Chapter 3: Observations: Surface and Atmospheric Climate Change

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Executive Summary

Global mean temperatures continue to rise, and have risen 0.8 ± 0.2°C since the late 19th century. Global-mean temperatures averaged over land and ocean surfaces, from three different estimates, each of which has been independently adjusted for various homogeneity issues, show consistent warming trends over the 1901–2005 period. The linear trends are 0.071, 0.064 and 0.060°C decade^{-1}, for estimates compiled by CRU/UKMO, NCDC and GISS, respectively, with ±2 standard error ranges of 0.02°C decade^{-1} suggesting that it is very likely that the linear warming was 0.65°C over the 20th century. However, the trend is not linear, and the warming from the first 70 years of instrumental record (1850–1919) to the last 5 years (2001–2005) is 0.78 ± 0.18°C.

Rates of surface temperature rise are greater after the mid-1970s. From 1979 to 2005 the linear trend is 0.16 to 0.18ºC decade^{-1}, for a total warming of 0.46ºC since 1979.

2005 is one of the warmest two years on record. The warmest years in the instrumental record are 1998 and 2005, with 1998 ranking first in CRU/UKMO estimate, but with 2005 slightly ahead in the NCDC and GISS estimates. 2002 to 2004 are the 3rd, 4th and 5th warmest years in the series since 1850 and 10 of the last 11 years (1995 to 2005) – the exception being 1996 – are among the 11 warmest years. Surface temperatures in 1998 were enhanced by the major 1997–1998 El Niño but no such strong anomaly was present in 2005.

Land regions have warmed at a faster rate than the oceans. Warming has occurred in both land and ocean domains, and in both sea surface temperature (SST) and night-time marine air temperature (NMAT) over the oceans. However, for both hemispheres, land temperatures have risen at about double the ocean rate after 1979 (over 0.25°C decade^{-1} versus 0.13°C decade^{-1}), with the greatest warming during winter (DJF) and spring (MAM) in the NH.

Changes in extremes of temperature are also consistent with warming of the climate. A widespread reduction in the number of frost days in mid-latitude regions, an increase in the number of warm extremes and a reduction in the number of daily cold extremes are observed. The most marked changes are for cold (lowest 10%) nights, which have become rarer over the 1951–2003 period for 76% of the land regions studied. Warm (highest 10%) nights have become more frequent across 72% of the same regions. Diurnal temperature range (DTR) averaged over the 71% of the land surface where data are available decreased by 0.07°C decade^{-1} averaged over 1950–2004, but had virtually zero change from 1979–2004. The record breaking heat wave over western and central Europe in the summer of 2003 is an example of an exceptional recent extreme. That summer (JJA) was the warmest since comparable instrumental records began around 1780 (1.4°C above the previous warmest in 1807) and is very likely to have been the warmest since 1500.

Recent warming is strongly evident at all latitudes in SSTs over each of the oceans. There are interhemispheric differences in warming in the Atlantic; the Pacific is punctuated by El Niño and Pacific decadal variability that is more symmetric about the equator, while the Indian Ocean exhibits steadier warming. These characteristics lead to important differences in regional rates of surface ocean warming that affect the atmospheric circulation.

Urban heat island effects are real but local, and have not biased the large-scale trends. A number of recent studies indicate that effects of urbanization and land-use change on the land-based temperature record (since 1950) are negligible as far as hemispheric- and continental-scale averages are concerned, because the very real but local effects are accounted for in the datasets used. In any case they are not present in the SST component of the record. Increasing evidence suggests that urban heat island effects extend to changes in precipitation, cloud and also DTR with the latter detectable as a “weekend effect” owing to lower pollution and other effects on weekends.

Lower-tropospheric temperatures have slightly greater warming rates than those at the surface over 1958–2005. The radiosonde record is markedly less spatially complete than the surface record and increasing evidence suggests that a number of records are unreliable, especially in the tropics. While there remain disparities among different tropospheric temperature trends estimated from satellite microwave sounder unit (MSU and advanced MSU, AMSU) measurements since 1979, and all likely still contain residual errors, estimates have been substantially improved (and dataset differences reduced) through adjustments for issues of changing satellites, orbit decay, and drift in local crossing time (diurnal cycle effects). It appears that the satellite tropospheric temperature record is physically consistent with surface temperature trends provided that the stratospheric influence on MSU channel 2 is accounted for, and is also in accord with ERA-40 reanalysis estimates of surface and lower-tropospheric temperature relationships. The range (due to different datasets) of global surface warming since 1979 is 0.16 to 0.18 compared to 0.12 to 0.19°C decade^{-1} for MSU estimates of tropospheric temperatures. It is likely that there is increased warming with altitude from the surface throughout the troposphere in the tropics, and a higher tropopause due also to pronounced cooling in the stratosphere.

Lower stratospheric temperatures feature cooling since 1979. Estimates from adjusted radiosondes, satellites (MSU channel 4) and reanalyses are in qualitative agreement, suggesting a lower stratospheric cooling of between 0.3 and 0.6°C decade^{-1} since 1979. Longer radiosonde records (back to 1958) also indicate cooling but the rate of cooling has been significantly greater since 1979 than between 1958 and 1978. It is likely that radiosondes overestimate stratospheric cooling, owing to changes in sondes not yet accounted for. Because of stratospheric warming episodes following major volcanic eruptions, the trends are far from being linear.

Precipitation increased over land north of 30°N over the period 1901–2005. From 10 to 30°N, precipitation increased markedly from 1901 to the 1950s, but declined after about 1970. Downward trends are present in the deep tropics from 10°N to 10°S, especially after 1976/77. Tropical values dominate the global mean. Precipitation trends were strongly up over South America south of 30°S. Patterns of precipitation change are more spatially and seasonally variable than temperature change, but where significant precipitation changes do occur they are consistent with measured changes in streamflow.

Substantial increases are found in heavy precipitation events. Consistent with a warming climate and observed significant increasing amounts of water vapour in the atmosphere, it is deemed likely that there have been increases in the number of heavy precipitation events (e.g., 95th percentile) within many land regions, even in those where there has been a reduction in total precipitation amount. Increases have also been reported for rarer precipitation events (1 in 50 year return period), but only a few regions have sufficient data to assess such trends reliably.

Droughts have become widespread in various parts of the world since the 1970s. The regions where droughts have occurred seem to be determined largely by changes in SSTs, especially in the tropics, through associated changes in the atmospheric circulation and precipitation. In the western United States, diminishing snow pack and subsequent reductions in soil moisture also appear to be a factor. In Australia and Europe, direct links to global warming have been inferred through the extreme nature of high temperatures and heat waves accompanying recent droughts. More generally, decreased precipitation and increased temperatures that enhance evapotranspiration and drying are important factors that have contributed to more regions being in droughts, as measured by the Palmer Drought Severity Index (PDSI).

Tropospheric water vapour is increasing. Surface specific humidity has generally increased after 1976 in close association with higher temperatures over both land and ocean. Total column water vapour has increased over the global oceans by 1.2 ± 0.3% (95% confidence limits) from 1988 to 2004, consistent in pattern and amount with changes in SST and a fairly constant relative humidity. Strong correlations with SST suggest that total column water vapour has increased by 4% since 1970. Similar trends have been detected in the upper troposphere from 1982–2004.

"Global dimming" is neither global in extent nor has it continued after 1990. Reported decreases in solar radiation at the Earth's surface from 1970 to 1990 have an urban bias and have reversed in sign. Although records are sparse, pan evaporation is estimated to have decreased in many places due to decreases in surface radiation associated with increases in clouds, changes in cloud properties, and/or increases in air pollution (aerosol) in different regions, especially from 1970 to 1990. However, in many of the same places actual evapotranspiration inferred from surface water balance exhibits an increase in association with enhanced soil wetness from increased precipitation, as the actual evapotranspiration becomes closer to the potential evaporation measured by the pans. Hence there is a trade-off in evapotranspiration between less solar radiation and increased wetness, with the latter generally dominating.

Cloud changes are dominated by ENSO and appear to be opposite over land and ocean. Widespread (but not ubiquitous) decreases in continental DTR since the 1950s coincide with increases in cloud amounts. Total and low-level cloud changes over the ocean disagree between surface and satellite observations. However, radiation changes at the top-of-the atmosphere from the 1980s to 1990s, possibly related in part to the El Niño Southern Oscillation (ENSO) phenomenon, appear to be associated with reductions in tropical upper-level cloud cover, and are linked to changes in the energy budget at the surface and in observed ocean heat content.

Changes in the large-scale atmospheric circulation are apparent. Atmospheric circulation variability and change is largely determined by relatively few major patterns. ENSO is the dominant mode of global-scale variability on interannual time scales although there have been times when it is less apparent. The 1976–1977 climate shift, related to the phase change in the Pacific Decadal Oscillation (PDO) and more El Niños, has affected many areas, including most tropical monsoons. For instance, over North America, ENSO and Pacific-North American (PNA) teleconnection-related changes appear to have led to contrasting changes across the continent, as the west has warmed more than the east, while the latter has become cloudier and wetter. There are substantial multi-decadal variations in the Pacific sector over the 20th century with extended periods of weakened (1900–1924; 1947–1976) as well as strengthened circulation (1925–1946; 1976–2005). Multi-decadal variability is also evident in the Atlantic as the Atlantic Multi-decadal Oscillation (AMO) in both atmosphere and ocean.

Mid-latitude westerly winds have increased in both hemispheres. These changes in atmospheric circulation are predominantly observed as “annular modes” (related to the zonally averaged mid-latitude westerlies) which have strengthened in most seasons from 1979 to the late 1990s, with poleward displacements of corresponding Atlantic and southern polar front jetstreams and enhanced storm tracks. These are accompanied by a tendency toward stronger wintertime polar vortices throughout the troposphere and lower stratosphere. In September 2002 a major stratospheric warming was observed for the first and only time in the southern hemisphere (SH) following an anomalously weak wintertime polar vortex. On monthly time scales, the southern and northern annular modes (SAM and NAM, respectively) and the North Atlantic Oscillation (NAO) are the dominant patterns of variability in the extratropics and the NAM and NAO are closely related. The increasing westerlies in the northern hemisphere (NH), as part of NAO and NAM changes, alter the flow from oceans to continents and are a major cause of the wintertime observed changes in storm tracks and related patterns of precipitation and temperature anomalies, especially over Europe. In the SH, SAM changes are associated with contrasting trends of the strong warming over the Antarctic Peninsula, and cooling over much of continental Antarctica. Wind and significant wave height analyses support reanalysis-based evidence for an increase in extratropical storm activity in the NH in recent decades.

Tropical cyclones have increased in intensity and duration since the 1970s. Variations in tropical cyclones, hurricanes and typhoons are dominated by ENSO and decadal variability, which result in a redistribution of tropical storm numbers and their tracks, so that increases in one basin are often compensated by decreases over other oceans. Trends are apparent in SSTs and other critical variables that influence tropical thunderstorm and tropical storm development. Globally, estimates of the potential destructiveness of hurricanes show a substantial upward trend since the mid-1970s, with a trend toward longer lifetimes and greater storm intensity, and such trends are strongly correlated with tropical SST. These relationships have been reinforced by findings of a large increase in numbers and proportion of hurricanes reaching categories 4 and 5 globally since 1970 even as total number of cyclones and cyclone days decreased slightly in most basins. The largest increase was in the North Pacific, Indian and Southwest Pacific Oceans. However, numbers of hurricanes in the North Atlantic have also been above normal (based on 1981–2000 averages) in 9 of the last 11 years, culminating in the record-breaking 2005 season. Moreover, the first recorded tropical cyclone in the South Atlantic occurred in March 2004 off the coast of Brazil.

3.1 Introduction

This chapter assesses the observed surface and atmospheric climate to place new observations and new analyses made during the past six years (since the TAR) in the context of the previous instrumental record. In previous IPCC reports, paleo-observations from proxy data for the pre-instrumental past and observations from the ocean and ice domains were included within the same chapter. This helped the overall assessment of the consistency among the various variables and their synthesis into a coherent picture of change. However, the amount of information became unwieldy and is now spread over Chapters 3 to 6. Nevertheless, a short synthesis and scrutiny of the consistency of all the observations is included here (see Section 3.9).

In the TAR, surface temperature trends were examined over 1860–2000 globally, for 1901 to 2000 as maps, and for three sub-periods 1910–1945, 1946–1975, and 1976–2000. The first and third sub-periods are of rising temperatures, while the second sub-period had relatively stable global mean temperatures. The 1976 divide is the date of a widely acknowledged "climate shift" (e.g., Trenberth, 1990) and seems to mark a time (see Chapter 10) when global mean temperatures began a discernible upward trend that has been attributed to increases in greenhouse gas concentrations in the atmosphere (see the TAR, IPCC, 2001). The picture prior to 1976 has essentially not changed and is therefore not repeated in detail here. However, it is more convenient to document the sub-period after 1979, rather than 1976, owing to the availability of satellite data since then (in particular TOVS data) in association with the Global Weather Experiment (GWE) of 1979. The post-1979 period allows, for the first time, a global perspective on many fields of variables, such as precipitation, that was not previously available. For instance, the reanalyses of the global atmosphere from National Centers for Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) (NCEP/NCAR, referred to as NRA) and European Centre for Medium Range Weather Forecasts (ECMWF, referred to as ERA-40) are markedly more reliable after 1979, and spurious discontinuities are present in the analyzed record at the end of 1978 (Santer et al., 1999; Bromwich and Fogt, 2004; Bengtsson et al., 2004; Trenberth et al., 2005a). Therefore the availability of high quality data has led to a focus on the post-1979 period, although physically this new regime seems to have begun in 1976–1977.

Documentation of the climate has traditionally analyzed global and hemispheric means, and land and ocean means, and has presented some maps of trends. However, climate varies on all space and time scales (from the diurnal cycle, to El Niño, to multi-decadal and millennial variations). The nature of atmospheric waves naturally creates regions of temperature and moisture of opposite-signed anomalies (departures from a base period) as moist warm conditions are favoured in poleward flow while cool dry conditions occur in equatorward flow. Although there is an infinite variety of weather systems, one area of recent substantial progress is recognition that a few preferred patterns (or modes) of variability determine the main seasonal and longer-term climate anomalies. These patterns arise from the differential effects on the atmosphere of land and ocean, mountains, and anomalous heating, such as occurs during El Niño events. The response is generally felt in regions far removed from the anomalous forcing through atmospheric teleconnections, associated with large-scale waves in the atmosphere. In this chapter we therefore document some aspects of temperature and precipitation anomalies associated with these preferred patterns, as they are vitally important for understanding regional climate anomalies and why they differ from global means. Changes in storm tracks, the jet streams, regions of preferred blocking anticyclones, and changes in monsoons all occur in conjunction with these preferred patterns and other climate anomalies. Therefore the chapter not only documents changes in variables, but also changes in phenomena, in order to increase understanding of the character of change.

Extremes of climate, such as droughts and wet spells, are very important because of their large impacts on society and the environment; but they are merely an expression of the variability. The nature of variability on different space and time scales is vital to our understanding. The global means of temperature and precipitation are most readily linked to global-mean radiative forcing and are important because they clearly indicate if unusual change is occurring. But the local or regional response can be complex and perhaps even counter-intuitive, such as global warming-induced changes in planetary waves in the atmosphere that result in regional cooling. As an indication of the complexity associated with temporal and spatial scales, Table 3.1 provides measures of the magnitude of natural variability of surface temperature in which climate signals are embedded. The measures used are indicators of the range: the mean range of the diurnal and annual cycles, and four times the standard deviation, which contains about 95% of the observations for a normal distribution. A normal distribution is a reasonable approximation in most places for temperature, with the exception of continental interiors in the cold season, which have strongly negatively skewed temperature distributions owing to cold extremes. For the global mean, the variance is somewhat affected by the observed trend, which inflates this estimate of the range slightly. The comparison highlights the large diurnal cycle and daily variability. Daily variability is, however, greatly reduced by either spatial or temporal averaging that effectively averages over synoptic weather systems. Nevertheless, even continental-scale averages contain much greater variability than the global mean in association with planetary-scale waves and events such as El Niño.

Table 3.1. Typical ranges of surface temperature on different space and time scales for a sample mid-latitude mid-continental station (Boulder, Colorado; based on 80 years of data) and for monthly mean anomalies (diurnal and annual cycles removed) for the United States as a whole and the globe for the 20th century. For the diurnal and annual cycles the monthly mean range is given, while other values are the difference between ±2 times the standard deviation.

Time and space scale	Range of temperature °C
Diurnal cycle	13.1 (December) to 15.1 (September)
Annual cycle	23
Daily anomalies	18
Monthly anomalies	8.5
United States monthly anomalies	4.7
Global mean monthly anomalies	0.96

Throughout the chapter we try to consistently indicate the degree of confidence and uncertainty in trends and other results. Quantitative estimates of uncertainty include, for the mean, twice the standard error; or for trends, statistical significance at the 0.05 (5%) level. This allows us to assess what is really unusual. We use the word "trend" to designate a generally monotonic change in the level of a variable. Where numerical values are given, they are equivalent linear trends, though more complex changes in the variable will often be clear from the description. We also assess if possible the physical consistency among different variables, which helps to provide additional confidence in trends. Where this is not possible, we use the following words to indicate judgmental estimates of confidence: virtually certain (>99% chance that a result is true); very likely (≥90% but ≤99% chance); likely (>66% but <90% chance); about as likely as not (>33% but ≤66% chance), unlikely (>10% but ≤33% chance); very unlikely (≥1% but ≤10% chance) and exceptionally unlikely (<1% chance).

3.2 Changes in Surface Climate: Temperature

3.2.1 Background

Improvements have been made to both land surface air temperature and sea surface temperature (SST) data bases during the 6 years since the TAR was published. Jones and Moberg (2003) revised and updated the Climatic Research Unit (CRU) monthly land surface air temperature record, improving coverage particularly in the SH in the late 19 th century. Minor revisions have additionally been made by Brohan et al. (2006), including a comprehensive reassessment of errors, together with an extension back to 1850. Under the auspices of the World Meteorological Organization (WMO) and the Global Climate Observing System (GCOS), daily temperature (together with precipitation and pressure) data for an increasing number of land stations have also become available, allowing more detailed assessment of extremes (see Section 3.8), as well as potential urban influences on both large-scale temperature averages and microclimate. A new gridded dataset of monthly maximum and minimum temperatures has updated earlier work (Vose et al., 2005a). For the oceans, the International Comprehensive Ocean-Atmosphere Data Set (ICOADS) has been extended by blending the former COADS with the United Kingdom's Marine Data Bank and newly digitized data, including the U.S. Maury Collection and Japan's Kobe Collection. As a result, coverage has been improved substantially before 1920, especially over the Pacific, with further modest improvements up to 1950 (Rayner et al., 2006; Worley et al., 2005). Improvements have also been made in the bias reduction of satellite-based infrared (Reynolds et al., 2002) and microwave (Reynolds et al., 2004; Chelton and Wentz, 2005) retrievals of SST for the 1980s onwards. These data represent ocean skin temperature, not air or sea surface temperature, and so must be adjusted to match the latter. Satellite infrared and microwave imagery can now also be used to monitor land surface temperature (Kwok and Comiso, 2002b; Jin and Dickinson, 2002; Peterson et al., 2000). Microwave imagery must be compensated for variations in surface emissivity and cannot act as surrogate for air temperature over either snow-covered (Peterson et al., 2000) or sea-ice areas. As satellite-based records are still short in duration, all regional and hemispheric temperature series shown in this section are based on conventional surface-based datasets, except where stated.

Despite these improvements, substantial gaps in data coverage remain, especially in the tropics and the SH, particularly Antarctica. These gaps are largest in the 19th century and during the two world wars. Accordingly, advanced interpolation and averaging techniques have been applied when creating global datasets and hemispheric and global averages (Smith and Reynolds, 2005), and these techniques have also been used in the estimation of errors (Brohan et al., 2006), both locally and on a global basis (see Appendix 3.B.1). These errors, as well as the influence of decadal and multi-decadal variability in the climate, have been taken into account when estimating linear trends and their uncertainties (see Appendix 3.A). Estimates of surface temperature from ERA-40 reanalyses have been shown to be of climate quality (i.e., without major time-varying biases) on large scales from 1979 (Simmons et al., 2004). Improvements in the ERA-40 over NRA arose both from improved data sources and better assimilation techniques (Uppala et al., 2005). Lack of satellite data before the mid-1970s and inadequate collection of sub-daily surface data before 1967 degraded the ERA-40 performance then (see Appendix 3.B.5).

3.2.2 Temperature in the Instrumental Record for Land and Oceans

3.2.2.1 Land-surface air temperature

Figure 3.2.1 shows annual global land surface air temperatures, relative to 1961–1990, from the improved analysis (CRUTEM3) of Brohan et al. (2006). Warming since 1979 has been 0.27°C decade^{-1} for the globe, but 0.33 and 0.13°C decade^{-1} for the NH and SH respectively (Table 3.2). The long-term variations are in general agreement with those from the operational version of the Global Historical Climatology Network (GHCN) dataset (NCDC, Smith and Reynolds, 2005, and Smith et al., 2005) and the Goddard Institute for Space Studies (GISS, Hansen et al., 2001) and Lugina et al. (2005) analyses (Figure 3.2.1). The NCDC analysis (which begins in 1880) is higher than the Brohan et al. (2006) analysis by between 0.1 and 0.2°C in the first half of the 20th century, very likely because it has been interpolated to be spatially complete. Most infilling techniques tend to move the analysis towards the modern climatology (1961–1990) used (Hurrell and Trenberth, 1999). CRUTEM2v (Jones and Moberg, 2003) and NCDC are in excellent agreement if the infilling by NCDC is omitted. Vose et al. (2005b) show almost identical trends when the Jones and Moberg (2003) data are gridded using the NCDC technique and *vice versa*. Global averages from CRUTEM3 and CRUTEM2v are almost identical (Brohan et al., 2006). The number of station series used by CRUTEM3, NCDC and GISS is 4349, 7230 and >7200 respectively, and much of the basic station data are in common. Differences in station numbers relate principally to CRUTEM3 requiring series to have a sufficient number of years in the 1961–90 base period in order to calculate anomalies (Brohan et al., 2006). Differences also arise from differing homogeneity adjustments by the three groups to the individual station series. The GISS analysis uses a population-based method to estimate the effects of urbanization (see also Section 3.2.2.2). For the United States, GHCN and GISS both use the 1200 stations from the U.S. Historic Climate Network (HCN), but GHCN uses a homogeneity adjusted version (Peterson et al., 1998), while GISS use a subset of these adjustments (Hansen et al., 2001). Interannual variability is lower in the GISS analysis because the gridding method favours isolated island and coastal sites, but long-term trends are similar in the NH. Lugina et al. (2005) also find reduced recent trends owing to their optimal interpolation method (cf. Hurrell and Trenberth, 1999). The major difference between CRUTEM2v and GHCN for the global average relates to whether it was calculated as one domain or as the average of the two hemispheres (Vose et al., 2005b and earlier discussion in Wigley et al., 1997). In Figure 3.2.1, the global average (for CRUTEM3 and GHCN) is a weighted sum of the two hemispheres (weights determined by the proportion of land in each hemisphere: NH 0.68 and SH 0.32).

Trends and low-frequency variability of large-scale surface air temperature from the ERA-40 reanalysis and from CRUTEM2v (Jones and Moberg, 2003) are in general agreement from the late 1970s onwards (Simmons et al., 2004). Correlations between monthly hemispheric- and continental-scale averages exceed 0.96 but trends in ERA-40 are 0.03 and 0.07°C decade^{-1} (NH and SH respectively) weaker than Jones and Moberg (2003). ERA-40 is more homogeneous than previous reanalyses (see Section 3.2.1 and Appendix

3.B.5.3) but is not completely independent of the Jones and Moberg data (Simmons et al., 2004). The warming trends continue to be greatest over the continents of the NH (see later maps of spatial trends in Figures 3.2.9 and 3.2.10), in line with the TAR. Issues of homogeneity of terrestrial air temperatures are discussed in Appendix 3.B.2.

From 1950–2004 the annual trends in minimum and maximum land surface air temperature averaged over regions with data were respectively 0.20°C decade^{-1} and 0.14°C decade^{-1}, with a trend in diurnal temperature range (DTR) of –0.07°C decade^{-1} (Vose et al., 2005a) (Figure 3.2.2). This is consistent with the TAR; spatial coverage is now 71% of the terrestrial surface instead of 54% in the TAR, although tropical areas are still under-represented. Prior to 1950, insufficient data are available to develop global-scale maps of maximum and minimum temperature trends. For 1979–2004, the corresponding linear trends for the land areas where data are available were 0.29°C decade^{-1} for both maximum and minimum temperature with no trend for DTR. DTR is particularly sensitive to observing techniques, and monitoring it requires adherence to GCOS monitoring principles (see Appendix 3.B). A map of the trend of annual DTR over the 1979–2004 period is discussed later (see Figure 3.2.11).

Table 3.2 provides trend estimates from a number of hemispheric and global temperature databases. Determining the statistical significance of a trend line in geophysical data is difficult, and many oversimplified techniques will tend to overstate the significance. Zheng and Basher (1999), Cohn and Lins (2005) and others have used time series methods to show that failure to properly treat the pervasive forms of long-term persistence and autocorrelation in trend residuals can make erroneous detection of trends a typical outcome in climatic data analysis.

[INSERT FIGURE 3.2.1 HERE]

Table 3.2. Linear trends of temperature (°C decade^{-1}) in hemispheric and global land surface air temperatures, SST and NMAT. Trends with ±2 standard error ranges and significances (**bold: <1%**; *italic, 1%–5%*) were estimated by REML (see Appendix 3.A), which allows for first-order serial correlation (AR1) in the residuals of the data about the linear trend. Annual averages, with estimates of uncertainties for CRU and HadSST2, were used to estimate trends.

	1850–2005	1901–2005	1910–1945	1946–1978	1979–2005
Land: Northern Hemisphere					
CRU (Brohan et al., 2006)	**0.063 ± 0.018**	**0.089 ± 0.030**	**0.142 ± 0.057**	–0.038 ± 0.064	**0.330 ± 0.108**
GHCN (Smith and Reynolds, 2005)		**0.072 ± 0.031**	**0.127 ± 0.065**	–0.040 ± 0.074	**0.344 ± 0.121**
GISS		**0.083 ± 0.030**	**0.166 ± 0.061**	–0.053 ± 0.062	**0.294 ± 0.090**
Lugina et al. (2005) up to 2004		**0.074 ± 0.032**	**0.144 ± 0.074**	–0.051 ± 0.061	**0.278 ± 0.096**
Land: Southern Hemisphere					
CRU (Brohan et al., 2006)	*0.034 ± 0.033*	**0.078 ± 0.054**	*0.091 ±0.076*	0.031 ±0.063	**0.135 ± 0.087**
GHCN (Smith and Reynolds, 2005)		**0.057 ± 0.020**	*0.091 ± 0.069*	0.054 ± 0.072	**0.220 ± 0.114**
GISS		**0.056 ± 0.015**	0.033 ± 0.042	*0.060 ± 0.052*	*0.085 ± 0.067*
Lugina et al. (2005) up to 2004		**0.056 ± 0.013**	**0.064 ± 0.046**	0.014 ± 0.052	*0.074 ± 0.062*
Land: Globe					
CRU (Brohan et al., 2006)	**0.054 ± 0.020**	**0.084 ± 0.026**	**0.125 ±0.042**	-0.016 ±0.055	**0.268 ± 0.084**
GHCN (Smith and Reynolds, 2005)		**0.068 ± 0.029**	**0.116 ±0.057**	–0.013 ±0.061	**0.315 ± 0.108**
GISS		**0.069 ± 0.020**	**0.102 ± 0.041**	0.003 ± 0.046	**0.188 ± 0.084**
Lugina et al. (2005) up to 2004		**0.065 ± 0.024**	**0.108 ± 0.043**	- 0.021 ± 0.059	**0.183 ± 0.075**

Ocean: Northern Hemisphere					
UKMO HadSST2 (Rayner et al., 2006)	**0.042 ± 0.019**	**0.071 ± 0.032**	**0.195 ±0.034**	0.014 ±0.082	*0.193 ± 0.146*
UKMO HadMAT1 (Rayner et al., 2003) from 1861	**0.038 ± 0.013**	**0.065 ± 0.024**	**0.176 ±0.045**	–0.047 ±0.073	**0.186 ± 0.073**
Ocean: Southern Hemisphere					
UKMO HadSST2 (Rayner et al., 2006)	**0.036 ± 0.014**	**0.068 ± 0.018**	*0.144 ±0.116*	**0.117 ±0.049**	**0.087 ± 0.049**
UKMO HadMAT1 (Rayner et al., 2003) from 1861	**0.040 ± 0.015**	**0.069 ± 0.013**	**0.100 ±0.046**	*0.075 ±0.055*	**0.092 ± 0.060**
Ocean: Globe					
UKMO HadSST2 (Rayner et al., 2006)	**0.038 ± 0.014**	**0.067 ± 0.018**	**0.153 ±0.037**	**0.071 ±0.048**	**0.133 ± 0.057**
UKMO HadMAT1 (Rayner et al., 2003) from 1861	**0.039 ± 0.012**	**0.067 ± 0.016**	**0.134 ±0.041**	0.021 ±0.050	**0.135 ± 0.054**

[INSERT FIGURE 3.2.2 HERE]

3.2.2.2 Urban heat islands and land-use effects

The modified land surface in cities affects the storage and radiative and turbulent transfers of heat and its partition into sensible and latent components, see Chapter 7, Section 7.2 and Box 7.2. The relative warmth of a city compared with surrounding rural areas, known as the Urban Heat Island (UHI) effect, arises from these changes and may also be affected by changes in water runoff, pollution and aerosols. UHIs are often very localized and depend on local climate factors such as windiness and cloudiness (which in turn depend on season), and on proximity to the sea. Section 3.3.2.4 discusses impacts of urbanization on precipitation.

Many local studies have demonstrated that the microclimate within cities is on average warmer, with smaller DTR than if the city were not there. However, the key issue from a climate change standpoint is whether urban-affected temperature records have significantly biased large-scale temporal trends. Studies that have looked at hemispheric and global scales conclude that any urban-related trend is an order of magnitude smaller than decadal and longer timescale trends evident in the series (e.g., Jones et al., 1990), a result that could partly be attributed to the omission from the gridded dataset of a small number of sites (<1%) with clear urban-related warming trends. In a worldwide set of about 270 stations, Parker (2006) noted that warming trends in night minimum temperatures over 1950–2000 were not enhanced on calm nights, which would be the time most likely to be affected by urban warming. Thus, the global land warming trend discussed is very unlikely to be influenced significantly by increasing urbanization (Parker, 2006). Over the conterminous USA, rural station trends were almost indistinguishable from series including urban sites (Peterson, 2003; and Figure 3.2.3 from Peterson and Owen, 2005), and the same is true of China from 1951–2001 (Li et al., 2004). One possible reason for the patchiness of UHIs is the location of observing stations in parks where urban influences are reduced (Peterson, 2003). In summary, although some individual sites may be affected, even some small rural locations, the UHI effect is not pervasive as all global-scale studies indicate it is a negligible component of large-scale averages.

Comparing surface temperature estimates from the NRA with raw station time series, Kalnay and Cai (2003) concluded that more than half of the observed decrease in DTR in the eastern USA since 1950 was due to changes in land use, which are partly related to urbanization. This conclusion was based on the fact that the reanalysis did not include these factors which would affect the observations. But the reanalysis also did not include many other natural and anthropogenic effects, such as increasing greenhouse gases and observed changes in clouds or soil moisture (Trenberth, 2004). Vose et al. (2004) show that the adjusted station data for the region (for various issues affecting homogeneity, see Appendix 3.B.2) do not support Kalnay and Cai's conclusions. Nor are the results reproduced in the surface temperature fields from the ERA-40 reanalyses (Simmons et al., 2004). Instead most of the changes appear related to abrupt changes in the type of data assimilated into the reanalysis, rather than to gradual changes over the period arising from land-use and urbanization changes. Reanalyses may be reliable for estimating trends since 1979 (Simmons et al., 2004) but are in general unsuited for estimating longer-term global trends as discussed in Appendix 3.B.5.

Nevertheless, regional changes in land use can be important for DTR at the local-to-regional scale. For instance, land degradation in northern Mexico resulted in an increase in DTR relative to locations across the border in the United States (Balling et al., 1998), and cropland effects on maximum temperatures in the United States (Bonan, 2001). Desiccation of the Aral Sea since 1960 raised DTR locally around this region (Small et al., 2001). By processing maximum and minimum temperature data as a function of day of the week, Forster and Solomon (2003) found a distinctive "weekend effect" in DTR at stations examined in the United States, Japan, Mexico, and China. The weekly cycle in DTR has a distinctive large-scale pattern with geographically-varying sign and strongly suggests an anthropogenic effect on climate, likely through changes in pollution and aerosols (Jin et al., 2005). Chapter 7, Section 7.2 provides fuller discussion of the effects of land use changes.

[INSERT FIGURE 3.2.3 HERE]

3.2.2.3 Sea surface temperature and marine air temperature

Most analyses of SST estimate the sub-surface bulk temperature, i.e., the temperature in the uppermost few metres of the ocean, not the ocean skin temperature measured by satellites. For maximum resolution and data coverage, polar-orbiting infrared satellite data can be used since 1981 so long as the satellite ocean skin temperatures are adjusted to estimate bulk SST values through a calibration procedure (see e.g., Reynolds et al., 2002; Rayner et al., 2003, 2006 and Appendix 3.B.3). But satellite SST data alone have not been used as a major resource for estimating climate change because of their strong time-varying biases which are hard to completely remove e.g., as shown in Reynolds et al. (2002) for the Pathfinder polar orbiting satellite SST dataset (Kilpatrick et al., 2001). Figures 3.2.4b, 3.2.9 and 3.2.10 (see later) do, however, make use of adjusted satellite SST estimates after November 1981 to provide nearer-to-global coverage for the 1979–2004 period, and O'Carroll et al. (2006) have developed an analysis based on Along-Track Scanning Radiometers (ATSRs) with potential for the future. Even satellite data are unable to fill in estimates of surface temperature over or near to sea-ice areas. Recent bulk SSTs estimated using ship and buoy data also have time-varying biases (e.g., Christy et al., 2001; Kent and Kaplan, 2006), larger than originally estimated by Folland et al. (1993), but not large enough to prejudice conclusions about recent warming (see Appendix 3.B.3.).

A combined physical-empirical method (Folland and Parker, 1995) is mainly used, as reported in the TAR, to estimate adjustments to ship SST data obtained up to 1941 to compensate for heat losses from uninsulated (mainly canvas) or partly insulated (mainly wooden) buckets. The adjustments are independent of land-surface air temperature or night marine air temperature (NMAT) data measured by ships. Confirmation that these spatially and temporally complex adjustments are realistic globally, in many ocean regions and also seasonally is shown by comparison of the Jones and Moberg (2003) land-surface air temperature anomalies with simulations using the Hadley Centre atmospheric climate model HadAM3 forced with observed SST and sea-ice extents since 1871 (Folland, 2005). Smith and Reynolds (2002) have independently bias-adjusted updated COADS (Slutz et al., 1985) SST anomalies to agree with COADS NMAT anomalies before 1942 (see also Appendix 3.B.3) and derive rather similar spatiotemporal adjustments to Folland and Parker (1995), although there are seasonal differences. Overall, they recommend use of the Folland and Parker (1995) adjustments as these are independent of any changes in NMAT data and more fully take into account evaporation errors in uninsulated buckets, especially in the tropics. Smith and Reynolds (2004) analysis of ICOADS (formerly COADS Release 2.0, Woodruff et al., 1998) requires SST bias adjustments before 1942 similar to those of Smith and Reynolds (2002), except in 1939–1941 when ICOADS contains a new data source which clearly has many more engine intake data that do not need adjustment. Rayner et al. (2006), in a new analysis of the ICOADS data with no interpolation, adapt the Folland and Parker (1995) adjustments in 1939–1941 in a similar way to Smith and Reynolds (2004) but, unlike Smith and Reynolds (2005), do not widen the error bars because the new adjustments are compatible with well-understood changes in the data.

The Smith and Reynolds (2004) analysis is interpolated to fill missing data areas, like that of Rayner et al. (2003). The main problem for estimating climate variations in the presence of large data gaps is underestimation of change, as most interpolation procedures tend to bias the analysis towards the modern climatologies used in these datasets (Hurrell and Trenberth, 1999). To deal with non-stationary aspects, Rayner et al. (2003) extracted the leading global covariance pattern, which represents long-term changes, before interpolating using reduced-space optimal interpolation (see Appendix 3.B.1); and Smith and Reynolds removed a smoothed, moving 15-year average field before interpolating by a related technique.

Figure 3.2.4a shows annual and decadally smoothed anomalies of global SST from the new, uninterpolated HadSST2 analysis (Rayner et al., 2006). NMAT (referred to as HadMAT1), used to avoid daytime heating of ship decks (Bottomley et al., 1990), is also shown. The HadMAT1 analysis includes limited optimal interpolation (Rayner et al., 2003) and has been chosen because of the demonstration by Folland et al. (2003) of its skill in the sparsely observed South Pacific from the late 19th century onwards, but major gaps e.g. the Southern Ocean are not interpolated. Although HadMAT1 data have been corrected for warm biases in World War II they may still be too warm in the NH and too cool in the SH at that time (Figures 3.2.4c,d). HadSST2 and HadMAT1 generally agree well, especially after the 1880s. The SST analysis in the TAR is included in Figure 3.2.4a. The changes in SST since the TAR are generally fairly small, though the new SST analysis is warmer around 1880 and cooler in the 1950s. The peak warmth in the early 1940s is likely to have arisen partly from closely-spaced multiple El Niño events (Brönnimann et al., 2004, see also 3.6.2) and also due to the warm phase of the Atlantic Multidecadal Oscillation (AMO, see Section 3.6.6.1). HadMAT1 generally confirms the hemispheric SST trends in the 20th century (Figures 3.2.4c, d and Table 3.2). Overall, the SST data should be regarded as more reliable because averaging of fewer samples is needed for SST than for HadMAT1 to remove synoptic weather noise. However, the relative changes of SST compared to NMAT since 1991 in the tropical Pacific may be partly real (Christy et al., 2001). As the atmospheric circulation changes, the relationship between SST and surface air temperature anomalies can change along with surface fluxes. Interannual variations in the heat fluxes into the atmosphere can exceed 100 W m^{-2} locally in individual months, but the main prolonged variations occur with ENSO, where changes in the central tropical Pacific exceed ±50 W m^{-2} for many months during major ENSO events (Trenberth et al., 2002a).

Figure 3.2.4b shows three time series of changes in global SST. The HadSST2 series (as in Figure 3.2.4a) does not include polar orbiting satellite data because of possible time-varying biases that remain difficult to correct fully (Rayner et al., 2003), though the data (Reynolds et al., 2002) used by NCDC do include satellite data from 1981. The Japanese (Ishii et al., 2005, referred to as COBE-SST) series is also *in situ* except for the specification of sea-ice. The warmest year globally in each SST record was 1998 (HadSST2 0.44°C, NCDC 0.38°C, COBE 0.37°C above the 1961 to 1990 average). The 5 warmest years in all analyses have occurred after 1995. The NCDC analysis is in principle affected by artificially reduced trends in the satellite data (Hurrell and Trenberth, 1999), though the data we show include recent attempts to reduce this.

Our understanding of the variability and trends in different oceans is still developing, but it is already apparent that they are quite different. The Pacific is dominated by ENSO and modulated by the PDO, which may provide ways of moving heat from the tropical ocean to higher latitudes and out of the ocean into the atmosphere (Trenberth et al., 2002a), thereby greatly altering how trends are manifest. In the Atlantic, observations reveal the role of the AMO (see Section 3.6.6.1 and Figure 3.6.8 later) (Folland et al., 1999; Delworth and Mann, 2000; Goldenberg et al., 2001, Enfield et al., 2001) which probably has wider influence than just the North Atlantic (Minobe, 1997; Folland et al., 1999; Chao et al., 2000; Knight et al., 2005). The AMO is likely to be associated with the Thermohaline Circulation (THC), which transports heat northwards, thereby moderating the tropics and warming the high latitudes. In the Indian Ocean, interannual variability is small compared with the trend. So we present in Figure 3.2.5 latitude-time sections from 1900 for SSTs (from HadSST2) for the zonal mean across each ocean, filtered to remove fluctuations less than 6 years or so, including the ENSO signal. In the Pacific the long-term warming is clearly evident, but punctuated by cooler episodes centred in the tropics, and no doubt linked to the PDO. The prolonged El Niño of 1939–1942 shows up as a warm interval. In the Atlantic, the warming from the 1920s to about 1940 in the NH was focussed on higher latitudes, with the SH remaining cool. This interhemispheric contrast is believed to be one signature of the THC. The subsequent relative cooling in the NH extratropics and the more recent intense warming in NH mid-latitudes was predominantly a multi-decadal variation of SST; only in the last decade is an overall warming signal clearly emerging. So the recent strong warming appears to be related in part to the AMO plus a global warming signal. The cooling in the north-western North Atlantic just south of Greenland, reported in the SAR, has now been replaced by strong warming (see also Figures 3.2.9 and 3.2.10; also Figures 5.2.1 and 5.2.2 for ocean heat content). The Indian Ocean also reveals a warm interval, poorly observed, in the early 1940s, and further shows the fairly steady warming in recent years. The multi-decadal variability in the Atlantic is much longer in time scale than that in the Pacific, but it is noteworthy that all oceans exhibit a warm period around the early 1940s.

[INSERT FIGURE 3.2.4 HERE]

[INSERT FIGURE 3.2.5 HERE]

3.2.2.4 Land and sea combined temperature: globe, NH, SH and zonal means

Gridded datasets combining land surface air temperature and SST anomalies have been developed and maintained by three groups: CRU with the UKMO Hadley Centre in the UK (HadCRUT3, Brohan et al., 2006), NCDC (Smith and Reynolds, 2005) and GISS (Hansen et al., 2001) in the United States. Although the component datasets differ slightly (see Sections 3.2.2.1 and 3.2.2.3) and the combination methods also differ, trends are similar. Comparative estimates of linear trends are given in Table 3.3. Overall warming since 1901 has been a little less in the NCDC data than in the HadCRUT3 data. All series indicate that the warmest 5 years have occurred after 1997, although there is slight disagreement about the ordering. HadCRUT3 have 1998 as the warmest, while NCDC and GISS have 2005. The GISS analysis of 2005 extrapolated over the Arctic Ocean to the exceptionally warm conditions in the extreme north of Eurasia and North America (see Figure 3.2.5). If the GISS data for 2005 are averaged only south of 75°N then 2005 is cooler than 1998. Also there were relative cool anomalies in 2005 in HadCRUT3 in parts of Antarctica and the Southern Ocean, where sea-ice coverage (see Chapter 4) has not declined.

Hemispheric and global series based on Brohan et al. (2006) are shown in Figure 3.2.6 and tropical and polar series in Figure 3.2.7. The recent warming is strongest in the NH extratropics, while El Niño events are clearly evident in the tropics, particularly the 1997/1998 event giving the warmest year. Overall the Arctic (north of 65°N) average annual temperature has increased since the 1960s and is now warmer (at the decade timescale) than during the 1920–1945 period. 2005 is the warmest year in the Arctic series. However, this period of recent warmth is not yet quite as long as that in the early-to-mid 20th century. Spatial patterns of Arctic warmth in the two periods (1920–1945 and since 1990) are also quite different, the latter being the more consistent with changes in the NAM (see Section 3.6.4) (Polyakov et al., 2003). Temperatures over mainland Antarctica (south of 65°S) have decreased slightly since 1966 (Doran et al., 2002), but there has very likely been strong warming over the last 50 years in the Antarctic Peninsula region (Turner et al., 2005), see Figure 3.6.7.

3.2.2.5 Consistency between land and ocean surface temperature changes

The course of temperature change over the 20th century, revealed by the independent analysis of land air temperatures, SST and NMAT, is remarkably consistent (Figure 3.2.8). Warming occurred in two distinct phases, 1915–1945 and since 1975; it has been substantially stronger over land than over the oceans in the later phase, as shown also by the trends in Table 3.2. The land component has also been more variable from year to year (compare Figures 3.2.1 and 3.2.4 a,c,d). Much of the recent difference in trend between global SST (and NMAT) and global land air temperature has arisen from accentuated warming over the continents in the mid-latitude NH (Figures 3.2.9 and 3.2.10), which is likely related to greater evaporation and heat storage in the ocean, and in particular to atmospheric circulation changes in the winter half year due to the NAO/NAM (see discussion in Section 3.6.4).

Table 3.3. Linear trends (°C decade^{-1}) in hemispheric and global combined land surface air temperatures and SST. Trends are estimated and presented as in Table 3.2. Annual averages, along with estimates of uncertainties for CRU/UKMO, were used to estimate trends. R^2 is the squared trend correlation in percent. The Durbin Watson D-statistic (not shown) for the residuals, after allowing for first-order serial correlation, never indicates significant positive serial correlation.

	1850–2005	1901–2005	1910–1945	1946–1978	1979–2005
Northern Hemisphere					
CRU/UKMO (Brohan et al., 2006)	**0.047 ± 0.016 R^2=54**	**0.075 ± 0.028 R^2=64**	**0.184 ±0 .040 R^2=79**	–0.005 ± 0.057 R^2=0	**0.234 ± 0.085 R^2=70**
NCDC (Smith and Reynolds, 2005)		**0.063 ± 0.027 R^2=55**	**0.157 ± 0.041 R^2=77**	–0.029 ± 0.050 R^2=6	**0.245 ± 0.075 R^2=72**
Southern Hemisphere					
CRU/UKMO (Brohan et al., 2006)	**0.038 ± 0.017 R^2=51**	**0.068 ± 0.020 R^2=74**	**0.144 ± 0.101 R^2=63**	**0.101 ± 0.050 R^2=47**	**0.095 ± 0.046 R^2=49**

NCDC (Smith and Reynolds, 2005)		**0.066 ± 0.010 R^2=82**	**0.101 ± 0.054 R^2=64**	**0.076 ± 0.049 R^2=42**	**0.096 ± 0.046 R^2=58**
Globe					
CRU/UKMO (Brohan et al., 2006)	**0.042 ± 0.014 R^2=57**	**0.071 ± 0.021 R^2=74**	**0.155 ± 0.031 R^2=81**	*0.048 ± 0.044 R^2=19*	**0.165 ± 0.056 R^2=67**
NCDC (Smith and Reynolds, 2005)		**0.064 ± 0.020 R^2=71**	**0.132 ± 0.033 R^2=79**	0.019 ± 0.044 R^2=4	**0.174 ± 0.062 R^2=72**
GISS		**0.060 ± 0.017 R^2=70**	**0.117 ± 0.031 R^2=75**	0.003 ± 0.036 R^2=0	**0.170 ± 0.058 R^2=67**

[INSERT FIGURE 3.2.6 HERE]

[INSERT FIGURE 3.2.7 HERE]

[INSERT FIGURE 3.2.8 HERE]

[INSERT FIGURE 3.2.9 HERE]

[INSERT FIGURE 3.2.10 HERE]

[INSERT FIGURE 3.2.11 HERE]

3.2.2.6 Temporal variability of global temperatures and recent warming

The standard deviation of the HadCRUT3 annual average temperatures for the globe for 1850–2005 in Figure 3.2.6 is 0.24°C. The greatest difference between two consecutive years in the global average since 1901 is 0.29°C between 1976 and 1977, demonstrating the importance of a 0.65°C warming (linear trend estimate) in a century time-scale context. However, 0.65°C is small compared with interannual variations at one location, and much smaller than day-to day variations (Table 3.1). The long-term warming is also not uniform but instead is characterised by two warming episodes punctuated by a period of little change in between. A linear trend is therefore a poor approximation to the actual course of the record over the 105 years.

The principal conclusion from this subsection is that the global average surface temperature linear trend has very likely been slightly more than 0.65 ± 0.2°C over the period from 1901 to 2005 (Table 3.3), a warming greater than any since at least the 11th Century (see Chapter 6). However, the record can also be characterized as level prior to about 1915, a warming to about 1945, levelling out or even slightly decreasing until the 1970s, and a fairly linear upward trend since then (see Figures 3.2.6 and Question 3.1). The overall warming from the average of the first 70-year 1850–1919 period through 2001 to 2005 is 0.78 ± 0.18°C, with 0.5°C increase occurring after the mid-1970s. Clearly the world's surface temperature has continued to increase since the TAR and the trend when computed in the same way as in the TAR remains 0.6°C over the 20th Century. In view of Section 3.2.2.2 and the dominance of the globe by ocean, the influence of urbanization on these estimates is expected to be small. The last 11 complete years (1995 to 2005) now contain ten of the eleven warmest years since comparable records can be developed from 1850. Only 1996 is not in this list – replaced by 1990. 2002–2005 are the 3rd, 4th, 5th and 2nd warmest years in the series with 1998 the warmest. The HadCRUT3 surface warming trend over 1979–2005 was 0.17°C decade^{-1}, i.e. a total warming of 0.46 ± 0.15°C (error bars overlap those of NCDC and GISS). During 2001 to 2005 the global average temperature anomaly has been 0.44ºC above the 1961–1990 average.

3.2.2.7 Spatial patterns of trend

Figure 3.2.9 illustrates the spatial patterns of annual surface temperature changes for 1901–2005 and 1979–2005, and Figure 3.2.10 shows seasonal trends for 1979–2005. All maps clearly indicate that differences in trends between locations can be large, particularly for shorter time periods. For the century-long period, warming is evident over most of the world's surface with the exception of an area south of Greenland and three smaller regions over the southeastern United States and parts of Bolivia and the Congo basin. The lack of significant warming at about 20% of locations (Karoly and Wu, 2005) is likely to be a result of changes in atmospheric circulation (see Section 3.6). Warming was strongest over the continental interiors of Asia and northwestern North America and also over some mid-latitude ocean regions of the SH as well as southeastern Brazil. In the recent period, some regions have warmed substantially while a few have cooled slightly on an annual basis (Figure 3.2.9). Southwest China has cooled since the mid-20th Century (Ren et al., 2005), but most of the cooling locations since 1979 have been oceanic and in the SH, possibly through changes in atmospheric and oceanic circulation related to the Pacific Decadal Oscillation (PDO) and Southern Annular Mode (SAM) (see discussion in Section 3.6.5). Warming dominates most of the seasonal maps for the period 1979 onwards, but weak cooling has affected a few regions, especially the mid-latitudes of the SH oceans, but also over eastern Canada in spring, possibly in relation to the strengthening NAO (Figure 3.6.5). Warming in this period was strongest over western North America, northern Europe and China in DJF, Europe and northern and eastern Asia in MAM, Europe and North Africa in JJA and northern North America, Greenland and eastern Asia in SON (Figure 3.2.10).

No single location follows the global average, and the only way to monitor the globe with any confidence is to include observations from as many diverse places as possible. We can, however, assess the importance of regions without adequate records by studying complete model reanalysis fields to give us clues as to the size of missing regions we can tolerate and still produce a reliable global or hemispheric series. The importance of these missing areas for hemispheric and global averages is incorporated into the errors bars in Figure 3.2.6a (see Smith and Reynolds, 2005 and Brohan et al., 2006). Error bars are generally larger in the data-sparser SH than the NH; they are larger before the 1950s and largest of all in the 19th century.

Figure 3.2.11 shows annual trends in DTR over 1979–2004. The decline in DTR since 1950 reported in the TAR has now ceased, as confirmed by Figure 3.2.2. Trends in DTR tended to be largest in magnitude in the winter in the Northern Hemisphere (Vose et al., 2005a). Daily minimum temperature increased in most areas except western Australia and southern Argentina, and parts of the western Pacific Ocean and daily maximum temperature also increased in most regions except northern Peru, northern Argentina, northwestern Australia, and parts of the North Pacific Ocean (Vose et al., 2005a). The changes reported here appear inconsistent with Dai et al. (2006) who report decreasing DTR in the United States, but this arises partly because Dai et al. (2005) included the high DTR years 1976–1978. Furthermore, Figure 3.2.11 is supported by many other recent regional-scale analyses.

Changes in cloud cover and precipitation explained up to 80% of the variance in historical DTR series for the United States, Australia, mid-latitude Canada, and the former Soviet Union during the 20th century (Dai et al., 1999). Cloud cover accounted for nearly half of the change in the DTR in Fennoscandia during the 20th century (Tuomenvirta et al., 2000). Variations in atmospheric circulation also affect DTR. Changes in the frequency of certain synoptic weather types resulted in a decline in DTR during the cold half-year in the Arctic (Przybylak, 2000). A positive phase of the NAM (see Section 3.6.4) is associated with increased DTR in the northeastern United States and Canada (Wettstein and Mearns, 2002). Variations in sea level pressure patterns and associated changes in cloud cover partially accounted for increasing trends in cold-season DTR in the northwestern United States and decreasing trends in the south-central United States (Durre and Wallace, 2001). The relationship between DTR and anthropogenic forcings is complex, as these forcings can affect atmospheric circulation, as well as clouds through both greenhouses gases and aerosols.

3.3 Changes in Surface Climate: Precipitation, Drought and Surface Hydrology

3.3.1 Background

Temperature changes are one of the more obvious and easily measured changes in climate, but atmospheric moisture, precipitation and atmospheric circulation also change, as the whole system is affected. Radiative forcing alters heating, and at the Earth's surface this directly affects evaporation as well as sensible heating, see Box 7.1. Further, increases in temperature lead to increases in the moisture holding capacity of the atmosphere at a rate of about 7% K^{-1}. Together these effects alter the hydrological cycle, especially characteristics of precipitation (amount, frequency, intensity, duration, type) and extremes (Trenberth et al., 2003). The extremes are dealt with in Section 3.8.2.2. Expectations for changes in overall precipitation amounts are complicated by aerosols. Because aerosols block the sun, surface heating is reduced. Absorption of radiation by some, notably carbonaceous, aerosols directly heats the aerosol layer that may otherwise have been heated by latent heat release following surface evaporation, thereby reducing the hydrological cycle. As aerosol influences tend to be regional, the net expected effect on precipitation over land is especially unclear. This section discusses most aspects of the surface hydrological cycle, except that surface water vapour is included with other changes in atmospheric water vapour in Section 3.4.2.

Difficulties in the measurement of precipitation remain an area of concern in quantifying the extent to which global and regional scale precipitation has changed (see Appendix 3.B.4). *In situ* measurements are especially impacted by wind effects on the gauge catch, especially for snow but also for light rain. For remotely-sensed measurements (radar and space-based), the greatest problems are that only measurements of instantaneous rate can be made, together with uncertainties in algorithms for converting radiometric measurements (radar, microwave, infrared) into precipitation rates at the surface. Because of measurement problems, and because most historical *in situ* based precipitation measurements are taken on land leaving the majority of global surface area under sampled, it is useful to examine the consistency of changes in a variety of complementary moisture variables, including both remotely-sensed and gauge-measured precipitation, drought, evaporation, atmospheric moisture, soil moisture and stream flow.

3.3.2 Changes in Large-scale Precipitation

3.3.2.1 Global land areas

Trends in global annual land precipitation were analyzed using data from the Global Historical Climatology Network (GHCN), using anomalies with respect to the 1981–2000 base period (Peterson and Vose, 1997; Vose et al., 1992). The observed GHCN linear trend (Figure 3.3.1) over the 105-year period from 1901–2005 is statistically insignificant, as is the CRU linear trend up to 2002 (Table 3.4b). However, the global mean land changes (Figure 3.3.1) are not at all linear, with an overall increase until the 1950s, a decline until the early 1990s then a recovery. Also, the global land mean is not a very meaningful quantity as it is made up of much larger regional anomalies of opposite sign.

There are several other global land precipitation data sets covering more recent periods and Table 3.4a gives their characteristics. The linear trends and their significance are given in Table 3.4b. There are a number of differences in processing, data sources and time periods that lead to the differences in the trend estimates. All but one data set (GHCN) are spatially infilled by either interpolation or the use of satellite estimates of precipitation. The PREC/L data (Chen et al., 2002) include both GHCN and synoptic data from the NOAA/Climate Prediction Center's Climate Anomaly Monitoring System (CAMS), and the Global Precipitation Climatology Project (GPCP) data (Adler et al., 2003) are a blend of satellite and gauge data. The Global Precipitation Climatology Centre (GPCC) (updated from Rudolf et al., 1994) provides monthly data from surface gauges on several grids constructed using GPCC sources (including data from CRU, GHCN, an FAO database and many nationally provided datasets). The dataset designated GPCC VasClim uses only those quasi-continuous stations whose long-term homogeneity can be assured, while GPCC v.3 has used all available stations to provide more complete spatial coverage. Gridding schemes also vary and include optimal interpolation and grid-box averaging of areally weighted station anomalies. The CRU dataset is from Mitchell and Jones (2005).

For 1951–2004 trends range from –7 to +2 mm decade^{-1} and standard errors range from 2.0 to 3.2 mm decade^{-1}. Only the updated PREC/L series (Chen et al., 2002) trend and the GPCC v.3 trend appear to be statistically significant, but the uncertainties, as seen in the different estimates, undermine that result. For 1979–2004, GPCP is added and trends range from –16 to +13 mm decade^{-1} but none is significant. Nevertheless, the discrepancies in trends are substantial, and highlight the difficulty of monitoring a variable such as precipitation which has large variability in both space and time. On the other hand, Figure 3.3.1 also suggests that interannual fluctuations have some overall reproducibility for land as a whole. The lag-1 autocorrelation of the residuals from the fitted trend (i.e., the detrended persistence) is in the range 0.3 to 0.5 for the PREC/L, CRU and GHCN series but 0.5 to 0.7 for the two GPCC and the GPCP series. This suggests that either the limited sampling by *in situ* gauge data adds noise, or systematic biases lasting a few years (the lifetime of a satellite) are afflicting the GPCP data, or a combination of the two.

Table 3.4. a. Characteristics and references of the six global land area precipitation data sets used to calculate trends.

Series	Period of Record	Gauge only	Satellite and gauge	Spatial infilling	Reference
GHCN	1900–2005	x		No	Vose et al., 1992
PREC/L	1948–2002	x		Yes	Chen et al., 2002
GPCP	1979–2002		x	Yes	Adler et al., 2003
GPCC VasClim	1951–2000	x		Yes	Rudolf et al., 1994
GPCC v.3	1951–2002	x		Yes	Rudolf et al., 1994
CRU	1901–2002	x		Yes	Mitchell and Jones, 2005

b. Global land precipitation trends (mm decade^{-1}). Trends, ±2 standard error ranges and significances (*italic, 1%–5%*) were estimated by REML (see Appendix 3.A) which allows for serial correlation in the residuals of the data about the linear trend. All trends are based on annual averages without estimates of intrinsic uncertainties.

Series	1901–2005	1951–2005	1979–2005
PREC/L		*–5.10 ± 3.95* [a]	–6.38 ± 10.68 [a]
CRU	1.10 ± 1.82[a]	–3.87 ± 4.73 [a]	–0.90 ± 19.75 [a]
GHCN	1.08 ± 2.27	–4.56 ± 5.28	4.16 ± 15.13
GPCC VasClim		1.82 ± 6.47 [b]	12.82 ± 26.08 [b]
GPCC v.3		*–6.63 ± 6.30* [a]	–14.64 ± 14.19 [a]
GPCP			–15.6 ± 24.12 [a]

Notes:
(a) Series ends at 2002
(b) Series ends at 2000

[INSERT FIGURE 3.3.1 HERE]

3.3.2.2 Spatial patterns of precipitation trends

The spatial patterns of trends in annual precipitation (% per century or decade) during the periods 1901–2005 and 1979–2005 are shown in Figure 3.3.2. The observed trends over land areas were calculated using GHCN station data interpolated to a 5° × 5° latitude/longitude grid. For most of North America, and especially over high latitude regions in Canada, annual precipitation has increased during the 105 years. The primary exception is over the southwest United States, northwest Mexico and the Baja Peninsula, where the trend in annual precipitation has been negative (1 to 2% decade^{-1}) as drought has prevailed in recent years. Across South America, increasingly wet conditions were observed over the Amazon Basin and southeastern South America, including Patagonia, while negative trends in annual precipitation were observed over Chile and parts of the western coast of the continent. The largest negative trends in annual precipitation was observed over western Africa and the Sahel. After having concluded that the effect of changing rainfall gauge networks on Sahel rainfall time series is small, Dai et al. (2004a) note that Sahel rainfall in the 1990s has recovered considerably from the severe dry years in the early 1980s (see Figure 3.7.4 and also Sections 3.7.1.4 and 3.8.3.3). A drying trend is also evident over southern Africa. Over much of northwestern India the 1901–2005 period shows increases of more than 20%, but the same area shows a strong decrease in annual precipitation for the 1979–2005 period. Northwestern Australia shows areas with moderate to strong increases in annual precipitation over both periods. Over most of Eurasia, increases in precipitation outnumber decreases for both periods.

[INSERT FIGURE 3.3.2 HERE]

[INSERT FIGURE 3.3.3 HERE]

To assess the expected large regional variations in precipitation trends, Figure 3.3.3 presents time series of annual precipitation (expressed as % of the 1961–1990 mean, given in the top left of each plot). The regions are 19 of the 23 defined in Chapter 11.3.1 and include: Central North America, Western North America, Alaska, Central America, Eastern North America, Mediterranean, Northern Europe, North Asia, East Asia, Central Asia, Southeast Asia, Southern Asia, Northern Australia, Southern Australia, Eastern Africa, Western Africa, Southern Africa, Southern South America, and the Amazon. The GHCN precipitation from NCDC was used for the annual green bars and the black decadal curve, and for comparison the CRU decadal values are given in magenta. This allows the reproducibility to be assessed and, based on this, plots for four additional regions (Greenland, Sahara, Antarctica and the Tibetan Plateau) are not included as precipitation data for these were not considered sufficiently reliable, and nor was the first part of the Alaskan series, prior to 1935. In most panels the range is +30% to –30%, except for the two Australian panels. Some discrepancies are still evident at times between the decadal variations, mostly owing to different subsets of stations and also some stations coming in or dropping out, but overall the confidence in what is presented is quite high. A latitude-time series of zonal averages over land is also presented (Figure 3.3.4).

In the tropics, precipitation is highly seasonal, consisting of a dry season and a wet season in association with the summer monsoon. These aspects are discussed in more detail in Section 3.7.1. Downward trends are strongest in the Sahel (discussed in Section 3.7.1.4) but occur in both Western and Eastern Africa in the past 50 years, and are reflected in the zonal means. The downward trends in this zone are also found in Southern Asia. The linear trend rainfall decreases for 1900–2005 were 7.5% in both Western Africa and Southern Asia regions (significant at <1%). This last region is much greater than India, whose rainfall features strong variability but little in the way of a century-scale trend. Also featuring a strong overall downward trend is Southern Africa, although with strong multi-decadal variability present. Often the change in rainfall in these regions occurs fairly abruptly, and in several cases occurs around the same time in association with the 1976-1977 climate shift (Wang and Ding, 2006). The timing is not the same everywhere, however, and the downward shift occurred earlier in the Sahel (see also Figure 3.7.4 and associated discussion). The main location with different trends in low latitudes is over Australia, but it is clear that large interannual variability, mostly ENSO-related, is dominant (note also the expanded vertical scales for Australia). The apparent upward trend occurs from two rather wet spells in northern Australia in the early 1970s and 1990s, when it was dry in Southeast Asia, see also Section 3.7.1.2. Also of note in Australia is the marked downward trend in the far southwest (Figure 3.3.2).

At higher latitudes, quite distinct upward trends are evident in many regions and these are reflected in the zonal means (Figure 3.3.4): Central North America, Eastern North America, Northern Europe, Northern Asia and Central Asia (east of the Caspian Sea) all experienced upward linear trends of between 6 and 8% from 1900 to 2005 (all significant at <5%). These regions all experience snowfall (see also Section 3.3.2.3) and part of the upward trend may arise from changes in efficiency of catching snow, especially in Northern Asia. However, there is ample evidence that these trends are real (see Section 3.3.4), and they extend from North America to Europe across the North Atlantic as evidenced by ocean freshening, documented in Chapter 5, Sections 5.2.3, and 5.3.2. Western North America shows longer timescale variability, principally due to the severe drought in the 1930s and lesser events more recently. Note the tendency for inverse variations between Northern Europe and the Mediterranean, associated with changes in the NAO (see Section 3.6.4). Central and southern Europe is characterized by a drier winter (DJF) during the positive phase of the NAO, while the reverse is true in the British Isles, Fennoscandia and northwestern Russia.

In the SH, Amazonia and Southern South America feature opposite changes, as the South American monsoon features shifted southwards (see Section 3.7.1.3), also in association with changes in ENSO and the 1976–1977 climate shift. The result is a pronounced upward trend in Argentina and the La Plata river basin, but not in Chile (where the main declines in precipitation are evident in the austral summer (DJF) and autumn (MAM)) (Figure 3.3.2). Decadal-scale variations over Amazonia are also out of phase with the Central American region to the north, which in turn has out-of-phase variations with Western North America, again suggestive of latitudinal changes in monsoon features. East and Southeast Asia show hardly any long-term changes, with both having plentiful rains in the 1950s. On the interannual timescales there are a number of surprising correlations: Amazonia is correlated with Northern Australia (0.44, significant at <1%) and also Southeast Asia (0.55, <1%), while Southern South America is inversely correlated with Western Africa (–0.51, <1%). They are surprising because they are based on high-frequency relationships and the correlations barely change when the smoothed series are used.

3.3.2.3 Changes in snowfall

Winter precipitation has increased in high latitudes, although uncertainties exist because of changes in undercatch, especially as snow changes to rain. Annual precipitation for the circumpolar region north of 50°N has increased during the past 50 years (not shown) by approximately 4% but this increase has not been homogeneous in time and space (Groisman et al., 2003, 2005). Statistically significant increases were documented over Fennoscandia, coastal regions of northern North America (Groisman et al., 2005), most of Canada (particularly, northern regions of the country) up till 1995 (Stone et al., 2000), the permafrost-free zone of Russia (Groisman and Rankova, 2001), and the entire Great Russian Plain (Groisman et al., 2005, 2006). However, there were no discernible changes in summer and annual precipitation totals over northern Eurasia, east of the Ural Mountains (Gruza et al., 1999; Sun and Groisman, 2000; Groisman et al., 2005, 2006). The rainfall (liquid precipitation) has increased during the past 50 years over western portions of North America and Eurasia north of 50°N by ~6%. Rising temperatures have generally resulted in rain rather than snow in locations and seasons where climatological-average (1961–1990) temperatures were close to 0°C. The liquid-precipitation season has become longer by up to 3 weeks in some regions of the boreal high latitudes over the last 50 years (Groisman et al., 2001; Cayan et al., 2001; Easterling, 2002; Groisman et al., 2005,2006) owing, in particular, to an earlier onset of spring. So in some regions (southern Canada and western Russia), snow has provided a declining fraction of total annual precipitation (Groisman et al., 2003, 2005, 2006). In other regions, in particular north of 55°N, the fraction of annual precipitation falling as snow in winter has changed little.

Berger et al. (2002) found a trend toward fewer snowfall events during winter across the lower Missouri river basin from 1948 to 2002, but little or no trend in snowfall occurrences within the plains region to the south. In New England, there has been a decrease in the proportion of precipitation occurring as snow at many locations, caused predominantly by a decrease in snowfall, with a lesser contribution from increased rainfall (Huntington et al., 2004). By contrast, Burnett et al. (2003) have found large increases in lake-effect snowfall since 1951 for locations near the North American Great Lakes, consistent with the observed decrease in ice cover for most of the Great Lakes since the early 1980s (Assell et al., 2003). In addition to snow data, they used lake sediment reconstructions for locations south of Lake Ontario to indicate that these increases have been ongoing since the turn of the 20th century. Ellis and Johnson (2004) found that the increases in snowfall across the regions to the lee of Lakes Erie and Ontario are due to increases in the frequency of snowfall at the expense of rainfall events, an increase in the intensity of snowfall events, and to a lesser extent an increase in the water equivalent of the snow. In Canada, the frequency of heavy snowfall events has decreased since the 1970s in the south and increased in the north (Zhang et al., 2001a).

3.3.2.4 Urban areas

As noted in Section 3.2.2.2 and see Chapter 7, Box 7.2, the micro-climates in cities are clearly different than in neighbouring rural areas. The presence of a city affects runoff, moisture availability and precipitation. Crutzen (2004) points out that while human energy production is relatively small globally compared with the sun, it is not locally in cities, where it can reach 20 to 70 W m^{-2}. Results from the Metropolitan Meteorological Experiment (METROMEX) in the United States in the 1970s showed that urban effects lead to increased precipitation during the summer months within and 50–75 km downwind of the city, reflecting increases of 5–25% over background values (Changnon et al., 1981). Balling and Brazel (1987) observed more frequent late afternoon storms in Phoenix during recent years of explosive population growth. Jauregui and Romales (1996) observed that the daytime heat island seemed to be correlated with intensification of rain showers during the wet season (May–October) in Mexico City and that the frequency of intense rain showers has increased in recent decades at a similar rate as the growth of the city. More recent observational studies (Bornstein and Lin, 2000; Shepherd et al., 2002; Changnon and Westcott, 2002; Diem and Brown, 2003; Shepherd and Burian, 2003; Fujibe, 2003; Dixon and Mote, 2003; Burian and Shepherd, 2005; Inoue and Kimura, 2004; Shepherd et al., 2004) have continued to link urban-induced dynamic processes to precipitation anomalies. Nor is it confined to urban areas (see Chapter 7, Section 7.2). Other changes in land use can also affect precipitation, and a notable example is in the Amazon arising from deforestation, where Chagnon and Bras (2005) find large changes in local rainfall, with increases in deforested areas, associated with local atmospheric circulations that are changed by gradients in vegetation. Changes are also found in seasonality.

Suggested mechanisms for urban-induced rainfall include one or a combination of the following: (1) enhanced convergence due to increased surface roughness in the urban environment (e.g., Changnon et al., 1981; Bornstein and Lin, 2000; Thielen et al., 2000); (2) destabilization due to urban heat island (UHI)-thermal perturbation of the boundary layer and resulting downstream translation of the UHI circulation or UHI-generated convective clouds (e.g., Shepherd et al., 2002; Shepherd and Burian, 2003); (3) enhanced aerosols in the urban environment for cloud condensation nuclei sources (e.g., Diem and Brown, 2003;

Molders and Olson, 2004); or (4) bifurcating or diverting of precipitating systems by the urban canopy or related processes (e.g., Bornstein and Lin, 2000; Loose and Bornstein, 1977). The "weekend effect" noted in Section 3.2.2.2 likely arises from these mechanisms. The diurnal cycle in precipitation, which varies over the United States from late afternoon maxima in the Southeast, to nocturnal maxima in the Great Plains (Dai and Trenberth, 2004), may be modified in some regions by urban environments. Dixon and Mote (2003) found that a growing urban heat island effect in Atlanta, Georgia (United States) enhanced and possibly initiated thunderstorms, especially in July (summer) just after midnight. Low-level moisture was found to be a key factor.

3.3.2.5 Ocean precipitation

Remotely-sensed precipitation measurements over the ocean are based on several different sensors in the microwave and infrared that are combined in different ways. Many experimental products exist. Operational merged products seem to perform best in replicating island-observed monthly amounts (Adler et al., 2001). This does not mean they are best for trends or low-frequency variability, because of the changing mixes of input data. The main global datasets available for precipitation, and which therefore include ocean coverage, have been the GPCP (Huffman et al., 1997; Adler et al., 2003) and NOAA Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) (Xie and Arkin, 1997). Comparisons of these datasets and others (Adler et al., 2001; Yin et al., 2004) reveal large discrepancies over the ocean; however there is better agreement among the passive microwave products even using different algorithms. Over the tropical oceans, mean amounts in CMAP and GPCP differ by 10 to 15%. Calibration using observed rainfall from small atolls in CMAP was extended throughout the tropics in ways that are now recognized as incorrect. However, evaluation of GPCP reveals that it is biased low by 16% at such atolls (Adler et al., 2003), also raising questions about the ocean GPCP values. Differences arise due to sampling and algorithms. Polar-orbiting satellites each obtain only 2 instantaneous rates per day over any given location, and thus suffer from temporal sampling that is offset by using geostationary satellites. However, only less accurate infrared sensors are available with the latter. Model-based (including reanalysis) products perform poorly in the evaluation of Adler et al. (2001) and are not currently suitable for climate monitoring. Robertson et al. (2001b) examined monthly anomalies from several satellite-derived precipitation datasets (using different algorithms) over the tropical oceans. The expectation in the TAR was that measurements from TRMM radar (PR) and passive microwave imager (TMI) would clarify the reasons for the discrepancies, but this has not yet been the case. Robertson et al. (2003) documented poorly correlated behaviour (0.12) between the monthly, tropical ocean-averaged precipitation anomalies from the PR and TMI sensors. Although the TRMM PR responds directly to precipitation size hydrometeors, it operates with a single attenuating frequency (13.8 GHz) that necessitates significant microphysical assumptions regarding drop-size distributions for relating reflectivity, signal attenuation, and rainfall, and uncertainties in microphysical assumptions for the primary TRMM algorithm (2A25) remain problematic.

The large regional signals from monsoons and ENSO that emphasize large-scale shifts in precipitation are reasonably well captured in GPCP and CMAP, see Section 3.6.2, but cancel out when area averaged over the tropics, and the trends and variability of the tropical average are quite different in the two products. At present, therefore, documenting the amplitudes of interannual variations in precipitation over the oceans as a whole remains a challenge. A 25-year plot of monthly global precipitation from GPCP (updated from Adler et al., 2003, but not shown) indicates monthly variability with a standard deviation of about 2% of the mean. The variability in the ocean and land areas when examined separately is larger, about 3%. There are also variations during the 25 years related to ENSO events (Curtis and Adler, 2003). Precipitation averaged over the ocean increases with El Niño but decreases over land.

Although the trend over 25 years in global total precipitation in the GPCP dataset (Adler et al., 2003) is very small (see Table 3.4), there is a small increase (about 4% over the 25 years) over the latitude range 25°S–25°N (over the ocean), with a partially compensating decrease over land (2%) in the same latitude belt. Northern mid-latitudes show a decrease over land and ocean. Over a slightly longer timeframe, precipitation increased over the North Atlantic between 1960–1974 and 1975–1989 (Josey and Marsh, 2005). The inhomogeneous nature of the datasets and the large ENSO variability limit what can be said about the validity of changes, both globally and regionally.

3.3.3 Evapotranspiration

There are very limited direct measurements of actual evapotranspiration over global land areas. Over oceans, estimates of evaporation depend on bulk flux estimates that contain large errors. Evaporation fields from the ERA-40 and NRA are not considered reliable because they are not well constrained by precipitation and radiation (Betts et al., 2003; Ruiz-Barradas and Nigam, 2005). The physical processes related to changes in evapotranspiration are discussed in Chapter 7, Section 7.2 and Box 3.2.

Decreasing trends during recent decades are found in sparse records of pan evaporation (measured evaporation from an open water surface in a pan) over the United States (Peterson et al., 1995; Golubev et al., 2001; Hobbins et al., 2004), India (Chattopadhyay and Hulme, 1997), Australia (Roderick and Farquhar, 2004), New Zealand (Roderick and Farquhar, 2005), China (Liu et al., 2004a; Y. Qian et al, 2006) and Thailand (Tebakari, et al., 2005). Pan measurements do not represent actual evaporation (Brutsaert and Parlange, 1998); any trends being more likely caused by decreasing surface solar radiation over the United States, parts of Europe and Russia (Abakumova et al., 1996; Liepert, 2002) and decreased sunshine duration over China (Kaiser and Qian, 2002) that may be related to increases in air pollution and atmospheric aerosols (Liepert et al., 2004; T. Qian et al; 2006) and increases in cloud cover (Dai et al., 1999). Whether actual evapotranspiration decreases or not also depends on how surface wetness changes, see Box 3.2. Changes in evapotranspiration are often calculated using empirical models, as a function of precipitation and surface net radiation (Milly and Dunne, 2001), or land surface models (e.g., van den Dool et al., 2003; T. Qian et al., 2006).

The TAR reported that actual evapotranspiration increased during the second half of the 20th century over most dry regions of the United States and Russia (Golubev et al., 2001), resulting from greater availability of surface moisture due to increased precipitation and larger atmospheric moisture demand due to higher temperature. One outcome is a larger surface latent heat flux (increased evapotranspiration) but decreased sensible heat flux (Trenberth and Shea, 2005). Using observed precipitation, temperature, cloudiness-based surface solar radiation and a comprehensive land surface model, T. Qian et al. (2006) found that global land evapotranspiration closely follows variations in land precipitation. It peaked in the early 1970s and then decreased steadily until the early 1990s when it started increasing again. Changes in evapotranspiration depend not only on moisture supply but also energy availability, as summarized in Box 3.2.

3.3.4 Changes in Soil Moisture, Drought, Runoff and River Discharge

Historical records of *in situ* measured soil moisture content are available only for a few regions and often are very short (Robock et al., 2000). A rare 45-year record of soil moisture over agricultural areas of the Ukraine shows a large upward trend, which was stronger during the first half of the period (Robock et al., 2005). Among over 600 stations from a large variety of climates, including the former Soviet Union, China, Mongolia, India, and the United States, Robock et al. (2000) showed an increasing long-term trend in surface soil moisture (top 1 m) content during summer for the stations with the longest records.

Since the *in-situ* observational record and global estimates of remotely sensed soil moisture data are limited, global soil moisture variation during the 20th century has been estimated by an offline simulation of a land surface model (LSM) (e.g. Hirabayashi et al., 2005). Observed precipitation and maximum and minimum temperature were used to drive the LSM for 1901 through 2000 to estimate the energy and water balance over global land areas at 1° × 1° resolution. LSM estimates agree well with corresponding *in-situ* observations in their inter-annual variation of surface soil moisture during summer. Decadal variations dominate in the 100-year LSM estimates of soil moisture during the 20th century, and there is no statistically significant long-term trend. However, it is too soon to conclude there was no long-term trend, as the accuracy of LSM estimates depends on the forcing data, notably the number of raingauge stations (Oki et al., 1999). The raingauge data used (New et al., 1999) were very limited in some regions in the early 20th century. There are a number of other LSM-simulated soil moisture content data (e.g., Dirmeyer et al., 1999; Maurer et al., 2002; van den Dool et al., 2003; Berg et al., 2003; Fan and van den Dool, 2004; Mitchell et al. 2004; Ngo-Duc et al., 2005; T. Qian et al., 2006) that have also been used to study changes and variations in surface moisture conditions, but results are model dependent as the LSMs differ between studies.

Most studies of long-term changes in soil moisture use calculations based on formulae or models. The primary approach has been to calculate Palmer Drought Severity Index (PDSI), see Box 3.1, values from observed precipitation and temperature (e.g., Dai et al., 2004b). In some locations much longer proxy-extensions have been derived from earlier tree-ring data (see Chapter 6, Section 6.5.6; e.g., Cook et al., 1999). The longer instrumental-based PDSI estimations are used to look at trends and some recent extreme PDSI events in different regions are placed in a longer-term context (see specific cases in Box 3.6). As with LSM-based studies the version of the PDSI used is crucial, and it can partly determine some aspects of the results found (Box 3.1).

Using the PDSI, Dai et al. (2004b) found a large drying trend over NH land since the middle 1950s, with widespread drying over much of Eurasia, northern Africa, Canada and Alaska. In the SH, land surfaces were wet in the 1970s and relatively dry in the 1960s and 1990s; and there was a drying trend from 1974 to 1998 although trends over the entire 1948–2002 period were small. Overall patterns of trends in PDSI are given in Question 3.2, Figure 1. Although the long-term (1901–2004) land-based precipitation trend shows a small increase (Figure 3.3.1), decreases in land precipitation in recent decades are the main cause for the drying trends, although large surface warming during the last 2–3 decades has likely contributed to the drying. Dai et al. (2004b) show that globally very dry areas, defined as land areas with the PDSI less than –3.0, more than doubled (from ~12% to 30%) since the 1970s, with a large jump in the early 1980s due to an ENSO-induced precipitation decrease over land and subsequent increases primarily due to surface warming. However, results are dependent on the version of the PDSI model used, since the empirical constants used in a global PDSI model may not be adequately adjusted for the local climate (see Box 3.1).

In Canada, the summer PDSI averaged for the entire country indicates dry conditions during the 1940s and 1950s, generally wet conditions from the 1960s to 1995, but much drier after 1995 (Shabbar and Skinner, 2004). They also show a relationship between summer droughts in Canada with global SSTs suggesting that the warming trend in SST is resulting in more summer drought in Canada. Groisman et al. (2006) found increased dryness based on the Keech-Byram forest-fire drought index in northern Eurasia, a finding supported by Dai et al. (2004b) using the PDSI. Long European records (van der Schrier et al., 2006) reveal no trend in areas affected by extreme PDSI values (either thresholds of ±2 or ±4) over the 20th century. Nevertheless, recently Europe has suffered prolonged drought, including the 2003 episode associated with the severe summer heat wave (see Box 3.6.5).

Although there was no significant trend over 1880–1998 during summer (JJA) in eastern China, precipitation for 1990–1998 was the most on record (Gong and Wang, 2000). Zou et al., (2005) found that for China as a whole there were no long-term trends in the percentage areas of droughts (defined as PDSI <–1.0) during 1951–2003. However, increases of drought areas were found in much of northern China (but not in northwest China, Zou et al., 2005), aggravated by warming and decreasing precipitation (Wang and Zhai, 2003; Ma and Fu, 2003), consistent with Dai et al. (2004b).

A severe drought affecting central and southwest Asia in recent years (see Box 3.6.1) appears to be the worst since at least 1980 (Barlow et al., 2002). In the Sahel region of Africa, rainfall has recovered somewhat in recent years, after large decreasing rainfall trends from the late 1960s to the late 1980s (Dai et al., 2004a; see also Section 3.3.2.2); see Figure 3.7.4. Large multi-year oscillations appear to be more frequent and extreme after the late 1960s than previously in the century. A severe drought affected Australia in 2002–2003; precipitation deficits were not as severe as during a few episodes earlier in the 20th century, but higher temperatures exacerbated the impacts (see Box 3.6.2). Severe drought, stemming from at least three years of rainfall deficits, continued into 2005 especially in the eastern third of Australia, although rains brought some relief in June 2005.

A multi-decadal period of relative wetness characterized the latter portion of the 20th century in the continental United States, both in terms of precipitation (Mauget, 2003a), streamflow (Groisman et al., 2004) and annual moisture surplus (precipitation minus potential evapotranspiration) (McCabe and Wolock, 2002). Despite this overall national trend towards wetter conditions, a severe drought affected the western United States from 1999 to November 2004 (see Box 3.6.3).

Available streamflow gauge records cover only about two-thirds of the global actively-drained land areas and they often have gaps and vary in record length (Dai and Trenberth, 2002). Estimates of total continental river discharge are therefore often based on incomplete gauge records (e.g., Probst and Tardy, 1987, 1989; Guetter and Georgakakos, 1993), reconstructed streamflow time series (Labat et al., 2004), or methods to account for the runoff contribution from the unmonitored areas (Dai and Trenberth, 2002). These estimates show large decadal to multi-decadal variations in continental and global freshwater discharge (excluding groundwater) (Guetter and Georgakakos, 1993; Labat et al., 2004).

Streamflow records for the world's major rivers show large decadal to multi-decadal variations, with small secular trends for most rivers (Cluis and Laberge, 2001; Lammers et al., 2001; Pekárová et al., 2003; Mauget, 2003b; Dai et al., 2004b). Increased streamflow during the later half of the 20th century has been reported over regions with increased precipitation, such as many parts of the United States (Lins and Slack, 1999; Groisman et al., 2004) and southeastern South America (Genta et al., 1998). Decreased streamflow was reported over many Canadian river basins during the last 30–50 years (Zhang et al., 2001b) where precipitation has also decreased during the period. Déry and Wood (2005) also found decreases in river discharge into the Arctic and North Atlantic from high latitude Canadian rivers with potential implications for salinity levels in these oceans and possibly the North Atlantic thermohaline circulation. These changes are consistent with observed decreases in precipitation in high latitude Canada from 1963 to 2000. Further, Milly et al. (2002) show significant trends towards more extreme flood events from streamflow measurements on 29 very large basins. The global increase in both severe drought and large floods suggests that hydrologic conditions have become more extreme. Recent extreme flood events in central Europe (on the Elbe and some adjacent catchments) are discussed in Box 3.6.4.

Because large dams and reservoirs have been built along many of world's major rivers, increasing low flow and reducing peak flow, dramatic changes in seasonal flow rates (Cowell and Stoudt, 2002; Ye et al., 2003; Yang et al., 2004) and trends in seasonal streamflow rates (e.g., Lammers et al., 2001) should be interpreted cautiously. Nevertheless, there is evidence that the rapid warming since the 1970s has induced earlier snowmelt and associated peak streamflow in the western United States (Cayan et al., 2001) and New England, USA (Hodgkins et al., 2003) and earlier breakup of river-ice in Russian Arctic rivers (Smith, 2000) and many Canadian rivers (Zhang et al., 2001b).

River discharges in the La Plata River basin in southeastern South America exhibit large interannual climate variability. Consistent evidence linking the Paraná and Uruguay streamflows and ENSO has been found (Bischoff et al., 2000; Berri et al., 2002; Camilloni and Barros, 2000, 2003; Robertson et al., 2001a; Krepper et al., 2003) indicating that monthly and extreme flows during El Niño are generally larger than those observed during La Niña events. For the Paraguay River, most of the major discharges at the Pantanal wetland outlet occurred in the neutral phases of ENSO, but in the lower reaches of the river the major discharge events occurred during El Niño events (Barros et al., 2004). South Atlantic SST anomalies also modulate regional river discharges through effects on rainfall in southeastern South America (Camilloni and Barros, 2000). The Paraná River shows a positive trend in its annual mean discharge since the 1970s in accordance with the regional rainfall trends (García and Vargas, 1998; Barros et al., 2000; Liebmann et al., 2004), as do the Paraguay and Uruguay Rivers since 1970.

Yang et al. (2002) used monthly records of temperature, precipitation, streamflow, ice thickness, and active layer depth for the 1935–1999 period in the Lena River basin in Siberia and found significant increases in temperature leading to increases in streamflow and decreases in ice thickness during the cold season. Strong springtime warming has resulted in an earlier snowmelt with a reduced maximum streamflow pulse in June. During the warm season, smaller streamflow increases are related to an observed increase in precipitation. Streamflow during the latter half of the 20th century for the Yellow River basin in China decreased significantly, even after accounting for increased human consumption (Yu et al., 2004). Temperatures have increased over the basin, but precipitation has shown no change, suggesting an increase in evaporation.

In Africa for 1950–1995, Jury (2003) found that the Niger and Senegal rivers show the effects of the Sahel drying trend with a decreasing trend in flow. The Zambezi also exhibits reduced flows, but rainfall over its catchment area appears to be stationary. Other major African rivers, including the Blue and White Nile, Congo, and inflow into Lake Malawi show high variability, consistent with interannual variability of SSTs in the Atlantic, Indian, and Pacific oceans. A composite index of riverflow for these rivers shows the five highest flow years occurred prior to 1979, and the five lowest flow years occurred after 1971.

In summary, the PDSI shows a large drying trend since the mid-1950s over NH land areas with widespread drying over much of Eurasia, Africa, Canada and Alaska. In the SH, there was a drying trend from 1974 to 1998 although trends over the entire 1948–2002 period are small. Seasonal decreases in land precipitation since the 1950s are the main cause for some of the drying trends, although large surface warming during the last 2–3 decades has also likely contributed to the drying. Based on the PDSI data, globally very dry areas, defined as land areas with the PDSI less than –3.0, have more than doubled since the 1970s, with a large jump in the early 1980s due to an ENSO-induced precipitation decrease over land and subsequent increases primarily due to surface warming. River discharges in many tropical areas of Africa and South Americas are strongly affected by the ENSO phase, with greater discharges on the Parana River after the climate shift in 1976/1977, while some major African rivers have been lower since this time.

Box 3.1: Drought Terminology and Determination

In general terms, drought is a “prolonged absence or marked deficiency of precipitation”, a “deficiency of precipitation that results in water shortage for some activity or for some group,” or a “period of abnormally dry weather sufficiently prolonged for the lack of precipitation to cause a serious hydrological imbalance” (Heim, 2002). Drought has been defined in a number of ways. *Agricultural drought* relates to moisture deficits in the topmost 1 metre or so of soil (the root zone) that impacts crops, *meteorological drought* is mainly a prolonged deficit of precipitation, and *hydrologic drought* is related to below normal streamflow, lake and groundwater levels.

Drought and its severity can be numerically defined using indices that integrate temperature, precipitation and other variables that impact evapotranspiration and soil moisture. The most commonly used index is the PDSI (Palmer 1965, Heim 2002). Although PDSI is not an optimal index, since it does not normally include variables such as wind speed, solar radiation, cloudiness, and water vapour, it is widely used and can be calculated across many climates as it requires only precipitation and temperature data for the calculation of potential evapotranspiration (PET) using Thornthwaite’s (1948) method. Because these data are readily available for most parts of the globe, the PDSI provides a consistent measure of drought for comparison across many regions.

However, PET is considered to be more reliably calculated using Penman (1948) type approaches that incorporate the effects of wind, water vapour, and solar and longwave radiation. There has been criticism of most Thornthwaite-based estimates of PDSI because the empirical constants have not been re-computed for each climate (Alley, 1984). Hence a self-calibrating version of the PDSI has recently been developed to ensure consistency with the climate at any location (Wells et al., 2004). Differences between Thornthwaite-based and Penman-based PET will be ameliorated for by the use of the self-calibrating PDSI. Also, studies that compute changes or trends in PDSI effectively remove influences of biases in the absolute values. As the effects of temperature anomalies on the PSDI are small compared to precipitation anomalies (Guttman, 1991), PDSI is largely controlled by precipitation changes.

3.3.5 Consistency and Relationships between Temperature and Precipitation

Observed changes in temperature and precipitation should provide a physically-consistent picture and here we assess basic temperature-precipitation relationships and trend consistencies. Significant large-scale correlations between observed monthly mean temperature and precipitation (Madden and Williams, 1978) for North America and Europe have stood up to the test of time and been expanded globally (Trenberth and Shea, 2005). In the warm season over continents, higher temperatures accompany lower precipitation amounts and vice versa. Hence, over land, strong negative correlations dominate, as dry conditions favour more sunshine and less evaporative cooling, while wet summers are cool. However, at latitudes polewards of 40° in winter, positive correlations dominate as the water-holding capacity of the atmosphere limits precipitation amounts in cold conditions and warm air advection in cyclonic storms is accompanied by precipitation. Where ocean conditions drive the atmosphere, higher surface air temperatures are associated with precipitation, as in El Niño. For South America, Rusticucci and Penalba (2000) show that warm summers are associated with low precipitation, especially in northeast and central-western Argentina, southern Chile, and Paraguay. Cold season (JJA) correlations are weak but positive to the west of 65°W, as stratiform cloud cover produces a higher minimum temperature. For stations in coastal Chile, the correlation is always positive and significant, as it is over the ocean, especially in the months of rainfall (May to September), showing that high SSTs favour convection.

This relationship of higher warm-season temperatures going with lower precipitation appears to apply also to trends (Trenberth and Shea, 2005). An example is Australia which exhibits evidence of increased drought severity, consistent with the observed warming during the latter half of the 20th century (Nicholls, 2004). Mean maximum and minimum temperatures during the 2002 Australian drought were much higher than during the previous droughts in 1982 and 1994, suggesting enhanced potential evaporation as well; see Box 3.6.2. Record high maximum temperatures also accompanied the dry conditions in 2005.

3.4 Changes in the Free Atmosphere

3.4.1 Temperature of the Upper Air: Troposphere and Stratosphere

Within the community that constructs and actively analyses satellite and the radiosonde-based temperature records there is agreement that the uncertainties in long-term change are substantial. Changes in instrumentation and protocols pervade both sonde and satellite records, obfuscating the modest long-term trends. Historically there is no reference network to anchor the true record and establish uncertainty in the effects of these changes – many of which are both barely documented and poorly understood. Therefore, investigators have to make seemingly reasonable choices of how to handle these sometimes known but often unknown influences. It is difficult to make quantitatively defensible judgments as to which, if any, of the multiple, independently derived, estimates is closer to the true climate evolution. This reflects almost entirely upon the inadequacies of the historical observing network and points to the need for future network design that provides the reference sonde-based ground-truth. A comprehensive review of this whole issue is given by CCSP (2006).

3.4.1.1 Radiosondes

Since the TAR considerable effort has been devoted to assessing and improving the quality of the radiosonde temperature record (see Appendix 3.B.5.1). A particular aim has been to reduce artificial changes arising from instrumental and procedural developments during the seven decades (1940s-2000s) of the radiosonde record (Free and Seidel, 2005; Thorne et al., 2005b; CCSP, 2006). However, a range of approaches yielded disparate results when applied to the identification of spurious jumps in a common set of temperature measurements at radiosonde sites (Free et al., 2002). An approach based on the physics of heat transfer within radiosondes performed poorly when evaluated against independently measured temperatures from satellite records (Durre et al., 2002).

The LKS (Lanzante et al., 2003a,b) dataset has subjectively derived bias adjustments throughout the length of its record. LKS is restricted to a select global network of 87 stations and terminates in 1997. The LKS dataset has been updated using the Integrated Global Radiosonde Archive (IGRA), Durre et al. 2006) by applying a different bias adjustment technique (Free et al., 2004b) to the data after 1997 to create a new archive (Radiosonde Atmospheric Temperature Products for Assessing Climate, RATPAC) with improved quality control (Free et al., 2005). However, Angell (2003) found problems with some tropical stations and Randel and Wu (2006) used collocated MSU data to show that cooling biases remain in some of the LKS/RATPAC radiosonde data for the tropical stratosphere and upper troposphere due to changes in instruments and radiation correction adjustments.

Data adjusted using satellite data (HadRT, Parker et al., 1997) have been used extensively for climate studies, but the adjustments reduce independence and can only be applied 1979 to present. In addition, the HadRT adjustments are conditional on local metadata, so spatial consistency is lost. Accordingly, a new radiosonde record, HadAT2 (Thorne et al., 2005a), has recently been constructed. HadAT2 uses a neighbour comparison approach to build spatial as well as temporal consistency throughout the record (see Appendix 3.B.5.1).

A comprehensive intercomparison (Seidel et al., 2004) showed that 5 radiosonde datasets yielded consistent signals for higher frequency events such as ENSO, QBO and volcanic eruptions. However, for long-term trends differences among datasets were apparent. So these authors recommended that multiple independent datasets be used in the assessment of longer-term change.

A new development in approaches to improve radiosonde data has used the bias-adjustments estimated during data assimilation into model-based reanalyses (Haimberger, 2005). Despite the risk of contamination by other biased data in the assimilation, or by model biases, the adjustments are found to agree with those estimated by existing methods. In another major new development, Sherwood et al. (2005) have found substantial changes in the diurnal cycle in radiosonde data. These changes are probably a consequence of improved sensors and radiation error adjustments. Relative to night-time values, they found a daytime warming of sonde temperatures prior to 1971 that is likely spurious and then a spurious daytime cooling from 1979 to 1997. They estimated that there was likely a spurious overall downward trend in sonde temperature records during the satellite era (since 1978) throughout the atmosphere of order 0.1°C decade^{-1} globally: the assessed spurious cooling is greatest in the tropics at 0.16°C decade^{-1} for the 850 to 300 hPa layer, and least in the NH extratropics of 0.04°C decade^{-1}. Night-time values are also affected by inhomogeneities, and Randel and Wu (2006) identify problems in many tropical radiosonde records such that in general they likely have a spurious negative bias.

The radiosonde dataset is limited to land areas, and coverage is poor over the tropics and SH. Accordingly, when global estimates based solely on radiosondes are presented, there are considerable uncertainties (Hurrell et al., 2000; Agudelo and Curry, 2004) and denser networks – which perforce still omit oceanic areas – may not yield more reliable “global” trends (Free and Seidel, 2005). Radiosonde records have an advantage of starting in the 1940s regionally and near-globally from about 1958. They monitor the troposphere and lower stratosphere; layers analysed are described below and in Figure 3.4.1. Radiosonde-based global mean temperature estimates are given in Figure 3.4.2.

[INSERT FIGURE 3.4.1 HERE]

3.4.1.2 The satellite MSU record

3.4.1.2.1 Summary of satellite capabilities and challenges

Satellite-borne microwave sounders estimate the temperature of thick layers of the atmosphere by measuring microwave emissions (radiances) that are proportional to the thermal state of emission of oxygen molecules from a complex of emission lines near 60 GHz. By making measurements at different frequencies near 60 GHz, different atmospheric layers can be sampled. A series of 9 instruments called microwave sounding units (MSUs) began making this kind of measurement in late 1978. Beginning in mid-1998, a follow-on series of instruments, the Advanced MSUs (AMSUs) began operation. Unlike infrared sounders, microwave sounders are not affected by most clouds, although some effects are experienced from precipitation and clouds with high liquid water content. Illustrated in Figure 3.4.1 are the lower troposphere (referred to as $T2_{LT}$), troposphere, and MSU channel 2 (referred to as T2) and channel 4 (lower stratosphere, referred to as T4) layers.

The main advantage of satellite measurements, compared to radiosondes, is the excellent global coverage of the measurements, with complete global coverage every few days. But like radiosondes, temporal continuity is a major challenge for climate assessment, as data from all the satellites in the series must be merged together. The merging procedure must accurately account for a number of error sources. The most important are: (1) offsets in calibration between satellites; (2) orbital decay and associated long-term changes in the time of day that the measurements are made at a particular location, which combine with the diurnal cycle in atmospheric temperature to produce diurnal drifts in the estimated temperatures; (3) drifts in satellite calibration that are correlated with the temperature of the on-board calibration target. Since the calibration target temperatures vary with the satellite diurnal drift, the satellite calibration and diurnal drift corrections are intricately coupled together (Fu and Johanson 2005). Independent teams of investigators have used different methods to determine and correct for these “structural” and other sources of error (Thorne et al., 2005b). Appendix 3.B.5.5 discusses adjustments to the data in more detail.

[INSERT FIGURE 3.4.2 HERE]

3.4.1.2.2 Progress since the TAR

Since the TAR, several important developments and advances have occurred in the analysis of satellite measurement for atmospheric temperatures. Scrutiny of existing datasets and identification of problems, leading to new versions, are described below. A number of new data records have been constructed from the MSU measurements, as well as from global reanalyses (see Section 3.4.1.3). Further, new insights have come from statistical combinations of the MSU records from different channels that have minimised the influence of the stratosphere on the tropospheric records (Fu et al., 2004a,b; Fu and Johanson, 2004, 2005).

These new datasets and analyses are very important because the differences highlight assumptions and it becomes possible to estimate the uncertainty in satellite-derived temperature trends that arises from different methods and approaches to the construction of temporally-consistent records.

Three main analyses of MSU channels 2 and 4 have been conducted by the University of Alabama, Huntsville (UAH) (Christy et al., 2000, 2003), Remote Sensing Systems (RSS: Mears et al., 2003; Mears and Wentz, 2005) and by Vinnikov and Grody (2003) version 1 (VG1), now superseded by version 2 (VG2: Grody et al., 2004; Vinnikov et al., 2006). MSU channel 2 (T2) measures a thick layer of the atmosphere, with approximately 75–80% of the signal coming from the troposphere and surface, 15% from the lower stratosphere, and the remaining 5–10% from the surface. MSU channel 4 (T4) is primarily sensitive to temperature in the lower stratosphere (Figure 3.4.1).

Global time series from each of the MSU records are shown in Figure 3.4.2 and global trends calculated are depicted in Figure 3.4.3. These show a global cooling of the stratosphere (T4) of –0.32 to –0.47 °C decade^{-1} and a global warming of the troposphere from T2 of 0.04 to 0.20°C decade^{-1} for the period 19792004 for the MSU records. The large spread in T2 trends stems from differences in the inter-satellite calibration and merging technique, the orbital drift and diurnal-cycle change corrections and the hot point calibration temperature corrections (Christy et al., 2003; Mears et al., 2003; Mears and Wentz, 2005; Grody et al., 2004; Christy and Norris, 2004; Fu and Johanson, 2005; Vinnikov et al., 2006; see also Appendix 3.B.5.5)

The RSS results for T2 indicate about a 0.1°C decade^{-1} more warming in the troposphere than UAH (see Figure 3.4.3) and most of the difference arises from the use of different amounts of data to determine the parameters of the calibration target effect. The UAH group used only satellite pairs with periods of simultaneous observation longer than one year and focused on reducing low-frequency differences, while the RSS group used all satellite pairs with simultaneous observations and minimized differences. Statistically the latter has many more degrees of freedom, whereas the former reduces relative trends between satellites, but further yields parameters for NOAA-9 (1985–1987) outside of the physical bounds expected by Mears et al. (2003). Hence the large difference in the calibration parameters for the single instrument mounted on the NOAA-9 satellite accounted for a substantial part of the trend difference between the UAH and RSS T2 results. The rest arises from differences in merging parameters for other satellites, differences in the correction for the drift in measurement time (Mears et al., 2003; Christy and Norris, 2004), and ways the hot point temperature is corrected for (Grody et al., 2004; Fu and Johanson, 2005). In the tropics, these accounted for differences in T2 of about 0.07°C decade^{-1} in trend after 1987 and discontinuities were also present in 1992 and 1995 at times of satellite transitions (Fu and Johanson, 2005).

The new T2 data record of Grody et al. (2004) and Vinnikov et al. (2006) (VG2) shows slightly more warming in the troposphere than the RSS data record (Figure 3.4.3). VG2 created a latitude-dependent analysis that allows for errors that depend on both the calibration target temperature and the atmospheric temperature being measured, although because temporal averaging is used to reduce noise in overlapping satellite measurements, issues remain in accounting for temporal variations in calibration target temperatures on individual satellites. The need to account for the target effect as a function of latitude, which was not done by UAH or RSS, is related to the diurnal cycle correction. The VG2 method does not, however, fully address the correction for diurnal drift before merging and does not produce maps, so that differences between land and ocean remain to be evaluated.

Although the T4 from RSS has about 0.1°C decade^{-1} less cooling than the UAH product (Figure 3.4.3), both datasets support the conclusions that the stratosphere has undergone strong cooling since 1979. Because about 15% of the signal for T2 comes from the lower stratosphere, the observed cooling causes the reported T2 trends to be underestimates of tropospheric warming. By creating a weighted combination of T2 and T4, this effect has been greatly reduced (Fu et al., 2004a) (see Figure 3.4.1 for troposphere-UW (for University of Washington)). This technique for the global mean temperature implies small negative weights at some stratospheric levels, but because of vertical coherence these merely compensate for other positive weights nearby and it is the integral that matters (Fu and Johanson, 2004). From 1979 to 2001 the stratospheric contribution to the trend of T2 is about –0.08°C decade^{-1}. Questions about this technique (Tett and Thorne, 2004) have led to further clearer interpretation of its application to the tropics (Fu et al., 2004b). The technique has also been successfully applied to model results (Gillett et al., 2004; Kiehl et al., 2005), although model biases in depicting stratospheric cooling can affect results. In a further development, weighted combinations of T2, T3 (from channel 3) and T4 since 1987 have formed tropical series for the upper, lower and whole troposphere (Fu and Johanson, 2005).

By differencing T2 measurements made at different slant angles, the UAH group produced an updated data record weighted for the lower and mid troposphere, $T2_{LT}$ (Christy et al., 2003). This retrieval also has the effect of removing the stratospheric influence on long-term trends but its uncertainties are augmented by the need to compensate for orbital decay and by computing a small residual from two large values (Wentz and Schabel, 1998). $T2_{LT}$ retrievals include a large signal from the surface and so are adversely affected by changes in surface emissivity, including changes in sea ice cover (Swanson, 2003). Fu and Johanson (2005) found that the $T2_{LT}$ trends were physically inconsistent compared with those of the surface, T2, and T4, even if taken from the UAH record and they showed that the large trend bias is largely attributed to the periods when a satellite had large local equator crossing time drifts that cause large changes in calibration target temperatures and large diurnal drifts. Mears and Wentz (2005) further found that the adjustments for diurnal cycle corrections required from satellite drift had the wrong sign in the UAH record. Corrections have now been made (version 5.2, Christy and Spencer, 2005) and are reflected in Figure 3.4.3, but the trend in the tropics is still significantly smaller than those both in the troposphere using T2 and T4 and at the surface. After 1987, when MSU channel 3 is available, Fu and Johanson (2005) find a systematic increasing temperature trend with altitude throughout the tropics. Mears and Wentz (2005) computed their own alternative $T2_{LT}$ record and find a $T2_{LT}$ trend 0.1°C decade^{-1} larger than the revised UAH.

Comparisons of tropospheric radiosonde station data with collocated satellite data (Christy and Norris, 2004) show considerable scatter and root mean square differences of UAH satellite data with radiosondes are substantial (Hurrell et al., 2000). Although Christy and Norris (2004) found good agreement between median radiosonde temperature trends and UAH trends, comparisons are more likely to be biased by spurious cooling than by spurious warming in unhomogenised (Sherwood et al., 2005) and even homogenised (Randel and Wu, 2006) radiosonde data (see Section 3.4.1.1 and Appendix 3.B.5.1). In the stratosphere, radiosonde trends are more negative than both MSU retrievals, especially when compared with RSS, and this is very likely due to changes in sondes and their processing for radiation corrections (Randel and Wu, 2006).

[INSERT FIGURE 3.4.3 HERE]

Geographical patterns of the linear trend in tropospheric temperature 1979–2004 (Figure 3.4.4) are qualitatively similar in the RSS and UAH MSU datasets. Both show coherent warming over most of the NH but UAH shows cooling over parts of the tropical Pacific and tropospheric temperature trends differ south of 45°S where UAH indicate more cooling than RSS.

[INSERT FIGURE 3.4.4 HERE]

3.4.1.3 Reanalyses

A comprehensive global reanalysis, ERA-40 (Uppala et al., 2005) completed since the TAR extends from September 1957 to August 2002. Reanalysis is designed to prevent changes in the analysis system from contaminating the climate record, as occurs with global analyses from operational numerical weather prediction, and it compensates for some but not all of the effects of changes to the observing system (see Appendix 3.B.5.3). Unlike the earlier NRA which assimilated satellite retrievals, ERA-40 assimilated bias-adjusted radiances including MSU data (Harris and Kelly, 2001; Uppala et al., 2005), and the assimilation procedure itself takes account of orbital drift and change in satellite height, factors that have to be addressed in direct processing of MSU radiances for climate studies (e.g., Christy et al., 2003; Mears et al., 2003; Mears and Wentz, 2005). Onboard calibration biases are treated indirectly via the influence of other datasets. Nonetheless, the veracity of low-frequency variability in atmospheric temperatures is compromised in ERA-40 by residual problems in bias corrections.

Trends and low-frequency variability of large-scale surface air temperature from ERA-40 and from the monthly climate station data analysed by Jones and Moberg (2003) are in generally good agreement from the late 1970s onwards (see also Section 3.2.2.1). Temperatures from ERA-40 vary quite coherently throughout the planetary boundary layer over this period, and earlier for regions with consistently good coverage from both surface and upper-air observations (Simmons et al., 2004).

Processed MSU records of layer temperature have been compared with equivalents derived from the ERA-40 analyses (Santer et al., 2004). The use of deep layers conceals disparate trends at adjacent tropospheric levels in ERA-40. Relatively cold tropospheric values before the satellite era arose from a combination of scarcity of radiosonde data over the extratropical SH and a cold bias of the assimilating model, giving a tropospheric warming trend that is clearly too large when taken over the full period of the reanalysis (Bengtsson et al., 2004; Simmons et al., 2004; CCSP, 2006). ERA-40 also exhibits a middle-tropospheric cooling over most of the tropics and subtropics since the 1970s, that is certainly too strong owing to a warm bias in the analyses for the early satellite years.

Tropospheric patterns of trends from ERA-40 are similar to Figure 3.4.4, with coherent warming over the NH, although over the SH ERA-40 indicates no net cooling. These differences are not fully understood, although the treatment of surface emissivity anomalies over snow- and ice-covered surfaces may contribute (Swanson, 2003). The large-scale patterns of stratospheric cooling are similar in ERA-40 and the MSU datasets (Santer et al., 2004). However, the ERA-40 analyses in the lower stratosphere are biased cold relative to radiosonde data in the early satellite years reducing downward trends. At high southern latitudes ERA-40 shows strong temperature trends in 1979–2001, in good accord with Antarctic radiosonde data. Section 3.5 relates the trends to atmospheric circulation changes.

3.4.1.4 The Tropopause

The tropopause marks the division between the troposphere and stratosphere and generally a minimum in the vertical profile of temperature. The height of the tropopause is affected by the heat balance of both the troposphere and the stratosphere. For example, when the stratosphere warms owing to absorption of radiation by volcanic aerosol, the tropopause is lowered. Conversely, a warming of the troposphere raises the tropopause, as does a cooling of the stratosphere. The latter is expected as a result of increasing greenhouse gas concentrations and stratospheric ozone depletion. Accordingly, changes in the height of the tropopause provide a sensitive indicator of human effects on climate. Inaccuracies and spurious trends in NRA preclude their use in determining tropopause trends (Randel et al., 2000) although they were found useful for interannual variability. Over 1979 to 2001, tropopause height increased by nearly 200 meters (as a global average) in ERA-40, partly due to tropospheric warming plus stratospheric cooling (Santer et al., 2004). Atmospheric temperature changes in the UAH and RSS satellite MSU datasets (see Section 3.4.1.2) were found to be more highly correlated with changes in ERA-40 than with those in NRA, illustrating the improved quality of ERA-40 and satellite data. The Santer et al. (2004) results provide support for warming of the troposphere and cooling of the lower stratosphere over the last four decades of the 20th century, and indicate that both of these changes in atmospheric temperature have contributed to an overall increase in tropopause height. The radiosonde-based analyses of Randel et al. (2000), Seidel et al. (2001) and Highwood et al. (2000) also show increases in tropical tropopause height.

3.4.1.5 Synthesis and comparison with the surface temperatures

Figure 3.4.2 presents the radiosonde and satellite global time series and Figure 3.4.3 gives a summary of the linear trends for 1979–2004. Values at the surface are from NOAA (NCDC), NASA (GISS), UKMO/CRU (HadCRUT2v), and the NRA and ERA-40 reanalyses. Trends aloft are for the lower troposphere corresponding to $T2_{LT}$, T2, T4 and also the linear combination of T2 and T4 to better depict the entire troposphere as given by Fu et al (2004a). In addition to the reanalyses, the results from the satellite-based methods from UAH, RSS and VG2 are given along with radiosonde estimates from HadAT2 and RATPAC. Panels show the global mean and the tropical mean from 20°N to 20°S. In both regions the radiosonde coverage is incomplete, causing spatial sampling error. The ERA-40 trends only extend through August 2002. VG2 is available only for T2. The error bars plotted here are 95% confidence limits associated with sampling a finite record where an allowance has been made for temporal autocorrelation in computing degrees of freedom (Appendix 3.A). However, the error bars do not include spatial sampling uncertainty, which increases the noise variance. Noise typically cuts down on temporal autocorrelation and reduces the temporal sampling error bars, which is why the RATPAC error bars are often smaller than the rest. Other sources of "structural" and "internal" errors of order 0.09°C for 95% levels (Mears and Wentz, 2005) (see Appendix 3.B.5) are also not explicitly accounted for here. Structural uncertainties (Thorne et al., 2005b) reflect divergence between different datasets after the common climate variability has been accounted for. Hence, use of difference time series better brings out structural and parametric errors, as seen for instance in T2 for RSS vs UAH in Fu and Johanson (2005); see also CCSP (2006).

From Figure 3.4.2 the first dominant impression is that overall, the records agree remarkably well, especially in the timing and amplitude of interannual variations. This is especially true at the surface, and even the tropospheric records from the two radiosonde datasets agree reasonably well, although HadAT2 has lower values in the 1970s. In the lower stratosphere, all records replicate the dominant variations and the pulses of warming following the volcanic eruptions that occur as indicated on the figure. The sonde records differ prior to 1963 in the lower stratosphere when fewer observations were available, and differences also emerge among all datasets after about 1992, with the sonde values lower than the satellite temperatures. The focus on linear trends tends to emphasize these relatively small differences.

A linear trend over the long term is often not a very good approximation of what has occurred (Seidel and Lanzante, 2004; Thorne et al., 2005a,b), and alternative interpretations are to factor in the abrupt 1976–1977 climate regime shift (Trenberth, 1990) and episodic stratospheric warming and tropospheric cooling for the 2 years following major volcanic eruptions. Hence the confidence limits for linear trends (Figure 3.4.3) are very large in the lower stratosphere owing to the presence of the large warming perturbations from volcanic eruptions, and a linear trend is not a good fit to the data. In the troposphere the confidence limits are much wider in the tropics than globally, reflecting the strong interannual variability associated with ENSO, so again a linear fit is not a good representation of the record.

Radiosonde, satellite observations and reanalyses agree that there has been global stratospheric cooling since 1979 (Figures 3.4.2, 3.4.3), although radiosondes almost certainly still overestimate the cooling owing to residual effects of changes in instruments and processing (such as for radiation corrections) (Lanzante et al., 2003b; Sherwood et al., 2005; Randel and Wu, 2006) and possibly increased sampling of cold conditions owing to stronger balloons (Parker and Cox, 1995). As the stratosphere is cooling and T2 has a 15% signal from there, it is virtually certain that the troposphere must be warming at a significantly greater rate than indicated by T2 alone. Thus, the tropospheric record adjusted for the stratospheric contribution to T2 has warmed more than T2 in every case. The differences range from 0.06°C decade^{-1} for ERA-40 (which has a warm-biased stratospheric trend) to 0.09°C decade^{-1} for both radiosonde and NRA datasets. For UAH and RSS the difference is 0.07°C decade^{-1}. The weakest tropospheric trends occur for NRA, which, unlike ERA-40 (Trenberth, 2004), did not allow for changes in greenhouse gas increases over the record, so that the NRA trends are unreliable (Randel et al., 2000). Quite aside from the radiative forcing of the model, this affects the satellite retrievals in the infrared, as carbon dioxide has increased. Upward trends at high surface mountain stations are stronger than NRA free atmosphere temperatures at nearby locations (Pepin and Seidel, 2005). The records suggest that since 1979 the global tropospheric trends are similar to those at the surface, although RSS, and by inference VG2, indicate greater tropospheric than surface warming, whereas UAH and the radiosonde record – but note its imperfections discussed above – suggest the reverse. This appears to also be the case in the tropics, although the scatter is greater there. Amplification occurs for the RSS fields, especially after 1987 when there are increasing trends with altitude throughout the troposphere based on T2, T3 and T4 (Fu and Johanson, 2005). In the tropics, the theoretically expected amplification of temperature perturbations with height is borne out by interannual fluctuations (ENSO) in radiosonde, RSS, UAH, and model data (Santer et al., 2005), and only the radiosonde records and UAH are at odds for trends. If the radiosondes were corrected for radiation effects (Sherwood et al., 2005), then they too would probably show increased warming with altitude.

The global mean trends since 1979 disguise many regional differences. In particular, in winter much larger temperature trends are present at the surface over northern continents than at higher levels (CCSP, 2006) (see Figures 3.2.9, 3.2.10 and Question 3.1, Figure 1). These are associated with weakening of shallow winter-time temperature inversions and the strong stable surface layers, that have little signature in the main troposphere. Such changes are related to changes in surface winds and atmospheric circulation (see Section 3.6.4).

In summary, for the period since 1958, overall global and tropical tropospheric warming estimated from radiosondes has slightly exceeded surface warming (Figure 3.4.2 and CCSP (2006). The climate shift of 1976 appeared to yield greater tropospheric than surface warming (Figure 3.4.2); such variations of climate make differences between the surface and tropospheric temperature trends since 1979 unsurprising. After 1979, there has also been global and tropical tropospheric warming; however it is uncertain whether tropospheric warming has exceeded that at the surface because the spread of trends among tropospheric data sets encompasses the surface warming trend. A further complexity is that surface trends have been greater over land than over ocean. Substantial cooling has occurred in the lower stratosphere. Compensation for the effects of stratospheric cooling trends on the T2 record (a cooling of about 0.08°C decade^{-1}) has been an important development. However, a linear trend is a poor fit to the data in the stratosphere and the tropics at all levels. The overall global variability is well replicated by all records, although small relative trends exacerbate the differences between records. Inadequacies in the observations and analytical methods result in structural uncertainties that still contribute to the differences between surface and tropospheric temperature trends, and revisions continue to be made. Changes in the height of the tropopause since 1979 are consistent with overall tropospheric warming as well as stratospheric cooling.

3.4.2 Water Vapour

Water vapour is a key climate variable. In the lower troposphere, condensation of water vapour into precipitation provides latent heating which dominates the structure of tropospheric diabatic heating (Trenberth and Stepaniak, 2003a,b). Water vapour is also the most important gaseous source of infrared opacity in the atmosphere, accounting for about 60% of the natural greenhouse effect for clear skies (Kiehl and Trenberth, 1997), and provides the largest positive feedback in model projections of climate change (Held and Soden, 2000).

Water vapour at the land surface has been measured since the late-19th century, but it is only since the 1950s that a database sufficient for climate studies has been compiled. The concentration of surface water vapour is typically reported as the vapour pressure, dewpoint temperature or relative humidity. Using physical relationships, it is possible to convert from one to the other, but the conversions are exact only for instantaneous values. As the relationships are non-linearly related to air temperature, errors accumulate as data are averaged to daily and monthly periods. Slightly more comprehensive data exist for oceanic areas, where the dewpoint temperature is included as part of the ICOADS database, but few analyses have taken place for periods before the 1950s.

The network of radiosonde measurements provides the longest record of water vapour measurements in the atmosphere, dating back to the mid-1940s. However, early radiosonde sensors suffered from significant measurement biases, particularly for the upper troposphere, and changes in instrumentation with time often lead to artificial discontinuities in the data record (e.g., see Elliott et al., 2002). Consequently, most of the analysis of radiosonde humidity has focused on trends for altitudes below 500 hPa and are restricted to those stations and periods for which stable instrumentation and reliable moisture soundings are available.

Additional information on water vapour can be obtained from satellite observations and reanalysis products. Satellite observations provide near-global coverage and thus represent an important source of information over the oceans, where radiosonde observations are scarce, and in the upper troposphere, where radiosonde sensors are often unreliable.

3.4.2.1 Surface and lower troposphere water vapour

Boundary layer moisture strongly determines the longwave radiative flux from the atmosphere to the surface. It also accounts for a significant proportion of the direct absorption of solar radiation by the atmosphere. The TAR reported widespread increases in surface water vapour in the NH. The overall sign of these trends have been confirmed from analysis of specific humidity over the United States (Robinson, 2000) and over China from 1951–1994 (Wang and Gaffen, 2001), particularly for observations made at night. Differences in the spatial, seasonal and diurnal patterns of these changes were found with strong sensitivity of the results to the network choice. Philipona et al. (2004) infer rapid increases in surface water vapour over central Europe from cloud-cleared LW radiative flux measured over the period 1995–2003. Subsequent analyses (Philipona et al., 2005) confirm that changes in integrated water vapour for this region are strongly coupled to the surface temperature, with regions of warming experiencing increasing moisture and regions of cooling experiencing decreasing moisture. For Central Europe, Auer et al. (2006) demonstrate increasing moisture trends. Their vapour pressure series from the Greater Alpine Region closely follow the decadal to centennial scale warming at both urban lowland and rural summit sites. In Canada, van Wijngaarden and Vincent (2005) found a decrease in relative humidity of several percent in the spring for 75 stations, after correcting for instrumentation changes, but little change in relative humidity elsewhere or for other seasons. Ishii et al. (2005) report that globally averaged dew points over the ocean have risen by about 0.25°C between 1950 and 2000. Increasing extremes in summer dew points, and increased humidity during summer heat waves, were found at three stations in northeastern Illinois (Sparks et al., 2002; Changnon et al., 2003) and attributed in part to changes in agricultural practices in the region.

Dai (2006) analyzed near global (60°S–75°N) synoptic data for 1976 to 2005 from ships and buoys and over 15,000 land stations for specific humidity, temperature and relative humidity. Nighttime relative humidity was greater than daytime by 2 to 15% over most land areas, as temperatures undergo a diurnal cycle, while moisture does not change much. The global trends of near-surface relative humidity were small, although there were statistically-significant decreases of –0.1 to –0.2% decade^{-1} over the global ocean. Trends in specific humidity tended to follow surface temperature trends with a global-average increase of 0.06 g/kg decade^{-1} (1976–2004). The rise in specific humidity corresponded to about 4.9%, 4.3% and 5.7% per 1°C warming over the globe, land and ocean respectively. Over the ocean this is close to constant relative humidity and the Clausius-Clapeyron relationship.

For the lower troposphere, water vapour information has been available from the TIROS series of Operational Vertical Sounder (TOVS) since 1979 and also from the Scanning Multichannel Microwave Radiometer (SMMR) from 1979–1984. However, the main improvement occurred with the introduction of the SSM/I in mid-1987 (Wentz and Schabel, 2000). Retrievals of column-integrated water vapour from SSM/I are generally regarded as providing the most reliable measurements of lower tropospheric water vapour over the oceans, although issues pertaining to the merging of records from successive satellites do arise (Trenberth et al., 2005a; Sohn and Smith, 2003).

Significant interannual variability of column-integrated water vapour has been observed using TOVS, SMMR and SSM/I data. In particular column water vapour over the tropical oceans increased by 1–2mm during the 1982/1983, 1986/1987 and 1997/1988 El Niño events (Soden and Schroeder, 2000; Allan et al., 2003: Trenberth et al., 2005a) and reduced by a smaller magnitude in response to global cooling following the eruption of Mt. Pinatubo in 1991 (Soden et al., 2002; Trenberth and Smith, 2005) (see also Chapter 8, Section 8.6.3.1). The linear trend based on monthly SSM/I data over the oceans was 0.40 ± 0.09 mm decade^{-1} or about 1.3% decade^{-1} for 1988–2003, and 1.2 ± 0.3% decade^{-1} for 1988–2004 (Figure 3.4.5). Since the trends are similar in magnitude to the interannual variability, it is likely that the latter impacts the magnitude of the linear trends. The trends are overwhelmingly positive in spatial structure, but also suggestive of an ENSO influence. As noted by Trenberth et al. (2005a), most of the patterns associated with the interannual variability and linear trends can be reproduced by scaling the observed SST changes over this period by 7.8% K^{-1}, which is consistent with a constant relative humidity increase in water vapour mixing ratio. Given observed SST increases, this implies an overall increase in water vapour of order 5% over the 20th century and 4% since about 1970.

An independent check on globally vertically-integrated water vapour amounts is whether the change in water vapour mass is reflected in the surface pressure field, as this is the only significant influence on the global atmospheric mass to within measurement accuracies. As Trenberth and Smith (2005) show, such checks indicate considerable problems prior to 1979 in reanalyses, but results are quite good thereafter for ERA-40. Evaluations of column integrated water vapour from NVAP (Randel et al., 1996), and reanalyses datasets from NRA, NCEP-2 reanalysis and ERA-15/ERA-40 (see Appendix 3.B.5.3) reveal several deficiencies and spurious trends, which limit their utility for climate monitoring (Zveryaev and Chu, 2003; Trenberth et al., 2005a; Uppala et al., 2005). The spatial distributions, trends and interannual variability of water vapour over the tropical oceans are not always well reproduced by reanalyses, even after the 1970s (Allan et al., 2002; Trenberth et al., 2005a; Allan et al., 2004).

To summarize, global, local and regional studies all indicate increases in moisture in the atmosphere near the surface, but highlight differences between regions and between day and night. Satellite observations of oceanic lower tropospheric water vapour reveal substantial variability during the last two decades. This variability is closely tied to changes in surface temperatures, with the water vapour mass changing at roughly the same rate at which the saturated vapour pressure does. A significant upward trend is observed over the global oceans and some northern hemisphere land areas although the calculated trend is likely influenced by large interannual variability in the record.

[INSERT FIGURE 3.4.5 HERE]

3.4.2.2 Upper-tropospheric water vapour

Water vapour in the mid and upper troposphere accounts for a large part of the atmospheric greenhouse effect and is believed to be an important amplifier of climate change (Held and Soden, 2000). Changes in upper-tropospheric water vapour in response to a warming climate have been the subject of significant debate.

Due to instrumental limitations, long-term changes of water vapour in the upper troposphere are difficult to assess. Wang et al. (2001) found an increasing trend of 1–5% decade^{-1} in relative humidity, with the largest increases in the upper troposphere, using 17 radiosonde stations in the tropical west Pacific. Conversely, a combination of Microwave Limb Sounder (MLS) and Halogen Occultation Experiment (HALOE) measurements at 215 hPa suggested a reduction in moisture with increasing temperature (Minschwaner and Dessler, 2004) on interannual time scales, but no clear trend in relative humidity was evident.

Maistrova et al. (2003) report an increase in specific humidity at 850 hPa and a decrease from 700–300 hPa for 1959–2000 in the Arctic, based on data from ships and temporary stations as well as permanent stations. In general the radiosonde trends are highly suspect owing to the poor quality and changes over time in the humidity sensors (e.g., Wang et al., 2002a). Comparisons of water vapour sensors during recent intensive field campaigns have produced a renewed appreciation of random and systematic errors in radiosonde measurements of upper-tropospheric water vapour and of the difficulty in developing accurate corrections for these measurements (Guichard et al., 2000; Revercombe et al., 2003; Wang et al., 2003; Turner et al., 2003; Soden et al., 2004; Miloshevich et al., 2004).

Information on the decadal variability of upper-tropospheric relative humidity (UTH) is now provided by 6.7 micron thermal radiance measurements from Meteosat (Picon et al., 2003) and the High-resolution Infrared Sounder (HIRS) series of instruments flying on NOAA operational polar orbiting satellites (Bates and Jackson, 2001; Soden et al., 2005). These products rely on the merging together of many different satellites to ensure uniform calibration. The HIRS channel 12 (T12) data have been most extensively analysed for variability and show linear trends in relative humidity of order ±1% decade^{-1} at various latitudes (Bates and Jackson, 2001) but these trends are difficult to separate from larger interannual fluctuations due to ENSO (McCarthy and Toumi, 2004) and are negligible when averaging over the tropical oceans (Allan et al., 2003).

In the absence of large changes in relative humidity, the observed warming of the troposphere (see Section 3.4.1) implies that the specific humidity in the upper troposphere should have increased. As the upper troposphere moistens the emission level for T12 increases due to the increasing opacity of water vapour along the satellite line of sight. However, the emission level for the MSU T2 remains constant because of its dependence on the concentration of oxygen that does not vary by any appreciable amount. Therefore, if the atmosphere moistens, the brightness temperature difference T2-T12 will increase over time due to the divergence of their emission levels (Soden et al., 2005). This radiative signature of upper tropospheric moistening is evident in the positive trends of T2-T12 for the period 1982–2004 (Figure 3.4.6). If the specific humidity in the upper troposphere had not increased over this period, the emission level for T12 would have remain unchanged and T2-T12 would show little trend over this period (dashed line in Figure 3.4.6).

Clear-sky OLR is also highly sensitive to upper-tropospheric water vapour and a number of scanning instruments have made well-calibrated but non-overlapping measurements since 1985 (see Section 3.4.3). Over this period, the small changes in clear-sky OLR can be explained by the observed temperature changes while maintaining a constant relative humidity (Wong et al., 2000; Allan and Slingo, 2002) and changes in well-mixed greenhouse gases (Allan et al., 2003). This again implies a positive relationship between specific humidity and temperature in the upper troposphere.

To summarize, the available data do not indicate a detectable trend in upper-tropospheric relative humidity. However, there is now evidence for global increases in upper-tropospheric specific humidity over the past two decades, which is consistent with the observed increases in tropospheric temperatures and the absence of any change in relative humidity.

[INSERT FIGURE 3.4.6 HERE]

3.4.2.4 Stratospheric water vapour

The TAR noted an apparent increase of roughly 1% per year in stratospheric water vapour content (~0.05 ppmv/yr) during the last half of the 20th century (Kley et al., 2000; Rosenlof et al., 2001). This is based on data taken at mid-latitudes, and from multiple instruments. However, the longest series of data come from just two locations in North America with no temporal overlap. The combination of measurement uncertainties and relatively large variability on time scales from months to years warrants some caution when interpreting the longer-term trends (Kley et al., 2000; Fueglistaler and Haynes, 2005). The moistening is more convincingly documented during the 1980s and most of the 1990s than earlier, due to a longer continuous record (the CMDL frost-point balloon record from Boulder, Colorado; Oltmans et al., 2000) and the availability of satellite observations during much of this period. However, discrepancies between satellite- and balloon-measured variations are apparent at decadal time scales, largely over the latter half of the 1990s (Randel et al., 2004).

An increase in stratospheric water vapour has important radiative and chemical consequences (see also Chapter 2, Section 2.3.8). These may include a contribution to the recent observed cooling of the lower stratosphere and/or warming of the surface (Forster and Shine, 1999, 2002; Smith et al., 2001), although the exact magnitude is difficult to quantify (Oinas et al., 2001; Forster and Shine, 2002). Some efforts to reconcile observed rates of cooling in the stratosphere with those expected based on observed changes in ozone and CO_2 since 1979 (Langematz et al., 2003; Shine et al., 2003) have found discrepancies in the lower stratosphere consistent with an additional cooling effect of a stratospheric water vapour increase. However, Shine et al. (2003) noted that because the water vapour observations over the period of consideration are not global in extent, significant uncertainties remain as to whether radiative effects of a water vapour change are a significant contributor to the stratospheric temperature changes. Moreover, other studies which account for uncertainties in the ozone profiles and temperature trends, and natural variability can reconcile the observed stratospheric temperature changes without the need for sizable water vapour changes (Ramaswamy and Schwarzkopf, 2002; Schwarzkopf and Ramaswamy, 2002).

Although methane oxidation is a major source of water in the stratosphere, and has been increasing over the industrial period, the noted stratospheric trend appears to be too large to attribute to methane oxidation alone (Oltmans et al., 2000; Kley et al., 2000). Therefore, other contributors to an increase in stratospheric water vapour are under active investigation. It is likely that different mechanisms are affecting water vapour trends at different altitudes. Aviation emits a very small amount of water vapour directly into the stratosphere (IPCC, 1999). Several indirect mechanisms have also been considered including: a) volcanic eruptions (Considine et al., 2001; Joshi and Shine, 2003); b) biomass burning aerosol (Sherwood, 2002; Andreae et al., 2004); c) tropospheric SO_2 (Notholt et al., 2005); and d) changes to methane oxidation rates from changes in stratospheric chlorine, ozone and OH (Röckmann et al., 2004). Other proposed mechanisms relate to changes in tropopause temperatures or circulation (Stuber et al., 2001; Dessler and Sherwood, 2004; Fueglistaler et al., 2004; Nedoluha et al., 2003; Roscoe, 2004; Rosenlof, 2002; Zhou et al., 2001).

It has been assumed that temperatures near the tropical tropopause control stratospheric water vapour according to equilibrium thermodynamics, importing higher water vapour values into the stratosphere when temperatures are warmer. However, tropical tropopause temperatures have cooled slightly over the period of the stratospheric water vapour increase (see Section 3.4.1 and Seidel et al., 2001; Zhou et al, 2001). This makes the mid-latitude lower stratospheric increases harder to explain (Fueglistaler and Haynes, 2005). Satellite observations (Read et al., 2004) show water vapour injected above the tropical tropopause by deep convective clouds, bypassing the traditional control point. Changes in the amount of condensate sublimating in this layer may have contributed to the upward trend, but to what degree is uncertain (Sherwood, 2002). Another suggested source for temperature-independent variability is changes in the efficiency with which air is circulated through the coldest regions before entering the stratosphere (Hatushika and Yamazaki, 2003; Fueglistaler et al., 2004; Bonnazola and Haynes, 2004; Dessler and Sherwood, 2004). However, it is not yet clear that a circulation-based mechanism can explain the observed trend (Fueglistaler and Haynes, 2005).

The TAR noted a stalling of the upward trend in water vapour during the last few years observed at that time. This change in behaviour has persisted, with a near-zero trend in stratospheric water vapour between 1996 and 2000 (Randel et al., 2004; Nedoluha et al., 2003). The upward trend of methane is also smaller and is currently close to zero (see Chapter 2, Section 2.3.2). Further, at the end of year 2000 there was a dramatic drop in water vapour in the tropical lower stratosphere as observed by both satellite and CMDL balloon data

(Randel et al., 2004). Temperatures observed near the tropical tropopause also dropped, but the processes producing the tropical tropopause cooling itself are currently not fully understood. The propagation of this recent decrease through the stratosphere should ensure flat or decreasing stratospheric moisture for at least the next few years.

To summarize, water vapour in the stratosphere has shown significant long-term variability and an apparent upward trend over the last half of the 20th century but with no further increases since 1996. It does not appear that this behaviour is a straightforward consequence of known climate changes. Although ideas have been put forward, there is no consensus as to what caused either the upward trend or its recent disappearance.

3.4.3 Clouds

Clouds play an important role in regulating the flow of radiation at the top of the atmosphere and at the surface. The response of cloud cover to increasing greenhouse gases currently represents the largest uncertainty in model predictions of climate sensitivity (see Chapter 8). Surface observations made at weather stations and onboard ships provide the longest available records of cloud cover changes dating back over a century. Surface observers report the all-sky conditions, which include the sides as well as bottoms of clouds, but are unable to report upper level clouds which may be obscured from the observer's view. Although limited by potential inhomogeneities in observation times and methodology, the surface-observed cloud changes are often associated with physically consistent changes in correlative data, strengthening their credibility. Since the mid-1990s, especially in the United States and Canada, human observations at the surface have been widely replaced with automated ceilometer measurements, which measure only directly overhead low clouds rather than all-sky conditions. In contrast, satellites generally only observe the upper-most level of clouds and have difficulty detecting optically-thin clouds. While satellite measurements do provide much better spatial and temporal sampling than can be obtained from the surface, their record is much shorter in length. These disparities in how cloud cover is observed contribute to the lack of consistency between surface and satellite measured changes in cloudiness. Condensation trails ("contrails") from aircraft exhaust may expand to form cirrus clouds and these and cosmic ray relations to clouds are dealt with in Chapter 2.

3.4.3.1 Surface cloud observations

As noted in the TAR and extended with more recent studies, surface observations suggest increased total cloud cover since the middle of the last century over many continental regions including the United States (Sun, 2003; Groisman et al., 2004; Dai et al., 2006); the former USSR (Sun and Groisman, 2000; Sun et al., 2001), Western Europe, mid-latitude Canada, and Australia (Henderson-Sellers, 1992). This increasing cloudiness since 1950 is consistent with an increase in precipitation and a reduction in DTR (Dai et al., 1997a; Dai et al., 1999; Dai et al., 2006). However, decreasing cloudiness has been reported over China during 1951–1994 (Kaiser, 1998). Cloud cover during the past 50 years has also decreased over Italy (Maugeri et al., 2001) and over Central Europe (Auer et al., 2006). If the analyses are restricted to after about 1971, changes in continental cloud cover become more mixed. For example, using a worldwide analysis of cloud data (Hahn and Warren, 2003; Minnis et al., 2004) regional reductions were found since the early 1970s over western Asia and Europe but increases over the United States.

Changes in total cloud cover along with an estimate of precipitation over global and hemispheric land (excluding North America) from 1976–2003 are shown in Figure 3.4.7. During this period, secular trends over land are small. The small variability evident in land cloudiness appears to be correlated with precipitation changes, particularly in the SH (Figure 3.4.7). Note that surface observations from North America are excluded in this figure due to the declining number of human cloud observations since the early 1990s over the United States and Canada, as human observers have been replaced with Automated Surface Observation Systems (ASOS) from which cloud amounts are less reliable and incompatible with previous records (Dai et al., 2006). However, independent human observations from military stations suggest an increasing trend (~1.4% of sky per decade) in U.S. total cloud cover.

The TAR also noted multi-decadal trends in cloud cover over the ocean. Updated analysis of this information (Norris, 2005a) has documented substantial decadal variability and decreasing trends in upper-level cloud cover over mid-latitude and low-latitude oceans since 1952. However, there are no direct observations of upper-level cloud from the surface and instead Norris (2005a) infers them from reported total and low cloud cover assuming a random overlap. These results partially reverse the finding of increasing trends in mid-level cloud amount in the northern mid-latitude oceans that was reported in the TAR although the new study does not distinguish between high and middle clouds. Norris (2005b) found that upper-level cloud cover had increased over the equatorial South Pacific between 1952 and 1997 and decreased over the adjacent subtropical regions, the tropical Western Pacific, and the equatorial Indian Ocean. This pattern is consistent with decadal changes in precipitation and atmospheric circulation over these regions noted in the TAR, which further supports, their validity. Deser et al. (2004) found similar spatial patterns in interdecadal variations of total cloud cover, SST, and precipitation over the tropical Pacific and Indian Oceans during 1900–1995. In contrast, low-cloud cover increased over almost all of the tropical Indo-Pacific Ocean, but this increase bears little resemblance to changes in atmospheric circulation over this period, suggesting that it may be spurious (Norris, 2005b). When averaged globally, oceanic cloud cover appears to have increased over the last 30 years or more (e.g., Ishii et al., 2005).

Large-scale cloud changes occur with ENSO, as during El Niño cloud generally decreases over land throughout much of the tropics and subtropics, but increases over the ocean in association with precipitation changes (Curtis and Adler, 2003), and even multi-decadal variations are affected by the 1976–1977 climate shift (Deser et al., 2004), and these dominate the low latitude trends from 1971–1996 found in Hahn and Warren (2003). Similarly, a positive trend in the NAO from 1971 to 1996 is associated with a negative trend in cloud over southern Europe and Northern Africa (see Section 3.6.4).

[INSERT FIGURE 3.4.7 HERE]

3.4.3.2 Satellite cloud observations

Since the TAR, there has been considerable effort in the development and analysis of satellite datasets for documenting changes in global cloud cover over the past few decades. The most comprehensive cloud climatology is that of the International Satellite Cloud Climatology Project (ISCCP), begun in June 1983. ISCCP shows an increase in globally-averaged total cloud cover of ~2% from 1983 to 1987, followed by a decline of ~4% from 1987 to 2001 (Rossow and Dueñas, 2004). Cess and Udelhofen (2003) documented decreasing ISCCP total cloud cover in all latitude zones between 40ºS and 40ºN. Norris (2005a) found that both ISCCP and ship synoptic reports show consistent reductions in middle or high-cloud cover from the 1980s to the 1990s over low- and mid-latitude oceans. Minnis et al. (2004) also found consistent trends in high-level cloud cover between ISCCP and surface observations over most areas except the North Pacific, where they differed by almost 2%/decade. In addition, analysis of SAGE II data reveal a decline in cloud frequency above 12 km between 1985 and 1998 (Wang et al., 2002b) that is consistent with the decrease in upper-level cloud cover noted in ISCCP and ocean surface observations. The decline in upper-level cloud cover since 1987 may also be consistent with a decrease in reflected SW radiation during this period as measured by the Earth Radiation Budget Satellite (ERBS) (see Section 3.4.4). Radiative transfer calculations, which use the ISCCP cloud properties as input, are able to independently reproduce the decadal changes in outgoing LW and reflected SW reported by ERBS (Zhang et al., 2004c).

Analyses of the spatial trends in ISCCP cloud cover reveal the influence of changes in satellite view angle and discontinuities associated with changes in satellite coverage which impact the global mean anomaly time series (Norris, 2000; Dai et al., 2006). The ISCCP spurious variability may occur primarily in low-level clouds with the least optical thickness (the ISCCP “cumulus” category) (Norris, 2005a), due to discontinuities in satellite view angles associated with changes in satellites. Such biases likely contribute to ISCCP’s negative cloud cover trend, although their magnitude and impact on radiative flux calculations using ISCCP cloud data are not yet known. Additional artefacts, including radiometric noise, navigation and rectification errors are present in the ISCCP data (Norris, 2000), but the effects of known and unknown artefacts on ISCCP cloud and flux data have not yet been quantified.

Other satellite data sets show conflicting decadal changes in total cloud cover. For example, analysis of cloud cover changes from the HIRS shows a slight increase in cloud cover between 1985 and 2001 (Wylie et al., 2005). However, spurious changes have also been identified in the HIRS dataset, which may impact its estimates of decadal variability. One important source of uncertainty results from the drift in Equatorial Crossing Time (ECT) of polar orbiting satellite measurements (e.g., HIRS and AVHRR) which aliases the large diurnal cycle of clouds into spurious lower-frequency variations. After correcting for ECT drift and other small calibration errors in AVHRR measurements of cloudiness, Jacobowitz et al. (2003) found essentially no trend in cloud cover for the tropics from 1981 to 2000.

While the variability in surface-observed upper-level cloud cover has been shown to be consistent with that observed by ISCCP (Norris, 2005a), the variability in total cloud cover is not, implying differences between ISCCP and surface-observed low cloud cover. Norris (2005a) shows that even after taking into account the difference between surface and satellite views of low-level clouds, the decadal changes between the ISCCP and surface datasets still disagree. The extent to which this results from differences in spatial and temporal sampling or differences in viewing perspective is unclear.

In summary, while there is some consistency between ISCCP, ERBS, SAGE II and surface observations for a reduction in high cloud cover during the 1990s relative to the 1980s, there are substantial uncertainties for decadal trends in all datasets and at present there is no clear consensus on changes in total cloudiness over decadal timescales.

3.4.4 Radiation

3.4.4.1 Top of atmosphere radiation

One important development since the TAR is the apparent unexpectedly large changes in tropical mean radiation flux reported by the Earth Radiation Budget Satellite (ERBS) (Wielicki et al., 2002a, 2002b). A recent reanalysis of the ERBS active cavity broadband data corrects for a 20 km change in satellite altitude between 1985 and 1999 and changes in the SW filter dome (Wong et al., 2006). Based upon the revised (Edition 3_Rev1) ERBS record (Figure 3.4.8), outgoing LW radiation over the tropics appears to have increased by about 0.7 W m^{-2} while the reflected SW radiation decreased by roughly 2.1 W m^{-2} from the 1980s to 1990s (Table 3.5).

These conclusions depend upon the calibration stability of the ERBS non-scanner record which is affected by diurnal sampling issues, satellite altitude drifts, and changes in calibration following a 3-month period when the sensor was powered off (Trenberth, 2002). Moreover, rather than a trend, the reflected SW change may stem mainly from a jump in the record in late 1992 in the ERBS record that is also observed in the ISCCP (version FD) record (Zhang et al., 2004c) but not in the AVHRR Pathfinder record (Jacobowitz et al., 2003). However, careful inspection of the sensor calibration revealed no known issues that can explain the decadal shift in the fluxes despite corrections to the ERBS time-series relating to diurnal aliasing and satellite altitude changes (Wielicki et al., 2002b; Wong et al., 2006).

As noted in Section 3.4.3, the low latitude changes in the radiation budget appear consistent with reduced cloud fraction from ISCCP. Detailed radiative transfer computations, using ISCCP cloud products along with additional global datasets, show broad agreement with the ERBS record of tropical radiative fluxes (Zhang et al., 2004c; Hatzianastassiou et al., 2004; Wong et al., 2006). However, the decrease in reflected SW from the 1980s to the 1990s may be inconsistent with the increase in total and low cloud cover over oceans reported by surface observations (Norris, 2005a) which show increased low cloud occurrence. The degree of inconsistency, however, is difficult to ascertain without information on possible changes in low-level cloud albedo.

While the ERBS satellite provides the only continuous long-term TOA flux record from broadband active cavity instruments, narrow spectral band radiometers have made estimates of both reflected SW and outgoing LW trends using regressions to broadband data, or using radiative transfer theory to estimate unmeasured portions of the spectrum of radiation. Table 3.5 shows the 1980s to 1990s TOA tropical mean flux changes for the ERBS Edition 3 data (Wong et al., 2006), the HIRS Pathfinder data (Mehta and Susskind, 1999), the AVHRR Pathfinder data (Jacobowitz et al., 2003), and the ISCCP FD data (Zhang et al., 2004c).

The most accurate of the datasets in Table 3.5 is believed to be the ERBS Edition 3 active cavity wide field of view data (Wielicki et al., 2005). The ERBS stability is estimated as better than 0.5 Wm^{-2} over the 1985 to 1999 period and the spatial and time sampling noise is less than 0.5 Wm^{-2} on annual time scales (Wong et al., 2006). The outgoing LW changes from ERBS are similar to the decadal changes in the HIRS Pathfinder and ISCCP FD records, but disagree with the AVHRR Pathfinder (Wong et al., 2006). The AVHRR

Pathfinder data also do not support the TOA SW trends. However, calibration issues, narrow-to-broad band conversion, and satellite orbit changes are thought to render the AVHRR record less reliable for decadal changes compared to ERBS (Wong et al., 2006). Estimates of the stability of the ISCCP time series for long-term TOA flux records are 3 to 5 W m^{-2} for SW flux and 1 to 2 W m^{-2} for LW flux (Brest et al., 1997), although the time series agreement of the ISCCP and ERBS records are much closer than these estimated calibration drift uncertainties (Zhang et al., 2004c).

Table 3.5. TOA radiative flux changes from the 1980s to 1990s in W m^{-2}. Values are given as tropical mean (20°S to 20°N) for the 1994–1997 period minus the 1985–1989 period. Dashes are shown where no data are available. From Wong et al. (2006).

Data Source	TOA LW	TOA SW	TOA Net
ERBS Edition 3	0.7	–2.1	1.4
HIRS Pathfinder	0.2	–	–
AVHRR Pathfinder	–1.4	0.7	0.7
ISCCP FD	0.5	–2.4	1.8

The changes in SW measured by ERBS Edition 3 are larger than the clear-sky flux changes due to humidity variations (Wong et al., 2000) or anthropogenic radiative forcing (see Chapter 2). If correct, the large decrease in reflected SW with little change in outgoing LW implies a reduction in tropical low cloud cover over this period. However, specific information on cloud radiative forcing is not available from ERBS after 1989 and, as noted in Section 3.4.3, surface datasets suggest an increase in low cloud cover over this period.

Since most of the net tropical heating of 1.4 W m^{-2} is a decrease in SW reflected flux, the change implies a similar increase in solar insolation at the surface which, if unbalanced by other changes in surface fluxes, would increase the amount of ocean heat storage. Wong et al. (2006) have shown that the changes in global net radiation are consistent with a new ocean heat storage data set from Willis et al. (2004), see Chapter 5 and Figure 5.2.5. Differences between the two datasets are roughly 0.4 W m^{-2}, in agreement with the estimated annual sampling noise in the ocean heat storage data.

Using astronomical observations of visible wavelength solar photons reflected from parts of the Earth to the moon and then back to the Earth at a surface-based observatory, Pallé et al. (2004) estimated a dramatic increase of Earth reflected SW flux of 5.5 W m^{-2} over 3 years. This is unlikely to be real, as over the same time period (2000–03), the CERES broadband data indicates a decrease in SW flux by almost 1 W m^{-2}, much smaller and the opposite sign (Wielicki et al., 2005), and changes in ocean heat storage are more consistent with the CERES data than with the Earthshine indirect observation.

The only long-term time series (1979–2001) of energy divergence in the atmosphere (Trenberth and Stepaniak, 2003b) are based on NRA which, although not reliable for depicting trends, are reliable on interannual times scales for which they show substantial variability associated with ENSO. Analyses by Trenberth and Stepaniak (2003b) reveal more convergence of energy into the deep tropics in the 1980s compared with the 1990s due to differences in ENSO, which may account for at least some of the changes discussed above.

In summary, although there is independent evidence for decadal changes in TOA radiative fluxes over the last two decades, the evidence is equivocal. Changes in the planetary and tropical TOA radiative fluxes are consistent with independent global ocean heat storage data, and are expected to be dominated by changes in cloud radiative forcing. To the extent that they are real, they may simply reflect natural low-frequency variability of the climate system.

[INSERT FIGURE 3.4.8 HERE]

3.4.4.2 Surface radiation

The energy balance at the surface requires net radiative heating to be balanced by turbulent energy fluxes and thus determines the evolution of surface temperature and the cycling of water, which are key parameters of climate change (see Chapter 7, Box 7.1). In recent years several studies have focused on observational evidence of changing surface radiative heating. Reliable shortwave radiative measurement networks exist since the International Geophysical Year in 1957–1958.

A reduction in downward solar radiation ("dimming") of about 1.3% decade^{-1} or about 7 W m^{-2} was observed from 1961 to 1990 at land stations around the world (Liepert, 2002; Gilgen et al., 1998). Additional studies also found declines in surface solar radiation in the Arctic and Antarctic (Stanhill and Cohen, 2001) as well as at sites in the former Soviet Union (Abakumova et al., 1996; Russak, 1990), around the Mediterranean Sea (Omran, 2000 and Aksoy, 1997), China (Ren et al., 2005), the United States (Liepert, 2002), and Southern Africa (Power and Mills, 2005). Stanhill and Cohen (2001) claim an overall reduction globally averaged of 2.7% decade^{-1} but used only 30 records. Hence, in spite of claims of "global dimming", the stations where these analyses have taken place are quite limited in domain and dominated by large urban areas, and the dimming is much less at rural sites (Alpert et al., 2005) or even missing altogether over remote areas, except for identifiable effects of volcanoes, such as Mount Pinatubo in 1991 (Schwartz, 2005). At the majority of 421 analyzed sites the decline in surface solar radiation ended around 1990 and a recovery of about 6 W m^{-2} occurred afterwards (Wild et al., 2004; 2005). The increase in surface solar radiation ("brightening") agrees with satellite and surface observations of reduced cloud cover (Wielicki et al., 2002a; Wang et al., 2002b; Rossow and Dueñas, 2004; Pinker et al., 2005; Norris, 2005b) although there is evidence that some of these changes are spurious (see Section 3.4.3), and these trends are not consistent with surface-observed cloud cover changes. Nor are they consistent with the continued decline in solar radiation at remote sites in the European Alps (Philipona et al., 2004) and China (Kaiser and Qian, 2002; Y. Qian et al., 2006). The continued decline in pan-evaporation (see Section 3.3.3) also relates directly to decreases in solar radiation.

From 1981 to 2003 over central Europe, Philipona and Dürr (2004) showed that decreases in solar radiation at the surface from increases in clouds were cancelled by opposite changes in longwave radiation in the net and that increases in net radiative flux were dominated by the clear-sky longwave radiation component relating to an enhanced water vapour greenhouse effect.

Alpert et al. (2005) provide evidence that a significant component of the reductions may relate to increased urbanisation and anthropogenic aerosol concentrations over the period; see also Chapter 7, Section 7.5.3. This has been detected in solar radiation reductions for polluted regions, e.g., China (Luo et al., 2001), but cloudiness changes must also play a major role, as shown at European sites and the United States (Liepert, 2002; Dai et al., 2006). In the United States, increasing cloud optical thickness and a shift from cloud-free to more cloudy skies are the dominating factors before the aerosol direct effects. Possible causes of the 1990s reversal are reduced cloudiness and also increased cloud-free atmospheric transparency due to the reduction of anthropogenic aerosol concentrations and recovery from the effects of the 1991 eruption of Mt. Pinatubo. See Box 3.2 for more discussion and a likely explanation of these aspects.

Box 3.2: The Dimming of the Planet and Apparent Conflicts in Trends of Evaporation and Pan Evaporation

Several reports have defined a term, "global dimming" (e.g., Cohen et al., 2004). This refers to a widespread reduction of solar radiation received at the surface of the Earth, at least up until about 1990 (Wild et al., 2005). At the same time there is considerable confusion in the literature over conflicting trends in pan evaporation and actual evaporation (Roderick and Farquhar, 2002, 2004, 2005; Ohmura and Wild, 2002; Hobbins et al., 2004; Wild et al., 2004, 2005) although the framework for explaining observed changes exists (Brutsaert and Parlange, 1998). Moreover, recent studies (Alpert et al., 2005; Schwartz, 2005) find that dimming is not global but is rather confined to only large urban areas.

Surface evaporation, or more generally evapotranspiration, depends upon two key components. The first is available energy at the surface, especially solar radiation. The second is the availability of surface moisture, which is not an issue over oceans, but which is related to soil moisture amounts over land. Evaporation pans provide estimates of the potential evaporation that would occur if the surface were wet. Actual evaporation is generally not measured, except at isolated flux towers, but may be computed using bulk flux formulae or estimated as a residual from the surface moisture balance.

The evidence is strong that a key part of the solution to the paradox of conflicting trends in evaporation and pan evaporation lies in changes in the atmospheric circulation and the hydrological cycle such that there has been an increase in cloud and precipitation, which reduce solar radiation available for actual evapotranspiration but also increase soil moisture and make the actual evapotranspiration closer to the potential evapotranspiration. An increase in both cloud and precipitation has occurred over many parts of the land surface (Dai et al., 1999; 2004b; 2006). This reduces solar radiation available for evapotranspiration, as observed since the late 1950s or early 1960s over the United States (Liepert, 2002), parts of Europe and Siberia (Peterson et al., 1995; Abakumova et al., 1996), India (Chattopadhyay and Hulme, 1997), and China (Liu et al., 2004a), and over land more generally (Wild et al., 2004). However, it also increases soil moisture and thereby increases actual evapotranspiration (Milly and Dunne, 2001). Moreover, increased cloud imposes a greenhouse effect and reduces outgoing longwave radiation (Philipona and Dürr, 2004), so that changes in net radiation can be quite small or even of reversed sign. Recent reassessments suggest increasing trends of evapotranspiration over southern Russia during the last 40 years (Golubev et al., 2001) or over the United States during the past 40 or 50 years (Golubev et al., 2001; Walter et al., 2004) in spite of decreases in pan evaporation. Hence, in most, but not all, places the net result has been an increase in actual evaporation but a decrease in pan evaporation. Both are related to observed changes in atmospheric circulation and associated weather.

It is an open question as to how much the changes in cloudiness are associated with other effects, notably impacts of changes in aerosols. Dimming seems to be predominant in large urban areas where pollution plays a role (Alpert et al., 2005). Increases in aerosols are apt to redistribute cloud liquid water over more and smaller droplets, brightening clouds, decreasing the potential for precipitation, and perhaps changing the lifetime of clouds (e.g., Rosenfeld, 2000; Ramanathan et al., 2001; Kaufman et al., 2002); see Chapter 2, Section 2.4 and Chapter 7, Section 7.5.3. Increases in aerosols also reduce direct radiation at the surface under clear skies (e.g., Liepert, 2002), and this appears to be a key part of the explanation in China (Ren et al., 2005).

Another apparent paradox raised by Wild et al. (2004) is that if surface radiation decreases then it should be compensated by a decrease in evaporation from a surface energy balance standpoint, especially given an observed increase in surface air temperature. Of course, back radiation from greenhouse gases and clouds operate in the opposite direction (Philipona and Dürr, 2004). Also, a primary change (not considered by Wild et al., 2004) is in the partitioning of sensible versus latent heat at the surface and thus in the Bowen ratio. Increased soil moisture means that more heating goes into evapotranspiration at the expense of sensible heating, reducing temperature increases locally (Trenberth and Shea, 2005). Temperatures are affected above the surface where latent heating from precipitation is realized, but then the full dynamics of the atmospheric motions (horizontal advection, adiabatic cooling in rising air and warming in compensating subsiding air) comes into play. The net result is a non-local energy balance.

3.5 Changes in Atmospheric Circulation

Changes in the circulation of the atmosphere and ocean are an integral part of climate variability and change. Accordingly regional variations in climate can be complex and sometimes counter-intuitive. For example, a rise in global mean temperatures does not mean warming everywhere, but can result in cooling in some places, due to circulation changes.

This section assesses research since the TAR on atmospheric circulation changes, through analysis of global-scale datasets of mean sea level pressure (MSLP), geopotential heights, jet streams and storm tracks. Related quantities at the surface over the ocean including winds, waves and surface fluxes are also considered. Many of the results discussed are based on reanalysis data sets. Reanalyses provide a global synthesis of all available observations, but are subject to spurious changes over time as observations change, especially in the late 1970s with the introduction of satellite observations. (See Appendix 3.B.5 for a discussion of the quality of reanalyses from a climate perspective).

3.5.1 Surface or Sea Level Pressure

MSLP maps synthesize the atmospheric circulation status. Hurrell and van Loon (1994) noted MSLP changes in the SH beginning in the 1970s while major changes were also occurring over the North Pacific in association with the 1976–1977 climate shift (Trenberth, 1990, Trenberth and Hurrell, 1994). More recently, analyses of sea level pressure from 1948 to 2005 for DJF found decreases over the Arctic, Antarctic and North Pacific, an increase over the subtropical North Atlantic, southern Europe and North Africa (Gillett et al., 2003, 2005), and a weakening of the Siberian High (Gong et al., 2001). The strength of mid-latitude MSLP gradients and associated westerly circulation appears to have increased in both hemispheres, especially during DJF, since at least the late 1970s; see Figure 3.5.1.

The increase in MSLP gradient in the NH appears to significantly exceed simulated internal and anthropogenically-forced variability (Gillett et al. 2003, 2005). However, the significance of changes over the SH is less clear, especially over the oceans prior to satellite observations in the late 1970s, as spurious trends are evident in both major reanalyses (NRA and ERA-40; Marshall, 2003; Bromwich and Fogt, 2004; Trenberth and Smith, 2005; Wang et al., 2006, see also Appendix 3.B.5). Consistent changes (validated with long-term station–based data) do, however, seem to be present since the mid-1970s and are often interpreted in terms of time-averaged signature of weather regimes (Cassou et al., 2004) or annular modes in both hemispheres (Thompson et al., 2000; Marshall, 2003; Bromwich and Fogt, 2004; see Section 3.6).

3.5.2 Geopotential Height, Winds and the Jet Stream

Mean changes in geopotential heights resemble in many ways their MSLP counterparts (Hurrell et al., 2004). Linear trends in 700 hPa height during the solstitial seasons, from ERA-40, are shown in Figure 3.5.1. The 700 hPa level was used as it is the first atmospheric level to lie largely above the east Antarctic ice sheet. NRA and ERA-40 trends agree closely between 1979 and 2001. Over the NH between 1960 and 2000, winter (DJF) and annual means of geopotential height at 850, 500, and 200 hPa decreased over high latitudes and increased over the mid-latitudes, as for MSLP, albeit westward shifted (Lucarini and Russell, 2002). Using NRA, Frauenfeld and Davis (2003) identified a statistically significant expansion of the NH circumpolar vortex at 700, 500, and 300 hPa from 1949–1970. But the vortex has contracted significantly at all levels since then (until 2000) and Angell (2006) found a downward trend in the size of the polar vortex from 1963 to 2001, consistent with warming of the vortex core and analysed increases in 850 to 300 hPa thickness temperatures.

In the NH for 1979–2001 during DJF, height rises occurred between 30° and 50°N at many longitudes, notably over the central north Pacific (Figure 3.5.1). North of 60°N, height changes are consistent with recent occurrences of more neutral phases of the mean polar vortex. Increases in 700 hPa height outweigh decreases in the northern summer (JJA) during 1979–2001. In SH high latitudes, the largest changes are seen in the solstitial seasons (Figure 3.5.1), with changes of opposite sign in many areas between DJF and JJA. Changes during DJF reflect the increasing strength of the positive phase of the SAM (see Marshall, 2003 and Section 3.6.5), with large height decreases over Antarctica and corresponding height increases in the mid-latitudes, through the depth of the troposphere and into the stratosphere. The corresponding enhancement of the near-surface circumpolar westerlies at ~60°S, and associated changes in meridional winds in some sectors, is consistent with a warming trend observed at weather stations over the Antarctic Peninsula and Patagonia (Thompson and Solomon, 2002; see also Sections 3.2.2.4 and 3.6.5). In winter (JJA), there have been height increases over the Antarctic continent since 1979, with a zonal wave 3–4 pattern of rises and falls in southern mid-latitudes. Trends up to 2001 are relatively strong and statistically significant, with annular modes in both hemispheres strongly positive during the 1990s, although less so in recent years. Hence, geopotential height trends in DJF in the SH through 2004 have weakened in the magnitude and significance, but with little change in spatial patterns of trend.

Hemispheric teleconnections are strongly influenced by jet streams, which alter waves and storm tracks (Branstator, 2002). Using NRA from 1979 to 1995, Nakamura et al. (2002) found a weakening of the North Pacific wintertime jet since 1987, allowing efficient vertical coupling of upper-level disturbances with the surface temperature gradients (Nakamura and Sampe, 2002; Nakamura et al., 2004). A trend from the 1970s to the 1990s towards a deeper polar vortex and Iceland Low associated with a positive phase of the NAM in winter (Hurrell, 1995; Thompson et al., 2000; Ostermeier and Wallace, 2003) was accompanied by intensification and poleward displacement of the Atlantic polar frontal jet and associated enhancement of the Atlantic storm track activity (Chang and Fu, 2002; Harnik and Chang, 2003).

[INSERT FIGURE 3.5.1 HERE]

3.5.3 Storm Tracks

A number of recent studies suggest that cyclone activity over both hemispheres has changed over the second half of the 20th century. General features include a poleward shift in storm track location, increased storm intensity, but a decrease in total storm numbers (e.g., Simmonds and Keay, 2000; Gulev et al., 2001; McCabe et al., 2001). In the NH, McCabe et al. (2001) found that there has been a significant decrease in mid-latitude cyclone activity and an increase in high-latitude cyclone frequency, suggesting a poleward shift of the storm track, with storm intensity increasing over the North Pacific and North Atlantic. In particular, Wang et al. (2006) found that the North Atlantic storm track has shifted about 180 km northward in winter (JFM) during the past half century. The above findings are corroborated by Zhang et al. (2004b), Paciorek et al. (2002), and Simmonds and Keay (2002).

Increases in storm track activity have also been found in eddy variance and covariance statistics, based on the NRA data. North Pacific storm track activity, identified as poleward eddy heat transport at 850 hPa, was significantly stronger during the late 1980s and early 1990s than during the early 1980s (Nakamura et al., 2002). A striking signal of decadal variability in the Pacific storm track activity was its midwinter enhancement since 1987, despite a concurrent weakening of the Pacific jet, concomitant with the sudden weakening of the Siberian High (Nakamura et al., 2002; Chang, 2003). Significant increasing trends over both the Pacific and Atlantic are found in eddy meridional velocity variance at 300 hPa and other eddy statistics (Chang and Fu, 2002; Paciorek et al., 2002). Since 1980 there was an increase in the amount of eddy kinetic energy in the NH due to an increase in the efficiency in the conversion from potential to kinetic energy (Hu et al., 2004). Graham and Diaz (2001) also found an increase in MSLP variance over the Pacific. All these results suggest that cyclone activity in the NH mid-latitudes has increased during the past 40 years.

NRA eddy meridional velocity variance at 300 hPa appears to be biased low prior to the mid-1970s, especially over east Asia and the western United States (Harnik and Chang, 2003). Hence the increases in eddy variance in the NRA reanalysis data are nearly twice as large as that computed from rawinsonde observations. Better agreement is found over the Atlantic storm track exit region over Europe. Major differences between radiosonde and NRA temperature variance at 500 hPa over Asia (Iskenderian and Rosen, 2000; Paciorek et al., 2002) also cast doubts on the magnitude of the increase in storm track activity, especially over the Pacific. Several studies (Bromirski et al., 2003; Chang and Fu, 2003) suggest that storm track activity during the last part of the 20th century may not be more intense than the activity prior to the 1950s.

Station pressure data over the Atlantic-European sector (where records are long and consistent) show a decline of storminess from high levels during the late-19th century to a minimum around 1960 and then a quite rapid increase to a maximum around 1990, followed again by a slight decline (Alexandersson et al., 2000; Bärring and von Storch, 2004; see also Section 3.8.4.1). However, changes in storm tracks are expected to be complex and depend on patterns of variability, and in practice the noise present in the observations makes the detection of long-term changes in extratropical storm activity difficult. A more relevant approach then seems to be the analysis of regional storminess in relation to spatial shifts and strength changes of teleconnections patterns (see Section 3.6).

Significant decreases in cyclone numbers, and increases in mean cyclone radius and depth over the southern extratropics over the last two or three decades (Simmonds and Keay, 2000; Keable et al., 2002; Simmonds, 2003; Simmonds et al., 2003) have been associated with the observed trend in the SAM. This decrease in mid-latitudes derived from NRA data may be related to reductions in winter rainfall from the shift in the storm track (e.g., the drying trend observed in southwestern Australia, Karoly, 2003). However, there are significant differences between ERA-40 and NRA in the SH: higher strong cyclone activity and less weak-cyclone activity over all oceanic areas south of 40°S in all seasons, and stronger cyclone activity over the subtropics in the warm season, in ERA-40, especially in the early decades (Wang et al., 2006).

3.5.4 Blocking

Blocking events, associated with persistent high-latitude ridging and a displacement of mid-latitude westerly winds lasting typically a week or two, are an important component of total circulation variability on intraseasonal time scales. In the NH, the preferred locations for the blocking are over the Atlantic and the Pacific (Tibaldi et al., 1994), with a spring maximum and summer minimum in the Atlantic-European region (Andrea et al., 1998; Trigo et al., 2004). Observations show that in the Euro-Atlantic sector long-lasting (>10 days) blockings are clearly associated with the negative NAO phase (Quadrelli et al., 2001; Barriopedro et al., 2006), whereas the blockings of 5–10 day duration exhibit no such relationship, pointing out to the dynamical links between the life cycles of NAO and blocking events (Scherrer et al, 2006; Schwierz et al., 2006). Wiedenmann et al. (2002) did not find any long-term statistically significant trends in NH blocking intensity. However, in the Pacific sector Barriopedro et al. (2006) found a significant increase from 1948 to 2002 in western Pacific blocking days and events (57% and 62% respectively). They also found less intense North Atlantic region blocking, with statistically significant decreases in events and days. Wiedenmann et al. (2002) found that blocking events, especially in the North Pacific region, were significantly weaker during El Niño years.

In the SH, blocking occurrence is maximised over the southern Pacific (Renwick and Revell, 1999; Renwick, 2005), with secondary blocking regions over the southern Atlantic and over the southern Indian Ocean and the Australian Bight. The frequency of blocking occurrence over the southeast Pacific is strongly ENSO-modulated (Rutllant and Fuenzalida, 1991; Renwick, 1998), while in other regions, much of the interannual variability in occurrence appears to be internally generated (Renwick, 2005). A decreasing trend in blocking frequency and intensity for the SH as a whole from NRA (Wiedenmann et al., 2002) is consistent with observed increases in zonal winds across the southern oceans. However, an overall upward trend in the frequency of long-lived positive height anomalies is evident in the reanalyses over the SH (Renwick, 2005), manifested largely as a step change in the late 1970s, and apparently related to the introduction of satellite observations at that time. Given data limitations, it may be too early to reliably define trends in SH blocking occurrence.

3.5.5 The Stratosphere

The dynamically stable stratospheric circulation is dominated in mid-latitudes by westerlies in the winter hemisphere and easterlies in the summer hemisphere, and the associated meridional overturning “Brewer-Dobson” circulation. In the tropics, zonal winds reverse direction approximately every two years, in the downward-propagating Quasi-Biennial Oscillation (QBO) (Andrews et al., 1987). Ozone is formed predominantly in the tropics and then transported to higher latitudes by the Brewer-Dobson circulation. Climatological stratospheric zonal-mean zonal winds (i.e., the westerly wind averaged over latitude circles) from different datasets show overall good agreement in the extratropics, whereas relatively large differences occur in the tropics (Randel et al., 2004).

The breaking of vertically-propagating waves, originating from the troposphere, decelerates the stratospheric westerlies (see Box 3.3). This sometimes triggers “sudden warmings” when the westerly polar vortex breaks down with an accompanying warming of the polar stratosphere, which can quickly reverse the latitudinal temperature gradient (Kodera et al., 2000). While no major warming occurred in the NH in nine consecutive winters during 1990–1998, seven major warmings occurred during 1999–2004 (Manney et al., 2005). As noted by Naujokat et al. (2002) many of the recent stratospheric warmings after 2000 have been atypically early and the cold vortex recovered in March. In September 2002 a major warming was observed for the first time in the SH (e.g., Krüger et al., 2005; Simmons et al., 2005). This major warming followed a relatively weak polar vortex in winter (Newman and Nash, 2005).

The analysis of past stratospheric changes relies on a combination of radiosonde information (available since the 1950s), satellite information (available from the 1970s), and global reanalyses. During the middle 1990s the NH exhibited a number of years when the Arctic wintertime vortex was colder, stronger (Kodera and Koide, 1997; Pawson and Naujokat, 1999) and more persistent (Waugh et al., 1999; Zhou et al., 2000). Some analyses show a downward trend in the NH wave forcing in the period 1979–2000, particularly in January and February (Newman and Nash, 2000; Randel et al., 2002). Trend calculations are, however, very sensitive to the month and period of calculation, so the detection of long-term change from a relatively short stratospheric data series is still problematic (Labitzke and Kunze, 2005).

In the SH, using radiosonde data, Thompson and Solomon (2002) report a significant decrease of the lower stratospheric geopotential height averaged over the SH polar cap in October–March and May between 1969 and 1998. ERA-40 and NRA stratospheric height reanalyses indicate a trend towards a strengthening Antarctic vortex since 1980 during summer (Renwick, 2004; and Section 3.5.2), largely related to ozone depletion (Ramaswamy et al., 2001; Gillett and Thompson, 2003). The ozone hole has led to a cooling of the stratospheric polar vortex in late spring (Randel and Wu, 1999), and to a 2–3 week delay in vortex breakdown (Waugh et al., 1999).

Box 3.3: Stratospheric-Tropospheric Relations and Downward Propagation

The troposphere influences the stratosphere mainly through planetary-scale waves that propagate upward during the extended winter season when stratospheric winds are westerly. The stratosphere responds to this forcing from below to produce long-lived changes to the strength of the polar vortices. In turn, these fluctuations in the strength of the stratospheric polar vortices are observed to couple downward to surface climate (Baldwin and Dunkerton, 1999, 2001; Kodera et al., 2000; Limpasuvan et al., 2004; Thompson et al., 2005). This relationship occurs in the zonal wind and can be seen clearly in annular modes which explain a large fraction of the intraseasonal and interannual variability in the troposphere (Thompson and Wallace, 2000) and most of the variability in the stratosphere (Baldwin and Dunkerton, 1999). Annular modes appear to arise naturally as a result of internal interactions within the troposphere and stratosphere (Limpasuvan and Hartmann, 2000; Lorenz and Hartmann, 2001; 2003).

The relationship between NAM anomalies in the stratosphere and troposphere can be seen in Figure 3.5.2 in which the NAM index at 10 hPa is used to define events when the stratospheric polar vortex was extremely weak (stratospheric warmings). On average, weak vortex conditions in the stratosphere tend to descend to the troposphere and are followed by negative NAM anomalies at the surface for more than two months. Anomalously strong vortex conditions propagate downwards in a similar way.

Long-lived annular mode anomalies in the lowermost stratosphere appear to lengthen the time scale of the surface NAM. The tropospheric annular mode timescale is longest during winter in the NH, but during late spring (November–December) in the SH (Baldwin et al., 2003). In both hemispheres the time scale of the tropospheric annular modes is longest when the variance of the annular modes is greatest in the lower stratosphere.

Downward coupling to the surface depends on having large circulation anomalies in the lowermost stratosphere. In such cases, the stratosphere can be used as a statistical predictor of the monthly-mean surface NAM on timescales of up to two months (Baldwin et al., 2003). Similarly, SH trends in temperature and geopotential height, associated with the ozone hole, appear to couple downward to affect high-latitude surface climate (Thompson and Solomon, 2002; Gillett and Thompson, 2003). As the stratospheric circulation changes with ozone depletion or increasing greenhouse gases, those changes will likely be reflected in changes to surface climate. Thompson and Solomon (2005) show that the springtime strengthening and cooling of the SH polar stratospheric vortex precedes similarly-signed trends in the SH tropospheric circulation by one month in the interval 1973–2003. They argue that similar downward coupling is not evident in the NH geopotential trends computed using monthly radiosonde data. An explanation for this difference may be that the stratospheric signal is stronger in the SH, mainly due to ozone depletion, giving a more robust downward coupling.

The dynamical mechanisms by which the stratosphere influences the troposphere are not well understood, but the relatively large surface signal implies that the stratospheric signal is amplified. The processes likely involve planetary waves (Song and Robinson, 2004) and synoptic-scale waves (Wittman et al., 2004), which interact with stratospheric zonal wind anomalies near the tropopause. The altered waves would be expected to affect tropospheric circulation and induce surface pressure changes corresponding to the annular modes (Wittman et al., 2004).

[INSERT FIGURE 3.5.2 HERE]

3.5.6 Winds, Waves and Surface Fluxes

Changes in atmospheric circulation imply associated changes in the winds, wind waves and surface fluxes. Meteorological observations, including surface winds, from Voluntary Observing Ships (VOS) became systematic around 150 years ago and are assembled in ICOADS (International Comprehensive Ocean-Atmosphere Data Set) (Worley et al., 2005). Apparent significant trends in scalar wind should be considered with caution because VOS wind observations are influenced by time-dependent biases, arising from the rising proportion of anemometer measurements, increasing anemometer heights, changes in definitions of Beaufort wind estimates (Cardone et al., 1990), growing ship size, inappropriate evaluation of the true wind speed from the relative wind (Gulev and Hasse, 1999) and time-dependent sampling biases (Sterl, 2001). Because ICOADS winds assimilated into reanalyses, these too will suffer biases. Consideration of the local surface pressure gradient time series (Ward and Hoskins, 1996) does not support the existence of the globally averaged trends in wind speeds, but reveals regional patterns of the upward trends in the tropical North Atlantic and extratropical North Pacific and downward trends in the tropical South Atlantic and subtropical North Pacific (see also Sections 3.5.1 and 3.5.3).

In contrast to marine winds, visual VOS observations of wind waves for more than a century, often measured as significant wave height (SWH, the highest one-third of wave (sea and swell) heights), have been less affected by changes in observational practice, although they may suffer from time-dependent sampling uncertainty, which was somewhat higher at the beginning of the record. Local wind speed directly affects only the wind sea component of SWH, while the swell component is largely influenced by the frequency and intensity of remote storms. Linear trends in the annual SWH from ship data (Gulev and Grigorieva, 2004) for 1900–2002 were significantly positive almost everywhere in the North Pacific, with a maximum upward trend of 8–10 cm decade^{-1} (up to 0.5% per year). These are supported by buoy records for 1978–1999 (Allan and Komar, 2000; Gower, 2002) for annual mean and winter (October to March) SWH and confirmed by the long-term estimates of storminess derived from the tide gauge residuals (Bromirski et al., 2003) and hindcast data (Graham and Diaz, 2001), although Tuller (2004) found primarily negative trends in wind off the west coast of Canada. In the Atlantic, centennial time series (Gulev and Grigorieva, 2004) show weak but statistically significant negative trends along the North Atlantic storm track, with a decrease of –5.2 cm decade^{-1} (0.25% per year) in the western Atlantic storm formation region. Regional model hindcasts (e.g., Vikebo et al., 2003; Weisse et al., 2005) show growing SWH in the northern North Atlantic over the last 118 years.

Linear trends for the period 1958–2002 (Figure 3.5.3) are statistically significant and positive over most of the mid-latitudinal North Atlantic and North Pacific, as well as in the western subtropical South Atlantic, the eastern equatorial Indian Ocean and the East China and South China seas. The largest upward trends of 14 cm decade^{-1} occur in the northwest Atlantic and the northeast Pacific. Statistically significant negative trends are observed in the western Pacific tropics, the eastern Indian Ocean, in the Tasman Sea, and in the south Indian Ocean (–11 cm decade^{-1}). Global and basin-scale model wave hindcasts of Wang and Swail (2001, 2002) and Sterl and Caires (2005), based respectively on NRA and ERA-40 winds, show an increasing mean SWH as well as intensification of SWH extremes during the last 40 years, with the 99% extreme of the winter SWH increased in the northeast Atlantic by a maximum of 0.4 m per decade. Wave height hindcasts driven with NRA surface winds shows that worsening wave conditions in the northeastern North Atlantic during the latter 20th Century were connected to a northward displacement in storm track, while the effect reduced wave heights in the central North Atlantic (Lozano and Swail, 2002). Increases of SWH in the North Atlantic mid-latitudes are further supported by a 14-year (1988–2002) time series of the merged TOPEX/Poseidon and ERS-1/2 altimeter data (Woolf et al., 2002).

[INSERT FIGURE 3.5.3 HERE]

Since the TAR, research into surface fluxes has continued to be directed at improving the accuracy of the mean air-sea exchange fields (particularly of heat) with less work on long-term trends. Significant uncertainties remain in global fields of the net heat exchange, stemming from problems in obtaining accurate estimates of the different heat flux components. Estimates of surface flux variability from reanalyses are strongly influenced by inhomogeneous data assimilation input, especially in the Southern Ocean. Sterl (2004) reported that variability of the surface latent flux in the Southern Ocean becomes much more reliable after 1979, when observations increased. Recent evaluations of heat flux estimates from reanalyses and *in situ* observations indicate some improvements but there are still global biases of several tens of W m^{-2} in unconstrained VOS observation-based products (Grist and Josey, 2003). Estimates of the implied ocean heat transport from the NRA, indirect residual techniques and some coupled models are in reasonable agreement with hydrographic observations (Trenberth and Caron, 2001; Grist and Josey, 2003). However, it should also be noted that the hydrographic observations contain significant uncertainties (see Chapter 5) due to both interannual variability and assumptions made in the computation of the heat transport, and these must be recognised when using them to evaluate the various flux products. For the North Atlantic, there are indications of positive trends in the net heat flux of 10 W m^{-2} $decade^{-1}$ in the eastern sub-polar gyre and coherent negative changes up to 15 W m^{-2} in the western subtropical gyre, closely correlated with the NAO variability (Marshall et al., 2001; Visbeck et al., 2003).

3.5.7 Summary

Changes from the late 1970s to the present generally reveal decreases of tropospheric geopotential heights over high latitudes of both hemispheres and increases over the mid-latitudes in DJF. The changes amplify with altitude up to the lower stratosphere, but remain similar in shape to lower atmospheric levels and are associated with the intensification and poleward displacement of corresponding Atlantic and southern polar front jet streams and enhanced storm track activity. Based on a variety of measures at the surface and in the upper troposphere, it is likely that there has been an increase and a poleward shift in NH winter storm track activity over the second half of the 20th century, but there are still significant uncertainties in the magnitude of the increase due to time-dependent biases in the reanalyses. Analysed decreases in cyclone numbers over the southern extratropics and increases in mean cyclone radius and depth over much of the SH over the last two decades are subject to even larger uncertainties. After the late 1990s in the NH, however, occurences of major sudden warmings seem to have increased in the polar stratosphere, associated with the occurrence of more neutral states of the tropospheric and stratospheric vortex. In the SH, there has been a strengthening tropospheric Antarctic vortex during summer in association with the ozone hole, which has led to a cooling of the stratospheric polar vortex in late spring and to a 2–3 week delay in vortex breakdown. In September 2002, a major warming was observed for the first and only time in the SH. The decreases in long-lasting blocking frequency over the North Atlantic–European sector over recent decades is dynamically consistent with NAO variability (see Section 3.6), but given data limitations, it may be too early to define the nature of any trends in SH blocking occurrence, despite observed trends in the SAM. Analysis of observed wind and SWH support the reanalysis-based evidence for an increase in storm activity in the extratropical NH in recent decades (see also Section 3.6) until the late 1990s. For heat flux there seems to have been a significant reduction in NAO-related heat loss over the Labrador Sea, which is a key region for deep water formation. In the eastern North Pacific it is likely that ocean heat loss to the atmosphere has decreased since 1977 due to the shift and variations in the strength of the Aleutian low.

3.6 Patterns of Circulation Variability

3.6.1 Teleconnections

The global atmospheric circulation has a number of preferred patterns of variability, all of which have expressions in surface climate. Box 3.4 discusses the main patterns and associated indices. Regional climates in different locations may vary out of phase, owing to the action of such “teleconnections” which modulate the location and strength of the storm tracks (Section 3.5.3), and poleward fluxes of heat, moisture and momentum. A comprehensive review (Hurrell et al., 2003) has been updated by new analyses, notably from Quadrelli and Wallace (2004) and Trenberth et al. (2005b). Understanding the nature of teleconnections and changes in their behaviour is central to understanding regional climate variability and change. Such seasonal and longer time-scale anomalies have direct human impacts, often being associated with droughts, floods, heat waves and other changes that can severely disrupt agriculture, water supply, and fisheries, and can modulate air quality, fire risk, energy demand and supply, and human health.

The analysis of teleconnections has typically employed a linear perspective, which assumes a basic spatial pattern with continuously varying amplitude that has mirror image positive and negative polarities (Hurrell et al., 2003; Quadrelli and Wallace, 2004). In contrast, nonlinear interpretations of atmospheric variability have also found applications within the climate framework (e.g., Palmer, 1999; Corti et al., 1999; Cassou and Terray, 2001; Monahan et al., 2001). From this perspective, preferred climate anomalies are identified as recurrent states of a specific amplitude and sign. Climate change may result through changes from one quasi-stationary state to another, as a preference for one sign of a pattern, or through a change in the nature or number or states (Palmer, 1999).

In the NH, one point correlation maps illustrate the PNA and NAO (Figure 3.6.1) but in the SH, wave structures do not emerge as readily owing to the dominance of the SAM. Although teleconnections are best defined over a grid, simple indices based on a few key station locations remain attractive as the series can often be carried back in time long before complete gridded fields are available (Figure 3.6.6); the disadvantage is increased noise from the reduced spatial sampling. For instance, Hurrell et al. (2003) find that the residence time of the NAO in its positive phase in the early 20th century is not as great as might be expected from the positive NAO index then.

Many teleconnections have been identified, but combinations of only a small number of patterns can account for much of the interannual variability in the circulation and surface climate. Quadrelli and Wallace (2004) found that many patterns of NH interannual variability can be reconstructed by simply rotating the first two EOFs of sea level pressure (approximately the NAM and the PNA). Trenberth et al. (2005b) analysed global atmospheric mass and found four key rotated EOF patterns; the two annular modes (SAM and NAM), a global ENSO-related pattern, and a fourth closely related to the North Pacific Index and the Pacific Decadal Oscillation, that in turn is closely related to ENSO and the PNA pattern.

All teleconnection patterns are affected by time of year, especially in the NH, tending to be most prominent in the wintertime, when the mean circulation is strongest. The strength of teleconnections, and the way they influence surface climate, has also varied over long time scales. For example, both the NAO and ENSO exhibit marked changes in their surface climate expressions over multi-decadal time scales during the 20th century (e.g., Jones et al., 2003; Power et al., 1999b). Hence multi-decadal changes of influence are often real and not just due to poorer data quality in earlier decades.

[INSERT FIGURE 3.6.1 HERE]

Box 3.4: Defining the Circulation Indices

An atmospheric teleconnection is made up of a fixed spatial pattern with an associated index time series showing the evolution of its amplitude and phase. Teleconnections are best defined by values over a grid, but it has generally been convenient to devise simplified indices based on key station values. A classic example is the Southern Oscillation (SO), encompassing the entire tropical Pacific, yet encapsulated by a simple SO Index (SOI), based on differences between Tahiti (eastern Pacific) and Darwin (western Pacific) MSLP anomalies. Using gridded fields to define indices provides a fuller picture of the true magnitude of fluctuations in a teleconnection pattern and reduces short term “noise”. However, an index defined in this way is more complicated to calculate, and relies on the existence of gridded data fields.

A number of teleconnections have historically been defined from either station data (SOI, NAO) or from gridded fields (NAM, SAM, PDO/NPI and PNA):

- **Southern Oscillation Index (SOI).** The MSLP anomaly difference Tahiti minus Darwin, normalised by the long-term mean and standard deviation of the MSLP difference (Troup, 1965; Trenberth, 1984; Können et al., 1998). Available from the 1860s. Darwin can also be used alone, as its data are more consistent than Tahiti prior to 1935.
- **North Atlantic Oscillation (NAO) Index.** The difference of normalized MSLP anomalies between Lisbon, Portugal and Stykkisholmur, Iceland has become the widest used NAO index and extends back in time to 1864 (Hurrell, 1995), and to 1821 if Reykjavik is used instead of Stykkisholmur and Gibraltar instead of Lisbon (Jones et al., 1997). When originally defined in the 1930s, Ponta Delgada, Azores and Stykkisholmur, Iceland were used and the series extended back to 1865, but this series is less easily updatable in real time.
- **Northern Annular Mode (NAM) Index.** The amplitude of the pattern defined by the leading empirical orthogonal function of winter monthly mean NH MSLP anomalies poleward of 20ºN (Thompson and Wallace, 1998, 2000). The NAM has also been known as the Arctic Oscillation (AO), and is closely related to the NAO.
- **Southern Annular Mode (SAM) Index.** The difference in average MSLP between Southern Hemisphere middle and high latitudes (usually 45°S and 65°S), either from gridded or station data (Gong and Wang, 1999; Marshall, 2003), or the amplitude of the leading empirical orthogonal function of monthly mean SH 850 hPa height poleward of 20°S (Thompson and Wallace, 2000). Formerly known as the Antarctic Oscillation (AAO) or High Latitude Mode (HLM).

- **Pacific-North American pattern (PNA) Index.** The mean of normalised 500 hPa height anomalies at 20°N, 160°W and 55°N, 115°W minus those at 45°N, 165°W and 30°N, 85°W (Wallace and Gutzler, 1981).
- **Pacific Decadal Oscillation (PDO) Index and North Pacific Index (NPI).** The NPI is the average MSLP anomaly in the Aleutian Low (AL) over the Gulf of Alaska for the region 30°N–65°N, 160°W–140°W from Trenberth and Hurrell (1994) and is an index of the PDO, which is also defined as the pattern and time series of the first empirical orthogonal function of SST over the North Pacific north of 20ºN (Mantua et al., 1997), see also Deser et al. (2004). The PDO has also been broadened to cover the whole Pacific Basin, as the Interdecadal Pacific Oscillation (IPO, Power et al, 1999b). The PDO and IPO exhibit virtually identical temporal evolution (Folland et al., 2002).

3.6.2 El Niño-Southern Oscillation and Tropical/Extra-tropical Interactions

3.6.2.1 El Niño-Southern Oscillation

El Niño-Southern Oscillation events are a coupled ocean-atmosphere phenomenon. El Niño involves warming of surface waters of the tropical Pacific in the region from the International Date Line to the west coast of South America, and associated changes in oceanic circulation. Its closely linked atmospheric counterpart, the Southern Oscillation (SO), involves changes in trade winds and associated tropical circulation. The total phenomenon is generally referred to as ENSO. El Niño is the warm phase of ENSO and La Niña is the cold phase. Historically, El Niño (EN) events occur about every 3–7 years and alternate with the opposite phases of below average temperatures in the tropical Pacific (La Niña). The strong SST gradient from the warm pool in the western tropical Pacific to the cold tongue in the eastern equatorial Pacific is maintained by westward-flowing trade winds, which drive the surface ocean currents and determine the pattern of upwelling of cold nutrient-rich waters in the east. The trade winds and atmospheric circulation organize tropical Pacific precipitation and associated atmospheric heating patterns through the release of latent heat that in turn set up teleconnections and determine the surface winds. Extratropical teleconnections are partly characterised by the PNA pattern for the NH (or a variant of that, Straus and Shukla, 2002), and the Pacific-South American (PSA) pattern in the SH (Renwick and Revell, 1999; Mo, 2000), and accompanied by changes to the jet stream and storm tracks in mid-latitudes (Chang and Fu, 2002).

ENSO has global impacts, manifested most strongly in the winter months in either hemisphere. MSLP anomalies are much greater in the extratropics while the tropics feature large precipitation variations. Associated patterns of surface temperature and precipitation anomalies around the globe are given in Trenberth and Caron (2000; see Figure 3.6.2), and the evolution of these patterns and links to global mean temperature perturbations is given by Trenberth et al. (2002b).

[INSERT FIGURE 3.6.2 HERE]

The nature of ENSO has varied considerably over time. Strong ENSO events occurred from the late 19th Century through the first 25 years of the 20th century and again after about 1950, but there were few events of note from 1925 to 1950 with the exception of the major 1939–1941 event (Figure 3.6.2). The climate shift in 1976–1977 (Trenberth, 1990) (see Figures 3.6.2 and 3.6.3) was associated with marked changes in El Niño evolution (Trenberth and Stepaniak, 2001) and with a shift to generally above normal SSTs in the eastern and central Pacific along the equator since then, i.e. more El Niños.

Since the TAR, there has been considerable work on decadal and longer-term variability of ENSO and Pacific climate. Such decadal atmospheric and oceanic variations (see Section 3.6.3) are even more pronounced in the North Pacific and across North America than in the tropics and are also present in the South Pacific, with evidence suggesting they are at least in part forced from the tropics (Deser et al., 2004). Because ENSO is involved in moving heat around the tropical Pacific and between the atmosphere and ocean, with typical exchanges of order 50 W m^{-2} over the central tropical Pacific (Trenberth et al., 2002a), it is likely that global climate change will interfere and alter El Niño, just as El Niño changes global mean temperatures. The 1997–1998 event was the largest on record in terms of SST anomalies and 1998 was the warmest year for the global mean. Trenberth et al. (2002b) estimate that global mean surface air temperatures were 0.17°C higher for the year centred on March 1998 owing to the El Niño. Also, extremes of the hydrological cycle such as floods and droughts are common with ENSO and are apt to be enhanced with global warming (Trenberth et al., 2003). For example, the modest El Niño of 2001–2002 was associated with a drought in Australia, made much worse by record-breaking heat (Nicholls, 2004; and see Box 3.6.2). Thus whether observed changes in ENSO are physically linked to global climate change is a research question of great importance.

3.6.2.2 Tropical-extratropical teleconnections: PNA and PSA

Circulation variability over the extratropical Pacific features wave-like anomaly patterns emanating from the subtropical western Pacific, characteristic of Rossby wave propagation associated with anomalous tropical heating (Horel and Wallace, 1981; Hoskins and Karoly, 1981). These are known as the PNA and PSA patterns and can arise through atmospheric dynamics as well as in response to heating. Over the NH in winter, the PNA pattern lies across North America from the subtropical Pacific, with four centres of action (Figure 3.6.1). While the PNA can be illustrated by taking a single point correlation (Figure 3.6.1), this is not so easy for the PSA (not shown), as its spatial centres of action are not fixed. However, the PSA pattern is present at all times of year, lying from Australasia over the southern Pacific and Atlantic (Kiladis and Mo, 1998; Mo and Higgins, 1998; Kidson, 1999; Mo, 2000).

The positive PNA, or a variant of that (Straus and Shukla, 2002), is associated with an enhanced Aleutian Low, a strengthened and extended Asian jet, and a tendency for the Pacific storm track to extend farther east and equatorward, resulting in enhanced precipitation in California and relatively dry, warm conditions over the northwest U.S. and southwestern Canada. During the negative PNA, the Pacific storm track curves north, favouring wintertime blocking events over the Alaskan region, and an increased frequency of cold air outbreaks over the western United States (Compo and Sardeshmukh, 2004). The PSA is associated with modulation of the westerlies over the South Pacific, effects of which include significant rainfall variations over New Zealand, changes in the nature and frequency of blocking events across the high latitude South Pacific, and interannual variations in Antarctic sea ice across the Pacific and Atlantic sectors (Renwick and Revell, 1999; Kwok and Comiso, 2002a; Renwick, 2002). While both PNA and PSA activity has varied with decadal modulation of ENSO, no systematic changes in their behaviour have been reported.

3.6.3 Decadal Pacific Variability

Decadal-to-interdecadal variability of the atmospheric circulation is most prominent in the North Pacific, where fluctuations in the strength of the wintertime Aleutian Low (AL) pressure system co-vary with North Pacific SST in the PDO. These are linked to decadal variations in atmospheric circulation, SST and ocean circulation throughout the whole Pacific Basin in the IPO (Trenberth and Hurrell, 1994; Gershunov and Barnett, 1998; Folland et al., 2002; McPhaden and Zhang, 2002; Deser et al., 2004). Key measures of Pacific decadal variability are the NPI, (Trenberth and Hurrell, 1994), PDO index (Mantua et al., 1997) and the IPO index (Power et al., 1999b; Folland et al., 2002); see Figures 3.6.3 and 3.6.4.

The PDO/IPO has been described by some as a long-lived El Niño-like pattern of Indo-Pacific climate variability (Knutson and Manabe, 1998; Evans et al., 2001; Deser et al., 2004; Linsley et al., 2004) or as a low frequency residual of ENSO variability on multidecadal time scales (Newman et al., 2003). Indeed, the symmetry of the SST anomaly pattern between the NH and SH may be a reflection of the common tropical forcing. However, Folland et al. (2002) showed that the IPO significantly affects the movement of the South Pacific Convergence Zone in a way independent of ENSO (see also Deser et al., 2004). Other results indicate the extratropical phenomena are generic components of the PDO (Deser et al., 1996; 1999; 2003; Gu and Philander, 1997). The extratropics may also contribute to the tropical SST changes via an “atmospheric bridge”, confounding the simple interpretation of a tropical origin (Barnett et al., 1999; Vimont et al., 2001).

The interdecadal timescale of tropical Indo-Pacific SST variability is likely due to oceanic processes. Extratropical ocean influences are also likely to play a role as changes in the ocean gyre evolve and heat anomalies are subducted and re-emerge (Deser et al., 1996; 1999; 2003; Gu and Philander, 1997). There is also the possibility that there is no well-defined coupled ocean-atmosphere “mode” of variability in the Pacific on decadal-to-interdecadal time scales, since instrumental records are too short to provide a robust assessment and paleoclimate records conflict regarding time scales (Biondi et al., 2001; Gedalof et al., 2002). Schneider and Cornuelle (2005) suggest that the PDO is not itself a mode of variability but is a blend of three phenomena. They showed that the observed PDO pattern and evolution can be recovered from a reconstruction of North Pacific SST anomalies based on a first-order autoregressive model and forcing by variability of the Aleutian low, ENSO, and oceanic zonal advection in the Kuroshio–Oyashio Extension. The latter results from oceanic Rossby waves that are forced by North Pacific Ekman pumping. The SST response patterns to these processes are not completely independent, but they determine the spatial characteristics of the PDO. Under this hypothesis, the key physical variables for measuring Pacific climate variability are ENSO and NPI (Aleutian Low) indices, rather than the PDO index.

Figure 3.6.4a shows a time series of the NPI for 1900–2005 (Deser et al., 2004). There is substantial low-frequency variability, with extended periods of predominantly high values indicative of a weakened circulation (1900 to 1924 and 1947 to 1976) and predominantly low values indicative of a strengthened circulation (1925 to 1946 and 1977 to 2005). The well-known decrease in pressure from 1976 to 1977 is analogous to transitions that occurred from 1946 to 1947 and from 1924 to 1925, and these earlier changes were also associated with SST fluctuations in the tropical Indian (Figure 3.6.4b) and Pacific Oceans although not in the upwelling zone of the equatorial eastern Pacific (Minobe, 1997; Deser et al., 2004). In addition the NPI exhibits variability on shorter time scales, interpreted in part as a bi-decadal rhythm (Minobe, 1999).

There is observational and modelling evidence (Pierce, 2001; Schneider and Cornuelle, 2005) suggesting the PDO/IPO does not excite the climate shifts in the Pacific area, but they share the same forcing. The 1976–1977 climate shift in the Pacific, associated with a phase change in the PDO from negative to positive, was associated with significant changes in ENSO evolution (Trenberth and Stepaniak, 2001) and with changes in ENSO teleconnections and links to precipitation and surface temperatures over North and South America, Asia, and Australia (Trenberth, 1990; Trenberth and Hurrell, 1994; Power et al., 1999a; Salinger et al., 2001; Mantua and Hare, 2002; Minobe and Nakanowatari, 2002; Trenberth et al., 2002b; Deser et al., 2004; Marengo, 2004). Schneider and Cornuelle (2005) add extra credence to the hypothesis that the 1976-1977 climate shift is of tropical origin.

[INSERT FIGURE 3.6.3 HERE]

[INSERT FIGURE 3.6.4 HERE]

3.6.4 The North Atlantic Oscillation (NAO) and Northern Annular Mode (NAM)

The only teleconnection pattern prominent throughout the year in the NH is the NAO (Barnston and Livezey, 1987). It is primarily a north-south dipole in sea level pressure characterized by simultaneous out-of-phase pressure and height anomalies between temperate and high latitudes over the Atlantic sector, and therefore corresponds to changes in the westerlies across the North Atlantic into Europe (Figure 3.6.5). The NAO has the strongest signature in the cold-season months (December to March) when its positive (negative) phase exhibits an enhanced (diminished) Iceland Low and/or Azores High (Hurrell et al., 2003). The NAO is the dominant pattern of atmospheric circulation variability over the North Atlantic, accounting for one-third of the total variance in monthly MSLP in winter. It is closely related to the NAM that has structure similar to the NAO over the Atlantic, but is more zonally symmetric. The leading wintertime pattern of variability in the lower stratosphere is also annular, but the MSLP anomaly pattern that is associated with it is confined almost entirely to the Arctic and Atlantic sectors and coincides with the spatial structure of the NAO (Deser, 2000; see also Section 3.5.5 and Box 3.3).

There is considerable debate over whether the NAO or the NAM is more physically relevant to the wintertime circulation (Deser, 2000; Ambaum et al., 2001; 2002), but the time series are highly correlated in winter (Figure 3.6.6). As Quadrelli and Wallace (2004) show, they are near neighbours in terms of their spatial patterns, and their temporal evolution. The annular modes appear to occur as a result of interactions between the eddies and the mean flow and external forcing is not required to sustain them (De Weaver and Nigam, 2000). In the NH, stationary waves provide most of the eddy momentum fluxes, although synoptic transient eddies are also important. As the intrinsic excitation of NAO/NAM pattern is limited to a period less than a few days (Feldstein, 2002), it should not exhibit year-to-year autocorrelation in conditions of constant forcing. Proxy and instrumental data, however, show evidence for intervals with prolonged positive and negative NAO index in the last few centuries (Cook et al., 2002; Jones et al., 2003). In winter, a reversal occurred from the minimum index values in the late 1960s to strongly positive NAO index values in the mid-1990s. Since then NAO values have declined to near the long-term mean (Figure 3.6.6).

For summer, Hurrell et al. (2001, 2002) identified significant interannual to multi-decadal fluctuations in the NAO pattern and the trend toward persistent anticyclonic flow over northern Europe has contributed to anomalously warm and dry conditions in recent decades (Rodwell, 2003).

Feldstein (2002) suggested that the trend and increase in the variance of NAO/NAM index during 1968–1997 was greater than would be expected from internal variability alone, while NAO behaviour during the first 60 years of the 20th century was consistent with atmospheric internal variability, although results are not so clear if based on just 1975–2004 (Overland and Wang, 2005). Although monthly-scale NAO variability appears to be largely unpredictable, and internal variability is strong (Czaja et al., 2003; Thompson et al., 2003), there may be predictability from stratospheric influences (Thompson et al., 2002; Scaife et al., 2005; see Box 3.3). There is mounting evidence that the recent observed interdecadal NAO variability comes from tropical ocean influences (Hurrell et al., 2004), land surface forcing (Gong et al., 2003; Bojariu and Gimeno, 2003) and from other external factors.

The NAO exerts a dominant influence on wintertime surface temperatures across much of the NH (Figure 3.6.5), and on storminess and precipitation over Europe and North Africa. When the NAO index is positive, enhanced westerly flow across the North Atlantic in winter moves warm moist maritime air over much of Europe and far downstream, with dry conditions over southern Europe and northern Africa and wet conditions in northern Europe, while stronger northerly winds over Greenland and northeastern Canada carry cold air southward and decrease land temperatures and SST over the northwest Atlantic. Temperature variations over North Africa and the Middle East (cooling), as well as the southeastern United States (warming), associated with the stronger clockwise flow around the subtropical Atlantic high-pressure centre are also notable. Following on from Hurrell (1996), Thompson et al. (2000) showed that for JFM over 1968–97, the NAM accounted for 1.6°C out of 3.0°C warming in Eurasian surface temperatures, 4.9 out of the 5.7 hPa decrease in sea level pressure from 60°N–90°N; 37% out of the 45% increase in Norwegian precipitation (55°N–60°N, 5°E–10°E), and 33% out of the 49% decrease in Spanish rainfall (35°N–45°N, 10°W–0°W). There were also significant effects on ocean heat content, sea ice, ocean currents and ocean heat transport.

[INSERT FIGURE 3.6.5 HERE]

[INSERT FIGURE 3.6.6 HERE]

Positive NAO index winters are associated with a northeastward shift in the Atlantic storm activity, with enhanced activity from Newfoundland into northern Europe and a modest decrease to the south (Hurrell and van Loon, 1997; Alexandersson et al., 1998). Positive NAO index winters are also typified by more intense and frequent storms in the vicinity of Iceland and the Norwegian Sea (Serreze et al., 1997; Deser et al., 2000). The upward trend toward more positive NAO index winters from the mid-1960s to the mid-1990s has been associated with increased wave heights over the northeast Atlantic and decreased wave heights south of 40°N (Carter, 1999); see also Section 3.5.6.

The NAO/NAM modulates the transport and convergence of atmospheric moisture and the distribution of evaporation (E) and precipitation (P), (Dickson et al., 2000). E exceeds P over much of Greenland and the Canadian Arctic and more precipitation than normal falls from Iceland through Scandinavia during high NAO index winters, while the reverse occurs over much of central and southern Europe, the Mediterranean and parts of the Middle East (Dickson et al., 2000). Severe drought has persisted throughout parts of Spain and Portugal as well (Hurrell et al., 2003). As far eastward as Turkey, river runoff is significantly correlated with NAO variability (Cullen and deMenocal, 2000). There are many NAO-related effects in ocean circulation, such as the freshwater balance of the Atlantic Ocean (see Chapter 5), the cryosphere (see Chapter 4), and in many aspects of the north Atlantic/European biosphere (see WGII report).

3.6.5 The Southern Hemisphere and Southern Annular Mode (SAM)

The principal mode of variability of the atmospheric circulation in the SH extratropics is now known as the SAM, see Figure 3.6.7. It is essentially a zonally-symmetric structure, but with a zonal wave number three pattern superimposed. It is associated with synchronous pressure or height anomalies of opposite sign in mid- and high-latitudes, and therefore reflects changes in the main belt of sub-polar westerly winds. When pressures are below average over Antarctica and westerly winds are enhanced over the southern oceans, the SAM is said to be in its high index or positive phase. The SAM appears as the leading SH EOF in many atmospheric fields at many levels (Thompson and Wallace, 2000; Trenberth et al., 2005b). As for the NAM, the structure and variability of the SAM results mainly from the internal dynamics of the atmosphere (e.g., Hartmann and Lo, 1998; Limpasuvan and Hartmann, 2000; and Box 3.3), although with ozone depletion also playing a role (Sexton, 2001; Thompson and Solomon, 2002). Poleward eddy momentum fluxes interact with the zonal mean flow to sustain latitudinal displacements of the mid-latitude westerlies (Limpasuvan and Hartmann, 2000; Rashid and Simmonds, 2004; 2005). The SAM contributes a significant proportion of SH mid-latitude circulation variability on many time scales (Hartmann and Lo, 1998; Kidson, 1999; Baldwin, 2001). Trenberth et al. (2005b) show that the SAM is the leading mode in an EOF analysis of monthly mean global atmospheric mass, accounting for around 10% of total global variance.

Gridded reanalysis datasets have been utilised to derive time series of the SAM, particularly the NRA (e.g., Gong and Wang, 1999; Thompson et al., 2000) and more recently ERA-40 (Renwick, 2004; Trenberth et al., 2005b). However, a declining positive bias in pressure at high southern latitudes in both reanalyses before 1979 (Hines et al., 2000; Trenberth and Smith, 2005) means that derived trends in the SAM are too strong. Marshall (2003) produced a SAM index based on appropriately-located station observations. His index revealed a general positive increase in the SAM beginning in the 1960s (Figure 3.6.7) resulting in a strengthening of the circumpolar vortex and intensification of the circumpolar westerlies, as observed in northern Antarctic Peninsula radiosonde data (Marshall, 2002).

[INSERT FIGURE 3.6.7 HERE]

The trend in the SAM, which is statistically significant annually, and in summer and autumn (Marshall et al., 2004), has contributed to Antarctic temperature trends (Thompson and Solomon, 2002; Kwok and Comiso, 2002b; van den Broeke and van Lipzig, 2003; Schneider et al., 2004); specifically a strong summer warming in the Peninsula region and a cooling over much of the rest of the continent (Turner et al., 2005), see Figure 3.6.7. Through the wave component, the positive SAM is associated with low pressure west of the Peninsula (e.g., Lefebvre et al., 2004) leading to increased northerly flow, warming, and reduced sea ice in the region (Liu et al., 2004b). Orr et al. (2004) propose that this scenario yields a higher frequency of warmer maritime air masses passing over the Peninsula, leading to the marked north-east Peninsula warming observed in autumn and summer. The positive trend in the SAM has led to more cyclones in the circumpolar trough (Sinclair et al., 1997) and hence a greater contribution to Antarctic precipitation from these near-coastal systems that is reflected in $\delta^{18}O$ levels in the snow (Noone and Simmonds, 2002). The SAM also impacts the spatial patterns of precipitation variability in Antarctica (Genthon et al., 2003) and southern South America (Silvestri and Vera, 2003). Jones and Widmann (2004) reconstruct century-scale records based on proxies of SAM which indicates that the recent trend is not unprecedented, even during the 20th century, although some studies suggest otherwise (Gillett et al., 2003; Marshall et al., 2004).

The imprint of SAM variability on the Southern Ocean system is observed as a coherent sea level response around Antarctica (Aoki, 2002; Hughes et al., 2003) and by its regulation of Antarctic Circumpolar Current flow through the Drake Passage (Meredith et al., 2004). Changes in oceanic circulation impact directly on the thermohaline circulation (Oke and England, 2004) and may explain recent patterns of observed temperature change at SH high latitudes described by Gille (2002). Diminished summer sea ice may in turn feed back into a more positive SAM (Raphael, 2003).

3.6.6 Other Indices

As noted earlier, many patterns of variability (sometimes mislabelled as “modes”) in the climate system have been identified over the years, but relatively few stand out as robust and dynamically significant features in relation to understanding regional climate change. This section discusses three climate signals that have recently drawn the attention of scientific community: the Atlantic Multi-decadal Oscillation, the Antarctic Circumpolar Wave, and the Indian Ocean Dipole.

3.6.6.1 Atlantic Multi-decadal Oscillation (AMO)

Over the instrumental period (since the 1850s) North Atlantic SSTs show a 65–75 year variation (0.4°C range), with apparent warm phases at roughly 1860–1880 and 1930–1960 and cool phases during 1905–1925 and 1970–1990 (Schlesinger and Ramankutty, 1994), and this feature has been termed the AMO (Kerr, 2000), as shown in Figure 3.6.8. The cycle appears to have returned to a warm phase beginning in the mid-1990s and tropical Atlantic SSTs were at record high levels in 2005. Instrumental observations capture only two full cycles of the AMO, so the robustness of the signal has been addressed using proxies. Similar oscillations in a 60–110 year band are seen in North Atlantic paleoclimatic reconstructions through the last four centuries (Delworth and Mann, 2000; Gray et al., 2004). Both observations and model simulations implicate changes in the strength of the thermohaline circulation as the primary source of the multi-decadal variability, and suggest a possible oscillatory component to its behaviour (Delworth and Mann, 2000; Latif, 2001; Sutton and Hodson, 2003; Knight et al., 2005).

The AMO has been linked to multi-year precipitation anomalies over North America, and appears to modulate ENSO teleconnections (Enfield et al., 2001; Shabbar and Skinner, 2004; McCabe et al., 2004). Multi-decadal variability in the North Atlantic also plays a role in Atlantic hurricane formation (see Section 3.8.3.2), African drought frequency, winter temperatures in Europe, sea ice concentration in the Greenland Sea and sea level pressure over high northern latitudes (e.g., Venegas and Mysak, 2000; Goldenberg et al., 2001). Walter and Graf (2002) identified a non-stationary relationship between the NAO and the AMO. During the negative phase of the AMO, the North Atlantic SST is strongly correlated to the NAO index. In contrast, the NAO index is only weakly correlated to the North Atlantic SST during the AMO positive phase. Chelliah and Bell (2004) defined a tropical multi-decadal pattern related to the AMO, the PDO and wintertime NAO with coherent variations in tropical convection and surface temperatures in the West African monsoon region, the central tropical Pacific, the Amazon basin, and the tropical Indian Ocean.

[INSERT FIGURE 3.6.8 HERE]

3.6.6.2 Antarctic circumpolar wave

The Antarctic circumpolar wave (ACW) is described as an approximately 4-year period pattern of variability in the southern high-latitude ocean-atmosphere system characterized by the eastward propagation of anomalies in Antarctic sea ice extent in a wave train, coupled to anomalies in SST, sea surface height, MSLP, and wind (White and Peterson, 1996; Jacobs and Mitchell, 1996; White and Annis, 2004). Since its initial formulation (White and Peterson, 1996), questions have arisen concerning many aspects of the ACW: the robustness of the ACW on interdecadal timescales (Carril and Navarra, 2001; Simmonds, 2003; Connolley, 2003), its generating mechanisms (Cai and Baines, 2001; Venegas, 2003; White et al., 2004; White and Simmonds, 2006) and even its very existence (Park et al., 2004).

To account for the changes in time of the ACW, Venegas (2003) suggested that it may involve both a self-sustained oscillation with a period of around 3.3 yr generated locally by an ocean-atmosphere coupling with a dominant zonal wave-number three, and an ENSO-forced component with a periodicity of around 5 years and a dominant zonal wave-number 2 across the southern ocean. The ENSO-forced component of the ACW is confirmed in sea level anomalies from TOPEX/Poseidon (Pottier et al., 2004). Using NRA, White and Annis (2004) found a shift or an equatorward (poleward) expansion (retreat) of the ACW toward a warmer (cooler) subtropical South Indian Ocean after (before) 1977, consistent with multi-decadal changes in El Niño evolution.

Based on Antarctic surface temperatures from *in situ* and satellite infrared measurements, Comiso (2000) found that the anomalies over the sea ice region around Antarctica in the last four decades are correlated, especially during winter months, with the ACW. Near-decadal temperature and precipitation variability over Australia and New Zealand appear to be modulated in part by the ACW (White and Cherry, 1999; White, 2000).

3.6.6.3 Indian Ocean Dipole

Large interannual variability of SST in the Indian Ocean has been associated with the Indian Ocean Dipole (IOD) or the Indian Ocean Zonal Mode (IOZM) (Saji et al., 1999; Webster et al., 1999). This pattern manifests itself through a zonal gradient of tropical SST, which in one extreme phase in boreal fall, shows cooling off Sumatra and warming off Somalia in the west, combined with anomalous easterlies along the equator. The magnitude of the secondary rainfall maximum from October to December in East Africa is strongly correlated with IOZM events (Xie et al., 2002). Several recent IOZM events have occurred simultaneously with ENSO events and there is a significant debate on whether the IOZM is an Indian Ocean pattern or whether it is triggered by ENSO in the Pacific Ocean (Allan et al, 2001). One argument for an independent IOZM was the large episode of 1961 when no EN occurred (Saji et al., 1999). Saji and Yamagata (2003), analyzing observations from 1958– 1997, concluded that 11 out of the 19 episodes identified as moderate to strong IOZM events occurred independently of ENSO but this is disputed by Allan et al. (2001) who find that accounting for varying lag correlations removes the apparent independence with ENSO. Trenberth et al. (2002b) showed that Indian Ocean SSTs tend to rise about 5 months after the peak in EN in the Pacific. The strongest ever observed IOZM episode occurred in 1997-98 and was associated with catastrophic flooding in East Africa. Further, at interdecadal timescales, the SST patterns associated with the Indian monsoon rainfall are very similar to the SST patterns associated with the interdecadal variability of ENSO indices (Krishnamurthy and Goswami, 2000) and with the North Pacific interdecadal variability (Deser et al., 2004). Krishnamurthy and Goswami (2000) find that the interannual variances of the ENSOs and Indian monsoon rainfall increase and decrease simultaneously, with the interannual variances of both, the monsoon rainfall and ENSO (Niño-3 SST regions) being high for the warm phase of the interdecadal SST mode in the Eastern Pacific.

3.6.7 Summary

Decadal variations in teleconnections considerably complicate the interpretation of climate change. Since the TAR, it has become clear that a small number of teleconnection patterns account for much of the seasonal to interannual variability in the extratropics. On monthly time scales, the SAM, NAM and NAO are dominant in the extratropics. NAM and NAO are closely related, and mostly independent from SAM, except perhaps on decadal time scales. Many other patterns can be explained through combinations of NAM and PNA in the NH, and SAM and PSA in the SH, plus ENSO-related global patterns. Both the NAM/NAO and the SAM have exhibited trends towards their positive phase (strengthened mid-latitude westerlies) over the last 3–4 decades, although both have returned to near their long-term mean state in the last five years. In the NH, this trend has been a major factor in the wintertime observed change in storm tracks, precipitation and temperature patterns. In the SH, SAM changes are identified with contrasting trends of strong warming in the Antarctic Peninsula, and cooling over the interior of Antarctica, and the increasing positive phase of the SAM has been linked to stratospheric ozone depletion and to greenhouse gas increases. Multi-decadal variability is also evident in the Atlantic, and appears to be related to the thermohaline circulation. Other teleconnection patterns discussed (PNA, PSA) exhibit decadal variations, but have not been shown to have systematic long-term changes.

ENSO has exhibited considerable interdecadal variability in the past century, in association with the PDO (or IPO). Systematic changes in ENSO behaviour have also been observed in the observational record as the 1976–1977 climate shift, which changed the evolution of ENSO events and seems to have enhanced the El Niño phase. Over North America, ENSO and PNA-related changes appear to have led to contrasting changes across the continent, as the west has warmed more than the east, while the latter has become cloudier and wetter. The tropical Pacific variability is influenced by interactions with the tropical Atlantic and Indian Oceans, and also from the extratropical North and South Pacific. Responses of the extratropical ocean become more important as the time scale is extended, and processes such as subduction, gyre changes, and the thermohaline circulation come into play.

3.7 Changes in the Tropics and Subtropics

3.7.1 Monsoons

Monsoons are generally referred to as tropical and subtropical seasonal reversals in both the surface winds and associated precipitation. The strongest monsoons occur over the tropics of southern and eastern Asia and northern Australia and parts of western and central Africa. Rainfall is the most important monsoon variable because the associated latent heat released drives atmospheric circulations and because of its critical role in the global hydrological cycle and its vital socio-economical impacts. Thus, other regions that only have an annual reversal in precipitation with an intense rainy summer and a dry winter have been recently recognized as monsoon regions, even though these regions have no explicit seasonal reversal of the surface winds (Wang, 1994; Webster et al., 1998). The latter regions include Mexico and the southwest United States, and parts of South America and South Africa. Owing to the lack of sufficiently reliable and long-term oceanic observations, analyses of observed long-term changes have mainly relied on land-based rain gauge data.

The global monsoon system embraces an overturning circulation that is intimately associated with the seasonal variation of monsoon precipitation over all major continents and adjacent oceans (Trenberth et al., 2000). Because the variability of regional monsoons is often the result of interacting circulations from other regions, simple indices of monsoonal strength in adjacent regions may give contradictory indications of strength (Webster and Yang, 1992; Wang and Fan, 1999). Decreasing trends in precipitation over the Indonesian Maritime Continent, equatorial parts of western and central Africa, Central America, Southeast Asia, and eastern Australia have been found for 1948–2003 (Chen et al., 2004); see Figure 3.3.2, while increasing trends were evident over the United States and northwestern Australia (see also Section 3.3.2.2 and Figure 3.3.3), consistent with Dai et al. (1997b). Although Chase et al. (2003) found diminished monsoonal circulations after 1950 and reported that the trends since 1979 did not indicate any change in monsoon circulations, results based on NRA suffer severely from artefacts arising from changes in the observing system (Kinter et al., 2004).

Two precipitation datasets (Chen et al., 2002; GHCN, see Section 3.3 and Figure 3.3.3) yield very similar patterns for change in the seasonal precipitation contrasts between 1976–2003 and 1948–1975 (Figure 3.7.1 based on analysis by Wang and Ding, 2006), despite some differences in details and discrepancies in northwest India. Significant decreases in the annual range (wet minus dry season) are observed over the NH tropical monsoon regions (e.g., Southeast Asia, and Central America). Over the East Asian monsoon region, the change over these periods involves increased rainfall in the Yangtze River valley and Korea but decreased rainfall over the lower reaches of the Yellow River and northeast China. However, the total precipitation for the land area of East Asia shows no appreciable change. In the Indonesian-Australian monsoon region, the change between the two periods is characterized by an increase in northwest Australia and Java but a decrease in northeast Australia and a northeastward movement in the SPCZ (Figure 3.7.1). However, the average rainfall shows no long-term trend but significant interannual and interdecadal variations, and the same applies over South America for area-averaged summer precipitation. In the South African monsoon region there is a slight decrease in the annual range of rainfall (Figure 3.7.1), and there is also a decreasing trend in area-averaged precipitation over the South African monsoon region (Figure 3.3.3).

Monsoon predictability depends on many factors, from regional air-sea interaction and land processes (e.g., snow cover fluctuations) to teleconnection influences (ENSO, NAO/NAM, PDO). New evidence, relevant to climate change, indicates that increased loading of aerosols may have strong impacts on monsoon evolution (Menon et al., 2002) through changes in local heating of the atmosphere and land surface (see also Box 3.2 and Chapter 2).

[INSERT FIGURE 3.7.1 HERE]

3.7.1.1 Asia

The Asian monsoon can be divided into the East Asian and the South Asian or Indian monsoon systems (Ding et al., 2004). Based on a summer monsoon index derived from MSLP gradients between land and ocean in the East Asian region, Guo et al. (2003) found a systematic reduction in the East Asian summer monsoon during 1951–2000, with a stronger monsoon dominant in the first half of the period and a weaker monsoon prevailing in the second half (Figure 3.7.2). This long-term change in the East Asian monsoon index is consistent with a tendency for a southward shift of the summer rain-belt over eastern China (Zhai et al., 2004). However, Figure 3.7.2, based on the newly developed HadSLP2 data set (Allan and Ansell, 2006), suggests that the weakening trend, which started in the 1920s, is not representative of the longer record extending back to the 1850s, which shows marked decadal-scale variability before the 1940s.

[INSERT FIGURE 3.7.2 HERE]

Several changes in the Asian monsoon occurred about the time of the 1976–1977 climate shift along with changes in ENSO (Qian et al., 2003; Huang et al., 2003) and declines in land precipitation are evident in Southern Asia and, to some extent, in Southeast Asia (see Figure 3.3.3). Gong and Ho (2002) suggested that the change in summer rainfall over the Yangtze River valley was due to a southward rainfall shift and Ho et al. (2003) also noted a sudden change in Korea. These occurred about the same time as a change in the 500 hPa geopotential height and typhoon tracks in summertime over the northern Pacific (Gong et al., 2002) (see Section 3.6.3) related to the enlargement, intensification and southwestward extension of the northwest

Pacific subtropical high. When the equatorial central and eastern Pacific is in a decadal warm period, summer monsoon rainfall is stronger in the Yangtze River valley but weaker in North China. A strong tropospheric cooling trend is found in East Asia during July and August. Accompanying this summer cooling the upper-level westerly jet stream over East Asia shifts southward and the East Asian summer monsoon weakens, which results in the tendency toward increased droughts in northern China and flood in Yangtze River Valley (Yu et al., 2004).

The Indian monsoon season occurs from June to September, when about 70% of annual rainfall is received and monsoon rainfall exhibits decadal variability. Observational studies have shown that the impact of El Niño is more severe during the below normal epochs, while the impact of La Niña is more severe during the above normal epochs (Kripalani and Kulkarni, 1997a; Kripalani et al., 2001, 2003). Such modulation of ENSO impacts by the decadal monsoon variability is also observed in the rainfall regimes over Southeast Asia (Kripalani and Kulkarni, 1997b). Links between monsoon-related events (rainfall over South Asia, rainfall over East Asia, NH circulation, tropical Pacific circulation) weakened between 1890 and 1930 but strengthened during 1930–1970 (Kripalani and Kulkarni, 2001). The strong inverse relationship between El Niño events and Indian monsoon rainfalls that prevailed for over a century prior to about 1976 has weakened substantially since then (Krishnamurthy and Goswami, 2000; Kumar et al., 1999; Sarkar et al., 2004) and involves large-scale changes in atmospheric circulation. Shifts in the Walker circulation and enhanced land-sea contrasts appear to be countering effects of increased El Niño activity. Ashok et al. (2001) also find that the Indian Ocean Dipole (see Section 3.6.6.3) plays an important role as a modulator of Indian rainfall. ENSO is also related to atmospheric fluctuations both in the Indian sector and in northeastern China (Kinter et al., 2002).

3.7.1.2 Australia

The Australian monsoon occupies the northern third of continental Australia and surrounding seas and, considering its closely coincident location and annual evolution, is often studied in conjunction with the monsoon over the islands of Indonesia and Papua New Guinea. The Australian monsoon exhibits large interannual and intraseasonal variability, largely associated with the effects of ENSO, the Madden-Julian Oscillation (MJO), and tropical cyclone (TC) activity (McBride, 1998; Webster et al., 1998; Wheeler and McBride, 2005). Using rain-gauge data, Hennessy et al. (1999) found an increasing trend in calendar-year total rainfall in Northern Territory of 18% from 1910 to 1995, attributed mostly to enhanced monsoon rainfall in the 1970s and coincident with an almost 20% increase in the number of rain days. With data updated to 2002, Smith (2004) demonstrated that increased monsoon rainfall has become statistically significant over northern, western, and central Australia. Northern Australian wet-season rainfall updated through 2004/2005 (Figure 3.7.3) (Jones et al., 2004) shows the positive trend and the contribution to it from the anomalously wet period of the mid-1970s, as well as the more recent anomalously wet period around 2000 (see also Smith, 2004). These two wet periods also constitute a large amount of the decadal variability present in the monsoon. Wardle and Smith (2004) have argued that the upward rainfall trend is consistent with the upward trend in land-surface temperatures that has been observed in the south of the continent, independent of changes over the oceans. Strong decadal variations in Australian precipitation have also been noted (Figure 3.7.3). Using northeastern Australian rainfall, Latif et al. (1997) has shown that rainfall was much increased during decades when the tropical Pacific was anomalously cold in the 1950s and 1970s. This strong relationship does not extend to the Australian monsoon as a whole; however, as the rainfall time series (Figure 3.7.3) has only a weak negative correlation (~–0.2) with the IPO. The fact that the long-term trends in rainfall and Pacific SSTs are both positive, and hence opposing their interannual relationship (Power et al., 1998), explains only a portion of why the correlation is diminished at decadal time scales. Trends in CAPE in northern Australia are weak and not significant (Gettelman et al., 2002; DeMott and Randall, 2004).

[INSERT FIGURE 3.7.3 HERE]

3.7.1.3 The Americas

The North American Monsoon System (NAMS) is characterized by ocean-land contrasts including summer heating of higher elevation mountain and plateau regions of Mexico and the southwestern United States, a large-scale upper level anticyclonic circulation, a lower level thermal low, and a strong subsidence region to the west in the cool stratus regime of the eastern North Pacific (Vera et al., 2006). The NAMS contains a strong seasonal structure (Higgins and Shi, 2000), with rapid onset of monsoon rains in southwestern Mexico in June, a later northward progression into the southwest United States later during its mature phase in July and August, and a gradual decay in September and October.

Timing of the start of the northern portion of the NAMS has varied considerably, with some years starting as early as mid-June and others starting as late as early August (Higgins and Shi, 2000). Since part of NAMS variability is governed by larger-scale climate conditions, it is susceptible to interannual and multi-decadal variations. Higgins and Shi (2000) further suggest that the northern portion of the NAMS may be affected by the PDO, wherein anomalous winter precipitation over western North America is correlated with North American monsoon conditions in the subsequent summer.

The South American Monsoon System (SAMS) is evident over South America in the austral summer (Barros et al., 2002; Nogués-Paegle et al., 2002; Vera et al., 2006). It is a key factor for the warm season precipitation regime. In northern Brazil, different precipitation trends (see Figure 3.3.3 for the Amazon and Southern South America regions) have been observed over northern and southern Amazonia, showing a dipole structure (Marengo, 2004) suggesting a southward shift of the SAMS. This is consistent with Rusticucci and Penalba (2000), who found a significant positive trend in the importance of the annual precipitation cycle, indicating a long-term climate change of the monsoon regime over the frontier across the semi-arid region of the La Plata Basin. Also, the long-term variability of the mean wind speed of the low level jet, a component of the SAMS that transports moisture from the Amazon to the south and southwest, showed a positive trend (Marengo et al., 2004).

3.7.1.4 Africa

Since the TAR, significant advances have been made in the investigation of the dominant modes of climate variability over eastern Africa. A variety of studies have firmly established that ENSO and SSTs in the Indian Ocean are the dominant sources of climate variability over eastern Africa (Goddard and Graham, 1999; Yu and Rienecker, 1999; Indeje et al., 2000; Clark et al., 2003). Further, Schreck and Semazzi (2004) isolated a secondary but significant pattern of regional climate variability based on seasonal (OND) rainfall data. In distinct contrast with the ENSO-related spatial pattern, the trend pattern in their analysis is characterized by positive rainfall anomalies over the northeastern sector of eastern Africa (Ethiopia, Somalia, Kenya and northern Uganda) and opposite conditions over the southwestern sector (Tanzania, southern parts of the Democratic Republic of the Congo and southwestern Uganda). This signal significantly strengthened in recent decades. Warming is associated with an earlier onset of the rainy season over the northeastern Africa region and a late start over the southern sector.

West Africa experiences marked variability in rainfall (e.g., Le Barbe et al., 2002; Dai et al., 2004a). Wet conditions in the 1950s and 1960s gave way to much drier conditions in the 1970s, 1980s and 1990s. The rainfall deficit in this region during 1970 to 1990 was uniform across the region implying that the deficit was not due to a spatial shift in the peak rainfall (Le Barbe et al., 2002) and was mainly linked to a reduction in the number of significant rainfall events occurring during the peak monsoon period (JAS) in the Sahel and during the first rainy season south of about 9°N. The decreasing rainfall and devastating droughts in the Sahel region during the last three decades of the 20th century (Figure 3.7.4) are among the largest climate changes anywhere. Dai et al. (2004a) provided an updated analysis of the normalised Sahel rainfall index based on the years 1920–2003 (Figure 3.7.4). Following the major El Niño event of 1982–1983, the rainfall reached a minimum of 170 mm below the long-term mean of (~506 mm). Since 1982 there is some evidence for a recovery but despite this the mean of the last decade is still well below the pre-1970 level. These authors also noted that large-multi-year oscillations appear to be more frequent and extreme after the late 1980s than previously.

ENSO impacts the West African monsoon and the correlation between Sahel rainfall and ENSO during JJA varied between 1945 and 1993 (Janicot et al., 2001). The correlation is always negative but was not significant during the 1960s to mid-70s when the role of the tropical Atlantic was relatively more important. Years when ENSO has a larger impact tend to be associated with same-signed rainfall anomalies over the west African region whereas years when the tropical Atlantic is more important tend to have a so-called anomalous “dipole” pattern, with the Sahel and Guinea Coast having opposite signed rainfall anomalies (Ward, 1998). Giannini et al. (2003) suggest that both interannual and decadal variability of the rainfall in the Sahel results from the response of the African summer monsoon to oceanic forcing, amplified by land-atmosphere interaction, based on model results.

While other parts of Africa have experienced statistically significant weakening of the monsoon circulation, analyses of long-term southern African rainfall totals in the wet season (JFM) have consistently reported no trends (Fauchereau et al., 2003). Decreases in rainfall are evident in analyses of shorter periods, such as the decade 1986–1995 which was the driest of the 20th century. New et al. (2006) report a decrease in average rainfall intensity and an increase in dry spell length (consecutive dry day length) for 1961–2000.

[INSERT FIGURE 3.7.4 HERE]

3.7.2 The Hadley and Walker Circulations, ITCZ, and Subtropical Highs

The Hadley Circulation (HC) is commonly defined as the zonal mean meridional overturning mass flow between the tropics and subtropics. In this zonally symmetric view, equatorward-moving air within the trade winds of both hemispheres converges in the lower troposphere and rises within the Intertropical Convergence Zone (ITCZ). The air then diverges and flows poleward in the upper troposphere into the subtropics, where it descends within the subtropical regions. This thermally driven direct circulation results in heavy precipitation within the ITCZ, and dry conditions in the subtropics under the influence of the subtropical high. The HC is strongest during the solstice seasons, when the ITCZ is located farthest from the equator and the cross-equatorial heating gradient is largest (Trenberth et al., 2000; Trenberth and Stepaniak, 2003a, b). The ITCZ and HC exhibit strong zonal asymmetries, due primarily to the distribution of land and ocean at low latitudes. The South Pacific Convergence Zone (SPCZ) is a semi-permanent cloud band extending from around the Coral Sea southeastward toward the extratropical South Pacific, while the South Atlantic Convergence Zone (SACZ) is a more transient feature over Brazil that transports moisture originating over the Amazon into the South Atlantic (Liebmann et al., 1999).

From the Tropics to about 31° latitude, the primary energy transport mechanism in the atmosphere is the Hadley and Walker overturning circulations (Trenberth and Stepaniak, 2003b). Tropical SSTs determine where the upward branch of the HC is located over the oceans. The dominant variations in the energy transports by the Hadley cell, which reflect on the Hadley cell strength itself, occur with ENSO (Trenberth et al., 2002a; Trenberth and Stepaniak, 2003a). During El Niño, elevated SST causes an increase in convection and relocation of the ITCZ and SPCZ to the equator over the central and eastern tropical Pacific, with a tendency for drought conditions over Indonesia. There follows a weakening of the Walker Circulation (WC), and a strengthening of the HC (Oort and Yienger, 1996; Trenberth and Stepaniak, 2003a). A strengthened local HC leads to drier conditions over many subtropical regions during El Niño, especially over the Pacific sector. As discussed in Section 3.4.4.1, increased divergence of energy out of the tropics in the 1990s relative to the 1980s (Trenberth and Stepaniak, 2003a) is associated with more El Niño events and especially the major 1997–1998 El Niño event, so these conditions clearly play a role in interdecadal variability. Examination of the HC in several datasets (Mitas and Clement, 2005) suggests some strengthening, although discrepancies among reanalysis datasets and known deficiencies raise questions about the robustness of the strengthening, especially prior to the satellite era (1979).

Deser et al. (2004) related epochs of high sea level pressure over the North Pacific (1900–1924 and 1947–1976) and epochs of low pressure (1925–1946 and 1977–1997) to climate variables throughout the tropical Indo-Pacific region. SST anomalies in the tropical Indian Ocean and southeast Pacific Ocean, rainfall and cloudiness anomalies in the vicinity of the SPCZ, stratus clouds in the eastern tropical Pacific, and sea level pressure differences between the tropical southeast Pacific and Indian Oceans all exhibit prominent interdecadal fluctuations that are coherent with those in sea level pressure over the North Pacific, implying also changes in the Hadley and Walker circulations. The spatial patterns of the interdecadal tropical climate anomalies are similar to but not identical with ENSO. Mu et al. (2002) also found 40-year oscillations in the northwestern Pacific subtropical high, while Gong and Ho (2002) noted a significant decadal shift of the Northwest Pacific subtropical high in summer about 1979–1980, after which the western North Pacific subtropical high has been enlarged, intensified, and shifted southwestward.

Positive SST anomalies in the western subtropical South Atlantic are associated with positive rainfall anomalies over the SACZ region (Doyle and Barros, 2002; Robertson et al., 2003). Barros et al. (2000) found that, during summer, the SACZ was displaced northward (southward) and more intense (weaker) with cold (warm) SST anomalies to its south. The ITCZ is modulated in part by surface features, like the gradient of SST over the equatorial Atlantic (Chang et al., 1999; Nogués-Paegle et al., 2002), and it modulates the interannual variability of seasonal rainfall over eastern Amazonia and northeastern Brazil (Nobre and Shukla, 1996). An interdecadal pattern of tropical convection and surface temperatures in the West African monsoon region, the central tropical Pacific, the Amazon basin, and the tropical Indian Ocean has been documented by Chelliah and Bell (2004).

3.7.3 Summary

Multi-time scale variability strongly affects monsoon systems. Large interannual variability associated with ENSO dominates the Hadley Circulation, Walker Circulation, ITCZ, and monsoons. There is also good evidence for decadal changes associated with monsoonal rainfall changes in many monsoon systems, especially across the 1976–1977 climate shift, but data uncertainties compromise evidence for trends. Some monsoons, especially the East Asian Monsoon System, have experienced a dipole change in precipitation with increases in one region and decreases in the other during the recent 50 years. However, the physical mechanism of such changes in monsoon system can not be clearly explained.

3.8 Changes in Extreme Events

3.8.1 Background

There is increasing concern that extreme events may be changing in frequency and intensity as a result of human influences on climate. Climate change may be perceived most through the impacts of extremes although these are to a large degree dependent on the system under consideration, including its vulnerability, resiliency and capacity for adaptation and mitigation; topics addressed by IPCC WGII. Improvements in technology mean that we hear about extremes in most parts of the world within a few hours of their occurrence. Pictures shot by camcorders on the news may foster a belief that weather-related extremes are increasing in frequency. An extreme weather event becomes a disaster when society and/or ecosystems are unable to effectively cope with it. Growing human vulnerability (due to growing numbers of people living in exposed and marginal areas or due to the development of more high-value property in high-risk zones) is increasing the risk, while human endeavours (such as by local governments) try to mitigate possible effects.

The assessment of extremes in this section is based on long-term observational series of weather elements. As in the TAR extremes refer to rare events based on a statistical model of particular weather elements and changes in extremes may relate to changes in the mean and variance in complicated ways. Changes in extremes are assessed on a range of time and space scales; e.g., from extremely warm years globally to peak rainfall intensities locally, and examples are given in Box 3.6. To span this entire range, data are required at a daily (or less) time scale. However, the availability of observational data restricts the type of extremes that can be analysed. The rarer the event, the more difficult it is to identify long-term changes, simply because there are fewer cases to evaluate (Frei and Schär, 2001; Klein Tank and Können, 2003). Identification of changes in extremes is also dependent on the analysis technique employed (Zhang et al., 2004a). To avoid excessive statistical limitations, trend analyses of extremes have traditionally focused on standard and robust statistics that describe moderately extreme events. In percentile terms these are events occurring between 1% and 10% of the time, at a particular location in a particular reference period (generally 1961–1990).

Global studies of daily temperature and precipitation extremes over land (e.g., Frich et al., 2002; see also the TAR) suffer from both a scarcity of data and regions with missing data. The main reason is that in various parts of the globe, there is a lack of homogeneous observational records with daily resolution covering multiple decades that are part of integrated digitized datasets (Mason et al., 2003). In addition, existing records are often inhomogeneous; for instance as a result of changes in observing practices or urban heat island effects (Vincent et al., 2002; DeGaetano and Allen, 2002; Wijngaard et al., 2003). This affects, in particular, our understanding of extremes, because changes in extremes are often more sensitive to inhomogeneous climate monitoring practices than changes in the mean (see Appendix 3.B.2 and 3.B.4). Consistent observing is also a problem when assessing long-term changes in the frequency and severity of tropical and extra-tropical storms. Similar difficulties are encountered when trying to find worldwide observational evidence for changes in severe local weather events like tornadoes, hail, thunderstorms and dust storms. Analyses of trends in extremes are also sensitive to the analysis period; e.g., the inclusion of the exceptionally hot European summer of 2003 may have a marked influence on results if the period is short.

Since the TAR, the situation with observational datasets has improved, although efforts to update and exchange data must be continued (e.g., GCOS, 2004). Results are now available from newly established regional- and continental-scale daily datasets; from denser networks, from temporally more extended high-quality time series, and from many existing national data archives, which have been expanded, to cover longer time periods. Moreover, the systematic use and exchange of time series of standard indices of extremes (with common definitions) provides an unprecedented global picture of changes in daily temperature and precipitation extremes (Alexander et al., 2006 which updates the results of Frich et al., 2002 presented in the TAR).

As an alternative, but not independent, data source, reanalyses can also be analysed for changes in extremes (see Appendix 3.B.5.3). Although spatially and temporally complete, under-representation of certain types of extremes (Kharin and Zwiers, 2000) and spurious trends in the reanalyses (especially in the tropics and in the SH) remain problematic, in particular before the start of the modern satellite era in 1979 (Sturaro, 2003; Marshall, 2002, 2003; Sterl, 2004; Trenberth et al., 2005a). For instance, Bengtsson et al. (2004) found that analysed global kinetic energy rose by almost 5% in 1979 as a direct consequence of the inclusion of improved satellite information over the oceans, which is expected to significantly affect analysed storm activity over the southern oceans, where ship data are sparse.

In this section observational evidence for changes in extremes is assessed for temperature, precipitation, tropical and extratropical cyclones and severe local weather events. Most studies of extremes consider the period since about 1950 with even greater emphasis on the last few decades (since 1979), although longer datasets exist for a few regions enabling more recent events to be placed in a longer context. We discuss mostly the changes observed in the daily weather elements, where most progress has been made since the TAR. Droughts (although they are considered extremes) are covered in Section 3.3.4 as they are more related to longer periods of anomalous climate.

3.8.2 Evidence for Changes in Variability or Extremes

3.8.2.1 Temperature

For temperature extremes in the 20th century, the TAR highlighted the lengthening of the growing or freeze-free season in most mid- and high-latitude regions. The reduction in the frequency of extreme low monthly and seasonal average temperatures, and smaller increases in the frequency of extreme high average temperatures. In addition, there was evidence to suggest a decrease in the intra-annual temperature variability with consistent reductions in frost days and increases in warm night-time temperatures across much of the globe.

Evidence for changes in observed interannual variability (such as standard deviations of seasonal averages) is still sparse. Scherrer et al. (2005) investigated standardized distribution changes for seasonal mean temperature in central Europe and found temperature variability to show a weak increase (decrease) in summer (winter) for the time period 1961 to 2004, but these changes are not statistically significant at the 10% level. On the daily time scale, a number of regional studies have been completed for southern South America (Vincent et al., 2005), Central America and northern South America (Aguilar et al., 2005), Caribbean (Peterson et al., 2002), North America (Kunkel et al., 2004; Vincent and Mekis, 2006), the Arctic (Groisman et al., 2003), central and northern Africa (Easterling et al., 2003), southern and western Africa (New et al., 2006), the Middle East (Zhang et al., 2005), Western Europe and east Asia (Kiktev et al., 2003), Australasia and southeast Asia (Griffiths et al., 2005), China (Zhai and Pan, 2003), and central and southern Asia (Klein Tank et al., 2006). They all show patterns of changes in extremes consistent with a general warming, although the observed changes of the tails of the temperature distributions are often more complicated than a simple shift of the entire distribution would suggest (see Figure 3.8.1). Also, uneven trends are observed for night-time and day-time temperature extremes. In southern South America, significant increasing trends were found in the occurrence of warm nights and decreasing trends in the occurrence of cold nights but no consistent changes in the indices based on daily maximum temperature. In Central America and northern South America, warm extremes of both minimum and maximum temperature extremes have increased. Warming of both the night-time and day-time extremes was also found for the other regions where data have been analysed. For Australasia and Southeast Asia, the dominant distribution change at rural stations for both maximum and minimum temperature, involved a change in the mean, impacting on either one or both distribution tails, with no significant change in standard deviation (Griffiths et al., 2005). For urbanized stations, however, the dominant change also involved a change in the standard deviation. This result was particularly evident for minimum temperature.

Few other studies have considered mutual changes in both the warm and cold tail of the same daily (minimum, maximum or mean) temperature distribution. Klein Tank and Können (2003) analysed such changes over Europe using standard indices to find that the annual number of warm extremes (above the 90th percentile for 1961–1990) of the daily minimum and maximum temperature distributions increased 2 times faster during the last 25 years than expected from the corresponding decrease in the number of cold extremes (lowest 10%). Moberg and Jones (2005) found that both the warm and the cold tail (defined by the 90th and 10th percentile) of the daily minimum and maximum temperature distribution over Europe in winter warmed over the 20th century as a whole with the warm tail of minimum temperature warming significantly in summer. For an even longer period, Yan et al. (2002) found decreasing warm extremes in Europe and China up to the late-19th century; decreasing cold extremes since then and increasing warm extremes only since 1961, especially in summer (JJA). Brunet et al. (2006b) analysed 22 Spanish records for the period 1894–2003 and found greater reductions in the number of cold days than increases in warm days. Since 1973, though, warm days have been rising dramatically, particularly near the Mediterranean coast. Beniston and Stephenson (2004) showed that changes in extremes of daily temperature in Switzerland were due to changes in both the mean and the variance of the daily temperatures. Vincent and Mekis (2006) find progressively fewer extreme cold nights and cold days and conversely more extreme warm nights and warm days for Canada from 1900–2003 and Robeson (2004) find intense warming of the lowest daily minimum temperatures over North America. In Argentina, the strong positive changes in minimum temperature seen during 1959–1998 were caused by significant increases in warm nights; there were also decreases in cold days (Rusticucci and Barrucand, 2004).

Alexander et al. (2006) and Caesar et al. (2006) have brought all these and other regional results together, gridding the common indices or data for the period 1951–2003. Over 76% of the global land area sampled showed a significant decrease in the annual occurrence of cold nights; a significant increase in the annual occurrence of warm nights took place over 72% of the area (Table 3.6, Figure 3.8.1 and Question 3.3). This implies a positive shift in the distribution of daily minimum temperature throughout the globe. Changes in the occurrence of cold days and warm days show warming as well, but generally less marked. This is consistent with the increase in minimum as opposed to maximum temperature leading to a reduction in DTR since 1951 (see Section 3.2.2.1 and 3.2.2.7). The change in the four extremes indices (Table 3.6) also show that the distribution of minimum temperature and the distribution of maximum temperature have not only shifted, but also changed in shape. The indices for the number of cold and warm events have changed almost equally, which indicates that the cold tails of the distributions have warmed considerably more than the warm tails over the last 50 years.

Table 3.6. Global trends in extremes of temperature or precipitation as measured by the 10th and 90th percentiles (for 1961–1990). Trends, ±2 standard error ranges and significances (**bold: <1%**) were estimated by REML (see Appendix 3.A) which allows for serial correlation in the residuals of the data about the linear trend. All trends are based on annual averages. Values are % decade^{-1}.

Series	1951–2003	1979–2003
TN10	**–1.17 ± 0.24**	**–1.24 ± 0.54**
TN90	**1.43 ± 0.51**	**2.60 ± 0.99**
TX10	**–0.63 ± 0.20**	**–0.91 ± 0.58**
TX90	**0.71 ± 0.42**	**1.74 ± 0.88**
PREC	**0.21 ± 0.12**	0.41 ± 0.46

Notes:
TN10 % incidence of T_{min} below coldest decile.
TN90 % incidence of T_{min} above warmest decile.
TX10 % incidence of T_{max} below coldest decile.
TX90 % incidence of T_{max} above warmest decile.
PREC % contribution of very wet days (above the 95th percentile) to the annual precipitation total.

[INSERT FIGURE 3.8.1 HERE]

3.8.2.2 Precipitation

The conceptual basis for changes in precipitation has been given by Allen and Ingram (2002) and Trenberth et al. (2003), see Question 3.2. Issues relate to changes in type, amount, frequency, intensity and duration of precipitation. Observed increases in atmospheric water vapour (see Section 3.4.2) imply increases in intensity, but this will lead to reduced frequency or duration if the total evaporation rate from the Earth's surface (land and ocean) is unchanged. The TAR states that it is likely that there has been a statistically significant 2 to 4% increase in the frequency of heavy and extreme precipitation events when averaged across the mid and high latitudes. Since then a more refined understanding has been achieved of the observed changes in precipitation extremes.

Many analyses indicate that the evolution of rainfall statistics through the second half of the 20th century is dominated by variations on the interannual to inter-decadal time scale and that trend estimates are spatially incoherent, as would be anticipated with the relatively high spatial and interannual variability of precipitation (Manton et al., 2001, Peterson et al., 2002, Griffiths et al., 2003, Herath and Ratnayatke, 2004). In Europe, there is a clear majority of stations with increasing trends in the number of moderate and very wet days (defined as the exceedence of the 75% and 95% percentiles respectively) during the second half of the 20th century (Klein Tank and Können, 2003; Haylock and Goodess, 2004). Similarly, for the contiguous United States, Kunkel et al. (2003) and Groisman et al. (2004) confirm earlier results and find statistically significant increases in heavy (upper 5%) and very heavy (upper 1%) precipitation, by 14% and 20%, respectively. Much of this increase has occurred during the last three decades of the century and it is most apparent over the eastern parts of the country. Also there is new evidence for Europe and the United States that the relative increase in precipitation extremes is larger than the increase in mean precipitation, and this is manifested as an increasing contribution of heavy events to total precipitation (Klein Tank and Können, 2003; Groisman et al., 2004).

Despite a decrease in mean annual rainfall, an increase in the fraction from heavy events was inferred for large parts of the Mediterranean (Alpert et al., 2002; Brunetti et al., 2004; Maheras et al., 2004). Further, Kostopoulou and Jones (2005) note contrasting trends of heavy rainfall events between an increase in the central Mediterranean (Italy) and a decrease over the Balkans. Also in South Africa, Siberia, central Mexico, Japan and the northeastern part of the United States an increase in only heavy precipitation is observed while total precipitation and/or the frequency of days with an appreciable amount of precipitation (wet days) is either not changing or is decreasing (Easterling et al., 2000; Fauchereau et al., 2003; Sun and Groisman, 2004; Groisman et al., 2005).

A number of recent regional studies have been completed for southern South America (Haylock et al., 2006), Central America and northern South America (Aguilar et al., 2005), southern and western Africa (New et al., 2006), the Middle East (Zhang et al., 2005), and central and southern Asia (Klein Tank et al., 2006). For southern South America, the pattern of trends for extremes between 1960 and 2000 for the extremes was generally the same as that for total annual rainfall (Haylock et al., 2006). A majority of stations show a change to wetter conditions, related to the generally lower value of the SOI since 1976–1977, with the exception of southern Peru and southern Chile, where a decrease was observed in many precipitation indices. In the latter region, the change has led to a weakening of the continental trough giving a southward shift in storm tracks and an important effect on the observed rainfall trends. No significant increases in the total amounts are found over Central America and northern South America (see also Figure 3.3.3), but rainfall intensities have increased related to changes in SST of tropical Atlantic waters. Over southern and western Africa, and the Middle East there are no spatially coherent patterns of statistically significant trends in precipitation indices. Averaged over central and southern Asia, a slight indication of disproportionate changes in the precipitation extremes compared with the total amounts is seen. In the Indian sub-continent Sen Roy and Balling (2004) find that about two thirds of all considered time series exhibit increasing trends in indices of precipitation extremes and that there are coherent regions with increases and decreases.

Alexander et al. (2006) have also gridded the extreme indices for precipitation (as for temperature in Section 3.8.2.1). Changes in precipitation extremes are much less coherent than for temperature, but globally-averaged over the land area with sufficient data, the percentage contribution to total annual precipitation from very wet days (upper 5%) is greater in recent decades than earlier decades (Figure 3.8.2a). Observed changes in intense precipitation (with geographically varying thresholds between the 90th and 99.9th percentile of daily precipitation events) for more than a half of the land area of the globe indicate an increasing probability of intense precipitation events beyond that expected from changes in the mean for many extra-tropical regions (Groisman et al., 2005) (Figure 3.8.2b). This robust finding confirms the disproportionate changes in the precipitation extremes described in the majority of regional studies above, in particular for the mid latitudes since about 1950. It is still difficult to draw a consistent picture of changes for the tropics and the subtropics, where many areas are not analyzed and data are not readily available.

As well as confirming previous findings, the new analyses provide seasonal detail and insight into longer-term variations for the mid latitudes. Whilst the increase in the United States is found primarily in the warm season (Groisman et al., 2004), central and northern Europe exhibited changes primarily in winter (DJF) and changes were insignificant in summer (JJA) – but the studies did not include the extreme European summers of 2002 (very wet) and 2003 (very dry) (Haylock and Goodess, 2004; Osborn and Hulme, 2002; Schmidli and Frei, 2005). Although data are not as good, the frequencies of precipitation extremes in the United States were at comparable levels from 1895 into the early 1900s to those during the 1980s to 1990s (Kunkel et al., 2003). For Canada (excluding the high latitude Arctic), Zhang et al. (2001a) and Vincent and Mekis (2006) find that the frequency of precipitation days significantly increases during the 20th century but averaged for the country as a whole, there is no identifiable trend in precipitation extremes. Nevertheless, Groisman et al. (2005) find significant increases in the frequency of heavy and very heavy (between the 95th and 99.7th percentile of daily precipitation events) precipitation in British Columbia south of 55ºN for 1910 to 2001, and in other areas (Figure 3.8.2b).

[INSERT FIGURE 3.8.2 HERE]

Since the TAR, several regional analyses have been undertaken for statistics with return periods much longer than in the previous studies. For the UK, Fowler and Kilsby (2003a, b), using extreme value statistics, estimate that the recurrence of 10-day precipitation totals with a 50-year return period based on data for 1961–1990 has increased by a factor of 2 to 5 by the 1990s in northern England and Scotland. Their results for long return periods are qualitatively similar to changes obtained for traditional (moderate) statistics (Osborn and Hulme, 2002; Osborn et al., 2000), but there are differences in the relative magnitude of the change between seasons (Fowler and Kilsby, 2003b). For the contiguous United States, Kunkel et al. (2003) and Groisman et al. (2004) analyse return periods of 1 to 20 years, and interannual to interdecadal variations during the 20th century exhibit a high correlation between all return periods. Similar results were obtained for several extra-tropical regions (Groisman et al., 2005), including the central United States, the northwestern coast of North America, southern Brazil, Fennoscandia, the East European Plain, South Africa, southeastern Australia, and Siberia. In summary, from the available analyses there is evidence that the changes at the extreme tail of the distribution are consistent with changes inferred for more robust statistics based on 75th and 95th percentiles.

3.8.3 Evidence for Changes in Tropical Storms

Box 3.5: Tropical Cyclones and Climate Change

In the summer tropics, outgoing longwave radiative cooling from the surface to space is not effective in the high water vapour optically-thick environment of the tropical oceans. Links to higher latitudes are weakest in the summer tropics and transports of energy by the atmosphere, such as occur in wintertime, are also not an effective cooling mechanism, while monsoonal circulations between land and ocean redistribute energy in areas where they are active. However, tropical storms cool the ocean surface through mixing with cooler deeper ocean layers and through evaporation. When the latent heat is realized in precipitation in the storms, the energy is transported high into the troposphere where it can radiate to space, with the system acting somewhat like a Carnot cycle (Emanuel, 2003). Hence tropical cyclones appear to play a key role in ameliorating the heat from the summer sun over the oceans.

As the climate changes and SSTs continue to increase (see 3.2.2.3), the environment in which tropical storms form is changed. Higher SSTs are generally accompanied by increased water vapour in the lower troposphere (see Section 3.4.2.2 and Figure 3.4.5), thus the moist static energy that fuels convection and thunderstorms is also increased. Hurricanes and typhoons generally form from pre-existing disturbances only where SSTs exceed about 26°C and, as SSTs have increased, it thereby potentially expands the areas over which such storms can form. However, many other environmental factors also influence the generation and tracks of disturbances, and wind shear in the atmosphere greatly influences whether or not these disturbances can develop into tropical storms. ENSO and variations in monsoons as well as other factors also affect where storms form and track (e.g., Gray, 1984). Whether the large-scale thermodynamic environment and atmospheric static stability (often measured by Convective Available Potential Energy, CAPE) becomes more favourable for tropical storms depends on how changes in atmospheric circulation, especially subsidence, affect the static stability of the atmosphere, and how the wind shear changes. The potential intensity, defined as the maximum wind speed achievable in a given thermodynamic environment (e.g., Emanuel, 2003), similarly depends critically on SSTs and atmospheric structure. The tropospheric lapse rate is maintained mostly by convective transports of heat upwards, in thunderstorms and thunderstorm complexes, including mesoscale disturbances, various waves, and tropical storms, while radiative processes serve to cool the troposphere. Increases in greenhouse gases decrease radiative cooling aloft, thus potentially stabilizing the atmosphere. In models, the parameterization of sub-grid scale convection plays a critical role in determining whether this stabilization is realized and whether CAPE is released or not. All of these factors, in addition to SSTs, determine whether convective complexes become organized as rotating storms and form a vortex.

While attention has often been focussed simply on the frequency or number of storms, the intensity and duration likely matter more. The energy in a storm, such as measured by NOAA’s Accumulated Cyclone Energy (ACE) index (Levinson and Waple, 2004), accounts for the collective intensity and duration of tropical storms and hurricanes during a given season and is proportional to velocity squared. The power dissipation of a storm is proportional to the wind speed cubed (Emanuel, 2005a), as the main dissipation is from surface friction and wind stress effects, and is measured by a Power Dissipation Index (PDI). Consequently, the effects of these storms are highly nonlinear and one big storm may have much greater impacts on the environment and climate system than several smaller storms.

From an observational perspective then, key issues are the tropical storm formation regions, the frequency, intensity, duration and tracks of tropical storms, and associated precipitation. For land-falling storms, the damage from winds and flooding, as well as storm surges, are especially of concern, but often depend more on human factors, including whether people place themselves in harms way, their vulnerability, and their resiliency through such things as building codes.

The TAR noted that evidence for changes in tropical cyclones (both in number and intensity) across the various ocean basins is often hampered by classification changes. In addition, considerable inter-decadal variability reduces significance of any long-term trends. Careful interpretation of observational records is therefore required, see also Box 3.5. Traditional measures of tropical cyclones, hurricanes and typhoons have varied in different regions of the globe, and typically have required thresholds to be crossed in terms of estimated wind speed and organization to be called a tropical storm, named storm, cyclone, hurricane or typhoon, or major hurricane or super typhoon. Many other measures or terms exist such as “named storm days”, “hurricane days”, “intense hurricanes”, “net tropical cyclone activity”, and so on.

The ACE index, see Box 3.5, (Figure 3.8.3) is essentially a wind energy index, defined as the sum of the squares of the estimated 6-hourly maximum sustained wind speed (knots) for all named systems while they are at least tropical storm strength. Since this index represents a continuous spectrum of both system duration and intensity, it does not suffer as much from the discontinuities inherent in more widely used measures of activity such as the number of tropical storms, hurricanes, or major hurricanes. However, the ACE values reported here are not adjusted for known inhomogeneities in the record (discussed below). The ACE index is also used to define above-, near-, and below-normal hurricane seasons (based on the 1981-2000 period). The index has the same meaning in every region. It integrates over size and intensity. Figure 3.8.3 shows the ACE index for 6 regions (adapted from Levinson, 2005, and updated through 2005). Prior to about 1970, there was no satellite imagery to help estimate the intensity and strength of tropical storms, so the estimates of ACE are less reliable, and values are not given prior to about the mid- or late-1970s in the Indian Ocean, South Pacific or Australian regions. Values are given for the Atlantic, and two North Pacific regions after 1948, although reliability improves over time, and trends contain unquantified uncertainties.

The Potential Intensity (PI) of tropical cyclones (Emanuel, 2003) can be computed from observational data based primarily on vertical profiles of temperature and humidity, CAPE (Box 3.5), and SSTs. In analysing CAPE from selected radiosonde stations throughout the tropics for the period 1958 to 1997, Gettelman et al. (2002) found mostly positive trends. DeMott and Randall (2004) found more mixed results, although their data may have been contaminated by spurious adjustments (Durre et al., 2002). Further, Free et al. (2004a) found that trends in PI were small and statistically insignificant at a scattering of stations in the tropics. As all of these studies were likely contaminated by problems with tropical radiosondes (Sherwood et al., 2005; Randel and Wu, 2006) (see Section 3.4.1 and Appendix 3.B.5), definitive results are not available.

The PDI index of the total power dissipation (Emanuel, 2005a; see also Box 3.5) showed substantial upward trends beginning in the mid-1970s. Because the index depends on wind speed cubed, it is very sensitive to data quality, and the initial Emanuel (2005a) report has been revised to show the PDI increasing by about 75% (versus about 100%) since the 1970s (Emanuel, 2005b). The increase comes about because of longer storm lifetimes and greater storm intensity, and the index is strongly correlated with tropical SST. These relationships have been reinforced by Webster et al. (2005) who found a large increase in numbers and proportion of hurricanes reaching categories 4 and 5 globally since 1970 even as total number of cyclones and cyclone days decreased slightly in most basins. The largest increase was in the North Pacific, Indian and Southwest Pacific oceans.

These studies have been challenged by several scientists (e.g., Landsea, 2005) who have questioned the quality of the data and the start date of the 1970s. The historical record typically records the central pressure and the maximum winds, but these turn out not to be physically consistent in older records (mainly prior to about the early 1970s). However, attempts at mutual adjustments result in increases in some years and decreases in others, with little effect on overall trends. In particular, in the satellite era after about 1970, the trends found by Emanuel (2005a) and Webster et al. (2005) are robust (Emanuel, 2005b). There is no doubt that active periods have occurred in the more distant past, notably in the North Atlantic (see below), but the PDI was not as high in the earlier years (Emanuel, 2005a). In the Atlantic and west Pacific combined, the power dissipation index (Emanuel, 2005a) confirms the reality of higher values in recent decades in strong association with higher SSTs (Emanuel, 2005b).

There is a clear El Niño connection in most regions, and strong negative correlations between regions in the Pacific and Atlantic, so that the total tropical storm activity is more nearly constant than ACE values in any one basin. With El Niño, the incidence of hurricanes typically decreases in the Atlantic (Gray, 1984; Bove et al., 1998) and far western Pacific and Australian regions, while it increases in the central North and South Pacific and especially in the western North Pacific typhoon region (Gray, 1984; Lander, 1994; Chan and Liu, 2004; Kuleshov and de Hoedt, 2003), emphasizing the change in locations for tropical storms to preferentially form and track with ENSO. Formation and tracks of tropical storms favour either the Australian or South Pacific region depending on the phase of ENSO (Basher and Zheng, 1995; Kuleshov and de Hoedt, 2003), and these two regions have been combined.

It is also possible to sum the ACE values over all regions and produce a global value. Although this has been done, it is not shown, as it is not considered sufficiently reliable. However, the highest ACE year through 2005 is 1997, when a major El Niño event began and surface temperatures were subsequently the highest on record (see Section 3.2), and this is followed by 1992, a moderate El Niño year. Such years tend to contain low values in the Atlantic, but much higher values in the Pacific, and they highlight the critical role of SSTs in the distribution and formation of hurricanes. 1994 is third, while 2004 and 2005 are close to the 1981–2000 mean. Emanuel's (2005a) power dissipation index also peaks in the late 1990s about the time of the 1997–1998 El Niño for the combined Atlantic and West Pacific regions, although 2004 is almost as high. Webster et al. (2005) find that numbers of intense (cat. 4 and 5) hurricanes after 1990 are much greater than from 1970 to 1989.

[INSERT FIGURE 3.8.3 HERE]

3.8.3.1 Western North Pacific

In the western North Pacific, long-term trends are masked by strong inter-decadal variability for 1960 to 2003 (Chan and Liu, 2004), but results also depend on the statistics used. Further increases in activity have occurred in the last few years after Chan and Liu (2004) was completed (Figure 3.8.3). Tropical cyclones making landfall in China are a small fraction of the total storms, and no obvious long-term trend can be discerned (He et al., 2003; Liu and Chan, 2003; Chan and Liu, 2004). Emanuel (2005a) and Webster et al. (2005), however, indicate that the typhoons have become more intense in this region, with more than a doubling of values of the PDI since the 1950s and an increase of order 30% in number of category 4 and 5 storms from 1990–2004 compared with 1975–1989.

The main modulating influence on tropical cyclone activity in the western North Pacific appears to be ENSO (Liu and Chan, 2003; Chan and Liu, 2004). However, the change in the atmospheric circulation associated with ENSO is the dominant factor in hurricane activity and not local SSTs (Chan and Liu, 2004). In El Niño years tropical cyclones tend to be more intense and longer-lived than in La Niña years (Camargo and Sobel, 2004) and occur in different locations. In the summer (JJA) and fall (SON) of strong El Niño years, tropical cyclone numbers increase markedly in the southeastern quadrant of the western North Pacific (0ºN–17ºN, 140°E–180°E) and decrease in the northwestern quadrant (17°N–30°N, 120°E–140°E) (Wang and Chan, 2002). In SON of El Niño years from 1961 to 2000 significantly fewer tropical cyclones made landfall in the western North Pacific compared with neutral years although in Japan and the Korean Peninsula no statistically significant change was detected. In contrast, in SON of La Niña years significantly more landfalls have been reported in China (Wu et al., 2004). Overall in 2004, when a weak El Niño occurred, the number of tropical depressions, tropical storms and typhoons was slightly above the 1971–2000 median. The number of typhoons (21), however, was well above the median (17.5) and second highest to 1997, when 23 developed. Moreover, a record number 10 tropical cyclones or typhoons made landfall in Japan; the previous record was 6 (Levinson, 2005). The ACE index was very close to normal for the 2005 season (Fig 3.8.3).

3.8.3.2 North Atlantic

The North Atlantic hurricane record begins in 1851 and is the longest among cyclone series. Values are considered fairly reliable, however, only after about 1950 when measurements from reconnaissance aircraft began. Methods of estimating wind speed from aircraft have evolved over time and, unfortunately, changes were not always well documented. The record is most reliable after the early 1970s (Landsea, 2005). The North Atlantic record shows a fairly active period from the 1930s to the 1960s followed by a less active period in the 1970s and 1980s, similar to the fluctuations of the AMO (Figure 3.6.8).

Beginning with 1995 all but two Atlantic hurricane seasons have been above normal (relative to the 1981–2000 base period). The exceptions are the two El Niño years of 1997 and 2002. As noted in Section 3.8.3, El Niño acts to reduce activity and La Niña acts to increase activity in the North Atlantic. The increased activity after 1995 contrasts sharply with the generally below-normal seasons observed during the previous 25-year period 1970–1994. These multi-decadal fluctuations in hurricane activity result nearly entirely from differences in the number of hurricanes and major hurricanes forming from tropical storms first named in the tropical Atlantic and Caribbean Sea. The change from the negative phase of the AMO in the 1970s and 1980s (see Section 3.6.6.1) to the post-1995 period has been a contributing factor to the increased hurricane activity (Goldenberg et al., 2001) and is well depicted in Atlantic SSTs (Figure 3.6.8), including those in the tropics. Nevertheless, it appears that most of the recent warming is associated with global SST increases rather than the AMO.

During 1995–2004, hurricane seasons averaged 13.6 tropical storms, 7.8 hurricanes, 3.8 major hurricanes, and have an average ACE index of 159% of the median. The record-breaking 2005 season is documented in more detail in Box 3.6.6. NOAA classifies all but two of these seasons (the exceptions being the El Niño years of 1997 and 2002) as above normal (the average over 1981–2000). In contrast, during the preceding 1970–1994 period, hurricane seasons averaged 8.6 tropical storms, 5 hurricanes, and 1.5 major hurricanes, and had an average ACE index of only 70% of the median. NOAA classifies twelve (almost one-half) of these 25 seasons as being below normal, and only three as being above normal (1980, 1988, 1989), with the remainder as normal. The positive phase of the AMO was also present during the above-normal hurricane decades of the 1950s and 1960s, as indicated by comparing Atlantic SSTs (Figure 3.6.8) and seasonal ACE values (Figure 3.8.3). In 2004, there were 15 named storms, of which 9 were hurricanes and an unprecedented four hit Florida, causing extensive damage (Levinson, 2005). In 2005, record high SSTs (Figure 3.6.8) and favourable atmospheric conditions enabled the most active season on record (by many measures), but this was not fully reflected in the ACE index; see Box 3.6.6. In 2005 the North Atlantic ACE was 3rd highest since 1948.

Key factors in the recent increase in Atlantic activity (Chelliah and Bell, 2004) include (1) warmer SSTs across the tropical Atlantic, (2) an amplified subtropical ridge at upper levels across the central and eastern North Atlantic, (3) reduced vertical wind shear in the deep tropics over the central North Atlantic, which results from an expanded area of easterly winds in the upper atmosphere and weaker easterly trade winds in the lower atmosphere, and (4) a configuration of the African easterly jet that favours hurricane development from tropical disturbances moving westward from the African coast. The vertical shear in the main development region where most Atlantic hurricanes form (Aiyyer and Thorncroft, 2006) fluctuates interannually with ENSO, and also with a multi-decadal variation that is correlated with Sahel precipitation. The latter switched sign around 1970 and remained in that phase until the early 1990s, consistent with the AMO variability. It has been argued that the QBO is also a factor in interannual variability (Gray, 1984). The most recent decade has the highest SSTs on record in the tropical North Atlantic (Figure 3.6.8), apparently as part of global warming and a favourable phase of the AMO. Generally in the Atlantic, the changing environmental conditions (Box 3.5) have been more favourable for tropical storms to develop in the past decade. In 2004 the Power Dissipation Index (Emanuel, 2005a) was by far the highest on record since 1930 for the North Atlantic (not counting 2005).

3.8.3.3 Eastern North Pacific

Tropical cyclone activity (both frequency and intensity) in this region is related especially to SSTs, the phase of ENSO, and the phase of the QBO in the tropical lower stratosphere. Above normal tropical cyclone activity during El Niño years and lowest activity typically associated with La Niña years is the opposite of the North Atlantic basin (Landsea et al., 1998). Tropical cyclones tend to attain a higher intensity when the QBO is in its westerly phase at 30 hPa in the tropical lower stratosphere. A well-defined peak in the seasonal ACE occurred in early 1990s, with the largest annual value in 1992 (Figure 3.8.3), but values are unreliable prior to 1970 in the pre-satellite era. In general, seasonal hurricane activity, including the ACE index, has been below average since 1995, with the exception of the El Niño year of 1997, and is inversely related to the observed increase in activity in the North Atlantic basin over the same time period. This pattern is associated with the AMO (Levinson, 2005) and ENSO. There has, nevertheless, been an increase in category 4 and 5 storms (Webster et al., 2005).

3.8.3.4 Indian Ocean

The North Indian Ocean tropical cyclone season extends from May-December, with peaks in activity during May-June and November when the monsoon trough lies over tropical waters in the basin. Tropical cyclones are usually short-lived and weak, quickly moving into the subcontinent. Tropical storm activity in the northern Indian Ocean has been near normal in recent years (Figure 3.8.3).

The tropical cyclone season in the South Indian Ocean is normally active from December through April and thus the data are summarized by calendar year in Figure 3.8.3, rather than by season. The basin extends from the African coastline, where tropical cyclones impact Madagascar, Mozambique and the Mascarene Islands, including Mauritius, to 110°E (tropical cyclones east of 110°E are included in the Australian summary), and from the Equator southward, although most cyclones develop south of 10°S. The intensity of tropical cyclones in the South Indian Ocean is reduced during El Niño events (Figure 3.8.3) (Levinson, 2005). The estimated ACE index for four of the last five years has been close to or slightly above average (Figure 3.8.3). Lack of historical record keeping severely hinders trend analysis.

3.8.3.5 Australia and the South Pacific

The tropical cyclone season in the South Pacific-Australia region typically extends over the period November through April, with peak activity from December through March. Tropical cyclone activity in the Australian region (105°E–160°E) has declined somewhat over the past decade (Figure 3.8.3); although this may be partly due to improved analysis and discrimination of weak cyclones that previously were estimated at minimum tropical storm strength (Plummer et al., 1999). Increased cyclone activity in the Australian region has been associated with La Niña years; while below normal activity has occurred during El Niño years (Plummer et al., 1999; Kuleshov and de Hoedt, 2003). In contrast in the South Pacific, the opposite signal has been observed, and the most active years have been associated with El Niño events, especially during the strong 1982–1983 and 1997–1998 El Niños (Levinson, 2005), and maximum ACE values occurred in January-March 1998 (Figure 3.8.3). Webster et al. (2005) found more than a doubling in the numbers of category 4 and 5 hurricanes in the southwest Pacific region from 1975–1989 to 1990–2004.

3.8.3.6 South Atlantic

In late March 2004 in the South Atlantic, off the coast of Brazil, the first and only documented hurricane in that region occurred (Pezza and Simmonds, 2005). It came ashore in the Brazilian state of Santa Catarina on 28 March 2004 with winds, estimated by the U.S. National Hurricane Center, of near 40 m s^{-1}, causing much damage to property and some loss of life (see Levinson, 2005). The Brazilian meteorologists dubbed it 'Catarina'. This event appears to be unprecedented although records are poor before the satellite era. Pezza and Simmonds (2005) suggest that a key factor in the hurricane development was the more favourable atmospheric circulation regime associated with the positive trend in SAM (see Section 3.6).

3.8.4 Evidence for Changes in Extratropical Storms and Extreme Events

3.8.4.1 Extratropical cyclones

Extratropical cyclones are low pressure systems fuelled by horizontal temperature gradients in the mid-latitudes of both hemispheres: they act to reduce these gradients. Intense extratropical cyclones are generally accompanied by severe windstorms. Significant increases in the number or strength of intense extra-tropical cyclone systems have been documented in a number of studies (e.g., Lambert, 1996; Gustafsson, 1997; McCabe et al., 2001; Wang et al., 2006) with associated changes in the preferred tracks of storms as described in Section 3.5.3. As with tropical cyclones, detection of long-term changes in cyclone measures is hampered by incomplete and changing observing systems. Some earlier results have been questioned because of changes in the observation system (e.g., Graham and Diaz, 2001).

Results from NRA and ERA-40 show that an increase in the number of deep cyclones is apparent over the North Pacific and North Atlantic (Graham and Diaz, 2001; Gulev et al., 2001), but only the North Pacific trend is statistically significant (Simmonds and Keay, 2002; Wang et al., 2006). Geng and Sugi (2001) find that cyclone density, deepening rate, central pressure gradient, as well as translation speed, have all been increasing in the winter North Atlantic. Caires and Sterl (2005) compare global estimates of 100-year return values of wind speed and significant wave height in ERA-40, with linear bias corrections based on buoy data, for three different 10-year periods. They show that the differences in the storm tracks can be attributed to decadal variability in the NH, linked to changes in global circulation patterns, most notably to the NAO; see also Section 3.5.3.

Significant decreases in cyclone numbers, and increases in mean cyclone radius and depth over the southern extratropics over the last two or three decades (Simmonds and Keay, 2000; Keable et al., 2002; Simmonds, 2003; Simmonds et al., 2003) have been associated with the observed contraction and strengthening of the southern polar vortex, and are likely related to decreased rainfall along the mid-latitude storm track axis and a circumpolar signal of increased precipitation off the Antarctic coast (Cai et al., 2003) and a drying trend observed in southwestern Australia (Karoly, 2003). Using NCEP-2 reanalysis data, Lim and Simmonds (2002) show that for 1979–1999, trends in the annual number of explosively-developing extra-tropical cyclones are significant in the SH and over the globe (0.56 and 0.78 more systems per year, respectively), while the positive trend does not achieve significance in the NH. Simmonds and Keay (2002) obtained similar results for the change in the number of cyclones in the decile for deepest cyclones averaged over the North Pacific and over the North Atlantic in winter over the period 1958–1997.

Besides reanalysis data, station data may also be used to find evidence for changes in extra-tropical cyclone activity. Instead of direct station wind measurements, which may suffer from a lack of consistency of instrumentation, methodology and exposure, values based on pressure gradients have been derived which are more reliable for discerning long-term changes. Alexandersson et al. (2000) used station pressure observations for 21 stations over northwestern Europe back to about 1880, from which geostrophic winds were calculated using 'pressure-triangle' methods. They found a decline of storminess expressed by the 95 and 99 percentiles from high levels during the late-19th century to a minimum around 1960 and then a quite rapid increase to a maximum around 1990, followed again by a decline (Figure 3.8.4). Positive NAO winters are typically associated with more intense and frequent storms (see Section 3.6.4). Similar results were obtained by Schmith et al. (1998) using simpler indices based on pressure tendency. Bärring and von Storch (2004) using both pressure tendencies and the number of very low pressure values, confirm these results on the basis of two especially long station series in southern Sweden dating back to 1800. A study of rapid pressure changes at stations indicates an increase in the number and intensity of severe storms over the southern U.K. since the 1950s, but a decrease over Iceland (Alexander et al., 2005). Thus the station pressure data for parts of the North Atlantic region show a modest increase in severe storms in recent decades. However, decadal-scale fluctuations of similar magnitude have occurred earlier in the 19th and 20th century.

Direct surface wind measurements have, however, been used in a few studies. An analysis of extreme pressure differences and surface winds (Salinger et al., 2005) over the southern part of New Zealand and the oceans to the south showed a significant increasing trend over the last 40 years in westerly winds extremes. The trends are consistent with the increased frequency of El Niño events in recent decades, associated with Pacific decadal variability (see Section 3.6.3). While the zonal pressure gradient and extreme westerly wind frequency have both increased over southern New Zealand, the frequency of extreme easterly winds has also increased there, suggesting more variability in the circulation generally. However, trends in pressure differences (based on the NRA and station data) are not always consistent with changes in surface windiness (e.g., Smits et al., 2005). Based on observed 10 m winds over the Netherlands, they find a decline in strong wind events over the last 40 years. Differences cannot entirely be explained by changes in surface aerodynamic roughness, and Smits et al. (2005) conclude that inhomogeneities in the reanalyses are the cause. However, local differences can be important and intensity and severity of storms may not always be synonymous with local extreme surface winds and gusts.

[INSERT FIGURE 3.8.4 HERE]

3.8.4.2 Tornadoes, hail, thunderstorms, dust storms, and other severe local weather

Evidence for changes in the number or intensity of tornadoes entirely relies on local reports. In the United States, databases for tornado reporting are well established, although Brooks et al. (2003) raise doubt whether the tornado record is complete enough to render the observed weak increase in the number of tornado days significant. Trapp et al. (2005) also question the completeness of the tornado record and argue that about 12% of squall-line tornadoes remain unreported. In many European countries, the number of tornado reports has increased considerably over the last five years (Snow, 2003; Tyrrell, 2003), which led to a much higher estimate of tornado activity (Dotzek, 2003). The increase in Germany between 1950 and 2003 mainly concerns weak tornadoes (F0 and F1 on the Fujita scale), thus paralleling the evolution of tornado reports in the United States after 1950 (cf. Dotzek et al., 2005) and making it likely that the increase in reports in Europe is at least dominated (if not solely caused) by enhanced detection and reporting efficiency. Meehl et al. (2000), Easterling et al. (2000) and Doswell et al. (2005) highlight the difficulties encountered when trying to find observational evidence for changes in extreme events on local scales connected to severe thunderstorms. In the light of the very strong spatial variability of small-scale severe weather phenomena, the density of surface meteorological observing stations is too coarse to measure all such events. Moreover, homogeneity of existing station series is questionable. While remote sensing techniques allow detection of thunderstorms even in remote areas, they do not always uniquely identify severe weather events from these storms.

Although a decreasing trend of dust storms was observed from mid-1950s to mid-1990s in northern China, the number of dust storm days increased from 1997 to 2002 (Li and Zhai, 2003; Zhou and Zhang, 2003). The decreasing trend appears linked to the reduced cyclone frequency and increasing winter (DJF) temperatures (Qian et al., 2002). The recent increase is associated with vegetation degradation and drought, plus increased surface wind speed. (Zou and Zhai, 2004; Wang and Zhai, 2004).

Box 3.6: Specific Extreme Events

Single extreme events cannot be simply and directly attributed to anthropogenic climate change, as there is always a finite chance the event in question might have occurred naturally. When a pattern of extreme weather persists for some time, it may be classed as an extreme climate event, perhaps associated with anomalies in SSTs (such as El Niño). In the following, examples are given of some recent (post-TAR) notable extreme climate events. A lack of long and homogeneous observational data often makes it difficult to place some of these events in a longer-term context. The odds may have shifted to make some of them more likely than in an unchanging climate, but attribution of the change in odds typically requires extensive model experiments; a topic taken up in Chapter 9. It may be possible, however, to say that the occurrence of these events is consistent with expectations arising from climate change.

Box 3.6.1 Drought in Central and Southwest Asia, 1998–2003

Between 1999 and 2003 a severe drought hit much of southwest Asia, including Afghanistan, Kyrgyzstan, Iran, Iraq, Pakistan, Tajikistan, Turkmenistan, Uzbekistan and parts of Kazakhstan (Waple and Lawrimore, 2003; Levinson and Waple, 2004). Most of the area is a semiarid steppe, receiving precipitation only during winter and early spring through orographic capture of eastward propagating mid-latitude cyclones from the Atlantic Ocean and the Mediterranean Sea (Martyn, 1992). Precipitation between 1998 and 2001 was on average less than 55% of the long-term average, making the drought conditions in 2000 the worst in 50 years (Waple et al., 2002). By June 2000, some parts of Iran had reported no measurable rainfall for 30 consecutive months. In December 2001 and January 2002 snowfall at higher altitudes brought relief for some areas, although the combination of above-average temperatures and early snowmelt, substantial rainfall, and hardened ground desiccated by prolonged drought resulted in flash flooding during spring in parts of northern Afghanistan, Tajikistan, and central and southern Iran. Other regions in the area continued to experience drought through 2004 (Levinson, 2005). In these years, an anomalous ridge in the upper-level circulation was a persistent feature during the cold season in central and southern Asia. The pattern served to both inhibit the development of baroclinic storm systems and deflect eastward-propagating storms to the north of the drought-affected area. Hoerling and Kumar (2003) have linked the drought in certain areas of the mid-latitudes to common global oceanic influences. Both the prolonged duration of the 1998–2002 cold phase ENSO (La Niña) event and the unusually warm ocean waters in the western Pacific and eastern Indian Oceans appear to contribute to the severity of the drought (Nazemosadat and Cordery, 2000; Barlow et al., 2002; Nazemosadat and Ghasemi, 2004).

Box 3.6.2 Drought in Australia, 2002–2003

A severe drought affected much of Australia during 2002, associated with the moderate El Niño event (Watkins, 2002). However, droughts in 1994 and 1961 were about as dry as the 2002 drought. Earlier droughts at the start of the 20th century may well have been even drier. Also, the 2002 drought came after several years of good rainfall (averaged across the country) rather than during an extended period of low rainfall such as occurred in the 1930s. Thus the 2002 drought does not provide evidence of droughts becoming more extreme, if only rainfall deficit is considered. However, daytime temperatures during the 2002 drought did set records for protracted warmth and were much higher than previously during droughts. The mean annual maximum temperature for 2002 was 0.5°C warmer than during the 1994 drought and 0.9°C warmer than during the 1961 drought. So, in this sense, the 2002 drought and associated heat waves was more extreme than the earlier droughts, because the impact of the low rainfall was exacerbated by high potential evaporation (Karoly et al., 2003; Nicholls, 2004). The very high temperatures during 2002 could not simply be attributed to the low rainfall, although there is a strong negative correlation between rainfall and temperature. Severe drought, stemming from at least three years of rainfall deficits, continued during 2005 especially in the eastern third of Australia, even as rains brought some relief in June 2005. These conditions also have been accompanied by record high maximum temperatures over Australia during the first half of 2005 (a comparable national series is only available since 1951).

Box 3.6.3 Drought in Western North America, 1999–2004

The western United States, southern Canada, and northwest Mexico experienced a recent pervasive drought (Lawrimore et al., 2002), with dry conditions commencing as early as 1999 and persisting through the end of 2004 (Figure 3.8.5). Drought conditions were recorded by several hydrologic measures including precipitation, streamflow, lake and reservoir levels and soil moisture (Piechota et al., 2004). The period 2000–2004 was the first instance of five consecutive years of below average flow on the Colorado River since the beginning of modern records in 1922 (Pagano et al., 2004). Cook et al. (2004) provide a longer-term context for this drought. In the western conterminous United States, the area under moderate to extreme drought, as given by the PDSI, rose above 20% in November 1999 and stayed above this level persistently until October 2004. At its peak (August 2002), this drought affected 87% of the West (Rockies westward), making it the second most extensive and one of the longest droughts in the last 105 years. The impacts of this drought have been exacerbated by depleted or earlier than average melting of the mountain snowpack, due to warm springs, as observed changes in timing from 1948 to 2000 trended earlier by one to two weeks in many parts of the West (Cayan et al., 2001; Stewart et al., 2005; Regonda et al., 2005). Within this episode, the spring of 2004 was unusually warm and dry, resulting in record early snowmelt in several western watersheds (Pagano et al., 2004).

Hoerling and Kumar (2003) attribute the drought to changes in atmospheric circulation associated with warming of the western tropical Pacific and Indian oceans, while McCabe et al. (2004) have produced evidence suggesting that the confluence of both Pacific decadal and Atlantic multidecadal fluctuations is involved. In the northern winter of 2004-05, the weak El Niño was part of a radical change in atmospheric circulation and storm track across the United States, ameliorating the drought in the Southwest, although lakes remain low.

[INSERT FIGURE 3.8.5 HERE]

Box 3.6.4 Floods in Europe, Summer 2002

A catastrophic flood occurred along several central European rivers in August 2002. The floods resulting from extraordinary high precipitation were enhanced by the fact that the soils were completely saturated and the river water levels were already high because of previous rain (Ulbrich et al., 2003ab; Rudolf and Rapp, 2003). Hence it was part of a pattern of weather over an extended period. In the flood, the water levels of the Elbe at Dresden reached a maximum mark of 9.4 m, which is the highest level since records began in 1275 (Ulbrich et al., 2003a). Some small villages in the Ore Mountains (on tributaries of the Elbe) were hit by extraordinary flash floods. The river Vltava inundated the city of Prague before contributing to the Elbe flood. A return period of 500 years was estimated for the flood levels at Prague (Grollmann and Simon, 2002). The central European floods were caused by two heavy precipitation episodes. The first, on 6–7 August was situated mainly over Lower Austria, the southwestern part of the Czech Republic and southeastern Germany. The second took place on 11–13 August 2002 and most severely affected the Ore Mountains and western parts of the Czech Republic. A persistent low pressure system moved slowly from the Mediterranean Sea to central Europe on a path over or near the eastern Alps and led to large-scale, strong and quasi-stationary frontal lifting of air with very high liquid water content. Additional to this advective rain were convective precipitation processes (showers and thunderstorms) and a significant orographic lifting (mainly over the Ore Mountains). A maximum 24-hour-precipitation total of 353 mm was observed at the German station Zinnwald-Georgenfeld, a new record for Germany. The synoptic situation leading to floods is well known to meteorologists of the region. Similar situations led to the summer floods of the River Oder in 1997 and the River Vistula in 2001 (Ulbrich et al., 2003b). Average summer precipitation trends in the region are negative but barely significant (Schönwiese and Rapp, 1997) and there is no significant trend in flood occurrences of the Elbe within the last 500 years (Mudelsee et al., 2003). However, the observed increase in precipitation variability at a majority of German precipitation stations during the last century (Trömel and Schönwiese, 2005) is indicative of an enhancement of the probability of both floods and droughts.

Box 3.6.5 Heat Wave in Europe, Summer 2003

The heat wave that affected many parts of Europe during the course of summer 2003 produced record-breaking temperatures particularly during June and August (Beniston, 2004; Schär et al., 2004), see Figure 3.8.6. Absolute maximum temperatures exceeded the record temperatures observed in the 1940s and early 1950s in many locations in France, Germany, Switzerland, Spain, northern Italy and the United Kingdom according to the information supplied by national weather agencies (WMO, 2004). Gridded instrumental temperatures (from HadCRUT2v for the region 35–50°N, 0–20°E) show that the summer was the warmest since comparative records began in 1780 (3.8°C above the 1961–1990 average) and 1.4°C warmer than any other summer in this period (next warmest 1807). Based on early documentary records, Luterbacher et al. (2004) estimate that 2003 is likely to have been the warmest summer since 1500. The 2003 heat wave was associated with a very robust and persistent blocking high pressure system that may be a manifestation of an exceptional northward extension of the Hadley Cell (Fink et al., 2004; Black et al., 2004). Already a record month in terms of maximum temperatures, June exhibited high geopotential values that penetrated northwards towards the British Isles, with the greatest northward extension and longest persistence of record-high temperatures observed in August. An exacerbating factor for the temperature extremes was the lack of precipitation in many parts of western and central Europe, leading to much-reduced soil moisture and surface evaporation and evapotranspiration, and thus to a strong positive feedback effect (Beniston and Diaz, 2004).

[INSERT FIGURE 3.8.6 HERE]

Box 3.6.6. *The 2005 Tropical Storm season in the North Atlantic*

The 2005 North Atlantic hurricane season (1 June to 30 November) was the most active on record by several measures, surpassing the very active season of 2004 (e.g., Levinson, 2005) and causing an unprecedented level of damage. Even before the peak in the seasonal activity, the 7 tropical storms in June and July were the most ever, and hurricane Dennis was the strongest on record in July and the earliest ever fourth named storm. The record 2005 North Atlantic hurricane season featured the largest number of named storms (27) (sustained winds over 17 m s^{-1}) and is the only time names have ventured into the Greek alphabet. It had the largest number of hurricanes (14) (sustained winds >33 m s^{-1}) recorded, and is the only time there has been three category 5 storms (maximum sustained winds >67 m s^{-1}). These included the most intense Atlantic storm on record (Wilma, recorded surface pressure in the eye 882 hPa), the most intense storm in the Gulf of Mexico (Rita, 897 hPa), and the most damaging storm on record (Katrina). Tropical storm Vince was the first to ever make landfall in Portugal and Spain. In spite of these metrics, the ACE index, although very high and surpassing the 2004 value (Figure 3.8.3), was not the highest on record, as several storms were quite short lived. Six of the eight most damaging storms on record for the United States occurred from August 2004 to September 2005 (Charlie, Ivan, Francis, Katrina, Rita, Wilma) while another storm in 2005 (Stan) caused severe flooding and mudslides as well as about 2000 fatalities in central America (Guatemala, El Salvador and southern Mexico).

In addition, SSTs in the tropical North Atlantic region critical for hurricanes (10° to 20°N) were at record high levels in the extended summer (June to October) of 2005 at 0.9°C above the 1901–1970 normal (Figure 3.6.8) and these high values were a major reason for the very active hurricane season, along with favourable atmospheric conditions (see Box 3.5). A substantial component of this warming was the global mean SST increase (see Section 3.2).

3.8.5 Summary

Even though our archived datasets are not yet sufficient for determining long-term trends in extremes, there are new findings on observed changes for different types of extremes. The definitions of the phenomena are summarized in Table 3.7. A summary of the changes in extremes by phenomena, region and time is given in Table 3.8 along with an assessment of the confidence.

New analyses since the TAR confirm the picture of a gradual reduction of the number of frost days over most of the mid-latitudes in recent decades. In agreement with this warming trend, the number of warm nights increased between 1951 and 2003 in 72% of the global land area sampled. Cold night decreases were slightly greater at 76%. Changes in the number of cold and warm days also show warming, but the trends are less marked than at night.

For precipitation, analysis of updated trends and results for regions that were missing at the time of the TAR show increases in heavy events for the majority of observation stations, with some increase in flooding (Milly et al., 2002). This result applies both for areas where total precipitation has increased and for areas where total precipitation has even decreased. Increasing trends are also reported for more rare precipitation events, although results for such extremes are available only for a few areas. Mainly because of lack of data, it remains difficult to draw a consistent picture of changes in extreme precipitation for the tropics and subtropics.

Tropical cyclones, hurricanes and typhoons exhibit large variability from year to year in individual ocean basins. Limitations in the quality of data in most basins shorten series and compromise evaluations of trends. Nonetheless, clear evidence exists for increases in category 4 and 5 storms globally since 1970 along with increases in the PDI due to increases in intensity and duration of storms. The records are probably most complete in the Atlantic, where higher decadal mean SSTs have accompanied increased hurricane activity in the past decade, but this is offset by decreased activity in the eastern North Pacific associated with the AMO and ENSO. The 2005 season in the North Atlantic broke many records: most named storms, the most intense storms in the Atlantic (Wilma) and in the Gulf of Mexico (Rita) and the most damaging storm on record (Katrina). The global view of tropical storm activity highlights the important role of ENSO in all basins, and the most active year is 1997, when a very strong El Niño began, suggesting that observed record high SSTs played a key role.

For extratropical cyclones, positive trends in storm frequency and intensity dominate for recent decades in most regional studies performed. Longer records for the northeastern Atlantic suggest that the recent extreme period maybe similar in level to that of the late-19th century.

As noted in 3.3.4, the PDSI shows a large drying trend from the middle of the 20th century over NH land areas since the mid-1950s and a drying trend in the SH from 1974 to 1998. Decreases in land precipitation since the 1950s are the main cause for the drying trends, although large surface warming during the last 2–3 decades has also likely contributed to the drying. Globally very dry areas, defined as land areas with the PDSI less than –3.0, appear to have more than doubled since the 1970s, with a large jump in the early 1980s due to an ENSO-induced precipitation decrease and subsequent increases primarily due to surface warming.

Table 3.7. Definition of phenomena used to assess extremes in Table 3.8.

PHENOMENON	Definition
Cold days / cold nights / frost days	Percentage of days with temperature not exceeding some threshold, either fixed (frost days) or varying regionally (cold days / cold nights), based on the 10th percentile of the daily distribution in the reference period (1961–1990)
Warm days / warm nights	See cold days / cold nights, but now exceeding the 90th percentile
Cold spells	Episode of several consecutive cold days / cold nights / frost days
Warm spells (heat waves)	Episode of several consecutive warm days / warm nights
Cold seasons / warm seasons	Seasonal averages (rather than daily temperatures) exceeding some threshold
Heavy precipitation events (events that occur every year)	Percentage of days (or daily precipitation amount) with precipitation exceeding some threshold, either fixed or varying regionally, based on the 95th or 99th percentile of the daily distribution in the reference period (1961–1990)
More rare precipitation events (with return periods >~10yr)	As heavy precipitation events, but for extremes further into the tail of the distribution
Drought (season / year)	Precipitation deficit; or based on PDSI; see Box 3.1
Tropical cyclones *(frequency, intensity, track, peak wind, peak precipitation)*	Tropical storm with thresholds crossed in terms of estimated wind speed and organization. Hurricanes in categories 1 to 5, according to the Saffir-Simpson scale, are defined as storms with wind speeds of 33 to 42 m s^{-1}, 43 to 49 m s^{-1}, 50 to 58 m s^{-1}, 58 to 69 m s^{-1}, and >69 m s^{-1},respectively. NOAA's Accumulated Cyclone Energy (ACE) index is a measure of the total seasonal activity that accounts for the collective intensity and duration of Atlantic tropical storms and hurricanes during a given hurricane season.
Extreme extra tropical storms *(frequency, intensity, track, surface wind, wave height)*	Intense low pressure systems that occur throughout the mid-latitudes of both hemispheres fuelled by temperature gradients and acting to reduce them.
Small-scale severe weather phenomena	Extreme events on local scales, severe thunderstorms, hail storms, tornadoes, water spouts, dust storms,

Table 3.8. Change (column 2) in extremes for phenomena (column 1) over the region (column 3) for the period (column 4), with the confidence given (column 5) and where discussed in detail (column 6).

PHENOMENON	Change	Region	Period	Confidence	Section
Cold days / cold nights / frost days	Decrease, more so for nights than days	Over 70% of global land area	1951–2003 (last 150 years for Europe and China)	Very likely	3.8.2.1
Warm days / warm nights	Increase, more so for nights than days	Over 70% of global land area	1951–2003	Very likely	3.8.2.1

PHENOMENON	Change	Region	Period	Confidence	Section
Cold spells (episodes of several days)	Insufficient studies, but daily temperature changes imply a decrease.				
Warm spells (heat waves) (episodes of several days)	Increase: implicit evidence from changes of daily temperatures.	Global	1951–2003	Likely	Question 3.3
Cold seasons / warm seasons (seasonal averages)	Some new evidence for changes in inter-seasonal variability	Central Europe, Argentina	1961–2004	Likely	3.8.2.1
Heavy precipitation events (that occur every year)	Increase, generally beyond that expected from changes in the mean (disproportionate)	Many mid-latitude regions (even where reduction in total precipitation)	1951–2003	Likely	3.8.2.2
More rare precipitation events (with return periods > ~10yr)	Increase	Only a few regions have sufficient data for reliable trends (e.g. UK and U.S.)	Various since 1893	Likely (consistent with changes inferred for more robust statistics)	3.8.2.2
Drought (season / year)	Increase	Many land areas of the world	since 1970s	Very likely	3.3.4 and Question 3.3
Tropical cyclones	No trend in number; positive trends toward longer lifetimes and greater storm intensity	Tropics	Since 1970s	Likely; more confidence in frequency and intensity than track	3.8.3 and Question 3.3
Extreme extra tropical storms	Net increase frequency/intensity and poleward shift in track	NH land	Since about 1950	Likely	3.8.4, 3.5, and Question 3.3
Small-scale severe weather phenomena	Insufficient studies for assessment				

3.9 Synthesis: Consistency Across Observations

Here, we briefly compare variability and trends within and across different climate variables to see if a physically-consistent picture enhances our confidence in the realism of apparent recent observed changes. So we look ahead to following observational chapters on the cryosphere (Chapter 4) and oceans (Chapter 5), which focus on changes in those domains. The emphasis here is on inter-relationships. For example, increases in temperature should enhance the moisture-holding capacity of the atmosphere as a whole and changes in temperature and/or precipitation should be consistent with those evident in circulation indices. Increases in temperature should also reduce snow seasons, sea ice and cause glacier retreat in many areas. The main sections where more detailed information can be found are given in square brackets following each bullet.

- Decadally-smoothed global-mean surface temperatures show overall warming of 0.8°C over the 1850–2005 period; rates of temperature rise are much greater after 1979. Linear trend estimates indicate a warming of 0.65°C over the same time interval, but a linear rise in temperature is a poor approximation to what has happened. Both land surface air temperatures and SST show warming although land regions have warmed at a faster rate than the oceans for both hemispheres in the past few decades, consistent with the much greater thermal inertia of the oceans and differences in surface characteristics. A few areas have cooled in recent decades but not significantly. [Sections 3.2.2, 3.3.5; Question 3.1]

- The warming of the climate is consistent with a widespread reduction in the number of frost days in mid-latitude regions. The latter is due to an earlier last day of frost in spring rather than a later start to the frost season in autumn. Cold extremes have declined over more than 70% of land regions studied from 1951–2003, with slightly greater changes at night compared to daytime. Increases in the number of warm extremes have also occurred, but not to such a large extent. . [Section 3.8.2.1; Question 3.3]
- Widespread (but not ubiquitous) decreases in continental DTR since the 1950s occur with increases in cloud amounts, as expected from the impact of cloud cover on solar heating of the surface. The rate of decrease of DTR overall has likely reduced to negligible values when considered over the 1979–2004 period. [Sections 3.2.2, 3.4.3]
- The temperature increases are consistent with the observed nearly worldwide reduction in glacier and small ice cap (not including Antarctica and Greenland) mass and extent in the 20th century. Glaciers and ice caps respond not only to temperatures but also changes in precipitation, and both winter accumulation and summer melting have increased over the last half century in association with temperature increases. In some regions moderately increased accumulation observed in recent decades is consistent with changes in atmospheric circulation and associated increases in winter precipitation (e.g., southwestern Norway, parts of coastal Alaska, Patagonia, Karakoram, and Fjordland of the South Island of New Zealand) even though enhanced ablation has led to marked declines in mass balances in Alaska and Patagonia. Tropical glacier changes are synchronous with higher latitude ones and all have shown declines in recent decades. Local temperature records all show a slight warming, but not of the magnitude required to explain the rapid reduction in mass of such glaciers (e.g., on Kilimanjaro). Other factors in recent ablation include changes in cloudiness, water vapour, albedo due to snowfall frequency and the associated radiation balance. [Chapter 4, Section 4.5]
- Snow cover has decreased in many NH regions, particularly in spring, consistent with greater increases in spring as opposed to autumn temperatures in mid-latitude regions. These changes are consistent with changes in permafrost, whose temperature has increased by up to 3ºC since the 1980s in the Arctic and Subarctic and permafrost warming is also observed on the Tibetan Plateau and the European mountain permafrost regions. Active layer thickness has increased and seasonally frozen ground depth has decreased over the Eurasian continent. [Chapter 4, Sections 4.2.4, 4.8]
- Sea-ice extents have decreased in the Arctic, particularly in spring and summer, and patterns of the changes are consistent with regions showing a temperature increase, although changes in winds are also a major factor. Sea-ice extents were at record low values in 2005, which was also the warmest year since records began in 1850 for the Arctic north of 65°N. There have also been decreases in sea-ice thickness. In contrast to the Arctic, Antarctic sea ice does not exhibit any significant trend since the end of the 1970s, which is consistent with the lack of trend in surface temperature south of 65°S over that period. However, along the Antarctic Peninsula, progressive break up of ice shelves has occurred beginning in the late 1980s, culminating in the break up of the Larsen-B ice shelf in 2002. Decreases are found in the length of the freeze season of river and lake ice. [Section 3.2.2.3; Chapter 4, Sections 4.3, 4.4; Chapter 5, Section 5.3.3]
- Surface temperature variability and trends since 1979 are broadly consistent with those estimated by most analyses of satellite retrievals of lower-tropospheric temperatures, provided the latter are adequately adjusted for all issues of satellite drift, orbit decay, changes of satellites, and stratospheric influence on the records, and also with ERA-40 estimates of lower-tropospheric temperatures. The range from different datasets of global surface warming since 1979 is 0.16 to 0.18 compared to 0.12 to 0.19°C decade^{-1} for estimates of lower tropospheric temperatures. [Sections 3.2.2, 3.4.1; Question 3.1]
- Stratospheric temperature estimates from radiosondes, satellites and reanalyses all show a cooling of between 0.3 and 0.8°C decade^{-1} since 1979, but the likely true range is a cooling of 0.3 to 0.5°C decade^{-1} because of cooling biases affecting radiosondes and one reanalysis. [Section 3.4.1]
- Increasing evidence suggests increasing warming with altitude from 1979 to 2005 from the surface through much of the troposphere in the tropics, cooling in the stratosphere, and a higher tropopause. [Section 3.4.1]

- Radiation changes at the top of the atmosphere from the 1980s to 1990s, possibly ENSO-related in part, appear to be associated with reductions in tropical cloud cover, and are linked to changes in the energy budget at the surface and in observed ocean heat content in a consistent way. [Sections 3.4.3, 3.4.4, 3.6.2]
- Surface specific humidity has also generally increased after 1976 in close association with higher temperatures over both land and ocean. Consistent with a warmer climate, total column water vapour has increased over the global oceans by 1.2 ± 0.3% from 1988 to 2004, consistent in patterns and amount with changes in SST and a fairly constant relative humidity. Upper tropospheric water vapour has also increased in ways such that relative humidity remains about constant, providing a major positive feedback to radiative forcing. [Section 3.4.2]
- Over land a strong negative correlation is observed between precipitation and surface temperature in summer and in low latitudes throughout the year, and areas that have become wetter, such as the eastern United States, have not warmed as much as other land areas. Increased precipitation is associated with increases in cloud and surface wetness, and thus increased evapotranspiration. The inferred enhanced evapotranspiration and reduced temperature increase is physically consistent with enhanced latent versus sensible heat fluxes from the surface in wetter conditions. [Section 3.3.5; Question 3.2; Section 3.4.4.2]
- Widespread (but not ubiquitous) decreases in continental DTR since the 1950s coincide with increases in cloud amounts. Surface observations of cloud cover changes over land exhibit coherent variations on interannual to decadal time scales which are positively correlated with gauge-based precipitation measurements and changes in atmospheric circulation, such as the NAO. [Sections 3.2.2, 3.4.3, 3.6.4]
- Reported decreases in solar radiation from 1970 to 1990 at the surface have an urban bias and have reversed. Although records are sparse, pan evaporation is estimated to have decreased in many places due to decreases in surface radiation associated with increases in clouds, changes in cloud properties, and/or increases in air pollution (aerosol) in different regions, especially from 1970 to 1990. There is evidence to suggest that this has reversed in recent years. [Chapter 2, Sections 2.4.5, 2.4.6; Sections 3.3.3, 3.4.4; Chapter 7, Sections 7.2.1, 7.5.3]
- Consistent with rising amounts of water vapour in the atmosphere, increases in the numbers of heavy precipitation events (e.g., 95th percentile) have been reported from many land regions, even those where there has been a reduction in total precipitation. Increases have also been reported for rarer precipitation events (1 in 50 year return period), but only a few regions have sufficient data to assess such trends reliably. [Sections 3.4.2, 3.8.2.2]
- Patterns of precipitation change are much more spatially and seasonally-variable than temperature change, but where significant changes do occur they are consistent with measured changes in streamflow. [Section 3.3.4]
- Droughts have increased in various parts of the world. The regions where they have occurred seem to be determined largely by changes in SSTs, especially in the tropics, through changes in the atmospheric circulation and precipitation. Inferred enhanced evapotranspiration and drying associated with warming are additional factors in increases in droughts, but decreased precipitation is the dominant factor. In the western United States, diminishing snow pack and subsequent summer soil moisture reductions have also been a factor. In Australia and Europe, direct links to warming have been inferred through the extreme nature of high temperatures and heat waves accompanying drought. [Section 3.3.4, Question 3.2; Box 3.6, Chapter 4, Section 4.2.2]
- Changes in the freshwater balance of the Atlantic Ocean over the past four decades have been pronounced as freshening has occurred in the North Atlantic and also south of 25°S, while salinity has increased in the tropics and subtropics, especially in the upper 500 m. The implication is that there have been increases in moisture transport by the atmosphere from the subtropics to higher latitudes, in association with changes in atmospheric circulation, including the NAO, thereby producing the observed increases in precipitation over the northern oceans and in adjacent land areas. [Sections 3.3.2, 3.6.4; Chapter 5, Sections 5.3.2, 5.5.3]
- Changes in extra-tropical atmospheric circulation are predominantly observed as “annular modes”, related to the zonally averaged mid-latitude westerlies, which have strengthened in most seasons from 1979 to the late 1990s, together with poleward displacements of corresponding jet streams and enhanced storm tracks and a tendency toward stronger wintertime polar vortices throughout the troposphere and lower stratosphere. In the NH, the NAM and NAO change the flow from oceans to

continents and are a major cause of the wintertime observed change in precipitation and temperature patterns, especially over Europe and North Africa. In the SH, SAM changes, in association with the ozone hole, have been identified with recent contrasting trends of large warming in the Antarctic Peninsula, and cooling over interior Antarctica. [Sections 3.5, 3.6, 3.8.4]

- The 1976–1977 climate shift toward more El Niños has affected Pacific Ocean islands and Pacific rim countries and monsoons throughout the tropics. Changes are linked in diverse regions ranging from wetter conditions in southeastern South America to drier conditions in southern Asia and southern Africa. Over North America, ENSO and PNA-related changes have led to contrasting changes across the continent, as the west has warmed more than the east, while the latter has become cloudier and wetter. [Sections 3.6, 3.7]
- Variations in extratropical storminess are mostly strongly associated with changes and variations in ENSO, NAO, PDO, and SAM. Wind and significant wave height analysis support the evidence for an increase in extratropical storm activity in the NH in recent decades. After the mid-1990s, however, some of these variations have diminished. Longer European records suggest that high levels of storm activity there also occurred in the late 19th century. [Sections 3.5, 3.6, 3.8.4]
- Tropical cyclones have increased in intensity and duration since the 1970s with a large increase in numbers and proportion of hurricanes reaching categories 4 and 5 globally even as total number of cyclones and cyclone days decreased slightly in most basins. Changes are observed to occur in the number, distribution and tracks of tropical storms that are clearly related to ENSO phases and to a slightly lesser extent to the AMO and QBO modulations. Changes are strongly related with tropical SST. The 2005 hurricane season in the North Atlantic broke many records, with the most named storms and storms with the deepest recorded low pressures in the Atlantic Ocean and Caribbean Sea. [Section 3.8.3.1]
- Sea level likely rose about 18 ± 3 cm during the 20th century, but increased 3.1 ± 0.4 mm yr^{-1} after 1992, when confidence increases from global altimetry measurements. During this period, glacier melt has increased ocean mass by approximately 1.0 mm yr^{-1}. Increases in ocean heat content and associated ocean expansion are estimated to contribute 0.4 ± 0.1mm yr^{-1} in the last 50 years, increasing to an estimated range of 1.3 to 1.8 mm yr^{-1} for 1993 to 2003. Changes in land water storage are uncertain but may have reduced water in the ocean. Isostatic rebound contributes about 0.3 mm/year. This near balance gives increased confidence that the observed sea level rise is a strong indicator of warming, and an integrator of the cumulative energy imbalance at the top of atmosphere. [Chapter 4, Sections 4.5, 4.7, 4.9.8; Chapter 5, Sections 5.2, 5.5]

In summary, global mean temperatures have increased since the 19th century, especially since the mid-1970s. Temperatures have increased nearly everywhere over land, and SSTs have also increased, reinforcing the evidence from land. However, temperatures have neither increased monotonically, nor in a spatially uniform manner, especially over shorter time intervals. The atmospheric circulation has also changed: in particular increasing zonal flow is observed in most seasons in both hemispheres, and the mid/high latitude annular modes have strengthened. In the NH this has brought milder maritime air into Europe and much of high-latitude Asia from the North Atlantic in winter, enhancing warming there. In the SH, where the ozone hole has played a role, it has resulted in cooling over the interior of Antarctica but large warming in the Antarctic Peninsula region and Patagonia. Temperatures generally have risen more than average where flow has become more poleward, and less than average or even cooled where flow has become more equatorward, reflecting PDO and other patterns of variability.

The three main ocean basins are individually unique and contain very different wind systems, SST patterns and currents, leading to vastly different variability associated, for instance, with ENSO in the Pacific, and the THC in the Atlantic. Consequently the oceans have not warmed uniformly, especially at depth. SSTs in the tropics have warmed at different rates and help drive, through coupling with tropical convection and winds, teleconnections around the world. This has changed the atmospheric circulation through ENSO, the PDO, the AMO, monsoons, and the Hadley and Walker circulations. Changes in precipitation and storm tracks are not as well documented but clearly respond to these changes on interannual and decadal timescales. When precipitation increases over the ocean, as it has in recent years in the tropics, it decreases over land, although it has increased over land at higher latitudes. Droughts have increased over many tropical and mid-latitude land areas, in part because of decreased precipitation over land since the 1970s but also from increased evapotranspiration arising from increased atmospheric demand associated with warming.

Changes in the cryosphere, ocean and land strongly support the view that the world is warming through observed decreases in snow cover and sea ice, thinner sea ice, shorter freezing seasons of lake and river ice, glacier melt, decreases in permafrost extent, increases in soil temperatures and borehole temperature profiles (see Chapter 6), and sea level rise.

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Question 3.1: How are Temperatures on the Earth changing?

Generally temperatures at the surface have risen, but with important variations regionally and with time. For the global average, warming in the last century has occurred in two phases, from the 1910s–1940s (0.35°C) and more strongly from the 1970s to the present (0.55°C), but with 0.1°C cooling between. Balloon temperature estimates since 1958 for the lower atmosphere (~2–8 km) show slightly greater overall global warming rates than the surface, and the warming evolves differently. Estimates of global lower atmospheric warming since 1979 from satellites and balloons range from somewhat less to slightly greater than that at the surface; uncertainties in the observing system are not yet fully resolved.

Expressed as a global average, surface temperatures have increased by about 0.8°C since the late-19th century; see Question 3.1, Figure 1. A series is produced by combining air temperature observations taken at several thousand stations over the land areas of the world with measurements of sea surface temperature (SST) from the world's oceans. Since 1981, *in situ* SST measurements have been augmented with satellite estimates in data-sparse regions. The warming has been neither steady nor the same in each season or in different locations. A linear approximation over the 20th century is a very poor assumption. There was an increase (0.35°C) in the global average from the 1910s to the 1940s, a slight cooling (0.1°C) from then to the 1970s followed by a rapid warming (0.55°C) at least up to the end of 2005 (see Question 3.1, Figure 1). The warmest year of the series was recorded in 1998 and 10 of the 11 warmest years have occurred in the last eleven complete years (1995–2005). Warming, particularly in the most recent phase, has been greater over land regions than the oceans. Seasonally, warming has been slightly greater in the winter hemisphere, although this is very dependent on the specific period analysed.

A few areas have cooled since 1901, most notably the northern North Atlantic near southern Greenland. Warming during this time has been strongest over the continental interiors of Asia and northern North America. As these are areas with large year-to-year variability, the most significant warming has occurred in parts of the middle and lower latitudes, particularly the tropical oceans. Up to about 1990, warming had been most evident as a reduction in the number of anomalously cold months rather than more frequent warm monthly values.

A number of recent studies indicate that effects of urbanization and land-use change on the land-based temperature record (since 1950) are negligible as far as hemispheric- and continental-scale averages are concerned, because the very real but local effects are either accounted for or avoided by the selection of stations satisfying international siting standards. Increasing evidence suggests that urban effects extend to changes in precipitation, cloud and also diurnal temperature range with the latter detectable as a "weekend effect" owing to reduced pollution at weekends.

Analysis of long-term changes in extreme daily temperatures has only been possible for regions of the world (parts of North and southern South America, Europe, northern and eastern Asia, southern Africa and Australia) where long, reliable and digital records of daily temperature data are available for the entire period since 1901. There is considerable spatial variability, but these records generally show a decrease in the number of extreme cold days and nights and to a slightly lesser extent an increase in the number of extreme warm days and nights. Daily data are much more spatially extensive and digitally available from about 1950 and show similar features for most but again not all regions. Since 1950, the length of frost-free season has increased in most mid-to-high latitude regions of both hemispheres. In the NH this is mostly manifest as an earlier start to spring rather than later frosts in the autumn.

Global temperature trends aloft since 1958 have been estimated from improved analyses of balloon data which have good vertical resolution, though coverage of the Southern Hemisphere oceans has remained sparse and the data in general have required adjustment for multiple changes in instruments and observing practices. Satellite estimates for deep layers from MSU instruments provide complete global coverage since 1979, but have come from 13 different instruments, and despite several new analyses with improved cross-calibration of those instruments and compensation for changes in observing time and satellite altitude, there remain uncertainties regarding long-term changes. Although lower atmospheric temperatures differ from surface temperatures, they are generally physically well-connected and are expected to show related trends. Estimates of global lower atmospheric (~2-8 km) warming since 1979 from adjusted satellite and balloon data range from somewhat less to slightly greater than that at the surface, owing to incomplete resolution of the uncertainties due to inadequacies in the observing system: see Question 3.1, Figure 1. The balance of evidence suggests that the tropical lower atmosphere has warmed slightly less than the surface since 1979, though some estimates show equal warming. Balloon data show warming in the tropical upper troposphere (~10 km), and both radiosonde and MSU data indicate cooling in the tropical and global stratosphere, consistent with model results. Balloon temperature estimates for the lower atmosphere since 1958 show slightly greater overall global warming rates than the surface, but the warming evolves differently.

[INSERT QUESTION 3.1, FIGURE 1 HERE]

Question 3.2: How is Precipitation Changing?

Observations show that changes are occurring in some parts of the globe in the amount, intensity, frequency, and type of precipitation. These aspects of precipitation generally show a large natural variability, and El Niño, and changes in atmospheric circulation patterns such as the North Atlantic Oscillation (NAO), have a substantial influence. Pronounced long-term trends from 1900 to 2005 have been observed in some places: significantly wetter in Argentina, eastern North America, northern Europe and northern Asia, but drier in the Sahel, southern Africa, and southern Asia. Widespread increases in heavy precipitation have been observed, even in places where total amounts have decreased. These changes are associated with increased water vapour in the atmosphere arising from the warming of the world's oceans, especially in lower latitudes. There are also increases in some regions in the occurrences of both droughts and floods, and shifts in some regions to wetter or drier conditions.

Precipitation is the general term for rainfall, snowfall, and other forms of frozen or liquid water falling from clouds. Precipitation is intermittent, and the character of the precipitation when it occurs depends greatly on temperature and the weather situation. The latter determines the supply of moisture through winds and surface evaporation, and how it is gathered together in storms as clouds. Precipitation forms as water vapour is condensed, usually by rising air that expands and hence cools. The upward motion comes from air rising over mountains, warm air riding over cooler air (warm front), colder air pushing under warmer air (cold front), convection from local heating of the surface, and other weather and cloud systems. Hence changes in any of these aspects will alter precipitation. Overall trends in precipitation are indicated by the Palmer Drought Severity Index (See Question 3.2, Figure 1, and Box 3.1), and this also factors in crude estimates of changes in evaporation.

[INSERT QUESTION 3.2, FIGURE 1 HERE]

As climate changes, several direct influences alter precipitation amount, intensity, frequency and type. A consequence of increased heating from the anthropogenic greenhouse effect is enhanced evaporation provided that adequate surface moisture is available (as it always is over the oceans and other saturated surfaces). Hence surface moisture effectively acts as an "air conditioner", as heat used for evaporation acts to moisten the air rather than warm it. An observed consequence of this is that summers often tend to be either warm and dry or cool and wet. Warming accelerates land-surface drying and increases the potential incidence and severity of droughts, which is observed to be happening worldwide. A well established physical law (the Clausius-Clapeyron relation) determines that the water holding capacity of the atmosphere increases by about 7% for every 1°C rise in temperature. Observations suggest that relative humidity remains about the same overall, from the surface throughout the troposphere, and hence increased temperatures result in increased water vapour, in part from the increased drying at the surface. Over the 20th century, based on changes in sea surface temperatures, it is estimated that atmospheric water vapour increased by about 5% in the atmosphere over the oceans. Because precipitation comes mainly from weather systems that feed on the water vapour stored in the atmosphere this generally means enhanced precipitation intensity and risk of heavy precipitation events. Basic theory, climate model simulations, and empirical evidence all confirm that warmer climates, owing to increased water vapour, lead to more intense precipitation events even when the total annual precipitation reduces slightly, and with prospects for even stronger events when the overall precipitation amounts increase. The warmer climate therefore increases risks of both drought and floods, but at different times and/or places. For instance, the summer of 2002 in Europe brought widespread floods but was followed a year later by record breaking heat waves and drought. The distribution and timing of floods and droughts is most profoundly affected by the cycle of El Niño events, particularly in the tropics and over much of the mid-latitudes of North and South America.

In areas where aerosol pollution masks the ground from direct sunlight, decreases in evaporation reduce the overall moisture supply to the atmosphere. Hence even as the potential for heavier precipitation occurs, the duration and frequency of events may be curtailed, as it takes longer to recharge the atmosphere with water vapour. Local and regional changes in the character of precipitation also depend a great deal on atmospheric circulation patterns determined by El Niño, the NAO, and other patterns of variability. An associated shift in the storm track makes some regions wetter and some – often nearby – drier, making for complex patterns of change. For instance in the European sector, a more positive NAO in the 1990s led to wetter conditions in northern Europe and drier conditions over the Mediterranean and Northern African regions. The prolonged drought in the Sahel (See Question 3.2, Figure 1), which was pronounced from the late 1960s to the late 1980s, continues although it is not quite as intense as it was; it has been linked to changes in tropical sea surface temperature patterns in the Pacific, Indian and Atlantic basins. Drought has become widespread throughout much of Africa and more common in the tropics and subtropics.

As temperatures rise, the likelihood of precipitation falling as rain rather than snow increases, especially in fall and spring at the beginning and end of the snow season, and in areas where temperatures are near freezing. Such changes are observed in many places, especially over land in middle and high latitudes of the Northern Hemisphere, leading to increased rains but reduced snow-packs, and diminished water resources in summer, when they are most needed. Nevertheless, the often spotty and intermittent nature of precipitation means observed patterns of change are complex. The long-term paleo record emphasizes that patterns of precipitation vary somewhat from year to year, and even prolonged multi-year droughts are usually punctuated by a year of heavy rains; for instance as El Niño influences are felt. An example may be the wet winter of 2004–2005 in the southwestern United States following a 6-year drought and below normal snow-pack.

Question 3.3: Has there Been a Change in Extreme Events like Heat Waves, Floods, Droughts, and Hurricanes?

Since 1950 heat waves have increased and widespread increases have occurred in warm nights. Droughts also have increased as precipitation over land has marginally decreased while evapotranspiration has been enhanced by warmer conditions. Generally, heavy daily precipitation events have increased but not everywhere. Tropical storm and hurricane frequencies vary considerably from year to year, but evidence suggests substantial increases in intensity and duration since the 1970s. In the extratropics, variations in tracks and intensity of storms reflect variations in major features of the atmospheric circulation, such as the North Atlantic Oscillation.

In several regions of the world, indications of a change in various types of extreme climate events have been found. The distribution of many variables, such as temperature and daily precipitation amount, exhibit some kind of bell curve when the frequency of values in narrow intervals is plotted. The extremes are considered to be the values at the low and high ends (or tails) of the distribution, and it is common to consider the percentiles exceeded 1%, 5% and 10% of the time along with the 90%, 95% and 99% values. The 50th percentile is the median, which for a symmetric distribution also constitutes the mean. In this Question, the focus is on how these extremes are changing rather than the mean. The warm nights or days (discussed below) are those exceeding the 90th percentile, while cold nights or days are those less than the 10th percentile. Heavy (very heavy) precipitation is defined as daily amounts greater than the 95th (99th) percentile.

Over 70% of the global land area sampled has shown a significant decrease in the annual occurrence of cold nights and a significant increase in the annual occurrence of warm nights (Question 3.3, Figure 1) associated with a positive shift in the distribution of daily minimum temperature. Decreases in the occurrence of cold days and increases in warm days are generally less marked. The distributions of minimum and maximum temperatures have not only shifted to higher values, consistent with overall warming, but also changed in shape such that the cold extremes have warmed more than the warm extremes over the last 50 years (Question 3.3, Figure 1). Despite the greater increase in the cold-tail of the distribution, more warm extremes imply an increased frequency of heat waves. Further indications of a robust change include the observed trend to fewer frost days associated with the average warming in most mid-latitude regions.

A prominent indication of a change in extremes is the evidence of increases in moderate to heavy precipitation events over the mid-latitudes in the last 50 years, even in places where mean precipitation amounts are not increasing (see also Question 3.2). For more rare precipitation events, increasing trends are reported as well but results for such extremes are available only for few areas.

Drought is one of the simplest extremes to measure because of its long duration and thus less dependence on large volumes of high-frequency data. Most studies use monthly precipitation totals and temperature averages combined into a measure called the Palmer Drought Severity Index (PDSI). The PDSI calculated from the middle of the 20th century shows a large drying trend over NH land areas since the mid-1950s, with widespread drying over much of Eurasia, northern Africa, Canada and Alaska (Question 3.2, Figure 1a), but a reverse trend in eastern North America. In the SH, land surfaces were wet in the 1970s and relatively dry in the 1960s and 1990s; and there was a drying trend from 1974 to 1998 although trends over the entire 1948–2002 period were small. Longer duration records for Europe for the whole of the 20th century indicate few significant trends. Decreases in precipitation over land since the 1950s are the likely main cause for the drying trends, although large surface warming during the last 2–3 decades has also likely contributed to the drying. Based on the PDSI data, one study shows that the extent of very dry land areas across the globe, defined as areas with PDSI less than –3.0, have more than doubled since the 1970s, with a large jump in the early 1980s due to an El Niño/Southern Oscillation (ENSO)-induced precipitation decrease over land and subsequent increases primarily due to surface warming.

Trends in tropical storm and hurricane frequency and intensity are masked by large natural variability on multiple timescales. El Niño events greatly affect the location and activity of tropical storms around the world. Globally, estimates of the potential destructiveness of hurricanes show a substantial upward trend since the mid-1970s, with a trend toward longer lifetimes and greater storm intensity, and the index is strongly correlated with tropical SST. These relationships have been reinforced by findings of a large increase in numbers and proportion of hurricanes reaching categories 4 and 5 globally since 1970 even as total numbers of cyclones and cyclone days decreased slightly in most basins. The largest increase was in the North Pacific, Indian and Southwest Pacific oceans. However, numbers of hurricanes in the North Atlantic have also been above normal (based on 1981–2000) in 9 of the last 11 years, culminating in the record breaking 2005 season.

Based on a variety of measures at the surface and in the upper troposphere, it is likely that there has been an increase and a poleward shift in NH winter storm track activity over the second half of the 20th century. The variations that have occurred in the NH are related to the North Atlantic Oscillation. Observations from 1979 to the late 1990s reveal a tendency toward stronger DJF polar vortices throughout the troposphere and lower stratosphere, together with poleward displacements of jetstreams and enhanced storm track activity. Observational evidence for changes in small-scale severe weather phenomena (such as tornadoes, hail and thunderstorms) is mostly local and too scattered to draw general conclusions; increases in many areas simply arise because there are more people to observe these phenomena.

[INSERT QUESTION 3.3, FIGURE 1]

Appendix 3.A: Low Pass Filters and Linear Trends

The time series used in this report have undergone diverse quality controls which have, for example, led to removal of outliers, thereby building in some smoothing. In order to highlight decadal and longer timescale variations and trends, it is often desirable to apply some kind of low-pass filter to the monthly, seasonal or annual data. In the literature cited for the many indices used in this chapter, a wide variety of schemes were employed. Here we have used the same filter wherever reasonable to do so. The desirable characteristics of such filters are 1) they should be easily understood and transparent; 2) they should avoid introducing spurious effects such as ripples and ringing (Duchon, 1979); 3) they should remove the high frequencies, and 4) they should involve as few weighting coefficients as possible, in order to minimize end effects. The classic low-pass filters widely used have been the binomial set of coefficients which remove 2Δt fluctuations, where Δt is the sampling interval. However, combinations of binomial filters are usually more efficient, and we have chosen to use these here, for their simplicity and ease of use. Mann (2004) discusses smoothing time series and especially how to treat the ends. We choose to use the 'minimum slope' constraint at the beginning and end of all time series, which effectively reflects the time series about the boundary. If there is a trend, then this will be conservative in the sense that it will underestimate the anomalies at the end.

The first filter (e.g., Fig. 3.2.5) is used in situations where only the smoothed series is shown and it is designed to remove interannual fluctuations and those on El Niño timescales. It has 5 weights 1/12[1-3-4-3-1] and its response function (ratio of amplitude after to before) is 0.0 at 2 and 3Δt, 0.5 at 6Δt, 0.69 at 8Δt, 0.79 at 10Δt, 0.91 at 16Δt, and 1 for zero frequency, so for yearly data (Δt =1) the half amplitude point is for a 6-year period, and the half power point is near 8.4 years.

The second filter used in conjunction with annual values (Δt =1) or for comparisons of multiple curves (e.g., Fig. 3.2.8) is designed to remove less than decadal fluctuations. It has 13 weights 1/576 [1-6-19-42-71-96-106-96-71-42-19-6-1]. Its response function is 0.0 at 2, 3 and 4Δt, 0.06 at 6Δt, 0.24 at 8Δt, 0.41 at 10Δt, 0.54 at 12Δt, 0.71 at 16Δt, 0.81 at 20Δt, and 1 for zero frequency, so for yearly data the half amplitude point is about a 12-year period, and the half power point is 16 years. This filter has a very similar response function to the 21-term binomial filter (used in the TAR).

Another low pass filter, widely used and easily-understood, is to fit a linear trend to the time series. Nevertheless, there is generally no physical reason why trends should be linear, especially over long periods. The overall change in the time series is often inferred from the linear trend over the given time period, but can be quite misleading. Also, such measures are typically not stable and are sensitive to beginning and end points, so that adding or subtracting a few points can result in marked differences in the estimated trend. Furthermore as the climate system exhibits highly non-linear behaviour, alternative perspectives of overall change are provided whereby a comparison is done of low-pass-filtered values (see above) near the beginning and end of the major series.

The linear trends are estimated by Restricted Maximum Likelihood regression (REML, Diggle et al., 1999), and estimates of statistical significance assume that the terms have serially uncorrelated errors and that the residuals have an AR1 structure. The error bars, shown as ±2 standard error ranges, are therefore wider and more realistic than those provided by the standard ordinary least squares technique. If, for example, a century-long series has multi-decadal variability as well as a trend, the deviations from the fitted linear trend will be autocorrelated. This will cause the REML technique to widen the error bars, reflecting the greater difficulty in distinguishing a trend when it is superimposed on other long-term variations, and the sensitivity of estimated trends to the period of analysis in such circumstances. Clearly, however, even the REML technique cannot widen its error estimates to take account of variations outside the sample period of record. While more sophisticated and non-linear methods are available, they are not as transparent. Robust methods for the estimation of linear and nonlinear trends in the presence of episodic components became available recently (Grieser et al., 2002).

As some components of the climate system respond slowly to change, the climate system naturally contains persistence, so that the REML AR1-based linear trend statistical significances are likely to be overestimated (Zheng and Basher, 1999; Cohn and Lins, 2005). Nevertheless, the results depend on the statistical model used, and more complex models are not as transparent and often lack physical realism.

Appendix 3.B: Techniques, Error Estimation and Measurement Systems

3.B.1 Methods of Temperature Analysis: Global Fields and Averages

The first step in creating representative global gridded datasets is to take account of the number and error-characteristics of the observations within individual grid-boxes, reducing the variance of grid-box values if they are based on sparse or unreliable data, and yielding uncertainty estimates for each grid-box value (Jones et al., 2001; Rayner et al., 2006; Smith and Reynolds, 2005; Brohan et al., 2006). Grid-box values have been (a) used to create maps of trends over specified periods and (b) combined with areal weighting to derive regional, continental, hemispheric and global time series. A number of maps and time series are shown in Section 3.2, all with temperatures expressed as anomalies or departures from 1961–1990. Estimates of actual temperatures can be retrieved by adding back the climatologies to the anomaly data (Jones et al., 1999). Estimates of uncertainties of time series values must involve an estimate of the number of spatial degrees-of-freedom, as only a fraction of all the observations are statistically independent (see Jones et al., 1997, 2001; Rayner et al., 2006; Brohan et al., 2006). Vinnikov et al. (2004) have also presented a new technique for analysis of diurnal and seasonal cycles and trends, in which anomalies are calculated only implicitly.

The effects of changes in coverage over the instrumental period (now since 1850 for global scales) were first assessed by the “frozen grid” and theoretical approaches (see Jones et al., 1997, 1999, 2001). Subsequently, Reduced-Space Optimal Interpolation (RSOI) has been used to infill incomplete and noisy fields and to provide local error estimates (Kaplan et al., 1997; Rayner et al., 2003). Optimal averaging (OA) yields large-area averages with error-bars (Folland et al., 2001). Global estimates are less reliable before 1900 (by a factor of two) than since 1951, but this is principally expressed on the interannual timescale. The sparser grids of the late-19th century estimate decadal and longer-timescale averages for periods since 1940 very reliably. RSOI and OA use the major patterns of variability (such as that associated with El Niño), to account for areas with no observations. The patterns are derived using data for recent, well-sampled years, and the technique relies on the assumption that the same patterns occurred throughout the record. Hence it depends on the stationarity of the record and this is a questionable assumption given known climate change. If the regions affected by a pattern are sparsely sampled, the pattern is accorded reduced weight in the analysis and error estimates are augmented. Neither RSOI nor OA can reproduce trends reliably (Hurrell and Trenberth, 1999); the data must therefore first be detrended by, for example, using the covariance matrix to estimate the temperature anomaly pattern associated with global warming, and removing the projection of this pattern from the data. After the techniques have been applied to the residuals, the trend component is restored.

Vose et al. (2005b) show that estimates of global land surface air temperature trends are affected less by local data coverage than by the choice between a weighted grid-box average for the globe and the average of the weighted grid-box averages for the two hemispheres. This underscores the value of the OA technique which takes optimal account of unsampled regions. However, because OA assumes zero anomaly in the absence of information, it yields global anomalies of smaller magnitude than other techniques when data are sparse (Hurrell and Trenberth, 1999).

In addition to errors from changing coverage and from random measurement and sampling errors, errors arise from biases (see Section 3.B.2 on homogenizing records). Major efforts have been made to adjust for known systematic biases, but some adjustments nonetheless are quite uncertain. Nevertheless, recent studies have estimated all the known errors and biases to develop error bars (Brohan et al., 2006). For example, for SSTs, the transition from taking temperatures from water samples from uninsulated or partially-insulated buckets to engine intakes near or during World War II is adjusted for, even though details are not certain (Rayner et al., 2006).

3.B.2 Adjustments to Homogenize Land Temperature Observations

Long-term temperature data from individual climate stations almost always suffer from inhomogeneities, owing to non-climatic factors. These include sudden changes in station location, instruments, thermometer housing, observing time, or algorithms to calculate daily means; and gradual changes arising from instrumental drifts or from changes in the environment due to urban development or land use. Most abrupt changes tend to produce random effects on regional and global trends, and instrument drifts are corrected by routine thermometer calibration. However, changes in observation time (Vose et al., 2004) and urban development are likely to produce systematic biases; for example, relocation may be to a cooler site out of town (Böhm et al., 2001). Urbanization usually produces warming, although examples exist of cooling in arid areas where irrigation effects dominate.

When dates for discontinuities are known, a widely used approach is to compare the data for a target station with neighbouring sites, and the change in the temperature data due to the non-climatic change can be calculated and applied to the pre-move data to account for the change, if the discontinuity is statistically significant. However, often the change is not documented, and its date must be determined by iterative tests. The procedure moves through the time series checking the data before and after each value in the time series (this works for monthly or longer means, but not daily values owing to greater noise at weather timescales) (e.g., Easterling and Peterson, 1995; Vincent, 1998; Menne and Williams, 2005). An extensive review is given by Aguilar et al. (2003).

Trends averaged over small regions, in particular, may be biased by systematic heterogeneities in the data (Böhm et al., 2001). However, the impact of random discontinuities on area-averaged values typically becomes smaller as the area or region becomes larger, and is negligible on hemispheric scales (Easterling et al., 1996). Nevertheless, the impact of non-random discontinuities can be important even with large averaging areas. The time-of-observation bias documented by Karl et al. (1986) shows a significant impact even with time series derived for the entire contiguous United States. Adjustments for this problem also remove an artificial cooling that occurs due to a switch from afternoon to morning observation times for the U.S. Cooperative Observer Network (Vose et al., 2004).

Estimates of urban impacts on temperature data have included approaches such as linear regression against population (Karl et al., 1988), and analysis of differences between urban and rural sites defined by vegetation (Gallo et al., 2002) or night lights (Peterson, 2003) as seen from satellites. Urbanization impacts on global and hemispheric temperature trends (Karl et al., 1988; Jones et al., 1990; Easterling et al., 1997; Peterson, 2003; Parker, 2004, 2006) have been found to be small. Furthermore, once the landscape around a station becomes urbanized, long-term trends for that station are consistent with nearby rural stations (Böhm, 1998; Easterling et al., 2005, Peterson and Owen, 2005). However, individual stations may suffer marked biases and require treatment on a case-by-case basis (e.g., Davey and Pielke, 2005); the influence of urban development on temperature depends on local geography and climate so that adjustment algorithms, based on e.g., population, developed for one region may not be applicable in other parts of the world (Hansen et al., 2001; Peterson, 2003).

Homogenization of daily temperature series requires much more metadata than monthly assessment (see the extensive discussion in Camuffo and Jones, 2002) and only a few series can be classed as totally homogeneous. Daily minima and maxima, and consequently also DTR and analysis of extremes, are particularly sensitive to non-climatic heterogeneities, including changes in height above ground, housing and ventilation of instruments (Auer et al., 2001; Brunet et al., 2006a). The ongoing automation of measuring networks is typically accompanied by a change from large and unventilated screens to small and continuously ventilated ones. Assessment of potential homogeneity problems in a network of 60 daily maximum and minimum temperature series, for Europe for the 20th century by Wijngaard et al. (2003), suggests that 94% of series should be classed as of doubtful homogeneity. The percent of doubtful series reduces to 61% when considering 158 series for 1946–1999. Vincent et al. (2002) in a Canadian study of over 200 daily temperature series, develop daily adjustments by smooth interpolation of monthly adjustments.

3.B.3 Adjustments to Homogenize Marine Temperature Observations

Owing to changes in instrumentation, observing environment and procedure, SSTs measured from modern ships and buoys are not consistent with those measured before the early 1940s using canvas or wooden buckets. SST measured by canvas buckets, in particular, generally cooled during the sampling process. Systematic adjustments are necessary (Folland and Parker, 1995; Smith and Reynolds, 2002; Rayner et al., 2006) to make the early data consistent with modern observations that have come from a mixture of buoys, engine inlets, hull sensors and insulated buckets. The adjustments are based on the physics of heat-transfer from the buckets (Folland and Parker, 1995) or on historical variations in the pattern of the annual amplitude of air-sea temperature differences in unadjusted data (Smith and Reynolds, 2002). The adjustments increased between the 1850s and 1940 because the fraction of canvas buckets increased and because ships moved faster, increasing the ventilation. By 1940 the adjustments were 0.4°C for the global average and approached 1°C in winter over the Gulf Stream and Kuroshio where surface heat fluxes are greatest. An atmospheric model, driven by the adjusted SSTs, simulated decadal and longer-term variations of land surface air temperatures on global and continental scales much better than when it was driven by unadjusted SSTs, thus providing strong support to the adjustments (Folland, 2005).

There are smaller biases between modern SSTs taken separately by engine inlets and insulated buckets (Kent and Kaplan, 2006) and between overall ship and buoy observations (Rayner et al., 2006). These biases may arise from the different measurement depths (buckets, typically 30 cm; buoys, typically 1 m; engine inlets, typically 10 m) and from heat inputs from the ship near engine inlets. Biases can also vary by nation. The biases are not large enough to prejudice conclusions about recent warming. The increasing amount of buoy data, although in principle more accurate than most ship measurements, introduces further inhomogeneities (Kent and Challenor, 2006; Kent and Kaplan, 2006), which may have caused an underestimate of recent warming (Rayner et al., 2006). The exact effect on trends of the changes in the methods of measurement in recent decades has not yet been assessed.

Modern observations of SST made *in situ* have been supplemented by satellite-based data since about 1980 giving much better geographical coverage. However, satellite estimates are of skin (infrared) or sub-skin (typically 1 cm, microwave) temperatures, and the infrared data are also affected by biases, especially owing to dust aerosol and to misinterpretation of thin clouds and volcanic aerosols as cool water. Also, instruments on successive satellites are not identical, and instruments in orbit can degrade slowly or show spurious jumps. *In situ* observations have been used to provide calibration for the satellite measurements, which can then been used to fill in the spatial patterns for areas where there are few ships or buoys (Reynolds et al., 2004).

Some efforts have been made to monitor SST from satellite data alone. Lawrence et al. (2004) have compared SSTs from the Pathfinder dataset, which uses the Advanced Very High Resolution Radiometer (AVHRR) with SSTs from the ATSR. The analysed data are not truly global because of problems in distinguishing SST from cloud top temperatures in many regions. Also, the Pathfinder data have time varying biases (Reynolds et al., 2002), and the method for combining data from two different ATSR instruments may need more scrutiny. Nevertheless, the Pathfinder dataset shows similar rates of warming to *in situ* data over 1985–2000. These rates are insignificantly different from the global trend over 1979–2005 from *in situ* data (0.13ºC decade^{-1}) (see Table 3.2). ATSR data also show warming but the period available (1991–2004 with some gaps) is too short to assess a reliable trend (O'Carroll et al., 2006). In future, satellite SST data may be improved by combining infrared and microwave data to provide global coverage where clouds make infrared data unreliable (Wentz et al., 2000; Donlon et al., 2002; Reynolds et al., 2004). The new Global Ocean Data Assimilation Experiment (GODAE) high-resolution SST pilot project (GHRSST-PP) will establish uncertainty estimates (bias and standard deviation) for all satellite SST measurements by careful reference to *in situ* SST observations, accounting for the mixed layer and differences in different bulk and skin temperatures.

Air temperatures taken on board ship have also been biased, mainly because ships have become larger, so that the temperatures were measured typically 6m above the sea in the late-19th century, 15m in the mid-20th century, and over 20m today. In addition, observing practices were irregular during the Second World War and in the 19th century. The data have been adjusted by Rayner et al. (2006). Owing to biases arising from solar heating of ships' fabric, marine air temperature analyses have so far been based on night-time data (Rayner et al., 2003), though Berry et al. (2004) have developed a model for correcting the daytime data. Note that surface air temperatures do not bear a fixed relation to SST: thus, surface heat fluxes in the tropics change with the phase of ENSO, and surface fluxes in the N Atlantic vary with the NAO. However, in many parts of the world oceans and on larger space scales, air temperature and SST anomalies follow each other closely on seasonal and longer time scales (Bottomley et al., 1990).

Many historical *in situ* marine data still remain to be digitized and incorporated into ICOADS (Worley et al., 2005), to improve coverage and reduce the uncertainties in our estimates of past marine climatic variations, but progress has been made since the TAR. The CLIWOC project (Garcia et al., 2005) has digitized an additional 40,000 marine air temperature (MAT) and SST data for the period before 1850. These data, and those of Chenoweth (2000) which have had quality control and bias adjustment, might allow NMAT to be extended back usefully to the early 19th century. Coverage would also be improved if daytime values could be corrected for time-varying daytime biases consistently through the whole dataset (Berry et al., 2004).

3.B.4 Solid/Liquid Precipitation: Undercatch and Adjustments for Homogeneity

3.B.4.1 Precipitation undercatch (snow and rain)

Studies of biases in precipitation measurements by *in-situ* rain gauges (Poncelet 1959; Sevruk 1982; Sevruk and Hamon, 1984; Legates and Wilmott, 1990; Goodison et al., 1998; Golubev et al., 1995, 1999; Bogdanova et al., 2002a,b) find that (a) light rainfall and snowfall are strongly underestimated owing to wind-induced acceleration and vertical motion over the rain gauge orifice (for snowfall, the resulting biases can be as high as 100% of "ground truth" precipitation). The main physical reasons for the observed systematic undercatch of conventional raingauges when exposed to the wind, including the considerably more severe losses of snowfall, were modelled and compared to field observations by Folland (1988). To fix this deficiency, wind-scale correction factors have been developed (cf. Sevruk, 1982; Goodison et al., 1998); (b) most precipitation gauges have trouble reporting the full amount of precipitation that reaches the gauge owing to gauge precision problems (traces), losses (retention, evaporation) and accumulation (condensation) of water in/from the gauge; to fix these deficiencies additive corrections have been developed (cf. Sevruk, 1982; Golubev et al., 1995, 1999) and (c) in windy conditions with snow on the ground, blowing snow enters the gauges causing "false" precipitation; only recently has this factor started to be taken into account in major precipitation datasets in high latitudes (Bryazgin and Dement'ev, 1996; Bogdanova et al., 2002a,b).

After the completion of the International Solid Precipitation Intercomparison Project (Goodison et al., 1998), several attempts to adjust precipitation in high latitudes and create new regional climatologies (Mekis and Hogg, 1999; Yang, 1999; Yang et al., 1999; Yang and Ohata, 2001) and global datasets (Adam and Lettenmaier, 2003) accounted for problems identified in items (a) and (b). However, when this approach was applied to high latitudes (e.g., Yang, 1999; Mekis and Hogg, 1999; Yang and Ohata, 2001), unrealistically high precipitation estimates caused confusion among hydrologists. Critical reassessment of the problem was conducted by Golubev et al. (1995, 1999), Golubev and Bogdanova (1996), and Bogdanova et al. (2002a,b). Universal adjustments have emerged from their studies using parameters of wind speed, gauge and precipitation types, wetting and evaporation adjustments, and flurry and blow-in adjustments. Measured precipitation values are ignored when wind at 10 meters above the snow-covered ground reaches a 10 m s^{-1} threshold and are replaced with estimates of mean regional snowfall intensity and its duration, although this could introduce biases if snowfall rate is correlated with wind strength. A precipitation climatology over the Arctic Ocean (Bogdanova et al., 2002a) using this approach replaces measured annual totals of 128 mm with adjusted annual totals of 165 mm, an increase of 28% over measured values. This climatology corresponds broadly with independent estimates over the Arctic Ocean from aerological and snow cover measurements but is much less than proposed by Yang (1999) for the same region using the same data.

All correction routines suggest higher (in relative terms) adjustments for frozen than for liquid precipitation undercatch. If rising temperature increases the chances for rainfall rather than snowfall, then unadjusted gauges will show precipitation increases owing to the better catch of liquid precipitation. This mechanism was shown to be a major cause of artificially inflated trends in precipitation over the Norwegian Arctic (Førland and Hanssen-Bauer 2000) but it is estimated to have a small effect on the measured precipitation trends in the European Alps (Schmidli et al. 2002).

3.B.4.2 Homogeneity adjustments

Precipitation series are affected by the same sort of homogeneity issues as temperature: random ones due to relocations (both in position and height above the ground), gauge changes, and more spatially consistent effects such as nationwide improvements to gauges and observation practices (Auer et al., 2005). Adjustment of precipitation series at the monthly, seasonal and annual timescale is much more demanding than for temperature, as the spatial correlation of precipitation fields is much weaker. Similar approaches have been tried as for land temperatures, looking at time series of the ratio of the catches at a candidate station to those of neighbours. In many regions, however, the networks are not dense enough to find many statistically significant differences. Auer et al. (2005) for the Greater Alpine Region give typical distances above which adjustments are not possible, these being timescale dependent. Although these are seasonally dependent, distances range from 150 km separation at the monthly to 40 km at the daily timescale. Only a few networks are, therefore, dense enough to consider homogeneity assessment of daily precipitation totals and large-scale studies have rarely been undertaken. In the Wijngaard et al. (2003) study for Europe, the quality of daily precipitation series appears higher than for temperature, perhaps because there were fewer tests that could be applied than for temperature owing to larger natural variability. Only 25% of 88 stations with near-complete records for the 20th century were classed as doubtful, falling to 13% (of 180) for 1946–1999. The reliability of estimated trends in daily extreme precipitation depends on the completeness as well as the homogeneity of the record and is seriously degraded if more than about one third of the daily data are missing (Zolina et al., 2005).

3.B.5 The Climate Quality of Free-Atmosphere and Reanalysis Datasets

3.B.5.1 Evolution of the observing system: radiosondes

Radiosondes measure temperature, humidity and wind speed as they ascend, generally reaching the lower stratosphere before balloons burst. The quality of radiosonde measurements has improved over the past 5 decades, but oceanic coverage has declined owing to the demise of ocean weather ships: spatial and temporal coverage over land has also declined in the 1990s. However, counts of standard-level (e.g., 50 hPa) stratospheric measurements have risen, likely due to better balloons, though there may be remaining biases as balloon bursts still occur more frequently when cold (Parker and Cox, 1995). Many stations have closed, and only a subset of current stations has sufficiently long records to be directly useful for climate monitoring, except through reanalysis. There have been many changes to instrument design and observing practices to improve the accuracy of weather forecasts, and many manufacturers have released multiple radiosonde models. There have also been changes in the radiation corrections applied to account for insolation, in ground equipment, and in calculation methods. Only some of these changes have been documented (Gaffen, 1996 and subsequent updates), and rarely have simultaneous measurements been made to accurately quantify their effects. Developers of Climate Data Records (CDRs) from radiosondes have, therefore, to cope with a highly heterogeneous and poorly documented raw dataset. Since the TAR, efforts have been made to improve global digital databases incorporating more thorough homogeneity and outlier checks (e.g., Durre et al., 2006). Two major efforts to form homogeneous temperature CDRs from these records illustrate the range of possible approaches. Lanzante et al. (2003a, b) (LKS) homogenised data from 87 well-spaced stations using a manually intensive method. They used indicators from the raw data and metadata to try to identify the times of artificial jumps resulting from non-climatic (see above) influences. The resulting homogenised station data series were closer to the only available satellite-based MSU time series at the time (Christy et al., 1998). Thorne et al. (2005a), in contrast, created a global database, HadAT2, containing 676 stations. They used LKS and the GCOS Upper Air Network (GUAN) to define an initial set of 477 adequate stations and then a neighbour comparison technique and metadata to homogenise their data. Subsequently the data from the remaining stations were incorporated in a similar way. The quality control identified an average of about 6 breakpoints per station that required adjustments, 70% of which were not identified with any known change in procedures, while about 29% were identified with changes in sonde or equipment. Most recently, effort has focused on reduction of possibly enhanced biases in daytime radiosonde data (Sherwood et al., 2005). Moisture data from radiosondes generally contain even more complex problems, and no climate quality homogenised databases have yet been produced.

3.B.5.2 Evolution of the observing system: buoys, aircraft and satellite data

Other types of observations have compensated for the decline in radiosonde coverage. New data by 1979 included MSU, HIRS and SSU soundings from satellites. In 1979, winds derived by tracking features observed from geostationary satellites first became available in significant numbers and there were substantial increases in buoy and aircraft data. Overall observation counts declined for a while after 1979, but recovered during the 1980s. The frequency and coverage of wind and temperature measurements from aircraft increased substantially in the 1990s. The launch of the Earth Radiation Budget Experiment (ERBE) in 1984 began a series of satellite instruments that provided the first climate quality record of top-of-atmosphere radiative fluxes. Beginning in 1987, newer satellite-based data from microwave instruments provided improved observations of total water-vapour content, surface wind speed, rain rate, and atmospheric soundings (Uppala et al., 2005).

3.B.5.3 Reanalysis and climate trends

Comprehensive reanalyses from NRA (Kalnay et al., 1996; Kistler et al., 2001), NCEP-2 reanalysis (Kanamitsu et al., 2002) and ERA-15/ERA-40 (Uppala et al., 2005) derived by processing multi-decadal sequences of past meteorological observations using modern data assimilation techniques have found widespread application in many branches of meteorological and climatological research. Care is needed, however, in using them to document and understand climatic trends and low-frequency variations. Atmospheric data assimilation comprises a sequence of analysis steps in which background information for a short period, typically of 6 or 12 hours duration, is combined with observations for the period to produce an estimate of the state of the atmosphere (the "analysis") at a particular time. The background information comes from a short-range forecast initiated from the most-recent preceding analysis in the sequence. Problems for climate studies arise partly because the atmospheric models used to produce these "background forecasts" are prone to biases. If observations are abundant and unbiased, they can correct the biases in background forecasts when assimilated. In reality, however, observational coverage varies over time, observations are themselves prone to bias, either instrumental or through not being representative of their wider surroundings, and these observational biases can change over time. This introduces trends and low-frequency variations in analyses that are mixed with the true climatic signals, making long-timescale trends over the full length of the reanalyses potentially unreliable (Bengtsson et al., 2004; see also http://wwwt.emc.ncep.noaa.gov/mmb/rreanl/). Better representation of trends by reanalysis systems requires progress on identifying and correcting model and observational biases, assimilating as complete a set of past observations as possible, and general improvements to the methods of data assimilation: in this regard the second-generation ERA-40 reanalysis represents a significant improvement over the earlier first generation analyses produced in Europe and the United States.

3.B.5.4 Bias correction for reanalysis

Reliable depiction of temperature trends by a reanalysis requires that changes over time in the biases of the assimilated observations be taken into account, just as they have to be when deriving trend information from radiosonde or MSU data alone. For satellite data, trends in the ERA-40 reanalysis have been affected adversely by difficulties in radiance bias adjustment for the early satellite data. Correcting older radiosonde data for reanalysis is also demanding owing to large, spatially and temporally variable biases and a lack of metadata. In ERA-40 no corrections were applied prior to 1980, but statistics of the difference between the observations and background forecasts are now being used to derive corrections for application both in future reanalyses and in direct trend analysis (Haimberger, 2005).

3.B.5.5 Analysis of tropospheric and stratospheric temperature using microwave radiances

The MSU that has been used for climate monitoring as well as in reanalyses, has been flown continuously since late 1978 (AMSU since 1998) on polar orbiting satellites. Two retrieval channels have been used to create CDRs. MSU channel 2 and its AMSU near-equivalent measure a thick layer of the atmosphere, with approximately 75–80% of the signal coming from the troposphere, 15% from the lower stratosphere, and the remaining 5–10% from the surface. MSU Channel 4 and its AMSU sequel receive their signal almost entirely from the lower stratosphere (see Figure 3.4.1). Each satellite has lasted several years, and usually at least two satellites have been monitoring at roughly 6-hour intervals. Although the instruments are designed to the same specifications for each satellite, MSU instruments have had relative biases of the order 1–2°C. As the orbits have tended to drift, MSU instruments measure at systematically later local times over a satellite's lifetime requiring adjustments to be made for the diurnal cycle, a procedure accommodated automatically in ERA-40 by inserting the observation at the appropriate time. Satellite orbits also tend to decay, affecting the limb soundings of Channel 2 used by UAH to gain a lower tropospheric retrieval (Spencer and Christy, 1992). Finally, there is a suspected, time-varying systematic effect of the instrument body temperature upon the retrievals.

The original set of MSU data records produced by UAH (Christy et al., 2000) has undergone improvement of the correction for diurnal drift, although the effect on trends was small; an error analysis was made and the record was extended to include AMSU measurements (Christy et al., 2003). A new set of data records for channel 2 was constructed by RSS (Mears et al., 2003). Despite starting with identical raw satellite radiances, differences arise between RSS and UAH from the choice of data used to determine the parameters of the calibration target effect. RSS utilizes pentad-mean intersatellite-difference data without further averaging for calculation of the target temperature coefficients. UAH averages daily data into periods of at least 60-days and focuses on reducing low-frequency differences. RSS employs all difference-data, i.e., data from all co-orbiting, overlapping spacecraft, which seeks the statistically best consensus for intersatellite bias determination. UAH omitted very small segments (e.g., 45 days or so) which occur at the tail-ends of the satellites' operational periods, to avoid the use of data segments which are too short for the averaging technique and are near the end of a satellite' lifetime when its biases may be unrepresentative of its full span. The resulting parameters from the UAH procedure for NOAA-9 (1985–1987) were reported to be outside of the physical bounds expected (Mears et al., 2003). Hence the large difference in the calibration parameters for the single instrument mounted on the NOAA-9 satellite accounts for a substantial part (~50%) of the global trend difference between the UAH and RSS results. The rest arises from differences in merging parameters for other satellites, differences in the correction for the drift in measurement time (Mears et al., 2003; Christy and Norris, 2004; Mears and Wentz, 2005), and ways the hot point temperature is corrected for (Grody et al., 2004; Fu and Johanson, 2005). In the tropics, these account for T2 trend differences of order 0.1°C decade^{-1} after 1987 and discontinuities are also present in 1992 and 1995 at times of satellite transitions (Fu and Johanson, 2005).

The T2 data record of Grody et al. (2004) (VG2, see also Vinnokov et al., 2006), which supersedes that of Vinnikov and Grody (2003), uses a zonal mean latitude-dependent analysis that allows for errors that depend on both the calibration target temperature and the atmospheric temperature being measured. Accordingly they point out the need to account for the target effect as a function of latitude, which was not done by UAH or RSS. However they did not account for temporal variations in target temperatures on individual satellites during overlap periods. Furthermore the VG2 method does not fully address the correction for diurnal drift and cannot distinguish between land and ocean.

A new benchmark method for measuring atmospheric temperatures is based on a time measurement using Radio Occultation (RO) from Global Positioning System (GPS) satellites. The promise of this method is revealed by Schroder et al. (2003) who found that UAH T4 retrievals in the Arctic lower stratosphere in winter were biased high relative to temperatures derived from GPS RO measurements.