changes are at present poorly projected. 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 few downscaling studies compounded by insufficient observed data over most of the region limit the capacity to develop strong regional scale statements of change. Most IPCC AR4 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 potential for abrupt changes in biogeochemical systems in AMZ remain as a source of uncertainty. Large differences in the projected climate sensitivities in the climate models incorporating these processes and lack of understanding of processes were identified (Friedlingstein et al., 2003).

The high and sharp Andes mountains is unresolved in low resolution models. The skill of IPCC AR4 models at simulating the dominant patterns of oscillations affecting South America (ENSO, SAM, PSA) and their changes under anthropogenic forcing is mostly undiagnosed. Lack of knowledge/information on the changes in extremes and in frequency and intensity of mid-latitude cyclones.

11.3.7 Australia – New Zealand

11.3.7.1 Key processes

Australia lies within the latitude range 12 to 43 degrees south, between the South-eastern Pacific and western Indian oceans. Its stretches between the tropical and mid-latitude climate zones and contains a wide range of regional climates. Key processes that influence the climate of Australia 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 imbedded disturbances. Due to its higher latitude location (34 to 46 degrees south) New Zealand is primarily influenced by only the latter two systems. Climatic variability in Australia and New Zealand is also strongly affected by the El Niño-Southern Oscillation system. In Australia, El Niño occurrences are the primary cause of major drought events, and in New Zealand rainfall and temperature patterns are affected by the swing to more south westerly winds across the Islands (McBride and Nicholls, 1983; Mullan, 1995). The influence of El Niño in the region is also modulated by the Interdecadal Pacific Oscillation (IPO) (Power et al., 1999; Salinger et al., 2002). Tropical cyclones occur in the region, and are a major source extreme rainfall and wind events in northern coastal Australian, and, more rarely, in the north island of New Zealand (Holland, 1984; Sinclair, 2002).

Tropical north-to northwest Australia lies under the influence of the monsoon and has a well-defined wet season between December and March. The tropical north-east is also monsoonal, but with substantial rains throughout the year due to disturbances in the trade winds. Tropical cyclones affect the entire northern coast of Australia. In the subtropics, the coastal zone east of the Dividing Range forms a distinct climate regime, with reasonably abundant rainfall with a summer maximum. Extreme rainfall events can (rarely) be associated with tropical cyclones in the lower latitudes, but a more common source of extreme rainfall in the region are east coast lows (Holland et al., 1987). The southern coastline of Australia forms another major zone, receiving most of its rainfall in winter (June – August) when the midlatitude westerlies and their embedded disturbances are furthest north. In the warmer months this zone lies under the influence of subtropical high pressure and tends to be dry. The tendency toward a Mediterranean climate is most marked the southwest, while in the southeast, summer rainfall is more common. The entire South coast, but especially the Southwest, is sensitive to drying caused by poleward displacement of the midlatitude storm track. The extensive arid- to semi-arid interior experiences sporadic extreme rainfall events (Roshier et al., 2001), primarily in summer and due to systems of tropical origin.

New Zealand's climate is influenced by the position of the westerlies and the accompanying subtropical high and subpolar low pressure belts, and especially disturbances embedded in the westerlies. Tropical cyclones occasionally impact the North Island (Holland, 1984; Sinclair, 2002). Rainfall patterns in New Zealand are also strongly influenced by the interaction of the predominantly westerly circulation with its very mountainous topography. For example average annual rainfalls on the western side of the Southern Alps commonly exceed 4000mm, whereas the eastern side can be less than 700mm. The interaction of variations in the atmospheric circulation with the topography of New Zealand results in complex patterns of rainfall variation from year to year. Much of the precipitation over the mountains falls as snow, but at lower elevations, snow is uncommon, particularly in the North Island. (Salinger et al., 2004; Sturman and Tapper, 1996)

Apart from the general increase in temperature that the region will share with most other parts of the globe, the particularities 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.

11.3.7.2 How well is the climate of the region currently simulated?

There are as yet relatively few studies of the quality of the AR4 global models in the Australia/New Zealand area. With regard to the circulation, reference to Chapter 8 shows that the composite model still has systematic low pressure bias near 50°S at all longitudes in the Southern hemisphere, including the Australia/NZ sector, corresponding to an equatorward displacement of the midlatitude westerlies. A study of the midlatitude storm track eddies (Yin, 2005) also indicates a consistent equatorward displacement on average. A study of current climate circulation patterns over southwest Western Australia (Hope, 2005a) found that deep winter troughs over the region were over-represented in the AR4 runs. 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 stormtrack 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 wamer 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, where the cold bias is larger than 2°C. Table 11.3.7.1 gives seasonal biases averaged over Southern and Northern Australasian regions. On large scales, the precipitation also has some systematic biases. Averaged across Northern Australian, models on average simulate 21% more precipitation than observed, but the range of biases in individual models in very large (–71 to +133%). This is discouraging with regard to confidence in many of the individual models. The average annual bias in the southern Australian region is negative 5%, and the range of biases more moderate (–58% to +35%) Inspection of the the model maps indicates that the Northwest is too wet and the Northeast and East coast too dry. The central arid zone is insufficiently arid in most models.

The Australasian simulations in the AOGCMs utilized in the TAR report have, in the intervening years, been scrutinized more closely in this region, in part as a component of 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., 2004, Walsh et al., 2004). Some high resolution regional simulations were also considered in this process, which included examination of quantitative skill scores such as RMS error and pattern correlations as well as qualitative evaluation. The general conclusion has been that the large-scale features of Australian climate are quite well simulated in nearly all current models. In winter, temperature patterns were poorer in the south where topographic variations more strongly influence the temperature patterns, although this was alleviated in the higher resolution simulations. A set of the TAR AOGCM simulations were also assessed for the New Zealand region by Mullan et al. (2001) with similar conclusions (broadscale features of mean climate captured, but with shortcomings in the detail). Our preliminary assessment of the AR4 global models is similar, but with concern about the disparate simulations of the monsoonal rainfall in the North.

There have been a number of studies that have considered the ability of AOGCMs and the CSIRO regional model DARLAM to simulate aspects of current climate variability. Mullan et al. (2001) examined AOGCM ability to represent ENSO-related variability in the Pacific. Most models adequately simulated 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 (Timball and McAvaney, 2001; Timball, 2004; Charles et al., 2004) based on broadscale 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.

11.3.7.3 Projected regional climate change

In addition to the models collected for the Fourth Assessment, numerous studies have been conducted with earlier models. Recent regional average projections are provided in Giorgi et al. (2001), Rousteenoja et al. (2003). CSIRO (1992, 1996) and Whetton et al. (1996) included assessment of subregional pattern of change, and some aspects of extremes. The most recent national climate change projections of CSIRO (2001) were based on the results of eight AOGCMs plus one higher resolution regional simulation. The methodology 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., 2004, Walsh et al., 2004, IOCI 2005). This work has focussed on temperature and precipitation, although additional variables such as potential evaporation and winds have been included in the more recent assessments.

A range of dynamically downscaled simulations 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 12 km for Tasmania (McGregor, 2004). These simulations use recent CSIRO simulations for background forcing. Downscaled projected climate change has also beeen undertaken for part of Australia recently using statistical methods (e.g., Timball and McAvaney, 2001; Charles et al., 2003; Timball, 2004; Timball and Jones, 2005).

Due its small size and complex topography, assessment of projected climate change over New Zealand has been undertaken using downscaling methods. Recent projections have used used statistical methods which used AOGCM projected changes in precipitation, temperature and sea level pressure as predictors (Mullan et al., 2001a; Ministry of the Environment, 2004).

11.3.7.3.1 Temperature

The temperature projection of the AR4 global models (comparing the period 2070–2099 in the A1B scenario to 1979–1999 in the 20C3M integrations) varies between 2 and 4.5°C (see Table 11.3.7.2), with the smaller values in the coastal regions, Tasmania, and the South Island of New Zealand, and with the largest values in Central and Northwest Australia (see Figure 11.3.7.1). The warming is larger than the surrounding oceans, but only comparable to, or slightly larger than the global mean warming. As can be seen in Table 11.3.7.2 averaging over the region south of 30°S (SAU), the mean warming among all of the models is 2.6 K (compared to a global mean warming of 2.5 K) whereas the warming averaged over the region north of 30°S (NAU) is 3.2 K. The seasonal cycle in the warming is weak, but with larger values (and larger spread amongst model projections) in summer. Across the models in the AR4 archive, 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. For example, in SAU the range is 2.0–3.9, as compare to the global mean range of 1.8–4.1, while in NAU the range is 2.3–4.5 K. The warming over the same time period in the B2, A1B, and A2 scenarios is close to the ratios of the global mean responses, and linear rescaling from one scenario to another and to different time-periods according to the magnitude of global mean warming seems well-justified.

[INSERT FIGURE 11.3.7.1 HERE]

These results are broadly (and in many details) similar to those described in earlier studies, so other aspects of these earlier studies can plausibly 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. 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 north-west. By 2070, annual average temperatures increase by 1 to 6°C over most of Australia with spatial variations similar to those for 2030. Dynamical 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).

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 assessment of changes in extreme temperatures have been based on adding a projected mean temperature change to each day of an 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 (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. Given the similarity in the AR4 model projections and the results from these earlier sets of models, we believe that these results will be similar in essence when repeated with this new set of models.

Model temperature projections are reasonably consistent with 20th century trends. All-Australian mean maximum and minimum daily temperatures have increased 0.06°C/decade 0.11°C/decade respectively since 1910 (Della-Marta et al., 2004). Models show relatively small difference between maximum and minimum temperatures trends (Whetton et al., 2002; see Chapter 9), a continuing cause for concern. Karoly and Braganza (2005) argue that part of the observed regional warming can be attributed to greenhouse gases using statistical attribution techniques. New Zealand has warmed by 0.9°C between 1900 and the 1990s (Folland et al., 2003).

11.3.7.3.2 Precipitation

[INSERT FIGURE 11.3.7.2 HERE]

Figure 11.3.7.2 shows the mean over all models in the AR4 database of the fractional change in precipitation between 2079–2099 in the A1B projections as compared to the 1970–1999 base. Also shown are the number of models (out of 20) projecting increases or decreases in precipitation. Simulated changes in precipitation averaged for the northern and Southern Australia regions are shown in Table 11.3.7.1. The most robust feature is the reduction in rainfall along the south coast in JJA and the annual mean. As may be seen in the regional averages (Table 11.3.7.1) decrease is also strongly evident in SON. 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., 2003; Miller et al., 2005; Yin, 2005; Chapter 10), but this reduction extends over land during the winter when the storm track is placed furthest equatorward. Due to the shape of the storm track, which drifts polewards as it crosses Australian longitudes, the strongest effect is in the Southwest, where the ensemble mean drying is in the 15–20% range. Hope (2005a) has shown that there is a southward shift in storm tracks in the AR4 runs over south-west Australia. 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.

[INSERT FIGURE 11.3.7.3 HERE]

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, but consistency amongst the models on this feature is not strong. In the monsoonal regime, there is a slight tendency for summer increase, except in the northwest. However consistency amongst models is weak and, as seen above, discrepancies in the current climate simulation in this region are large.

These results are broadly consistent with results published based on earlier GCM simulations. In the CSIRO (2001) projections (see Figure 11.3.7.3) 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. 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.

Compared to the GCM patterns of change, higher resolution regional modelling results for rainfall change differ in detail, particularly near the coast and in areas of more marked topography (Whetton et al., 2001; BTE, 2004). Whetton et al. (2001) demonstrated that rainfall inclusion of high resolution topography could reverse the simulated direction of rainfall change in parts of Victoria. In a region of strong rainfall decrease as simulated directly by the GCMs, two different downscaling methods (Charles et al., 2004; Timball, 2004) have been applied to obtain to characteristics of rainfall change at stations (Timball, 2004; IOCI, 2005). The downscaled results continued to show the simulated decrease, although the magnitude of the changes was moderated relative to the GCM in the Timball (2004) study. Downscaled rainfall projections for New Zealand (incorporating differing results of some six GCMs) showed a strong variation across the Islands (Ministry of the Environment, 2004). The picture that emerges is that the pattern of precipitation changes decribed above in the global simulations is still present, but with the precipitatipon 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.

There has been a marked decreasing winter rainfall trend in southwestern Australia since the 1970s (discussed in Chapters 3 and 9) which is in qualitative agreement with model projections for the 20th century (Section 9.5.3.2) and 21st century. This observed trend and has been demonstrated to be related to changes in large scale changes in circulation and moisture (Timball, 2004; Hope, 2005b; IOCI, 2005), particularly a poleward displacement of the westerlies, although there is evidence that regional land clearing may have enhanced the trend (Pitman et al., 2004). The regional circulation changes may be related to the impact on the Southern Annular Mode of the Antarctic ozone hole (Section 9.5.2.3), but that link has been established primarly for the southern summer and not the season of rainfall decline. There may also be contributions from the response to enhanced greenhouse gases in the 20th century (see Miller et al., 2005) and regional natural fluctuations (Timball et al., 2005; IOCI, 2001; Cai et al., 2005). Dry conditions in winter in southeastern Australia since the the mid –1990s (Timball and Jones, 2005) also appear to be related to similar large scale circulation changes. In recent decades New Zealand has become drier in the north of the North Island and wetter in the north, west south and south east of the South Island. This has been attributed to more frequent southwesterly flow as a consequence of a shift in the Interdecadal Pacific Oscillation (Salinger and Mullan, 1999), but it is also the pattern expected from an equatorward shift in the circulation, whether driven by the ozone hole or other mechanisms..

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). 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 Timball, 2004). Recently (Abbs, 2004) dynamically downscaled current and enhanced greenhouse sets of extreme daily rainfall occurrence in northern NSW and southern Queensland as simulated by the CSIRO GCM to a resolution of 7km. The downscaled extreme events for a range of return periods compared well with observations and the enhanced greenhouse results for 2040 showed increased 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 (Bureau of Meteorology, 1999) spatially for the case of 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 of frequency 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., 2004). Mullan et al. (2005) has shown that by 2080s in New Zealand, there may be significant increase in drought frequency in the east of both islands.

11.3.7.3.3 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). However, marked decreases on average snow water over New Zealand (60% by 2040 under the A1B scenario) have been simulated by Ghan and Shippert ( ) using a high resolution subgridscale orography in a global model.

11.3.7.3.4 Potential evaporation

Using the method of Hobbins et al. (2004) 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., 2004; Walsh et al., 2004). In all cases increases in potential evaporation were simulated, and in almost all cases the moisture balance deficit became stronger. This is strong indication of the Australian environment becoming drier under enhanced greenhouse conditions.

Roderick and Farquhar (2004) have noted that pan evaporation has decreased over recent decades at most measurement sites in Australia. This is potentially inconsistent with projected future increases in potential evaporation, and may be related to past changes in solar radiation and winds. Gifford et al. (2005) has shown that the downward trend reversed after 1996 and that historical pan evaporation variations are partly related to rainfall variability.

11.3.7.3.5 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 windspeed 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 Section 10.3.5) contributes to uncertainty with respect tropical cyclone behaviour (Walsh, 2004).

11.3.7.3.6 Winds

[INSERT FIGURE 11.3.7.4 HERE]

The ensemble mean projected change in wintertime sea level pressure is shown in Figure 11.3.7.4. Much of Australia lies to the north of the center 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. Wind changes are much more dramatic over New Zealand, where the increase in pressure gradient from the Northern to the Southern tip is roughly 2.6 mb in this A1B ensemble mean. The pressure gradient increases in every model, after averaging over each model's individual 20C3M and A1B realizations (see Figure 11.3.7.5), ranging from a minimum in CCSM3.0 (0.6mb) and FGOALSg1.0 (0.7 mb) to a maximum in GFDL-CM2.0 (5.1 mb) and ECHAM5/MPI-OM (4.8 mb) In the A2 ensemble mean, the increase is 3.4 mb. An assumption of a 60% increase, assuming no change in the variability about the mean implies a doubling of the frequency of daily wind speeds over 30 m/s (Ministry of the Environment, 2004).

A concern is that many of the models generate pressure gradients in this season that are too large, with only half the models simulating a pressure gradient within a factor of two of the observed value (roughly 4 mb from the northern to the southern tip of New Zealand). The split-jet structure and blocking activity east of Australia is difficult to simulate in models of this resolution. However, if we just average over those models with control pressure gradients that are within a factor of two of the observed, the change in the pressure drop is even larger (3.0 as opposed to 2.6 mb for A1B).

[INSERT FIGURE 11.3.7.5 HERE]

11.3.7.3.7 Storm surge

There have been relatively few studies that address the impact of climate change on storm surge and waves in the Australian region. In tropical Australia, Hardy et al. (2004) utilised storm surge and wave models to study the change to storm tide return periods at two locations on the tropical east coast of Australia, approximately 100 and 200 km north of Brisbane respectively. The climate change scenarios used were a 10% increase in the intensity of all cyclones combined with a southward shift of cyclone tracks of 1.3°, a 10% increase in frequency of tropical cyclones and a 0.3 m sea level rise. The increase in the 100 year storm tide event at both locations was around 0.45 and 0.5 m respectively with the changes dominated by the sea level rise, and the frequency changes being almost insignificant.

In eastern Bass Strait in southeast Australia, changes to storm surge return periods were determined under different climate change scenarios in McInnes et al. (2005). Scenarios of average and 95th percentile wind speed changes were determined from 13 global climate models using the method described in Whetton et al. (2005), which yielded annual low, mid, high and wintertime high changes in average wind speed of –5, +3, +10 and +14% and 95th percentile wind speed changes of –6, +3, +11 and +19% by 2070 compared with 1961 to 1990 values. Under the worst case and wintertime worst case scenarios, storm surge increases along the coastline considered increased in the range of 0.10 to 0.13 and 0.16 to 0.22 m respectively indicating that in this region, sea level rise scenarios in the range of 0.07 to 0.49 m will generally have the dominant effect.

11.3.7.4 Uncertainties

Major uncertainties concerning projected climate change for this region are:

- Uncertainty regarding the future behaviour ENSO contributes significantly to uncertainty about rainfall and drought in the region and regional tropical cyclone behaviour.
- Monsoon rainfall simulations and projections vary substantially from model to model. As a result, we have little confidence in model precipitation projections for Northern Australia. However, few models predict very large fractional changes in rainfall in this region.
- More broadly across the continent summer rainfall projections vary substantially from model to model reducing confidence in our abililty to project summer rainfall change
- To date, no detailed assessment of AR4 model performance over Australia or New Zealand is available. This means that the current range of projected changes will include the results of models that may be eventally viewed as unreliable in the region.
- Downscaled results of the AR4 simulations are not yeat available for New Zealand, but much needed because of the strong topographical control of New Zealand rainfall.

11.3.8 Polar

11.3.8.1 Arctic

11.3.8.1.1 Key processes

The Arctic climate is characterized by a distinctive complexity due to numerous nonlinear interactions between and within the different components (atmosphere, cryosphere, ocean, land) which generate a variety of internal feedbacks. Sea ice plays, through the albedo-temperature feedback and feedbacks associated with humidity and clouds, a critical role for the Arctic climate. Sea ice, ocean and atmosphere are closely coupled to each other. Examples are the following: Changes in sea ice concentrations influence the surface heat fluxes and surface albedo, both affecting the atmosphere. In return, weather systems and surface heat flux changes impact the sea ice thickness by determining the thermodynamic growth and ice dynamics. Changes in the oceanic heat transport (e.g., driven by atmospheric circulation pattern changes) affect the sea ice thickness and concentration and hence the climate sensitivity (Steele et al., 2004; Kauker et al., 2003). Strong low-frequency variability is evident in various atmosphere and ice parameters (Polyakov et al., 2003a,b), complicating the detection and attribution of Arctic changes. The natural decadal and multi-decadal variability, e.g., as possibly expressed by the warming in the 1920s–1940s (Johannessen et al., 2004; Bengtsson et al., 2004) followed by cooling until the 1960s, is in the Arctic large. In both models and observations, the interannual variability of monthly temperatures is a maximum in high latitudes (Räisänen, 2002).

Natural atmospheric modes of variability on annual and centennial time scales play an important role for the Arctic climate. Such modes include for example the NAO/AO and the North Pacific Index (see Section 3.6). The influence of NAO/AO on Arctic temperature is directly opposed in the western and eastern Arctic. A positive NAO/AO index is associated with warmer and wetter winters in northern Europe and Siberia and cooler and drier winters in western Greenland and north-eastern Canada. A positive AO index is associated with warmer temperatures in Alaska and a reduction of blocking events and the associated severe weather throughout Alaska. The North Pacific Index is a more regionally restricted signal. In its negative phase, a deeper and eastward shifted Aleutian low pressure system 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 (Stone, 1997; Curtis et al., 1998; Lynch et al., 2004) found relationships between the Alaskan and Beaufort-Chukchi region's climate and Northern Pacific variability.

11.3.8.1.2 Present climate: regional simulation skill

The above described complexity includes many processes that are still poorly understood and thus pose still a challenge for climate models (ACIA, 2005). Generally, individual GCMs show still large biases in the simulated Arctic temperature, precipitation, and sea ice. Substantial across-model scatter exists. But the evaluation of the model simulations in the Arctic generally contains a relatively high uncertainty as, except for the sea ice cover, the few available observations are sparsely distributed in space and time and the different data sets often differ considerably (Serreze and Hurst, 2000). This holds especially for the precipitation measurements with its problems in cold environments (Goodison et al., 1998; Bogdanova et al., 2002).

Few pan-Arctic atmospheric RCMs are in use. Notwithstanding their dependence on the boundary data used, they capture the geographical variation of temperature and precipitation in the Arctic more realistically than the GCMs. Further, driven by analyzed boundary conditions, RCMs tend to show smaller temperature and precipitation biases in the Arctic compared to the GCMs indicating that sea ice simulation biases and biases originating from lower latitudes contribute to the contamination of GCM results in the Arctic (Dethloff et al., 2001; Wei et al., 2002; Lynch et al., 2003; Semmler et al., 2005). However, even under a very constrained experimental RCM design, there can be considerable across-model scatter in the simulations as shown by the ARCMIP experiment (Tjernström et al., 2005; Rinke et al., 2005). 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., 2004).

Temperature

The simulated spatial patterns of the AR4 model ensemble mean temperatures agree closely with those of the observations throughout the annual cycle. Generally, the simulations are 1–2°C colder than the observations with the exception of a cold bias maximum of 6–8°C in the Barents Sea (particularly in winter and spring) caused by over-simulated sea ice in this region (Chapman and Walsh, 2005; see Chapter 8 and Figure 11.3.8.1). Compared with previous TAR models (Walsh et al., 2002), the annual temperature simulations improved in the Barents and Norwegian Seas and Sea of Okhotsk, but also worsening is noted in the central Arctic Ocean and the high terrain areas of Alaska and northwest Canada (Chapman and Walsh, 2005). Over the Arctic Ocean, the cold bias is largest (lowest) in winter (summer) (Table 11.3.8.2). The annual mean root-mean-squared error by the individual AR4 models ranges from 2°C to 7°C (Chapman and Walsh, 2005).

The mean model ensemble bias is relatively small compared to the across-model scatter (ACIA, 2005). However, difference between the coldest and warmest model is large during most of the year. Over the Arctic Ocean, the across-model scatter shows the same seasonality as the bias and is consistent with the wide range of simulated sea ice margins from autumn to spring. The across-model scatter of annual and seasonal temperatures is generally larger than the interannual variability, but the key features of the spatial patterns are similar connected with the sea ice variability. Compared with previous models, the AR4 temperatures are more (less) consistent across the models in winter (summer) (Chapman and Walsh, 2005).

There is considerable agreement between the modelled and observed interannual variability both in magnitude and spatial pattern of the variations and the seasonality of the variability is also well-simulated (Chapman and Walsh, 2005). A large subset of AR4 models are able to replicate such major warming events as occurred in the Arctic in the past (1920–1950 and 1978-present; see Chapter 8) (Wang et al., 2005).

Precipitation

The AR4 model simulated monthly precipitation varies substantially among the models throughout the year. To give one example, the simulated mean July precipitation averaged over the area north of 70°N ranges from 0.7 mm/d to 1.2 mm/d (Kattsov et al., 2005). But, the model ensemble mean is throughout the year within the range between different data sets which indicates an improvement compared to earlier overestimation (Walsh et al., 2002; ACIA, 2005). The seasonal cycle of the model ensemble mean is again in agreement with the observed climatology, but the mean precipitation is improved from autumn to spring (Kattsov et al., 2005). The ensemble mean bias varies with the season and remains greatest in spring and smallest in summer. The bias pattern (positive bias over the central Arctic and particular over the North American sector, negative bias over the north-eastern North Atlantic and eastern Arctic) persists throughout the year and can be partly attributed to coarse orography, biased atmospheric circulation (i.e., storm tracks) and sea ice cover.

The AR4 models show the same (positive) sign of the annual precipitation 20th century trend as that observed (Kattsov et al., 2005).

Sea ice and ocean

There is a considerable range of Arctic sea ice conditions in present-day AR4 simulations, particularly on the regional scale (Arzel et al., 2005; Zhang and Walsh, 2005) as in previous CMIP simulations (Flato et al., 2004; Hu et al., 2004). However, the Arctic- and multi model averaged sea ice extent and its trend are in agreement with observations. The AR4 models generally underestimate sea ice concentrations in the interior Arctic while they overestimate it in the Greenland and Barents Seas (Figure 11.3.8.1). The spatial distribution of the simulated sea ice thickness varies considerably among the models (Figure 11.3.8.3). Chapter 8 discusses these AR4 model skills in detail (see Chapter 8.3.3 for sea ice and Chapter 8.3.2 for ocean).

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, e.g., 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 hindcast simulations show a reduction in the Arctic ice volume over recent decades with an especially remarkable decline from mid-1980s to the mid-1990s and the simulated long term loss of Arctic sea ice is usually less than corresponding observational estimates (Holloway and Sou, 2002). 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 probably due to biased simulation of arctic shelf processes which differ widely in these models or biased wind forcing.

11.3.8.1.3 Climate projection

Temperature

The maximum northern high-latitude warming ("polar amplification") is consistently found in all GCM intercomparison studies (see recent review by Serreze and Francis, 2005). The simulated annual mean Arctic warming exceeds the global mean warming by 2 times in the AR4 models. Comparable magnitudes are known from previous studies (Holland and Bitz, 2003, ACIA, 2005). It is not clear whether the polar amplification signal depends on the model's resolution: some lower resolution GCMs show a larger polar amplification signal (Dixon et al., 2003; May and Roeckner, 2001), some not (Govindasamy et al., 2003). The consistency between observations and near-future (20102029) model projections (characterized by initial ice retreat and thinning) supports the concept of Arctic amplification (Serreze and Francis, 2005).

At the end of the 21st century, the annual warming in the Arctic is estimated to be 5°C (with a considerable across-model range of 2.8–7.8°C between the lowest and highest projection) by the AR4 models under the A1B scenario (Table 11.3.8.1). Larger (smaller) mean magnitudes are found for the A2 (B1) scenario with 5.6°C (3.4°C) but with a same across-model range of ~4°C. Comparable magnitudes have been found in earlier estimates (ACIA, 2005). The across-model and across-scenario variabilities in the projected temperatures are comparable.

The largest (smallest) warming is projected in autumn/winter (summer) both over ocean and land (Table 11.3.8.1, Figure 11.3.8.1). But, the seasonal amplitude of the temperature change is over ocean (7°C) much larger than over land (4°C) due the presence and melt of sea ice over the ocean in summer keeping the temperatures close to the freezing point. The Arctic Ocean region is generally warmed more than the land area (except in summer) (Table 11.3.8.1). The range between the individual simulated changes is large. For Arctic land by the end of the century, the warming ranges from 3.7°C to 9.5°C in winter, and from 1.6°C to 5.5°C in summer under A1B scenario. The across-model scatter can be attributed to the different description of the physical processes in the individual models, whereby the present-day sea ice state is one important factor. Internal variability, which is large particularly over land (Table 11.3.8.2), contributes also to the across-model differences.

[INSERT FIGURE 11.3.8.1 HERE]

The annual temperature response pattern (Figure 11.3.8.2) is characterized by a large warming over the central Arctic Ocean (5–7°C) and caused by the warming in winter and autumn associated with the reduced sea ice. The maximum warming is near the Barents Sea where the present-day model bias is also greatest. Further, a region of reduced warming (<2°C, slight cooling in several models) is projected over the northern North Atlantic which is also consistent among the models. This is caused by deep ocean mixing, weakening of the THC and reduction of heat transport into these regions (see Chapter 10.3.4) and is in agreement with earlier studies (Holland and Bitz, 2003).

[INSERT FIGURE 11.3.8.2 HERE]

Within the first half of the 21st century, the projected temperature changes do not exceed the internal variability, i.e. are not significant (Chapman and Walsh, 2005). At the end of the 21st century, the projected changes over the Arctic Ocean are clearly discernable from natural variability. However, the projected large warming over northern Alaska in winter cannot be discerned from natural variability as the simulated (and observed) temperature variability in this region is so large (Chapman and Walsh, 2005).

The regional temperature responses are largely determined by changes in the synoptic circulation patterns. The AR4 models project in winter circulation changes consistent with an increasingly positive AO (see Chapter 10.3.5.3) which corresponds to warm anomalies in Eurasia and western North America, while in summer, circulation patterns are more likely that favor warm anomalies north of Scandinavia and extending into the eastern Arctic and cold anomalies over much of Alaska (Cassano et al., 2005). But, this projected cooling is in disagreement with the recent strong warming trend in Alaska (ACIA, 2005; Hinzman et al.,

2005) indicating a decreased confidence in the summer projections (associated with the models inability to accurately simulate the present-day summer synoptic patterns).

The patterns of temperature changes simulated by RCMs are quite similar to those simulated by GCMs. However, the RCMs simulate regional structures which can be ascribed to the higher resolution and therefore often related to better topographical heights. RCMs show an increased warming along the sea ice margin due to a stronger response to sea ice changes associated with a better description of the non-linear energy cascade connected with mesoscale weather system developments. Less warming is simulated over most of the central Arctic and Siberia, particular in summer, which is due to a more realistic present-day snow pack simulation (ACIA, 2005). The warming is modulated by the topographical height, snow cover and connected albedo feedback as shown for the region of northern Canada and Alaska (Laprise et al., 2003; Plummer et al., 2005). Additionally, the regional warming pattern can be masked by temperature changes associated with changes in the large-scale circulation like changes in the NAO phase. Dorn et al. (2003) found that northern Europe and Eastern Arctic can be cooled by up to 5°C in a time slice (2039–2046) characterized by a negative NAO phase (and high GHG level) compared with an earlier time slice (2013–2020) characterized by a positive NAO and lower GHG levels.

Precipitation

The AR4 models simulate a consistent general increase in precipitation over the Arctic at the end of the 21st century (Figure 11.3.8.3). The precipitation increase is robust among the models and qualitatively well understood, attributed to the projected warming and related increased moisture convergence (ACIA, 2005; Kattsov et al., 2005). The spatial pattern of the projected change shows greatest percentage increase over the Arctic Ocean (30–40%) and smallest (and even slight decrease) over the northern North Atlantic (<5%). The correlation between the temperature and precipitation changes over the Arctic Ocean is strong and the magnitude of the precipitation response is consistent among the models (ca. 5% precipitation increase per degree warming).

[INSERT FIGURE 11.3.8.3 HERE]

By the end of the 21st century, the projected change in the annual mean Arctic precipitation varies between the lowest and highest projection from 10% to 29%, with an AR4 model ensemble mean of 19% for the A1B scenario (Table 11.3.8.1). Larger (smaller) mean magnitudes are found for the A2 (B1) scenario with 22% (13%) but with a same inter-model range. The differences between the projections for different scenarios are small in the first half of the 21st century, but increase after. However, towards the end of the 21st century, the differences between different scenarios are smaller than the across-model scatter (ACIA, 2005; Kattsov et al., 2005). For each scenario, the across-model scatter of the projections is substantial, but smaller than the across-model scatter under present-day conditions (Kattsov et al., 2005). The percentage precipitation increase is largest in winter and autumn and smallest in summer, accordingly to the projected warming (Table 11.3.8.1, Figure 11.3.8.1).

The range between the individual simulated changes is large. For Arctic land by the end of the century, the precipitation increase ranges from 13% to 44% in winter and from 3% to 21% in summer under A1B scenario (Table 11.3.8.1). The differences increase rapidly as the spatial domain becomes smaller (ACIA, 2005). To give one example, 6 AR4 models project a decrease in summer precipitation for the Ob basin, while the rest of the 14 models project an increase under the A2 scenario at the end of the 21st century (Kattsov et al., 2005). The local precipitation anomalies are determined largely by changes in the synoptic circulation patterns. During winter, the AR4 models project a decreased (increased) frequency of occurrence of strong Arctic high (Icelandic low) pressure patterns which favor precipitation increases along the Canadian west coast, southeast Alaska and North Atlantic extending into Scandinavia (Cassano et al., 2005). The regional precipitation patterns, e.g., along the North Atlantic storm track and close to complex topography and coast lines are more detailed in RCM simulations due to the higher resolution (ACIA, 2005).

The across-model scatter in the precipitation projections can be attributed to the different description of the physical processes in the individual models and to internal variability. At end of the 21st century under A1B scenario, the AR4 model averaged signal-to-noise ratio starts exceeding the factor 2 in the annual mean and in winter/autumn, and mostly over ocean (Kattsov et al., 2005), indicating that the projected increase is discernable from natural variability. However, local precipitation changes (particularly in the Atlantic sector and generally in summer) remain difficult to discern from natural variability even at the end of the 21st century (ACIA, 2005; Kattsov et al., 2005).

The following table summarizes the AR4 model ensemble mean projections for temperature and precipitation in the Arctic, and provides information on the model spread.

Extremes of temperature and precipitation. Very little work has been done in analyzing future changes in extreme events in the Arctic. Taken the values that represent the 95% of the present-day mean climate distribution, and looking at the fraction of the future distributions that are beyond it, Table 11.3.8.2 gives the chance of extreme temperature and precipitation in future AR4 model projections for the Arctic under A1B scenario. (The PDFs are calculated by the method of Tebaldi et al., 2005; see Chapter 11.2.2). A dramatic increase in the probability of extreme warm and wet seasons is likely (Table 11.3.8.2), arisen by a shift of the temperature (precipitation) distribution to warmer (wetter) values. Weisheimer and Palmer (2005) suggest a similar high (60–80%) frequency of occurrence of extreme warm winter over the Arctic Ocean, but a small (10–19%) for Alaska (ALA) and Greenland (GRL) at the end of the 21st century.

Sea ice. The Arctic sea ice is projected to decrease, both in its extent and thickness, consistently among models. The annual mean northern hemisphere sea ice extent (averaged over the AR4 models) is estimated to be reduced by 31% at the end of the 21st century under the A1B scenario (Zhang and Walsh, 2005). The reduction in the annual mean sea ice volume is about twice that (Arzel et al., 2005). The projected sea ice changes vary strongly between models, particularly at the regional scale. This scatter is largely caused by differences among the simulated present-day sea ice (see Chapter 8.3.3, Figure 11.3.8.1 and 11.3.8.3). Chapter 10 discusses the sea ice projections in detail (see Chapter 10.3.3.1 and Figures 10.3.10, 10.3.11, and 10.3.12).

Snow. Associated with the warming, the beginning of the snow accumulating season (the end of the snow melting season) is projected to be later (earlier), and the fractional snow coverage (calculated based on snow-water equivalent SWE) will decrease during the snow season. However, the projected snow coverage changes are small and of comparable or smaller order than the present-day model bias (Hosaka et al., 2005). The snow amount (SWE) is projected to increase over the Arctic northern regions (northern Siberia and North America) attributed to the increase of snowfall from autumn to winter (Hosaka et al., 2005). The regions of northern Canada and Alaska are projected by one RCM to receive more snowfall in winter due to decreased sea ice off the north coast leading to increased convective precipitation (Laprise et al., 2003; Plummer et al., 2005). Detailed information about northern hemisphere snow changes is presented in Chapter 10.

Frozen soil and permafrost. For all of the Arctic regions for which projections are available, the models (which most are off-line soil models using GCM input) predict an increase of the permafrost temperature (by 0.5°C to 2.5°C) and of the active layer depth (by 20% to >50%) by the mid of the 21st century and a zone with thawing permafrost at the end of the 21st century (ACIA, 2005). The increase of active layer depth is likely not uniform either in time nor geographically as relatively cold/warm periods associated with natural fluctuations in air temperature and precipitation are superimposed on the background warming trend. The simulated changes clearly vary among the models and the regions and depend on assumptions about soil, vegetation and snow (ACIA, 2005).

Glaciers and Greenland ice sheet. Detailed information is presented in Chapter 10. Only a small reduction in surface mass balance (SMB) is projected for the glaciated areas in the high Arctic (Svalbard, Severnaja and Novaja Semlja, Franz Josef Land, Baffin and Ellesmere Islands) due to the generally low temperatures in these areas (Van de Wal and Wild, 2001; Schneeberger et al., 2003). For the Greenland ice sheet, most of the models estimate a reduction of the SMB (Table 10.x) associated with sea level rise (see Chapter 10.6). As the GCMs poorly resolve the ice sheet due to their coarse resolution, the SMB calculations contain substantial uncertainties (Kiisholm et al., 2003; Wild et al., 2003; Huybrechts et al., 2004; see Section 10.6).

Arctic Ocean. A systematic analysis of future projections of the Arctic Ocean is still lacking due to still unsatisfactory present-day simulations. The coarse resolution is not adequate to resolve important processes in the Arctic Ocean (As example, the missing convection in the Greenland Sea prevents heat discharge of Atlantic water). The AR4 models project a reduction in the meridional overturning circulation in the Atlantic Ocean (see Section 10.3.4). Correspondingly, the northward oceanic heat transport decreases south of 60°N in the Atlantic. However at higher latitudes, the oceanic heat transport is projected to increase which might be due to stronger horizontal gyre circulations in the models (Holland and Bitz, 2003). The poleward ocean north of 60°N is generally warmed and freshened (Wu et al., 2003).

11.3.8.1.4 Uncertainties

Probability of changes. PDFs were derived by the method of Tebaldi et al. (2005) (see Section 11.2.2 for the method and its assumptions) for Arctic temperature and precipitation changes (Figure 11.3.8.4; Table 11.3.8.2). The probability that the increase in temperature (precipitation) exceeds 2°C (20%) is very unlikely in 2011–2030 (except the winter warming), increases dramatically afterwards, and is likely by the end of the century (except for summer precipitation change which is still very unlikely to exceed 20%), under the A1B scenario.

[INSERT FIGURE 11.3.8.4 HERE]

Model issues. The understanding of the Arctic climate system is still incomplete due to its complex atmosphere-ice-ocean interactions involving a lot of feedbacks. Processes which are not particularly well represented in neither, GCMs nor RCMs, are clouds, planetary boundary layer processes, and sea ice (ACIA, 2005). The Arctic Ocean and its exchanges with lower latitude seas are still particularly challenging for coupled climate models (Drange et al., 2005). Additionally, the simulations contain an implicit uncertainty based on the effects of internal nonlinear processes. In Arctic RCMs, the uncertainties in lateral and initial conditions generate strong internal model variability (Caya and Biner, 2004; Rinke et al., 2004; Wu et al., 2005). As the internal variability is large, it remains difficult to project significant temperature and precipitation changes particularly on the regional scale (Chapman and Walsh, 2005; Kattsov et al., 2005). However, the uncertainties in the projected changes by the two sources (model, scenario) are comparable.

Large-scale flow changes and natural variability. Arctic climate changes involve natural variability and major phenomena contributing to this are NAO/AO and PNA, but their projections contain distinct uncertainty. The projected NAO/AO changes are strongly model-dependent and nonlinear (Gillett et al., 2003; Osborn, 2004; see Section 10.3.5). The projection of PNA is difficult because of the uncertainty over mechanisms of mode shift, which may include internal instabilities as well as ENSO (Risbey et al., 2002). Generally, the large-amplitude natural decadal and multi-decadal climate variability impacting the Arctic may confound the detection and attribution of far-future climate changes.

11.3.8.2 Antarctic

11.3.8.2.1 Key processes

A permanent ice sheet covers the entire continent and dominates the climate of the Antarctic atmosphere. The processes that determine the distribution of the accumulation of the ice sheet are mainly the potential precipitable water content of the atmosphere and the precipitation from air masses travelling onto the continent. Sea ice cover varies greatly during the year (seasonal variation is six times greater than in the Arctic) with a maximum found during September, effectively doubling the continental area. About half of the Antarctic coast line is covered by floating ice shelves. Since they are floating changes in their mass do not alter global sea level.

The dominant factors controlling the atmospheric seasonal to interannual variability of the Southern Hemisphere (SH) extra-tropics are the SAM and ENSO (see Section 3.6) and their signature involving the Antarctic have been revealed in many studies (reviews by Carleton, 2003 and Turner, 2004). The variability of the East Antarctic climate is tied to the SAM over a large area, while that of the West Antarctic is strongly linked to the circulation variability in the South Pacific which in turn is teleconnected to the tropical Pacific during strong El Niño and La Niña events (Bromwich et al., 2000; Bertler et al., 2004). The positive phase of the SAM is associated with cold anomalies over most of the Antarctic (with the maximum in the Ross Sea area, over the East Antarctic plateau). The exception is the Antarctic Peninsula, with warm anomalies due to increased warm advection from the Southern Ocean. During El Niño periods, positive temperature anomalies are noted in the Pacific sector. Warmer (cooler) SSTs off the Ross Sea are associated with negative (positive) ENSO index, with the opposite behaviour in the other regions (Kwok and Comiso, 2002). The ENSO signal in Antarctic precipitation is still somewhat uncertain (Bromwich et al., 2000; Genthon and Cosme, 2003; Guo et al., 2004; Bromwich et al., 2004a).

11.3.8.2.2 Present climate: regional simulation skill

Major challenges still are the representation of the atmospheric conditions of the polar desert in the high interior of East Antarctica (Guo et al., 2003; Pavolonis et al., 2004) and of the precipitation patterns (Van de Berg et al., 2004). However, the evaluation of the temperature and precipitation simulations in the Antarctic contains significant uncertainty. Reanalyses and satellite monthly temperature data agree with weather station data to within 3°C (Bromwich and Fogt, 2004; Simmons et al., 2004; Comiso, 2000). Precipitation evaluation is more problematic (Connolley and Harangozo, 2001; Zou et al., 2004) as there are no reliable precipitation gauge data, few detailed snow accumulation time series, and major challenges exist in utilizing satellite observations to infer precipitation (e.g., Xie and Arkin, 1998).

On the regional scale, RCMs generally capture the large cyclonic events affecting the coast with fidelity (Adams, 2004) and the associated synoptic variability of temperature and precipitation (Bromwich et al., 2004b). Notwithstanding their dependence on the boundary data used, they capture the geographical variation of temperature and precipitation in the Antarctic more realistically than the GCMs. Further, 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; Van den Broeke and Van Lipzig, 2003; Bromwich et al., 2004c).

Temperature

The AR4 ensemble annual surface temperatures are warmer than the observations in the Southern Ocean. The bias is in the range of 2–6°C (Carril et al., 2005) which indicates a slight improvement compared to previous CMIP models (Covey et al., 2003) caused by a better simulation of the position and depth of the Antarctic trough (Carril et al., 2005; Raphael and Holland, 2005). Errors are largest over the Ross Sea and generally larger over the western than the eastern Antarctic seas (Carril, 2005). The biases over the continent are locally very different, ranging from –6°C to +6°C. A different model formulation (e.g., cloud and radiation parameterizations) has been shown to change the temperature simulation significantly (Hines et al., 2004). A lateral nudging of a GCM (getting the right synoptic cyclones from 60°S and lower latitudes) generally but not systematically brings the model in better agreement with observations (Genthon et al., 2002).

In contrast to previous TAR models (Vaughan et al., 2003), a subset of AR4 models qualitatively capture the observed enhanced warming trend over the Antarctic Peninsula in the past 50 years (Carril et al., 2005; Lynch et al., 2005). The general improvements in resolution, sea ice models and cloud-radiation packages contribute to an improved atmospheric circulation which is the key.

Precipitation

The precipitation simulation remains difficult both in GCMs and RCMs, and that on all timescales (Covey et al., 2003; Van de Berg et al., 2004; Bromwich et al., 2004b,c) as a result of model physics limitations. All atmospheric models (including reanalyses) have incomplete parameterizations of polar cloud microphysics and (clear-sky) precipitation. The across-model scatter is large in GCMs (Covey et al., 2003). The simulated precipitation depends on the simulated sea ice concentrations (Weatherly, 2004).

Sea ice

There is a considerable range of SH sea ice conditions in present-day AR4 simulations, particularly on the regional scale (Arzel et al., 2005; Holland and Raphael, 2005; Carril et al., 2005). However, the Antarctic- and multi model averaged sea ice extent is in agreement with observations, while its trend is not. The majority of AR4 models produce too little sea ice cover as known from previous CMIP models (Flato et al., 2004). The AR4 models generally overestimate the amplitude of the seasonal cycle of sea ice extent (excessive winter bias), particularly in the Amundsen and Weddell Seas (Figures 11.3.8.2). Chapter 8 discusses these AR4 model skills in detail (see Chapter 8.3.3).

11.3.8.2.3 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, and estimated to be 2.7°C (with of 1.4–4.9°C) by the AR4 models under the A1B scenario (Table 11.3.8.3, Figures 11.3.8 and 11.3.8_f). Larger (smaller) mean magnitudes are found for the A2 (B1) scenario with 3.0°C (1.8°C) but with a same inter-model range of ~2.5°C. The magnitudes are similar as in previous studies (Covey et al., 2003). Over the continent, neither the magnitude of temperature change nor the across-model scatter shows any seasonal dependency. However over ocean, the temperature change as well as the across-model scatter is largest in winter (JJA) (Table 11.3.8.3, Figure 11.3.8.6). The latter can primarily be attributed to the different sea ice simulations in the individual models (see 11.3.8.2.4 and Chapter 10.3.3).

[INSERT FIGURE 11.3.8.5 HERE]

[INSERT FIGURE 11.3.8.6 HERE]

The changes in surface air temperature do not project substantially onto any of the dominant patterns of variability. Stone et al. (2001) as well as Shindell and Schmidt (2004) report on a decoupling of the surface temperature trend from the SAM (Chapter 10.3.5). Surface temperature changes are largely determined by changes in the radiation balance rather than by altered atmospheric circulation patterns as the latter are quite small (Shindell and Schmidt, 2004). However, Lynch et al. (2005) show a systematic trend in the AR4 models towards stronger cyclonic events over the southern oceans. Connected with this, cooling of sub-Antarctic seas and warming of Antarctic Peninsula is expected in summer and winter. This regional pattern of temperature change is consistent with that observed over the past 50 years (Turner et al., 2005). The associated temperature changes are stronger in winter (JJA) than in summer.

Precipitation

The AR4 models simulate a precipitation increase at the end of the 21st century (Figure 11.3.8.3); the projected increase is robust among the models. The pattern shows greater increase over the Southern Ocean compared to the continent which is projected to be wetter by <0.25 mm/d (or 5–30%) in all seasons, under A1B scenario. The relative precipitation increase is largest (smallest) in winter (summer), but shows a considerable scatter among the individual models (Table 11.3.8.3). By the end of the 21st century, the projected change in the annual precipitation over the Antarctic continent varies from –1% to 35%, with an AR4 model ensemble mean of 14% for the A1B scenario (Table 11.3.8.3). Similar (smaller) mean magnitudes are found for the A2 (B1) scenario with 14% (9%) but with a same large inter-model range.

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 (DJF) and winter (JJA), a systematic shift towards strong cyclonic events is projected in the AR4 models. Particularly, the frequency of occurrence of deep Bellingshausen to Ross Sea cyclones is increased by 20–40% (63%) in summer (winter) by the mid of the 21st century. Related to this, the precipitation over the sub-Antarctic seas and Antarctic Peninsula are projected to increase. Associated with the reduction in strong anti-cyclonic conditions in summer (Antarctic high), anomalous low precipitation events will be reduced over the inner continent (Lynch et al., 2005).

Table 11.3.8.3 summarizes the AR4 model ensemble mean projections for temperature and precipitation in the Antarctic, and provides information on the model spread.

Extremes of temperature and precipitation

Very little work has been done in analyzing future changes in extreme events in the Antarctic. Taken the values that represent the 95% of the present-day mean climate distribution, and looking at the fraction of the future distributions that are beyond it, Table 11.3.8.4 gives the chance of extreme temperature and precipitation in future AR4 model projections for the Antarctic under the A1B scenario. (The PDFs are calculated by the method of Tebaldi et al., 2005; see Chapter 11.2.2). A dramatic increase in the probability of extreme warm (wet) seasons is likely by the mid (end) of the 21st century over Antarctica and the adjacent oceans (Table 11.3.8.4).

Sea ice. The SH sea ice is projected to decrease, both in its extent and thickness, consistently among models. The annual mean SH sea ice extent (averaged over the AR4 models) is estimated to be reduced by about 25% at the end of the 21st century under the A1B scenario (Arzel et al., 2005). The reduction in the annual mean sea ice volume is about of same order of that. The projected sea ice changes vary strongly between models, particularly at the regional scale. Chapter 10 discusses the sea ice projections in detail (see Chapter 10.4.1 and Figures 10.3.9, 10.3.10, and 10.4.1).

Antarctic ice sheet. Detailed information is presented in Chapter 10. For the Antarctic ice sheet, the models estimate an increase of the SMB (Table 10.B) contributing negatively to sea level (see Section 10.6.4). The summer temperatures are still too low to cause any significant melt, and the annual accumulation is estimated to increase due to increased temperature and atmospheric moisture as well as atmospheric circulation changes. As the GCMs poorly resolve the ice sheet due to their coarse resolution, the SMB calculations contain substantial uncertainties (Genthon and Krinner, 2001; Van Lipzig et al., 2002b; Wild et al., 2003; Huybrechts et al., 2004; see Section 10.6.4).

11.3.8.2.4 Uncertainties

Probability of changes. PDFs were derived by the method of Tebaldi et al. (2005) (see Section 11.2.2 for the method and its assumptions) for Antarctic temperature and precipitation changes (Figure 11.3.8.4; Table 11.3.8.4). The probability that the increase in precipitation exceeds 20% is very unlikely within the whole 21st century. The probability that the temperature increases more than 2°C is very unlikely within the first half of the 21st century, however likely over the Antarctic continent by the end of the 21st century, under the A1B scenario.

11.3.9 Small Islands

Climate change scenarios for small islands of the Caribbean Sea, Indian Ocean and Pacific Ocean are included in the fourth assessment for a number of reasons. The choice of islands was based of the availability of AOGCM projections for these regions. Because of their small size and orography most small islands do not generate their own climate, unlike larger landmasses that interact with the atmosphere. Furthermore even if small islands are mountainous enough to create their own climate, these interactions are not simulated on global atmospheric models, which do not have sufficiently fine resolutions to see these islands. Since models do not include atmosphere and land interaction over small islands, their simulations are given over ocean surfaces rather than over land, which is what is done for larger land masses. Many small islands are sufficiently removed from large landmasses so that atmospheric circulation may be different over the smaller islands than their larger neighbours, e.g., in the Pacific Ocean. For the Caribbean that is close to large landmasses in Central American and Northern South America, some islands share some features of Central America, while others share features of Northern South America. 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. Apart from the consideration of climatic features, most small islands have different degrees of concern about global climate change than their larger neighbours. Two such concerns are about sea level rise that threaten their way of life, and rising sea surface temperatures that affect the health of coral reefs. Finally, separating scenarios for small islands highlights the deficiencies in modelling and statistical downscaling for small islands. Very little of this is done and coupling the small islands with their larger neighbours would tend to mask these deficiencies.

In the following sections the key regional processes governing the climatology of the islands will be introduced, and the ability of the global climate models to simulate the climatology will be discusses. This will be followed by projections of temperature and precipitation taken from PCMDI models using A1B SRES emission scenarios. Because of the clear absence of regional modelling or statistical downscaling results, except for a few studies, the projections will be augmented by climate trends. Climate trends however are limited in scope because they are usually based on limited data sets, and do not necessarily reflect changes due to greenhouse gas emissions. Recent model results for tropical cyclones in the Atlantic and Pacific and trends in sea level rise will also be discussed.

11.3.9.1 Key processes

11.3.9.1.1 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 Alafro, 2005). The dominant synoptic influence is the North Atlantic subtropical high (NAH). During the winter the NAH is southernmost with strong easterly trades on its equatorial flank. Coupled with a strong trade inversion, a cold ocean and reduced atmospheric humidity, the region generally is at its driest during the winter. With the onset of the spring, the NAH moves northward, the trade wind intensity decreases and the southern 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 10ºN-20ºN latitudinal band referred to as the main development region. They represent the primary rainfall source and their onset in June and demise in November roughly coincides with the mean Caribbean rainy season. During the rainy season the rainfall is at a minimum around July in the northern Caribbean, and this relative minimum, known as Mid Summer Drought (MSD) has been attributed to air-sea interactions and teleconnections between the eastern Pacific warm pool and the Gulf of Mexico and the Caribbean Sea (Magaña et al., 1999).

For small islands, differences in size, shape, topography and orientation with respect to the trade wind influence the amount of rainfall received by the various islands. Cuba, Jamaica, Hispaniola and Puerto Rico, the larger and more mountainous islands of the Greater Antilles in the north, receive heavier rainfall at higher elevations, with a rain-shadow effect on their southern coasts that are distinctively arid. The smaller islands to the southeast tend to receive less rainfall, with Barbados and Trinidad in the South receiving more rainfall than the rest. The dry belt of the Caribbean is found over the south-western islands of the Netherlands Antilles.

Inter annual variability of the rainfall is influenced mainly by ENSO events. The late rainfall season tends to be drier in El Niño years and wetter in La Niña years and tropical cyclone activities diminish over the Caribbean diminishes during El Niño summers (Gray 1984). However the early rainfall season in the Southern Caribbean tends to be wetter in the year after an El Niño and drier in a La Niña year (Chen and Taylor, 2002; Taylor et al., 2002).

11.3.9.1.2 Indian Ocean

For climate model comparison purposes 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 by the Asian monsoons (See section 11.3.4.2.1). Around the end of September the summer monsoon, called southwest monsoon, retreats from India. The northeast monsoon then sets in the southeast Peninsula of India (about 10°N, in the neighbourhood of the Maldives). It is marked by a trough of low pressure from south Bay of Bengal to south Arabian Sea across the south Peninsula of India. This trough of low pressure very slowly slides southwards and remains close to the latitude of 7°N approximately during December to February. From March to May, this trough of low pressure again crawls back northwards and is about 10°N during May. This trough of low pressure remains a zone of cloud and precipitation throughout this period. A series of easterly waves move in its vicinity from southeast Bay of Bengal to southwest Arabian Sea. During the period of October to May, this trough of low pressure is not ITCZ since the ITCZ is to the south of the equator and the flow over this part of the Indian Ocean is from the Northern Hemisphere. The trough of low pressure to the north of the equator in the period October to May is called the Near Equatorial Trough (NET).

From October, the NET south of the equator assumes the role of the ITCZ. On the western part of the Indian Ocean (along the coast of East Africa), it moves southwards from 2°S in October to about 12°S by end of December. It remains in this extreme position up to about end of January and then starts its northward journey, slowly. By end of April, it is back to about 2°S, is about to give up its role as the ITCZ and to function again as the NET south of the equator. At this stage, the NET north of the equator assumes the role of the ITCZ, moves northwards and takes the monsoon northwards, again to India, via the Maldives (Asnani, 1993). As a consequence of the seasonal N-S characteristics of the ITCZ/NET, the likely periods for cyclones over the Maldive, and over the Seychelles are October to June

11.3.9.1.3 Pacific

This region refers to equatorial, tropical and subtropical region of the Pacific in which there is 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 intertropical convergence zone (ITCZ), the South Pacific Convergence zone (SPCZ, see Vincent, 1994), the easterly trade winds (both north and south of the equator) and the southern hemisphere high pressure belt. 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. During El Niño conditions, rainfall increases in the zone Northeast of the SPCZ (Vincent, 1994). The SPCZ extends from the ICTZ near the equator due north of New Zealand south-eastward to at least 21°S, 130°W. 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. For example Nadi on the eastern end of Viti Levu (Fiji) has only a third of the winter rainfall of Suva on the eastern end of the island.

11.3.9.2 Skill of models in simulating present climate

The ability of AOGCM's to simulate present climate in the Caribbean, Indian Ocean and North and South Pacific Ocean are summarized in Table 11.3.9.1, which give the average, minimun and maximum biases of the PCMDI models in simulating present day temperature and precipitation (1979–1998) on a seasonal and annual basis. The annual values will be discussed in detail below. For PCMDI model results the regions are defined by the following coordinates:

Caribbean: 10°N to 25°N and 85°W to 60°W
Indian Ocean: 35°S to 17.5°N and 50°E to 100°E
Northern Pacific Ocean: 0° to 40°N and 150°E to 120°W
Southern Pacific: 0° to 55°S and 150°E to 80°W

11.3.9.2.1 Caribbean

Simulations of the annual Caribbean temperature in the 20th century (1979–1998) by PCMDI models give an average that agrees closely with climatology, differing by approximately 0.1°C. The deviations of individual the models from the climatology ranged from –1.2°C (-4%) to +1.5°C (+5%). Thus the models have good skill in simulating temperature.

Global Climate Models approximately simulate 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 MSD (Magaña and Caetano, 2005). This is reflected in the PCMDI simulations, the average of which underestimate the mean precipitation by approximately 30%. The deviation in simulations of precipitation by individual models ranges from –64% to +20%, which is greater than the deviation in temperature simulations. Santer ( ) presented similar conclusions for the simulations from CMIP2 project.

11.3.9.2.2 Indian Ocean

For annual temperature in the Indian Ocean in the 20th century (1979–1998), the mean value of the PCMDI model outputs overestimated the climatology by 0.7°C, with values ranging from –0.3°C to 2.0°C. For rainfall the PCMDI consensus was only slightly below the mean precipitation by 3%, and the model deviations ranged from –22% to +20%. Thus the models have better skill in simulating present climate for the Indian Ocean than for the Caribbean.

11.3.9.2.3 Pacific

Climate model simulation 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 pattern of temperature and precipitation across the region. The AOGCM performance at simulating precipitation patterns was more variable in the models considered. 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. Rainfall amounts varied between the models, with some significantly underestimating or overestimating the intensity of rainfall in the high rainfall zones.

Analysis of the PCMDI simulations show that the average model value overestimated the mean annual temperature from 1979–1998 by 0.8°C over a Southern Pacific region, with deviations ranging from –0.1°C to 2.1°C. Over the North Pacific, the consensus temperature simulation for the period of 1979-1998 was only 0.5°C above the climatology, with model deviations from climatology ranging from –0.5°C to 1.3°C. Average precipitation was overestimated by 10% with values ranging from –8% to 31% in the southern Pacific region, whereas in the north Pacific the mean model output for precipitation differed from climatology by only –2%. The individual models deviated from –13% to 12%.

On a smaller scale, Lal et al. (submitted) have used the stretched grid C-CAM model nested in NCEP analyses to simulate the current climate of the Fiji at a horizontal resolution of 10 km. They were able to model seasonal cycles of temperature and precipitation realistically, and were also able to reproduce climatic contrasts between the western and eastern ends of Viti Levu.

11.3.9.3 Temperature and precipitation projections

Projections of temperature and precipitation changes from 1979–1988 to 2079–2098 are summarized in Table 11.3.9.2, which gives the average, minimum and maximum changes that are simulated by the PCMDI models on a seasonal and annual basis using the SRES A1B scenario. The annual values will be discussed in detail below.

11.3.9.3.1 Caribbean

Figure 11.3.9.1 summarizes the temperature and precipitation change scenarios for the Caribbean at the end of the 21st century (2079–2098) simulated by PCMDI models using A1B emission scenarios. The models displayed temperature increases ranging from 1.2 to 3.1°C with an average increase of 2°C. Statistical downscaling of HadCM3 results using A2 and B2 greenhouse gas emission scenarios gives around 2°C rise in temperature by 2080's, approximately the same as the HadCM3 model (Chen et al., 2004). Thus there was agreement between the AOGCM and the downscaling analysis. The downscaling was performed with the use of the SDSM model developed by Wilby et al. (2002)

Figure 11.3.9.1 shows most models giving decreases in precipitation and a few giving increases. The changes in precipitation range from –37% to +11%, with an average of –12%. The model results gave greater decreases in the summer than at other times. However this is around the time of the mid-summer drought (MSD), which models do not simulate well. The uncertainty in the precipitation scenario was emphasized when the HadCM3 results were downscaled for A2 and B2 emission scenarios using SDSM, since the statistical downscaling projected an increase of approximately 2 mm per day in annual precipitation by the 2080's, while the HadCM3 gives decreases in precipitation by lesser amounts. Thus there is more consistency in the temperature results than in the precipitation results. There were no regional modelling results available.

[INSERT FIGURE 11.3.9.1 HERE.]

11.3.9.3.2 Indian Ocean

Figure 11.3.9.2 gives the temperature and precipitation change scenarios for the Indian Ocean at the end of the 21st century (2079–2098) as simulated by the PCMDI models using A1B emission scenarios. Based on model consensus the annual temperature will increase by about 2.1°C and the precipitation by 4%. The individual models showed temperature increases ranging from 1.3 to 3.6°C. The precipitation changes for individual models varied from –2% to 20%. No regional modelling or downscaling result was available. (See also Section 11.3.4.2.3, Future Projections for South Asia)

[INSERT FIGURE 11.3.9.2 HERE.]

11.3.9.3.3 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).Simulated mean precipitation change shows a more variable pattern. Across the region as a whole the pattern is mixed with both increases and decreases simulated (Ruosteenoja et al., 2003; Lal, 2004). However the GCM simulations analysed by Jones et al. (2000, 2002), and Whetton and Suppiah (2003) showed a pattern of rainfall increases in the northeast over northern Polynesia (up to 30% per degree of global warming), but much less change and possible decrease in other regions (Micronesia, Melanesia and South Polynesia).

The scenarios from the PCMDI models using A1B emission scenarios for the period 2079 to 2098 show an average increase in temperature of 1.8ºC and a precipitation increase of 3% over the South Pacific (Figure 11.3.9.3). The individual model values for temperature and precipitation vary respectively from 1.2°C to 3.0°C and –4% to +11%. Over the North Pacific, the simulations give an average increase in temperature of 2.2°C, with values ranging from 1.4°C to 3.7°C. (Figure 11.3.9.4) For the same period precipitation increases when averaged over all models was 6%, with individual models giving values from 0% to 16% increases. Most of these increases were in the latter half of the year.

[INSERT FIGURE 11.3.9.3 HERE]

[INSERT FIGURE 11.3.9.4 HERE]

Figure 11.3.9.5 illustrates the spatial distribution of average DJF and JJA rainfall change 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.

[INSERT FIGURE 11.3.9.5 HERE]

Change in rainfall variability in the South Pacific has not been examined using recent simulations (but see Jones et al., 2000). However, this will be strongly driven by changes to ENSO, but this is not well understood (see Sections 10.3.5).

11.3.9.4 Climate trends

11.3.9.4.1 Caribbean

Based on analysis of data from 1950's to 2000, Peterson and Taylor et al. (2002) deduced that the climate of the Caribbean is changing. Analysis of linear regression slopes significant at 1% showed that the percent of time that maximum and minimum temperature observations were at or above the 90th percentile is increasing, and the corresponding percentage at or below the 10th percentile is decreasing. They concluded that the number of very warm days and nights is increasing dramatically and the number of very cool days and nights are decreasing, while at the same time the extreme inter-annual temperature range is decreasing. Defining a dry day as one where precipitation is less than 1 mm, they also showed, from linear regression slopes significant at 1%, that the annual maximum number of consecutive dry days is decreasing. They also found that the greatest 5-day total of rainfall, a measure of extreme precipitation, is increasing. However because of the short sampling period the trends could be sensitive to the sampling period. The data were analyzed at a Caribbean Regional Climate Change Workshop held in Jamaica in January 2001 where participants from 18 of the 21 meteorological services in the region brought daily data with them for analysis.

11.3.9.4.2 Indian Ocean

As part of a workshop held in Casablanca, Morocco, similar to the workshop described above in Section 11.3.9.4.1, data from the Seychelles were used to calculate long term trends in a number of climate extreme indices (Easterling et al., 2003). The trend in all the temperature indices showed warming. The percentage of time where 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. Trends in the contribution of heaviest 5-day rainfall to the total, and trends in the percentage of annual total rainfall, due to events equal to or greater than the 95th percentile showed increases, indicating that extreme rainfall seemed to increase.

11.3.9.4.3 Pacific

Trends in extreme daily temperature and rainfall have been analyzed from 1961 to 1998 for Southeast Asia and the South Pacific (Manton et al., 2001; Griffiths et al., 2003). Significant increases were detected in the annual number of hot days and warm nights, with significant decreases in the annual number of cool days and colds nights. Almost all stations exhibited increases in the frequency of hot extremes and decrease in cold extremes, with many of these trends being statistically significant. Mean rainfall showed an increasing trend in and north-east of the SPCZ. Extreme rainfall trends were less spatially coherent, with some stations showing increases in the proportion of annual rainfall from extreme events and some showing decrease in the number of rain days. However because of the short sampling period the trends could be sensitive to the sampling period. Folland et al (2003) showed that the annual and seasonal ocean surface and island air temperatures have increased by 0.6 to 1.0°C since near 1910 throughout a large part of the South Pacific southwest of the South Pacific convergence zone (SPCZ). To the northeast of the SPCZ, decadal increases of 0.3°C to 0.5°C in annual temperature are only widely seen since 1970, preceded by some cooling after 1940, which is the beginning of the record. Objective estimates show that estimates of uncertainty in SST are quite wide in the earlier decades of the record.

A recent paper by Griffiths et al. ( ) shows that, compared to the trends in hot days over the period 1961–1998 (Manton et al., 2001), the spatial pattern for an expanded period from 1961–2003 remain consistent., and over the five additional years the number of stations with ‘significant’ hot day trends has increased. The largest increases in hot day were located in the South pacific islands from the Solomon Islands to Fiji, and also Papette in the French Polynesia. The updated trends in warm nights was also spatially consistent with those seen over the period 1961–1998. The biggest trend in increased warm night frequency was seen from the Solomon Islands to Papeete. The updated trends in decreases in cool days and cold nights were also consistent with the previous period. Nearly all stations experience increase in minimum (Tmin) and maximum (Tmax) temperatures. There were also positive correlations between Tmax and the frequency of hot days, with very strong correlations in Papua New Guinea, Fiji, the Solomon Islands and French Polynesia. The correlation between warm nights and Tmin was consistently strong in the tropical Pacific Ocean and South East Asia. Conversely, negative correlations between cool days and Tmax, and between cold nights and Tmin were observed. Based on these correlations the authors suggest that temperature mean may be a useful predictor of changes in extreme climate.

11.3.9.5 Sea level rise

Church et al. (2004) used TOPEX/Poseidon altimeter data to estimate global empirical orthogonal functions which were then combined with historical tide gauge to estimate monthly distributions of large-scale sea level variability and change over the period 1950–2000. The best estimate of the rate of global averaged sea level rise was 1.8 ± 0.3 mm yr^{-1}. There was a maximum rate of rise in the northeastern Indian Ocean. A maximum was also observed in the central to eastern off-equatorial Pacific, spreading north and south to higher latitudes around the subtropical gyres of the Pacific Ocean near 90°E, mostly between 2 and 2.5 mm yr^{-1} but peaking at over 3 mm yr^{-1}. This maximum was split by a minimum rate of rise, less than 1.5 mm yr^{-1}, along the equator in the eastern Pacific linking to the western Pacific just west of 180°. The rise in the Caribbean appears to be near the mean. In a more recent paper (Church et al., submitted), the estimated rate of sea level rise in the Maldives over the period 1950–2001 was close to 1 mm yr^{-1} (see also Chapter 5).

11.3.9.6 Tropical cyclones

There have been a number of recent regional model-based studies of changes in tropical cyclone behaviour in the southeast Pacific (e.g., Walsh and Katzfey, 2000; Ngyuen and Walsh, 2001; Walsh and Ryan, 2000; Walsh et al., 2004; and see Walsh, 2004) which examined aspects of number, tracks and intensities. Using the DARLAM regional model, Nguyen and Walsh (2001) simulated a decrease in the frequency of tropical cyclone numbers in the south Pacific, but did show some poleward extension in their occurrence. Walsh et al 2004 obtained for $3 \times CO_2$ condition, a 56% increase in storms of maximum windspeed of greater than 30 m s^{-1}. However, in general Walsh (2004) concluded that 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 (ref to chapter 11) contributes to uncertainty with respect tropical cyclone behaviour (Walsh, 2004; Chapter 10).

One of the more recent studies on the impact of CO_2-induced warming on simulated hurricane intensity and precipitation in tropical basins (Knutson and Tuleya, 2004) supports the notion that, after about a century of climate warming in response to greenhouse gases, the upper limits on tropical cyclone intensity will be altered so as to allow for tropical cyclones with greater precipitation rates and higher intensity. However such induced increases are unlikely to be detected in present climate since the study employed sea surface temperature increases ranging from 0.8°C to 2.4°C (over a period of 80 years with CO_2 increasing at 1% per year compounded), while smaller SST changes have been observed over the last 50 years. Additionally variability in recent hurricane activity in the Atlantic can be explained in terms of natural variability (Gray et al., 1997).

[START OF BOX 11.3]

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., CO_2, radiation) and environmental factors, such as soil types. In some mountain regions, it has been shown that there is an elevation dependency on temperature trends and anomalies (Giorgi et al., 1997), a feature that is not, however, systematically observed in other 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, 2003). Recent simulations have incorporated mountain regions within larger domains of integration (e.g., the Alps or the Scandes in Europe), thereby enabling some measure of climatic change in mountains. High-resolution RCM simulations (5-km and 1-km scales) are used for specific investigations of processes such as surface runoff, infiltration, and evaporation (e.g., Arnell, 1999; Bergström et al., 2001), extreme events such as precipitation (Frei et al., 1998), and damaging wind storms (Goyette et al., 2003, but these simulations are too costly to operate in a "climate mode".

Projections of changes in precipitation patterns in mountains are tenuous in most climate models 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), they are still not well predicted by climate models.

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 apparently minor changes in temperature. As warming progresses 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 rise by about 150 m. 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 a strong emissions scenario (the IPCC SRES A2 emissions future), snow duration may be reduced by 50% at altitudes 2000 m to 95% at levels below 1000 m. Where some models predict an increase in wintertime precipitation, this increase does not compensate for the change 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, 2004), 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 hydrology are likely to have significant repercussions not only in the mountains themselves but also in 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., Graham et al, 2005).

[END OF BOX 11.3]

[START OF BOX 11.4]

Box 11.4: 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 time averaged sea level over the next century and beyond. Time averaged 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 either independently of, or by substantially enhancing, the time averaged changes. There is insufficient reliable 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. Therefore evaluating changes to the current threat invariably requires firstly quantifying the hazard presently posed by sea level extremes.

Detecting changes in observed records of extreme sea level is difficult because long records comprising high frequency measurements are needed but are sparse. 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.

Several recent studies have attempted to simulate extreme water levels for the present day and future climates for a limited number of sites. At sites where there are observations a present day simulation provides a means of validating model results.

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 11.2 for general details of these).

The dynamical approach is to use the results from global climate model simulations to drive higher resolution models over a limited region of interest. 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. These are then used to drive a storm surge model, again limited in extent to the region of interest (e.g., Lowe et al., 2001). This sequence of one-way coupled models is usually carried out for a present day

(Debenard et al., 2002) 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., 2002).

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 developed from the present day to the future large-scale atmospheric synoptic conditions simulated by a global climate model (e.g., von Storch and Reichardt, 1997).

In the stochastic sampling method synoptic weather events responsible for extreme sea levels are identified and the key characteristics (intensity and movement) are represented by frequency distributions which can be randomly sampled to generate a population of severe weather events. For each event simple models, such as cyclone vortex models in the case of tropical cyclones, are used to generate the surface wind and pressure fields and these are applied to the storm surge model (i.e. as with the dynamical approach above, e.g., Hubbert and McInnes, 1999). Frequency distributions are modified to represent changes under enhanced greenhouse conditions to determine storm surge characteristics under enhanced greenhouse climates. These changes can be derived from analysis of results of the dynamical techniques, e.g., by sampling from available storm surge simulations, and in this way related back to the large scale changes provided by global climate models.

The above approaches all have particular strengths and weaknesses. The major advantage of the dynamical approach is that it attempts to physically model the processes which may lead to changes in extreme level. Thus it does not make use of statistical relationships between large scale synoptic conditions and local storm surges derived from historic conditions which may change in the future. The major disadvantage is the computational complexity which means that simulation periods may be too short to adequately sample extreme behaviour. The statistical approach has the advantage that it is computationally less expensive and, when observations are employed, can account for very fine scale local behaviour. However the assumption that the statistical relationships are constant over time may not be valid, for instance, if there are large shifts in the tracks of storms. The major advantage of the stochastic method is that, within a given climate, it is straightforward to generate results representing hundreds of years and to describe well the distribution of extremes. The major disadvantages are that it may be difficult to capture the full range of synoptic forcing using simple models and it is not obvious how the frequency distribution should be changed in a future climate.

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. (2003) 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. An extreme value distribution was fitted to the cyclone intensity, cyclone size was assumed constant and cyclones were selected either to cross the coast non-preferentially between 16°S and 17°S or travel parallel to it. Relative frequencies of the events were calculated from the observations with an average of one every five years.

Cyclone intensity distribution was 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. In this study, analysing 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).

2. Europe
A number of recent predictions of climate driven changes in extreme water levels on the European shelf region have been carried out using the dynamic method. Woth et al. (2005) analysed changes in storm surges along the North Sea coasts, forcing a hydrodynamic storm surge model with pressure and wind data from four of the HadAM3H A2 scenario driven PRUDENCE simulations. They found up to a 20–30 cm increase in the 99.5th percentile of sea surface height (above the average sea level change) from 1961–1990 to 2071–2100 along the eastern coasts of the North Sea, but no change at the east coast of the UK. Using the Hadley Centre regional model (HadRM3H) driven HadAM3H to drive a storm surge model and including the effects of global mean sea level rise and vertical land movements, Lowe and Gregory (2005) found that increases in extreme sea level are positive around the entire UK coastline, with the largest rise in the Thames Estuary (Box 11.4, Figure 1). Meier (2005) used a Baltic Sea ocean model driven by data from four RCM simulations to study storm surges in the Baltic Sea. The simulations gave varying results but suggested a possibility of large changes, one of them indicating the 100-year surge in the Gulf of Riga to increase 41 cm more than the average sea level.

[INSERT BOX 11.4, 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 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.

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. Furthermore, the climate change can not be related easily to any particular emissions or socio economic scenario.

Lowe ( ) used 40 years of simulation from the HadCM2 model, downscaled to 50 km using HadRM2, to drive a 10km barotropic storm surge model. The first 20 year time slice represented present day conditions and the second period 2040–2060 conditions. A second future simulation was made with an increase in mean sea level plus some vertical land movement taken from observations. The simulated changes in storminess lead to a change in extreme water levels though not significantly different compared with natural variability. When the mean sea level rise and vertical land movement are included the changes in extreme water level are outside those expected by natural variability alone.

Uncertainty

At present, we can not reliably quantify the range of uncertainty in estimates of future coastal flooding as only a limited set of models have been used to assess these. At best we can make crude estimates of the minimum values of the uncertainty ranges (Lowe and Gregory, 2005a).

[END OF BOX 11.4]

[START OF QUESTION 11.1]

Question 11.1: How Useful are Regional Scale Projections?

Short Answer: Regional climate change is a direct function of global change affecting the regional atmospheric circulation, compounded with changes in local scale processes from land use change and other feedback mechanisms responding to the globally forced change. To the extent these aspects are understood and incorporated in the analysis methods, the regional projections are valuable. At present the robust statements of regional projections are based on consensus between GCMs (providing broad regional messages), and more detailed analyses through empirical and dynamical downscaling techniques, as well as interpretation of projected changes in large-scale processes relevant to the regions. In general the regional statements are a combination of multiple sources of information. Although dependent on region and variable, the messages of regional change are thus viable for adaptation and response strategies. See Box 11.1 for details

[END OF QUESTION 11.1]

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Tables

Table 11.2.1. Methods for generating probabilistic information from future climate simulations at continental and sub-continental scales, SRES – scenario specific.

	Input Type			Methodological Assumptions	
Reference	**Experiment**	**Spatial Scale**	**Time Resolution**	**Synthesis Method and Results**	**Model Performance Evaluation**
Furrer et al. (2005)	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 (Bias and Convergence) implicitly brought to bear through likelihood assumptions
Giorgi and Mearns (2003)	Multimodel Ensemble	Regional averages (Giorgi and Francisco)	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.
Greene et al. (2005)	Multimodel Ensemble	Regional averages (Giorgi and Francisco)	Annual (seasonal and year-average) time series, smoothed to extract low frequency trend.	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). Coefficients estimates applied to future simulations. PDFs at regional level	Model performance evaluated through R-square statistics, and "best models" chosen a-priori to enter the regression model.
Raisanen (2005)	Multimodel Ensemble	Grid points (after interpolation to common grid)	Seasonal multidecadal averages	Non-parametric quantiles estimation. Models are assumed independent. Information from all grid points is pooled across space. PDFs at the gridpoint level independently derived, not accounting for spatial dependency	Either no model performance evaluation (all models contribute equally to the quantile estimation) or "bad" models discarded a priori as a sensitivity test.
Tebaldi et al. (2004, 2005)	Multimodel Ensemble	Regional averages (Giorgi and Francisco)	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

Stott et al. (2005)	Single Model (HADCM3)	Continental averages	Original integration (HADCM3)	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 onto as additional uncertainty component. PDFs at the continental scale level	Not applicable
Harris et al. (2005)	Perturbed Physics Ensemble	Grid points	Original integration (EBM)	Simple (linear) pattern scaling applied to bridge equilibrium response of slab-models in the PPE (climate feedback parameter and spatial patterns) and time-dependent response under transient climate change scenarios from EBM. PDFs at arbitrary level of aggregation	No model performance evaluation.

Table 11.3.2.1.

		Temperature (°C)			Precipitation (%)		
		Mean	Min	Max	Mean	Min	Max
SAH	DJF	3.3	2.2	4.9	–16		
	MAM	3.5	2.3	5.1	–18		
	JJA	4.0	2.6	5.8	–3		
	SON	3.8	2.7	5.4	4		
	Ann	3.7	2.6	5.3	–5		
WAF	DJF	3.2	2.3	4.5	6	–16	24
	MAM	3.3	1.7	4.6	0	–9	11
	JJA	3.2	1.5	4.6	1	–17	16
	SON	3.2	1.9	4.8	4	–11	15
	Ann	3.2	1.8	4.7	2	–9	13
EAF	DJF	3.0	2.0	4.2	12	–4	34
	MAM	3.1	1.7	4.4	6	–11	19
	JJA	3.2	1.7	4.3	2	–18	16
	SON	3.1	1.9	4.3	9	–10	36
	Ann	3.1	1.8	4.3	7	–4	24
SAF	DJF	3.1	1.8	4.7	0	–7	9
	MAM	3.3	1.8	4.5	–3	–23	11
	JJA	3.3	1.9	4.7	–21	–44	–2
	SON	3.6	2.1	5.0	–16	–40	3
	Ann	3.3	1.9	4.8	–4	–12	5

Table 11.3.3.1. Biases in present-day (1979–1998) temperature and precipitation in NEU (land 10°W–40°E, 48–75°N) and SEU (land 10°W–40°E, 30–48°N) in the AR4 AOGCM simulations.

		Temperature (°C)			Precipitation (% of observed)		
		Mean	Min	Max	Mean	Min	Max
NEU	DJF	–3.9	–22.7 (–6.3)[a]	1.1	26	–5	69
	MAM	–2.9	–11.5 (–5.1)[a]	1.0	25	–13	54
	JJA	-0.6	–3.4	3.0	–10	–58 (–39)[a]	15
	SON	–2.3	–10.0 (–4.9)[a]	1.2	10	–11	36
	Ann	–2.4	–11.0 (–4.1)[a]	1.6	10	–18	30
SEU	DJF	–1.9	–5.0	1.2	8	–10	65 (36)[b]
	MAM	–1.3	–3.3	0.8	11	–21	78 (42)[b]
	JJA	–0.1	–2.7	3.9	7	–53	65
	SON	–2.1	–4.2	0.2	–7	–32	30
	Ann	–1.4	–3.3	0.9	4	–21	59 (27)[b]

Notes:
(a) Excluding iap_fgoals
(b) Excluding giss_aom

Table 11.3.3.2. Simulated area mean temperature and precipitation changes from 1979–1998 to 2079–2098 in NEU (land 10°W–40°E, 48°–75°N) and SEU (land 10°W–40°E, 30°–48°N) under the SRES A1B scenario.

		Temperature (°C)			Precipitation (%)		
		Mean	Min	Max	Mean	Min	Max
NEU	DJF	4.7	2.4	7.9 (6.7)[a]	17	10	25
	MAM	3.4	2.1	5.4	11	0	19
	JJA	3.0	1.4	5.0	0	–20	17
	SON	3.3	1.8	5.3	7	–3	14
	Ann	3.6	2.3	5.2	9	0	17
SEU	DJF	2.8	1.5	4.5	–7	–16	5
	MAM	3.1	1.7	4.4	–13	–25	–2
	JJA	4.3	2.6	6.5	–25	–53	–2
	SON	3.4	2.1	5.1	–12	–27	–1
	Ann	3.4	2.0	5.0	–13	–26	–3

Notes:
(a) Excluding one model (iap_fgoals) with a very large cold bias in 1979–1998.

Table 11.3.4.1. Biases in present-day (1979–1998) temperature and precipitation in the Asian regions in the AR4 AOGCM simulations.

		Temperature (°C)			Precipitation (% of observed)		
		Mean	Min	Max	Mean	Min	Max
Northern Asia NAS	DJF	–2.1	–9.9	2.1	20	–16	92
	MAM	–2.6	–6.2	0	58	2	110
	JJA	–0.7	–4.5	2.3	14	–38	61
	SON	–2.2	–6.6	1.3	20	–14	48
	Ann	–2.0	–5.7	0.8	24	–11	54
Central Asia CAS	DJF	–1.4	–4.9	3.1	18	–34	76
	MAM	–1.2	–4.2	1.9	22	–37	78
	JJA	0.4	–4.5	6.0	–16	–71	62
	SON	–1.6	–4.6	1.5	1	–49	45
	Ann	–1.0	–3.9	2.1	10	–45	50
Tibetan Plateau TIB	DJF	–3.4	–10.3	1.3	220	13	666
	MAM	–4.0	–7.8	–0.1	226	132	482
	JJA	–2.2	–7.4	0.9	46	3	147
	SON	–3.4	–6.6	–0.7	155	69	331
	Ann	–3.2	–6.0	–0.4	119	50	242
Eastern Asia EAS	DJF	–3.1	–6.6	1.8	56	–23	138
	MAM	–1.8	–5.2	0.4	49	0	106
	JJA	–1.0	–3.8	0.8	7	–15	27
	SON	–2.5	–5.9	–0.3	17	–17	76
	Ann	–2.1	–5.3	0.3	23	–7	60
Southern Asia SAS	DJF	–1.7	–6.7	2.8	33	–28	123
	MAM	0.3	–4.6	3.6	3	–46	73
	JJA	0.4	–2.0	2.6	–11	–70	28
	SON	–0.9	–4.7	3.6	2	–25	42
	Ann	–0.5	–4.2	3.2	–4	–49	33
Southeast Asia SEA	DJF	–2.0	–3.6	-0.1	6	–37	51
	MAM	–1.2	–3.2	0.4	13	–30	62
	JJA	–1.5	–3.1	0.4	6	–28	47
	SON	–1.7	–3.3	0.3	5	–36	53
	Ann	–1.6	–3.1	0.2	8	–27	45

Table 11.3.4.2. Simulated mean temperature and precipitation changes from 1979–1998 to 2079–2098 in the Asian regions under the A1B scenario. Mean is the mean change averaged over the 20 AR4 models, while Min/Max are the slightest/largest changes by individual model ensembles.

		Temperature (°C)			Precipitation (%)		
		Mean	Min	Max	Mean	Min	Max
Northern Asia NAS	DJF	5.6	2.9	8.6	28	13	55
	MAM	3.9	2.0	6.9	18	2	27
	JJA	3.5	2.0	5.6	8	–1	15
	SON	4.7	2.8	6.9	17	8	29
	Ann	4.4	2.7	6.4	15	10	24
Central Asia CAS	DJF	3.4	2.2	5.1	4	–10	23
	MAM	3.7	2.3	4.8	–10	–24	4
	JJA	4.3	2.7	5.7	–17	–59	20
	SON	3.7	2.5	4.9	2	–18	26
	Ann	3.8	2.6	5.1	–4	–19	6
Tibetan Plateau TIB	DJF	4.3	3.0	6.8	19	1	34
	MAM	3.8	2.5	6.2	10	–2	34
	JJA	3.9	2.7	5.4	5	–11	27
	SON	3.9	2.9	6.3	6	–8	22
	Ann	4.0	2.8	6.1	9	–2	28
Eastern Asia EAS	DJF	3.8	2.1	5.2	13	–5	42
	MAM	3.3	2.2	4.6	10	0	20
	JJA	3.3	2.0	5.0	8	–1	17
	SON	3.4	2.4	5.0	6	–12	27
	Ann	3.4	2.4	4.9	9	2	19
Southern Asia SAS	DJF	3.5	2.7	4.7	–6	–36	15
	MAM	3.4	2.1	5.2	5	–31	34
	JJA	2.8	1.2	4.4	10	–4	23
	SON	3.0	2.0	4.4	12	–13	26
	Ann	3.2	2.0	4.7	8	–16	20
Southeast Asia SEA	DJF	2.6	1.6	3.6	6	–4	13
	MAM	2.7	1.5	3.9	6	–4	18
	JJA	2.5	1.5	3.8	6	–3	17
	SON	2.5	1.6	3.6	6	–4	23
	Ann	2.6	1.5	3.7	6	–3	15

Table 11.3.6.1. Biases in present-day (1979-1998) temperature and precipitation in AMZ and SSA in the AR4 AOGCM simulations. Between brackets, number of models with negative and positive biases.

Model Bias: (1979–1998) - Obs.		Temperature (deg)			Precipitation (%)		
		Mean	Min	Max	Mean	Min	Max
AMZ	DJF	–0.7	–1.7 (18)	2.0 (2)	–2.2	–33.7 (10)	31.4 (10)
	MAM	–0.9	–1.8 (18)	1.6 (2)	–11.8	–28.6 (16)	10.7 (4)
	JJA	–0.8	–3.1 (13)	0.7 (7)	–22.1	–56.4 (17)	43.3 (3)
	SON	0.5	–1.6 (10)	2.9 (10)	1.2	–56.7 (8)	37.3 (12)
	ANN	–0.5	–1.7 (15)	1.8 (5)	–7.7	–30.6 (12)	25.5 (8)
SSA	DJF	0.6	–1.2 (8)	4.8 (12)	2.4	–42.4 (8)	41.6 (12)
	MAM	–0.3	–1.8 (15)	3.3 (5)	–14.4	–49.9 (16)	11.2 (4)
	JJA	–1.4	–3.7 (19)	1.3 (1)	3.8	–29.4 (10)	64.8 (10)
	SON	–0.2	–2.8 (11)	2.2 (9)	–1.4	–43.8 (11)	51.6 (9)
	ANN	–0.3	–1.9 (13)	2.9 (7)	–2.8	–38.0 (12)	32.2 (8)

Table 11.3.6.2. Simulated area mean temperature and precipitation response (2079–2098 minus 1979–1998) in AMZ and SSA under the SRES A1B scenario in the AR4 AOGCM simulations. The number of models with area mean response greater than 2°C and 4°C and the number of models with wetter climate is also given.

Scenario Response:		Temperature (deg)					Precipitation (%)			
A1B (2079–2098) minus 20C3M (1979–1998)		Mean	Min	Max	Nb. Models with ΔT>2	Nb. Models with ΔT>4	Mean	Min	Max	Nb. Models with ΔP>0
AMZ	DJF	3.0	1.7	4.6	19	2	4.5	–13.3	17.2	13
	MAM	3.0	1.7	4.6	19	2	1.8	–12.9	14.3	14
	JJA	3.4	2.0	5.7	19	4	–4.7	–36.9	14.4	7
	SON	3.5	1.8	5.3	19	5	–2.8	–33.6	21.3	8
	ANN	3.3	1.8	4.8	19	4	0.9	–20.9	13.7	11
SSA	DJF	2.8	1.5	4.3	19	2	1.8	–16.0	10.0	11
	MAM	2.7	1.8	4.1	18	1	0.7	–10.2	7.4	12
	JJA	2.5	1.7	3.6	16	0	–1.5	–20.6	17.8	10
	SON	2.7	1.7	3.9	18	0	–2.0	–20.4	10.9	12
	ANN	2.7	1.7	3.9	18	0	0.3	–11.7	7.0	13

Table 11.3.7.1. Biases in present-day (1979–1998) temperature and precipitation in NAU and SAU in the AR4 AOGCM simulations.

		Temperature (°C)			Precipitation (% of observed)		
		Mean	Min	Max	Mean	Min	Max
NAU	DJF	–0.1	–1.9	2.5	22	–78	124
	MAM	–0.2	–2.9	2.4	11	–61	107
	JJA	1.0	–4.6	2.7	19	–43	169
	SON	0.3	–2.0	3.7	41	–86	233
	Ann	–0.2	–2.3	2.8	21	–71	133
SAU	DJF	0.9	–1.6	4.1	24	–53	70
	MAM	–0.5	–2.5	3.5	–7	–53	39
	JJA	–2.2	–5.3	0.5	–17	–61	30
	SON	0.1	–4.0	2.5	–13	–67	53
	Ann	–0.4	–3.1	2.6	–5	–58	35

Table 11.3.7.2. Simulated area mean temperature and precipitation changes from 1979–1998 to 2079–2098 in NAU and SAU under the SRES A1B scenario. For precipitation change, the number of simulations (out of 20) showing increase is also given.

		Temperature (°C)			Precipitation (%)			
		Mean	Min	Max	Mean	Min	Max	No. incr
NAU	DJF	3.1	2.2	4.5	0	–22	22	9
	MAM	3.1	2.1	4.2	3	–28	44	10
	JJA	3.0	2.0	4.3	–13	–54	37	5
	SON	3.3	2.5	5.0	–16	–46	21	5
	Ann	3.2	2.3	4.5	–2	–26	24	9
SAU	DJF	2.8	2.0	4.2	–1	–22	29	8
	MAM	2.5	2.0	3.8	–1	–29	38	8
	JJA	2.3	1.7	3.5	–12	–37	11	3
	SON	2.8	2.0	4.1	–17	-40	8	2
	Ann	2.6	2.0	3.9	–7	–27	11	6

Table 11.3.8.1. Simulated mean temperature and precipitation changes from 1979–1998 to 2079–2098 in the Arctic (averaged over 60–90°N) under the A1B scenario. Mean is the mean change averaged over the 20 AR4 models, while Min/Max are the slightest/largest changes by individual model ensembles.

		Temperature (°C)			Precipitation (%)		
		Mean	Min	Max	Mean	Min	Max
Arctic land	DJF	6.4	3.7	9.5	28.6	13.2	43.5
	MAM	4.1	2.3	7.0	19.1	10.2	34.5
	JJA	3.0	1.6	5.5	12.5	3.4	20.7
	SON	5.1	2.8	7.2	22.8	13.0	33.9
	Ann	4.7	2.8	7.0	19.4	12.1	28.9
Arctic land+ocean	DJF	7.1	4.3	11.4	24.5	11.2	38.2
	MAM	4.4	2.3	7.3	17.5	8.3	32.1
	JJA	2.5	1.2	5.3	12.6	3.8	20.4
	SON	5.9	2.9	8.9	22.2	9.5	31.0
	Ann	5.0	2.8	7.8	18.8	9.9	28.6

Table 11.3.8.2. Present-day bias, natural variability and probability of projected changes in Arctic temperature and precipitation.

	Present-Day		Climate Change Under A1B Scenario			
	Bias[a]	Natural variability[b]	Median 2080–2099	IQR[c] 2080–2099	% chance of exceeding threshold[d]	% chance of increase of extremes[e]
Temperature	K	K				
Arctic land DJF	3.1	0.98 (4.38)	7.2	0.9	0.48/1/1	1/1/1
Arctic land JJA	3.9	0.44 (1.97)	3.7	1.6	0/0.72/0.95	1/1/1
Arctic land+ocean DJF	4.7	0.66 (2.95)	8.2	1.1	0.83/1/1	1/1/1
Arctic land+ocean JJA	2.8	0.34 (1.52)	3.3	1.6	0/0.62/0.92	1/1/1
Precipitation	mm/d	mm/d				
Arctic land DJF	0.05	0.06 (0.27)	25.8	3.8	0/0.03/0.98	0.94/1/1
Arctic land JJA	0.05	0.08 (0.36)	13.2	2.4	0/0/0	0.92/1/1
Arctic land+ocean DJF	0.2	0.08 (0.36)	26.0	5.9	0/0.14/0.92	0.95/1/1
Arctic land+ocean JJA	0.03	0.08 (0.36)	13.2	1.9	0/0/0	0.9/1/1

Notes:
(a) "20C3M AR4 model mean minus observation", based on period 1980–1999. Used "observations" are ERA40 re-analysis data.
(b) Natural variability of the 20-year means computed on the basis of the time series of seasonal observed values based on the 1980–1999 observations. The inter-annual variability values are in parenthesis.
(c) IQR=interquartile range =range of 25–75%=variability of distribution
(d) chance of exceeding 2 degrees temperature or 20% precipitation increase; for 3 time slices 2011–2030/2046–2065/2080–2099
(e) Method: Taking the values that represent the 95% of the current mean climate distribution, and looking at the fraction of the future distributions that are beyond it. A value of 0.3 means a 30% chance of exceeding the 95th quantile of current climate distribution; for the 3 time slices 2011–2030/2046–2065/2080–2099.

Table 11.3.8.3. Simulated mean temperature and precipitation changes from 1979–1998 to 2079–2098 in the Antarctic (averaged over 60–90°S) under the A1B scenario. Mean is the mean change averaged over the 20 AR4 models, while Min/Max are the slightest/largest changes by individual model ensembles.

		Temperature (°C)			Precipitation (%)		
		Mean	Min	Max	Mean	Min	Max
Antarctic land	DJF	2.5	0.9	4.6	9.1	–10.6	30.2
	MAM	2.7	1.3	5.4	14.0	1.5	39.4
	JJA	2.9	1.4	5.0	18.6	5.1	39.0
	SON	2.5	1.2	4.7	13.5	–2.4	35.2
	Ann	2.6	1.4	4.9	13.8	–1.4	35.0
Antarctic land+ocean	DJF	1.7	0.5	3.6	10.9	–0.8	19.6
	MAM	2.4	0.9	4.8	13.7	6.3	26.6
	JJA	3.0	1.4	5.9	14.8	7.3	31.6
	SON	2.2	0.8	4.6	11.9	5.4	25.1
	Ann	2.3	0.8	4.7	13.0	6.2	25.5

Table 11.3.8.4. Present-day bias, natural variability and probability of projected changes in Antarctic temperature and precipitation.

	Present-Day		Climate Change Under A1B Scenario			
	Bias[a]	Natural variability[b]	Median 2080–2099	IQR[c] 2080–2099	% chance of exceeding threshold[d]	% chance of increase of extremes[e]
Temperature	K	K				
Antarctic land DJF	–0.4	1.05 (4.70)	2.3	0.2	0/0/0.97	0.88/1/1
Antarctic land JJA	–2.7	1.29 (5.77)	2.9	0.2	0/0.39/1	0.87/1/1
Anarctic land+ocean DJF	1.6	0.49 (2.19)	1.7	0.2	0/0/0.02	0.99/1/1
Antarctic land+ocean JJA	4.1	0.65 (2.91)	2.8	0.5	0/0.05/0.97	0.91/1/1
Precipitation	mm/d	mm/d				
Anarctic land DJF	–0.26	0.04 (0.18)	10.9	3.6	0/0/0	0.61/0.92/0.99
Antarctic land JJA	0.38	0.06 (0.27)	20.1	3.0	0/0/0.51	0.79/1/1
Antarctic land+ocean DJF	–0.21	0.05 (0.22)	5.1	3.1	0/0/0	0.33/0.77/0.97
Antarctic land+ocean JJA	0.29	0.06 (0.27)	12.4	2.7	0/0/0	0.78/1/1

Notes:
(a) "20C3M AR4 model mean minus observation", based on period 1980–1999. Used observations are AVHRR derived surface temperatures (Comiso, 2000) and hindcast PolarMM5 simulations for precipitation (Bromwich et al., 2005). For Antarctic temperature, period is 1982–2001.
(b) Natural variability of the 20-year means computed on the basis of the time series of seasonal observed values based on the 1980–1999 observations. The inter-annual variability values are in parenthesis.
(c) IQR=interquartile range =range of 25–75%=variability of distribution
(d) chance of exceeding 2 degrees temperature or 20% precipitation increase; for 3 time slices 2011–2030/2046–2065/2080–2099
(e) Method: Taking the values that represent the 95% of the current mean climate distribution, and looking at the fraction of the future distributions that are beyond it. A value of 0.3 means a 30% chance of exceeding the 95th quantile of current climate distribution; for the 3 time slices 2011–2030/2046–2065/2080–2099.

Table 11.3.9.1. Biases in present day (1979–1998) temperature and precipitation for the Caribbean, Indian Ocean and North and South Pacific Ocean in PCMDI AOGCM simulations.

		Temperature (°C)			Precipitation (% of observed)		
		Mean	Min	Max	Mean	Min	Max
CAR	DJF	0.6	–0.8	2.0	0	–43	131
	MAM	–0.3	–1.8	1.2	–43	–76	10
	JJA	–0.4	–1.8	1.0	–36	–76	44
	SON	0.4	–1.0	2.1	–28	–64	25
	Ann	0.1	–1.2	1.5	–30	–64	20
IND	DJF	0.6	–0.3	1.7	4	–23	39
	MAM	0.7	–0.4	1.8	–7	–31	27
	JJA	0.8	–0.2	2.5	–5	–30	15
	SON	0.7	–0.3	2.1	–3	–26	31
	Ann	0.7	–0.3	2.0	–3	–22	20
NPAC	DJF	0.8	–0.1	1.8	–2	–15	13
	MAM	0.3	–0.6	1.3	–11	–28	10
	JJA	0.2	–1.0	1.0	6	–15	31
	SON	0.8	–0.3	1.7	0	–12	17
	Ann	0.5	–0.5	1.3	–2	–13	12
SPAC	DJF	0.8	–0.3	2.6	1	–25	31
	MAM	1.2	0.1	2.8	13	–6	28
	JJA	1.1	0.1	2.5	17	1	44
	SON	0.5	–0.5	1.8	8	–17	38
	Ann	0.8	–0.1	2.1	10	–8	31

Table 11.3.9.2. AR4 Simulations of temperature and precipitation changes from 1979–1988 to 2079–2098 under the SRES A1B scenario. For precipitation changes, the number of simulations (out of 20) showing increase is also given in the last column.

		Temperature (°C)			Precipitation (%)			
		Mean	Min	Max	Mean	Min	Max	No. incr
CAR	DJF	2.1	1.4	3.2	–6	–20	11	5
	MAM	2.1	1.3	3.3	–12	–27	6	3
	JJA	2.1	1.3	3.2	–20	–56	8	2
	SON	2.2	1.6	3.4	–8	–37	18	7
	Ann	2.1	1.4	3.2	–12	–37	11	3
IND	DJF	2.2	1.4	3.8	5	–4	20	17
	MAM	2.3	1.4	3.7	6	1	20	20
	JJA	2.2	1.4	3.7	3	–3	19	13
	SON	2.1	1.4	3.5	4	–5	20	17
	Ann	2.2	1.4	3.7	4	–2	20	17
NPAC	DJF	2.3	1.5	3.6	4	–3	16	16
	MAM	2.2	1.4	3.5	1	–18	17	13
	JJA	2.4	1.4	3.9	10	2	23	20
	SON	2.4	1.6	3.9	8	1	20	20
	Ann	2.3	1.5	3.7	6	0	16	20
SPAC	DJF	1.9	1.3	3.2	4	–7	14	17
	MAM	1.9	1.3	3.2	6	–3	17	19
	JJA	1.9	1.4	3.1	3	–2	13	17
	SON	1.8	1.4	3.0	0	–8	5	13
	Ann	1.9	1.3	3.1	3	–4	11	17