Chapter 11: Regional Climate Projections

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Executive Summary

Regional climate change projections presented here are primarily based on four information sources (although not of equal weight in each region): global atmosphere-ocean climate models; downscaling techniques used to enhance regional detail; our level of physical understanding of the factors controlling regional responses; and recent climate change.

Global climate models remain the primary source of regional information on the range of possible future climates. Although some model deficiencies persist, a clearer picture of the robust aspects of regional climate change is emerging due to steady improvement in model resolution, the simulation of processes of importance for regional change, and the expanding set of model results available.

Downscaling methods have matured since the IPCC WGI Third Assessment Report (IPCC, 2001) (hereafter TAR) and have been more widely applied. However, systematic downscaling studies remain limited. In some regions, large-scale coordination of multi-model downscaling climate change simulations has been achieved. Research on the co-ordinated multi-model downscaling studies lags that of equivalent GCM studies, and it remains an ongoing activity to develop probabilistic information on the distribution of possible climate responses and the sources of uncertainty, including the sensitivity to the global model input.

The growing insight into key physical processes that underlie regional climate responses increases confidence in the robust aspects of the model projections. A number of important themes have emerged:

- Warming generally increases precipitation gradients, and contributes to a reduction of rainfall in the subtropics and an increase in higher latitudes. Regions of large uncertainty in the precipitation response are often associated with boundaries between regions of robust increases and decreases, as there is little agreement between models on the accurate location of these boundaries.
- The poleward expansion of the subtropical highs, combined with the general tendency towards subtropical reduction in precipitation, creates especially robust projections of a reduction in precipitation on the poleward edges of the subtropics. Most of the regional projections of reductions in precipitation in the 21st century are associated with the land areas adjacent to these subtropical highs.
- Monsoonal circulations tend to weaken and yet result in increased precipitation, while the pattern of warming over the tropical oceans exerts strong control on precipitation changes within the tropics.

Previous chapters describe observed climate change on regional scales (Chapter 3) and compare model simulations with these changes (Chapter 9). In general, these comparisons are more useful for temperature than for precipitation, due to the smaller signal to noise ratio for the latter. For precipitation change there is a greater dependency on assessing model convergence in both global and downscaling models along with physical insights. Where there is lack of model convergence, further research into sources of model deficiencies is clearly needed before any robust conclusions can be reached. This lack of convergence especially in the tropics is highlighted, as the impacts of climate change may be large. Where there is near unanimity among models with good supporting physical arguments, as is typically the case for middle and higher latitudes, these factors encourage strong statements as to the likelihood of a regional climate change. However, these must be carefully weighed against the small sample of models, the lack of true independence among the models, and the absence, in many cases, of clear observational verification that this change is already occurring.

Within the limits of the available evidence, the summary likelihood statements on projected regional climate are as follows:

- *Temperature projections:* These are comparable in magnitude to those of the TAR, however the confidence in the regional projections is now higher due to larger number and variety of simulations, improved models, a better understanding of the role of model deficiencies, and more detailed 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:* Overall patterns of change are comparable to those of TAR, with greater confidence in the projections for some regions. The regions for which the model projections are robust are now more clearly defined. For some regions there are grounds for stating the projected

precipitation changes as likely or possibly even very likely. For other regions confidence in the projected change is weak, even in terms of the direction of precipitation change.

- *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 supplies a higher level of confidence. Notable improvements in confidence relate to the regional statements concerning heat waves, heavy precipitation, and droughts, while changes in storminess seem highly dependent on detailed regional changes in atmospheric circulation, where detailed convergence between AOGCMs is still lacking.

It is very likely that the following changes will occur within this century:

- *Africa:* decreases in annual rainfall in portions of Northern Sahara and the Mediterranean coast, and in winter rainfall for regions of south western Africa.
- *Mediterranean and Europe:* In northern Europe, winter minimum temperatures increase more than mean temperatures; Higher than average increase for the highest temperatures in Southern Europe. Annual precipitation increases in most of northern Europe and decrease in most of the Mediterranean area; Extremes of daily precipitation increasing in northern Europe; A decrease in the annual number of precipitation days is in the Mediterranean area; A decrease in snow season and depth.
- *Asia:* Warming well above the global mean in Central Asia, Tibetan Plateau and Northern Asia, above the global mean in East Asia and South Asia, and similar to the global mean in Southeast Asia. Heat waves / hot spells in summer of longer duration, more intense, and more frequent in East Asia, and fewer very cold days in East Asia and South Asia. Winter precipitation increases in Northern Asia, East Asia and the Tibetan Plateau, with increases in the return frequency of intense precipitation events in parts of South Asia, East Asia, and Southeast Asia.
- *North America:* Increases in lowest winter temperatures higher than the increase in average winter temperature in northern North America; increases in annual precipitation in northern parts of North America with decreases in the length of the snow season and snow depth.
- *Central and South America:* Decreases in annual precipitation along the southern Andes and increase in summer in south eastern South America.
- *Australia - New Zealand:* An increase in rainfall in the west of the South Island of New Zealand and increase in drought frequency in the east; Increased frequency of extreme high daily temperatures, decrease in the frequency of cold extremes, and increase in the frequency of extreme precipitation; Increased risk of drought in southern areas of Australia.
- *Polar:* Arctic warming for most areas, with the annual mean warming clearly exceeding the warming of the global mean; Annual Arctic precipitation increases; Arctic sea ice decreases in its extent and thickness. For the Antarctic, sea ice cover decreases more slowly, and temperature increases more slowly, than in the Arctic.
- *Small Islands:* Islands in regions of enhanced sea level rise to be vulnerable to coastal erosion and flooding. Models indicate that sea level rise during the 21st century will not be geographically uniform; sea level rise is projected to be larger than average in the Arctic, and in a pronounced but narrow band stretching across the southern Atlantic, Indian and Pacific Oceans.

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 observed current and future projected global-scale change, and are aimed at identifying actions that can be taken by multiple nations or regions. In contrast, adaptation decisions and actions tend to be more of a local and regional scale issue, and are limited by the measure of confidence in the projected changes over smaller spatial scales. It is at regional scales that the need for credible information on probable climate change and the associated uncertainties is the greatest. The possible consequences of climate change within some regions may also motivate countries to commit to and argue for further mitigation practises.

In view of this clear need, much effort has been expended in recent years on developing regional projections. Global Climate Models (GCMs) only provide information at the scale they are able to resolve, at best, but important aspects of model performance in many regions of the World rely on details related to the treatment of processes at the unresolved scales, Therefore, alternative methods have been developed to derive detailed regional information at finer scales than that resolved by GCMs. Through nested Regional Climate Models (RCMs) or empirical downscaling, these developments have generated new ways to assess important regional processes central to climate change. However, to date, much of the work remains at the level of methodological development. Downscaled climate change projections that are tailored to the needs of the impacts community, and which are based on projections across different forcing GCMs, are only starting to become more available.

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 a new effort compared to previous assessment reports, most of the effort was spent on assessing the strength and weaknesses of these methods, building on illustrative examples chosen from various geographical locations. At the time only limited efforts had been made to analyse regional climate change projections in a coordinated fashion, so the actual projections assessed were limited. The central results regarding projected changes in seasonal temperature and precipitation were almost entirely based on analysis of 9 coarse resolution AOGCMs which had performed transient experiments over the period 1960–2100 with the specifications for the A2 and B2 emission scenarios. However, in contrast to previous IPCC reports where only broad continental-scale regions were assessed, 23 sub-continental regions were considered.

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 operating at substantially higher resolution than AOGCMs consistently improve the spatial details of the simulated climate. Likewise statistical downscaling of AOGCM simulations was assessed to provide enhanced performance for many applications.

The assessment in the TAR was that it is very likely all land areas will warm more than the global average (with the exception of Southeast Asia and South America in JJA), with amplification at high latitudes. The following changes in precipitation were assessed to be likely: an increase over northern mid-latitude regions in winter and over high latitude regions in both winter and summer; in DJF, an increase in tropical Africa, little change in Southeast Asia, and a decrease in Central America; an increase or little change in JJA over South Asia and a decrease over Australia and the Mediterranean region. The TAR also warned that studies with regional models indicate that changes at finer scales may be substantially different in magnitude from these large sub-continental findings.

Information available for assessment regarding climate variability and extremes at the regional scale was too sparse for it to be meaningful to draw it together in a systematic manner. However, some statements of a more generic nature could be made, but with somewhat lower confidence than for the changes in the mean. For example it was assessed that the variability of 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; 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

The climate of a region is determined by the interaction between regional forcings and atmospheric and oceanic circulations that occur at many spatial scales, and for a 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, snow, sea ice, and ocean current distribution. Moreover, teleconnection patterns such as those associated with El Nino Southern Oscillation (ENSO) and North Atlatnic Oscillation (NAO) can strongly influence climate variability and the regional climate responses to forcing. The difficulties related to the simulation of regional climate and climate change are therefore quite apparent. 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

As GCMs have steadily improved their general performance (e.g., Chapter 8), many of them have been run for an increasing range of forcing scenarios (e.g., Chapter 10) and much more attention is being paid to the regional climate change response of these models. The 21-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 now also exist high resolution time-slice studies with uncoupled atmospheric models, ranging up to the 20 km resolution. Although coordinated multi-model experiments are needed to optimize the value of these high resolution studies for general assessments, these studies are promising, for example, as an approach towards convincing simulations of the climatology of tropical cyclones (e.g., May et al. 2004a; Mizuta et al. 2005).

11.1.3.2 RCMs

While most of the RCM work assessed in the TAR consisted of simulations of limited duration (months to a decade), experiments with RCMs of 20–30 year duration have become standard for many groups around the world (e.g., Christensen et al., 2002; Leung et al., 2004; Plummer et al., 2006). 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). The need for comparative studies using different RCMs to downscale climate change information from GCMs has also been emphasized by the scientific community. Christensen et al. (2001) with later updates by Rummukainen et al. (2003) combined the information from four RCM climate change experiments for Scandinavia, and demonstrated that it is feasible to explore not only uncertainties related to projections in the mean climate state, but also for higher order statistics.

In the European initiative PRUDENCE (Christensen et al., 2002; 2006) as many as 10 RCMs were applied to explore the uncertainties in regional climate change projections. This enabled some rough quantitative estimates to be made regarding the sources of uncertainty in regional climate change projection generation (Rowell 2005; Deque et al., 2005, 2006; Frei et al. 2005a; Graham et al. 2006; Beniston et al., 2006).

Another significant change compared to TAR is that many RCMs have been adjusted to operate at 20 km or finer horizontal scales (e.g., Leung et al., 2003,2004; Christensen & Christensen, 2004; Kleinn et al. 2005; Kurihara et al. 2005; Yasunaga et al. 2006). This development follows naturally from that of numerical weather prediction; where many centres on an operational basis apply non-hydrostatic regional models with less than 5 km inter grid distance. Figure 11.1.1 demonstrates that in order to depict essential geographical details in the precipitation patterns in the Alps, inter grid distances below 20 km may be desirable.

[INSERT FIGURE 11.1.1 HERE]

Coupled modelling is the norm in global climate modelling. Steps towards coupled modelling have also been taken in regional climate modelling since TAR (Bailey and Lynch 2000; Bailey et al. 2004; Döscher et al., 2002; Rummukainen et al., 2004; Schrum et al., 2003; Sasaki et al. 2005). 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, regional oceans and ocean current distribution (e.g., Döscher and Meier, 2004; Meier et al., 2004; Sato, 2005).

A few RCMs have been applied in full transient experiments, throughout the whole 21st Century (i.e. Whetton et al., 2000; Kwon et al., 2003; Kjellström et al., 2006). Transient RCM-runs help in 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 distinctive 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, Chapter 10, Appendix 10.4) has greatly expanded in the subsequent years. Empirical techniques are 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., stream flow or aquatic ecosystems; Cannon and Whitfield, 2002; Blenckner and Chen, 2003). There has been little development of coherent multi-technique research programmes assessing the relative merits of different empirical techniques, however, with the European STARDEX (Goodess et al., 2006) and MICE (Hanson et al., 2006) initiatives offering new contributions.

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 (i.e., Wilby et al., 2002a; Salathe, 2003, or Mehrotra et al., 2004). There remains, however, much downscaling work that goes unreported, where it is implemented for the pragmatic purpose of serving a project need, rather than explicitly for use by a broader scientific community. This is especially the case in developing nations. In some cases this work is only found within the soft literature, for example, the AIACC project (http://www.aiaccproject.org/), which supports impact studies in developing nations.

11.2 Assessment of Regional-Climate Projection Methods

11.2.1 Methods for Generating Regional-Climate Information

Coupled Global Climate Models (CGCMs) constitute the primary tool for simulating the global climate system, and to study the processes responsible for maintaining the general circulation and natural variability (see Chapter 8), and its response to external forcing (see Chapter 10). Because of their significant complexity and the need to integrate these models for many centuries horizontal resolutions of the atmospheric components of the CGCMs in the AR4 range from 400 km to 125 km.

The process of regional-scale climate-change assessment begins of necessity with an evaluation of the ability of CGCMs to simulate the current climate. Contrary to numerical weather predictions where spread in an ensemble of forecasts is to a large extent the result of natural variability and predictability limits, the spread in climate-change projections across members of an ensemble of CGCMs reflects also different responses of individual models to a prescribed forcing. There is no established practice on how to best weight individual model results in an ensemble (see Chapter 10, Section 10.5). Several different approaches to weighting have

[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).

been proposed and these are discussed in Section 11.2.2. While some responses are robust in CGCMs simulations, for others the spread is large, particularly at regional scales. A large spread may be linked to regions with important feedbacks; it does not mean that this information cannot be used, but simply that there are large associated uncertainties. Convergence (small spread) in an ensemble of projections does not, of necessity, guaranty reliability (small uncertainty) of the projected climate changes. Because in general a climate-change projection from a single model provides no sense of associated uncertainties, such projections are of little practical use in an assessment. However, the response in a simulation acknowledging all known forcing within the last century can be validated against the observed response of the climate system and this way constrain the likelihood of future climate change projection (Stott et al., 2006, see also Section 11.2.2). Information about the spread in CGCMs' projections for each of the regions is presented in Sections 11.3.2-11.3.9.

11.2.1.1 Downscaling methods

Generating information below the grid scale of CGCMs is referred to as downscaling; the two main approaches are the dynamical and empirical downscaling methods. Dynamical downscaling is achieved through high-resolution numerical climate models that use as boundary condition some data from CGCM simulations. The models are atmosphere-only GCMs, of uniform or variable horizontal resolution, and nested regional climate models (RCMs). Empirical downscaling also uses data from climate model simulations and applies to these statistical relationships derived from observed data or a statistical analysis of model behaviour. Dynamical downscaling has the potential for capturing mesoscale nonlinear effects under perturbed forcing conditions and providing coherent information between multiple climate variables. Confidence in the method to downscale realistically future climates comes from the ability of the models to faithfully reproduce widely varying climates around the world with the same set of equations. The main drawback of such models is their computational cost. Empirical downscaling has the ability to access scales finer than the dynamical methods and to make use of high resolution observations, where available, to provide information. The methods are computationally inexpensive though they have the drawback that they require long time series of reliable, homogeneous station data and assume that the derived statistical relationships will remain unaltered under perturbed climate.

11.2.1.1.1 Dynamical downscaling methods

High-resolution AGCMs

AGCMs can employ finer meshes than CGCMs. They include fully interactive land-surface processes as in CGCMs but their sea surface temperature and sea-ice (SSTI) are prescribed by interpolation of CGCMs' results. In some AGCMs, observed SSTI are used, either on their own for present day simulations of in combination with CGCM-simulated changes for future climate simulations. Model 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 (Bengtsson, 1996; May and Roeckner, 2001; Déqué and Gibelin, 2002; Govindaswamy, 2003). The Earth Simulator now allows global computations on 20 km grid mesh (Mizuta et al., 2005), although for short time slices.

Evaluated on the scale typical of current CGCMs, nearly all quantities simulated by higher resolution models agree better with observations, but the effect of increased resolution on skill actually varies significantly with region (Duffy et al., 2003). Notable improvements occur in orographic precipitation, and due to improved dynamics of mid-latitude weather systems (see Chapter 10, Section 10.3) and resolved tropical cyclones (see Chapter 10, Section 10.3).

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 (Douville, 2005; Inatsu and Kimoto, 2005). 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 boundary condition (May and Roeckner, 2001; Govindasamy et al., 2003). In AGCMs that derive their SSTI by combining changes in SSTI with analysed SSTI, there is an even greater risk of inconsistencies. While the large-scale responses appear to be similar in many regions, further research is required to determine if the similarity is accurate enough for the time-slice approach with AGCMs to be considered a valid downscaling technique.

An alternative to uniform high-resolution is variable-resolution (including stretched-grid) AGCMs (VRGCM; e.g., Déqué and Piedelievre, 1995; Krinner et al., 1997; Fox-Rabinovitz et al., 2001; McGregor et al., 2002; Gibelin and Déqué, 2003). The VRGCM approach is attractive as it permits to achieve, within a unified modelling framework, a regional increase of resolution while retaining the full interaction of all regions of the globe. Numerical artefacts due to stretching have been shown to be small when using modest stretching factors (e.g., Lorant and Royer, 2002). 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.

Nested RCMs

The principle behind regional climate models (RCMs) 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) are supplied as lateral boundary conditions (BC) and SSTI are supplied as lower BC; (2) the lateral BC exert sufficient control on the RCM large-scale circulation to keep it consistent with the driving large-scale atmospheric circulation; and (3) subgrid-scale physical processes are suitably parameterised, including fine-scale surface forcings such as orography, land-sea contrast and land use. The first successful demonstration was realised by Dickinson et al. (1989) and Giorgi and Bates (1989). Recently a two-way nested RCM has been developed (Lorenz and Jacob, 2005) that allows feedback from the RCM onto the GCM. RCMs are increasingly coupled interactively with other components of the climate system, such as regional ocean and sea ice (e.g., Maslanik et al., 2000; Döscher et al., 2002; Rinke et al., 2003; Debernard et al., 2003; Schrum et al., 2003; Meier et al., 2004; Rummukainen et al., 2004;), hydrology, and some work has been initiated with interactive vegetation (Gao and Yu, 1998; Xue et al., 2000)

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; Laprise 2003). The difficulties associated with the implementation of lateral BC are well documented (e.g., Warner et al., 1997). The mathematical interpretation is that nested models represent a fundamentally ill-posed boundary-value problem (Staniforth, 1997). These difficulties can be compounded in climate application 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 location and season (e.g., Caya and Biner 2004). In some applications, the flow developing within the RCM domain may become incoherent with the driving BC. This may (Jones et al., 1997) or may not (Caya and Biner, 2004) impact on climate statistics.

An important issue concerns the predictability of nested models: Can RCMs generate meaningful fine-scale structures that are absent in the lateral BC? de Elía et al. (2002) found that nested models are incapable of maintaining deterministic temporal coherence of small-scale features (at the right place at the right time) beyond a day or so, even if these were present initially and in the lateral BC. On the other hand, the climate statistics of small-scale features can be recreated with the right amplitude and spatial distribution, even if these small scales are absent in lateral BC (Denis et al., 2002, 2003; Antic et al., 2005; Dimitrijevic and Laprise, 2005). These results imply that RCMs can contribute added value at small scales to climate statistics when driven by CGCMs with accurate large scales

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. In multi-year ensemble simulations driven by atmospheric reanalyses at the lateral boundary, Vidale et al. (2003) have shown that RCMs have skill in reproducing interannual variability in precipitation and surface air temperature, though this is weakest in summer over continents. 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. Recently climate-change projections have been completed on the Earth Simulator with a 5-km mesh non-hydrostatic RCM over for East Asia (Kanada et al., 2005; Yoshizaki et al., 2005), for 10 years of June and July, driven by the outputs of a 20-km AGCM.

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 (see Chapter 8). 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 (see Chapter 8), which will lead to higher quality boundary conditions for RCMs.

11.2.1.1.2 Statistical downscaling methods

A complementary technique to RCMs is the use of statistically derived relationships linking large-scale atmospheric variables (predictors) and local/regional climate variables (predictands) and commonly referred to as empirical or statistical downscaling (hereafter SD). The local/regional scale climate-change information is obtained by applying the cross-scale relationships to equivalent predictor variables from GCM simulations. The IPCC Task Group on Data and Scenario Support for Impact and Climate Analysis **(**TGICA) guidance document (Wilby et al., 2004) provides a comprehensive background to use this approach with extensive examples from the literature, and covers the important issues to be addressed in any robust SD downscaling.

Important developments in SD research since the TAR are: increased availability of generic downscaling tools for the impact community (e.g., SDSM, Wilby et al., 2002b; clim.pact package, Benestad, 2004b); the use of downscaling techniques to address exotic variables such as phenological series (Matulla et al., 2003), extreme heat-related mortality (Hayhoe et al, 2004), ski season (Scott et al., 2003), land-use (Solecki and Oliveri, 2004); the downscaling of climate extremes (e.g., Katz et al., 2002; Wang et al., 2004a; Seem, 2004, Wang and Swail, 2004); inter-comparison studies evaluating statistical methods (e.g., STARDEX, Goodess et al., 2006; Schmidli et al., 2006; Haylock et al., 2006); 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, 2006; Wang and Swail, 2004); addressing non-stationarity in climate relationships with conservative methodologies (Hewitson and Crane, 2006); and spatial interpolation based on GIS-approach utilising geographical dependencies (Benestad, 2005).

SD techniques cover regression-type models including both linear or nonlinear relationships between predictands and large-scale predictors, weather generators (WGs) which are mature SD methods for generating synthetic sequences of local variables that replicate their observed statistical attributes, techniques based on the weather classification which draw on the more skilful attributes of GCMs to simulate circulation patterns, , analogue methods which seek equivalent weather states from the historical record, or some combination of these. In an extension to these the statistical-dynamical downscaling (SDD) (e.g., Fuentes and Heimann, 2000) technique combines weather classification with RCM simulations. A possibly valuable development of the above approaches could be the application of the SD techniques to the high resolution CGCMs/RCMs (Lionello et al., 2003; Imbert and Benestad, 2005). For example, Lionello et al. (2003) found that the surface wind fields derived from T106 ECHAM-4 sea level pressure fields by statistical downscaling model based on CCA are much improved with respect to the T106 fields.

In some cases SD may be used to predict statistical attributes as opposed to predicting the raw values of the predictand, for example the probability of rainfall occurrence, precipitation / wind distribution parameters, frequency of extreme events, and 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, Pryor et al., 2005ab). 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., Bartman et al., 2003; Busuioc et al., 2001a; Trigo and Palutikof, 2001; Hansen-Bauer et al., 2003). Stationarity remains a concern with SD, as to some degree it may be with RCMs, as to whether the cross scale relationships are valid under future climate regimes. This is only weakly assessed through cross-validation tests, although convergence of the climate-change signals across CGCMs, RCMs and SDs can further strengthen the results (e.g. , Hewitson and Crane, 2006, Busuioc et al., 2006). More recently, the degree of non-stationarity in a projected climate change has been assessed as part of a SD application

(Hewitson and Crane, 2006). Most appropriate are methods that combine 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., 2004a; Lionello et al., 2003; Hewitson and Crane, 2006; Wilby et al., 2003; Hansen and Mavromatis, 2001). Regarding the predictors, the best choice is to combine dynamical and moisture variables (e.g. Wilby et al., 2003).

11.2.1.1.3 Pattern scaling of climate model simulations

Pattern-scaling methods allow development of 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 model (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. More details are presented in TAR (Chapter 13). Some developments were made using various versions of scaling techniques (e.g., Christensen et al., 2001; Mitchell, 2003; Ruosteenoja et al., 2006; Salathé , 2005). For example, Ruosteenoja et al. (2006) developed a 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, Chapter 9, Appendix 9.1). In order to reduce the noise induced by the GCM internal variability (a common problem to all scaling methods), the scaling was carried out using an ensemble mean instead of an individual GCM response.

11.2.1.1.4 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 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). More details about these methods are presented in the TAR (Chapter 13).

11.2.1.1.5 Inter-comparison of downscaling methods

Any studies comparing several SD techniques (Bartman et al., 2003; Buishand et al., 2004; Diaz-Nieto and Wilby, 2005; Goodess et al., 2006; Matulla et al., 2003; Huth, 2002, 2003; Schoof and Pryor, 2001; Widmann et al., 2003; Wilby et al., 2002a, 2003; Wood et al., 2004) as well as SD with CGCMs/dynamical downscaling (e.g., Huth et al., 2001; Hansen-Bauer et al., 2003, 2005; Wood et al., 2004, Busuioc et al., 2006; Schmidli et al., 2006; Haylock et al., 2006) 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, 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 may not be better than simpler linear / slightly non-linear approaches for some applications. However, Haylock et al. (2006) found that models based on non-linear artificial neural networks are best at modelling the inter-annual variability of heavy precipitation but underestimate extremes.

Since the TAR a few studies have systematically compared the SD and RCM approaches. These mainly related to the similarity of the climate change signal (e.g. Hanssen-Bauer et al., 2003). A more complex study considered using additional information about the RCM skill in simulating the current regional climate features for fitting the SD models (Busuioc et al., 2006). Other studies resulted from the STARDEX project (e.g. Schmidli et al., 2006; Haylock et al., 2006) compared the two approaches in terms of their skill in reproducing current climate features as well as the future climate change scenarios, focusing on climate extremes and complex topography over Europe. The conclusion of the TAR that SD and RCM downscaling techniques are comparable for simulating current climate appears to still hold, even while both methodological approaches have matured and become more skilful. It is thus recommended that more studies be undertaken to leverage the relative strengths of both statistical and dynamical downscaling.

11.2.2 Quantifying Uncertainties

11.2.2.1 Sources of regional uncertainty

Most sources of uncertainty on regional scales are similar to those on the global scale (Chapter 10, Section 10.5), but there are both changes in emphasis and new issues that arise in the regional context. Of the climate forcing agents, uncertainty in aerosol forcing adds especially to regional uncertainty because of the spatial inhomogeneity of the forcing and the response. Land use/cover change has an inherently regional scope as well (De Fries et al., 2002; Chapter 2;, and Box 11.5). When analyzing studies involving further layers of models too add local detail, the cascade of uncertainty through the chain of models used to generate regional or local information has to be considered.

A major component of uncertainty is the representation in climate models of the response of the climate system to anthropogenic emissions and other perturbations to drivers of the system. These include uncertainties in: the conversion of projected future emissions into concentrations of radiatively active species (i.e., via atmospheric chemistry and carbon-cycle models); the radiative forcing for known concentrations (particularly large for aerosols); other response of the physical climate system to these forcings resulting from incomplete representations of resolved processes (e.g., moisture advection) and parameterizations of sub-grid-scale processes (e.g., clouds, precipitation, planetary boundary layer, land surface), e.g. the strength of feedback mechanisms on the global and regional scale. The property of the climate system, and of climate models, that integrates a large fraction of these sources of uncertainty is global climate sensitivity, and Chapter 10, Box 10.2 is dedicated to its in-depth treatment. The degree to which these uncertainties influence the projections of different climate variables is not uniform. For example models agree more readily on the sign and magnitude of temperature changes than of precipitation changes.

The regional impact of these uncertainties in the response of the climate system has been illustrated by several authors. Incorporating a model of the carbon-cycle into a coupled AOGCM gave a dramatically enhanced response to climate change over the Amazon basin (Cox et al. 2000; Jones et al. 2003) and Borneo (Kumagi et al. 2004). The scale of the resolved processes in a climate model can significantly affect its simulation of climate over large regional scales (Pope and Stratton 2002; Lorenz and Jacob 2005). Frei et al. (2003) show that models with the same representation of resolved processes but different representations of sub-grid-scale processes can represent the climate differently. The regional impact of changes in the representation of the land-surface feedback is demonstrated by, for example, Oleson et al. (2004) and Feddema et al. (2005b). See also Box 11.5 on land use.

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

11.2.2.2 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 GCMs were qualitatively portrayed (e.g., large or small increases/decreases in precipitation) based only on simple agreement heuristics (e.g., seven of the nine models showed increases). 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 and attaching probabilities to a group of scenarios on a regional scale (New and Hulme, 2000; Jones, 2000)

Although, 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 10.5). For statistical reasons, large ensembles of projections from full GCMs are necessary to produce formally robust probabilistic estimates of sub- continental scale regions; and until very recently, sufficient computer resources have not been available for such studies

11.2.2.2.2 Using multi-model ensembles

A number of studies have taken advantage of multi-model ensembles formed by GCMs that have run the same climate experiments to generate quantitative measures of uncertainty, particularly probabilistic information on a regional scale. Table 11.1 summarizes aspects of the methods reviewed below and in Section 11.2.2.2.3. The results highlighted in Chapter 10, Section 10.5 and Box 10.2 on climate sensitivity, demonstrate t that multi-model ensembles explore only a limited range of the uncertainty that may exist. Also, the distribution of GCM sensitivities is arbitrary and does not form a representative sample from the probability distributions derived for climate sensitivity. Thus, regional probabilities generated using multi-model ensembles should be viewed as relatively conservative, i.e., they underestimate the width of the PDFs of future regional climate change.

[INSERT TABLE 11.1 HERE]

Räisänen and Palmer (2001) used 17 GCMs forced with an idealised but physically plausible annual increase in CO_2 of 1% to calculate the probability of exceedance of thesholds of temperature increase (e.g., >1°C) and precipitation change (e.g., <–10%). These were used to demonstrate that a probabilistic interpretation of climate change has advantages over conventional deterministic interpretations by demonstrating the economic value of a probabilistic assessment of future climate change. Giorgi and Mearns (2002) developed measures of uncertainty for regional temperature and precipitation change by weighting model results according to biases in their simulation of present-day climate and convergence of their projections to the central tendency of the aggregated model projections. These were applied to the 9 GCMs assessed in the TAR to provide uncertainty estimates separately for the A2 and B2 SRES emission scenarios for 22 large sub-continental regions. Benestad (2002b,2004) used a multi-model ensemble coupled to statistical downscaling to derive tentative probabilistic scenarios at a regional scale.

Tebaldi et al. (2004a, 2005) used a Bayesian approach to define a formal statistical model for deriving probabilities from an ensemble of projections forced by a given SRES scenario. In this, current and future regional climate signals and model reliabilities are treated as uncertain quantities which start with uninformative (i.e., flat) prior distributions that are updated using data (model projections and observations) via Bayes' theorem. These data are applied similarly to Giorgi and Mearns (2002 and 2003) so posterior PDFs of temperature and precipitation change signals are obtained, from the models' biases with respect to current climate observations and models' convergence. The choice of applying the observed and model data this way is a matter of expert judgement as are the relative weights they should have within the method.

Greene et al. (2005) used a Bayesian framework to model an ensemble of GCM projections under individual SRES scenarios by an extension of the methods for seasonal forecasting. The set of GCM simulations of the observed period 1902–1998 are jointly calibrated through a linear model to the observed trend. The coefficient estimates and their uncertainty are derived and then applied to the projections to provide probabilistic forecast of future trends. The assumption of the applicability of the relationship between observed and modelled historical regional trends to the models' projections produces a stricter constraint than the bias criterion in Tebaldi et al. (2004, 2005). Also, in some regions, particularly at lower latitudes, the PDFs are significantly shifted from the location of the ensemble's individual projections implying that calibrating the model trends to fit the historical trends significantly reshapes them. Finally, the approach does not make allowance for uncertainties in historical forcings and not all models incorporate all focrings.

Figure 11.2.1 compares the Tebaldi et al. (2004, 2005) and Greene et al. (2006) methods for the Giorgi regions with the raw model projections. Each bar represents the range of values covering 90% of the probability of temperature change (2080–2099 vs. 1980–1999) in December, January and February. A major factor in the differences are the differing criteria for weighting models and emphasize the key role played by these assumptions in this kind of analysis. Understanding which aspects of a model are most crucial for its climate projection, and, therefore, which comparisons to observations are of most relevance in weighting or adjusting models projections so as to refine the raw model output remains an open research problem. Maps of temperature change under A1B in June, July and August and of precipitation change under A1B for both seasons (Tebaldi et al. (2004, 2005) and empirical PDFs only for precipitation) are included among the supplementary material (Supplementary material Figures S11.2.1–3).

[INSERT FIGURE 11.2.1 HERE]

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). Their work is focused on measuring the changes in PDFs as a function of the different sources of uncertainty. In this analysis, the impact of the SRES scenarios turns out to be the most relevant for temperature changes, particularly in the upper tail of the distributions while the GCM weighting does not produce substantial differences. Climate sensitivity has an impact mainly in the lower tail of the distributions. For precipitation changes, all sources of uncertainty seem relevant but the results are very region-specific and thus difficult to generalize. However, the use of pattern scaling will likely underestimate the range of projections that would be obtained by running a larger ensemble of GCMs (Murphy et al., 2004).

The work described above has involved either large area averages of temperature and precipitation change or statistical modelling at the grid box scale. Good and Lowe (2006) show that trends for large area and grid box average projections for precipitation are often very different. This demonstrates the inadequacy of inferring the behaviour at fine-scales from that for large-area averages However, the study finds stable, region-dependent relationship between inter-model variability at the sub-regional and regional scales, in a framework similar to pattern-scaling.

11.2.2.2.3 Using perturbed physics ensembles

Another method for exploring uncertainties in regional climate projections is the use of large perturbed physics ensembles (described in detail in Chapter 10). These allow a characterisation of the uncertainty due to poorly constrained parameters within the formulation of a model. Collins et al. (2006) applied this method to produce a 17-member ensemble of GCM projections under the idealised scenario of 1% per year CO_2 increase. The study offers preliminary results in terms of mean and standard deviation of global fields of temperature and precipitation change and opens the way to more formal Bayesian approaches to the evaluation of perturbed physics experiments at the regional scales. Harris et al. (2006) have combined the results from this study with a larger perturbed physics ensemble investigating the equilibrium climate response to a doubling of CO_2 (Murphy et al., 2004). They developed a bridge between spatial patterns of the transient and equilibrium climate response by way of simple pattern scaling (Santer et al. 1990) allowing results from the large ensemble to be translated into PDFs of time dependent regional changes. Uncertainties in surface temperature and precipitation changes are derived (Supplementary material Figures S11.2.4 and S11.2.5), which arise from the poorly-constrained atmospheric model parameters, internal variability and pattern scaling errors. The latter are calibrated by matching the transient and equilibrium responses of the 17 model versions with corresponding parameter settings. Scaling errors are largest when the transient response varies non-linearly with global temperature, as is the case for precipitation in certain regions. Again, a key assumption in these methodologies is the use of GCM present-day simulation biases to provide a weighting function for the ensemble of projections.

11.2.2.2.4 Other approaches to quantifying regional uncertainty

As described in Chapter 10, Stott and Kettleborough (2002) provide pdfs of future climate change by making use of the robust observational constraints on a climate model's response to greenhouse gas and sulphate aerosol forcings that underpin the attribution of recent climate change to anthropogenic sources. The study by Stott et al. (2005) is the first to adapt this method for the regional (or continental in this case) scales. This method uses the linear scaling factors which demonstrate the link between a GCM's response to observed forcings and changes in climate and the skill of a GCM to reproduce these observed changes. Differing from the studies described in Section 11.2.2.2, this strain of work uses projections from a single GCM (HadCM3) though Stott et al. (2006) have confirmed the results of this methodology with other models. The regional projections 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 which forms the basis of the estimated scaling factors, is larger. The second approach incorporates more information, hence reducing the uncertainty, but assumes the GCM represents correctly the relationship between global mean and regional temperature change.

An approach to a process-based assessment of the reliability of modelled climate change responses and thus uncertainties in its future projections has been proposed by Rowell and Jones (2006). They perform an assessment of the physical and dynamical mechanisms responsible for a specific future outcome, in their case European Summer drying. Their analysis isolates the contribution of the four major mechanisms analysed, spatial pattern of warming, other large-scale changes, reduced spring soil moisture and summer soil moisture feedbacks. In certain regions the second process makes a minor contribution with the first and third dominating. This leads to the conclusion that the sign of the change is robust as confidence in the processes underlying these mechanisms is high.

In general, the regional sections of this chapter can be seen as an application of these same ideas: providing a likelihood statement of change based on expert opinion from understanding the climate processes relevant to a region and evaluation of the projected changes by different models together with assessment of observational evidence to support the model-projected changes.

11.2.2.2.5 Combined uncertainties: GCMs, emissions, and downscaling techniques

It is important to quantify the relative importance of the uncertainty from the downscaling step (from one's RCM formulation or the assumptions underlying one statistical model) against the other sources of uncertainty. The PRUDENCE project provided the first opportunity to weigh these various sources of uncertainty for simulations over Europe. Rowell (2005) evaluated a 4 dimensional matrix of climate modelling experiments that included two different emissions scenarios, 4 different GCM experiments, and 9 different RCMs, for the area of the British Isles. He found that the dynamical downscaling added a small amount of uncertainty compared to the other sources for temperature evaluated as monthly/seasonal averages. For precipitation the relative contributions of the four sources of uncertainty are more balanced. Deque et al. (2005, 2006) show similar results for the whole of Europe, as do Ruosteenja et al. (2006) for subsections of Europe. Kjellstrom et al. (2006) found that the differences among different RCMs driven by the same GCM become comparable to those among the same RCM driven by different GCMs when evaluating daily maximum and minimum temperatures. However, the spread in the responses of the PRUDENCE RCMs compared to that of the driving GCM suggests that some of this variability of the RCM responses may be spurious (Jones et al., 1997). It should be also noted that only few of the RCMs in PRUDENCE were driven by more than one GCM which adds further uncertainty regarding these conclusions. Other programs similar to PRUDENCE have begun for other regions of the world, such as NARCCAP over North America (Mearns et al., 2005), and CREAS over South America.

11.3 Regional Projections

11.3.1 Introduction to Regional Projections

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

Each of these continental-scale regions encompasses a broad range of climates; they are generally too large to be used as a basis for conveying quantitative regional climate change information. Therefore, each of these is subdivided into a number of sub-continental or oceanic regions. For example, Africa is comprised of the Saharan, East African, West African and South African regions. These regions are used for presenting area-averaged precipitation and temperature change information from the AR4 GCM simulations. The region boundaries are defined in Table S11.1 in Supplementary Material. They are very close to those initially devised by Giorgi and Francesco (2000) with some minor modifications similar to those of and Ruosteenoja et al. (2003). The objectives behind the original Giorgi and Francesco (2000) regions were that they have simple shape, be no smaller than the horizontal scales on which current GCMs are thought to be useful for climate simulations (typically judged to be a few thousand kilometres), and recognise where possible distinct climatic regimes.

These regional averages have some deficiencies for discussion of the GCM projections. In several instances the boundaries of these boxes are oriented awkwardly with respect to the mean AR4 GCM hydrological responses, averaging together areas in which precipitation is projected to increase with regions in which it is projected to decrease. South America provides one such example. There are also areas of smaller scale than these regions, where the case can be made for a robust and physically plausible hydrological response in the AR4 GCMs, such as South-western Australia or the central Andes, which are washed out in the regional averages. Partially to help in discussing these features, we also use maps of temperature and precipitation responses, interpolated to a 128 (longitude) x 64 (latitude) grid typical of many of the lower resolution atmospheric models in the AR4/PCMDI Archive.

In most of the regional discussion to follow, the focus is on temperature and precipitation, both seasonal means and on extremes on various time scales. The focus on precipitation is problematic, in that it provides a limited view of hydrological changes, but was deemed necessary to limit the amount of information presented. Supplementary material Figure S11.3.1.1 illustrates this issue by comparing the annual mean response in precipitation to the annual mean response in precipitation minus evaporation, over the 21st century in the A1B scenario across the AR4 GCM ensemble. Over North America and Europe, in particular, the region of drying in the sense of precipitation minus evaporation (or runoff, over land) is shifted polewards compared to the region of reduced precipitation. This distinction should be kept in mind in the following discussion.

For each region we gather onto a single graph: 1) the observed time series of the evolution of the surface air temperature anomaly during the 20th century with respect to the century average; 2) the spread of the 20th century simulations by the AR4 GCMs that contain a full set of historical forcings of the same quantity as displayed for the observations; 3) the evolution of the range of this temperature anomaly as represented in the 21 AR4 projections for the A1B scenario between 2000 and 2100, and 4) the spread of the projected anomaly for the last decade of the 21st century for the B1, A1B, and A2 scenarios. Averages are taken over all realizations for each model before they are used as input into these figures to emphasize the spread in estimates of the forced response and minimize internal variability. For an example, see Figure 11.3.2.2 for the African regions. These plots serve to place the temperature projections in the context of the observed trends and help one visualize the regional definitions. The 20th century segments of these plots are displayed in more detail and discussed in Chapter 9.

Table 11.2 provides detailed information for each region generated by the AR4 global models focusing on the change in climate between the 1980-1999 period in the “20C3M” integrations and the 2080–2099 period using the A1B scenario. The distribution of the annual and seasonal mean surface air temperature and fractional precipitation responses are described by the median, the 25% and 75%, or quartile, values (half of the models lie between these two values) and the maximum and minimum values in the model ensemble. Information on model biases in these regional averages for the 1980–1999 simulations is provided in Supplementary material Table S11.2 in a similar format. We also include in the discussion to follow temperature time-series plots for each of these regions similar to those shown in Box 11.1 for the continental averages.

Most of the discussion focuses on the A1B scenario. The global mean near-surface temperature responses (between the period 1980–1999 of the 20C3M integrations and the period 2080–2099) in the ensemble mean of the AR4 GCMs are in the ratio 0.69:1:1.17 for the B1:A1B:A2 scenarios. The local temperature responses in nearly all regions closely follow the same ratio, as discussed in Chapter 10 and as illustrated in Supplementary material Figures S11.3.1.2-4, the high latitude oceans departing most significantly from this caling. Therefore, little is gained by repeating discussion of the A1B scenario for the other cases. The ensemble mean local precipitation responses also roughly scale with the global mean temperature response, although not as precisely as the temperature itself. Given the substantial uncertainties in hydrological responses, the generally smaller signal/noise ratio, and the similarities in the basic structure of the GCM precipitation responses in the different scenarios, a focus on A1B seems justified for the precipitation as well. The overall regional assessments, however, do rely on all available scenario information.

The evolution of the local temperature response in the mean model A1B projection is typically very linear in time. There is little to be gained by adding discussion of different time periods other than 2080–2099 when discussing the mean climate. There is no indications in the ensemble mean GCM projections of abrupt climate change or even substantial nonlinearity. In the literature on the individual global models one can find instances of apparent nonlinearity in the 21st century scenarios integrations (i.e., Held, et al. 2005), but in no instance does this appear to be robust across models. While the possibility exist that the models are missing sources of abruptness, most likely involving ocean circulation or land surface/vegetation feedbacks, having little basis to judge the plausibility of these factors (see Chapter 10), we base all of our discussion on this linear picture.

Table 11.2 also provides information on the signal/noise ratio. The noise in this case is an estimate of the internal variability of the 20 year means of the seasonal or annual mean temperature or precipitation, as generated by the models. The signal-to-noise ratio is converted into the time interval that is required before the signal is *clearly discernable*, assuming that the signal grows linearly according to the rate of the ensemble mean A1B projection. Because this noise estimate is solely based on the models, it must be treated with caution, but it would be wrong to assume that models invariably underestimate this internal variability. Some models overestimate and some underestimate the amplitude of ENSO, for example, thereby over- or under-estimating the most important source of interannual variability in the tropics, and some models are documented as clearly overestimating the interannual variability of land surface temperatures in mid-latitudes (Chapter 8). On the other hand, few models capture the range of decadal variability of rainfall in West Africa (Hoerling, et. al. 2006).

Also included in Table 11.2 is an estimate of the probability of *extremely warm, extremely wet, and extremely dry seasons*, once again for the A1B scenario, for the time period 2080–2099. *An extremely warm* summer is defined as follows. Examining all of the summers simulated in a particular realization of a model in the 1980–1999 control period, one can compute the warmest of these 20 summers, as an estimate of the temperature of the warmest 5% of all summers in the control climate. One then examines the period 2080–2099, and determines what fraction of the summers exceed this warmth. This is referred to as the probability of extremely warm summers. The results are tabulated after averaging over models, and similarly for both extremely low and extremely high seasonal precipitation amounts.

Reference in the following is made to the probabilistic results of Tebaldi et al. (2004,2005) in order to provide an example of quantifying uncertainty in regional climate change from multi-model ensembles. Quantiles of the PDFs presented in Supplementary material Table S11.3 summarize the probabilistic uncertainty ranges for change in seasonal temperature and percent precipitation under the A1B scenario. As discussed in Section 11.2.2, this area of research is evolving and these results should be viewed as illustrative.

11.3.1.2 Some unifying themes

The basic pattern of the projected warming is little changed from previous assessments, as described in Chapter 10. Examining the spread across the AR4 GCMs, one finds that temperature projections in many regions are strongly correlated with the global mean projections, with the most sensitive models from a globally averaged perspective often the most sensitive locally. While differing treatments of regional processes are responsible for some spread, a substantial part of the spread in regional temperature projections is also due to differences in the total sum of the global feedbacks that control global transient climate sensitivity.

The response of the hydrological cycle is controlled in part by fundamental consequences of warmer temperatures and the increase in water vapor in the atmosphere (Chapter 3). Water is 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 more water will converge into regions of climatological convergence and more will diverge out of regions of climatological divergence. We see the consequences of this increased moisture transport in plots of the global response of precipitation described in Chapter 10 where, on average, precipitation increases in the intertropical convergence zones, decreases in the subtropics, and increases in sub-polar and polar regions. One expects to see this pattern imprinted on the salinity distribution in the world oceans, as described in Chapter 5. This pattern is also described in Chapter 8, which assesses the extent that this pattern is visible over land during the 20th century in observations and in model simulations.

Over North America and Europe, this pattern of subpolar moistening and subtropical drying dominates the 21st century projections. Regions of large uncertainty often lie near the boundaries between these robust moistening and drying regions, with different models placing these boundaries differently.

Another important theme in the models 21st century projections is the poleward expansion of the subtropical highs, and the poleward displacement of the midlatitude westerlies and associated storm tracks. This circulation response is often referred to 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 poleward boundaries of the 5 subtropical oceanic high centers in the South Indian, South Atlantic, South Pacific, North Atlantic and, less robustly, the North Pacific (where a tendency towards El-Niño-like conditions in the Pacific trends to counteract this expansion) . Most of the regional projections of strong drying tendencies over land in the 21st century are immediately downstream of these centers (Southwestern Australia, the Western Cape Provinces of South Africa, the central Andes, the Mediterranean, and Mexico). The robustness of this large-scale circulation signal is discussed in Chapter 10, while Chapters 3, 8, and 9 describe the observed poleward shift in the Southern Hemisphere in the late 20th century and the ability of models to simulate this shift.

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 an earlier spring snowmelt, and, more generally, the earlier reduction in soil moisture (Manabe and Wetherald, 1987) is a continuing theme in discussion of summertime continental climates.

The well-known control that sea surface temperature anomalies exerts on tropical rainfall variability provides an important unifying theme for tropical climates. Models can differ in their projections of small changes in tropical ocean temperature gradients and in there simulation of the potentially large shifts in rainfall that are forced by these oceanic changes. Chou and Neelin (2003) provides a guide to some of the complexity involved in diagnosing and evaluating hydrological responses in the tropics. With a few exceptions the spread in projections of hydrological changes is still too large to make strong statements about the future of tropical climates. The difficulty of making projections for tropical storm frequency adds to this uncertainty.

Assessments of the regional and sub-regional climate change projections have primarily been based on 1) the GCM projections summarized in Table 11.2 and an analysis of the biases in the GCM simulations, 2) regional downscaling studies available for some regions with either physical or statistical models or both, and 3) reference to plausible physical mechanisms which have attained a sufficient level of credibility in the community.

To assist the reader in placing the various regional assessments in a global context, Box 11.1 visualises many of the detailed assesments documented in the following regional sections. Likewise, a global overview of projected changes in various types of extreme weather statistics are summarised in Table 11.3. This table not only contain information extracted from the assessments within this chapter, but also holds information extracted from Chapter 10. Thus the details of the assessment that lead to each individual statements can all be found in either Chapter 10, or the respective regional sections, and clear links for each statement are identifiable from Table 11.3.

[INSERT TABLE 11.3 HERE]

Box 11.1: Summary of Regional Responses

The discussion on regional projections is organized according to the same regions adopted for discussion of impacts in WG II in the AR4 and in earlier assessments: *Africa, Europe and Mediterranean, Asia, North America, Central and South America, Australia-New Zealand, Polar Regions, and Small Islands*. As an introduction, we illustrate how continental scale warming is projected to evolve in the 21st century using the AR4 models. We also put this warming into the context of the observed warming during the 20th century and the ability of that subset of the AR4 models using all known forcings to simulate the observed evolution (see Chapter 9 for more details). Box 11.1, Figure 1 shows each continental region: 1) the observed time series of the evolution of the surface air temperature anomaly during the 20th century with respect to the century average; 2) the spread of the 20th century simulations by the AR4 GCMs that contain a full set of historical forcings of the same quantity as displayed for the observations; 3) the evolution of the range of this temperature anomaly as represented in the 21 AR4 projections for the A1B scenario between 2000 and 2100, and 4) the spread of the projected anomaly for the last decade of the 21st century for the B1, A1B, and A2 scenarios.

[INSERT BOX 11.1, FIGURE 1 HERE]

Figure Box11.1, Figure 2 serves to illustrate some of the more significant hydrological changes, with the two panels corresponding to the months of Dec-Jan-Feb and Jun-Jul-Aug. The backdrop to these figures is the fraction of the GCMs (out of the 21 considered for this purpose) that predict an increase in mean precipitation in that grid cell (using the A1B scenario and comparing the period 2080–2099 with the control 1980–1999). Aspects of this pattern is examined more closely in the separate regional discussions. Robust findings on regional climate change for mean and extreme precipitation, drought, snow, sea-ice, extreme winds and tropical cyclones are highlighted.

[INSERT BOX 11.1, FIGURE 2 HERE]

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

There are many pathways through which changes in the surrounding oceans can alter African climates. The Indian Ocean supplies most of the water for rainfall in Southern and Eastern Africa, and anomalies in Indian Ocean temperatures strongly affects these regions in GCMs (Bader and Latif, 2003). The North Atlantic, with its variable and potentially sensitive overturning circulation, together with the waters of the Gulf of Guinea (Vizy and Cook, 2001), 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 Pacific ocean 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. The field has moved steadily away from explanations for rainfall variations in this region as due primarily to land use changes and towards explanations based on changes in sea surface temperatures (SSTs). The early SST perturbation GCM experiments (Folland et. al., 1986) have being updated with impressive results from the most recent models (Giannini, et. al., 2003; Hoerling, et al, 2006). Haarsma et al. (2005) showed the mechanism by which the global SST distribution affects atmospheric circulation and also how that mechanism leads to an increase of Sahel rainfall in response to anthropogenic warming. 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, although a subset of models also dry the Sahel in response to uniform warming of SSTs (Held, et. al., 2005). 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 dry the Sahel. The work of Rotstayn and Lohmann (2002), supports this picture, as do Held, et al. (2005) and Paeth and Feichter (2006).

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. Evidence has been presented for strong links 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., 2003; Tadross et al., 2005ab; Anyah and Semazzi, 2004), it is plausible that these land-surface feedbacks are also important for climate change simulations in Southern Africa

Increasing 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, and the increase in days with stable inversion layers over southern Africa (Freiman and Tyson, 2000; Tadross et al., 2005b, 2006) in the late-20th century suggests that the same process (possibly linked to increases in Indian ocean SSTs) plays a role in this trend, as well as in related positive trends in southern African daytime temperatures and consecutive dry days (New et al., 2006).

There is little doubt that vegetation patterns help shape the climatic zones throughout much of Africa (e.g., Wang and Eltahir, 2000; 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 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. The possibility of multiple stable modes of African climate due to vegetation/climate interactions has been raised, especially in the context of discussions of the very wet Sahara during the mid-Holocene 6–8 K yr BP (Foley et al., 2003; Claussen et al., 1999). One implication is that feedbacks associated with vegetation patterns may make climate changes less reversible.

11.3.2.2 Skill of models in simulating present and past climates

The precipitation generated by the ensemble mean of 21 of the models in the PCMDI/AR4 database, averaged over the years 1979–1999 from the 20C3M integrations, is displayed in Supplementary material Figure S11.3.2.1. Average biases for four African sub-regions are also provided in Supplementary material Table S11.2. There are biases that are systematic across the ensemble, an overestimate of rainfall in Southern Africa being of special concern. Of these models, 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. Models often generate the largest fractional precipitation responses in dry or semi-arid regions, so this bias raises a concern that the sensitivity of southern African precipitation could be underestimated. Simulated surface temperatures across Africa in the AR4 models are too cold on average, by about 1K, with larger cold biases in drier areas, but these temperature biases in themselves are not large enough to affect the credibility of the model projections.

The intertropical convergence zone in the Atlantic is displaced equatorward in nearly all AR4 models, and ocean temperatures are too warm by an average of 1–2 K in the Gulf of Guinea, and typically by 3 K 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 contribute to the difficulties many of the models have in simulating West African and Sahel rainfall, as critically analyzed by Cook and Vizy (2006). In several of the models the summer rains in West Africa fail to move from the Gulf onto land, so there is effectively no West

African Monsoon, but many of the models do have a monsoonal climate albeit with some distortion. Moderately realistic interannual variability of SSTs in the Gulf of Guinea and the associated dipolar rainfall variations in the Sahel and the Guinean Coast is, by the criteria of Cook and Vizy, only present in 4 of the 18 models examined. Tennant (2003) examines three GCMs from the TAR in terms of their simulation of southern Africa regional synoptic and inter-annual variability, describes systematic biases such as the equatorward displacement of the midlatitude jet in austral summer, a deficiency that persists in the AR4 global models (Chapter 8), and notes that the models with the best synoptic variability do not necessarily generate the most realistic responses to the interannual variability in SSTs.

The multi-model analysis of Hoerling, et al. (2006) 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 (e.g. Bader and Latif, 2003 Giannini et al., 2003; Kamga et al., 2005; Haarma et al., 2005) using simulations of this type to study interannual variability. However, there is less confidence in the ability of coupled GCMs 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 AR4 models produce droughts comparable in magnitude to the Sahel drought of the 1970's and 1980's (Hoerling, et. al., 2006). There are exceptions, but what distinguishes these from the bulk of the AR4 models is not understood.

The very wet Sahara in the mid-Holocene (6–8 thousand years ago) is thought to be the climatic response to the increased summer insolation due to alignment of the perihelion of the Earth's orbit with summer solstice. These studies 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. The fact that GCMs continue to have difficulty in simulating the full magnitude of the mid-Holocene wet period, especially in the absence of vegetation feedbacks, may indicate a lack of sensitivity to other kinds of forcing. (Jolly et al., 1996; Kutzbach et al., 1997)

11.3.2.3 Regional downscaling

Regional climate simulations using dynamical models with a specific focus on Africa are very limited, and only in recent years has simulation quality been rigorously evaluated. In view of the biases noted above, the boundary conditions provided by global GCMs are unlikely to be adequate for many detailed regional issues, but the finer resolution in RCMs should still result in qualitatively useful information on the effects of local orography and sharp gradients in land surface properties. In East Africa, some studies have focused on how regional climate dynamics are influenced by the Great Lakes, (Anyah and Semazzi, 2004; Song et al., 2004, following earlier work of Indeje, 2001) however, the simulations are too short to draw meaningful conclusions about climate sensitivity.

The bulk of African regional climate modelling has focused on southern Africa. Some of the problems encountered are shared with the global models. For example, Engelbrecht et al. (2002) and Arnell et al (2003) both simulate excessive rainfall in parts of southern Africa, reminiscent of the bias in the AR4 global models. Hewitson et al. (2004) and Tadross et al. (2006), find sensitivity of both the frequency and diurnal cycle of rainfall to the choice of convective parameterisation, a familiar problem in GCMs. Tadross et al. (2005b) and New et al. (2003) explore the sensitivity of this model to changes in soil moisture and vegetative cover, reinforcing the view (Rowell, et al, 1995) that land surface feedbacks enhance regional climate sensitivity over Africa's arid and semi-arid region. Sensitivity of the simulated precipitation to the model design is found to be particularly large under high pressures systems, the frequency of which has increased in recent decades (Tadross et al., 2005b), increasing the importance of this problem for simulation of rainfall trends. When optimized and forced with observed flows at the lateral boundaries, these models can improve on the climatologies generated by global models.

Over West Africa the number of RCM investigations is even more limited (Jenkins et al., 2002), with a focus typically on the simulation of regional phenomena, including African easterly waves (Druyan et al., 2001), and the African easterly Jet (Hsieh and Cook, 2005). Vizy and Cook (2002) have studied the southward shift of the ITCZ in response to warm SSTs in the Gulf of Guinea, resulting in realistic positive rainfall anomalies along the coast and a drying over the Sahel. The quality of the 25-year simulation undertaken by Paeth et al.

(2005) is encouraging, emphasizing the role of regional SSTs and changes in the land surface in forcing West African rainfall anomalies.

Analyses of African climate change in high resolution time-slice simulations are also very limited (e.g., Coppola and Giorgi et. al., 2005) and difficult to utilize until a larger range of models are available at these resolutions.

Empirical downscaling has been applied over southern Africa for a number of different applications. For example, Landman and Goddard (2002) used empirical techniques to enhance seasonal forecasting products. For longer simulation periods Hewitson and Crane (2005) have developed empirical downscaling for point scale precipitation at sites spanning the continent, as well as a 0.1° resolution grid over South Africa. The downscaled precipitation forced by reanalysis data provide a close match to the historical climate record, including regions such as the eastern escarpment of the sub-continent that have proven difficult for RCMs.

11.3.2.4 Climate projections
11.3.2.4.1 Mean temperature

[INSERT FIGURE 11.3.2.1 HERE]

Focusing on the differences in near surface temperature between years 2080–2099 in the A1B scenario and the years 1980–1999 in the 20C3M 20th century simulations, averages over the West African (WAF), East African (EAF), South African (SAF), and Saharan (SAH) sub-regions are provided in Table 11.2. The Mediterranean coast is discussed together with Southern Europe in Section 11.3.3. The upper panels in Figure 11.3.2.1 show the geographical structure of the ensemble mean projected warming in more detail.

Global models predict a relatively uniform warming over the continent. In most regions the ensemble mean response is between 3 and 4 K, with smaller values in equatorial and coastal areas and larger values in the Western Sahara. This African temperature response is about 50% larger on average than the global mean response. The table shows that half of the models project warmings within about 0.5K of the median values. The total range of the regional warming is comparable in percentage terms to the range of global mean warming. There is a strong correlation across the AR4 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. The pdf constructed by Tebaldi et al, (2004, 2005) (Supplementary material Table S11.3), have a very similar half width for temperature but reduce the likelihood of the extreme high limit as compared to the raw quartiles in Table 11.2.

The largest temperature responses in North Africa are projected to occur in June-July-August, while the largest responses in Southern Africa occur in September-October-November. But the seasonal structure in the temperature response over Africa is modest as compared to extratropical regions. The basic structure of the pattern of projected warming has been robust to changes in models since the TAR, as indicated by comparison with Hulme et al. (2001).

To date there is insufficient evidence from RCMs to modify the large scale temperature projections from GCMs, although Tadross, et al (2005b) project changes in the A2 scenario for southern Africa that are near the low end of the spread in the AR4 global models, likely due to a weaker drying tendency than in most of the global models.

The observed rate of warming over the African continent is generally consistent with the model consensus, as shown in Figure 11.3.2.2. As is true for most regions, one can predict rather accurately the ensemble mean temperature response in other time periods, and for the A2 and B1 scenarios, from these temperature responses for A1B in the 2080–2099 time frame by a simple linear rescaling according to the ensemble mean global mean response. The signal/noise ratio is very large for 20 year mean temperatures. Using the models' internal variability, the A1B temperature change over the 21st century, and the assumption of a linearly growing signal in time, 10 years is typically adequate to obtain a clearly discernible signal, as indicated in Table 11.2.

[INSERT FIGURE 11.3.2.2 HERE]

11.3.2.4.2 Mean precipitation

Figure 11.3.2.1 and Table 11.2 also illustrate some of the robust aspects of the precipitation response over Africa in the AR4 models. The middle panels in the Figure show the percent change in precipitation averaged over the ensemble of models, once again between years 2080–2099 of the A1B scenario and the years 1980–1999 of the 20C3M historical integrations. The lower panels show the number of models (out of 21) that predict moistening at a particular location. The fractional changes in annual mean precipitation in each of these 21 models is provided in Supplementary material Figure S11.3.2.2. With respect to the most robust features (drying in the Mediterranean and much of Southern Africa, and increases in rainfall in East Africa) there is qualitative agreement with the results in Hulme et al. (2001) and 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 a plausible hydrological response to a warmer atmosphere, a consequence of the increase in water vapour and the resulting increase in vapour transport in the atmosphere from regions of moisture divergence to regions of moisture convergence (see Chapter 9 and Section 11.3.2.1).

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 down the West coast as far as 15°N. The processes involved include increased moisture divergence as well as a systematic poleward shift of the storm tracks affecting the winter rains, with positive feedback from decreasing soil moisture in summer (see Section 11.3.3).

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 circulation 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 occurs in the spring (September-October-November) and is mirrored in some RCM simulations for this region (see below), and roughly speaking can be thought of as a delay in the onset of the rainy season. This springtime drying contributes to the springtime maximum in the temperature response in this region, 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 21 models projecting an increase in the core of this region, east of the Great Lakes. This East African increase was also evident in the TAR models. The Guinean coastal rain belts and the Sahel do not show as robust a response. (The ensemble mean increase at 20°N in the East Sahara is generated by 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. The composite model has a weak drying trend in the Sahel in the 20th century that does not continue in the future projections (Hoerling, et al 2006), implying that the 20th century drying trend in the composite model is very likely forced by aerosols and not greenhouse gases.

Individual models generate large, but disparate, responses in the Sahel. Two interesting outliers are GFDL/CM2.1, which projects very strong drying in the Sahel and throughout the Sahara, and MIROC3.2_midres which shows a very strong trend towards increased rainfall (see Supplementary Figure

S11.3.2.2). Vizy and Cook (2005) find moderately realistic interannual variability in the Gulf of Guinea and Sahel in both models. While the drying in the GFDL model is extreme within the ensemble, it generates a plausible simulation of 20th century Sahel rainfall trends (Held et al., 2005, Hoerling, et. al, 2006). More research is clearly needed to understand the variety of modelled precipitation responses in the Sahel and elsewhere in the tropics. Progress is being made in developing new methodologies for this purpose (e.g., Chou and Neelin, 2004; Lintner and Chiang, 2005; Chou et al , 2006), leading to better appreciation of the sources of model differences.

It has been argued (e.g., Paethe and Hense, 2004) that the partial amelioration of the Sahel drought since the 90's may be a sign of a greenhouse-gas driven increase in rainfall, providing support for those models that moisten the Sahel into the 21st century (e.g., Kamga et al., 2005; Haarma et al , 2005). Although the mechanism of Haarma et al. (2005) is consistent with the warming projected by most of the models leading to robust conclonsion in temperature increase, our view is that it is premature to take this partial amelioration as evidence of a global warming signature, given the likely influence of decreasing aerosol forcing and internal variability on inter-hemispheric SST gradients, and, through these gradients, on Sahel rainfall.

As one moves northwards in the Sahara, one eventually enters the latitudes to which the Mediterranean drying penetrates robustly (see Figure 11.3.2.1). 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.

Table 11.2 provides information on the spread of model projections for the fractional change in precipitation in the 4 African sub-regions. The regions/seasons for which the central half (25–75%) of the projections are uniformly of one sign are EAF where there is an increase in DJF, MAM, SON, and annual mean, SAF where there is a decrease in austral winter and spring, and SAH where there is a decrease in boreal winter and spring. The Tibaldi et al (2004, 2005) pdfs estimates (Supplementary material Table S11.3) do not change this distinction between robust and non-robust regions/seasons. The time required for emergence of a clearly discernible signal in these robust regions/seasons is typically 50–100 years, except in the Sahara where even longer times are required.

Land use changes cannot be ignored as a potential contributor to drying in the 21st century. Taylor et al. (2002) project drying over the Sahel of 4% between 2015 and 1996 due to changing land use, but suggest that the magnitude could grow substantially further into the century. Maynard and Royer (2004a) suggest that estimated land use change scenarios for the mid 21st century would have only a modest compensating effect on the greenhouse gas induced moistening in their model. In neither of these studies is there a dynamic vegetation model.

Several regional climate change projections based on RCM simulations are available for southern Africa but are much scarcer for other regions. For example, Tadross et al. (2005b) examine two RCMs (PRECIS and MM5) nested for Southern Africa in the HadAM3H GCM for SRES A2. During summer, both models respond to the increase in high pressure systems entering from the west generated by the global model. During the early summer season, October-December, 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. The drying in the west continues into late summer, but there are increases in total rainfall towards the east in January and February, a feature barely present in the consensus AR4 global model. Given the variety of responses in Southern Africa among the AR4 models (Supplementary material Figure S11.3.2.2), downscaling of a larger range of models will be needed to assess the robustness of the new information provided by the regional models.

Empirical downscaling of projections has been pursued by Hewitson and Crane (2006), who provide projections for daily precipitation as a function of 6 GCM simulations of climate change (3 from the TAR, 3 from the AR4/PCMDI archive). The downscaled results for the SRES A2 emissions scenario near the end of the 21st century, show convergence in broad scale patterns and in some spatial details, suggesting more commonality in GCM projected changes in daily circulation, on which the downscaling is based, than in the GCM precipitation responses. Figure 11.3.2.3 shows the response 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 pattern of change, with some differences in magnitude. The consensus of these downscaling results shows increased precipitation in east Africa 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. There is also drying along the Mediterranean coast, and, in most models, drying in the western portion of southern Africa. The downscaling also shows marked local scale variation in the projected changes, for example, the contrasting changes on the west and east of Madagascar, and on the coastal and inland borders of the Sahel.

[INSERT FIGURE 11.3.2.3 HERE]

While this result is generally consistent with the underlying GCM and the composite AR4 GCM projection, there is a clear tendency for greater Sahel drying than in the underlying GCM, 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

11.3.2.4.3 Extremes

Using the definition of "extreme" seasons given in Section 11.3.1.1, the probability of encountering extremely warm, wet, and dry seasons as estimated by the AR4 models is provided in Table 11.2. As in most tropical regions, all seasons are extremely warm by the end of the 21st century, with near certainty, in the A1B scenario.

Changes in extreme wet and dry seasons are not as dramatic as the changes in extreme warmth, but still substantial. We focus on the robust (colored) regions/seasons in Table 11.2, In East Africa, the frequency of extremely wet seasons ranges from 9% in JJA to 24% in DJF, with the frequency of extremely dry seasons generally decreasing. In South Africa, in contrast, the frequency of extremely dry austral winter and springs increases to 20%. Although the mean response in West Africa is less robust than in East Africa, the increase in the number of extremely wet seasons is comparable.

On shorter time scales, regional modelling and downscaling results (Tadross, et al, 2005b) both suggest some increase in the rainfall intensity in Southern Africa. In regions of drying, there is generally a proportionally larger decrease in the number of rain days, indicating some compensation between intensity and frequency of rain. In the downscaling results of Hewitson and Crane (2006) and Tandross et. al. (2005b), changes in the median precipitation event magnitude at the station scale do not always mirror the projected changes in seasonal totals.

With regard to tropical cyclones impacting the Southeast coast of Africa, there is little modelling guidance. The 20km global time-slice simulation by (Mizuta et al. 2005) indicates that intensive precipitation associated with these systems could increase, but the robustness of these results remains uncertain. Thermodynamic arguments for increases in intensity (see Chapter 10), are applicable to these Indian Ocean storms just as for other regions.

11.3.2.5 Robust conclusions and uncertainties

Conclusions about projected climate change for Africa (with types of evidence indicated according to Section 11.3.1) are:

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 2.
2. Annual rainfall is very likely to decrease in much of North Africa and Northern Sahara. Based on: 1 and 3.
3. Winter rainfall will very likely decrease in much of Southern Africa. Based on: 1, 2, and 3.
4. There will likely be an increase in annual mean rainfall in tropical and East Africa. Based on: 1, 2, and 3.

5. It is uncertain how rainfall in the Sahel and the Southern Sahara will evolve in this century. Based on: 1 and 2.

Major uncertainties concerning projected climate change for this region are:

- It is difficult to judge the consequences for climate responses of the systematic errors across the ensemble of global models (excessive rainfall in the south, southward displacement of Atlantic ITCZ, insufficient upwelling off the West Coast).
- 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 the global models. Land surface modification is also not taken into account in the projections.
- RCMs are still being developed for different African regions; experience as to the extent to which current models can successfully downscale precipitation is limited.
- Empirical downscaling schemes are conservative in character, and cannot capture changes in local feedback mechanisms.
- Absence of realistic variability in Sahel in most 20th century simulations casts doubt on the reliability of coupled models in this region.
- There is insufficient information on which to assess possible changes in the distribution of tropical cyclones impacting Africa, but thermodynamic arguments for increases in intensity are applicable here as in other regions.

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 increased water vapour transport from low to high latitudes (Box 11.1), a number of other factors may shape future climate changes in Europe and the Mediterranean area. Variations in the atmospheric circulation influence the European climate both on interannual and longer time scales. Recent examples include the central European heat wave in the summer 2003, characterized by a long period of anticyclonic weather (e.g., Fink et al., 2004), and the strong warming of winters in northern Europe from the 1960’s to 1990’s and the simultaneous decrease in winter precipitation in the Mediterranean area that were both affected by an upward trend in the NAO (e.g., Hurrell and van Loon, 1997; Räisänen and Alexandersson, 2003; Xoplaki et al., 2004; Scaife et al., 2005). On fine geographical scales the effects of atmospheric circulation are modified by topography particularly in mountainous areas (Bojariu and Giorgi, 2005).

Europe, particularly its northwestern parts, owes its relatively mild climate partly 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 models (see Chapter 10, Section 10.3), this will act to reduce the warming in Europe but is in the light of our present understanding very unlikely to reverse the warming to cooling (see Section 11.3.3.3.1).

Local thermodynamic factors also affect the European climate and are potentially important for its future changes. In the northeastern 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. 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 (Black et al., 2004; 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 (Supplementary material Table S11.2). Excluding one model with extremely cold winters in northern Europe, the seasonal area mean temperature biases in NEU vary from –5°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 large 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 (Supplementary material Table S11.2), but the interpretation of the difference is complicated by the observational uncertainty associated with the undercatch of, in particular, solid precipitation (e.g., Adam and Lettenmaier, 2003). In summer, most models simulate too little precipitation, particularly in the eastern parts of the area. In 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 well in many but not all current AOGCMs. 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. The resulting biases in the near-surface atmospheric flow may explain a substantial fraction of the biases in temperature and precipitation (van Ulden and van Oldenborgh, 2005).

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

The ability of RCMs to simulate climate extremes in Europe has been addressed in several studies. In the PRUDENCE simulations (Box 11.2), the biases in the tails of the temperature distribution were generally larger than the biases in average temperatures (Kjellström et al., 2006). The biases also varied substantially between the RCMs, not only in magnitude but in most parts of Europe also in sign. Inspection of the individual models showed some similarity between the biases in daily and interannual variability, suggesting that similar mechanisms may be affecting both.

The magnitude of precipitation extremes in RCMs is model-dependent. In a comparison of the PRUDENCE RCMs, Frei et al. (2006) found the area mean 5-year return values of maximum one-day precipitation in the vicinity of the European Alps to vary by up to a factor of two between the models. However, except for too low extremes in the southern parts of the area in summer, the set of models as a whole showed no systematic tendency to over- or underestimate the magnitude of the extremes. The models also showed skill in simulating the mesoscale patterns of extreme precipitation within the topographically complicated Alpine area. A similar level of skill has been found in 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).

Evidence of model skill in simulation of wind extremes is mixed. Weisse et al. (2005) found an RCM to simulate a very realistic wind climate over the North Sea, including the number and intensity of storms, when driven by analysed boundary conditions. However, most PRUDENCE RCMs, while quite realistic over sea, severely underestimate the occurrence of very high wind speeds (17.2 m/s or more) over land and coastal areas (Rockel and Woth, 2006). 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 the two models that had a gust parameterization.

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 uncertainty in these projections. Four sources of uncertainty were studied: (i) *Sampling uncertainty* due to the fact that model climate is estimated as an average over a finite number (30) of years, (ii) *Regional model uncertainty* due to the fact that regional climate models use different techniques to discretize the equations and to represent sub-grid effects, (iii) *Emission uncertainty* due to choice of 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. More details are provided in e.g. Christensen et al. (2006), Déqué et al., 2005) 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[b]			1		1	1	1				
ECHAM4 +A2				1	1						
ECHAM4 +B2				1	1						
ARPEGE +A2[a]		1									
ARPEGE +B2[b]		3									

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

11.3.3.3 Climate projections

11.3.3.3.1 Mean temperature

The observed evolution of European temperatures in the 20th century, characterised by a warming trend modulated by multidecadal variability, was well within the envelope of the AR4 simulations (Figure 11.3.3.1).

In this century, the warming is projected to continue at a rate somewhat greater than its global mean, with temperatures rising above the background of natural variability within the next few decades Table 11.2. Under the A1B scenario, the simulated area and annual mean warming from 1980–1999 to 2080–2099 varies from 2.3 to 5.3°C in NEU and from 2.2 to 5.1°C in SEU, with a mean (median) of 3.6°C (3.2°C) in NEU and 3.4 (3.5°C) in SEU. Ensemble mean temperature changes for other periods and emissions scenarios scale approximately linearly with the global mean warming (Supplementary material Figures S11.3.1.2-4).

[INSERT FIGURE 11.3.3.1 HERE]

In northern Europe, particularly its northeastern parts, the warming is likely to be largest in winter, in the Mediterranean area in summer (Figure 11.3.3.2). Seasonal mean temperature changes typically vary by a factor of three among the AR4 models Table 11.2; however the very high upper end of the range in NEU in DJF (8.1°C) is reduced to 6.7°C when one model with an extreme cold bias in present-day winter climate is excluded. The probabilistic scheme of Tebaldi et al. (2005) suggests 5–95% uncertainty ranges slightly narrower than the full range of the model results, with a larger difference in the upper than in the lower end of the range (Figure 11.2.1, Supplementary material Figure S11.2.1, and Supplementary material Table S11.3).

[INSERT FIGURE 11.3.3.2 HERE]

Although changes in atmospheric circulation have a significant potential to affect temperature in Europe (e.g., Dorn et al., 2003), they are not the main cause of the projected anthropogenic warming (e.g., Rauthe and Paeth, 2004; van Ulden et al, 2006; Stephenson et al., 2006). For example, van Ulden and van Oldenborgh (2005) estimated the contribution of circulation changes for western central Europe using a regression method and seven AOGCM simulations of climate change from 1971–2000 to 2071–2100 under the SRES A2 scenario. In most models, circulation changes enhanced the warming 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 circulation contribution typically ranged from –1°C to 1.5°C. Most of the warming, 1–5°C depending on model and season, was unrelated to the circulation.

Most AOGCMs simulate a decrease in the North Atlantic THC with increasing greenhouse gas forcing (see Chapter 10, Section 10.3). In spite of this, all the AR4 simulations indicate warming in all of Europe, as the direct atmospheric effects of increased greenhouse gases dominate over the changes in ocean circulation. The same is true for earlier increased greenhouse gas simulations except for a very few (Russell and Rind, 1999; Schaeffer et al., 2004) that have showed slight cooling along the northwestern coastlines of Europe but warming over the rest of the continent. The impact of THC changes depends on the regional details of the change, being largest if ocean convection is suppressed in high latitudes where the sea-ice feedback may amplify atmospheric cooling (Schaeffer et al., 2004). AOGCM sensitivity studies with an artificial shutdown of the THC, with no changes in greenhouse gas concentrations, indicate a 1–3°C annual mean cooling in Europe (e.g., Manabe and Stouffer, 1997; Vellinga and Wood, 2002), with possibly larger cooling in the extreme northwestern parts (Rind et al., 2001).

Various SDMs have been used to derive projections of local temperature change, applying data from several AOGCMs including the AR4 models, especially for northern Europe (e.g., Benestad, 2005; Hanssen-Bauer et al., 2003, 2005). These studies have shown a similar large-scale warming as dynamical models, but with finer-scale regional details. For example, Hanssen-Bauer et al. (2005) found that, in most of Scandinavia, the warming during the 21th century would increase with distance from the coast and with latitude. Comparing RCM and SDM projections downscaled from the same GCM, Hanssen-Bauer et al. (2003) found the largest differences between the two approaches in winter and/or spring at localities with frequent temperature inversions in the present climate. A larger warming at these localities in the SDM projections was found consistent with increased winter wind speed in the driving GCM and reduced snow cover, both of which disfavour ground inversions.

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 simulations, the interannual summertime temperature variability in central Europe doubled from 1961–1990 to 2071–2100 under the A2 scenario, while others showed almost no change (Schär et al., 2004; Vidale et al., 2006). Possible reasons for the increase in temperature variability are reduced soil moisture, which reduces the capability of evaporation to damp temperature variations, and increased land-sea contrast in average summer temperature (Rowell, 2005; Lenderink et al., 2006). In qualitative agreement with these RCM results, most of the AR4 simulations indicate the interannual standard deviation of summer mean temperature to increase in both northern Europe and the Mediterrenean area (Giorgi and Bi, 2005). The increased variability may have played a role in producing the European heatwave in summer 2003 (Schär et al., 2004). The PRUDENCE simulations suggest that temperature conditions similar to those observed in 2003 may occur in an average summer in the late 21st century (Beniston, 2004).

Kjellström et al. (2006) analysed daily temperature variability in the PRUDENCE simulations and found the intermodel differences in the simulated change to increase towards the extreme ends of the 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 by Shkolnik et al. (2006). These RCM results are supported by GCM studies of Hegerl et al. (2004) and Meehl and Tebaldi (2004).

[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 et al., 2004; Giorgi and Bi, 2005; Rowell, 2005) and daily time scales (Hegerl et al., 2004; Kjellström et al., 2006). In the PRUDENCE simulations, the lowest winter minimum temperatures increased more than the median minimum temperature especially in eastern, central and northern Europe, although the magnitude of this change was strongly model-dependent (Figure 11.3.3.3). The geographical patterns of the change indicate a connection to reduced snow cover, with a large warming of the cold extremes where snow retreats but a more moderate warming in southwestern Europe which is mostly snow-free even today (Rowell, 2005; Kjellström et al., 2006). Reduced temperature variability in Europe in winter is consistent with long-term observed trends (Yan et al., 2002).

Along with the overall warming, the number of frost days is very likely to decrease. In the PRUDENCE simulations under the A2 scenario, the largest absolute decreases of about 60 days per year occurred in northern and eastern Europe and in the Alps (Jylhä et al., 2006), whereas larger relative decreases occurred further southwest. The same study also indicated a general decrease in the number of days with temperature intersecting 0°C, except for northernmost Europe where fewer such days were simulated 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 1980–1999 to 2080–2099 in the AR4 A1B simulations varies from 0 to 16% in NEU and from –4% to –27% in SEU (Table 11.2). The largest increases in northern and central Europe are simulated in winter. In summer, the NEU area mean changes vary in sign between models, although most models simulate increased (decreased) precipitation north (south) of about 55°N. In SEU, the most consistent and in per cent terms largest decreases occur in summer, but the area mean winter precipitation also decreases in most models. More detailed seasonal statistics are given in Table 11.2; the 5–95% uncertainty ranges from the Tebaldi et al. (2005) method are similar to or slightly narrower than the full range of the model results (Supplementary material Table S11.3). Note that increasing evaporation makes the simulated decreases in annual precipitation minus evaporation to extend a few hundred kilometres further north in central Europe than decreasing precipitation (Supplementary material Figure S11.3.1.1).

Changes in precipitation may vary substantially on relatively small horizontal scales, particularly in areas of complex topography. However, the details of this variation depend on changes in the atmospheric circulation, as shown in Figure 11.3.3.4 for two PRUDENCE simulations that only differ with respect to the driving global model. In one of these, an increase in westerly flow from the Atlantic Ocean (caused by a large 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 other simulation, with little change in the average pressure pattern, the increase is only 0–10%. When compared with circulation changes in the more recent AR4 simulations, these two cases fall in the opposite ends of the range. Most AR4 models indicate increased north-south pressure gradient across northern Europe, but the change is generally smaller than in the top row of Figure 11.3.3.4.

[INSERT FIGURE 11.3.3.4 HERE]

The importance of circulation changes was also demonstrated by van Ulden and van Oldenborgh (2005), who studied precipitation changes in western central Europe in seven AR4 AOGCMs. They found that increases in winter precipitation were in most models enhanced by increased westerly winds, whereas the general decrease in summer precipitation was largely due to a more easterly and anticyclonic flow type. The residual precipitation change that was unexplained by changes in circulation varied much less with season and (with the exception of summer) between models than the actual precipitation change. For most months and models, the residual change from 1971–2000 to 2071–2100 was a modest increase of 0–15%, consistent with the increased moisture transport capacity of a warmer atmosphere.

Rowell and Jones (2006) used a regional version of the HadAM3P model to isolate the mechanisms that led to reduced summer precipitation in the global version of the same model in southern and central Europe. Although they found changes in the 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. Because changes in 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.

SDM based projections of precipitation change in Europe tend to support the large-scale picture from dynamical models (e.g., Busuioc et al., 2001a; Beckmann and Buishand, 2002; Hanssen-Bauer et al., 2003, 2005; Benestad, 2005; Busuioc et al., 2006), although variations between SDM methods and the dependence on the GCM data sets used (see Section 11.2.1.1.2) make it difficult to draw quantitative conclusions. However, SDMs have suggested a larger small-scale variability of precipitation changes than indicated by GCM and RCM results, particularly in areas of complex topography (Hellström et al., 2001).

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

Time scale also matters. Although there are some indications of increased interannual variability particularly in summer precipitation (Räisänen, 2002; Giorgi and Bi, 2005; Rowell, 2005), changes in long-term (monthly to annual) extremes are generally expected to follow the changes in mean precipitation more closely than those in short-term extremes (Räisänen, 2005).

An illustration of the possible characteristics of precipitation change, based on Frei et al. (2006), is given in Figure 11.3.3.5. The eight models in this PRUDENCE study indicated an increase in mean precipitation in winter both in southern Scandinavia and central Europe, due to both increased wet day frequency and increased mean precipitation for the wet days. In summer, a decrease in the number of wet days led to a decrease in mean precipitation particularly in central Europe. Changes in extreme short-term precipitation were broadly similar to the change in average wet-day precipitation in winter. In summer, extreme daily precipitation increased in most models despite the decrease in mean precipitation, but the magnitude of the change was highly model-dependent. Note that this study only covered the uncertainty associated with the choice of the RCM, not those associated with the driving GCM and the emissions scenario.

[INSERT FIGURE 11.3.3.5 HERE]

Much larger changes are expected in the recurrence frequency of precipitation extremes than in the magnitude of extremes. For example, Frei et al. (2006) estimated that, in Scandinavia under the A2 scenario, the highest 5-day winter precipitation totals occurring once in 5 years in 2071–2100 would be similar to those presently occurring once in 8–18 years (the range reflects variation between the PRUDENCE models). Analysing another RCM simulation, 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. In the AR4 simulations, large increases occur in the frequencies of both high winter precipitation in northern Europe and low summer precipitation in the Mediterranean area (Table 11.2).

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

The decrease in precipitation together with enhanced evaporation in spring and early summer is very likely to lead to reduced summer soil moisture in the Mediterranean region and parts of central Europe (e.g., Douville et al., 2002). In northern Europe, where increased precipitation competes with earlier snowmelt and increased 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 northwestern Europe.

A key factor are the changes in the large-scale atmospheric circulation. Simulations with increased north-south pressure gradient across northern Europe (e.g., top of Figure 11.3.3.4) tend to indicate stronger winds in northern Europe, both because of the larger time-averaged pressure gradient and a northward shift in cyclone activity. Conversely, the northward shift in cyclone activity tends to reduce windiness in the Mediterranean area. Such a change in the pressure pattern, resembling a shift towards the positive phase of the NAO, occurs in some form in most current AOGCM simulations (see Chapter 10, Section 10.3), but there are also simulations from which this change is largely absent. The HadAM3H simulations used to drive most PRUDENCE RCMs (e.g., bottom of Figure 11.3.3.4) exemplified the latter. Thus, these RCM simulations only showed relatively small changes in windiness, although the changes varied seasonally and included a tendency towards increased average and extreme wind speeds in western and central Europe in winter (Räisänen et al., 2004; Beniston et al., 2006; Leckebusch et al., 2006; Rockel and Woth, 2006).

Extreme wind speeds in Europe are mostly associated with strong winter cyclones (e.g., Leckebush and Ullbrich, 2004), the occurrence of which is only indirectly related to the time mean circulation. Nevertheless, models suggest a general similarity between the changes in average and extreme wind speeds (Knippertz et al., 2000; Räisänen et al., 2004). A caveat to this conclusion is that, even in most RCMs, the extremes of wind speed over land tend to be too low (see Section 11.3.3.2).

11.3.3.3.6 Mediterranean cyclones

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

11.3.3.3.7 Snow and sea-ice

Increased melting and decreased fraction of solid precipitation due to warmer climate will very likely reduce the amount of snow and the length of the snow season in most if not all of 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. (2006) found the average annual number of days with snow cover in northern Europe (55–75°N, 4–35°E) to decrease by 43–60 from 1961–1990 to 2071–2100 under the A2 scenario. The average DJF mean snow water equivalent decreased by 45–60%. Further south, smaller absolute but larger relative decreases occurred in both quantities. Results from other studies (e.g., Rowell, 2005) are qualitatively similar. Snow conditions in the coldest parts of Europe, such as northern Scandinavia and northwestern Russia (Räisänen et al., 2003; Shkolnik et al., 2006) and the highest peaks of the Alps (Beniston et al., 2003) appear to be less sensitive to the temperature and precipitation changes projected for this century than those at lower latitudes and altitudes (see also Box 11.3).

The Baltic Sea is likely to lose a large part of its seasonal ice cover during this century. Based on temperature changes simulated by six AOGCMs, Jylhä et al. (2006) estimated 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. In simulations with a regional atmosphere-Baltic Sea model (Meier et al., 2004), 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).

11.3.3.8 Robust conclusions and uncertainties

Conclusions about projected climate change for Europe (with types of evidence indicated according to Section 11.3.1) are:

1. Annual mean temperatures in Europe are likely to increase at a rate somewhat greater than the global mean. In northern Europe, warming is likely to be largest in winter, and in the Mediterranean area in summer. Based on: 1, 2, and 3. The uncertainty in the Atlantic THC suggests, however, a small (less than 10%) possibility of cooling in extreme 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 3.
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, and 3.
4. Extremes of daily precipitation will very likely increase in northern Europe. Based on: 1, 2, and 3, 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, and 3.
6. Risk of summer drought is likely to increase in central Europe and in the Mediterranean area, because of reduced summer precipitation and increased spring evaporation. Based on: 1, 2, 3, and process studies (increasing saturation deficit with increasing temperature).
7. It is uncertain whether and how wind storm frequency and/or intensity will change, although a majority of evidence suggests increased wind speeds in northern Europe. Based on: 1.
8. Snow season length and snow depth are very likely to decrease in most of Europe. Based on: 1, 2, and 3.

Although many features of the simulated climate change in Europe and the Mediterranean area are qualitatively consistent between models and qualitatively well-understood in physical terms, substantial uncertainties remain. Simulated seasonal mean temperature changes vary even on the subcontinental scale by a factor of 2–3 among the current generation of AOGCMs. Similarly, while agreeing on a large-scale increase in winter-half-year precipitation in the northern and decrease in summer-half-year precipitation in the southern parts of the area, models disagree on the magnitude and geographical details of precipitation change. 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 THC. 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

As monsoons are the dominant phenomena over much of Asia, the factors that influence the monsoonal flow and precipitation are of central importance for understanding climate change in this region. Precipitation is affected both by the strength of the monsoonal flows and the amount of water vapor carried by the flow. Monsoonal flows and the tropical large-scale circulation often weaken in global warming simulations, a counterintuitive result that is understandable from the reasoning of Knutson and Manabe (1995). But there is an emerging consensus that the effect of enhanced moisture convergence in a warmer moister atmosphere dominates over any such weakening of the circulation, resulting in increased monsoonal precipitation (Douville et al., 2000; Giorgi, et. al., 2001ab; Stephenson et al., 2001).

There is an association of the phase of ENSO with the strength of the summer monsoons (Pant and Rupa Kumar, 1997), so changes in ENSO will have an impact on these monsoons. Indeed there is evidence of secular variation in the ENSO/South Asian monsoon connection (Krishna Kumar et al., 1999; Sarkar et al., 2004; see Chapter 3, Section 3.7). Moreover, there is a link between Eurasian snow cover and the strength of the monsoon (see Chapter 3, Section 3.7) which might tend to strengthen the monsoon if snowcover retreats. 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 (e.g., Chen et al., 2004), are additional issues. However, although aerosol effects may have been large as compared to the impacts of changing greenhouse forcing in the 20th century, most emission scenarios suggest that future changes in regional climate will be dominated by increasing greenhouse forcing rather than changes in sulphate and absorbing aerosols.

For South Asia, the monsoon depressions and tropical cyclones generated over the Indian seas modulate the monsoon anomalies. For East Asia, the monsoonal circulations are strengthened by extratropical cyclones energized in the lee of the Tibetan plateau and by the strong temperature gradient along the East Coast. ENSO's influence on the 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), suggesting that spatial structure of the warming in the Pacific will be relevant for changes in these features. The Meiyu-Changma-Baiu rains in the early summer, which derive from disturbances of baroclinic character but are 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.

Issues related to monsoonal controls are also central 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 (e.g., 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). The pattern of ocean temperature change across the Pacific will be of centrasl importance to climate change in this region.

In *Central Asia*, including the *Tibetan Plateau*, the temperature response to greenhouse gas increases is strongly influenced by changes in winter and spring snowcover, the isolation from maritime influences, and diffusion of the larger wintertime Arctic warming into the region by eddies. With regard to precipitation, a key issue is related to the moisture transport in summer penetrating eastward through the southern rim of Central Asia (from Iran to Pakistan), and from the northwest during winter. The same processes control winter precipitation over the northern part of South Asia and Tibet. How far the drying of the Mediterranean in global warming simulations penetrates into these regions is likely to be strongly dependent on accurate simulation of these sources of moisture. The dynamics of climate change in the *Tibetean Plateau*, and also downstream over East China, are further complicated by the high altitude of this region and its complex topography with large elevation differences.

11.3.4.2 Skill of models in simulating present climate

Simulated regional mean temperature and precipitation in the AR4 AOGCMs show biases when compared with observed climate (Supplementary material Table S11.2). The model mean shows a cold and wet bias in all regions and in most seasons. The annual average bias ranges from –3.2ºC over the Tibetan Plateau to -0.5ºC over South Asia. For most regions there is a 6–7ºC range in the biases from the individual models. For Southeast Asia the range is 3.3ºC. The mean bias in precipitation is small (less than 10%) in Central Asia, Southeast Asia, and South Asia, larger for Northern Asia and East Asia (around +24%), and very large for the Tibetan Plateau (+120%). Annual biases in individual models are in the range of –50% to +60% across all regions except the Tibetan plateau, where some models show annual precipitation three times the observed and even larger seasonal biases occur in winter and spring. These global models clearly have limited credibility over Tibet, due to the difficulty in simulating the effects of the dramatic tropographic relief. The consistent cold bias throughout the continent is also of concern, especially if futher research suggests distorted albedo feedbacks due to excessive snowcover.

South Asia

Over South Asia, the summer is dominated by the southwest monsoon, which spans the four months June through September, and dominates the seasonal cycles of precipitation, temperature, wind and many other climatic parameters. 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 (Lal and Harasawa, 2001, Rupa Kumar and Ashrit, 2001, Rupa Kumar et al., 2003). This are likely linked to the coarse resolution of the models as the heavy rainfall over these regions is generally associated with the steep orography. However, the simulated annual cycles in South Asian mean precipitation and surface air temperature are reasonably close to the observed (Figure 11.3.4.1). The AR4 models capture the gross regional features of the monsoon such as low rainfall amounts coupled with high variability over northwest India. However, there has not yet been sufficient analysis of whether finer details of regional significance, which were not represented in some of the earlier models analysed by Rupa Kumar et al. (2002), are simulated more adequately in the AR4 models.

Recent work indicates that time slice experiments using atmospheric GCMs with prescribed SSTs are not able to accurately capture the South Asian monsoon response simulated in a coupled system (Douville, 2005). Thus, neglecting the high-frequency SST feedback and variability seems to have a significant impact on the projected monsoon response to global warming, complicating the regional downscaling problem. 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 (Douville, 2005).

INSERT FIGURE 11.3.4.1 HERE]

The Hadley Centre's regional climate model PRECIS has recently been used to simulate the South Asian climate with a horizontal resolution of 50 km. Three-member ensembles of baseline simulations (1961–1990) have confirmed that significant improvements in the representation of regional processes over South Asia can be achieved (Rupa Kumar et al., 2006). For example, the steep gradients in monsoon precipitation with a maximum along the western coast of India are well-represented in PRECIS. 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

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.

East Asia

Simulated temperatures in most AR4 models are too low in all seasons over East Asia; the mean cold bias is largest in winter and smallest in summer (Supplementary material Table S11.2) The annual precipitation exceeds the observed estimates in almost all models and the rain band in mid-latitudes is shifted northward in seasons other than summer. This bias in the placement of the rains in Central China also occurred in earlier models (e.g., Gao et al., 2001; Gao et al., 2004). In winter, the area mean precipitation is overestimated by over 50% on average due to strengthening of the rain band associated with extrratropical systems over Southern China. The bias and inter-model differences in precipitation are smallest in summer but the northward shift of this rain band results in large discrepancies in summer rainfall distribution over Korea, Japan and adjacent seas. In summer, the Northwest Pacific High is typically stronger than observed and this could lead to the premature northward shift of the rains, resulting in the precipitation deficit in this area.

Kusunoki et al. (2006) find that the simulation of these Meiyu-Changma-Baiu rains in the East Asian Monsoon is improved substantially with increasing horizontal resolution. Confirming the importance of resolution, RCMs simulate more realistic climatic characteristics over East Asia than AOGCMs (e.g., Ding et al. 2003; Oh et al. 2004; Sasaki et al. 2005; Fu et al. 2005; Gao et al. 2006). Several studies reproduce the fine-scale climatology of small areas using a multiply-nested RCM and a very high resolution RCM (Yasunaga et al. 2006). Gao et al. (2006) reported that simulated East Asia large-scale precipitation patterns are very significantly affected by resolution, particularly during the mid to late monsoon months, when smaller scale convective processes dominate. Figure 11.3.4.2 shows the spatial correlation between the simulated and observed annual mean precipitation from the simulations of Gao et al. (2006). In general, the correlation increases with increasing resolution.

[INSERT FIGURE 11.3.4.2 HERE]

Southeast Asia

The broadscale spatial distribution of temperature and precipitation in DJF and JJA averaged across the AR4 models 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. (2004c) 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 was poorest in the Southeast Asian portion of the domain, where observed SST- rainfall links were often reversed in the models. This represented a shortcoming in model processes that is likely to be relevant to the reliability of enhanced greenhouse simulations. Since current AOGCMs continue to have some significant shortcomings in representing ENSO variability (see Chapter 8, Section 8.4), the difficulty of projecting changes in ENSO-related rainfall in this region are compounded.

Rainfall simulation across the region at finer scale has been examined in some studies. McGregor et al. (1998) reported that a ten-year regional simulation with DARLAM at 44 km 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 80 km 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. (2004b) 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.

Central Asia and Tibet

Due to the complex topography and the associated meso-scale weather systems of the high altitude and arid areas, GCMs typically perform poorly over the region. Importantly, they tend to overestimate the precipitation over arid and semi arid areas in the north (e.g., Small et al., 1999; Gao et al., 2001.)

Over Tibet, the few available RCM simulations generally exhibit improved performance in the simulation of present day climate compared to GCMs (e.g., Gao et al., 2003a, b; Zhang et al., 2005a). The GCM simulation of Gao et al. (2003a) overestimated the precipitation over the northwestern Tibetan Plateau by a factor of 5–6, while in an RCM nested in this model the overestimate was less than a factor of 2. Due to the lack of observation data, complex topography, and a large portion of solid precipitation, observations could substantially underestimate the true precipitation in this area.

11.3.4.3 Climate projections

11.3.4.3.1 Temperature

The temperature projections for the 21st century based on AR4 AOGCMs (Figure 11.3.4.3 and Table 11.2) represent a significant acceleration of warming over that observed in the 20th century. Warming is least rapid, similar to the global mean warming, in Southeast Asia (mean warming from 1980–1999 to 2080–2099 2.6°C under the A1B scenario), stronger over South Asia (3.2°C) and East Asia (3.4°C) and greatest in the continental interior of Asia (3.8°C in Central Asia, 4.0°C in Tibet and 4.5°C in Northern Asia). In four out of the six regions, the largest warming occurs in DJF, but in Central Asia the maximum occurs in JJA. In Southeast Asia, the warming is nearly the same throughout the year. Model to model variation in warming is typically about three quarters of the mean warming (e.g., 2.0–4.7°C for annual mean warming in South Asia). The 5–95% ranges based on Tebaldi et al. (2005) suggest a slightly smaller uncertainty than the full range of the model results (Supplementary material Table S11.3).

[INSERT FIGURE 11.3.4.3 HERE]

Because the projected warming is large compared to interannual temperature variability, a large majority, or in some parts of Asia virtually all, individual years and seasons in the late 21st century are likely to be extremely warm by present standards (Table 11.2). The projections for changes in mean temperature and, where available, temperature extremes, are discussed below in more detail for individual Asian regions.

South Asia

For the A1B scenario, the AR4 models indicate an increase of 2.0–4.7ºC in annual mean temperature in the region by the end of the 21st century, with half of the models in the range 2.7–3.6ºC and a median of 3.3ºC (Table 11.2). The median warming varies seasonally from 2.7ºC in JJA to 3.6ºC in DJF. The warming is likely to increase northward in the area, particularly in winter, and from sea to land (Figure 11.3.4.4). Studies based on earlier AOGCM simulations (Douville et al., 2000; Lal and Harasawa, 2001; Lal et al., 2001; Rupa Kumar and Ashrit, 2001; Rupa Kumar et al., 2002, 2003; Ashrit et al., 2003; May, 2004b) support this picture. The tendency of the simulated warming to be more pronounced during winter and post-monsoon months compared to the rest of the year is also a conspicuous feature of the observed temperature trends from instrumental data analyses over India (Rupa Kumar et al., 2002, 2003).

[INSERT FIGURE 11.3.4.4 HERE]

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

East Asia

For the A1B scenario, the AR4 models indicate an increase of 2.3–4.9ºC in annual mean temperature in EAS by the end of the 21st century, with half of the models in the range 2.8–4.1ºC and a median of 3.3ºC (Table

11.2). The median warming varies seasonally from 3.1ºC in JJA to 3.6ºC in DJF. The warming tends to be largest in winter, especially in the northern inland area (Figure 11.3.4.4) but the area mean difference from the other seasons is not large. There is no obvious relationship between model bias and the magnitude of the warming. The ensemble median change of annual mean temperature based on the higher SRES A2 scenario is 4.3°C, similar to the earlier model result of Min et al. (2004). The spatial pattern of larger warming over northwest EAS (Figure 11.3.4.4.) is very similar to the ensemble mean of the earlier models. RCM simulations show mean temperature increases similar to that in AOGCMs (Gao et al., 2001; 2002; Kwon et al., 2003; Kanada et al., 2005; Xu et al 2005;).

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

Southeast Asia

For the A1B scenario, the AR4 models indicate an increase of 1.5–3.7ºC in annual mean temperature in SEA by the end of the 21st century, with half of the models in the range 2.3–3.0ºC and a median of 2.5ºC, with little seasonal variation (Table 11.2). Simulations by the DARLAM regional model (McGregor et al. 1998) and more recently by the CSIRO stretched grid model (McGregor and Dix, 2001) centred on the Indochina Peninsula (AIACC 2004, at a resolution of 14 km) have demonstrated the potential for significant local variation in warming, particularly the tendency for warming to be significantly stronger over the interior of the landmasses than over the surrounding coastal regions. A tendency for the warming to be stronger over Indochina and the larger landmasses of the archipelago is also visible in the AR4 models (Chapter 10, Figure 10.3.5 and Figure 11.3.4.4). As in other regions, the magnitude of the warming depends on the forcing scenario. 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.

Although few studies have been undertaken for Southeast Asia on how temperature variability and extremes may change, it seems very likely that the region would share in the global tendency for increased daily extreme high temperatures as the climate warms (see Chapter 10, Section 10.3).

Central Asia and *Tibet*

For the A1B scenario, the AR4 models indicate an increase of 2.6–5.2ºC in annual mean temperature in Central Asia by the end of the 21st century, with half of the models in the range 3.2–4.4ºC and a median of 3.7ºC (Table 11.2). The median warming varies seasonally from 3.2ºC in DJF to 4.1ºC in JJA. The 5th to 95th quantile range using the probabilistic approach of Tebaldi et al. (2004, 2005) is 2.2 to 4.5ºC in winter and 2.9 to 5.6ºC in summer. For the Tibetean Plateau, the corresponding range in annual mean warming is 2.8–6.1ºC, half of the models are within 3.2–4.5ºC and the median is 3.8ºC. The seasonal variation in the simulated warming in Tibet is modest, the median varying from 3.6ºC in MAM to 4.1ºC in DJF. The 5th to 95th quantile range using the probabilistic approach of Tebaldi et al. (2004, 2005) is 3.3 to 5.6ºC in winter and 2.8 to 5.0ºC in summer. Findings from earlier multi-model studies (Xu et al. 2003a,b; Meleshko et al., 2004) are consistent with the AR4 results.

An RCM study by Gao et al. (2003a) indicated greater warming over the Plateau compared to surrounding areas, with the largest warming at highest altitudes, e.g., over the Himalayas. The higher temperature increase over high altitude areas can be explained by the decrease in surface albedo associated with the melting of snow and ice (Giorgi et al., 1997). This phenomenon is found to different extents in some (e.g., the two versions of MIROC3.2) although not all (e.g., ECHAM5/MPI-OM) of the AR4 models, and it is visible in the multi-model mean changes particularly in the winter half-year (Figure 11.3.4.4).

11.3.4.3.2 Precipitation and associated circulation systems

The consensus of AR4 models indicates an increase in annual precipitation in most of Asia during this century, the relative increase being largest and most consistent between models in North and East Asia

(Figure 11.3.4.4, Table 11.2). The main exception is Central Asia, particularly its western parts, where most models simulate reduced precipitation in the summer half-year. Based on these simulations, sub-continental mean winter precipitation will increase very likely in Northern Asia and the Tibetan Plateau (where all models agree on an increase under the A1B scenario) and likely in Central, Southeast and East Asia (16 to 19 out of 21 models agree on an increase). Summer precipitation will likely increase in North, South, Southeastern, and East Asia (18 to 19 models agree on an increase) but decrease in Central Asia (17 models agree on a decrease). Probability estimates from Tebaldi et al. (2005) (Supplementary material Table S11.3) support these judgements.

The projected decrease in mean precipitation in Central Asia is accompanied by an increase in the frequency of very dry spring, summer and autumn seasons; conversely, where and when models project increases in the mean precipitation seasons with very high precipitation become more common (Table 11.2). Below, the projections for changes in mean precipitation and, where available, precipitation extremes, are discussed in more detail for individual Asian regions. Where appropriate, the connection to changes in precipitation-bringing circulation systems is also discussed. Where not specifically noted, the numeric values refer to changes from 1980–1999 to 2080–2099 under the A1B scenario. Smaller (slightly larger) changes are generally projected for the B1 (A2) scenario, but the inter-scenario differences are small compared with the inter-model differences.

South Asia

Most of the AR4 models project a decrease of precipitation in DJF (the dry season), and an increase during the rest of the year. The median change and the full range of the model results (in parentheses) under the A1B scenario in the end of the 21st century are –5% (–35% to 15%) in DJF, 11% (–3% to 23%) in JJA and 11% (–15% to 20%) in the annual mean (Table 11.2). The probabilistic method of Tebaldi et al. (2005) calculates a 90% confidence interval for winter of –32% to 23% and in summer –6% to 26%. Only 3 of the 21 models project a decrease in annual precipitation. This qualitative agreement on increasing precipitation is also supported by earlier AOGCM simulations (Lal and Harasawa, 2001; Lal et al., 2001; Rupa Kumar and Ashrit, 2001; Rupa Kumar et al., 2002, 2003; Ashrit et al., 2003; May, 2004b).

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

ECHAM4 time slice experiments indicate a general increase in the intensity of heavy rainfall events in the future, with large increases over the Arabian Sea and the tropical Indian Ocean, in northern Pakistan and northwest India as well as in northeast India, Bangladesh and Myanmar (May, 2004a). The regional climate model HadRM2 shows an overall decrease in the annual number of rainy days up to ~15 days over a large part of South Asia, under IS92a scenario in the 2050s, but an increase in the precipitation intensity as well as extreme precipitation (Krishna Kumar et al., 2003). 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., 2006).

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

East Asia

The consensus of AR4 models indicates an increase in precipitation in East Asia in all seasons. The median change and the full range of the model results (in parentheses) under the A1B scenario at the end of the 21st century are 10% (–4% to 42%) in DJF, 9% (–2% to 17%) in JJA and 9% (2% to 20%) in the annual mean (Table 11.2). Based on the probabilistic methods of Tebaldi et al. (2004, 2005), the 90% confidence interval for DJF is –11 to 24% and in summer 1% to 15%. In winter this increase contrasts with a decrease in precipitation over the ocean to the southeast, where reduced precipitation corresponds well with increased mean sea level pressure. These projections with a good qualitative agreement but large quantitative differences between the models are consistent with previous studies (e.g., Giorgi et al., 2001; Hu et al., 2003; Min et al., 2004).

The increase in rainfall in summer is associated with changes in atmospheric circulation in East Asia and the Northwestern Pacific. Using 17 AOGCM experiments with increased CO_2, Kimoto (2005) suggested increased Meiyu-Changma-Baiu activity associated with the strengthening of anticyclonic cells to its south and north. Based on eight AR4 simulations, Kwon et al. (2005) concludes that the increased East Asia summer precipitation is contributed by the effect of the enhanced monsoon circulation in the decaying phase of El Niño. A time-slice experiment with 20 km MRI/JMA AGCM shows that Meiyu-Changma-Baiu rainfall increases over the Yangtze River valley, the East China Sea, and western Japan, while rainfall decreases to the north of these areas mostly due to the lengthening of the Meiyu-Changma-Baiu (Oouchi et al. 2006). A northward shift of the Meiyu-Changma-Baiu front is not clear in the warming climate, and its termination tends to be delayed until August.

Kitoh and Uchiyama (2006) investigated the onset and withdrawal times of the Asia summer rainfall season in 15 AR4 simulations (Figure 11.3.4.5). They found a delay in early summer rain withdrawal over the region extending from Taiwan to Ryukyu Islands to the south of Japan, but an earlier withdrawal over the Yangtze Basin, although the latter is not significant due to large inter-model variation. Changes in onset dates are smaller. These later withdrawals may be related to higher mean surface pressure anomalies in the tropical western Pacific, associated with the projected El Nino-like mean SST change.

[INSERT FIGURE 11.3.4.5 HERE]

Yasunaga et al. (2006) used a 5 km mesh cloud resolving RCM, driven by boundary data for a 20-km mesh AGCM to investigate summer rainfall in Japan. They found no changes in rainfall in June but increased rainfall in July in a warmer climate. Precipitation systems with an area larger than 900,000 km^2 were more frequently simulated in July in the warmer climate than in the present climate, resulting in more rainfall. The occurrence of these large systems increased particularly in the vicinity of Kyushu Island, where an increase in baroclinicity was simulated.

Intense precipitation events will very likely increase in East Asia, consistent with the historical trend in this region (Fujibe et al. 2005; Zhai et al., 2005). Kanada et al. (2005) showed using a time-slice experiment with a 5 km mesh non-hydrostatic model that the confluence of disturbances from the Chinese Continent and from the East China Sea would often cause extremely heavy precipitation over Kyushu Island of Japan in July in a warmer climate. An increase in the frequency and intensity of heavy precipitation events also occurs in Korea in the long RCM simulation of Boo et al. (2006), with the largest change in the northern regions. Similarly based on RCM simulations, Xu et al. (2005) reported more extreme precipitation events in the future over China. Gao et al. (2002) found a simulated increase in the number of rainy days in Northwest China and parts of inner Mongolia, and a larger number of days with heavy rains over some regions in Southeast and Southwest China.

High-resolution simulations have also been used to study the specific kinds of disturbances that give extremely heavy precipitation. A simulation with the high-resolution MIROC3.2 AOGCM suggests that frequencies of non-precipitating and heavy ($\geq$30 mm day^{-1}) rainfall days would increase significantly at the expense of relatively weak (1–20 mm day^{-1}) rainfall days in Japan under the 21st century (Kimoto et al., 2005). More non-precipitating days would occur in winter, while heavy rainfall would become more frequent mainly in warm seasons. Similarly, Mizuta et al. (2005) find significantly more days with heavy precipitation and stronger average precipitation intensity in western Japan and Hokkaido Island. Hasegawa and Emori (2005) showed from a time-slice climate change experiment with a T106 resolution AGCM that daily precipitation associated with tropical cyclones over western North Pacific would increase due to increased water vapour in a warmer climate.

Southeast Asia

The area mean precipitation over Southeast Asia increases in most AR4 models, with a median change of about 7% in all seasons (Table 11.2), but the projected seasonal changes vary strongly within the region. The seasonal confidence intervals based on the methods of Tebaldi et al. (2004, 2005) are similar for DJF and JJA (roughly –6% to 16%. The strongest and most consistent increases broadly follow the ITCZ, lying over northern Indonesia and Indochina in JJA, and over southern Indonesia and Papua New Guinea in DJF (Figure 11.3.4.4). Away from the ITCZ, precipitation decrease is often simulated. The pattern is broadly one of wet season rainfall increase and dry season decrease.

Earlier studies of precipitation change in the area have in some cases suggested a worse intermodel agreement than found for the AR4 models. Both Giorgi et al. (2001) and Ruosteenoja et al. (2003) found inconsistency in the simulated direction of precipitation change in the region, but a relatively narrow range of possible changes. Similar results were found over an Indonesian domain by Boer and Faqih (2004). Compositing the projections from a range of earlier simulations forced by the IS92a scenario, Hulme and Sheard (1999a,b) found a pattern of rainfall increase across Northern Indonesia and the Philippines, and decrease over the southern Indonesian archipelago. More recently Boer and Faqih (2004) compared patterns of change across Indonesia from five AOGCMS and obtained highly contrasting results. Their concluded that ‘no generalisation could be made on the impact of global warming on rainfall’ in the region.

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

Rainfall variability will be affected by changes to ENSO and its effect on monsoon variability, but this is not well understood (see Chapter 10, Sections 10.3). However, as Boer and Faqih (2004) noted, those parts of Indonesia that experience mean rainfall decrease are likely to also experience increases in drought risk. It is also likely that the region will share the general tendency for daily extreme precipitation to become more intense under enhanced greenhouse conditions, particularly where the mean precipitation is projected to increase. This has been demonstrated in a range of global and regional studies (see Chapter 10, Section 10.3.6.1), 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 Chapter 10, Section 10.3 there is evidence in general of likely increases in tropical cyclone intensity, but less consistency about how occurrence will change (see also Walsh, 2004). The likely increase in intensity (precipitation and winds) has been supported for the NW Pacific (and other regions) by the recent modelling study of Knutson and Tuleya (2004). The high resolution time-slice modelling experiment of Hasegawa and Emori (2005) also demonstrated an increase in tropical cyclone precipitation in the western North Pacific, but not an increase in tropical cyclone intensity. Wu and Wang (2004) examined possible changes in tracks in the NW Pacific due to changes in steering flow in two GFDL enhanced greenhouse 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 Chapter 10, Section 10.3).

Central Asia and Tibet

Precipitation over Central Asia increases in most AR4 models in winter but decreases in the other seasons. The median change and the full range of the model results (in parentheses) under the A1B scenario in the end of the 21st century are 4% (–10% to 22%) in DJF, –13% (–59% to 21%) in JJA (the dry season) and –3% (–18% to 6%) in the annual mean (Table 11.2). This seasonal variation in the changes is broadly consistent with the earlier multi-model study of Meleshko et al. (2004), who, however, found an increase in summer precipitation in the northern part of the area.

Over the Tibetean Plateau, all AR4 models simulate increased precipitation in DJF (median 19%, range from 1% to 36%). Most but not all models also simulate increased precipitation in the other seasons (Table 11.2). Earlier studies by Xu et al. (2003a, 2003b) and Gao et al. (2003b) are consistent with these findings. Given the large biases in precipi6tation in the AR4 models, the quantitative results from the global models are suspect, but there is qualitative agreement with the regional modelling.

11.3.4.3.2 Robust conclusions and uncertainties

Conclusions about projected climate change for Asia (with types of evidence indicated according to Section 11.3.1) are:

1. All of Asia is very likely to warm during this century, the warming is likely to be well above the global mean in Central Asia, Tibetan Plateau and Northern Asia, above in Eastern Asia and South Asia, and similar to in Southeast Asia. Based on: 1, 2, and 3.
2. DJF precipitation will increase very likely in Northern Asia and the Tibetan Plateau, and likely in Eastern Asia and the southern parts of Southeastern Asia. Based on: 1, 2 and 3.
3. JJA precipitation will likely increase in Northern Asia, East Asia, South Asia and most of Southeast Asia, but it will likely decrease in Central Asia. Based on: 1, 2 and 3.
4. It is very likely that heat waves / hot spells in summer will be of longer duration, more intense, and more frequent in East Asia. Based on: 1, 2 and 3.
5. Fewer very cold days are very likely in East Asia and South Asia. Based on: 1,2 and 3.
6. There is very likely an increase in return frequency of intense precipitation events in parts of South Asia, East Asia, and Southeast Asia. Based on: 1, 2 and 3.
7. Extreme rainfall and winds associated with tropical cyclones are likely to increase in East Asia, Southeast Asia, and South Asia. Result may be affected or offset by changes in tropical cyclone numbers. Based on: 1 and 2.

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 of 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), land-sea contrast and ocean current distribution.
- Model biases in representing monsoon processes lead to substantial inter-model differences in precipitation projections, resulting in uncertainties in the quantitative estimates.
- Projections based on time slice experiments, including dynamical downscaling using regional climate models, are subject to uncertainties arising out of the lack of a realistic air-sea interaction in the simulated monsoon variability.

11.3.5 North America

11.3.5.1 Key Processes

North America spans several climatic zones, from subtropical to arctic, through mid-latitudes, the region from roughly 30° to 60° N lying in the westerlies. The North Pacific storm track terminates on the West Coast. Under the permanent influence of the Aleutian low pressure, the coastal regions from Alaska to Oregon receive the largest annual precipitation amounts, while the Rocky Mountain cordillera acts as a moisture barrier for the entire continent. On the eastern side, the thermal contrast in winter between the cold continent 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; as a result the regions northeast of the Gulf of Mexico up to Labrador receive substantial precipitation amounts. Most of North America, with the exception of southwest USA and northern Mexico, is under the influence of convergence of atmospheric moisture transported by travelling weather systems. The southwest USA and northern Mexico region is very arid, under the general influence of a subtropical ridge of high pressure. Climate-change projections indicate a slight northward displacement and intensification of the westerly flow, and an increase in the number of intense mid-latitude weather systems but a decrease in the total number. Consequent with the projected warming, the atmospheric moisture transport and the intensity of its convergence and divergence are projected to increase, resulting in a widespread increase of annual precipitation over most of the continent except the south and southwestern part.

The Pacific North America (PNA) index characterises the meandering of the jet stream: its positive phase corresponds to an intensified Aleutian low and its negative phase a more zonal flow. North America is affected by the several important patterns of oscillations (see Chapter 3): the El Niño – Southern Oscillation (ENSO), the Pacific Decadal Oscillation (PDO) and the North Atlantic / Arctic Oscillation (NAO/AO). The positive phase of ENSO produces above-normal rainfall over large regions of the USA, from southern California, the central and Gulf Coast states, and even Florida (Hagemeyer and Almeida, 2003). ENSO effects over North America however are very strongly modulated by the PDO (e.g., Gutzler et al., 2002). The positive phase of NAO/AO is characterised by stronger westerly flow and eastward displacement of the storm track, with cooling and drying over eastern Canada due to the strengthened advection of cold Arctic air masses in winter. Projections of the future changes in these oscillations are rather uncertain. In several CGCMs projections the changes over the Pacific look roughly like the El Niño phase of the ENSO cycle. The fact that PDO reverses phases at interval of a few decades poses a serious modelling challenge for projections of changes in ENSO. The future variations of the PNA are uncertain because of the limited understanding of mechanisms of mode shift, which may include internal instabilities (Dole and Black, 1990) as well as ENSO (Horel and Wallace, 1981). Several CGCMs project circulation changes reminiscent of the positive phase of the NAO, but the details of the circulation changes are model-dependent and some models do not show characteristic NAO-like circulation changes (see Chapter 10).

The North America monsoon system (NAMS) is a circulation that develops in early July over north-western Mexico and the south-western USA (Arizona, New Mexico, Utah, Colorado, Nevada, California) (e.g., Higgins et al., 1997). Similar to but of smaller scale and intensity 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 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 NAMS is strongly affected by the thermal contrast between the North American continent and adjacent tropical and North Pacific Ocean SSTs. Climate-change projections indicate a smaller warming over the Pacific Ocean than over the North American continent, increasing the thermal contrast between land and ocean in summer. In some CGCMs, this results in an amplification of the subtropical anticyclone off the West Coast of USA, inducing a decrease of annual precipitation for southwestern USA and northern Mexico.

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. The LLJ is a dynamical feature that is confined to the low levels of the atmosphere. Several factors, including the land-sea thermal contrast, 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 precipitation in the northern plains of USA (such as Nebraska, Iowa) (e.g., Augustine and Caracena, 1994). Climate-change projections indicate an increased land-sea thermal contrast in summer, with anticipated repercussions on the LLJ; CGCMs however have insufficient resolutions to adequately capture the details of the LLJ.

11.3.5.2 Simulation skill at regional scale

11.3.5.2.1 CGCMs

Current-climate simulations of AR4-generation CGCMs indicate the following characteristics over North America. While individual models vary in their ability to reproduce the observed patterns of pressure, surface air temperature and precipitation over North America, there are also several systematic aspects to their performance. The ensemble mean of CGCMs reproduces very well the annual-mean mean sea level pressure distribution (see Chapter 8, Section 8.4). 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 western part of the Canadian North-West Territories, probably due to the inability of coarse-resolution models to adequately resolve the high topography of the Rocky Mountains and to properly block incoming cyclones in the Gulf of Alaska. Conversely the pressure trough over the Labrador Sea is not deep enough; this annual-mean error pattern arises mostly from the winter biases (±4 hPa). The depth of the thermal low pressure over the southwest states in summer is somewhat excessive.

AR4 CGCMs simulate successfully the overall pattern of surface air temperature over North America (Supplementary material Table S11.2), with reduced biases compared to TAR. Ensemble-mean surface air temperature biases vary from –1.9°C to +0.6°C for all regions and seasons, and the annual-mean biases vary between –1.9°C to –0.3°C depending on the region. Over the Rocky Mountains simulated temperatures are too cold by 1.9°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 (Supplementary material Table S11.2), but almost all models overpredict precipitation for western and northern regions; the ensemble-mean excess reaches 1 to 2 mm/day over high terrain in the West of the continent. Individual model precipitation biases vary in sign over central and eastern regions; the ensemble-mean relative precipitation biases are small, ranging from –13% to +16% depending on the region and season, and the annual-mean biases vary between –1% and +8%. The ensemble-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 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 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. The ensemble mean fails to represent the region of high precipitation over southeastern USA, while the northeastern states are too wet in summer. The wet region in the Midwest is displaced westward, and summer precipitation is incorrectly simulated over Mexico and the Gulf of Mexico. An important reason for CGCMs deficiency in warm-season precipitation over North America is the prevalence of mesoscale convective systems that propagate over long distances, often 1000 km or more; these systems are much smaller than CGCM-node spacing and are fundamentally different from current subgrid-scale parameterizations of convection. There is a suggestion that there may be some relationship between horizontal resolution of atmospheric models and their ability to simulate surface air temperature throughout the year and precipitation in winter (e.g., Duffy et al. 2003). The reason appears to be that winter precipitation is mainly stratiform and depends crucially on the details of the atmospheric circulation and its interaction with topography, while summertime precipitation is mainly convective and needs to be parameterised in all climate models.

11.3.5.2.2 RCMs

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

Driven by atmospheric analyses, RCMs succeed in reproducing the overall climate. For a roughly 10° x 10° Southern Plains region, an ensemble of six RCMs in the North American Regional Climate Change Assessment Program (NARCCAP; Mearns et al., 2004, 2005) had 76% of all monthly temperature biases within ±2°C and 82% of all monthly precipitation biases within ±50%, based on preliminary results for a single year. Strong regional forcing, such as fine-scale features forced by resolved topography and land-sea contrasts, improves the skill of regional model simulations (e.g., Wang et al., 2004a). RCMs' simulations over North America exhibit rather high sensitivity to parameters such as domain size (e.g., Juang and Hong, 2001; Pan et al., 2001; Vannitsem and Chomé, 2005) and the intensity of the large-scale nudging (e.g., von Storch et al., 2000; Miguez-Macho et al., 2004).

At their typical grid-mesh of a few tens of km, 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) than corresponding warm-season values since the warm-season climate is more controlled by mesoscale and convective-scale precipitation events (Giorgi et al., 2001; Liang et al., 2004; Leung et al., 2003). On the other hand Gutowski et al. (2004) found that spatial patterns of monthly precipitation for the USA were better simulated in summer than winter in their results. 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. 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. Leung et al. (2003) examined 95th percentile of daily precipitation and found generally good agreement across many areas of the Western USA, despite important remaining methodological issues related to comparing precipitation extremes from station observations with model grid-point values. Studies targeted at the representation of convection, such as the EUROCS project, indicate that convection parameterizations usually fail 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 convection parameterizations arises from an incorrect sensitivity of the 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.

The RCMs' simulations generally inherit several biases of the driving CGCMs. 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 the North American Regional Climate Change Assessment Program (NARCCAP) (Mearns et al., 2004, 2005). The sensitivity of simulated surface air temperature to changing lateral boundary conditions from reanalyses to CGCMs appears high in winter and low in summer; for precipitation, however, the sensitivity appears to be much higher in summer than in winter (e.g., Han and Roads, 2004; Plummer et al., 2006). Improvements and increased resolution of the driving CGCMs compared to those used to drive RCMs in the TAR will lead to higher quality boundary conditions for driving RCMs; it is important to note however that, unless otherwise indicated, RCMs results reported in this section are based on simulations driven by TAR-generation CGCMs.

11.3.5.3 Projected climate changes

In this section, unless otherwise stated, CGCMs' climate-change projections refer to results of the latest AR4-generation CGCMs under the SRES scenario A1B – a middle-range scenario comprised between SRES A2 (high) and B1 (low) – for 20-year projections for the period 2079–2098, using the 20-year simulation period 1979–1998 as reference. For all regions of North America, the magnitude of the climate changes is projected to increase almost linearly with time (Figure 11.3.5.1). Unless otherwise stated, RCMs' projections refer to simulations driven by earlier TAR-generation CGCMs. Until the recent advent of the NARCCAP, climate-change projections over North America using high-resolution AGCMs and RCMs have been undertaken without a coordinated effort to produce ensembles under controlled experimental conditions.

[INSERT FIGURE 11.3.5.1 HERE]

11.3.5.3.1 Atmospheric circulation

In general the projected climate changes over North America follow the overall features of those over the Northern Hemisphere (NH) (see Chapter 10). CGCMs project northward displacement and strengthening of the mid-latitude westerly flow and its associated storm tracks, with decreasing surface pressure in the northern portion of North America and a slight increase in the south (<0.5 hPa); this tendency is most pronounced in autumn and winter. The northward displacement of the westerly flow is associated with a northward displacement of the Aleutian low-pressure centre and a northwestward displacement of the Labrador Sea trough. 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. On an annual basis, the pressure decrease in the north exceeds the spread amongst models by a factor 3 on an annual-mean basis and 1.5 in summer, so it is significant. 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 southwestern North America. The pressure increase in the south, on the other hand, is small compared to the spread amongst models, so this projection is rather uncertain.

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 NH mid-latitudes in all seasons, due to a reduction in the number of weak- and medium-strength cyclones, while strong cyclones increase in summer and decrease in winter in NH including over the East Coast of North America.

11.3.5.3.2 Surface air temperature

The ensemble mean of AR4 CGCMs projects a generalised warming for the entire continent, the annual-mean surface air temperatures warming varying from 2 to 3°C along the western, southern and eastern continental edges (there at least 16 out of the 20 models projecting a warming in excess of 2°C), up to more than 5°C in the northern region (where 16 out of the 20 CGCMs project a warming in excess of 4°C). This warming is highly significant, exceeding the spread amongst models by a factor of 3 to 4 over most of the continent. The warming in the USA is projected to exceed 2°C by nearly all models, and to exceed 4°C by more than 5 CGCMs. The largest warming is projected to occur in wintertime over northern parts of Alaska and Canada, reaching 10°C in the northernmost parts. 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. In summertime, projected warming ranges between 3 and 5°C over most of the continent, with weaker values near the coasts.

The climate-change response of RCMs is sometimes different from that of the driving CGCMs. This appears to be the result of a combination of factors, including the use of different parameterisations (convection and land-surface processes are particularly important over North America in summer) and resolution; the different resolution may also lead to differing behaviour of a same parameterisation package. For example, Chen et al. (2003) found that two RCMs projected larger temperature changes in summer than their driving CGCM. A particularly interesting contrast in the response of an RCM and its driving CGCM was found by Pan et al. (2004) and Liang et al. (2006) 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 absent in the driving model, may be due to a changing pattern of the low-level jet (LLJ) frequency and associated moisture convergence. The improved simulation of the LLJ in the RCM is made possible owing to its increased resolution.

Several RCM studies focused particularly on changes in extreme climate events. Bell et al. (2004) examined changes in temperature extremes in their simulations centred on California. They found increases in extreme temperature events, both as distribution percentiles and threshold events, prolonged hot spells and increased diurnal temperature range. Leung et al. (2003) examined changes in extremes in their RCM simulations of the western USA; in general they found increases in diurnal temperature range in six sub-regions of their domain in summer. Diffenbaugh et al. (2005) found that the frequency and magnitude of extreme temperature events changes dramatically under SRES A2, with increases in extreme hot events and decrease in extreme cold events.

Regional Statements for Surface Air Temperature

This subsection makes specific statements about anticipated temperature changes for individual regions. Unless otherwise stated, the quoted numbers refer to the 20 AR4-CGCMs ensemble results. For some fields climate-change values are also quoted from the probabilistic scheme of Tebaldi et al. (2004) as described in the uncertainty section, for the 5th and 95th percentiles of the distribution (these values are indicated in parentheses).

- *ALA:* Consequent with the general poleward amplification of the projected 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 increases between 4.5 and 11.0°C (5 and 9 °C), and smaller in summer, with warming between 1.3 and 5.6°C (1.7 and 3.7°C).

- *CGI:* Similarly to ALA, this region is expected to undergo a very large warming, with the largest warming occurring in winter, with temperature increases between 3.3 and 8.5°C (4.1 and 7.6°C), and smaller in summer, with temperature increases between 1.5 and 5.5°C (1.8 and 3.8°C).

- *WNA:* A general warming is projected for this region, with modest seasonal variations of warming. For example DJF warming spread ranges between 1.6 and 5.8°C (2.5 to 4.9°C) and JJA warming between 2.2 and 5.7°C (2.7 to 4.6°C). Consistent with a projected warming over the Pacific Ocean limited to 1 to 2°C, the projected warming is smallest near the West Coast, about 2 to 3°C, and larger inland. 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.

- *CNA:* A general warming is projected for this region, with modest seasonal variations of warming. The largest warming is expected to occur in July-August-September and smaller warming in March-April-May. For JJA projected warming spreads between 2.4 and 6.5°C (2.9 to 5.0°C); some RCMs project as much as 1.5°C less warming than their driving CGCM due to an effect referred to as a "warming hole" over the south-eastern part of the region, as discussed in Section 11.3.5.3.1. Projected warming in DJF spreads between 2.0 and 6.0°C (2.2 to 4.6°C), with smaller warming near the Gulf Coast, between 2 and 3°C, and larger values northward inland.

- *ENA:* A general warming is projected for the region with little seasonal variation of warming, ranging in DJF from 2.2 to 5.9°C (2.6 to 4.7°C) and in JJA from 2.2 to 5.4°C (2.5to 3.8°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.

11.3.5.3.3 Precipitation

The magnitude of projected precipitation changes over North America appears to scale directly with the precipitation amounts in current climate, hence it appears natural to describe precipitation projections in term of relative changes, as fraction of precipitation amounts in simulations of current climate, rather than as 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 and widely varying precipitation amounts over short distances, below the resolution of current climate models.

As a consequence of the temperature dependence of the saturation vapour pressure in the atmosphere, the projected warming is expected to be accompanied by an increase of atmospheric moisture flux and of its convergence / divergence intensity. This will result in a general increase of precipitation over most of the continent but the southwest-most part (Figure 11.3.5.2). The ensemble mean of AR4 CGCMs projects an increase of annual-mean precipitation in the North, reaching +20%, which is twice the inter-model spread, so likely significant; the projected increase reaches as much as +30% in wintertime. In the south the situation is more complex. As warming is projected to be smaller over the Pacific Ocean (+1 to +2°C) than over the continent (about +3°C over the western portion), the projected enhanced thermal contrast between land and ocean is expected to contribute to the amplification of the Pacific subtropical anticyclone off the West Coast of USA (e.g., Mote and Mantua, 2002). As a consequence of the broadening 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; this reduction is close to the inter-model spread, so it contains large uncertainty. It is noteworthy that 7 out of the 20 CGCMs do project an increase of precipitation there. In spring and summer there is a widespread projected decrease of precipitation in the South and Southwest part of the continent, with only 2 CGCMs projecting an increase of precipitation in spring there. Increased saturation vapour pressure can also yield greater evaporation, so projected increases in annual precipitation are partially offset by increases in evaporation; regions in central North America with increased precipitation may experience net surface drying as a consequence (see Supplementary material Figure S11.3.1.1).

Time-slice projections with high-resolution AGCMs can provide useful indications on the sensitivity of global models to resolution, resulting in important regional-scale differences due to better representation of topography and other factors at higher resolution. Averaged over the USA, Govindasamy (2003) found that AGCMs projected a larger (smaller) increase in precipitation than the CGCMs in winter (summer), although generally not statistically significant and averaging to negligible differences in the annual-mean precipitation responses.

Since the TAR there have been a number of RCM climate-change projections over various sub-regions of North America, using a variety of driving CGCMs, with a strong focus on changes in precipitation and water budget; these include projections over the western USA (Kim et al., 2002; Snyder et al., 2003; Bell et al., 2004; Leung et al., 2004), the north-eastern USA (Horgrefe et al., 2004), 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). In particular western USA 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. 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 several cases, RCMs responses differ significantly from one another, even when nested by the same CGCM (Kim et al., 2002; Snyder et al., 2003; Mearns et al., 2003; Liang et al., 2004; Diffenbaugh et al., 2005). For example, Chen et al. (2003) found that, in some areas downwind of the Great Lakes, some RCMs projected precipitation increases whereas the CGCM projected precipitation decreases. Han and Roads (2004) found in their results that precipitation response of an RCM differed significantly from its driving CGCM 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. On the other hand Plummer et al. (2006) found only small differences in precipitation responses using two sets of physical parameterisations in their RCM, despite the fact that one set of parameterisations corrected significant summertime precipitation excess.

[INSERT FIGURE 11.3.5.2 HERE]

Several RCM studies focused particularly on changes in extreme climate events. Bell et al. (2004) examined changes in precipitation extremes in their simulations centred on California. They found that 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. (2004) found that extremes in precipitation during the cold season increased in the northern Rockies, the Cascades, the Sierra and British Columbia by up to 10% for 2040–2060, although mean precipitation was mostly reduced, a result that was reported earlier in other climate-change projections (Giorgi et al., 2001). In a large river basin in the Pacific Northwest, increases in rainfall over snowfall and rain-on-snow events increased extreme runoff by 11%, which would contribute to more severe flooding. In their 25 km RCM simulations covering the entire USA, Diffenbaugh et al. (2005) found widespread increases in extreme precipitation events under SRES A2, which they determined to be significant.

Regional Statements for Precipitation

This subsection makes specific statements about anticipated fractional precipitation changes for individual regions. Unless otherwise stated, the quoted numbers refer to the 20 AR4-CGCMs ensemble results. For some fields climate-change values are also quoted from the probabilistic scheme of Tebaldi et al. (2004) as described in the uncertainty section, and include the 5th and 95th percentiles of the distribution (these values are indicated in parentheses).

- *ALA:* In keeping with the projected northward displacement of the westerlies and the intensification of the Aleutian low, the region is expected to undergo an increase of precipitation, particularly in winter with an increase ranging between +6 and +59% (+9 and +40%); in summer, the increase should be between +2 and +30% (+8 and +24%). The increase in precipitation could be larger on the windward slopes of the mountains as a result of increased orographic precipitation.

- *CGI:* Similarly to ALA, this region is projected to undergo an increase of precipitation, particularly in winter when the increase is expected to be between +6 and +41% (+7 and +33%). In summer, the increase is projected to be between 0 and +19% (+5 and +18%), August being the month with the smallest precipitation increase.

- *WNA:* Averaged over the region, modest annual-mean precipitation changes are projected, the majority of CGCMs indicating an increase in winter, –4 to +35% (–1 and +15%), and a decrease in summer, –19% to +11% (–14 and +7%). The uncertainty around the projected changes is large however: projections from different CGCMs produce a wide range of values (signal-to-noise ratio ≤1) and the changes do not scale well between different SRES scenarios. Also, CGCMs do not resolve well the region's important mesoscale convection dynamics. The averages for the entire region hide important north-south differences: the north is projected to experience precipitation increase and the south, a decrease. In the ensemble mean 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 +15% at the extreme north in winter, while south of the line decreases should reach –20% in summer in the ensemble mean. The line of zero change is also projected to lie further to the North under SRES scenarios with larger GHG amounts.

- *CNA:* Averaged over the region, precipitation changes are projected to be modest with little seasonal variation, ranging in DJF from –20 to +13% (–10 and +16%), and in JJA from –32% to +22% (–27 and +13%). The uncertainty around the projected changes is large however: projections from different CGCMs produce a wide range of values (signal-to-noise ratio ≤0.5) and the changes do not scale well across different SRES scenarios. The averages for the entire region hide important north-south differences: the north is projected to generally experience an increase of precipitation and the south, a decrease. The line of zero change is oriented more or less west-to-east, and projections move it meridionally with seasons, from around 35°N in winter to about 50° 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 changes could 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:* Averaged over the region, precipitation changes are projected to vary from a maximum increase in February-March-April, ranging in DJF from +2 to +26% (+3 and +21%), to modest changes in July-August-September, ranging in JJA from -18 to +14% (-7 and +8%). The uncertainty around the projected changes is large however particularly in summer: projections from different CGCMs produce a wide range of values (signal-to-noise ratio ≤ 0.2) and the changes do not scale well across different SRES scenarios. In winter the northern parts are expected to experience an increase of precipitation, reaching +25%, and the south negligible changes. Summertime precipitation is projected to decrease under SRES scenarios with larger GHG amounts, except for the Appalachian region where a small increase is projected.

11.3.5.3.4 Snowpack, snowmelt and river flow

The ensemble-mean AR4-CGCMs projections indicate a general decrease of snow depth (see Chapter 10), as a combined effect of delayed autumn snow fall and earlier spring snow melt in regions with temperatures not much below freezing, reduced accumulation as a result of increased rainfall at the expense of snow fall. The enhanced resolution of RCMs potentially allows for improved representation of certain cryospheric processes and their response under climate change, although an issue confounding comparisons between models is their widely different snow treatments (e.g., Slater et al., 2001). Because there is no consensus on how to best model snow, details in projected changes of snow depth contain large uncertainties. For some regions some models project an increase of snow depth despite climate warming, e.g. in CGCMs projections over some land around the Arctic Ocean (Chapter 10, Figure 10.3.12). In regions with well below freezing surface air temperatures, a projected increase of winter precipitation can more than make up for the shorter snow season and yield increased snow accumulation. Such conditions are met in the far north and some RCMs project snow-depth increase in the northern-most part of the North-West Territories (Figure 11.3.5.3). In principle a similar situation could arise at lower latitudes at high elevations over the Rocky Mountains; models do not agree on this aspect, and most models project a widespread decrease of snow depth over the Rocky Mountains. Several RCM studies concern projected changes in snow amount over western USA, particularly as a function of elevation (Kim et al., 2002; Snyder et al., 2003; Leung et al., 2004). Leung et al. (2003) examined changes in extremes in their RCM simulations of the western USA; they noted increases in rain-on-snow events that could contribute to more severe flooding.

[INSERT FIGURE 11.3.5.3 HERE]

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

11.3.5.4 Robust conclusions and uncertainties

Conclusions about projected climate change for North America (with types of evidence indicated according to Section 11.3.1) are:

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 3. 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 3.
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 3.
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 3.
5. Snow season length and snow depth are very likely to decrease in most of North America. Based on: 1, 2, and 3.

The uncertainties in regional climate changes over North America are strongly linked to the ability of CGCMs in reproducing the dynamical features affecting the region:

- The skill of AR4 CGCMs in simulating ENSO and NAO/AO, their projection under altered forcing, and their influence on North American climate, is largely unknown, due to the completion of the simulations shortly before this assessment;
- 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 are their influence on climate of surrounding areas;
- 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 very likely;
- 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 (Chapter 10);
- 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.

Some uncertainties listed above may be altered when the AR4-CGCMs simulations are better documented. 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. This situation stems from a combination of factors:
- All reported RCMs' projections were nested with TAR-generation CGCMs that exhibited larger biases than AR4-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);
- Several RCMs employ too few levels 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, which under samples the natural variability.
- Ensemble runs are seldom performed, occasionally few (e.g., 3) runs are made with one sometimes two RCMs, and very few RCMs have been driven systematically by several CGCMs to provide a representative sample of downscaled projections;
- RCM's projections were performed for a wide diversity of domains, periods and SRES scenarios, making it difficult or impossible to compare results.

11.3.6 Central and South America

11.3.6.1 Key processes

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

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

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

11.3.6.2 Skill of models in simulating present climate

In the Central America (CAM) and Amazonia (AMZ) regions, most AR4 models have a cold bias of 0–3°C, except in AMZ in SON (Supplementary material Table S11.2). In Southern South America (SSA) average biases are close to zero. The biases are unevenly geographically distributed (Supplementary material Figure S11.3.6.1). The AR4 models ensemble mean climate shows a warm bias around 30°S (particularly in summer) and in parts of central South America (especially in SON). Over the rest of South America (central and northern Andes, eastern Brazil, Patagonia) the biases tend to be predominantly negative. The SST biases along the western coasts of South America are likely related to weakness in oceanic upwelling.

For the CAM region, the multi-model scatter in precipitation is substantial, but half of the models lie in the range of (–15%, 25%) in the annual mean. The largest biases occur during the boreal winter and spring seasons, when precipitation is meagre (Supplementary material Table S11.2). For both AMZ and SSA, the ensemble annual mean climate exhibits drier than observed conditions, with about 60% of the models having a negative bias. Unfortunately, this choice of regions for averaging is particularly misleading for South America since it does not clearly bring out critical regional biases such as those related to rainfall underestimation in the Amazon and La Plata basins (Supplementary material Figure S11.3.6.2). Simulation of the regional climate is seriously affected by models' deficiencies at low latitudes. In particular, the AR4 ensemble tends to depict a relatively weak ITCZ, which extends southward of its observed position. The simulations have a systematic bias towards underestimated rainfall over the Amazon Basin. The simulated subtropical climate is typically also adversely affected by a dry bias over most of South Eastern South America and in the South Atlantic Convergence Zone, especially during the rainy season. In contrast, rainfall along the Andes and in NE Brazil is excessive in the ensemble mean.

AGCM simulations in tropical regions have improved in some aspects but remain a large challenge as there are systematic differences between the fine structure of the AOGCM simulated equatorial sea surface temperatures and observations that lead to differences in ocean-atmosphere interaction and thus tropical clouds and precipitation. AGCMs 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 so-called Mid Summer Drought (Magaña and Caetano 2005). Attempts to simulate tropical cyclone formation may become relevant to assess their impact on seasonal time scales (Camargo and Sobel 2004), although much is to be developed in this field.

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

Relatively few studies using RCMs for Central and South America exist, and those that do are constrained by too short simulation length. Some studies (Chou et al., 2000; Nobre et al., 2001; Druyan et al., 2002) examine the skill of experimental dynamic downscaling of seasonal predictions over Brazil. Results suggest that both more realistic GCM forcing and improvements in the RCMs are needed. Seth and Rojas (2003) performed seasonal integrations with emphasis on tropical South America applying reanalyses boundary forcing. The model was able to simulate the different rainfall anomalies and large-scale circulations but, as a result of weak low-level moisture transport from the Atlantic, rainfall over the western Amazon was undersimulated. Vernekar et al. (2003) followed a similar approach to study the low-level jets and reported that the RCM produces better regional circulation details than does the reanalysis because of its higher resolution, more realistic topography and coastal geometry, and because of its ability to realistically simulate the effects of mesoscale circulation on the time-mean flow.

Other studies (Rojas and Seth, 2003; Misra et al., 2003) analyse seasonal RCM simulations driven by AGCM simulations. Relative to the AGCMs, regional models generally improve the rainfall simulation and the tropospheric circulation over both tropical and subtropical South America. However, AGCM-driven RCMs degrade compared with the reanalyses-driven integrations and they could even exacerbate the dry bias over sectors of AMZ and perpetuate the erroneous ITCZ over the neighbouring ocean basins from the AGCMs. Menéndez et al. (2001) used a RCM driven by a stretched-grid AGCM with higher resolution over the southern mid-latitudes to simulate the winter climatology of SSA. They find that both the AGCM and the regional model have similar systematic errors but the biases are reduced in the RCM. Analogously, other RCM simulations for SSA have given too little precipitation over the subtropical plains and too much over elevated terrain (e.g., Saulo et al., 2000; Menéndez et al., 2004).

11.3.6.3 Climate projections

11.3.6.3.1 Temperature

The warming as simulated by the AR4 models for the SRES A1B scenario is projected to increase roughly linearly with time during this century, but the magnitude of the change and the inter-model range in it are greater over CAM and AMZ than over SSA (Figure 11.3.6.1). The annual mean warming under the A1B scenario from 1980–1999 to 2080–2099 varies in the CAM region from 1.8 to 5.0°C, with half of the models within 2.6–3.6°C and a median of 3.2°C. The corresponding numbers for AMZ are 1.7 to 5.0°C, 2.6–3.7°C and 3.3°C, and those for SSA 1.7 to 3.9°C, 2.3–3.1°C and 2.5°C (Table 11.2). The median warming is close to the global ensemble mean in SSA but about 30% above the global mean in the other two regions. As in the rest of the tropics, the signal to noise ratio is large for temperature, and it requires only 10 years for a 20 year mean temperature, growing at the rate of the median A1B response, to be clearly discernible above the models' internal variability.

[INSERT FIGURE 11.3.6.1 HERE]

The simulated warming is generally largest in the most continental regions, such as inner Amazonia and northern Mexico (Figure 11.3.6.2). Seasonal variation in the regional area mean warming is relatively modest, except in CAM where there is a difference of 1°C in median values between DJF (2.6°C) and MAM (3.6°C) (Table 11.2). On finer scales, the warming in central Amazonia tends to be larger in JJA than in DJF, while the reverse is true over the Altiplano where, in other words, the seasonal cycle of temperature is simulated to increase (Figure 11.3.6.2). Similar results were found by Boulanger et al. (2006) who studied the regional thermal response over South America by applying a statistical method based on neural networks and Bayesian statistics to find optimal weights for a linear combination of AR4 models.

[INSERT FIGURE 11.3.6.2 HERE]

For the variation of seasonal warming between the individual models, see Table 11.2. As an alternative approach to estimating uncertainty in the magnitude of the warming, the 5% and 95% quantiles for temperature change at the end of the 21st century, assessed from the method of Tebaldi et al. (2005) are typically within ±1°C of the median value in all three of these regions (Supplementary material Table S11.3).

11.3.6.3.2 Precipitation

The AR4 models suggest a general decrease in precipitation over most of Central America, where the median annual change by the end of the 21st century is –9% under the A1B scenario, and half of the models project area mean changes from –16% to –5% although the full range of the projections extends from –47% to 9%. Median changes in area mean precipitation in Amazonia and Southern South America are small and the variation between the models is also more modest than in Central America, but the area means hide marked regional differences (Table 11.2, Figure 11.3.6.2).

Area mean precipitation in Central America decreases in most models in all seasons. It is only in some parts of North Eastern Mexico and over the eastern Pacific, where the ITCZ forms during JJA that increases in summer precipitation are projected (Figure 11.3.6.2). However, since tropical storms can contribute a significant fraction of the rainfall in hurricane season in this region, these conclusions might be modified by the possibility of increased rainfall in storms not well captured by these global models. In particular, if the number of storms does not change, Knutson and Tuleya (2004) estimate nearly a 20% increase in average precipitation rate within 100 km of the storm centre at the time of CO_2 doubling.

For South America, the multi-model mean precipitation response (Figure 11.3.6.2) indicates marked regional variations. The annual mean precipitation is projected to decrease over northern South America near the Caribbean coasts, as well as over large parts of northern Brazil, Chile and Patagonia, while it is projected to increase in Colombia, Ecuador and Peru, around the equator and in South Eastern South America. The seasonal cycle modulates this mean change especially over the Amazon basin where monsoon precipitation increases in DJF and decreases in JJA. In other regions (e.g., Pacific coasts of northern South America, a region centred over Uruguay, Patagonia) the sign of the response is preserved throughout the seasonal cycle.

As seen in the bottom panels in Figure 11.3.6.2, most models foresee a wetter climate near the Rio de la Plata and drier conditions along much of the southern Andes, especially in DJF. However, when estimating the likelihood of this response, the qualitative consensus within this set of models must be weighed against the fact that most models are not able to reproduce the regional precipitation patterns in their control experiment with sufficient accuracy.

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 is known to influence the climate of Soth Eastern South America and the South Atlantic Convergence Zone (Robertson et al., 2003, Liebmann et al., 2004). The increase in rainfall in South Eastern South America is related with a corresponding poleward shift of the Atlantic storm track (Yin, 2005).

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

Coupled carbon-climate modeling suggests that drying of the Amazon has the potential to accelerate the rate of anthropogenic global warming by increasing atmospheric carbon dioxide (Cox et al., 2000; Jones et al., 2003, Friedlingstein et al., 2001; Dufresne et al., 2002). These models display large uncertainty in climate projections and differ in the timing and sharpness of the changes (Friedlingstein et al., 2003). Changes in carbon dioxide are related to changes in precipitation in regions such as northern Amazon (Zeng et al., 2004). A tendency to a more El Niño like state in the HADCM3 model give rise to reduced rainfall and vegetation dieback in the Amazon (Cox et al., 2004). This model projects by far the largest negative annual area-average rainfall response over AMZ among the AR4 (–21% for the A1B scenario), and is unrepresentative of the ensemble of AR4 models, stressing the necessity of being very cautious in interpreting carbon cycle results until there is more convergence among models on projections for rainfall in the Amazon with fixed vegetation.

11.3.6.4 Extremes

Little research is available on extremes of temperature and precipitation for this region. Table 11.2 provides estimates on how frequently the seasonal temperature and precipitation extremes as simulated in 1980–1999 are exceeded in using the A1B scenario. Essentially all seasons and regions are extremely warm by this criterion by the end of the century. In Central America, the projected time mean precipitation decrease is accompanied by more frequent dry extremes in all seasons. In Southern America, models anticipate extremely wet seasons in about 27% (in AMZ) and 13% (in SSA) of all DJF seasons in the period 2080–2099. The corresponding frequencies for extremely dry JJA seasons would be 16% (in AMZ) and 11% (in SSA). However, a more careful analysis is required to determine how often these wet and dry extremes are projected by the same model before concluding that both extremes are likely to increase. Austral winter (summer) seasons would not project significant changes in the frequency of extremely wet (dry) seasons.

On the daily time scale, Hegerl et al. (2004) analysed an ensemble of simulations from two AOGCMs and found that both models simulate a temperature increase in the warmest night of the year larger than the mean response over the Amazon Basin but smaller than the mean response over parts of SSA. Concerning extreme precipitation, both models foresee stronger wettest day per year over large parts of South Eastern South America and central Amazonia and weaker precipitation extremes over the coasts of NE Brazil.

11.3.6.5 Robust conclusions and uncertainties

Conclusions about projected climate change for Central and South America (with types of evidence indicated according to Section 11.3.1) are:

1. All of Central and South America is very likely to warm during this century. The annual mean warming is likely to be similar to the global mean warming in Southern South America but larger than the global mean warming in the rest of the area. Based on: 1 and 3.
2. Annual precipitation is likely to decrease in most of Central America, with the relatively dry boreal spring becoming drier. Based on: 1 and 3.
3. Annual precipitation is likely to decrease in Southern Andes. Based on: 1 and 3. 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. Precipitation is likely to increase in South Eastern South America during austral summer. Based on: 1 and 3.
5. It is uncertain how annual and seasonal mean rainfall will change over northern South America, including the Amazon forest. Based on: 1. Lack of understanding of biogeochemical feedbacks, and lack of confidence in the projections for changes in the pattern of equatorial Pacific temperatures In some regions there is qualitative consistency among the simulations (rainfall increasing in Ecuador and northern Peru, and decrease in the northern tip of the continent and in southern northeast Brazil.

The serious systematic errors in simulating current mean tropical climate and its variability (see Chapter 8, Section 8.4) and the large inter-model differences in future changes of El Niño amplitude (see Chapter 10, Section10.3) preclude a conclusive assessment of the regional changes over large areas of Central and South America. Most 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 high and sharp Andes mountains are unresolved in low resolution models, affecting the assessment over much of the continent. As with all land masses, the feedbacks from land use and land cover change are not well accommodated, and lend some degree of uncertainty. The potential for abrupt changes in biogeochemical systems in AMZ remains as a source of uncertainty (see Chapter 10, Box 10.1). Large differences in the projected climate sensitivities in the climate models incorporating these processes and lack of understanding of processes were identified (Friedlingstein et al., 2003).

Over Central America, tropical cyclones may become an additional source of uncertainty for regional scenarios of climate change, since the summer precipitation over this region may be affeceted by systematic changes in hurricane tracks and intensity.

Regional models are still being tested and developed. A major concern is the lack of knowledge/information on the changes in extremes and in frequency and intensity and 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 eastern 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 (McBride and Nicholls, 1983; Mullan, 1995 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 (Sinclair, 2002).

Tropical northern Australia lies under the influence of the monsoon and has a well-defined wet season between December and March 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. 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 (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. 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/New Zealand 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 (2006) 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 warmer than observed, but at most by 1°C in the composite. Despite this slight warm bias, the ensemble mean temperatures are biased cold over land, especially in winter in the Southeast and Southwest, where the cold bias is larger than 2°C. On large scales, the precipitation also has some systematic biases (see Supplementary material Table S11.2). Averaged across Northern Australia, the median model error is 20% more precipitation than observed, but the range of biases in individual models is large (–71% to +130%). This is discouraging with regard to confidence in many of the individual models. Consistent with this Moise et al (2006) identified simulation of Australian monsoon rainfall as a major deficiency of many of the AOGCM simulations included in CMIP2. The median annual bias in the southern Australian region is negative 6%, and the range of biases –59% to +36%. Inspection of 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., 2003a,). 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).

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. (2001a) 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 (Timbal and McAvaney, 2001; Timbal, 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. (2001b), 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 (and simulations) used in these projections is described in Whetton et al. (2005) and follows closely that described for earlier projections in Whetton et al. (1996). More detailed projections for individual states and other regions have also been prepared in recent years (Whetton et al., 2001; McInnes et al., 2003; Hennessy et al., 2004a; McInnes et al., 2004; Hennessy et al., 2004b, Cai et al., 2003a, , IOCI 2005). This work has focussed on temperature and precipitation, although additional variables such as potential evaporation and winds have been included in the more recent assessments. Moise et al (2006) analysed the results of 18 AOCM simulations included in CMIP2.

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 14 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., Timbal and McAvaney, 2001; Charles et al., 2004; Timbal, 2004;).

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 for the Environment, 2004).

11.3.7.3.1 Temperature

In both the southern and northern Australia regions, the projected warming in the 21st century under the A1B emission scenario in the AR4 AOGCMs represents a significant acceleration on warming over that observed in the 20th Century (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. Averaging over the region south of 30°S (SAU), the median 2100 warming among all of the models is 2.6 K (with an interquartile range of 2.4 to 2.9 K) whereas the median warming averaged over the region north of 30°S (NAU) is 3.0 K (range of 2.8 to 3.5 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. 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. The warming varies subregionally, 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 Chapter 10, Figure 10.3.5).

[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 2070. 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 and simulations (Mullan et al., 2001b). Decreases in the frequency of days below zero of 5–30 days per year by 2100 are projected for New Zealand, particularly for the lower North Island and the South Island. Increases in the number of days above 25°C of 10–50 days per year by 2100 are projected.

Model temperature projections are reasonably consistent with 20th century trends. All-Australian mean maximum and minimum daily temperatures have increased 0.06°C/decade and 0.11°C/decade respectively since 1910 (Della-Marta et al., 2003). 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 percentage change in precipitation between 2080–2099 in the A1B projections as compared to the 1970–1999 base. Also shown are the number of models projecting increases or decreases in precipitation. Simulated changes in precipitation averaged for the northern and Southern Australia regions are shown in Table 11.2 . The most robust feature is the reduction in rainfall along the south coast in JJA and in the annual mean. As may be seen in the regional averages (Table 11.2) decrease is also strongly evident in SON. The percentage JJA change in 2100 under the A1B scenario for Southern Australia has an interquartile range of –20% to –4%. (Table 11.2). By comparison the same range using the method probabilistic method of Tebaldi et al (2004) is –13% to –6%. There are large reductions to the south of the continent in all seasons, due to the poleward movement of the westerlies and embedded depressions (Cai et al., 2003b; Yin, 2005; Chapter 10), but this reduction extends over land during 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 (2006) has shown a southward or longitudinal shift in storms away form southwestern Australia in the AR4 simulations. To the east of Australia and over New Zealand, the primary storm track is more equatorward, and the north/south drying/moistening pattern associated with the poleward displacement is shifted equatorward as well. The result is a robust projection of increased rainfall in the South Island (especially its southern half), possibly accompanied by a decrease in the north part of the North island. The South Island increase is likely to be modulated by the strong topography, with the likelihood of it applying mainly up wind of the main range.

[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 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. Figure 11.3.7.3 shows rainfall projections over Australia using the approach of CSIRO (2001) (and described more fully in Whetton et al (2005)) but using 14 of the AR4 simulations. This shows a similar pattern to CSIRO (2001), although a slightly stronger drying tendency overall. Moise et al. (2006) also found a tendency for winter rainfall decreases across southern Australia and a slight tendency for rainfall increases in eastern Australia in 18 CMIP2 simulations under 1% per year CO_2 increase.

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;). 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; Timbal, 2004) have been applied to obtain to characteristics of rainfall change at stations (Timbal, 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 Timbal (2004) study. Downscaled rainfall projections for New Zealand (incorporating differing results of some six GCMs) showed a strong variation across the Islands (Ministry for the Environment, 2004). The picture that emerges is that the pattern of precipitation changes described above in the global simulations is still present, but with the precipitation changes focused on the upwind sides of the islands, with the increase in rainfall in the South concentrated in the West, and the decrease in the North concentrated in the East.

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 (see Chapter 9, Section 9.5) and 21st century. This observed trend and has been demonstrated to be related to changes in large scale changes in circulation and moisture (Timbal, 2004; Hope et al., 2006; IOCI, 2005), particularly a decrease in the frequency of rain-bearing systems over the region, although regional land clearing may have enhanced the trend (Pitman et al., 2004, Timbal and Arblaster, 2006). Timbal et al (2006) have demonstrated potential attribution of the change to the anthropogenic forcing. The regional circulation changes may be related to the impact on the Southern Annular Mode of the Antarctic ozone hole (see Chapter 9, Section 9.5). 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 (IOCI, 2001; Cai et al., 2005). 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 strengthened westerlies 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 Timbal, 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 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., 2004c).

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 (2006) using a high resolution subgridscale orography in a global model that simulates little change in precipitation.

11.3.7.3.4 Potential evaporation

Using the method of Walsh et al. (1999) changes to potential evaporation in the Australian region have been calculated for a range of enhanced greenhouse climate model simulation (Whetton et al., 2002; McInnes et al., 2003; Hennessy et al., 2004a; McInnes et al., 2004; Hennessy et al., 2004b; Cai et al., 2003a;). In all cases increases in potential evaporation were simulated, and in almost all cases the moisture balance deficit became stronger. Simulations with the CSIRO CGCM indicate the increases over central Australia are correlated with small increases in 10 M wind speeds; dynamically downscaled simulations with CCAM also support this relationship. 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 Chapter 10, Section 10.3) contributes to uncertainty with respect tropical cyclone behaviour (Walsh, 2004).

11.3.7.3.6 Winds

The ensemble mean projected change in wintertime sea level pressure may be seen in Chapter 10, Figure 10.3.6 based on the AR4 runs.. 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 hPa 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 11.3.7.4), ranging from a minimum in CCSM3.0 (0.6 hPa) and FGOALSg1.0 (0.7 hPa) to a maximum in GFDL-CM2.0 (5.1 hPa) and ECHAM5/MPI-OM (4.8 hPa). In the A2 ensemble mean, the increase is 3.4 hPa. 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^{-1} (Ministry for 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 hPa 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 hPa for A1B).

[INSERT FIGURE 11.3.7.4 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, with the frequency changes having little effect.

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 Robust conclusions and uncertainties

Conclusions about projected climate change for Australia and New Zealand (with types of evidence indicated according to Section 11.3.1) are:

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 3.
2. Rainfall is likely to decrease in Southern Australia in winter and spring. Based on: 1, 2 and 3.
3. Rainfall is very likely to decrease in Southwestern Australia in winter. Based on: 1, 2 and 3.
4. There will very likely be an increase in rainfall in the South Island of New Zealand. Based on: 1 and 3.
5. 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.
6. Increased mean windspeed across the southern island of New Zealand, particularly in winter, is likely. Based on: 1.
7. Increased frequency of extreme high daily temperatures, and decrease in the frequency of cold extremes is very likely. Based on: 1, 2, and 3.
8. Extremes of daily precipitation will very likely increase. Based on: 1, 2, and 3. The effect may be offset or reversed in areas of significant decrease in mean rainfall (southern Australian in winter and spring.)
9. Increase in potential evaporation is likely. Based on: 1. The effect is primarily due to increased temperature.
10. Increased risk of drought in southern areas of Australia is very likely. Based on: 1, 2, and 3.

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 eventually viewed as unreliable in the region.
- Downscaled results of the AR4 simulations are not yet 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 atmosphere, cryosphere, ocean, and land. Sea ice plays a crucial role in the Arctic climate, through the albedo-temperature feedback and feedbacks associated with the heat flux through the ice and with clouds. Substantial low-frequency variability is evident in various atmosphere and ice parameters (Polyakov et al., 2003a, b), complicating the detection and attribution of Arctic changes. Natural multi-decadal variability has been suggested as partly responsible for the large warming in the 1920s–1940s (Johannessen et al., 2004; Bengtsson et al., 2004) followed by cooling until the 1960s. In both models and observations, the interannual variability of monthly temperatures is a maximum in high latitudes (Räisänen, 2002).

Natural atmospheric patterns of variability on annual and decadal time scales play an important role in the Arctic climate. Such patterns include the NAM, NAO, and the North Pacific Index (see Chapter 3, Box 3.4 and Section 3.6). A positive NAO or NAM index is associated with warmer/wetter winters in northern Europe and Siberia and cooler/drier winters in western Greenland and north-eastern Canada. A positive NAM index is associated with warmer temperatures in Alaska and a reduction of blocking events and the associated severe weather throughout Alaska. Observations over past decades show a trend towards the positive phase of NAO/NAM (see Chapter 3, Section 3.6) that has proven difficult to simulate (see Chapter 8, Section 8.4). Despite this inconsistent record in the 20th century, models project a clear positive trend in the NAO/NAM in the 21st century (see Chapter 10, Section 10.3).

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) find relationships between the Alaskan and Beaufort-Chukchi region's climate and Northern Pacific variability. Patterns of variability in the Pacific sector, and their implications for climate change, are especially difficult to sort out due to the presence of several patterns (NAM, PDO, PNA) with potentially different underlying mechanisms.

11.3.8.1.2 Present climate: Regional simulation skill

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

Few pan-Arctic atmospheric RCMs are in use. When driven by analyzed lateral and sea-ice boundary conditions, RCMs tend to show smaller temperature and precipitation biases in the Arctic compared to

GCMs, indicating that sea ice simulation biases and biases originating from lower latitudes contribute substantially to the contamination of GCM results in the Arctic (e.g., Dethloff et al., 2001; Wei et al., 2002; Lynch et al., 2003; Semmler et al., 2005). However, even under a very constrained experimental RCM design, there can still be considerable across-model scatter in the simulations (Tjernström et al., 2004; Rinke et al., 2006). The construction of coupled atmosphere-ice-ocean RCMs for the Arctic is a recent development (Maslanik et al., 2000; Rinke et al., 2003; Debernard et al., 2003; Mikolajewicz et al., 2005).

Temperature

The simulated spatial patterns of the 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/spring) caused by overestimated sea ice in this region (Chapman and Walsh, 2006a; Chapter 8, Section 8.3). Compared with previous models, the annual temperature simulations improved in the Barents and Norwegian Seas and Sea of Okhotsk, but some deterioration is noted in the central Arctic Ocean and the high terrain areas of Alaska and northwest Canada (Chapman and Walsh, 2006a).

The mean model ensemble bias is relatively small compared to the across-model scatter of temperatures. The annual mean root-mean-squared error in the individual AR4 models ranges from 2°C to 7°C (Chapman and Walsh, 2006a). Compared with previous models, the AR4 model simulated temperatures are more consistent across the models in winter, but somewhat less so in summer, suggesting that studies of summertime climate change in this region using the AR4 ensemble of models would benefit from quality control and selection of the better performing models.

There is considerable agreement between the modelled and observed interannual variability both in magnitude and spatial pattern (Chapman and Walsh, 2006a).

Precipitation

The AOGCM simulated monthly precipitation varies substantially among the models throughout the year. But, the seasonal cycle of the multi-model ensemble mean is in qualitative agreement with the climatologies (Walsh et al., 2002; ACIA, 2005). The ensemble mean bias varies with season and remains greatest in spring and smallest in summer. The annual bias pattern (positive bias over most parts of the Arctic) can be partly attributed to coarse orography and to biased atmospheric storm tracks and sea ice cover. The AR4 models capture the observed increase of the annual precipitation through the 20th century (Chapter 3, Section 3.3).

Sea Ice and Ocean

The performance biases and the range of Arctic sea ice conditions in present-day AR4 model simulations are discussed in Chapter 8, Section 8.3. Arctic ocean-sea ice RCMs under realistic atmospheric forcing are increasingly capable of reproducing the known features of the Arctic Ocean circulation and observed sea ice drift patterns, 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 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 and/or wind forcing. Most hindcast simulations with these RCMs show a reduction in the Arctic ice volume over recent decades (Holloway and Sou, 2002).

11.3.8.1.3 Climate projections

Temperature

A northern high-latitude maximum in the warming ("polar amplification") is consistently found in all GCMs (see Chapter 10, Section 10.3). The simulated annual mean Arctic warming exceeds the global mean warming by roughly a factor of two in the AR4 models, while the wintertime warming in the central Arctic is a factor of 4 larger than the global annual mean when averaged over the models. These magnitudes are comparable to those obtained in previous studies (Holland and Bitz, 2003, ACIA, 2005). The consistency between observations and the ensemble mean 20th century simulations (Figure 11.3.8.1), combined with the fact that the near future projections (2010–2029) continue the late 20th century trends in temperature, ice extent and thickness with little modification (Serreze and Francis, 2006), increases confidence in this basic polar amplified warming pattern, despite the inter-model differences in the amount of polar amplification.

[INSERT FIGURE 11.3.8.1 HERE]

At the end of the 21st century, the projected annual warming in the Arctic is 5°C, estimated by the AR4 model mean under the A1B scenario (Figure 11.3.8.1). There is a considerable across-model range of 2.8–7.8°C between the lowest and highest projection (Table 11.2). Larger (smaller) mean warming is found for the A2 (B1) scenario with 5.9°C (3.4°C), with a proportional across-model range. Comparable magnitudes have been found in earlier estimates (ACIA, 2005). The across-model and across-scenario variability in the projected temperatures are both considerable and of comparable amplitude (Chapman and Walsh, 2006a).

Both over ocean and land, the largest (smallest) warming is projected in autumn/winter (summer) (Table 11.2, Figure 11.3.8.2). But, the seasonal amplitude of the temperature change is much larger over ocean than over land due the presence of melting sea ice in summer keeping the temperatures close to the freezing point. The surface air temperature over the Arctic Ocean region is generally warmed more than over Arctic land areas (except in summer). The range between the individual simulated changes remains large (Figure 11.3.8.2, Table 11.2). For the Arctic, by the end of the century, the warming ranges from 4.3°C to 11.4°C in winter (Tebaldi et al., 2005) 5th to 95th confidence interval of 4.4–10.5°C, Figure 11.2.1), and from 1.2°C to 5.3°C (1.7–3.4°C 5th to 95th confidence interval; Supplementary material Figure S11.2.1) in summer under the A1B scenario. In addition to the overall differences in global warming, difficulties in simulating sea ice, partly related to biases in the surface wind fields, as well as deficiencies in cloud prediction schemes, are likely responsible for much of the inter-model scatter. Internal variability plays a secondary role when examining these late 21st century responses.

[INSERT FIGURE 11.3.8.2 HERE]

The annual mean temperature response pattern at the end of the 21st century (Supplementary material Figures S11.3.8.1 and S11.3.8.2) is characterized by a robust and large warming over the central Arctic Ocean (5–7°C), dominated by the warming in winter/autumn associated with the reduced sea ice. The maximum warming is near the Barents Sea where, however, the present-day model bias is also greatest. So, the cold bias and excessive ice cover could suggest a risk of overestimating the warming there. A region of reduced warming (<2°C, even slight cooling in several models) is projected over the northern North Atlantic which is consistent among the models. This is caused by mixing into the deep ocean and reduction of northward heat transport into these regions due to weakening of the THC (see Chapter 10, Section 10.3).

While the natural variability in Arctic temperatures is large compared to other regions, the signals are still large enough to emerge quickly from the noise (Table 11.2). Looking more locally, as described by Chapman and Walsh (2006a), Alaska is perhaps the land region with the smallest signal-to-noise ratio, and is the only Arctic region in which the 20-year-mean 2010–2019 temperature is not clearly discernible from the 1980–1999 mean in the AR4 models. But even here the signal is clear by mid-century in all three scenarios.

The regional temperature responses are modified by changes in circulation patterns. In the Eastern Arctic, shifts in NAO phase can induce interdecadal temperature variations of up to 5 K (Dorn et al., 2003). The AR4 models project winter circulation changes consistent with an increasingly positive NAM (see Chapter 10, Section 10.3) which acts to enhance the warming in Eurasia and western North America. In summer, circulation patterns are projected to favor warm anomalies north of Scandinavia and extending into the eastern Arctic, with cold anomalies over much of Alaska (Cassano et al., 2006). But these circulation-induced temperature changes are not large enough to change the pattern of relatively uniform summer warming seen in the AR4 models. The deficiencies in the Arctic summertime synoptic activity in these models (as described by Cassano et al., 2006) reduce our confidence in the detailed spatial structure in these projections.

The patterns of temperature changes simulated by RCMs are quite similar to those simulated by GCMs. RCMs typically show an increased warming along the sea ice margin possibly due to a better description of the mesoscale weather systems and air-sea fluxes associated with the ice edge (ACIA, 2005). The warming over most of the central Arctic and Siberia, particular in summer, tend to be lower in RCMs (by up to 2 K) probably due to more realistic present-day snow pack simulations (ACIA, 2005).The warming is modulated by the topographical height, snow cover and connected albedo feedback as shown for the region of northern Canada and Alaska (Plummer et al., 2006; Section 11.3.5). Further systematic work with RCMs is needed to confirm and quantify these differences.

Precipitation
The AR4 models simulate a general increase in precipitation over the Arctic at the end of the 21st century (Table 11.2; Supplementary material Figure S11.3.8.3). The precipitation increase is robust among the models and qualitatively well understood, attributed to the projected warming and related increased moisture convergence (ACIA, 2005; Chapter 10, Section 10.3). The very strong correlation between the temperature and precipitation changes (~5% precipitation increase per degree warming) across the model ensemble is worth noting (Figure 11.3.8.3). Thus, both the sign and the magnitude (per degree warming) of the precipitation change are robust among the models.

[INSERT FIGURE 11.3.8.3 HERE]

The spatial pattern of the projected change (Supplementary material Figure S11.3.8.3) shows greatest percentage increase over the Arctic Ocean (30–40%) and smallest (and even slight decrease) over the northern North Atlantic (<5%). By the end of the 21st century, the projected change in the annual mean Arctic precipitation varies between the lowest and highest projection from 10% to 28%, with an AR4 model ensemble median of 18% for the A1B scenario (Table 11.2). Larger (smaller) mean precipitation increase is found for the A2 (B1) scenario with 22% (13%) but with the same inter-model range. The percentage precipitation increase is largest in winter/autumn and smallest in summer, consistent with the projected warming (Figure 11.3.8.2; Table 11.2). The Tebaldi et al. (2005) 5th to 95th quantile confidence interval of percentage precipitation change in winter is 13–36% and in summer 5–19% (Supplementary material Table S11.3).

For each scenario, the across-model scatter of the precipitation projections is substantial (Table 11.2).The differences between the projections for different scenarios are small in the first half of the 21st century, but increase after. The differences among the models increase rapidly as the spatial domain becomes smaller (ACIA, 2005). The geographical variation of precipitation changes is determined largely by changes in the synoptic circulation patterns. During winter, the 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., 2006).

Like for temperature, the large-scale patterns of precipitation changes simulated by RCMs are quite similar to those simulated by GCMs, but along the North Atlantic storm track and close to complex topography and coast lines regional details become visible in RCM simulations (ACIA, 2005).

By the end of the 21st century, under the A1B scenario, the AR4 model ensemble projected precipitation increase is significant (Table 11.2), particularly the annual and cold season (winter/autumn) precipitation. However, local precipitation changes in some regions and seasons (particularly in the Atlantic sector and generally in summer) remain difficult to discern from natural variability (ACIA, 2005).

Extremes of Temperature and Precipitation.
Very little work has been done in analyzing future changes in extreme events in the Arctic. However, the AR4 simulations indicate that the increase in mean temperature and precipitation will be combined with an increase in the frequency of very warm and wet winters and summers. Using the definition of extreme season in Section 11.3.1, every DJF and JJA seasons, in all models are "extremely" warm in the period 2080–2099 (Table 11.2). The corresponding numbers for extremely wet seasons are 89% and 83% for DJF and JJA. For the other scenarios, the frequency of extremes is very similar, except that for the wet seasons under B1 which is smaller (~63%).

Cryosphere.
Sea ice projections are discussed in Chapter 10, Section 10.3, Northern hemisphere snow projections in Chapter 10, Section 10.3, projected changes in the surface mass balance of Arctic glaciers and Greenland ice

sheet in Chapter 10 (Sections 10.3, 10.6 and 10.6), and frozen soil/permafrost changes by WGII (Chapter 15).

Arctic Ocean.
A systematic analysis of future projections for the Arctic Ocean circulation is still lacking. Coarse resolution in global models prevents the proper representation of local processes that are of global importance (such as the convection in the Greenland Sea which impacts the deep waters in the Arctic Oceans and the intermediate waters that form overflow waters). The AR4 models project a reduction in the meridional overturning circulation in the Atlantic Ocean (see Chapter 10, Section 10.3). Correspondingly, the northward oceanic heat transport decreases south of 60°N in the Atlantic. However, CMIP model assessment showed a projected increase of the oceanic heat transport at higher latitudes, associated with a stronger sub-Arctic gyre circulation in the models (Holland and Bitz, 2003). The Atlantic Ocean north of 60°N freshens during the 21st century, in pronounced contrast to the observed development in the late 20th century (Wu et al., 2003).

11.3.8.2 Antarctic
11.3.8.2.1 Key processes
Over Antarctica, there is special interest in changes in accumulation of snow that will accompany global climate change as well as the pattern of temperature change, particularly potential differences in warming over the peninsula and the interior of the icesheet. As in the Arctic, warming of the atmosphere is expected to increase precipitation, but circulation changes in both ocean and atmosphere can alter the pattern of air masses affecting the peninsula as well as the interior, modifying both precipitation and temperature patterns substantially.

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

Compared to the SAM, the Southern Oscillation (SO) shows weaker association with surface temperature over Antarctica, but the correlation with SST and sea ice extent variability in the Pacific sector of the Southern Ocean is significant (e.g., recently, Kwok and Comiso, 2002b; Renwick, 2002; Yuan, 2004; Bertler et al., 2004). Correlation between the SO index and Antarctic precipitation/accumulation has also been studied, but the persistence of the signal is being debated (Bromwich et al., 2000; Genthon and Cosme, 2003; Guo et al., 2004; Bromwich et al., 2004a; Genthon et al., 2005). Recent work suggests that this intermittence is due to nonlinear interactions between ENSO and SAM that vary on decadal time scales (Fogt and Bromwich, 2006; L'Heureux and Thompson, 2006). The SO index has a negative trend over the recent decades (corresponding to a tendency towards more El-Nino conditions in the Equatorial Pacific), associated with sea ice cover anomalies in the Pacific sector, namely negative (positive) anomalies in the Ross and Amundsen Seas (Bellingshausen and Weddell Seas) (Kwok and Comiso, 2002a). The possibility of trends in ENSO impacting sea ice extent in the future exists as well.

11.3.8.2.2 Present climate: Regional simulation skill
Major challenges face the simulation of the atmospheric conditions and precipitation patterns of the polar desert in the high interior of East Antarctica (Guo et al., 2003; Bromwich et al., 2004a; Pavolonis et al., 2004, Van de Berg et al., 2005). In addition, the evaluation of the temperature and precipitation simulations over Antarctica contains significant uncertainty. Surface temperature fields from different (re)analyses can contain large errors and are significantly different from each other (Connolley and Harangozo, 2001), with reanalyses and satellite monthly temperature data disagreeing with weather station data by as much as 3°C

(Bromwich and Fogt, 2004; Simmons et al., 2004; Comiso, 2000). Precipitation evaluation is even more challenging (Connolley and Harangozo, 2001; Zou et al., 2004). The different (re)analyses differ significantly. Very few direct precipitation gauge and detailed snow accumulation data are available, and these are uncertain to varying degrees.

Most of the AR4 global models displace the SH storm tracks and the associated surface westerlies equatorward from their observed position (see Chapter 9), with large resulting biases in subpolar latitudes. On the regional scale, RCMs generally capture the large cyclonic events affecting the coast with more fidelity (Adams, 2004) and the associated synoptic variability of temperature and precipitation (Bromwich et al., 2004a). 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., 2002ab; Van den Broeke and Van Lipzig, 2003; Bromwich et al., 2004b; Monaghan et al., 2006). Krinner et al. (1997) show the value of a stretched grid over the Antarctic as compared to standard GCM formulations. Despite these promising developments, since TAR there has been no coordinated comparison of the performance of GCMs, RCMs and other alternatives to global GCMs over Antarctica.

Temperature

The AR4 ensemble annual surface temperatures are in general slightly warmer than the observations in the Southern Ocean to the north of the sea ice region. The mean bias is predominantly less than 2°C (Carril et al., 2005) which may indicate a slight improvement compared to previous models caused by a better simulation of the position and depth of the Antarctic trough (Carril et al., 2005; Raphael and Holland, 2006). The temperature bias over sea ice is larger (e.g., it exceeds 10°C in the Ross Sea). The biases over the continent are on the order of several degrees where the model topography is erroneous (Turner et al., 2006), however the biases have to be also seen in the context of the above discussed uncertainty in the observed data sets. Changes in cloud and radiation parameterizations have been shown to change the temperature simulation significantly (Hines et al., 2004). A lateral nudging of a stretched-grid GCM (imposing the correct synoptic cyclones from 60°S and lower latitudes) brings the model in better agreement with observations (Genthon et al., 2002) but significant biases remain.

The spread in the individual global AR4 model-simulated patterns of surface temperature trends in the past 50 years is very large, but in contrast to previous models, the multi-model composite of the AR4 models qualitatively captures the observed enhanced warming trend over the Antarctic Peninsula (Chapman and Walsh, 2006b). The general improvements in resolution, sea ice models and cloud-radiation packages have evidently contributed to improved simulations. The ensemble-mean temperature trends show similarity to the observed spatial pattern of the warming, for both annual and seasonal trends (Chapman and Walsh, 2006b). For the annual trend, this includes the warming of the peninsula and near coastal Antarctica and neutral or slight cooling over the sea ice covered regions of the Southern Ocean. While the large spread among the models is not encouraging, this level of agreement suggests that some confidence in the ensemble mean 21st century projection is appropriate.

Precipitation

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

Sea Ice

The performance biases and the range of SH sea ice conditions in present-day AR4 model simulations are discussed in Chapter 8, Section 8.3.

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 but significant (Figure 11.3.8.1; Supplementary material Figure S11.3.8.4; Chapman and Walsh, 2006b). It is estimated to be 2.6°C by the median of the AR4 models under the A1B scenario, with a range from 1.4 to 5.0°C across the models (Table 11.2). Larger (smaller) warming magnitudes are found for the A2 (B1) scenario with mean values of 3.1°C (1.8°C) but with a same inter-model range of ~2.5°C. The magnitudes of the AR4 model projections are similar to previous models (Covey et al., 2003). Over the continent, the mean temperature change does not show a strong seasonal dependency; the ensemble mean A1B projections for winter (summer) are 2.9 (2.5) (Supplementary material Figure S11.3.8.4; Chapman und Walsh, 2006b). This is also illustrated by how close the Tebaldi et al. (2005) 5th to 95th confidence interval for the two seasons is: 0.1–5.7°C in summer and 1.0–4.8°C in winter (Figure 11.2.1 and Supplementary material Figure S11.2.2 and Supplementary material Table S11.3). However over the Southern Oceans, the temperature change is larger in winter/autumn than in summer/spring, which can primarily be attributed to the sea ice retreat (see Chapter 10, Section 10.3).

[INSERT FIGURE 11.3.8.4 HERE]

The annual mean AR4 model projections show a relative uniform warming over the entire continent (with a maximum in the Weddell Sea) (Figure 11.3.8.4; Carill et al., 2005; Chapman and Walsh, 2006b). They do not show a local maximum warming over the Antarctic Peninsula. This is a robust feature among the individual models (Supplementary material Figure S11.3.8.5). Thus, the pattern of observed warming and cooling trends in the last half of the 20th century is not projected to continue throughout the 21st century, despite a projected positive SAM trend (see Chapter 10, Section 10.3). It has been argued that two distinct factors have contributed to the observed SAM trend, greenhouse gas forcing and the ozone hole formation (Stone et al., 2001; Shindell and Schmidt, 2004). The relative importance of these two forcing agents for the peninsular warming requires further examination to better understand the pattern of projected warming and the implications of the healing of the ozone hole in the first half of the 21st century.

Precipitation

Almost all AR4 models simulate a robust precipitation increase in the 21st century (Supplementary material Figure S11.3.8.6; Table 11.2). By the end of the 21st century, the projected change in the annual precipitation over the Antarctic continent varies from –2% to 35%, with an AR4 model ensemble median of 14%, for the A1B scenario (Table 11.2). Similar (smaller) mean precipitation increase is found for the A2 (B1) scenario with 15% (10%) but with a same large inter-model range. The spatial pattern of the annual change is rather uniform (Supplementary material Figure S11.3.8.6). The projected relative precipitation change does not show a strong seasonal dependency, however is larger in winter than in summer (Supplementary material Figure S11.3.8.4). Generally, the Antarctic continent is projected to be wetter by 5–30%, assuming the A1B scenario. The scatter among the individual models is considerable (Table 11.2). The Tebaldi et al. (2005) 5th to 95th confidence interval for winter is –1–34% and in summer –6–22% (Supplementary material Table S11.3). It is notable that the most recent model studies of Antarctic precipitation show no significant contemporary trends (Monaghan et al., 2006; Van de Berg et al., 2005; Van den Broeke et al., 2006).

The moisture transport to the continent by synoptic activity represents a large fraction of net precipitation (Noone and Simmonds, 2002; Massom et al., 2004). During summer and winter, a systematic shift towards strong cyclonic events is projected in the AR4 models (see Chapter 10, Section 10.3) Particularly, the frequency of occurrence of deep cyclones in the Ross Sea to Bellingshausen Sea sector is increased by 20–40% (63%) in summer (winter) by the mid of the 21st century (Lynch et al., 2006). Related to this, the precipitation over the sub-Antarctic seas and Antarctic Peninsula are projected to increase.

Extremes of Temperature and Precipitation.

Very little work has been done in analyzing future changes in extreme events in the Antarctic. However, the AR4 simulations indicate that the increase in mean temperature and precipitation will be combined with an increase in the frequency of very warm and wet winters and summers. Using the definition of "extreme" seasons provided in Section 11.3.1, the AR4 models predict extremely warm seasons in about 84% of all DJF and 82% of all JJA seasons in the period 2080–2099, as averaged over all models (Table 11.2). The corresponding numbers for extremely wet seasons are 32% and 60%. For the B1 scenario, the frequency of extremes is smaller, as indicated in the table, with little difference between A1B and A2.

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

11.3.8.2.4 Robust conclusions and uncertainties
Conclusions about projected climate change for Polar regions (with types of evidence indicated according to Section 11.3.1) are:

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 3.
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 3.
3. It is likely that the Antarctic will be warmer and wetter although the magnitude is uncertain. Based on 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 3. 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.

Specific uncertainties related to polar climate change projections:

Arctic:
Arctic climate involves large natural variability, and major phenomena contributing to this are the NAO/NAM and PNA patterns; but projections of trends in these patterns contain substantial uncertainty (see Chapter 10, Section 10.3). Generally, the large-amplitude natural decadal and multi-decadal climate variability impacting the Arctic may confound the detection and attribution of climate changes for the next few decades. Further, our understanding of the Arctic climate system is still incomplete due to its complex atmosphere-land-ice-ocean interactions involving a variety of distinctive feedbacks. Processes which are not particularly well represented in either GCMs or 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). Pan-Arctic RCMs have a distinct uncertainty caused by uncertainties/biases in the driving forcings (Caya and Biner, 2004; Rinke et al., 2004; Wu et al., 2005). The uncertainties in the projected changes by the two sources (model, scenario) are of comparable order of magnitude.

Antarctic:
- large variability on interannual to interdecadal timescales
- projections of SAM and ENSO (Chapter 10, Section 10.3)
- future transient evolution of ozone forcing and its effect on SAM variability
- large model-to-model differences in present-day simulations (SH circulation, sea ice, 20th century surface temperature trend)

- some processes affecting the Antarctic climate are poorly represented or not presently included in current climate models (e.g., polar stratospheric clouds, interactive ozone and methane, high resolved stratosphere, ice-crystal precipitation)

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. Since AOGCM's do not have sufficiently fine resolutions to see the islands, the projections are given over ocean surfaces rather than over land. Very little work has been done in downscaling these projections to individual islands by dynamic or statistical means. However including the islands in the projections for neighbours with larger land masses would miss features peculiar to the islands themselves. Many small islands are sufficiently removed from large landmasses so that atmospheric circulation may be different over the smaller islands compared to their larger neighbours, e.g., in the Pacific Ocean. For the Caribbean that is close to large landmasses in Central America and northern South America, some islands partly share climate features of one, while others partly share features of the other. At the same time the Caribbean islands share many common features that are more important than those shared with the larger landmasses, such as the strong relationship of their climate to sea surface temperature. Apart from the consideration of climatic features, most small islands have concerns about global change of different emphasis than those of 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.

In the following sections the key regional processes governing the climatology of the islands which may be affected by climate change will be introduced, and the ability of the global climate models to simulate temperature and precipitation will be discussed. This will be followed by projections of these features by AR4/PCMDI models (herein referred to as AR4) using A1B SRES emission scenarios. Recent model results for tropical cyclones and sea level rise in a warming environment will also be discussed. Brief mention will be made of current climate trends which support the projections if the trends cannot be readily explained by natural variability. A discussion on ENSO changes in the tropics and ENSO- monsoon relationship, which affect climate variability in the tropics, is given in Chapter 10, Sections 10.3.

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 and the region is generally at its driest. With the onset of the spring, the NAH moves northward, the trade wind intensity decreases and the equatorial flank of the NAH becomes convergent. Concurrently easterly waves traverse the Atlantic from the coast of Africa into the Caribbean. These waves frequently mature into storms and hurricanes under warm sea surface temperatures and low vertical wind shear, generally within a 10–20ºN latitudinal band. They represent the primary rainfall source and their onset in June and demise in November roughly coincides with the mean Caribbean rainy season. In the coastal zones of Venezuela and Columbia, the wet season occurs later, from October to January (Martis et al., 2002). Inter annual variability of the rainfall is influenced mainly by ENSO events through their effect on sea surface temperatures in the Atlantic and Caribbean basins. The late rainfall season tends to be drier in El Niño years and wetter in La Niña years (Giannini et al., 1998, Martis et al., 2002, Taylor et al., 2002) and tropical cyclone activities diminish over the Caribbean during El Niño summers (Gray, 1984). However the early rainfall season in the Central and Southern Caribbean tends to be wetter in the year after an El Niño and drier in a La Niña year (Chen and Taylor, 2002).

11.3.9.1.2 Indian Ocean

The Indian Ocean region refers to the area between 35°S to 17.5°N and 50°E to 100°E. The climate of the region is influenced by the Asian monsoons (see Section 11.3.4) which is controlled by the ITCZ. In the NH (SH) summer, the ITCZ is located to the north (south) of the equator but at some distance away from it, and another trough of low pressure, called the Near Equatorial Trough (NET) is located to the south (north). Around the end of September the summer monsoon, called southwest monsoon, retreats from India as the ITCZ moves south of the Equator. 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 (the NET), from south Bay of Bengal to south Arabian Sea across the south Peninsula of India, which slowly slides southwards and remains close to the latitude of 7°N approximately during December to February. From March to May, the trough of low pressure again crawls back northwards and is about 10°N during May.

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 in the vicinity of the Seychelles. 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). Since tropical cyclones develop in the vicinity of the ITCZ or NET, cyclones are likely to originate over the Maldives and over the Seychelles from October to June due to the seasonal N-S characteristics of the ITCZ/NET.

11.3.9.1.3 Pacific

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

11.3.9.2 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 is summarized in Supplementary material Table S11.2, which gives the biases of the AR4 global models in simulating present day temperature (ºC) and precipitation (% of observed) for the period 1980–1999 on a seasonal and annual basis in terms of quartiles ranging from minimum to maximum biases. In general the biases in about half of the temperature simulations are less than 1ºC in all seasons, so that the model performances were, on the whole, satisfactoryl. There were however large spreads in precipitation simulations. For the 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

Recently, a fully coupled global climate model (Angeles et al., 2006) and a regional climate model (Martinez-Castro et al., 2006) were found to be capable of simulating the main climate features over the Caribbean region. Simulations of the annual Caribbean temperature in the 20th century (1980–1999) by AR4 models gave an average that agreed closely with climatology, differing by less than 0.1°C. The deviations of 50% individual models from the climatology ranged from –0.3°C to +0.3°C. Thus the models have good skill in simulating annual temperature. Angles et al (2006) found that the GCM underestimated the precipitation amounts. This is reflected in the AR4 simulations, the average of which underestimates the mean precipitation by approximately 30%. The deviations in the simulations of precipitation by individual models range from –64% to +20%, much greater than the deviations in temperature simulations, so that uncertainties can be expected in the simulation of Caribbean precipitation.

11.3.9.2.2 Indian Ocean

For annual temperature in the Indian Ocean in the 20th century (1980–1999), the mean value of the AR4 model outputs overestimated the climatology by 0.7°C, with 50% of deviations ranging from 0.2°C to 1.0°C. For rainfall the model consensus was only slightly below the mean precipitation by 3%, and the model deviations ranged from –22% to +20%. Thus the models have better skill in simulating precipitation 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 patterns of temperature and precipitation across the region, with the precipitation patterns more variable than for temperature in the models considered, and showing some significantly underestimating or overestimating of the intensity of rainfall in the high rainfall zones. All models simulated a broad rainfall maximum stretching across the SPCZ and ITCZ, but not all models resolved a rainfall minimum between these two regions.

Analysis of the AR4 simulations show that the average model value overestimated the mean annual temperature from 1980–1999 by 0.9°C over a southern Pacific region, with 50% of deviations varying from 0.6°C to 1.2°C. Over the North Pacific, the consensus temperature simulation for same the period was only 0.5°C above the climatology, with half of model deviations from climatology ranging from 0.2°C to 1.0°C. Average precipitation was overestimated by 10%, but individual model values varied from –7% to 31% in the southern Pacific region, whereas in the northern Pacific the mean model output for precipitation almost agreed with climatology. The individual models deviated from –13% to 13%. Thus the models were better at simulating rainfall in the northern Pacific than in the southern Pacific and the quality of the simulations, both north and south, were not much different than for the Indian Ocean.

11.3.9.3 Temperature and precipitation projections

Scenarios of temperature change (ºC) and percentage precipitation change from 1980–1999 to 2080–2099 are summarized in Table 11.2, which gives the median, the 25% and 75% (or quartile) values, and the maximum and minimum values that are simulated by the AR4 models on a seasonal and annual basis, using the SRES A1B scenario. Also shown in the table are the time interval in years (T) that is required before the signal becomes clearly discernable, and the relative frequency of extreme temperature and precipitation change. The table is described in detail in Section 11.3.1. T is a measure of the signal to noise ratio so that a small value of T implies a large signal to noise ratio. It can be seen that, in general, the signal to noise ratio is greater for temperature than for precipitation change and the probability of warming is 100% in all cases for the small islands so that the scenarios of warming are all very significant by the end of the century. Approximate results for A2 and B1 scenarios and for other future times in this century can obtained by scaling the A1B values, as described in Section 11.3.1.

[INSERT FIGURE 11.3.9.1 HERE]

The temporal evolution of temperature as simulated by AR4 models in the 20th and 21st centuries are also show in Figure 11.3.9.1, for the Caribbean (CAR), Indian Ocean (IND), North Pacific Ocean (NPA) and South Pacific Ocean (SPA). A detailed explanation of the diagrams is given in the section. The observed decadal temperature anomaly with respect to the mean temperature in the 20th century for each region (black line) can be seen to lie within the range of model anomalies when natural and anthropogenic forcings are included in the models (red shading). Thus although model biases exist, the observed anomaly lie within the range of the biases. The evolution in the 21st century is given by the green shading. In general it can be seen that the temperature increases for the small islands are less than for the continental regions. Also seen from the figures is the almost linear nature of the evolution. The ranges for the A2 and B1 scenarios at the end of the 21st century are given by the red and blue vertical lines resectively. Temperature and precipitation projections for the small island regions will be discussed below in the context of Table 11.2.

11.3.9.3.1 Caribbean

Angeles et al (2006) simulated 1ºC rise, approximately, in sea surface temperature up the 2050's using an IS92a scenario. The AR4 models simulated annual temperature increases at the end of the 21st century ranging from 1.4 to 3.2°C with an average increase of 2.1°C, somewhat below the global average. Fifty percent of the models give values differing from the mean by only ± 0.3°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. Thus there was agreement between the AOGCM and the downscaling analysis and there is a high level of confidence in the temperature simulations. The downscaling was performed with the use of the SDSM model developed by Wilby et al. (2002b) as part of an AIACC SIS06 project (http://www.aiaccproject.org). Figure 11.3.9.2(a) shows the average monthly increases projected by the the individual models with increases ranging from 1.2 to 3.4°C and no noticeable differences in monthly changes. Evidence of temperature increases in the Caribbean from 1950's to 2000 was provided by Peterson et al., (2002), who found that that the percent of time that maximum and minimum temperature observations were at or above the 90th percentile is increasing, and at the same time the corresponding percentage at or below the 10th percentile is decreasing. They also reported that the number of very warm days and nights is increasing dramatically and the number of very cool days and nights is decreasing.

Table 11.2 shows most models giving decreases in annual precipitation and a few giving increases, varying from –39% to +11%, with an average of –12%. Figure 11.3.9.3 (a) shows monthly percentage precipitation change at the end of the century. Individual models show a greater spread compared to the other regions (IND, NPA, SPA) and give greater decreases in the summer than at other times. However this is around the time of the mid-summer drought which models do not simulate well (Magana and Caetono, 2005). Note also the long time for a discernable signal. 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. Angeles et al (2006) also simulated an increase in rainfall production during the Caribbean wet season. Thus there is more consistency in the temperature results than in the precipitation results and the latter are uncertain. Peterson et al., (2002) found no statistically significant trends in mean precipitation amounts from 1950's to 2000.

[INSERT FIGURE 11.3.9.2 HERE.]

[INSERT FIGURE 11.3.9.3 HERE.]

11.3.9.3.2 Indian Ocean

Based on AR4 model consensus the annual temperature is projected to increase by about 2.2°C, somewhat below the global average with individual models ranging from 1.4 to 3.7° and at least half of the models giving values quite close to the mean. Figure 11.3.9.2 (b) gives the average monthly increases projected by the models. All models show temperature increases for all month with no significant seasonal variation. Evidence of temperature increases from 1961-90 in the Seychelles is provided by Easterling et al., (2003) who found that the percentage of time when the minimum temperature was below the 10th percentile is decreasing, and the percentage of time where the minimum temperature exceeded the 90th percentile is increasing. Similar results were obtained for the maximum temperatures.

The annual precipitation changes for individual AR4 models varied from –2% to 20% with a mean change of 4% and 50% of the models giving changes for 3% to 5%. Thus there is some level of confidence in the precipitation results although not as high as for temperature. The large number of years for a discernable signal is probably due to one outlier in the model results. Figure 11.3.9.3(b) show the monthly percentage precipitation changes given by the individual models at the end of the 21th century. All models show increases in March and April, and relatively few show decreases in the first half of the year. Easterling et al., (2003), found evidence that extreme rainfall tended to increase from 1961–1990. (See also Section 11.3.4.3, Future Projections for South Asia). Thus there is a likilhood of small precipitation changes especially in the first half of the year.

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).

The scenarios from the AR4 models using A1B emission scenarios for the period 2079 to 2098 show an average increase in temperature of 1.9ºC, somewhat below the global average over the South Pacific (Table 11.2). The individual model values vary respectively from 1.3°C to 3.1° and at least half of the models gave values very close to the mean. Figure 11.3.9.2 (d) show the monthly variation in temperature for all the models. All model show increases, slightly less in the second half of the year compared to the first. Over the North Pacific, the simulations give an average increase in temperature of 2.3°C, slightly below the global average with values ranging from 1.5°C to 3.7°C and 50% of the models within ±0.4°C of the mean. The monthly variation for each model is shown in Figure 11.3.9.2 (c), showing notable increases in the second half of the year.

A warming trend from 1961 to 2003 in Southeast Asia and the South Pacific has been found in data analyzed by Manton et al., (2001) and Griffiths et al., (2005). 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. 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 SPCZ.

For the same period, 2080 to 2099, precipitation increases over the Southern pacific when averaged over all AR4 models was 3%, with individual models giving values from –4% to +11% and 50% of the models showing increases from 3% to 6%. The time for a discernable signal is relatively low. (Table 11.2). Most of these increases were in the first half of the year as shown in Figure 11.3.9.3 (d) with all model showing increases in May and June. For precipitation in the Northern pacific an average increase of 6% was found, with individual models giving values from 0% to 19% increases and at least half of the model within ±4% of the mean. The time for a discernable signal is relatively large. Most of these increases were in the latter half of the year (Figure 11.3.9.3 (c)). Figure 11.3.9.4 illustrates the spatial distribution of annual 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 due to increased moisture transport described in Section 11.3.1.2. Change in rainfall variability in the South Pacific has not been examined using other recent simulations (but see Jones et al., 2000). However, this will be strongly driven by changes to ENSO, and this is not well understood (see Chapter 10, Section 10.3). Griffiths et al., (2003) found that there was in increasing trend from 1961–2000 in mean rainfall in and north-east of the SPCZ in the southern Pacific. As for the Indian Ocean, there is some level of confidence in the precipitation results for the Pacific, but not as high as for the temperature results.

[INSERT FIGURE 11.3.9.4 HERE]

11.3.9.4 Sea level rise

Projections of global average sea-level changes for the 21st Century due to thermal expansion, glacier and ice sheet mass changes with respect to 2000 is in the range 130–380 mm by 2100 (see Chapter 10, Section 10.6). Due to ocean density and circulation changes, the distribution will not be uniform and Figure 10.6.2 shows a distribution in local sea level change based on ensemble mean of 14 AOGCM's. A contrast of larger than average rise in the Artic and a lower than average rise in the Southern Ocean can be seen. Also obvious is a narrow band of pronounced sea-level rise stretching across the southern Atlantic and Indian Oceans at about 40ºS. This is also seen in the southern Pacific at about 30ºS. However large deviations among models make estimates of distribution across the Caribbean, Indian and Pacific Oceans uncertain.

Global sea-level rise over the 20th century is discussed in Chapter 5, Section 5.5. The increasing consensus is that the best estimate of rise lies nearer to 2 than 1 mm yr^{-1}. Observed sea-level rise in the Pacific and

Indian Oceans is discussed in Chapter 2. There have been large observed variations in sea-level rise in the Pacific Ocean mainly due to ocean circulations associated ENSO events. From estimates of observed sea level rise from 1950 to 2000 by Church et al., (2004), the rise in the Caribbean appeared to be near the global mean.

11.3.9.5 Tropical cyclones

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

11.3.9.6 Robust conclusions and uncertainties

Conclusions about projected climate change for Small Islands regions (with types of evidence indicated according to Section 11.3.1) are:

1. Sea levels will likely continue to rise on average during the century around the small islands of the Caribbean Sea, Indian Ocean and Northern and Southern Pacific Oceans. Models indicate that the rise will not be geographically uniform but large deviations among models make estimates of distribution across the Caribbean, Indian and Pacific Oceans uncertain.
2. All of Caribbean islands are very likely to warm during this century. The warming is likely to be somewhat smaller than the global, annual mean warming in all seasons. Based on: 1, 2 and 3.
3. Changes in seasonal and annual precipitation in the Caribbean islands are uncertain. Based on: 1 and 2
4. All of Indian Ocean islands are very likely to warm during this century. The warming is likely to be somewhat smaller than the global, annual mean warming in all seasons. Based on: 1 and 3.
5. Annual rainfall is likely to increase slightly in the Indian Ocean with increases likely in DJF, but changes in JJA are less certain. Based on: 1.
6. All of Northern Pacific islands are very likely to warm during this century. The warming is likely to be slightly below the global, annual mean warming in all seasons. Based on: 1 and 3.
7. Annual rainfall is likely to increase in the Northern Pacific with increases likely in JJA, but changes in DJF are less certain. Based on 1
8. All of Southern Pacific islands are very likely to warm during this century. The warming is likely to be somewhat below the global, annual mean warming in all seasons. Based on: 1 and 3.
9. Annual rainfall is likely to increase slightly in the Southern Pacific with increases likely DJF and JJA. Based on 1.

Limitations

- There is insufficient information on future simulated SST changes and insufficient model runs to determine regional distribution of cyclone changes.
- Uncertainty about future ENSO behaviour leads to uncertainty with respect changes in precipitation patterns and tropical cyclone behaviour.
- RCM's and statistical downscaling models are just being developed for many of the islands
- Large deviations among models make regional distribution of sea level rise uncertain.

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 dependence 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; the Japanese Islands in Asia), 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, extreme events such as precipitation (Kanada et al., 2005 and Yasunaga et al., 2006; Weisman et al. 1997; Walser et al. 2004), 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 GCMs because the controls of topography on precipitation are not adequately represented. In addition, it is now recognized that the superimposed effects of natural modes of climatic variability such as El Niño/Southern Oscillation (ENSO) or the North Atlantic Oscillation (NAO) can perturb mean precipitation patterns on time scales ranging from seasons to decades (Beniston and Jungo, 2001). Even though there has been progress in reproducing some of these mechanisms in coupled ocean-atmosphere models (Osborn et al., 1999), they are still not well predicted by climate models. However, considering the potential of todays downscaling techniques, several studies indicate that the higher resolution of RCMs can represent observed mesoscale patterns of the precipitation climate that are not resolved in GCMs (Kanada et al., 2005 and Yasunaga et al., 2006; Frei et al. 2005a; Schmidli et al. 2006).

Snow and ice are, for many mountain ranges, a key component of the hydrological cycle, and the seasonal character and amount of runoff is closely linked to cryospheric processes. In temperate mountain regions, the snow-pack is often close to its melting point, so that it may respond rapidly to 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 on average rise by about 150 m. Although the concept of defining the snowline is difficult to determine in the field, it is established that at lower elevations the snowline is very likely to rise by more than this simple average estimate (e.g., Martin et al., 1994; Vincent 2002; Gerbaux et al., 2006, see also Chapter 4, Section 4.2). Beniston et al. (2003) have shown that for a 4°C shift in mean winter temperatures in the European Alps, as projected by recent RCM simulations for climatic change in Europe under a strong emissions scenario (the IPCC SRES A2 emissions future), snow duration is likely to 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, 2006).

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. 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. This gives a poor baseline for assessing future changes and detecting changes in observed records. Using results from 141 sites worldwide for the last four decades Woodworth and Blackman (2004) found that at some locations extreme sea levels have increased and that the relative contribution from changes in mean sea level and atmospheric storminess depended on location.

Methods of simulating extreme sea levels

Climate driven changes in extreme sea level will come about because of the increases in mean sea level and changes in the track, frequency or intensity of atmospheric storms. (From the perspective of coastal flooding the vertical movement of land, for instance due to post glacial rebound, is also important when considering the contribution from mean sea level change.) To provide the large-scale context for these changes global climate models are required though their resolution (typically 150 to 300 km horizontally) is too coarse to represent the details of tropical cyclones or even the extreme winds associated with mid-latitude cyclones. However, some studies have used global climate model forcing directly to drive storm surge models to provide estimates of changes in extreme sea level (e.g., Flather and Williams, 2000). To obtain more realistic simulations from the large-scale drivers three approaches are used, dynamical and statistical downscaling and a stochastic method (see Section 11.2 for general details of these including their strengths and weaknesses).

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

Extreme sea level changes – sample projections from three regions

1. Australia

In a study of storm surge impacts in northern Australia, a region with only a few short sea level records, McInnes et al. (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). Similar increases for Cairns and other coastal locations were found by Hardy et al. (2004).

2. Europe

Several projections of climate driven changes in extreme water levels on the European shelf region have been carried out recently using the dynamic method. Woth (2005) explored the effect of two different GCMs and their projected climates changes due to two different emissions scenarios (SRES A2 and B2) on storm surges along the North Sea coast. She used data from one RCM downscaling the four GCMs simulations (Woth et al., 2006) using data from four RCMs driven by one GCM produced indistinguishable results) and demonstrated significant increases in the top 1% of events of 10-20cm above average sea-level change over the continental European North Sea coast. The changes from the different experiments were statistically indistinguishable though those from the models incorporating the A2 emissions were consistently larger. When including the effects of global mean sea level rise and vertical land movements Lowe and Gregory (2005) found increases in extreme sea level are projected for the entire UK coastline using a storm surge model driven by one of the RCMs analysed by Woth et al. (2006) (Box 11.4, Figure 1). A Baltic Sea ocean model driven by data from four RCM simulations indicated the possibility of large changes in storm surges, e.g., a 41cm increase above average sea-level in the100-year surge in the Gulf of Riga (Meier, 2006).

[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 statistically significant changes in the extreme surge level and a decrease in the extreme wave height with increased CO_2. An underestimation of the observed wave heights and surge levels calls for caution in the interpretation of these results. Wang et al. (2004b) used AOGCM projections to infer an increase in winter and autumn seasonal mean and extreme wave heights in the northeast and southwest North Atlantic, but a decrease in the mid-latitudes of the North Atlantic. However, the changes showed decadal fluctuations reflecting a low signal-to-noise ratio and in some regions (e.g. the North Sea) their sign was found to depend on the emissions scenario.

3. Bay of Bengal

Several dynamic simulations of storm surges have been carried out for the region but these have often involved using results from a small set of historical storms with simple adjustments (such as adding on a mean sea level or increasing wind speeds by 10%) to account for future climate change (e.g., Flather and Khandker, 1993). This technique has the disadvantage that by taking a relatively small and potentially biased set of storms it may lead to a biased distribution of water levels with an unrealistic count of extreme events. Furthermore, the climate change can not be related easily to any particular emissions or socio economic scenario. In one study using dynamical models driven by RCM simulations of current and future climates, Unnikrishnan et al. (2006) showed that despite no significant change in the frequency of cyclones there were large increases in the frequency of the highest storm surges.

Uncertainty

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

Box 11.5: Land-Use/Cover Change Experiments Related to Climate Change

Land use and land cover change (LUCC) significantly affect climate at the regional and local scales (e.g. Hansen et al, 1998; Kabat et al., 2002, Bonan, 2001; Foley et al, 2005). Recent modelling studies also show that in some instances these effects can extend beyond the areas where the land cover changes occurs, through climate teleconnection processes (e.g., Pielke et al., 2002; Marland et al., 2003). Changes in vegetation result in alteration of surface properties, such as albedo and roughness length, and alter the efficiency of ecosystems to exchange water, energy and carbon dioxide with the atmosphere (for more details see Chapter 7, Section 7.2). The effects differ widely based on the type of and location of the ecosystem altered. The effects of LUCC may be divided based on their source or origin and by the processes responsible for the transformation (Kabat et al., 2002). The effects of LUCC on climate can also be divided into biogeochemical and biophysical (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 (see Chapter 7, Section 7.2); 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 in large part due to albedo changes (Snyder et al., 2004). These LUCC changes lead to cooling by lowering average daily maximum temperatures, while daily minimum temperatures are little affected. The mean diurnal temperature range, thus also decreases. These effects are consistent with certain aspects of observed continental temperature increases: maximum temperatures remain relatively constant; i.e. the warming due to other causes (e.g., increased greenhouse gases) is roughly offset by cooling from land cover; but the minimum temperature increases are not offset, thereby leading to a net warming (Bonan, 2001; Mahmood et al, 2006). In contrast to direct cooling due to boreal deforestation, positive feedbacks associated with natural land cover change in the predominantly snow covered regions could amplify greenhouse gas warming further in the future (Chapin et al, 2005, Foley 2005).

These simulations of historical anthropogenic land-cover change effects 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; Matthews, 2003; Feddema et al. 2005a). In the future, agricultural areal expansion resulting in cooling could offset a portion of the expected warming due to greenhouse gas effects alone.

One significant land-cover conversion impact, not well 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 a considerable portion of the world population live in urban environments (and this proportion may very well increase), many people will be exposed to even warmer local climates due to increased urban heat island effects, especially through increases in mean daily minimum temperatures, a variable known to have health consequences (Meehl and Tebaldi, 2004).

Most areas well suited to large scale agriculture have already been converted to this land use/cover type. These areas include western Europe, the eastern U.S., eastern China, South America and portions of South Africa and southeastern Australia. Land-cover conversion to agriculture is likely to continue in the future, especially in parts of the western North America, tropical areas of south and central America, and arable regions in Africa and south and central Asia (IPCC, 2001; RIVM, 2002). In contrast, reforestation is expected to occur in eastern North America and the eastern portion of Europe, which is likely to continue in the future. In these areas climate impacts may include local warming associated with reforestation and decreased albedo values (Feddema, 2005b). In addition, high rates of urbanization in the same areas 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 land cover change results in a very different climate response compared to mid-latitude areas. Changes in plant cover and the reduced ability of the vegetation to transpire water to the atmosphere lead to warmer temperatures by as much as 2°C in regions of deforestation (Gedney and Valdes, 2000; Costa and Foley, 2000; De Fries et al., 2002). The decrease in transpiration acts to reduce precipitation, but this effect may be modified by changes in atmospheric moisture convergence. Most model simulations of Amazonian deforestation suggest reduced moisture convergence which would amplify the decrease in precipitation (e.g., McGuffie and Hendersson-Sellers, 1995). However, increased precipitation and moisture convergence in Amazonia during the last decades contrast with this expectation, suggesting that deforestation has not been the dominant driver of the observed changes (see Section 11.3.6.1).

Tropical regions also have the potential to affect climates beyond their immediate areal extent (Chase et al, 2000; Delire et al., 2002; Voldaire and Royer, 2004; Avissar and Worth, 2005; Feddema et al., 2005ab; Snyder, 2006). For example, changes in convection patterns can affect the Hadley circulation and thus propagate climate perturbations intothe midlatitudes. In addition, tropical deforestation in the Amazon has been found to affect sea surface temperatures in nearby Ocean locations, further amplifying teleconnections (Avissar and Worth, 2005; Feddema, 2005b; Neelin and Su, 2005; Voldoire and Royer, 2005). However, studies also indicate that there are significantly different responses to similar land use changes in other tropical regions and that responses are typically linked to dry season conditions (Voldoire and Royer, 2004a,b; Feddema et al, 2005b). Simulations of Amazonian deforestation typically show a strong climate response, both locally and in mid-latitude areas, especially North America and central Asia (Feddema et al, 2005b). However tropical land cover change in Africa and southeast Asia appear to have weaker local impacts in large part due to influences of the Asian and African monsoon circulation systems (Voldoire and Royer, 2005; Mabuchi et al., 2005a,b). While local effects are not as strong in the Indian Ocean region, land cover change in Africa, south Asia and Australia could have significant impacts on the Asian monsoon circulation and in regions where the path of Inter-Tropical Convergence Zone is affected by the monsoon (Lawrence, 2004; Feddema 2005b; Mabuchi et al., 2005ab).

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

These simulations suggest that the effects of future land-cover change will be a complex interaction of local land-cover change impacts combined with teleconnection effects due to land-cover change elsewhere, in particular the Amazon, and areas surrounding the Indian Ocean. However, projecting the potential outcomes of future climate effects due to land-cover change is difficult for two reasons. First, there is considerable uncertainty regarding how land cover will change in the future. In this context, the past may not be a good indicator of the types of land transformation that may occur in the future. 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.

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Question 11.1: Does Regional Climate Change Vary from Region to Region?

The regional response to global change is dependent on a variety of factors, including latitude, proximity to the oceans, and the dominant weather phenomena of interest. The combination of these factors is different for each region. While developing an understanding of the correct balance of regional factors remains a challenge, confidence in our regional projections has grown steadily.

Latitude is a good starting point for considering how global climate change will affect one's region. For example, while warming is expected everywhere over land, in nearly all climate models the amplitude of the warming generally increasing as one moves from the tropics to the poles. Precipitation is more complex, but also has some features that are latitutude-dependent: in subpolar latitudes precipitation is expected to increase, while decreases are expected in the many parts of the subtropics.

This general latitudinal pattern is modified, often very significantly, by one's location with respect to the oceans and mountain ranges. In many regions coastal zones are expected to warm less than the continental interiors. Precipitation responses are especially sensitive, not only to the continental geometry but the shape of nearby mountain ranges, and monsoons, extratropical cyclones, and hurricanes/typhoons are all influenced in different ways by these region-specific features. The general unifying themes as noted in Section 11.3.1.2 are developed in part around our understanding of these factors.

Some of the most difficult aspects of regional climate change relate to possible changes in the circulation of the atmosphere and oceans, and its patterns of variability. Although general statements covering a variety of regions with qualitatively similar climates can be made in some cases, nearly every region is idiosyncratic in some ways. This is true whether is be coastal zones surrounding the distinctive subtropical Mediterranean sea, or the distinctive extreme weather in the North American interior that depend on moisture transport from the Gulf of Mexico, or the interactions between vegetation patterns, oceanic temperatures, and the atmospheric circulation that help control the southern boundary of the Sahara. Many of these parts of the climate change puzzle remain to be resolved.

Tables

Table 11.1. 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. (2006)	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.
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. (2006a)	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 11.2. Temperature and precipitation projections by the AR4 global models

Averages over a number regions of the projections by a set of 21 AR4 global models for the A1B scenario. The mean temperature and precipitation responses are first averaged for each model over all available realizations of the 1980–1999 period from the 20C3M simulations and the 2080–2099 period of A1B. Computing the difference betwen these two periods, the table shows the minimum, maximum, median (50%), and 25% and 75% quartile values among the 21 models, for temperature in degrees Celsius and precipitation as a fractional change. Regions in which the middle half (25–75%) of this distribution is all of the same sign in the precipitation response are colored light brown for decreasing and light green for increasing precipitation. Signal-to-noise ratio for these values is indicated by first computing a consensus standard deviation of 20 yr means, using those models that have at least 3 realizations of the 20C3M simulations. The signal is assumed to increase linearly in time, and the time required for the median signal to reach 2.88 times the standard deviation is displayed as an estimate of when this signal is clearly discernable. The probability of extremely warm, wet, and dry seasons is also presented, as described in the text. For definitions of the regions see Table Sup. 11.2.2.1

REGION	SEASON	Temperature Response MIN 25 50 75 MAX	T YRS	% Precipitation Response MIN 25 50 75 MAX	T YRS	Extreme Seasons WARM	WET	DRY
Africa								
WAF	DJF	2.3 2.7 3.0 3.5 4.6	10	-16 -2 6 13 23	115	100	24	5
	MAM	1.7 2.8 3.5 3.6 4.8	10	-11 -7 -3 5 11	175	100	8	9
	JJA	1.5 2.7 3.2 3.7 4.7	10	-18 -2 2 7 16	>200	100	21	9
	SON	1.9 2.5 3.3 3.7 4.7	10	-12 0 1 10 15	>200	100	14	5
	ANN	1.8 2.7 3.3 3.6 4.7	10	-9 -2 2 7 13	170	100	25	9
EAF	DJF	2.0 2.6 3.1 3.4 4.2	10	-3 6 13 16 33	55	100	24	1
	MAM	1.7 2.7 3.2 3.5 4.5	10	-9 2 6 9 20	130	100	14	5
	JJA	1.6 2.7 3.4 3.6 4.7	10	-18 -2 4 7 16	150	100	9	6
	SON	1.9 2.6 3.1 3.6 4.3	10	-10 3 7 13 38	95	100	21	3
	ANN	1.8 2.5 3.2 3.4 4.3	10	-3 2 7 11 25	60	100	32	1
SAF	DJF	1.8 2.7 3.1 3.4 4.7	10	-6 -3 0 5 10	>200	100	8	6
	MAM	1.7 2.9 3.1 3.8 4.7	10	-25 -8 0 4 12	>200	98	3	8
	JJA	1.9 3.0 3.4 3.6 4.8	10	-43 -27 -22 -7 -3	70	100	1	21
	SON	2.1 3.0 3.7 4.0 5.0	10	-43 -20 -13 -8 3	90	100	2	19
	ANN	1.9 2.9 3.4 3.7 4.8	10	-12 -9 -4 2 6	115	100	2	13
SAH	DJF	2.4 2.9 3.2 3.5 5.0	15	-47 -31 -18 -12 31	>200	97	3	11
	MAM	2.3 3.3 3.6 3.8 5.2	10	-42 -37 -18 -10 13	190	100	3	21
	JJA	2.6 3.6 4.1 4.4 5.8	10	-53 -28 -5 16 74	>200	100	13	10
	SON	2.8 3.4 3.7 4.3 5.4	10	-52 -15 6 23 64	>200	100	5	6
	ANN	2.6 3.2 3.6 4.0 5.4	10	-44 -24 -6 3 57	>200	100	7	15
Europe								
NEU	DJF	2.6 3.6 4.3 5.5 8.1	40	9 13 15 22 25	50	82	44	0
	MAM	2.1 2.4 3.1 4.3 5.3	35	0 8 12 15 21	60	81	31	1
	JJA	1.4 1.9 2.7 3.3 5.0	25	-21 -5 2 7 16	>200	89	10	10
	SON	1.9 2.6 2.9 4.2 5.4	30	-5 4 8 11 13	80	86	20	2
	ANN	2.3 2.7 3.2 4.5 5.3	25	0 6 9 11 16	45	97	47	1
SEU	DJF	1.7 2.5 2.6 3.3 4.6	25	-16 -10 -6 -1 6	155	93	3	12
	MAM	2.0 3.0 3.2 3.5 4.5	20	-24 -17 -16 -8 -2	60	99	1	28
	JJA	2.7 3.7 4.1 5.0 6.5	15	-53 -35 -24 -14 -3	55	100	1	41
	SON	2.3 2.8 3.3 4.0 5.2	15	-29 -15 -12 -9 -2	90	99	1	21
	ANN	2.2 3.0 3.5 4.0 5.1	15	-27 -16 -12 -9 -4	45	100	0	45

REGION	SEASON	Temperature Response MIN	25	50	75	MAX	T YRS	% Precipitation Response MIN	25	50	75	MAX	T YRS	Extreme Seasons WARM	WET	DRY
Asia																
NAS	DJF	2.9	4.8	6.0	6.6	8.7	20	12	20	26	37	55	30	90	69	0
	MAM	2.0	2.9	3.7	5.0	6.8	25	2	16	18	24	26	30	88	65	1
	JJA	2.0	2.7	3.0	4.9	5.6	15	-1	6	9	12	16	40	100	53	1
	SON	2.8	3.6	4.8	5.8	6.9	15	7	15	17	19	29	30	99	63	0
	ANN	2.7	3.4	4.3	5.3	6.4	15	10	12	15	19	25	20	100	90	0
CAS	DJF	2.2	2.6	3.2	3.9	5.2	25	-11	0	4	9	22	>200	83	9	2
	MAM	2.3	3.1	3.9	4.5	4.9	20	-26	-14	-9	-5	3	140	91	3	17
	JJA	2.7	3.7	4.1	4.9	5.7	10	-58	-28	-13	-5	21	140	100	3	20
	SON	2.5	3.2	3.8	4.1	4.9	15	-18	-4	3	9	24	>200	99	9	10
	ANN	2.6	3.2	3.7	4.4	5.2	10	-18	-6	-3	2	6	>200	100	4	12
TIB	DJF	2.8	3.7	4.1	4.9	6.9	20	1	12	19	26	36	45	95	38	0
	MAM	2.5	2.9	3.6	4.3	6.3	15	-3	4	10	14	34	70	94	35	2
	JJA	2.7	3.2	4.0	4.7	5.4	10	-11	0	4	10	28	>200	100	27	3
	SON	2.7	3.3	3.8	4.6	6.2	15	-8	-4	8	14	21	100	100	20	4
	ANN	2.8	3.2	3.8	4.5	6.1	10	-1	2	10	13	28	45	100	46	2
EAS	DJF	2.1	3.1	3.6	4.4	5.4	20	-4	6	10	17	42	105	95	19	1
	MAM	2.1	2.6	3.3	3.8	4.6	15	0	7	11	14	20	55	97	36	2
	JJA	1.9	2.5	3.1	3.9	5.0	10	-2	5	9	11	17	45	100	34	1
	SON	2.2	2.7	3.3	4.2	5.0	15	-13	-1	9	15	29	95	100	20	2
	ANN	2.3	2.8	3.3	4.1	4.9	10	2	4	9	14	20	40	100	48	1
SAS	DJF	2.7	3.2	3.6	3.9	4.8	10	-35	-9	-5	1	15	>200	99	5	7
	MAM	2.1	3.0	3.5	3.8	5.3	10	-30	-2	9	18	26	150	100	13	5
	JJA	1.2	2.2	2.7	3.2	4.4	15	-3	4	11	16	23	45	96	31	0
	SON	2.0	2.5	3.1	3.5	4.4	10	-12	8	15	20	26	50	100	27	3
	ANN	2.0	2.7	3.3	3.6	4.7	10	-15	5	11	15	20	40	100	38	3
SEA	DJF	1.6	2.1	2.5	2.9	3.6	10	-4	3	6	10	12	80	99	24	3
	MAM	1.5	2.2	2.7	3.1	3.9	10	-4	2	7	9	17	75	100	26	2
	JJA	1.5	2.2	2.4	2.9	3.8	10	-3	3	7	9	17	70	100	25	1
	SON	1.6	2.2	2.4	2.9	3.6	10	-2	2	6	10	21	85	99	26	2
	ANN	1.5	2.3	2.5	3.0	3.7	10	-2	3	7	8	15	40	100	44	1
North America																
ALA	DJF	4.4	5.6	6.3	7.5	11.0	30	6	20	28	34	56	40	80	40	0
	MAM	2.3	3.2	3.5	4.7	7.7	35	3	13	17	23	38	40	64	44	0
	JJA	1.3	1.8	2.4	3.8	5.7	25	1	8	14	20	30	45	87	45	1
	SON	2.3	3.6	4.5	5.3	7.4	25	6	14	19	31	36	40	86	53	0
	ANN	3.0	3.7	4.5	5.2	7.4	20	6	13	21	25	32	25	97	82	0
CGI	DJF	3.3	5.2	5.9	7.2	8.5	20	6	15	26	32	42	30	93	60	0
	MAM	2.4	3.2	3.8	4.6	7.2	20	4	13	17	20	34	35	96	52	1
	JJA	1.5	2.1	2.8	3.7	5.6	15	0	8	11	12	19	35	100	49	1
	SON	2.7	3.4	4.0	5.7	7.3	20	7	14	16	22	37	35	100	60	0
	ANN	2.8	3.5	4.3	5.0	7.1	15	8	12	15	20	31	25	100	89	0
WNA	DJF	1.6	3.1	3.6	4.4	5.8	25	-4	2	7	11	36	105	79	17	2
	MAM	1.5	2.4	3.1	3.4	6.0	20	-7	2	5	8	14	130	86	13	4
	JJA	2.3	3.2	3.8	4.8	5.7	10	-18	-10	-1	2	10	>200	100	2	13

REGION	SEASON	Temperature Response MIN 25 50 75 MAX	T YRS	% Precipitation Response MIN 25 50 75 MAX	T YRS	Extreme Seasons WARM	WET	DRY
	SON	2.0 2.8 3.1 4.5 5.3	20	-3 3 6 12 18	105	94	18	2
	ANN	2.1 2.9 3.4 4.1 5.7	15	-3 0 5 9 14	70	100	20	2
CNA	DJF	2.0 2.9 3.5 4.2 6.1	30	-18 0 5 8 14	>200	74	6	5
	MAM	1.9 2.8 3.3 3.9 5.7	25	-17 2 7 12 17	125	83	18	4
	JJA	2.4 3.1 4.1 5.1 6.4	20	-31 -15 -3 4 20	>200	92	6	16
	SON	2.4 3.0 3.5 4.6 5.8	20	-17 -4 4 11 24	>200	92	11	8
	ANN	2.3 3.0 3.5 4.4 5.8	15	-16 -3 3 7 15	>200	98	12	6
ENA	DJF	2.1 3.1 3.8 4.6 6.0	25	2 9 11 19 28	85	82	24	3
	MAM	2.3 2.7 3.5 3.9 5.9	20	-4 7 12 16 23	60	86	22	2
	JJA	2.1 2.6 3.3 4.3 5.4	15	-17 -3 1 6 13	>200	99	9	10
	SON	2.2 2.8 3.5 4.4 5.7	20	-7 4 7 11 17	150	95	20	5
	ANN	2.3 2.8 3.6 4.3 5.6	15	-3 5 7 10 15	55	100	32	1
Central and South America								
CAM	DJF	1.4 2.2 2.6 3.5 4.6	15	-57 -18 -14 -9 0	105	96	2	25
	MAM	1.9 2.7 3.6 3.8 5.2	10	-46 -25 -16 -10 15	75	100	1	20
	JJA	1.8 2.7 3.4 3.6 5.5	10	-44 -25 -9 -4 12	90	100	5	24
	SON	2.0 2.7 3.2 3.7 4.6	10	-45 -10 -4 7 24	>200	100	7	16
	ANN	1.8 2.6 3.2 3.6 5.0	10	-48 -16 -9 -5 9	65	100	3	35
AMZ	DJF	1.7 2.4 3.0 3.7 4.6	10	-13 0 4 11 17	130	93	27	5
	MAM	1.7 2.5 3.0 3.7 4.6	10	-13 -1 1 4 14	>200	100	16	5
	JJA	2.0 2.7 3.5 3.9 5.6	10	-38 -10 -3 2 13	170	100	7	16
	SON	1.8 2.8 3.5 4.1 5.4	10	-35 -12 -2 8 21	>200	100	15	14
	ANN	1.8 2.6 3.3 3.7 5.1	10	-21 -3 0 6 14	>200	100	21	9
SSA	DJF	1.5 2.5 2.7 3.3 4.3	10	-16 -2 1 7 10	>200	100	13	4
	MAM	1.8 2.3 2.6 3.0 4.2	15	-11 -2 1 5 7	>200	98	9	7
	JJA	1.7 2.1 2.4 2.8 3.6	15	-20 -7 0 3 17	>200	95	8	11
	SON	1.8 2.2 2.7 3.2 4.0	15	-20 -12 1 6 11	>200	99	7	11
	ANN	1.7 2.3 2.5 3.1 3.9	10	-12 -1 3 5 7	125	100	10	9
Australia and New Zealand								
NAU	DJF	2.2 2.6 3.1 3.7 4.6	20	-20 -8 1 9 27	>200	87	7	4
	MAM	2.1 2.7 3.1 3.3 4.3	20	-24 -12 1 15 40	>200	91	12	2
	JJA	2.0 2.7 3.0 3.3 4.3	25	-54 -20 -14 3 26	>200	95	4	10
	SON	2.5 3.0 3.2 3.8 5.0	20	-58 -32 -12 2 20	>200	98	5	10
	ANN	2.3 2.8 3.0 3.5 4.5	15	-25 -8 -4 8 23	>200	99	9	5
SAU	DJF	2.0 2.4 2.7 3.2 4.2	20	-23 -12 -2 12 30	>200	95	9	6
	MAM	2.0 2.2 2.5 2.8 3.9	20	-31 -9 -5 13 32	>200	89	7	7
	JJA	1.7 2.0 2.3 2.5 3.5	15	-37 -20 -11 -4 9	120	96	4	18
	SON	2.0 2.6 2.8 3.0 4.1	20	-42 -27 -14 -5 4	140	94	5	14
	ANN	1.9 2.4 2.6 2.9 3.9	15	-27 -13 -4 3 12	>200	100	5	7
Polar Region								
ARC	DJF	4.3 6.0 6.9 8.4 11.4	15	11 19 26 29 39	25	100	89	0
	MAM	2.4 3.7 4.4 4.9 7.3	15	9 14 16 21 32	25	100	74	0
	JJA	1.2 1.7 2.1 3.0 5.3	15	4 10 14 17 20	25	100	83	0
	SON	2.9 4.8 6.0 7.2 8.9	15	9 17 21 26 35	20	100	95	0

REGION	SEASON	Temperature Response MIN 25 50 75 MAX	T YRS	% Precipitation Response MIN 25 50 75 MAX	T YRS	Extreme Seasons WARM	WET	DRY
	ANN	2.8 4.0 4.9 5.6 7.8	15	10 15 18 22 28	20	100	100	0
ANT	DJF	0.8 2.2 2.6 2.9 4.6	20	-11 5 9 14 31	50	84	32	2
	MAM	1.3 2.2 2.6 3.3 5.3	20	1 8 12 19 40	40	89	52	0
	JJA	1.4 2.3 2.8 3.3 5.2	25	5 14 19 24 41	30	82	60	0
	SON	1.3 2.1 2.3 3.2 4.8	25	-2 9 12 18 36	45	77	42	0
	ANN	1.4 2.3 2.6 3.0 5.0	15	-2 9 14 17 35	25	98	81	1
Small Islands								
CAR	DJF	1.4 1.8 2.1 2.4 3.2	10	-21 -11 -6 0 10	185	100	3	10
	MAM	1.3 1.8 2.2 2.4 3.2	10	-28 -20 -13 -6 6	115	100	3	18
	JJA	1.3 1.8 2.0 2.4 3.2	10	-57 -35 -20 -6 8	60	100	2	40
	SON	1.6 1.9 2.0 2.5 3.4	10	-38 -18 -6 1 19	180	100	5	21
	ANN	1.4 1.8 2.0 2.4 3.2	10	-39 -19 -12 -3 11	60	100	2	37
IND	DJF	1.4 2.0 2.1 2.4 3.8	10	-4 2 4 9 20	135	100	19	2
	MAM	1.5 2.0 2.2 2.5 3.8	10	0 3 5 6 20	80	100	24	1
	JJA	1.4 1.9 2.1 2.4 3.7	10	-3 -1 3 5 20	165	100	19	4
	SON	1.4 1.9 2.0 2.3 3.6	10	-5 2 4 7 21	110	100	19	2
	ANN	1.4 1.9 2.1 2.4 3.7	10	-2 3 4 5 20	65	100	29	2
NPA	DJF	1.5 1.9 2.4 2.5 3.6	10	-5 1 3 6 17	130	100	17	1
	MAM	1.4 1.9 2.3 2.5 3.5	10	-17 -1 1 3 17	>200	100	14	8
	JJA	1.4 1.9 2.3 2.7 3.9	10	1 5 8 14 25	55	100	42	0
	SON	1.6 1.9 2.4 2.9 3.9	10	1 5 6 13 22	50	100	32	0
	ANN	1.5 1.9 2.3 2.6 3.7	10	0 3 5 10 19	60	100	36	1
SPA	DJF	1.4 1.7 1.8 2.1 3.2	10	-6 1 4 7 15	80	100	20	4
	MAM	1.4 1.8 1.9 2.1 3.2	10	-3 3 6 8 17	35	100	36	1
	JJA	1.4 1.7 1.8 2.0 3.1	10	-2 1 3 5 12	70	100	29	3
	SON	1.4 1.6 1.8 2.0 3.0	10	-8 -2 2 4 5	135	100	14	15
	ANN	1.4 1.7 1.8 2.0 3.1	10	-4 -3 3 6 11	40	100	38	2

Notes:
ARC = land + ocan
ANT = land only

Table 11.3. Projected changes in climate extremes under SRES A1B for the period 2079–2098 compared to the period 1979–1998. VL: Very Likely; L: Likely; M: Medium confidence

Temperature-Related Phenomena	
Change in phenomenon	**Projected changes**
Higher maxTmax, more hot / warm summer days	VL (consistent across model projections) maxTmax increases at same rate as the mean or median[i] over northern Europe[ii], Australia and New Zealand[iii] L (fairly consistent across models, but sensitivity to land-surface treatment) maxTmax increases more than the median over southern and central Europe, and South-West USA[iv] L (consistent with projected large increase in mean temperature) Dramatic increase in probability of extreme warm seasons over most part of the world [v]
Longer duration, more intense, more frequent heat waves / hot spells in summer	VL (consistent across model projections) Over almost all continents[vi], but particularly central Europe[vii], California and regions of western USA[viii], East Asia[ix] and Korea[x]
Higher maxTmin; more warm and fewer cold nights	VL (consistent with higher mean temperatures) Over most continents[xi]
Higher minTmin	VL (consistent across model projections) minTmin increases more than the mean in many mid-and hi-latitude locations[xii], particularly in winter over eastern, central and northern Europe[xiii]
Higher minTmax, fewer cold days	L (consistent with warmer mean temperatures)
Fewer frost days	VL (consistent across model projections) Decrease in number of days with below freezing temperatures everywhere[xiv]
Fewer cold outbreaks; fewer, shorter, less intense cold spells / cold extremes in winter	VL (consistent across model projections) Northern Europe, East Asia[xv] L (consistent with warmer mean temperatures) For other regions L (some inconsistencies across model projections) Southern Europe, Australia, New Zealand[xvi]
Reduced diurnal temperature range (DTR)	L (consistent across model projections) Over most continental regions, night temperatures increase faster than the day temperatures[xvii]
Increase of heat index	VL (consistent with increased temperature and moisture) Over most land areas, heat index rises more than temperature
Temperature variability on interannual and daily time scales	L (general consensus across model projections) Reduced in winter over most of Europe[xviii] Increase in central Europe in summer[xix]

Moisture-Related Phenomena	
Change in phenomenon	**Projected changes**
Intense precipitation events	VL (consistent across model projections; empirical evidence, generally higher precipitation extremes in warmer climates) Much larger increase in the frequency than in the magnitude of precipitation extremes over most land areas in middle latitudes[xx], particularly over northern Europe[xxi], Australia and New Zealand[xxii] Large increase during the Indian summer monsoon season over Arabian Sea, tropical Indian Ocean, northern Pakistan, northwest and northeast India, Bangladesh and Myanmar[xxiii] Increase in summer over southeast and southwest China, Korea, and Japan[xxiv] L (some inconsistencies across model projections) Modest increase over central Europe in winter[xxv] Increase associated with tropical cyclones over Southeast Asia, Japan[xxvi] Uncertain Changes in summer over Mediterranean and central Europe [xxvii] L decrease (consistent across model projections) Iberian Peninsula[xxviii], northwest India[xxix], South Asia[xxx]

Wet days	L (consistent across model projections) Increase in number of days at high latitudes in winter, and over Western and central parts of South Asia, Himalayas foothills, northeast India, northwest China, parts of inner Mongolia[xxxi] Increase over the ITCZ[xxxii] Decrease in South Asia[xxxiii] and the Mediterranean area[xxxiv]
Dry spells (periods of consecutive dry days)	VL (consistent across model projections) Increase in length and frequency over the Mediterranean area[xxxv], southern areas of Australia, New Zealand[xxxvi] L (consistent across model projections) Increase in most subtropical areas Little change over northern Europe[xxxvii]
Increased continental drying and associated risk of drought	L (consistent across model projections; consistent change in P-E, but sensitivity to formulation of land-surface processes) In summer over many mid-latitude continental interiors, e.g. central[xxxviii] and southern Europe, Mediterranean area[xxxix], in boreal spring and dry periods of the annual cycle over Central America[xl] Uncertain response Over the Sahel region

Tropical Cyclones (typhoons and hurricanes)	
Change in phenomenon	**Projected changes**
Increase in peak wind intensities	L (high-resolution AGCM and embedded hurricane-model projections) Over most tropical cyclone areas[xli]
Increase in mean and peak precipitation intensities	L (high-resolution AGCM projections and embedded hurricane-model projections) East[xlii] and Southeast Asia[xliii], Australia and southeast Pacific [xliv]
Changes in frequency of occurrence	M (some high-resolution AGCM projections) Decrease in number of weak storms, increase in number of strong storms[xlv] M (several climate model projections) Globally averaged decrease in number, but specific regional changes dependent on SST change[xlvi] Possible increase over the North Atlantic in addition to changes due to natural variability[xlvii]
Longer mean duration	Insufficient studies for assessment

Extratropical Cyclones	
Change in phenomenon	**Projected changes**
Changes in frequency and position	L (consistent in CGCM projections) Decrease in the total number of extratropical cyclones[xlviii] Slight poleward shift of storm track and associated precipitation, particularly in winter[xlix]
Change in storm intensity and winds	L (consistent in most CGCM projections, but not explicitly analysed for all models) Increased number of intense cyclones[l] and associated strong winds, particularly in winter over the North Atlantic[li], northern and central Europe[lii], and Southern Island of New Zealand[liii] Reduced windiness in Mediterranean Europe[liv]
Increased wave height	L (based on projected changes in extratropical storms) Increased occurrence of high waves in most midlatitude areas analyzed, particularly the North Sea[lv]

Assessment basis and references:

[i] Kharin and Zwiers (2005a,b)
[ii] §11.3.3.3.2, Fig. 11.3.3.3, PRUDENCE, Kjellström et al. (2005)
[iii] §11.3.7.3.1, CSIRO (2001)
[iv] §11.3.1
[v] Tebaldi et al. (2006)
[vi] §11.3.3.3.2, Tebaldi et al. (2005), Meehl and Tebaldi (2004)
[vii] Gregory and Mitchell (1995), Zwiers and Kharin (1998), Hegerl et al. (2004), Meehl and Tebaldi (2004)
[viii] §11.3.5.3.2, Bell et al. (2004), Leung et al. (2003a)
[ix] §11.3.4.3.3, Gao et al. (2002)
[x] §11.3.4.3.3, Kwon et al. (2005), Boo et al. (2006)
[xi] §11.3.3.3.2, §11.3.4.3.3
[xii] Kharin and Zwiers (2005a,b)
[xiii] §11.3.3.3.2, Fig. 11.3.3.3, PRUDENCE
[xiv] Tebaldi et al. (2005), Meehl et al. (2004), §11.3.3.3.2, PRUDENCE, Jylhä et al. (2005) §11.3.7.3.1, CSIRO (2001), §11.3.7.3.1, Mullan et al. (2001b)
[xv] §11.3.3.3.2, PRUDENCE, Kjellström et al. (2005), §11.3.4.3.3, Gao et al. (2002), Krishna Kumar et al. (2003)
[xvi] §11.3.3.3.2, Vavrus et al. (2005), §11.3.1
[xvii] §11.3.5.3.2, Bell et al. (2004), Leung et al. (2003a), §11.3.4.3.3, Krishna Kumar et al. (2003), Mizuta et al. (2005)
[xviii] §11.3.3.3.2, Räisänen (2001), Räisänen et al. (2003), Giorgi and Bi (2005), Zwiers and Kharin (1998), Hegerl et al. (2004), Kjellström et al. (2005)
[xix] §11.3.3.2, PRUDENCE, Schär et al. (2004), Vidale et al. (2006)
[xx] Groisman et al. (2005), Kharin and Zwiers (2005a,b), Hegerl et al. (2004), Semenov and Bengtsson (2002), Meehl et al. (2005)
[xxi] §11.3.3.3.4, Räisänen (2002), Giorgi and Bi (2005), Räisänen (2005)
[xxii] §11.3.1, §11.3.7.3.2, Hennessy et al. (1997), Whetton et al. (2002), Watterson and Dix (2003), Suppiah et al. (2004), McInnes et al. (2003), Hennessy et al. (2004b), Abbs (2004), Semenov and Bengtsson (2002)
[xxiii] §11.3.4.3.3, May (2004a)
[xxiv] §11.3.4.3.3, Gao et al. (2002), Boo et al. (2006), Kimoto et al. (2005), Mizuta et al. (2005)
[xxv] §11.3.3.3.4, PRUDENCE, Frei et al. (2005), Christensen and Christensen (2003, 2004)
[xxvi] §11.3.1, §11.3.4.3.3, Kimoto et al. (2005), Mizuta et al. (2005), Hasegawa and Emori (2005), Kanada et al. (2005)
[xxvii] §11.3.3.3.4, PRUDENCE, Frei et al. (2005), Christensen and Christensen (2003, 2004), Tebaldi et al. (2005)
[xxviii] §11.3.3.3.4, PRUDENCE, Frei et al. (2005)
[xxix] §11.3.4.3.3
[xxx] §11.3.4.3.3, Krishna Kumar et al. (2003)
[xxxi] §11.3.4.3.3, Gao et al. (2002), Hasegawa and Emori (2005)
[xxxii] Semenov and Bengtsson (2002)
[xxxiii] §11.3.4.3.3
[xxxiv] §11.3.3.3.4, Semenov and Bengtsson (2002), Voss et al. (2002), Räisänen et al. (2003,2004); Frei et al. (2006)
[xxxv] §11.3.1
[xxxvi] §11.3.1, §11.3.7.3.2, Whetton and Suppiah (2003), McInnes et al. (2003), Walsh et al. (2002), Hennessy et al. (2004), Mullan et al. (2005)
[xxxvii] §11.3.3.3.4, Beniston et al. (2005), Tebaldi et al. (2005), Voss et al. (2002)
[xxxviii] Rowell and Jones (2006)
[xxxix] §11.3.1, §11.3.3.3.4, Voss et al. (2002), Räisänen (2005)
[xl] §11.3.1
[xli] Knutson and Tuleya (2004)
[xlii] §11.3.4.3.3, Gao et al. (2002), Hasegawa and Emori (2005)
[xliii] §11.3.1
[xliv] §11.3.7.3.5, §11.3.9.5, Walsh and Katzfey (2000), Walsh and Ryan (2000), Ngyuen and Walsh (2001), Walsh et al. (2004), Walsh (2004)
[xlv] Oouchi et al. (2006)
[xlvi] Hasegawa and Emori (2005)
[xlvii] Sugi et al. (2002), Oouchi et al. (2006)
[xlviii] §11.3.1, §11.3.3.3.5, Yin (2006), Lambert and Fyfe (2006), §11.3.3.3.5, Lionello et al. (2002), Leckebusch et al. (2005), Vérant (2004), Somot (2005)
[xlix] Yin (2006), Lambert and Fyfe (2006)
[l] §11.3.1, §11.3.3.3.5, Yin (2006), Lambert and Fyfe (2006)
[li] §11.3.3.3.6, Ulbrich et al. (2005a)

[lii] §11.3.3.3.5, Zwiers and Kharin (1998), Knippertz et al. (2000), Leckebusch and Ulbrich (2004), Pryor et al. (2005a), Lionello et al. (2002), Leckebusch et al. (2005), Vérant (2004), Somot (2005)
[liii] §11.3.1
[liv] §11.3.3.3.5, Lionello et al. (2002), Leckebusch et al. (2005), Vérant (2004), Somot (2005)
[lv] Wang et al. (2004)