Chapter 3: Observations: Surface and Atmospheric Climate Change

Coordinating Lead Authors: Philip D. Jones, Kevin E. Trenberth

Lead Authors: Peter G. Ambenje, Roxana Bojariu, David R. Easterling, Albert M. G. Klein Tank, David E. Parker, James A. Renwick, Fatemeh Rahimzadeh, Matilde M. Rusticucci, Brian J. Soden, Pan-Mao Zhai

Contributing Authors: R. Adler, L. Alexander, H. Alexandersson, R. P. Allan, M. P. Baldwin, M. Beniston, D. H. Bromwich, I. Camilloni, C. Cassou, D. R. Cayan, E. K. M. Chang, J. R. Christy, A. Dai, C. Deser, N. Dotzek, R.L. Fogt, C. K. Folland, P. Forster, M. Free, C. Frei, B. Gleason, J. Grieser, P. Y. Groisman, S. K. Gulev, J. W. Hurrell, M. Ishii, S. A. Josey, P. W. Kållberg, G. N. Kiladis, R.H. Kripalani, K. E. Kunkel, C-Y. Lam, J. R. Lanzante, J. H. Lawrimore, D. H. Levinson, B. G. Liepert, G. J. Marshall, C. A. Mears, P. W. Mote, H. Nakamura, N. Nicholls, J. R. Norris, T. Oki, F. R. Robinson, K. Rosenlof, F. H. Semazzi, D. J. Shea, J. M. Shepherd, T. G. Shepherd, S. C. Sherwood, A. J. Simmons, I. Simmonds, P. C. Siegmund, C. D. Thorncroft, P. D. Thorne, S. M. Uppala, R. S. Vose, B. Wang, S. G. Warren, R. Washington, M. C. Wheeler, B. A. Wielicki, T. Wong.

Review Editors: Brian J. Hoskins, Bubu P. Jallow, Tom R. Karl

Date of Draft: 12 August 2005

Notes: This is the TSU compiled version. References are highlighted where they are not yet published.

Table of Contents

Executive Summary

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 1901–2004 period. The linear trends are 0.062, 0.056 and 0.058 K decade^{-1}, for estimates compiled by CRU/UKMO, NCDC and GISS, respectively, with ±2 standard error ranges of 0.02 K decade^{-1} suggesting that it is very likely that a warming of 0.6 K occurred over the 20th century. However, the trend is not linear, and low pass filtered time series suggest that the warming is 0.75 K by 2004 relative to 1860–1900. Agreement is consistent between the datasets in their separate land and ocean domains, and between sea surface temperature (SST) and nightime marine air temperature (NMAT) for the oceans, over this period. Rates of temperature rise are greater after the mid-1970s (from 1979 to 2004 the linear trend is 0.15–0.18 K decade^{-1}), for a total warming of 0.5 K, and land regions have warmed at a faster rate than the oceans for both hemispheres (0.24 to 0.26 versus 0.14 K decade^{-1}), with the greatest warming in northern winter (DJF) and spring (MAM) in the NH. The warmest year remains 1998, since it was enhanced by the major 1997–1998 El Niño, 2002–2004 are the 2nd, 3rd and 4th warmest years in the series since 1861 and nine of the last 10 years (1995 to 2004) – the exception being 1996 – are among the ten warmest years.

The warming of the climate is consistent with 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. The most marked changes are for cold nights, which have been reduced over the 1951–2003 period for 76% of the land regions studied. Warm nights have increased across 72% of the same regions. The greater increase in nighttime as opposed to daytime temperatures has continued. Diurnal temperature range (DTR) averaged over the 71% of the land surface where data are available decreased by 0.07 K decade^{-1} for 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. The summer (JJA) was the warmest since comparable instrumental records began around 1780 (1.4 K above the previous warmest in 1807) and very likely to have been the warmest since 1500. Recent warming is strongly evident at all latitudes in SSTs over each of the oceans, although thermohaline signatures are apparent in the Atlantic through interhemispheric differences, the Pacific is punctuated by El Niño and Pacific decadal variability that is more symmetric about the equator, while the Indian Ocean exhibits more steady warming, leading to important differences in regional rates of surface ocean warming that affect the atmospheric circulation.

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. 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 alleviation of pollution and other effects on weekends.

Radiosonde measurements of lower-tropospheric temperature show similar warming rates to the surface temperature record over 1958–2004, but the radiosonde record is markedly less spatially complete than the surface and increasing evidence suggests 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 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 0.15 to 0.18 compared to 0.12 to 0.19 K decade^{-1} for MSU estimates of lower tropospheric temperatures. It is likely that there is increased warming with altitude from the surface throughout the troposphere in the tropics, pronounced cooling in the stratosphere, and a higher tropopause.

Stratospheric temperature estimates from adjusted radiosondes, satellites (MSU channel 4) and reanalyses are all in quantitative agreement, recording a cooling of between 0.3 and 0.4 K 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 this cooling, owing to changes in sondes not yet accounted for. Because of stratospheric warming episodes following major volcanic eruptions, the trends are not linear.

Patterns of precipitation change are more spatially- and seasonally-variable than temperature change, but where significant changes do occur they are consistent with measured changes in streamflow. Taken as a whole, average annual precipitation over global land areas has increased over the period 1901–2004 by 11 to 21 mm per century, but there are marked regional differences. Consistent with a warming climate and observed significant increasing amounts of water vapour in the atmosphere, it is deemed likely that there are increases in the numbers of heavy precipitation events (e.g., 95th percentile) from many land regions, even 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.

Despite an overall increase in global land precipitation, droughts have been widespread in various parts of the world since the 1970s, and 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. In the western United States, diminishing snow pack and subsequent reductions in soil moisture appears to also 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, increased temperatures and associated increased potential evapotranspiration that have enhanced evaporation and drying, and decreased precipitation are important factors that have led to increased regions under drought, as measured by the Palmer Drought Severity Index.

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. Similar trends have been detected in the upper troposphere from 1982-2004. Although records are sparse, continental-scale estimates of pan evaporation show decreases, 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 inferred evaporation exhibits an increase in association with enhanced soil wetness from increased precipitation, as the actual evaporation becomes closer to the potential evapotranspiration measured by the pans. Hence there is a trade-off on evaporation between less solar radiation and increased wetness, with the latter generally dominating.

Widespread (but not ubiquitous) decreases in continental DTR since the 1950s coincide with increases in cloud amounts. A hypothesis that changes in cosmic rays affect cloud and thus climate has not stood up to independent data. 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 El Niño Southern Oscillation (ENSO) related in part, 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 (dominated by relatively few major patterns and annular modes) are apparent and confidence is greatest after 1979. Increasing westerlies have occurred as annular modes in both hemispheres have strengthened in most seasons from 1979 to the late 1990s, together 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. Wind and significant wave height analysis support the reanalysis-based evidence for an increase in extratropical storm activity in the northern hemisphere (NH) in recent decades. ENSO is the dominant mode of global-scale variability on interannual time scales although there have been times when it is less apparent, and the 1976–1977 climate shift, related to the phase change in the Pacific Decadal Oscillation toward more El Niños, has affected many areas, including most tropical monsoons. For instance, over North America, ENSO and Pacific-North American 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. On monthly time scales, the southern and northern annular modes (SAM and NAM, respectively) and the North Atlantic Oscillation (NAO) are dominant in the extratropics and the latter two are closely related. The increasing westerlies in the NH change the flow from oceans to continents and are a major part of the wintertime observed changes in storm tracks and related patterns of precipitation and temperature anomalies, especially over Europe, as part of NAO and NAM changes. In the SH, SAM changes are identified with contrasting trends of the strong warming over the Antarctic Peninsula, and cooling over much of continental Antarctica. Multi-decadal variability is also evident in the Atlantic, and appears to be related to the thermohaline circulation.

Variations in tropical cyclones, hurricanes and typhoons are dominated by ENSO and decadal variability, which result in a redistribution of tropical storms and their tracks, so that increases in one basin are often compensated by decreases in other oceans. By far the most active tropical storm year is 1997, when a major El Niño event occurred and surface temperatures were subsequently the highest on record, and this is followed by 1992, a moderate El Niño year. Such years tend to contain low values in the North Atlantic, but much higher values in the Pacific. Linear trends are not significant. Nevertheless some trends are apparent in SSTs and other critical variables that influence tropical thunderstorm and tropical storm development. Moreover, the first recorded tropical cyclone occurred in the South Atlantic in March 2004 off the coast of Brazil. In the North Atlantic, tropical cyclone activity during 1995–2004 was considerably above normal, with eight out of ten seasons (the exception being El Niño years of 1997 and 2002) above normal, although with decreases in the East Pacific. The 21 typhoons in 2004 is second highest on record to 1997 (23) in the western North Pacific. Estimates of potential destructiveness of hurricanes show a substantial upward trend since the 1970s, with a trend toward longer lifetimes and greater storm intensity.

3.1 Introduction

This chapter assesses the observed surface and atmospheric climate to place the new observations and new analyses during the past six years (since the TAR) in the context of the previous instrumental record. In the previous IPCC reports, paleo-observations from proxy data on the distant 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 is enormous 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.

In the TAR temperature trends were examined over 1860–2000 globally and 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 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 not changed and is therefore not repeated in detail here. However, it proves to be 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 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. The reanalyses of the global atmosphere from National Centers for Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) (NCEP/NCAR) and European Centre for Medium Range Weather Forecasts (ECMWF, referred to as ERA-40), for instance, 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; Trenberth et al., 2005a). Therefore the availability of 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 presented some maps of trends. However, climate varies on all space and time scales. The nature of atmospheric waves naturally creates regions of temperature and moisture of opposite-signed anomalies (departures from normal) as moist warm conditions are favoured in poleward flow and cool dry conditions occur in equatorward flow. Although there are 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 the dominant 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 enormous impacts on society and the environment. But extremes 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 directly linked to global mean radiative forcing and are important because these indicate if unusual change is occurring. But the response locally can be complex and perhaps even counter-intuitive (such as cooling in response to global warming that changes planetary waves in the atmosphere). 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 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 approximation is somewhat affected by the observed trend, which increases 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 somewhat representative mid-latitude mid-continental station (Boulder, Colorado; based on 80 years of data) and for monthly mean anomalies 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 four times the standard deviation.

Time and space scale	Range of temperature ºC
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 very unusual, given the null hypothesis. 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); medium likelihood (>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 5 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 19th century. Minor revisions have additionally been made by Brohan et al. (2005), including a comprehensive reassessment of errors. 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 (Section 3.8), as well as potential urban influences on both large-scale temperature averages and microclimate. A new gridded dataset of 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 (Diaz et al., 2002; Rayner et al., 2005). Improvements have also been made in the bias reduction of satellite-based infrared (Reynolds et al., 2002) and microwave (Wentz et al., 2000; Reynolds et al., 2004) retrievals of SST for the 1980s onwards. 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). These data represent skin temperature, not air temperature, and so must be adjusted to match the latter. 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 length, 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., 2005), both locally and on a global basis (see Appendix 3.A.1.1). 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.1.2). Estimates of surface temperature from ERA-40 reanalyses have been shown to be of climate quality from 1979 (Simmons et al., 2004). Improvements in the ERA-40 over NRA arose both from improved data sources and better assimilation techniques, although problems remain. 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.A.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 (CRUTEM2v) of Jones and Moberg (2003) and Brohan et al. (2005). Warming since 1979 has been 0.26 K decade^{-1} for the globe, but 0.35 and 0.16 K 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) analyses (Figure 3.2.1). The NCDC analysis (which begins in 1880) is higher than the Jones and Moberg (2003) analysis by nearly 0.2 K 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 and NCDC are in excellent agreement if the infilling aspects for NCDC are ignored (see Vose et al., 2005b, which shows almost identical time series, when the Jones and Moberg (2003) data are gridded using the NCDC technique and *vice versa*). The number of station series used by CRUTEM2v, NCDC and GISS is 4167, 5985 and 6308 respectively, and much of the basic station data are in common. Differences in station numbers relate principally to CRUTEM2v requiring series to have a sufficient number of years in the 1961–1990 base period in order to calculate anomalies (see discussion in Jones and Moberg, 2003). Differences also relate to differing homogeneity adjustments that have been applied by the three groups to the individual station series. The GISS network is mostly the same as GHCN, but uses in addition a population-based method to estimate the effects of urbanization (see also Section 3.2.2.2). For the United States, GHCN and GISS use the 1200 stations from the U.S. Historic Climate Network (HCN). 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 of the large equal-area gridding method, but long-tern trends are similar in the NH. The major difference between CRUTEM2v and GHCN for the global average relates to whether it is calculated as one domain or as the average of the two hemispheres (Vose et al., 2005b and earlier discussion in Wigley et al., 1997). Here we use the latter method, which gives slightly more weight to the less well covered SH.

Trends and low-frequency variability of surface air temperature from the ERA-40 reanalysis and from the monthly climate station data analysed by Jones and Moberg (2003) are in generally excellent agreement (correlations >0.96 between hemispheric- and continental-scale averages) from the late 1970s onwards (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.A.2.

From 1950–2004 the annual trends in minimum and maximum land surface air temperature averaged over regions with data were respectively 0.20 K decade^{-1} and 0.14 K decade^{-1}, with a trend in diurnal temperature range (DTR) of –0.07 K decade^{-1} (Vose et al., 2005a) (Figure 3.2.2). This is consistent with the TAR; spatial coverage is now 71% instead of 54% of the terrestrial surface, although tropical areas are still somewhat under-represented. Prior to 1950, insufficient data are available to develop global-scale maps of maximum and minimum temperature trends. Coverage is generally inadequate over much of Latin America, Africa and much of southern Asia (Vose et al., 2005a). A map of the trend of change of annual DTR over the 1979–2004 period is discussed later (see Figure 3.2.11). For 1979–2004, the corresponding linear trends for the land areas where data are available are 0.29 K decade^{-1} for both maximum and minimum temperature with no trend for DTR.

[INSERT FIGURE 3.2.1 HERE]

Table 3.2. Temperature trends (K decade^{-1}) in hemispheric and global land surface air temperatures, SST and night marine air temperatures. Trends with ±2 standard error ranges and significances (**bold: <1%**; *italic, 1%–5%*) were estimated by Restricted Maximum Likelihood (Diggle et al., 1999) see Appendix 3.A.1.2, 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	1861–2004	1901–2004	1910–1945	1946–1978	1979–2004
NH Land, CRU (Jones and Moberg, 2003)	**0.066 ± 0.017**	**0.090 ± 0.028**	**0.135 ± 0.076**	–0.022 ± 0.070	**0.351 ± 0.122**
NH Land, GHCN		**0.073 ± 0.028**	**0.118 ± 0.066**	–0.053 ± 0.061	**0.320 ± 0.128**
NH Land,GISS		**0.079 ± 0.030**	**0.166 ± 0.061**	–0.053 ± 0.062	**0.280 ± 0.095**
SH Land, CRU (Jones and Moberg, 2003)	**0.049 ± 0.016**	**0.073 ± 0.018**	**0.110 ±0.080**	0.032 ±0.081	**0.163 ± 0.092**
SH Land, GHCN		**0.048 ± 0.015**	*0.064 ± 0.056*	0.019 ± 0.055	**0.145 ± 0.074**
SH Land,GISS		**0.053 ± 0.015**	0.030 ± 0.042	0.057 ± 0.049	0.066 ± 0.076
Globe Land, CRU (Jones and Moberg, 2003)	**0.058 ± 0.016**	**0.081 ± 0.021**	**0.122 ±0.066**	0.006 ±0.061	**0.257 ± 0.089**
Globe Land, GHCN		**0.066 ± 0.025**	**0.105 ±0.053**	–0.033 ±0.058	**0.274 ± 0.105**
Globe Land,GISS		**0.065 ± 0.021**	**0.100 ± 0.041**	0.002 ± 0.045	**0.173 ± 0.089**
NH SST, UKMO (Jones et al., 2001)	**0.037 ± 0.013**	**0.060 ± 0.021**	**0.155 ±0.044**	–0.038 ±0.077	**0.183 ± 0.062**
SH SST, UKMO (Jones et al., 2001)	**0.048 ± 0.010**	**0.068 ± 0.012**	**0.140 ±0.043**	*0.071 ±0.054*	**0.087 ± 0.047**
Globe SST, UKMO (Jones et al., 2001)	**0.045 ± 0.010**	**0.064 ± 0.014**	**0.153 ±0.043**	0.013 ±0.047	**0.136 ± 0.041**
NH NMAT, UKMO (Rayner et al., 2003)	**0.037 ± 0.014**	**0.063 ± 0.025**	**0.176 ±0.045**	–0.047 ±0.073	**0.183 ± 0.077**
SH NMAT, UKMO (Rayner et al., 2003)	**0.039 ± 0.015**	**0.069 ± 0.015**	**0.101 ±0.046**	0.076 ±0.055	*0.091 ± 0.067*
Globe NMAT, UKMO (Rayner et al., 2003)	**0.038 ± 0.013**	**0.066 ± 0.017**	**0.138 ±0.041**	0.011 ±0.053	**0.140 ± 0.057**

[INSERT FIGURE 3.2.2 HERE]

3.2.2.2 Urban temperatures and the urban heat island

The micro-climates in cities are clearly different than in neighbouring rural areas. The relative warmth of a city compared with surrounding rural areas is known as the Urban Heat Island (UHI), associated with urban-related effects on heat retention, changes in runoff, changes in albedo, changes in pollution and aerosols, and so on. Such changes are real but very localized. Section 3.3.2.4 discusses impacts of urbanization on precipitation. This section considers the UHI and the broader issues of the impacts of land-use change, both from a temperature perspective.

Many local studies have demonstrated that the microclimate within cities is, on average, warmer 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 temporal trends. The few studies that have looked at hemispheric and global scales conclude that any urban-related effect 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 omitting a small number of sites (<1%) with clear urban-related warming trends. In a worldwide set of about 260 stations, Parker (2004) 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 by increasing urbanization (Parker, 2004). Over the conterminous USA, rural station trends are almost indistinguishable from series including urban sites (Figure 3.2.3 from Peterson and Owen, 2005), and the same is true of China from 1951–2001 (Li et al., 2004).

Comparing surface temperature estimates from the NCEP/NCAR reanalysis with unadjusted (i.e., as measured) 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. However, there was no specific analysis of urban or rural effects and the conclusion was based on the fact that the reanalysis did not include these factors. Also the reanalysis did not include many other natural and anthropogenic effects, such as increasing greenhouse gases and observed changes in clouds or soil moisture (Trenberth, 2004), and Vose et al. (2004) show that the adjusted station data for the region (for various issues affecting homogeneity, see Appendix 3.A.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 can only be used for estimating longer timescale trends reliably since 1979 (Simmons et al., 2004) and their unreliability for estimating trends incorporating earlier periods is discussed in Appendix 3.A.5.

On the same regional scale, urban stations in the conterminous USA had slightly larger DTR declines than rural stations after 1950, but the difference was not statistically significant (Gallo et al., 1999). Radiosonde-based temperature measurements in the central USA had a decrease in DTR only below 850 hPa (Balling and Cerveny, 2003). The decrease at 850hPa was larger than that at the surface, suggesting that trends in the low troposphere were likely influenced by near-surface processes such as variations in cloud cover or precipitation but were not dominated by urban and land-use change effects. Regional changes in land use can still 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 USA (Balling et al., 1998) and croplands in the mid-western United States had lower maximum temperatures and a smaller DTR in comparison with forested areas in the northeastern part of the country (Bonan, 2001). Desiccation of the Aral Sea since 1960 raised DTR locally around this region (Small et al., 2001). By processing surface temperature maximum and minimum 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 and strongly suggests an anthropogenic effect on climate, most likely through changes in pollution and aerosols.

[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. However, 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 and Appendix 3.A.3). So far, 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). But Figures 3.2.4b, 3.2.9 and 3.2.10 (see later) do make use of adjusted satellite SST estimates after November 1981 to provide nearer-to-global coverage for the 1979–2004 period. Even satellite data are unable to fill in estimates of a surface temperature measured over or near to sea-ice areas. Post 1942 bulk SSTs estimated using ship and buoy data also have time-varying biases (e.g., Christy et al., 2001; Kent and Kaplan, 2005), larger than originally estimated by Folland et al. (1993), but not large enough to prejudice conclusions about recent warming (Appendix 3.A.3.).

A combined physical-empirical method (Folland and Parker, 1995) is mainly used, as 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 the COADS SST dataset (updated from Slutz et al., 1985) to agree with the NMAT (Bottomley et al., 1990) COADS data before 1942 (see also Appendix 3.A.3) and derive rather similar adjustments to Folland and Parker (1995), geographically as well as through time, 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. 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) extract the leading global covariance pattern, which represents long-term changes, before interpolating using reduced-space optimal interpolation (see Appendix 3.A.1); and Smith and Reynolds remove a smoothed, moving 15-year average field before interpolating by a related technique. Rayner et al. (2005), in a new analysis of the ICOADS data with no interpolation, find that they need to adopt 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.

Figure 3.2.4a shows annual and decadally smoothed values of global SST from this new UK Met Office (UKMO) analysis (Rayner et al., 2005) that does not fill regions of missing data. NMAT, used to avoid daytime heating of ship decks (Bottomley et al., 1990; Folland and Parker, 1995), is also shown. These generally agree well, especially after the 1880s. The SST analysis from the TAR is included. 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). Although NMAT 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). The NMAT analysis incorporates optimally interpolated data using independent spatial patterns (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. NMAT 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 less sampling is needed for SST than for NMAT to remove synoptic weather noise. However, the relative changes of SST compared to NMAT since 1991 in the tropical Pacific (Christy et al., 2001) may be partly real. 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 UKMO 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) series is also *in situ* except for the specification of sea-ice. The warmest year globally in each SST record was 1998 (UKMO 0.47 K, NCDC 0.38 K, Japan 0.37 K 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 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 manifested. In the Atlantic, observations reveal the AMO (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), and it 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. Accordingly, we present in Figure 3.2.5 latitude-time sections from 1900 for SSTs 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 is focussed on the 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 is 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 not only ceased but has been replaced by strong warming (see also Figures 3.2.9 and 3.2.10). 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 how 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 (HadCRUT2v, Parker et al., 2004, Brohan et al., 2005), NCDC (Smith and Reynolds, 2005) and GISS (Hansen et al., 2001) in the United States. Although the component datasets differ slightly (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 and recent warming has been a little less in the NCDC data than in the CRU/UKMO data. All series indicate that the warmest 5 years have occurred since 1998, although there is slight disagreement about the ordering of the years with the exception of the warmest (1998).

Hemispheric and global series based on Parker et al. (2004) 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 65ºN) average annual temperature has increased since the 1960s and is now warmer (at the decade timescale) than conditions experienced during the 1920–1945 period. However, this warming is not yet quite as long as that in the early-to-mid 20th century. Patterns of Arctic warmth in the two periods (1920–1945 and since 1990) are also quite different, as they are in almost all regions of the world with adequate data for the earlier period. The pattern of recent warmth is also more consistent with changes in the NAM (see Section 3.6.4) than those earlier in the 20th century (Polyakov et al., 2003). Temperatures over Antarctica (south of 65°S) have increased, but are only weakly statistically significant over the whole period from 1958, and not since 1979. Strong warming has likely occurred 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 was 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 partly related to atmospheric circulation changes in the winter half year due to the North Atlantic Oscillation (NAO)/Northern Annular Mode (NAM) (see discussion in Section 3.6.4). In contrast, some locations, mainly oceanic and in the SH, have cooled since 1979 (Figure 3.2.9), again possibly through changes in atmospheric circulation related to the Pacific Decadal Oscillation (PDO) and Southern Annular Mode (SAM) (see discussion in Section 3.6.5).

Table 3.3. Temperature trends (K decade^{-1}) in hemispheric and global combined land surface air temperatures and SST. Trends, ±2 standard error ranges and significances (**bold: <1%**; *italic, 1%-5%*) were estimated by Restricted Maximum Likelihood (Diggle et al., 1999), which allows for serial correlation in the residuals of the data about the linear trend. Trends are based on annual averages with estimates of intrinsic uncertainties.

Series	1861–2004	1901–2004	1910–1945	1946–1978	1979–2004
NH (CRU/UKMO)	**0.053 ± 0.019**	**0.072 ± 0.026**	**0.165 ±0 .041**	–0.036 ± 0.048.	**0.257 ± 0.074**
SH (CRU/UKMO)	**0.044 ± 0.011**	**0.061 ± 0.014**	**0.116 ± 0.049**	0.034 ± 0.051	**0.104 ± 0.042**
Globe (CRU/UKMO)	**0.047 ± 0.014**	**0.062 ± 0.019**	**0.144 ± 0.037**	0.002 ± 0.039	**0.176 ± 0.042**
NH (NCDC)		**0.058 ± 0.024**	**0.150 ± 0.043**	–0.025 ± 0.039	**0.232 ± 0.079**
SH (NCDC)		**0.057 ± 0.012**	**0.082 ± 0.043**	**0.068 ± 0.033**	**0.075 ± 0.043**
Globe (NCDC)		**0.057 ± 0.020**	**0.112 ± 0.023**	0.023 ± 0.030	**0.154 ± 0.051**
Globe (GISS)		**0.058 ± 0.015**	**0.117 ± 0.031**	0.004 ± 0.036	**0.159 ± 0.061**

[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 HadCRUT2v annual average temperatures for the globe for 1861–2004 in Figure 3.2.6a is 0.23 K and the greatest difference between two consecutive years in the global average since 1901 is 0.27 K between 1956/1957 and 1963/1964, demonstrating the importance of a 0.6 K warming in a century time-scale context. However, 0.6 K is small compared with interannual variations at one location, and much smaller than day-to day variations (e.g., see 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 Splitting the record into three parts, the average for 1971 to 2005 was warmer than 1936 to 1970 with the latter period warmer than 1901–1935. Indeed, a simple t-statistic shows that the three periods are unlikely to come from that same statistical distribution, a statement that could not be made when comparing 1901 to 1935 with 1866 to 1900, although data for the latter period are sparser than the records since 1901. Splitting the record into decades it is only the 1950s and 1960s decade that could not be said to be warmer than the decade before.

The principal conclusion from this section is that the global average surface temperature linear trend has very likely warmed by just over 0.6 ± 0.2 K over the period from 1901 to 2005 (Table 3.3), a warming that, considering historical precedents (Chapter 6) would be classed as extremely unlikely to have occurred by chance. However, the record can also be characterized as level prior to about 1920, a warming to 1940 or so, levelling out or even slightly decreasing until 1970, and a fairly linear upward trend since then (see Figure QACCS3.1). Using the low-pass filtered data, the overall warming through 2005 is 0.75 K, with 0.5 K increase occurring after the mid-1970s. Clearly the world's surface temperature has continued to warm since the TAR. The last 10 complete years (1995 to 2004) now contain nine of the ten warmest years since comparable records can be developed from 1856. Only 1996 is not in this list – replaced by 1990 instead. 2002–2004 are the 2nd, 3rd and 4th warmest years in the series with 1998 the warmest. The HadCRUT2v surface warming trend over 1979–2004 is 0.18 K decade^{-1}, i.e. a total warming of 0.41 ± 0.03 K (error bars which overlap those of NCDC and GISS).The global trend over the past 30 years has been fairly linear (see Figure 3.2.6a) and over the last 5 years the global average temperature anomaly has been 0.45 K 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 1901–2004 and 1979–2004, and Figure 3.2.10 shows seasonal trends for 1979–2004. 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 two smaller regions over the southeastern United States and a part of Bolivia. 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 is strongest over the continental interiors of Asia and northwestern North America and also over some mid-latitude ocean regions of the SH. In the recent period, some regions have warmed substantially while a few have cooled slightly (Figure 3.2.9). All of the NH shows annual warming (Figure 3.2.9) except for parts of the North Pacific. Warming over the SH is generally smaller in magnitude than for the NH, with cooling dominating over parts of the mid-latitude oceans, particularly over the southeastern Pacific. Warming dominates most of the seasonal maps, but weak cooling has occurred over a few regions, more generally over the mid-latitudes of the SH oceans in 1979–2004. Warming in this period is 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 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., 2005). Error bars are generally larger in the 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. As noted in Figure 3.2.2, for this period DTR trends overall are small and depend a lot on starting and ending years. Trends tend to be largest in the DJF and season (not shown). Daily minimum temperature (see Vose et al., 2005a) increased in most areas except parts of Australia and southern South America, and daily maximum temperature also increased in most regions. However, the changes reported here, are not consistent with some other datasets, where DTR is reported to be decreasing in the United States in association with cloud increases by Dai et al. (2005). These findings are, however, in close agreement with some recent regional-scale analyses for Poland (Wibig and Glowicki, 2002), Alaska (Stafford et al., 2000), and the Arctic in general (Tuomenvirta et al., 2000). The lack of a significant change in DTR over most of Canada since 1950 has also been noted (Bonsal et al., 2001), as has the increase in DTR in South Korea (Jung et al., 2002) and the decrease in maximum temperatures over northern Argentina (Rusticucci and Barrucand, 2004). Over India, DTR has not changed since 1979, and shows mostly non-significant regions of increases and decreases when looked at since 1901 (Kothawale and Rupa Kumar, 2005; Sen Roy and Balling, 2005). Changes in cloud cover and precipitation explained up to 80% of the variance in historic 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 cool half-year in the Arctic (Przybylak, 2000). Recent changes in the Northern Annular Mode (NAM, see Section 3.5) imply that warmer winters and springs were correlated with increases in DTR in the northeastern United States and Canada (Wettstein and Mearns, 2002). Moreover, variations in sea level pressure patterns partially accounted for increases in cold-season DTR in the northwestern United States and decreases in the south-central United States (Durre and Wallace, 2001). Therefore, the interannual and decadal fluctuations in DTR may have been influenced by these atmospheric circulation and cloud changes.

[START OF QUESTION 3.1]

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 has occurred in two phases, from 1920–1940 (0.3ºC) and more strongly from around 1970 (0.55ºC). Temperature estimates for the lower atmosphere (~2-8km), a distinctly different region than the surface, nevertheless show similar warming rates (within the error estimates of the two datasets) since 1958 (from radiosondes) and from 1979 (from satellites).

Expressed as a global average, surface temperatures have increased by about 0.75°C since the late-19th century, see Question 3.1, Figure 1. A series is produced by combining air temperature observations taken at about 2000 stations over the land areas of the world with measurements of sea surface temperature (SST) from the world's oceans. Since 1979, SST measurements can be 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 (0.3ºC) in the global average from about 1920 to about 1940, a slight cooling (0.1ºC) from then to about 1970 followed by a rapid warming (0.55ºC) at least up to the end of 2004 (see Figure QACCS3.1). The warmest year of the series was recorded in 1998 and 9 of the 10 warmest years have occurred in the last ten complete years (1995–2004). Warming, particularly 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.

Spatially, 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 fewer 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-space 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 heat island effects extend to changes in precipitation, cloud and also DTR with the latter detectable as a "weekend effect" owing to alleviation of pollution and other effects on weekends.

Extreme daily temperatures at any location have little effect on the global averages at the monthly, annual and decadal timescale. Local heat waves and cold spells mostly reflect weather events. The two extremes can occur simultaneously, separated by several hundred km due to teleconnection effects resulting from the circulation changes. One region may have more poleward (warm air) flow, while that downstream gets cooler equatorward flow. In only a few regions of the world (parts of North and southern South America, Europe, northern and eastern Asia, southern Africa and Australia) are long, reliable and digital records of daily temperature data available for the entire period since 1901. Although spatial variability is important, 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 digitally available and spatially extensive 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 averages aloft can also be estimated for recent decades from satellite data (since 1979) and from the radiosonde network (since 1958). These estimates are for various layers from the lower troposphere (e.g., 850–300 hPa or ~2–8 km) to the lower stratosphere. The tropospheric measures need not be the same as surface temperatures, but the two would be expected to show similar trends on long timescales. Satellite estimates are made from MSU instruments housed on polar-orbiting satellites, and so provide global coverage at least twice a day. Since 1979, measurements (of radiances which must be converted to temperatures) have been taken from 13 different satellites so that careful cross-calibration of the different sensors must be performed to account for the different instruments onboard (there can be as much as 2°C between the sensors), different equator crossing times (seeing the world at different times of the diurnal cycle), orbital decay and influences of the stratosphere, which has been cooling since 1979, on the raw radiances. After all the necessary adjustments have been made, lower tropospheric trends from the satellite instruments for the 1979–2004 period are similar to those from surface estimates (within the standard error estimates of the two measures), see Question 3.1, Figure 1. Evidence suggests increasing warming with altitude from the surface throughout the troposphere in the tropics, cooling in the stratosphere, and a higher tropopause, consistent with model results. Estimates from radiosondes since 1958 also agree with surface estimates, but the radiosonde network has markedly poorer coverage than the surface network and is not global in extent. At the surface over land, slightly greater warming is associated with greater increases in minimum temperatures that affect only shallow layers.

[INSERT QUESTION 3.1, FIGURE 1 HERE]

[END OF QUESTION 3.1]

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

3.3.1 Background

As the climate changes, temperature changes are one of the more obvious and easily measured changes, 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. 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) and extremes (Trenberth et al., 2003). However, expectations for overall precipitation amounts are not clear as they are complicated by aerosols, which act to short circuit the hydrological cycle even as increased greenhouse gases likely act to increase amounts overall. As aerosol influences tend to be regional, the net effect over land is especially unclear.

This section discusses most aspects of the surface hydrological cycle, except that surface water vapour is included with changes in the atmospheric column 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.A.4). These difficulties occur for *in situ* measurements owing to spatial sampling and wind effects on the gauge catch, especially for snow. For remotely sensed measurements (radar and space-based), the greatest problems are that only measurements of instantaneous rate can be made, and further there are uncertainties in algorithms for converting various radiometric measurements (radar, microwave, infrared) into precipitation rates at the surface. Because of these measurement problems, and also that most historical *in situ* based precipitation measurements are taken on land, leaving the majority of global surface area under sampled, it is most 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, stream flow, and soil moisture.

3.3.2 Changes in Large-scale Precipitation

3.3.2.1 Global land areas

Trends in global annual precipitation were analyzed using data from the Global Historical Climatology Network (GHCN), with anomalies determined with respect to the 1961–1990 base period (Peterson and Vose, 1997; Vose et al., 1992). The observed trend in global annual precipitation over land areas (Figure 3.3.1) indicates a statistically significant increase of approximately 0.98% decade^{-1} (2.17 mm decade^{-1}) over the 104-year period from 1901–2004 (Table 3.4). In 2003, the annual global precipitation anomaly was negative (dry), reaching approximately –0.90% (–10 mm) from the mean. This was a larger departure from normal than in 2002, when the observed anomaly was slightly negative, but very close to the 1961–1990 average. The current period of negative annual global precipitation anomalies began in 2001. Prior to this 3-year period of negative anomalies, global land areas had been dominated by a period of positive anomalies that began in the mid-1990s. Between 1996 and 2000 the global precipitation anomaly had been positive (wet) for 4 of these 5 years.

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 (Global Precipitation Climatology Project: GPCP) (Adler et al., 2003). The GPCP data are a blend of satellite and gauge data, and the PREC/L data include both GHCN and synoptic data from the NOAA/Climate Prediction Center's Climate Anomaly Monitoring System (CAMS). The CRU dataset is from Mitchell and Jones (2005). The Global Precipitation Climatology Centre (GPCC) (updated from Rudolf et al., 1994) provides monthly data from surface gauges on a 2.5° grid. This dataset has been constructed using GPCC sources (including data from CRU, GHCN, an FAO database and many nationally provided datasets). The dataset designated GPCC uses only those stations whose long-term homogeneity can be assured, while GPCCv3 has used more stations to provide more complete spatial coverage. Gridding schemes also vary and include optimal interpolation and grid-box averaging of areally weighted station anomalies.

For 1951–2004 trends range from –5 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 appears to be statistically significant, but the scatter and uncertainties undermine that result. For 1979–2004, GPCP is added and trends range from –16 to +13 mm decade^{-1} but all are well within the error bars, and thus not 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.

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
PREC/L	19482002	x		Yes	Chen et al., 2002
CRU	1901–2002	x		Yes	Mitchell and Jones, 2005
GHCN	1901–2004	x		No	Vose et al., 1992
GPCC	1948–2002	x		Yes	Rudolf et al., 1994
GPCC V. 3	1951–2002	x		Yes	Rudolf et al., 1998
GPCP V. 2	1979–2002		x	Yes	Adler et al., 2003

b. Global land precipitation trends (mm decade^{-1}). Trends, ±2 standard error ranges and significances (**bold: <1%**; *italic, 1%–5%*) were estimated by Restricted Maximum Likelihood (Diggle et al., 1999, see Appendix 3.A.1.2) 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	1901–2004	1951–2004	1979–2004
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	*2.17 ± 1.83*	–2.56 ± 4.91	3.94 ± 14.85
GPCC		1.82 ± 6.47 [b]	12.82 ± 26.08 [b]
GPCCv3		–3.89 ± 6.27 [b]	–6.08 ± 17.85 [b]
GPCP			–15.6 ± 24.12 [a]

Notes:
(a) Series ends at 2002
(b) Series ends at 2000.
The lag-1 autocorrelation of the residuals from the fitted trend (i.e., the detrended persistence) is in the range 0.3 to 0.4 for the PREC/L, CRU and GHCN series but 0.5 to 0.7 for the two GPCC series and the GPCP series. This suggests that the limited sampling by *in situ* gauge data adds noise, or that systematic biases lasting a few years (the lifetime of a satellite) are afflicting the GPCC data, or both.

[INSERT FIGURE 3.3.1 HERE]

3.3.2.2 Spatial patterns of precipitation trends

Spatially the trend in annual precipitation (% per century/decade) during the periods 1901–2004 and 1979–2004, respectively, is 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 104 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 southern 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 trend in annual precipitation was observed over western Africa and the Sahel, where the rainfall decrease over the 1901–2004 period was 9% per century. 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 3.7.5 and also Section 3.7.1.4 and 3.8.3.3). Dai et al. (2004a) examined the effect of changing rain-gauge networks on Sahel rainfall time series and concluded the effect is small. A drying trend is also evident over southern Africa. Over much of eastern India the 1901–2004 period shows increases of more than 20%, but the same area shows a strong decrease in annual precipitation for the 1979–2004 period. Similarly most of Australia shows moderate to strong increases in annual precipitation for the 1901–2004 period, but the 1979–2004 period shows little change, and over most of Eurasia, increases in precipitation outnumber decreases for both periods.

[INSERT FIGURE 3.3.2 HERE]

[INSERT FIGURE 3.3.3 HERE]

The seasonal trends in precipitation over the period 1979–2004 are shown in Figure 3.3.3. In general, high-latitude regions show increasing trends in seasonal precipitation. Similar to the trend in annual anomalies

(Figure 3.3.2b), western and equatorial Africa show a small increase in seasonal precipitation reversing the decreasing trends observed in all four seasons over 1901–2004 (not shown). This suggests a major change in large-scale circulation patterns during the 20th century. A decreasing trend in austral summer (DJF) affected southern Africa during 1979–2004, and over the Indian subcontinent and Southeast Asia, the trend was to a slight drying particularly during the southwest monsoon (JJA). Argentina saw increases in precipitation in most seasons, but central Chile experienced a drying, most evident in the austral summer (DJF) and autumn (MAM). Australia, especially the eastern and southwest parts, show a trend to more precipitation during the austral spring and summer, but increased drying during austral fall. Much like the annual map, most of Europe shows little change during this period.

3.3.2.3 Changes in snowfall

Winter precipitation has increased in high latitudes (Figure 3.3.2). 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, 2005a). Statistically significant increases were documented over Fennoscandia, coastal regions of northern North America (Groisman et al., 2005a), most of Canada (particularly, northern regions of the country) (Stone et al., 2000), the permafrost-free zone of Russia (Groisman and Rankova, 2001), and the entire Great Russian Plain (Groisman et al., 2005a,b). However, there were no discernable changes in summer and annual totals of precipitation over northern Eurasia, east of the Ural Mountains (Gruza et al., 1999; Sun and Groisman, 2000; Groisman et al., 2005a,b). The rainfall (liquid precipitation) has increased during the past 50 years over western portions of North America and Eurasia north of 50ºN on a circumpolar basis by ~6%. Rising temperatures have generally resulted in rain rather than snow in locations and seasons where (1961-1990) climatological-average 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., 2005a,b), 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, 2005a,b). In other regions, in particular north of 55°N, the fraction of annual precipitation falling in winter has changed little.

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 found for the period 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 snowfall to snow water equivalent ratio.

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). At Barrow, Alaska, the last day of snowmelt occurs earlier now by 8 days since the 1960s (Stone et al., 2002). At Cracow and Zakopane, Poland, there is no significant trend in snow variables over the period 1895–1999 (Falarz, 2002). Laternser and Schneebeli (2003) found a gradual increase in snow depth, snowfall days, and snow cover for the period 1931–1980 in the lower elevations of the Swiss Alps, then a significant decrease from 1981–1999. Examining the snowfall season length, Ye and Ellison (2003) found that the period of continuous snow cover in some areas of northern Eurasia has shown small increases, however in southern Siberia decreases in continuous snow cover are found. Further, much of Siberia showed large increases in the transitional snowfall season, as defined by the difference between the first (last) snowfall date and the first (last) date of continuous ground snow cover (Ye and Ellison, 2003). Similar types of information on snow duration and amount are generally lacking in other mountainous and polar areas. However, the East Antarctic ice-sheet interior north of 81.6°S appears to have increased in mass by 45 ± 7 billion metric tons according to satellite radar altimetry measurements (Davis et al., 2005) over the period 1992–2003, mainly due to increased snowfall (precipitation). It is suggested that this gain would be enough to slow sea-level rise by 0.12 ± 0.02 mm per year.

3.3.2.4 Urban areas

As noted in Section 3.2.2.2, the micro-climates in cities are clearly different than in neighbouring rural areas and 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; Shepherd, 2005) have continued to link urban-induced dynamic processes to precipitation anomalies. Nor is it confined to urban areas. 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 being modified in some regions by urban environments. Dixon and Mote (2003) found that a growing urban heat island effect in Atlanta, Georgia (USA) enhanced and possibly initiated thunderstorms, especially in July (summer) just after midnight. Low-level moisture was found to be a key factor. Events tended to occur under atmospheric conditions that were more unstable than those on rain-free days but not unstable enough to produce widespread convection.

3.3.2.5 Ocean precipitation

Precipitation measurement over the ocean is based on several different sensors in the microwave and infrared that are combined in different ways. Many experimental products exist, although operational products based on multiple methods that merge them in an optimal way 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, where continuity and the changing mixes of products are issues. 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 Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) (Xie and Arkin, 1997), and comparisons of these and others (Adler et al., 2001; Yin et al., 2004) reveal large discrepancies over the ocean. 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 deemed suitable for climate monitoring.

Robertson et al. (2001b) examined monthly anomalies from several satellite-derived precipitation datasets (using different algorithms) over the tropical oceans. When averaged over the domain of the tropical oceans (30°N/S), two of the algorithms with different methods of detecting precipitation, the Wentz and Spencer (1998) Special Sensor Microwave Imager (SSM/I) algorithm and a deep convective index (DCI) from Microwave Sounding Unit channel 2 (MSU2), showed strong temporal correlations between each other (0.86) and with SST (0.72 for each). However, monthly anomalies from the GPCP and CMAP datasets had markedly lower correlations with the MSU ice scattering signatures (0.47 and 0.06, respectively). These results highlight the uncertainty of integrated, interannual signals over the tropical oceans, but the expectation in the TAR was that measurements from TRMM radar (PR) and passive microwave imager (TMI) would clarify the reasons for these discrepancies. Unfortunately, this has not yet been the case. Robertson et al. (2003) have 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 these 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 that are related to ENSO events (Curtis and Adler, 2003), with the area-averaged land and ocean signals negatively correlated (Adler et al., 2003), therefore tending to cancel each other. Precipitation as a whole increases over the ocean with El Niño but decreases over land, although this is not the case regionally.

Although the trend over 25 years in global total precipitation in the GPCP dataset (Adler et al., 2003) is very small and not significant (see Table 3.4), some regional 25-year trends are noted in the dataset. There is a small increase (about 4% over the 25 years) over the latitude range 25ºS–25ºN (over 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 geographical pattern of changes from the GPCP dataset reveals general increases in the central Pacific Ocean and over and surrounding the maritime continent and in the Indian Ocean with a small area of negative change values along the Equator in the western Pacific Ocean. This pattern of changes in the tropics is similar to a combination of El Niño and La Niña anomaly patterns. This similarity may indicate that the greater frequency of ENSO events in the latter part of the 25-year period may be related to this pattern of change. The inhomogeneous nature of the dataset and the variability of the parameter require that very careful uncertainty analyses be carried out to determine the validity of these estimated changes, both globally and regionally.

3.3.3 Evaporation

There are very limited direct measurements of actual evaporation over global land areas. 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), India (Chattopadhyay and Hulme, 1997), Australia (Roderick and Farquhar, 2004), and China (Liu et al., 2004a). 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 the 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) and increases in cloud cover (Dai et al., 1999). Whether actual evaporation decreases or not depends on how surface wetness changes, see Box 3.1. 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). As a result, changes in evaporation are often calculated based on some empirical models, such as a function of precipitation and surface net radiation (Milly and Dunne, 2001), or using land surface models (e.g., van den Dool et al., 2003; Qian et al., 2005).

The TAR reported that actual evaporation 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. Using observed precipitation, temperature, cloudiness-based surface solar radiation and a comprehensive land surface model, Qian et al. (2005) found that global land evaporation 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, so that large increases in atmospheric moisture demand (i.e., potential evapotranspiration) induced by the rapid warming since the late 1970s may result in substantial drying over global land areas. Dow and DeWalle (2000) showed significant decreases in watershed evaporation and significant increases in sensible heating of the atmosphere with increased urban/residential development on watersheds in the eastern United States.

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). Regional inconsistencies in trends are common. A rare 45-year record of soil moisture over agricultural areas of the Ukraine shows little change over the last 3 decades (Robock et al., 2005). However, 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. However locally, observed soil moisture trends show mixed results. For instance, Iowa shows a decreasing trend, but nearby data from Illinois shows a consistent increasing trend (Robock et al., 2000).

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 was estimated by an offline simulation of a land surface model (LSM) (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 $1^o \times 1^o$ 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 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; Fan et al., 2003; van den Dool et al., 2003; Berg et al., 2003; Mitchell et al. 2004; Dai et al., 2004b) that have also been used to study changes and variations in surface moisture conditions, but results are model dependent and 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) 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 (e.g., Cole and Cook, 1998; Cook et al., 1999; Fye et al., 2003). This section firstly examines the longer instrumental-based PDSI estimations, to look at trends over the recent and the longest periods possible. Later, some recent extreme PDSI events in different regions of the world are used as a means to place them in a longer-term context (see specific cases in Box 3.5 within the extremes section 3.8). As with LSM-based studies the version of the PDSI used is crucial, and it can partly determine some aspects of the results found. Wells et al. (2004) have proposed an alternative to the original PDSI formulation (see discussion in Alley, 1984) that is self-calibrating and which could allow more consistent results across climate regimes with regard to soil moisture. Similar considerations were also included in the PDSI analyses of Dai et al. (2004b).

Using both the PDSI and Common Land Model (CLM – one of the LSMs available) simulated soil water content Dai et al. (2004b) show 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. Although the long-term (1901–2004) land-based precipitation trend shows a small increase, 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. Based on the PDSI data, Dai et al. (2004b) show that globally very dry areas, defined as land areas with the PDSI less than –3.0, have 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 and subsequent increases primarily due to surface warming. Longer European records (van der Schrier et al., 2005) reveal no trend in areas affected by extreme PDSI values (either thresholds of ±2 or ±4) over the 20th century. However, these 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. In Canada, the summer Palmer Drought Severity Index 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.

Centennial-scale oscillations of wet and dry conditions in eastern China observed in a 528-year record appear to migrate from north to south at about 3° latitude decade^{-1}, with similar behaviour found in the western U.S. (Hu and Feng, 2001). Although there was no significant trend over 1880–1998 in eastern China summer (JJA), precipitation for 1990–1998 was the wettest period 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 3.5.1) 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 3.7.5. 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 in the 20th century, but higher temperatures exacerbated the impacts (see Box 3.5.2). Severe drought, stemming from at least three years of rainfall deficits, continue during 2005 especially in the eastern third of Australia, even as 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 recent severe drought has affected the western United States from 1999 to 2004 (see Box 3.5.3).

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), or 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. Dery 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 3.5.4.

Because large dams and reservoirs were built along many of world's major rivers during the last 100 years and they can dramatically change the seasonal flow rates (e.g., by increasing low flow and reducing peak flows) (Cowell and Stoudt, 2002; Ye et al., 2003; Yang et al., 2004), 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).

The La Plata River basin in southeastern South America exhibits large interannual climate variability identifiable in river discharges. 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; Robertson et al., 2001a; Krepper et al., 2003) indicating that monthly flows during El Niño are larger than those observed during La Niña events. Extreme monthly river discharge events are clearly related to El Niño. Camilloni and Barros (2003) found that two thirds of the major discharge anomalies of the Paraná River occurred during El Niño 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). The Uruguay River also exhibits a relation between extreme discharge events and ENSO phases (Camilloni and Caffera, 2005) with major flow events occurring during El Niño events when large positive rainfall anomalies are observed in the region. 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) and it was also significant since 1970 in the Paraguay and Uruguay Rivers.

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 a drying trend, 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, one study shows that 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 and subsequent increases primarily due to surface warming.

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 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. In the coastal Chile stations, 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), and an example is recent drought in Australia which shows evidence of increased intensity, 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 evaporation as well; see Box 3.5.2. Record high maximum temperatures also accompany the dry conditions in 2005.

[START OF QUESTION 3.2]

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 has a substantial influence. Pronounced long-term trends have been observed in some places, with widespread increases in heavy rains, even in places where total amounts have decreased. There are also increases in some regions in the risks of both droughts and floods, and shifts of some regions to wetter or drier conditions. These changes are associated with increased water vapour in the atmosphere arising from global warming.

Precipitation is the general term for rainfall, snowfall, and other forms of frozen or liquid water falling from clouds. Most of the time precipitation is not formed, and the character of the precipitation when it does 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, or many other weather and cloud systems. Hence changes in any of these aspects will alter precipitation.

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). 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 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 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, and hence increased temperatures result in increased water vapour in the atmosphere, in part from the increased drying at the surface. Over the 20th century, based on changes in sea surface temperatures, it is estimated that water vapour increased by about 5% over the oceans in the atmosphere. 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 precipitation remains constant, and with prospects for even stronger events when the overall precipitation amounts increase. Ironically, 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 (see Question 3.2, Figure 1).

[INSERT QUESTION 3.2, FIGURE 1 HERE]

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 North Atlantic Oscillation (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 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. Overall trends in precipitation are indicated by the Palmer Drought Severity Index (See Question 3.2, Figure 1), and this also factors in crude estmates of changes in evaporation.

As temperatures rise, the likelihood of precipitation falling as rain rather than snow increases, especially in spring and fall 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 reduced snow-pack, 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.

[END OF QUESTION 3.2]

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 particularly 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 approaches to come up with a range of estimates of how to account for 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, although an informed commentary is given here. This reflects almost entirely upon the inadequacies of the historical observing network to observe long-term upper air changes and points to the need for future network design that provides the reference ground-truth. A comprehensive review of this whole issue is given by CCSP (2005).

3.4.1.1 Radiosondes

Since the TAR there has been considerable effort devoted to assessing, comparing and improving the quality of the radiosonde temperature record (see Appendix 3.A.5.1). A particular aim has been to reduce artificial changes arising from instrumental and procedural developments during the five decades of the radiosonde record (Free and Seidel, 2005; Thorne et al., 2005; CCSP, 2005).

A range of approaches yielded disparate results when applied to the identification of spurious jumps in a common set of temperature series 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). 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., 2005), has recently been constructed. HadAT uses a neighbour comparison approach to build spatial as well as temporal consistency throughout the record (see Appendix 3.A.5.1).

Another dataset (LKS, Lanzante et al., 2003a) 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. Its value has been verified using independent satellite temperatures (Lanzante et al., 2003b). The removal of widespread biases, especially at high altitudes, led to enhanced tropospheric warming and reduced stratospheric cooling from 1979-97, although the former change was not statistically significant. Similar warming adjustments were obtained by Angell (2003), who removed tropical stations with trends that were inconsistent compared with neighbours. Nevertheless, global trends from the adjusted LKS data are consistently lower than one satellite record (Lanzante et al., 2003b), which in turn is consistently lower than the other available satellite products (Section 3.4.1.2.2). The LKS dataset has been updated by applying a different bias adjustment technique (Free et al., 2004b) to the data after 1997. Data for the extension are taken from a newly created archive (Radiosonde Atmospheric Temperature Products for Assessing Climate: RATPAC) with improved quality control (Durre et al., 2005; Free et al., 2005). Randel and Wu (2005) have used collocated MSU data to show that instrument-related cooling biases remain in some of the LKS/RATPAC radiosonde data for the tropical stratosphere and upper troposphere.

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. In addition, a linear trend over the long term is a poor approximation of what has occurred (Seidel and Lanzante, 2004; Thorne et al., 2005), 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.

A new development in approaches to improve radiosonde data has been to use 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) has found substantial changes in the diurnal cycle as measured by sondes that are almost certainly a consequence of improved sensors, which have have become much smaller over time, reducing the radiation effects. Hence relative to nighttime values, they find a daytime warming of sonde temperatures prior to 1971 that is likely spurious and then a daytime cooling, especially from 1979 to 1997 during the satellite era, that is also spurious. Thus there is likely a spurious downward trend in sonde temperature records throughout the atmosphere after 1979 of order 0.1 K globally: the assessed spurious cooling is greatest in the tropics of 0.16 K decade^{-1} for the 850 to 300 hPa layer, and least in the NH extratropics of 0.04 K decade^{-1}.

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 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. The main layers referred to here are illustrated in Figure 3.4.1 for the lower troposphere, troposphere (or troposphere-UW), mid-troposphere (which corresponds to MSU channel 2), and lower stratosphere (which corresponds to MSU channel 4). Temperatures estimated from MSU channels 2 and 4 radiances are referred to as T2 and T4, while the lower troposphere based on channel 2 is referred to as $T2_{LT}$.

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 (inter-satellite offsets). (2) Orbital decay and associated long-term changes in the time of day that the measurements are made. When coupled with the diurnal cycle in atmospheric temperature, these changes produce drifts in the estimated temperatures (diurnal drift). (3) Drifts in satellite calibration that are correlated with the temperature of the on-board calibration target (calibration target effect). Independent teams of investigators have used different methods to determine and correct for these and other sources of error. Appendix 3.A.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 satellite measurement of atmospheric temperatures. A number of new data records have been constructed from the MSU measurements, as well as from global reanalyses (Section 3.4.1.3). Further, new insights have come from statistical combinations of the MSU records from different channels that have specified the influence of the stratosphere on the tropospheric records (Fu et al., 2004a,b; Fu and Johanson, 2004, 2005). These new datasets 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). 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 K decade^{-1} and a global warming of the troposphere from T2 of 0.04 to 0.20 K decade^{-1} for the period 1979-2004 for the MSU records. The large spread in T2 trends stem 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; see also Appendix 3.A.5.5)

The RSS results for T2 have about 0.1 K 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 seems preferable as it has many more degrees of freedom, and further, the resulting parameters from the UAH procedure for NOAA-9 (1985–1987) are 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 of the 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), and ways the hot point temperature is corrected for (Grody et al., 2004; Fu and Johanson, 2005). In the tropics, these account for differences in T2 of order 0.1 K decade^{-1} in trend after 1987 and discontinuities are 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) (VG2) shows slightly more warming in the troposphere than the RSS data record (Figure 3.4.3). VG2 create a latitude-dependent analysis that allows for errors that depend on both the calibration target temperature and the atmospheric temperature being measured, although they did not account for temporal variations in calibration target temperatures on individual satellites during overlap periods. 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 and cannot distinguish between land and ocean.

Although the T4 from RSS has about 0.1 K 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 low relative to those in the troposphere. 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 K decade^{1}, leaving a residual influence of less than ±0.01 K 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), although 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 MSU-2 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 can be adversely affected by changes in surface emissivity, as it includes a large signal from the surface. Thus, Swanson (2003) demonstrated that the $T2_{LT}$ retrieval fails to represent the seasonal cycle of temperature in the Antarctic lower troposphere owing to changes in sea-ice coverage, although effects on anomalies will be much smaller. 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 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. In the tropics (Figure 3.4.3), UAH $T2_{LT}$ featured a near zero trend in contrast to the positive trends at the surface and for the troposphere from UAH of about 0.1 K decade^{-1}. Corrections have now been made (version 5.2, J. Christy, personal communication) and are reflected in Figure 3.4.3. 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 0.2 K decade^{-1} larger than UAH.

While comparisons of radiosonde station data with collocated satellite data (Christy and Norris, 2004) suggest that the median trends of radiosonde temperatures in the troposphere are very close to UAH trends and a little less than RSS trends, comparisons of trends at individual radiosonde sites vary and root mean square differences of UAH satellite data with radiosondes are substantial (Hurrell et al., 2000). Moreover, radiosonde data contain diurnal cycle influences (Sherwood et al., 2005) that lead to spurious cooling throughout the atmosphere from 1979 to 1997, and residual spurious downward jumps (Randel and Wu, 2005), so that they are compromised by multiple problems (Section 3.4.1.1 and Appendix 3.A.5.1). In the stratosphere, radiosonde trends are more negative than both MSU retrievals, especially when compared with RSS, and this too is likely due to changes in sondes (Randel and Wu, 2005).

[INSERT FIGURE 3.4.3 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 operational global analyses, and it compensates for some but not all of the effects of changes to the observing system (see Appendix 3.A.5.3). Unlike the earlier NCEP/NCAR reanalysis (NRA) which assimilated satellite retrievals, ERA-40 assimilated bias adjusted radiances and included several other advances (see Appendix 3.A.5.3).

Trends and low frequency variability of 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. 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 MSU data used in ERA-40 are corrected for estimated biases (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. Agreement between ERA-40 and the RSS record is better than with the UAH record (Santer et al., 2004), although 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, 2005). ERA-40 also exhibits a middle-tropospheric cooling over most of the tropics and subtropics since the 1970s, that is almost certainly too strong owing to a warm bias in the analyses for the early satellite years.

Geographical patterns of the linear trend in tropospheric temperature 1979–2004 (e.g., Figure 3.4.4) are qualitatively similar in ERA-40 and the two MSU datasets. All show coherent warming over most of the NH but some show cooling over parts of the tropical Pacific. Tropospheric temperature trends differ substantially only south of 45ºS, where UAH indicate marked cooling, RSS indicate moderate cooling, and ERA-40 indicates no net cooling (even when adjusted for stratospheric influence). These differences are not fully understood, although the treatment of surface emissivity 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. At high southern latitudes (Bromwich and Fogt, 2004) 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.

[INSERT FIGURE 3.4.4 HERE]

3.4.1.4 The Tropopause

The tropopause marks the division between the troposphere and stratosphere and 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 including values at the surface from NOAA (NCDC), NASA (GISS), UKMO/CRU (HadCRUT2v), and the reanalyses from NRA and ERA-40. Other levels presented are 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 two estimates from radiosondes from HadAT 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 so some discrepancies arise. Also, the ERA-40 trends only extend through August 2002. VG2 is available only for the global values. The error bars plotted here are 95% confidence limits associated with sampling a finite record where an allowance has been made for autocorrelation in computing degrees of freedom. However, the error bar does not include spatial sampling uncertainty, which has two effects: incomplete sampling may increase the variance but also increases the noise. The former is reflected in the confidence limits but the latter typically cuts down on 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 K for 95% levels (Mears and Wentz, 2005) (see Appendix 3.A.5) are also not explicitly accounted for here. Structural uncertainties reflect uncertainties between different datasets as the common climate variability is then accounted for.

From Figure 3.4.2 the first dominant impression is that the overall records agree remarkably well. 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. Nevertheless, a focus has been on linear trends, which tend to emphasize these relatively small differences.

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. Similarly, 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 and satellite observations and reanalyses agree that there has been global stratospheric cooling since 1979 (Figures 3.4.2, 3.4.3), although radiosondes still overestimate the cooling owing to residual effects of instrumental changes (Lanzante et al., 2003b; Sherwood et al., 2005; Randel and Wu, 2005) and 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 obvious that the troposphere must be warming at a significantly greater rate than indicated by T2 alone. Given that so much attention has been given to T2 in the past, it is important to note that the troposphere adjusted for the stratospheric contribution to T2 has warmed more than T2 in every case. The differences range from 0.06 K decade^{-1} for ERA-40 (which has a warm-biased stratospheric trend) to 0.09 K decade^{-1} for both radiosonde and NRA datasets. For UAH and RSS the difference is 0.07 K decade^{-1}. Lowest tropospheric trends occur for NRA, which did not allow for changes in greenhouse gas increases over the record (Trenberth, 2004) so that the NRA trends are unreliable (Randel et al., 2000), and upward trends at high surface mountain stations are stronger than NRA free atmosphere temperatures at nearby locations (Pepin and Seidel, 2005). Quite aside from the radiative forcing of the model, this affects the satellite retrievals in the infrared, as carbon dioxide has increased. The records suggest that for this period, the global tropospheric trends are similar to those at the surface, although RSS, and by inference VG2, indicate greater tropospheric than surface warming. 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 in interannual fluctuations (ENSO) in radiosondes, RSS and with models (Santer et al., 2005), and only the radiosonde records are at odds for trends. If the latter were corrected for radiation effects (Sherwood et al., 2005), then they too show increased warming with altitude.

The global mean trends since 1979 disguise many differences regionally. In particular, much larger surface temperature trends are present over northern continents, especially in winter, than at higher levels (CCSP, 2005) (see Figures 3.2.9, 3.2.10 and Question 3.1, Figure 1). These are associated with surface trends manifested more strongly in minimum, typically nighttime, temperatures that often affect only a very shallow layer, and thus this issue is related to the downward trends in DTR (Section 3.2.2.1). Weakening of winter-time temperature inversions and the strong stable surface layers, that have little signature in the main troposphere, play an important role in the differential temperature change with height. Such changes are also related to changes in surface winds and atmospheric circulation (Section 3.6.4).

In summary, for the period since 1958, global and tropical radiosonde-based tropospheric temperature trends are in close accord with temperature trends at the surface. The climate shift of 1976 (Trenberth, 1990) appeared to yield greater tropospheric than surface warming in the tropics (see also Figure 3.4.2) (Seidel and Lanzante, 2004; Thorne et al., 2005); such variations of climate make differences between the surface and tropospheric temperature trends since 1979 unsurprising. After 1979, linear trends of over 0.1 K decade^{-1} are similar at the surface and troposphere, or slightly larger in the troposphere in some estimates, although trends are greater for the surface over land (particularly compared to ocean), where minimum temperatures have risen more than maxima. However, substantial cooling occurred in the lower stratosphere. Clarification of the effects of stratospheric cooling trends on the T2 record (a cooling of about 0.08 K 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 (compared with the variability) relative trends exacerbate the differences between records. It is apparent that inadequacies in the observations and analytical methods 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 water vapour acts as the main source of moisture for precipitation, providing latent heating in the process and dominating the structure of tropospheric diabatic heating (Trenberth and Stepaniak, 2003a,b). It is also the most important greenhouse gas (Kiehl and Trenberth, 1997) and provides the largest positive feedback in model projections of future climate change (Held and Soden, 2000).

Water vapour at the land surface has been measured (indirectly through the use of dewpoint and wet-bulb thermometers) since the late-19th century, but it is only since the 1950s that the network can be considered extensive, principally due to the international exchange of data from this time. However, the surface water vapour variable has altered over time and currently, depending somewhat on timescale, vapour pressure, dewpoint temperature and relative humidity are exchanged. Using physical relationships, it is possible to convert from one to the other, but these are exact only for instantaneous raw 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 (dewpoint temperature is part of the ICOADS database), but few analyses of data 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.

3.4.2.1 Surface water vapour

The TAR reported widespread increases in surface water vapour in the NH. Increasing trends in specific humidity and dew points in the United States were confirmed by Robinson (2000), but different spatial, seasonal and diurnal patterns of change were found, with the largest increases at night rather than in the day, and in winter and spring rather than in summer. Hence although the overall sign of the trends is robust, the details of humidity trends are sensitive to network choice, suggesting local influences, and time period. For 1961–1990, annual mean U.S. dew point temperature trends were 0.22 K decade^{-1} but data after 1990 were not included because of the likelihood of discontinuities caused by the introduction of automated measurement systems. 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. Similarly, Wang and Gaffen (2001) found increases in a number of humidity measures, including specific and relative humidity and dew point temperature in China from 1951 to 1994, particularly at night, as well as increases in heat stress. These increases are consistent with the positive trends in water vapour over China reported in the TAR. Reported trends in water vapour at the surface are, however, still limited almost entirely to regions north of 20ºN. Ishii et al. (2005) report that globally-averaged dew points over the ocean have risen by about 0.25 K between 1950 and 2000.

In an attempt to update near global (60ºS–75ºN) surface humidity trends, Dai (2005) analyzed synoptic data from over 15,000 ships and stations for 1976 to 2005. This analysis, based on individual measurements, analysed specific humidity, temperature and relative humidity. Nighttime relative humidity is greater than daytime by 2 to 15% over most land areas, as temperatures undergo a diurnal cycle, while moisture does not change much. Global trends of –0.1 to –0.2% decade^{-1} are statistically significant in surface relative humidity although small. Trends in specific humidity tend to follow surface temperature trends with an increase of 0.06 g/kg (1976–2004), corresponding to about 4.9%, 4.3% and 5.7% per 1 K warming over the globe, land and ocean respectively. Over the ocean this is close to constant relative humidity and the Clausius Clapeyron relationship.

To summarize, the global, local and regional studies all indicate increases in moisture in the atmosphere at the surface, but highlight differences between regions and between day and night measures and also difficulties in developing long and consistent series.

3.4.2.2 Lower-tropospheric 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 US (Robinson, 2000) and over China (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 and Durr (2004) infer rapid increases in surface water vapour over central Europe from cloud-cleared LW radiative flux measured over the period 1995–2003. Trends in surface observations of water vapour are still almost entirely limited to regions north of 20°N. Ishii et al. (2005) report that globally averaged dew points over the ocean have risen by about 0.25 K between 1950 and 2000.

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. Retrievals of column-integrated water vapour from SSM/I are generally regarded to provide 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 1997/1998 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 Section 8.6.3.1). Linear trends from 1988-2003 based on the SSM/I data over the oceans and time-series of monthly anomalies and linear trends for the global oceans have an upward trend of 0.40 ± 0.09 mm decade^{-1} or about 1.3% decade^{-1}, and this changes to 1.2 ± 0.3% decade^{-1} for 1979–2004 (Figure 3.4.5). The trends are of similar magnitude to the interannual variability relating to ENSO, which may impact the magnitude of the calculated 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 close to the rate at which the precipitable water vapour would increase if relative humidity were held constant. Given observed SST increases, this implies an overall increase in water vapour of order 5% over the 20th century.

An independent check on 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 NCEP/NCAR, NCEP-2 reanalysis and ERA-15/ERA-40 (see Appendix 3.A.5.3) reveal several deficiencies, which limit their utility for climate monitoring. The NVAP product exhibits large discontinuities during 1993 and 2000 when major changes in processing procedures were implemented, resulting in spurious trends (Trenberth et al., 2005a). Zveryaev and Chu (2003) examined changes in precipitable water in the tropics using the NCEP/NCAR reanalysis and found a widespread increase in the late 1970s, but concluded that this was most likely attributable to the introduction of satellite data to the reanalysis at that time, rather than to a real climate shift. However, changes after the 1970s have been shown to agree well with satellite observations over the tropical oceans (Allan et al., 2002) although the amplitude and structures of the variability are unrealistic (Trenberth et al., 2005a). The ERA-40 reanalyses have quite good spatial structures and more realistic low-frequency variability of column integrated water vapour when compared with SMMR and SSMI retrievals (Allan et al., 2004; Trenberth et al., 2005a; Uppala et al., 2005).

To summarize, satellite observations of oceanic low-altitude 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 northern hemisphere land although the calculated trend is likely influenced by large interannual variability in the record.

[INSERT FIGURE 3.4.5 HERE]

3.4.2.3 Upper-tropospheric water vapour

Water vapour in the mid and upper troposphere accounts for a large part of the atmospheric greenhouse effect. In the dry upper troposphere, small changes in the amount of water vapour contribute disproportionately to the greenhouse effect (Held and Soden, 2000). Changes in upper-tropospheric water vapour likely to accompany climate change 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 there is evidence of reduced relative humidity from alternative radiosonde measurements (Sun et al., 2001). Similarly, a combination of Microwave Limb Sounder (MLS) and Halogen Occultation Experiment (HALOE) measurements at 215 hPa suggest a reduction in moisture with increasing temperature (Minschwaner and Dessler, 2004). However, this dependence may stem from errors in the NCEP temperature field used in that study.

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 radiosonde 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 6.3–6.7 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 from many different satellites to ensure uniform calibration: the HIRS data have been most extensively analysed for variability and show linear trends 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). The trends appear to be too small to draw any conclusion concerning climate change and it is not clear whether they are larger than the bias uncertainty due to instrument changes. However, the lack of any significant changes in UTH combined with observed warming of the troposphere (Section 3.4.1) are consistent with an observed moistening of the upper troposphere over the past two decades (Soden et al., 2005).

Clear-sky OLR is also highly sensitive to UTH 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 and constant relative humidity (Wong et al., 2000; Allan and Slingo, 2002).

To summarize, the available data do not indicate a detectable trend in upper tropospheric relative humidity. However, there is widespread evidence for global increases in specific humidity, which is the expected outcome of increasing tropospheric temperatures in the absence of any change in relative humidity.

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 vapor has important radiative and chemical consequences (see also Section 2.3.8). These may include a contribution to the recent observed cooling of the lower stratosphere and/or warming of the surface comparable to those produced by the observed stratospheric ozone or CO_2

changes (Forster and Shine, 1999, 2002; Smith et al., 2001), although the exact magnitude is difficult to quantify due to differences between model parameterizations (Oinas et al., 2001; Forster and Shine, 2002). Recent 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. Shine et al., 2003) noted that because the water vapour observations over the period of consideration are not of a global nature, significant uncertainties remain as to whether radiative effects of a water vapour change are the actual reason for the differences between modelled and observed stratospheric temperature changes.

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). Evans et al. (1998) suggest that roughly 40% of the increase in water vapour in the early 1990s may result from this process. For the altitude levels in the mid-latitude lower stratosphere where balloon observations exist, a maximum of half of the methane in a given parcel should have oxidised (Kley et al., 2000). Methane increases during the 1990s are on the order of 0.01 ppmv/yr (Section 2.3.2). Prior to the 1990s, methane increases were larger, but still a conservative upper limit for the total increase of methane since 1951 is ~0.5 ppmv (Kley et al., 2000.). Assuming the 50% efficiency of oxidation limit, that would increase water a maximum of 0.5 ppmv in 50 years in the NH mid-latitude lower stratosphere. An increase of 1%/yr since 1950, as implied by the observations, leads to an increase of ~2.5 ppmv in water vapour, clearly much larger than that possible from simply increasing the methane entering the stratosphere.

Because of the disconnect between the methane trends and apparent stratospheric water vapour trends, 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 observed mid-latitude lower stratospheric increases harder to explain (Fueglistaler and Haynes, 2005). Recent satellite observations (Read et al., 2004) show water vapour injected above the tropical tropopause by deep convective clouds, bypassing the traditional control point. The sizes and numbers of ice crystals in the tops of these clouds vary, and when sizes are smallest (apparently due in part to biomass burning), water vapour mixing ratios observed entering the stratosphere increase by as much as a factor of two for a given temperature, presumably due to more rapid sublimation (Sherwood, 2002). This effect probably contributed to the upward trend, but to what degree is uncertain. 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. Some calculations indicate that relatively small portions of the tropical tropopause play a dominant role in establishing stratospheric water vapour (Fueglistaler et al., 2004; Bonnazola and Haynes, 2004). There are other studies that suggest some air entering the stratosphere can bypass the coldest part of the tropics (Dessler and Sherwood, 2004). So far it is not clear that a circulation-based mechanism can help 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 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 (see also 3.4.1.4), however 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, its recent disappearance, or what the future holds.

3.4.3 Clouds

Cloud observations have been made from the surface for many years, but observers report the all sky conditions, which therefore include sides as well as bottoms of clouds. In contrast, from satellite only the tops of clouds are seen while observations provide local values. Since the mid-1990s, especially in the U.S. and Canada, human observations at the surface were widely replaced with automated ceilometer measurements, which only measure directly overhead low clouds.

3.4.3.1 Surface cloud observations

Surface observations made at weather stations and onboard ships provide the longest available records of cloud cover changes. 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. As noted in the TAR and extended with more recent studies, these records 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., 2005); 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., 2005).

However, as noted in the TAR, decreasing cloudiness has been reported over China during 1951–1994 (Kaiser, 1998). Moreover, if the analyses are restricted to roughly the last 30 years, 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 3.4.6. During this period, secular trends over land are small. The little variability evident in land cloudiness appears to be correlated with precipitation changes, particularly in the SH (Figure 3.4.6). 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., 2005). However, independent data from military stations suggest an increasing trend (~3% of sky per decade) in U.S. total cloud cover accompanied by a physically consistent decreasing trend (~–0.4 K decade^{-1}) in DTR from 1976 to 2004.

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, though 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). The positive trend in oceanic cloud cover is also evident in both the NH and SH averages.

[INSERT FIGURE 3.4.6]

In Tibet (Yu et al., 2004) monthly mean anomalous cloudiness and surface temperature vary in tandem. Surface warming leads to destabilization and desaturation in the boundary layer, suggesting a positive feedback between the continental stratus clouds and surface temperature through changing lower-tropospheric relative humidity and stratification. The positive feedback mechanism is more robust during periods of surface cooling than during surface warming (Yu et al., 2004); see also Sections 3.2.2.1 and 3.2.2.7.

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. Analyses by Campbell and vonder Haar (2005) suggest that, because the current ISCCP data are not corrected for limb brightening, the ISCCP trend is largely an artefact of changes in satellite view angles (as more geosynchronous satellites have been added to the ISCCP cloud analysis), and discontinuities associated with changes in satellites. In particular, the decrease in cloudiness over the United States suggested by the ISCCP data is in direct contrast to the increase suggested by the U.S. military station records (Section 3.4.3.1) and the accompanying decrease in the DTR (Dai et al., 2005). The ISCCP spurious variability may occur primarily in low-level clouds with the least optical thickness (the ISCCP "cumulus" category) (Norris, 2005a). When this bias is removed, ISCCP's negative cloud cover trend should be smaller, although its 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; Campbell and vonder Haar, 2005; Robinson, 2005), but the effects of known and unknown artefacts on ISCCP cloud and flux data have not yet been quantified. These results suggest that the homogenization methods used by the ISCCP are inadequate for assessment of trends and climate change detection.

ISCCP shows an increase in globally-averaged total cloud cover of ~2% from 1983 to 1987, followed by a decline of ~5% 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, although Deser et al. (2004) suggest there is important structure present, not resolved by broad-scale averages. Analysis of SAGE II data also suggests a decline in cloud frequency above 12 km between 1985 and 1998 (Wang et al., 2002b), which appears to be 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).

Of great concern is the inconsistency in the observed decadal changes in total cloud cover reported from ISCCP and other datasets. For example, analysis of cloud cover changes from the High Resolution Infrared Sounder (HIRS) shows a slight increase in cloud cover between 1985 and 2001 (Wylie et al., 2005). The HIRS measurements indicate very little trend over land but a slight positive trend over ocean, in qualitative agreement with the trends from ship observations. However, spurious changes have also been identified in the HIRS dataset, which may impact its estimates of decadal variability. Another important independent record of cloud cover has recently emerged from the reprocessed AVHRR data under the PATMOS (Pathfinder Atmosphere) project (Jacobowitz et al., 2003). This shows a very strong diurnal cycle in tropical clouds. Hence orbital drift in Equator crossing time (ECT) must be accounted for when interpreting the variations in cloud cover from polar orbiting satellite measurements like AVHRR and HIRS. However, because the ECT drift increases with satellite lifetime, it is difficult to separate this effect from other secular trends in the dataset. After correcting for ECT drift and other small calibration errors in the PATMOS product, Jacobowitz et al. found essentially no trend in cloud cover for the tropics from 1981 to 2000. Also, 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.

The reported increase in surface-observed low-level cloud cover (Norris, 1999), previously described in the TAR, would presumably lead to an increase in reflected SW radiation, which is not consistent with the ISCCP and ERBS radiation flux records at low latitudes (Norris, 2005b). The degree of inconsistency, however, is difficult to ascertain without information on possible changes in low-level cloud albedo. Surface-observed and ISCCP low-level cloud records cannot yet be usefully compared since ISCCP low-level cloud amount also suffers from artefacts. Changes in overlap could be responsible for some of the ISCCP discrepancy with surface-based climatologies. Variability in surface-observed upper-level and low-level cloud cover appears most consistent with ERBS and ISCCP over northern mid-latitude oceans, the Indian Ocean, and the western and central tropical Pacific, and discrepancies are greatest over the tropical Pacific east of 130ºW and over the Atlantic Ocean between 30ºS and 30ºN. Reasons for the origin and geographical distribution of the discrepancies have not yet been identified.

Away from low-latitudes, Wylie and Menzel (1999) found increases in the high altitude cloud fraction from HIRS over northern mid-latitudes over the period 1989–1996. Using AVHRR data and surface observations, Wang and Key (2003) show reduced winter cloud coverage and increased spring and summer cloudiness in the Arctic over the period 1982–1999.

3.4.3.3 Observations related to potential causes of cloud cover changes

Several observational studies have attempted to identify non-climatic mechanisms which might alter cloud cover, including aircraft emissions, soot pollution, and cosmic rays. Of the three, only the aircraft emissions have been convincingly demonstrated to affect cloud cover over large areas.

Condensation trails (“contrails”) from aircraft exhaust may expand to form cirrus clouds. Analysis of surface-based visual cloud observations during the decade 1982–1991 showed increasing cirrus amounts in regions of flight paths (Boucher, 1999). By comparing trends in cirrus to trends in upper tropospheric humidity (UTH), Minnis et al. (2004) were able to assess the contribution of contrails to the cirrus amount. Surface observations for 1971–1995 indicate that cirrus increased over the United States by 1% decade^{-1}, did not change over Europe (contrails apparently compensating for reduced UTH), and decreased over other land areas. For cirrus, the longwave radiative forcing exceeds the shortwave forcing, giving a net warming, and Minnis et al. estimated the global-average radiative forcing of contrails (as of 1992) as extremely small, only 0.006–0.025 W m^{-2}, although Marquart et al. (2003) report that emissions from aircraft and associated generation of contrails contribute to radiative forcing of 0.035 W m^{-2} in 1992 that is expected to increase.

A modelling study (Ackerman et al., 2000) suggests that soot pollution from the Indian subcontinent could increase the absorption of sunlight in low clouds and hasten their evaporation. However, a study of cloud observations from ships in the region showed no decreasing trend in the amount of low clouds from 1952 to 1997, a time of increasing sources of atmospheric soot (Norris, 2001).

Svensmark and Friis-Christensen (1997) and Svensmark (1998) noted a close resemblance during 1984–1990 between the smoothed time series of cosmic ray flux and the smoothed time series of total cloud cover averaged over oceans poleward of 22.5º obtained from the ISCCP C-series (Rossow and Schiffer, 1991). Weaker correlations occurred between cloud cover and cosmic ray flux in the tropics, which they attributed to larger shielding by magnetic field lines at lower latitudes. Subsequently, Marsh and Svensmark (2000a, 2000b) produced a new analysis documenting a close resemblance during 1984–1994 between cosmic ray flux and globally averaged low-level cloud cover obtained from the ISCCP D-series using only IR detection. They now found that the best correlation occurred at low latitudes, a revision of their original hypothesis.

While lacking a specific microphysical mechanism, Svensmark, Friis-Christensen, and Marsh hypothesized that increased ionization of the atmosphere due to increased cosmic ray flux favoured more cloud cover.

The cosmic ray-cloud hypothesis has been criticized for a number of reasons. One is that the combined radiative forcing inferred from changes in individual ISCCP cloud types and cloud optical thickness opposes that posited by Svensmark and Friis-Christensen (Kuang et al., 1998; Kernthaler et al., 1999; Kristjánsson and Kristiansen, 2000). An additional difficulty is that the correlation between the cosmic ray flux and cloud cover was greatly diminished as more years of ISCCP data became available. The cosmic ray time series does not correspond to global total cloud cover after 1991 or to global low-level cloud cover after 1994 (Kristjánsson and Kristiansen, 2000; Sun and Bradley, 2002). Marsh and Svensmark (2003; 2004) asserted the presence of a calibration error in ISCCP had caused divergence of the cloud and cosmic ray time series after 1994.

Additional criticisms of the cosmic ray-cloud hypothesis have questioned whether the ISCCP data actually support a relationship between low-level cloud cover and cosmic ray flux prior to 1994. Kristjánsson et al. (2002) found that the correlation between cosmic ray flux and ISCCP oceanic low-level cloud cover becomes negative when a high-pass filter (less than 1 year) is applied to the data, whereas any cosmic ray modification of cloud condensation nuclei would presumably occur almost immediately. Sun and Bradley (2004) and Kristjánsson et al. (2002) both note that cosmic ray flux exhibits no correlation with the time series of global low-level cloud cover based on the combined visible and infrared detection which is considered to be more reliable than infrared-only detection of low clouds. Multidecadal (1952–1997) time series of cloud cover from ship synoptic reports also do not exhibit any relationship to cosmic ray flux (Kristjánsson and Kristiansen, 2000; Sun and Bradley 2002), although these data might also suffer from observational artefacts (Norris, 1999).

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., 2005). Based upon the revised (Edition 3_Rev1) ERBS record (Figure 3.4.7), 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 present in ISCCP 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, 2002b; Wong et al., 2005).

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; Wong et al., 2005). However, as noted in Section 3.4.3, 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 2 record (Wielicki et al., 2002a; Lee et al., 2004) which does not contain the satellite altitude or SW filter dome corrections, the corrected ERBS Edition 3 data (Wong et al., 2005), 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 W m^{-2} over the 1985 to 1999 period and the spatial and time sampling noise is less than 0.5 W m^{-2} on annual time scales (Wong et al., 2005). 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., 2005). 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., 2005). 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. Consistency of TOA radiative flux changes from the 1980s to 1990s. Values are given as tropical mean (20ºS to 20ºN) for the 1994–1997 period minus the 1985–1989 period. Dashes are shown where no data are available or relevant. From Wong et al. (2005).

Data Source	TOA LW	TOA SW	TOA Net
ERBS Edition 2	3.1	–2.4	–0.7
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 (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 the majority 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. (2005) 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 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, the CERES broadband data indicates a decrease in SW flux by almost 2 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 and energy divergence in the atmosphere (Trenberth and Stepaniak, 2003b) do not reveal trends of significance, but do show substantial interannual variability associated with ENSO. These computations were based on NRA which are not considered reliable for depicting trends. However, during El Niño events, divergence of energy of order 10 W m^{-2} out of the tropics occurs and this all comes from the Pacific where it can exceed 50 W m^{-2} over broad regions for several months (Trenberth et al., 2002b). The latitude time series of Trenberth and Stepaniak (2003b) reveal more convergence of energy into the deep tropics in the 1980s, compared with the 1990s. Conditions in the tropical Pacific were cool following the 1982–1983 El Niño and especially in 1988–1989 when a large La Niña occurred, in contrast to the prolonged weak El Niño conditions from 1990 to 1995 and the major El Niño event 1997–1998, so these conditions clearly play a role in interdecadal variability, and may account for at least some of the changes discussed above.

In summary, although there is increasing 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 3.4.7]

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. In recent years several studies have focused on observational evidence of changing surface radiative heating. Measurements of surface longwave radiation are generally only available since the 1990s. Considering this period over central Europe, Philipona and Dürr (2004) showed that increases in net flux were dominated by the clear-sky longwave radiation component relating to an enhanced water vapour greenhouse effect. Reliable shortwave radiative measurement networks exist since the International Geophysical Year in 1957–1958. A reduction in downward solar radiation of about 1.3% decade^{-1} or about 7 W m^{-2} was observed from 1961 to 1990 at stations worldwide (Liepert, 2002; Gilgen et al., 1998). Note that a global surface temperature increase of 0.5 K over this period could correspond to a global mean net longwave heating increase of about 2 W m^{-2} (e.g., Liepert et al., 2004). Stanhill and Cohen (2001) calculate a stronger reduction of 2.7% decade^{-1}. The reason for the discrepancy might be that only 30 records were used in the latter study compared to about 110 sites worldwide used in the former study. Surface solar radiation declined at 60% of the sites in the former Soviet Union (Abakumova et al., 1996; Russak, 1990), and around the Mediterranean Sea (Omran, 2000 and Aksoy, 1997). Stations in the United States (Liepert, 2002), and in Southern Africa (Power and Mills, 2005) experienced significant declines as well as sites in the Arctic and Antarctic (Stanhill and Cohen, 2001). On the other hand, the changes at sites in Europe were rather mixed from 1960 to 1990. It has been argued that increasing anthropogenic aerosol concentrations might be a dominant cause of these observed drops. 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., 2005). 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.

At 26 out of 32 analyzed sites worldwide the decline in surface solar radiation ("dimming") ended around 1990 and a recovery of about 6 W m^{-2} occurred afterwards (Wild et al., 2004; 2005). Possible causes are again: reduced cloudiness in the 1990s and also increased cloud-free atmospheric transparency due to the reduction of anthropogenic aerosol concentrations and also to the recovery from the effects of the 1991 eruption of Mt. Pinatubo. The increase in surface solar radiation (“brightening”) agrees with observations from ISCCP that show a globally reduced cloud amount of 3–4% from the end of the 1980s on to 2001 (Rossow and Dueñas, 2004), and with estimates of the surface solar radiation based upon ISCCP cloud data (Pinker et al., 2005). Globally reduced cloud cover in the 1990s is consistent with recent reports of top of the atmosphere increasing outgoing LW radiation and decreasing reflected SW radiation over the tropics (Wielicki et al., 2002a). However, 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 (Trepte and Winkler, 2004; Philipona and Durr, 2004) and China (Kaiser and Qian, 2002). Also puzzling is the continued decline in pan-evaporation in Australia from 1970 to 2002. This reduction has been interpreted as an indicator for ongoing “dimming” in the SH (Roderick and Farquhar, 2002; 2004). See Box 3.1 for more discussion and a likely explanation of these aspects.

In general, reduced radiative heating does not necessarily mean lower temperatures since part of the heating is taken away from the latent and sensible heat flux, suppressing evaporation and convection. The troposphere itself can still be warmer and moister because the moisture holding capacity is increased

(Clausius-Clapeyron equation) and water stays longer in the atmosphere (Liepert et al., 2004; Trenberth et al. 2003).

[START OF BOX 3.1]

Box 3.1: 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 till 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; 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).

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 measure 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 the main 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 is an increase in cloud and precipitation, which reduce solar radiation available for evaporation but also increase soil moisture and make the actual evaporation 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; 2005). This reduces solar radiation available for evaporation, 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 evaporation (Milly and Dunne, 2001). A recent reassessment of evaporation data from the United States and former Soviet Union (Golubev et al., 2001) suggests increasing trends in actual evaporation over southern Russia and most of the United States during the last forty years, and estimates of evapotranspiration over the United States by Walter et al. (2004) find increases in evapotranspiration over the past 50 years 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. 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 Section 2.4. Increases in aerosols also reduce direct radiation at the surface under clear skies (e.g., Liepert, 2002).

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. 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 evaporation 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.

[END OF BOX 3.1]

3.5 Changes in Atmospheric Circulation

Changes in the circulation of the atmosphere and ocean are an integral part of climate change. Regional variations in climate can often be related to such changes, referred to as *teleconnections* (see discussion in Section 3.6.1). Changes are often complex and sometimes counter intuitive. For example, a rise in global mean temperatures does not mean warming everywhere. Principally due to teleconnection effects resulting from the circulation changes, a few areas experience cooling. Changes in large-scale atmospheric waves alter the location and strength of the jet streams and associated storm tracks. The latter term is commonly used to refer to the main tracks of extratropical disturbances that occur as sequences of low (cyclonic) and high pressure (anticyclonic) systems carried eastward in the prevailing mid-latitude westerly winds. Temperature gradients and associated preferred spawning regions for extratropical cyclones are also altered, along with their development rate, track, and region of decay, thereby changing precipitation patterns, cloud and humidity. A shift of a storm track polewards, for instance, is apt to make one region wetter at the same time that it dries out another region, with a north-south dipole anomaly structure.

This section assesses atmospheric circulation changes (since the TAR) by examining global-scale datasets of sea level pressure, geopotential heights, jet streams and storm tracks. Many of the results are based on various 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.A.5 for a discussion of the quality of reanalyses from a climate perspective).

3.5.1 Surface or Sea Level Pressure

The sea level pressure (SLP) map depicts an analysis of weather systems, cold and warm fronts, and associated cloud, temperature and precipitation, and sea level pressure is often analyzed first to examine changes in atmospheric circulation. As the total mass of dry air is conserved, changes in sea level pressure necessarily involve mainly changes in distributions or in patterns of variability (Section 3.6).

Hurrell and van Loon (1994) noted circulation 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, Gillett et al. (2003) analyzed linear trends in three sea level pressure datasets for 1948 to 1998 for DJF. They found decreases in sea level pressure over the Arctic, Antarctic and North Pacific, and an increase over the subtropical North Atlantic, southern Europe and North Africa. A weakening trend of the Siberian High was also revealed (Gong et al., 2001). The strength of the mid-latitude westerly circulation appears to have increased in both hemispheres, since at least the late 1970s.

The increases for the NH were significant compared to simulated internal variability. However, changes over the SH are less clear, especially over the oceans prior to satellite observations in the late 1970s. Many spurious trends are evident in both major reanalyses (from NRA and ERA-40; Marshall, 2003; Bromwich and Fogt, 2004; Trenberth and Smith, 2005, see also Appendix 3.A.5). Real trends (validated with long-term station–based data), however, do 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.

In summary, the most notable features of the observed SLP trend from the late 1970s to the late 1990s are decreases in sea level pressure over high latitudes: the Arctic, Siberia, Antarctic and North Pacific, and an increase over the subtropical North Atlantic, southern Europe and North Africa. These changes imply increased gradients and stronger mid-latitude westerlies in both hemispheres, although some seasonal departures from this are evident.

3.5.2 Geopotential Height, Winds and the Jet Stream

Mean changes in geopotential heights tend to be equivalent barotropic (where the amplitude increases with altitude with similar spatial structure). Thus, the main changes in tropospheric geopotential heights resemble their SLP counterparts (Hurrell et al., 2004). NRA for 1960 to 2000 over the NH reveal that winter (DJF) and annual means of geopotential height at 850, 500, and 200 hPa decrease over high latitudes and increase over the mid-latitudes as for surface pressure, 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 been contracting significantly since then (until 2000) at all levels and Angell (2005) finds 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 temperatures. In DJF at 200 hPa, significantly greater heights were found over the Mediterranean Sea, eastern Asia, and near 30°N across the North Pacific and eastern North America in contrast with lower heights over northern North America and the northern North Pacific (Figure 3.5.1). Warming of the troposphere is also suggested at 200 hPa in JJA by significantly higher heights in the North Sea, eastern Mediterranean extending eastward along about 30°N to China, and in the North Pacific (Figure 3.5.1). These changes are not matched by significantly lower values elsewhere in the extratropics.

Fogt and Bromwich (2005) identified geopotential height trends in the SH using ERA-40 reanalyses during the satellite era 1979–2001 (see Figure 3.5.1). In the SH the seasonal cycle of changes is important in the height fields and, at least over Antarctica, the strongest trends are seen in the solstitial seasons, with nearly opposite trends in DJF and JJA. There are substantial and significant height increases in the polar winter upper troposphere over the Antarctic continent in the last 23 years. The changes in the summer reflect the increasing strength of the Southern Annular Mode (SAM) seen in the last 3 decades (see Marshall, 2003), with large height decreases over Antarctica and corresponding and nearly equal 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 ~60ºS 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).

Hemispheric teleconnections are strongly influenced by jet streams, which act like waveguides for quasi-stationary Rossby wavetrains (Ambrizzi et al., 1995; Branstator, 2002). Using NRA from 1979 to 1995, Nakamura et al. (2002) found a weakening of the North Pacific wintertime jet since 1987, which allowed efficient vertical coupling of upper-level eddies with the surface baroclinic zone anchored by the subarctic oceanic frontal zone (Nakamura and Sampe, 2002; Nakamura et al., 2004). A trend from the 1970s to the 1990s towards a deeper polar vortex and Icelandic 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 the associated enhancement of the Atlantic storm track activity (Chang and Fu, 2002; Harnik and Chang, 2003).

In 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 trends are larger with altitude in the troposphere, but qualitatively similar in shape to lower atmospheric levels and go along with the intensification and poleward displacement of corresponding Atlantic and southern polar front jet stream.

[INSERT FIGURE 3.5.1]

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. (2005) found that the North Atlantic storm track has shifted about 143 km northward in winter (JFM) during the past half century. The above findings are corroborated by Zhang et al. (2004a), 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 concomitantly with the sudden weakening of the Siberian High (Nakamura et al., 2002; Chang, 2003). In contrast to linear theories of baroclinic instability, the enhancement of the storm track activity occurred under a weakening of the Pacific jet (Nakamura et al., 2002). 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 SLP variance over the Pacific. All these results suggest that baroclinic wave activity in the NH mid-latitudes has increased during the past 40 years, consistent with an increase in intense cyclone activity during that period.

Observing system changes significantly impact the quality of the reanalysis products (see Appendix 3.A.5.3). 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 increasing trend in eddy variance in the reanalysis data is 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 reanalysis 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). 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 related 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 may be related to reductions in rainfall from the shift in the storm track (e.g., the drying trend observed in southwestern Australia, Karoly, 2003). However, significant differences between ERA-40 and NRA in the SH (fewer cyclones over the Antarctic but more cyclones over the southern oceans, especially in the warm season, in ERA-40) are amplified in the early decades (Wang et al., 2005).

In summary, it is likely that there has been a significant increase in NH winter storm track activity over the second half of the 20th century, but there are still significant uncertainties in the exact magnitude of the increase due to time dependent biases in the reanalysis data. 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 possibly even larger reanalyses uncertainties.

3.5.4 Blocking

Blocking events, involving a breakdown in mid-latitude westerly winds associated with persistent high-latitude ridging 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). The sign of the NAO-blocking relationship changes from west to east. Increased blocking is observed over the Europe (western Atlantic) in association with the positive (negative) phases of NAO in wintertime (Quadrelli et al., 2001; Scherrer et al., 2005). When the NAO is in the negative phase, blocking events over the western Atlantic last more than 11 days on average, significantly longer (95% confidence level) than during the positive phase (8 days) (Barriopedro et al., 2005). In the Pacific sector in spring and winter, Barriopedro et al. (2005) found a significant increase of 53.3% blocked days over its western part from 1948–2002.

In the SH, blocking occurrence is maximised over the southern Pacific, downstream of the subtropical-subpolar split jet region near the Dateline (Renwick and Revell, 1999; Renwick, 2005), with secondary blocking regions over the southern Atlantic and, to a lesser extent, 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.

In summary, the NAO change over recent decades is dynamically consistent with decreases in blocking frequency over the West Atlantic and increases in blocking frequency over the European mainland. 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.

3.5.5 The Stratosphere

The dynamically stable stratospheric circulation is dominated by westerlies in the winter hemisphere and easterlies in the summer hemisphere (Andrews et al., 1987). 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). Much of the forcing for the stratospheric circulation comes from below, through upward propagation of energy by large-scale tropospheric waves that is limited to the winter-spring period, or wave activity changes in the distribution of wave drag within the stratosphere (Shepherd and Shaw, 2004). There is a consequent variability in the winter-spring build-up of extratropical ozone (Fusco and Salby, 1999; Randel et al., 2002), which provides an ozone radiative feedback. The effects on ozone persist through summer and early fall, until the next year's dynamical variability resumes (Fioletov and Shepherd, 2003). In the Arctic, dynamical variability induces strong variability in chemical ozone loss (Rex et al., 2004). The stratospheric aerosol loading from strong volcanic eruptions is another source of year-to-year variability in stratospheric wave drag. It increases the local radiative equilibrium temperature and modifies the meridional gradients, implying changes in zonal wind, thus leading to changes in wave drag.

The breaking of vertically-propagating waves decelerates the stratospheric westerlies, see Box 3.2. The deceleration of the westerlies is particularly large in stratospheric "sudden warmings" when the westerly polar vortex is highly distorted by planetary waves and breaks down with an accompanying warming of the polar stratosphere, which can quickly reverse the latitudinal temperature gradient and turn the westerlies into easterlies (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). In this context, it should be noted that an increasing trend of wave activity during winter had been reported after the late 1970s in both hemispheres (Kuroda and Kodera, 2001).

The analysis of past stratospheric changes is limited by the length of the satellite record, with only sparse information being available prior to 1979. During the middle 1990s the NH exhibited a number of years when the Arctic wintertime vortex was colder and stronger (Kodera and Koide, 1997; Pawson and Naujokat, 1999) and more persistent (Waugh et al., 1999; Zhou et al., 2000). Any dynamically induced component of these changes requires weakened wave drag and Brewer-Dobson circulation. Observations 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. A comparison of ERA-40 and NCEP/NCAR stratospheric height reanalyses indicates a trend towards a strengthening Antarctic vortex since 1980 during summer and autumn (Renwick, 2004 and Section 3.5.2). The SH stratosphere has been perturbed far more dramatically than the NH through the ozone hole. 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). The increased strength of the Antarctic vortex in spring seems to be mainly due to ozone depletion (Ramaswamy et al., 2001; Gillett and Thompson, 2003).

In summary, observations from 1979 to the late 1990s reveal a tendency toward colder and stronger wintertime NH vortex. After 1997, however, the wave forcing seems to increase along with increased occurrences of major sudden warmings in the polar stratosphere. As for the SH, there has been a strengthening Antarctic vortex during summer and autumn 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.

[START OF BOX 3.2]

Box 3.2: 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 organizes this chaotic wave forcing from below to produce long-lived changes to the strength of the polar vortices. These fluctuations in the strength of the stratospheric polar vortices in both hemispheres 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). Although this relationship occurs in the zonal wind, it can be seen more clearly in annular modes, the leading patterns of geopotential variability at levels through the troposphere and stratosphere. Annular modes 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; Lorenz and Hartmann, 2003).

In the stratosphere the NAM (see Section 3.6 for definition) is a measure of the strength of the polar vortex (Thompson and Wallace, 1998). The relationship between NAM anomalies in the stratosphere and troposphere can be seen by examining averages (composites) of the weakest and strongest observed stratospheric anomalies. In Figure 3.5.2 the NAM index at 10 hPa is used to define events during which the stratospheric polar vortex was extremely weak (which correspond to 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. The composites illustrate that simulated NAM anomalies in the lower stratosphere are related to similar NAM fluctuations in the troposphere. The opposite is true for anomalously strong vortex conditions.

Long-lived annular mode anomalies in the lowermost stratosphere appear to lengthen the time scale of the surface NAM. Tropospheric annular mode timescale is longest during winter in the NH, but 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. A consequence of this downward coupling is that 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, 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) implying a more linear (i.e. robust) downward coupling.

The dynamical mechanisms by which the stratosphere influences the troposphere are not well understood, but the fairly 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 NAM (Wittman et al., 2004).

[INSERT FIGURE 3.5.2]

[END OF BOX 3.2]

3.5.6 Winds, Waves and Surface Fluxes

Changes in atmospheric circulation imply associated changes in the ocean winds, wind waves and surface fluxes. Meteorological observations, including surface winds, from Voluntary Observing Ships (VOS) became systematic 150 years ago and are assembled in ICOADS (International Comprehensive Ocean-Atmosphere Data Set) (Worley et al., 2005). Gulev et al. (2005) found significant trends in scalar wind, but these should be considered with caution because VOS wind observations are influenced by time-dependent biases, such as the ratio between the anemometer measurements and Beaufort wind estimates, which changes over time (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; Gulev et al., 2005). With ICOADS winds assimilated into reanalyses, inhomogeneities inherent in the raw VOS wind data will likely dominate. 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. Wind speed directly affects only the wind sea component of SWH, while the swell component is largely influenced by the frequency of storms rather than wind magnitude. Linear trends in the annual SWH from ship data processed by Gulev and Grigorieva (2004) for the period 1900–2002 are 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 show weak 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. 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}). Increases of SWH in the North Atlantic mid-latitudes is confirmed 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]

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 different sources (reanalyses, coupled models, in situ observations) indicate some improvement but there are still global biases of several tens of W m^{-2} (Grist and Josey, 2003). Only indirect estimates of the net from atmosphere-to-ocean heat budgets give reasonable implied ocean heat transports (Trenberth and Caron, 2001). For the North Atlantic Gulev et al. (2005) reported 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, see also Marshall et al. (2001) and Visbeck et al. (2003).

In summary, analysis of observed wind and significant wave height support the reanalysis-based evidence for an increase in storm activity in the extratropical NH in recent decades. 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.

3.5.7 Summary

Large-scale changes in the tropospheric and stratospheric circulation from 1979 to the present are dominated by relatively few major patterns (Section 3.6) and annular modes. 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 corresponding Atlantic and southern polar front jetstreams and enhanced storm tracks. Increasing trends are also visible in extratropical storm activity in the NH in recent decades until the late 1990s.

3.6 Patterns of Circulation Variability

3.6.1 Teleconnections

A number of preferred patterns of variability exist in the global atmospheric circulation, all of which have expressions in surface climate. Box 3.3 discusses the main patterns and indices used here. Regional climates in different locations may vary out of phase, due to the action of such “teleconnections” (see also Section 3.5). The simplest illustration is to take one point correlation maps to illustrate the PNA and NAO, for example (Figure 3.6.1). In the SH, wave structures such as the PNA do not emerge as readily owing to the dominance of the annular mode SAM. Teleconnections are often a consequence of large-scale wave motions, whereby energy is transferred from source regions along preferred paths in the atmosphere. Teleconnections modulate the location and strength of the storm tracks, 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 are central to our understanding of climate variability and change, as they encapsulate much of the seasonal and longer-timescale variability in the troposphere and determine, to a large extent, regional surface climate anomalies. Such anomalies have direct human impacts, often being associated with droughts, floods, heat waves and other changes that can severely disrupt agriculture and fisheries, and can modulate air quality, fire risk, energy demand and supply, and human health.

The analysis of teleconnections has typically evolved using a linear perspective, which assumes a basic spatial pattern that has mirror image positive and negative polarities (see Quadrelli and Wallace, 2004). In contrast, nonlinear interpretations of atmospheric variability have recently found applications within the climate framework (e.g., Palmer, 1999; Corti et al., 1999; Cassou and Terray, 2001; Monahan et al., 2001). In that case, preferred climate anomalies are identified as recurrent states of a specific amplitude and sign. Conceptually, the preferred states, as naturally occurring expressions of the dynamics, are likely to be relatively robust to modest changes in climate. Indeed, climate change may result through changes from one quasi-stationary state to another, or as a preference for one sign of a pattern, thereby changing the probability density function of circulation indicators (Palmer, 1999).

The precise nature and shape of the preferred patterns vary to some extent according to the statistical methodology and to sampling variability (Jones et al., 2003). Examples of spatial patterns are given in Figures 3.6.1 to 3.6.3, 3.6.5 and 3.6.7. Although teleconnections are best defined over a grid, simple indices based on a few key station locations remain attractive. The advantage is that such series can often be carried back in time long before complete gridded fields are available; the disadvantage is increased noise from the reduced spatial sampling (see Figure 3.6.6 for example). 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.

A large variety of 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 in various analyses 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 there 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). Researchers must be mindful of such multi-decadal variability, when looking at relationships over the last 30–40 years, since changes of influence are real and not just due to poorer data quality in earlier decades.

[INSERT FIGURE 3.6.1]

[START OF BOX 3.3]

Box 3.3: Defining the 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 values near the centres-of-action. A classic example is the Southern Oscillation (SO), that describes how mean sea-level pressure (MSLP) tends to vary out of phase on opposite sides of the tropical Pacific: pressures over the eastern Pacific tend to be anomalously high when pressures over northern Australia and Indonesia are anomalously low, and vice versa. Accordingly, a simple SO Index (SOI) is the difference between Tahiti (eastern Pacific) and Darwin (western Pacific) MSLP anomalies, appropriately normalized. The SOI was defined using Tahiti and Darwin MSLP because those sites have long high-quality instrumental records and are the two stations with strongest negative correlations (teleconnectivity). However, monthly MSLP anomalies are often influenced by local fluctuations, so seasonal or smoothed values are most appropriate to define the SO (Trenberth, 1984). 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. Such fields are available for the most part only from the mid-20th century, compromising the length of record, especially for studies of long-term climate trends. Use of different sets of stations to define indices, or use of gridded rather than station data, does not usually materially affect the definition of the teleconnection. For each index, alternate definitions are highly correlated in time.

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).** 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 is less easily updatable in real time.
- **Northern Annular Mode (NAM).** 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).
- **Southern Annular Mode (SAM).** The difference in average MSLP between 45ºS and 65ºS (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 20ºS (Thompson and Wallace, 2000). Formerly known as the Antarctic Oscillation (AAO) or High Latitude Mode (HLM).
- **Pacific-North American pattern (PNA).** The mean of normalised 500 hPa height anomalies at 20ºN, 160ºW and 55ºN, 115ºW minus those at 45ºN, 165ºW and 30ºN, 85ºW (Wallace and Gutzler, 1981).
- **Pacific Decadal Oscillation (PDO) 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 30ºN–65ºN, 160ºW–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).

[END OF BOX 3.3]

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

3.6.2.1 El Niño-Southern oscillation

El Niño events are a coupled ocean-atmosphere phenomenon, involving warming of surface waters of the tropical Pacific in the region from the International Dateline to the west coast of South America, and result in changes in oceanic circulation and in local and regional ecology. 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 atmospheric counterpart, the Southern Oscillation (SO), is closely linked with these ocean changes: the total phenomenon generally referred to as ENSO. El Niño is the warm phase of ENSO and La Niña is the cold phase.

El Niño events are a striking example of a phenomenon that would not occur without interactions between the atmosphere and ocean (Neelin et al., 1998; Trenberth et al., 2002a). 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 are the surface branch of the large-scale equatorial atmospheric overturning Walker Circulation (see Section 3.7.2), which is strongly influenced by SST. The circulation organizes tropical Pacific precipitation, in turn determining atmospheric heating patterns through the release of latent heat. The heating drives the large-scale monsoon-type circulations in the tropics, and consequently determines the winds. Through atmosphere-ocean feedbacks, perturbations in the winds or the upper ocean circulation and SST can quickly magnify into an El Niño or La Niña event.

ENSO has global impacts, manifested most strongly in the winter months in either hemisphere, as Rossby wave teleconnections to higher latitudes, featuring alternating sequences of high and low pressures accompanied by distinctive wave patterns in the jet stream and storm tracks in mid-latitudes (Chang and Fu, 2002). Associated patterns of surface temperature and precipitation anomalies around the globe are given in Trenberth and Caron (2000; see Figure 3.6.2). Extratropical teleconnections to ENSO are characterised by the Pacific-North American (PNA) pattern for the NH, and the Pacific-South American (PSA) pattern in the SH. Both may be forced by ENSO-related tropical heating anomalies, but are also triggered by other tropical forcings (Renwick and Revell, 1999; Mo, 2000; Straus and Shukla, 2002). For sea level pressure, anomalies are much greater in the extratropics while the tropics feature large precipitation variations. Although warming is generally associated with El Niño events in the Pacific and extends, for instance, into western Canada, cool conditions typically prevail over the mid-latitudes of the North and South Pacific Oceans. To a first approximation, reverse patterns occur during the opposite La Niña phase of the phenomenon. However, the latter is more like an extreme case of the normal pattern with a cold tongue of water along the Equator.

[INSERT FIGURE 3.6.2]

Each El Niño event evolves differently and the strength and importance of ENSO has varied over time. Strong ENSO events occurred in 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 and with a shift to generally above normal SSTs in the western 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 (see Section 3.6.2.2). 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 K higher for the year centred on March 1998 owing to the El Niño. Whether observed changes in ENSO are physically linked to global climate change is a research question of great importance. 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). An example is the modest El Niño of 2001–2002, which was associated with record-breaking heat accompanying a drought in Australia (Nicholls, 2004; and see Box 3.5.2).

3.6.2.2 Tropical-extratropical teleconnections: PNA and PSA

Circulation variability over the extratropical Pacific, on time scales from sub-seasonal to interannual, 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 the Pacific-North American (PNA) and Pacific South American (PSA) patterns.

Over the NH in winter, the PNA teleconnection pattern describes a Rossby wave train that arches across North America from the central subtropical Pacific, with four centres of action (Figure 3.6.1). Across the Pacific, the PNA pattern is related to the strength and extent of the subtropical jet stream, and hence to the Pacific storm track. While the PNA and NAO can be illustrated by taking a single point correlation (Figure 3.6.1), this is not so for the PSA (not shown). However, over the SH, the PSA pattern is present at all times of year, and is a SH analogue of the PNA pattern (Kiladis and Mo, 1998; Mo and Higgins, 1998; Kidson, 1999). The PSA pattern originates over Australia and the tropical Pacific, poleward of the subtropical jet maximum, and propagates towards the southeast Pacific before curving equatorward over the southern Atlantic. Unlike the PNA, the PSA is not fixed in space, and tends to be represented by pairs of orthogonal patterns with wave-3 zonal structures (e.g., PSA-1 and PSA-2; Mo, 2000) and are mainly brought out by filtering the data. Projections onto the PNA and PSA patterns may occur as mid-latitude responses to ENSO heating anomalies. However, significant variability of the PNA and PSA occurs on other time scales, even in the absence of ENSO, indicating that they may also be considered “internal” modes of atmospheric variability (Renwick and Revell, 1999; Mo, 2000), although Straus and Shukla (2002) suggest that the NH tropically forced pattern has distinctly different form.

Both PNA and PSA are associated with significant anomalies of temperature, precipitation, and synoptic weather activity over the extratropical Pacific. The positive PNA 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). PSA-related circulation anomalies are 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; Yuan and Martinson, 2001; Kwok and Comiso, 2002a; Renwick, 2002).

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, and are linked to decadal variations in atmospheric circulation, SST and ocean circulation throughout the whole Pacific Basin (Trenberth and Hurrell, 1994; Gershunov and Barnett, 1998; Folland et al., 2002; Deser et al., 2004). Key measures of Pacific decadal variability are the North Pacific Index (NPI, Trenberth and Hurrell, 1994), the Pacific Decadal Oscillation (PDO) index (Mantua et al., 1997) and the Interdecadal Pacific Oscillation (IPO) index (Power et al., 1999b; Folland et al., 2002); see Figures 3.6.3 and 3.6.4.

Decadal changes in the mean state of the Pacific have an important modulating influence upon ENSO behaviour and hence upon many aspects of the global circulation and climate. The 1976–1977 climate shift in the Pacific, associated with a phase change in the PDO, is marked by significant changes in ENSO evolution and with changes in ENSO teleconnections and links to precipitation and surface temperatures over North America, Asia, and Australasia (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., 2002a; Deser et al., 2004). Such effects have many environmental and human impacts, as discussed for ENSO in Section 3.6.2.

Figure 3.6.4a shows a time series of the NPI for 1900–2005 (Deser et al., 2004). There is substantial low-frequency variability since the beginning of the record, with extended periods of predominantly high values indicative of a weakened circulation (1900–1924 and 1947–1976) and predominantly low values indicative of a strengthened circulation (1925–1946 and 1977–2003). 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 including a bi-decadal rhythm (Minobe, 1999).

Many aspects of Pacific interdecadal climate variability are still under debate. Linkages between the extratropics of both hemispheres and the tropical Indo-Pacific region support a physical mechanism of tropical origin, similar to the paradigm for ENSO (Evans et al., 2001; Deser et al., 2004; Linsley et al., 2004). The decadal-to-interdecadal timescale of tropical Indo-Pacific SST variability is likely due to oceanic processes. However, 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). Extratropical ocean influences are also likely to play a role as changes in the ocean gyre evolve and heat anomalies are subducted and can 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). While the PDO/IPO may be the chance low frequency residual of ENSO variability on multidecadal time scales (Newman et al., 2003), Folland et al. (2002) showed that the IPO significantly affects the movement of the South Pacific Convergence Zone in a way independent of ENSO, suggesting that the IPO may have an independent component (see also Deser et al., 2004). There also remains ambiguity about inter-decadal Pacific-wide features dependence on global warming (Folland et al., 1999; Livezey and Smith, 1999).

[INSERT FIGURE 3.6.3]

[INSERT FIGURE 3.6.4]

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 North Atlantic Oscillation (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 Northern Annular Mode (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 SLP 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).

There is considerable debate over whether the NAO or the NAM is more physically relevant, but the time series are highly correlated (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 show evidence for intervals with prolonged positive and negative NAO index in the last few centuries (Cook et al., 2002). In winter, a reversal occurred from the minimum index values in the late 1960s to strongly positive NAO index values in the early and mid-1990s. Since then NAO values have marginally declined (see Figure 3.6.6). Hurrell et al. (2001, 2002) identified significant interannual to multi-decadal fluctuations in the summer NAO pattern. The trend toward persistent anticyclonic flow over northern Europe in summer has contributed to anomalously warm and dry conditions in recent decades (Rodwell, 2003). The increase in the trend and the variance of NAO/NAM index over roughly the last three decades is greater than would be expected from climate noise alone (Feldstein, 2002). This behaviour contrasts with that of the first 60 years of the 20th century, for which Feldstein (2002) showed that all of the NAO/NAM interannual variability is consistent with atmospheric internal variability. Although sub-seasonal atmospheric events appear to be largely non-predictable, and sampling variability of an internally generated process is strong (Czaja et al., 2003; Thompson et al., 2003), there may be predictability from stratospheric influences (Thompson et al., 2002) (see Box 3.2) and there is mounting evidence that the recent observed interdecadal NAO variability comes from boundary or forcing external to the atmosphere. Longer-term trends of the NAO/NAM appear to be due to external factors and slowly varying boundary conditions over the land surface (Gong et al., 2003; Bojariu and Gimeno, 2003) and especially over the tropical oceans (Hurrell et al., 2004).

The NAO exerts a dominant influence on wintertime temperatures across much of the NH (Figure 3.6.5). Surface air temperature and SST across wide regions of the North Atlantic Ocean, North America, the Arctic, Eurasia and the Mediterranean are significantly correlated with NAO variability. These changes, along with related changes in storminess and precipitation (Figure 3.6.5), ocean heat content, ocean currents and their related heat transport, and sea ice cover, have significant impacts on a wide range of human activities as well as on marine, freshwater and terrestrial ecosystems. 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 1968–1997, the NAM accounted for 1.6 K out of 3.0 K warming in Eurasian surface temperatures, 4.9 out of the 5.7 hPa decrease in sea level pressure from 60ºN–90ºN; 37% out of the 45% increase in Norwegian precipitation (55ºN–60ºN, 5ºE–10ºE), and 33% out of the 49% decrease in Spanish rainfall (35ºN–45ºN, 10ºW–0ºW).

[INSERT FIGURE 3.6.5]

[INSERT FIGURE 3.6.6]

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). Such changes have consequences for the regional ecology, as well as for the operation and safety of shipping, offshore industries such as oil and gas exploration, and coastal development.

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 (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 Southern Annular Mode (SAM), or sometimes as the high-latitude mode (HLM) or Antarctic Oscillation (AAO), see Figure 3.6.7. It is essentially a zonally-symmetric structure 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 is equivalent barotropic and appears as the leading EOF in many atmospheric fields (Thompson and Wallace, 2000; Trenberth et al., 2005b). Model experiments demonstrate that the structure and variability of the SAM result mainly from the internal dynamics of the atmosphere (e.g., Hartmann and Lo, 1998; Limpasuvan and Hartmann, 2000). Poleward eddy momentum fluxes interact with the zonal mean flow to sustain latitudinal displacements of the mid-latitude westerlies (Limpasuvan and Hartmann, 2000). 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 the ERA-40 reanalysis (Renwick, 2004; Trenberth et al., 2005b). However, a declining positive bias in pressure at high southern latitudes in the former reanalysis before 1979 (Hines et al., 2000) 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]

This trend in the SAM, which is statistically significant annually, and in summer and autumn (Marshall et al., 2004), has contributed to recent spatial variability in 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 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. However, an opposite SAM trend in southern winter (JJA) is towards weaker zonal winds over the southern oceans (Fogt and Bromwich, 2005), see Figure 3.5.1. Although the SAM is essentially zonal, a wave-number 3 pattern is superimposed, 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). Indeed, Orr et al. (2004) propose a blocking mechanism whereby the interaction of the increasing westerlies on the Peninsula produces this scenario and also 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 δ^{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).

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).

Attempts have been made to reconstruct century-scale records based on proxies of the SAM. Goodwin et al. (2004) use sodium concentrations from Law Dome while Jones and Widmann (2003) employ tree-ring width chronologies: both these studies stress the decadal variability in their derived SAM time series. The Jones and Widmann (2004), reconstruction indicates that the recent trend is not unprecedented, even during the 20th century, although other studies suggest otherwise (Gillett et al., 2003; Marshall et al., 2004). While natural forcings may have played a role (Marshall et al., 2004), and may act synergistically with anthropogenic forcing (Hartmann et al., 2000), attribution of the observed changes in SAM has been made to ozone depletion (Sexton, 2001; Thompson and Solomon, 2002), and increasing greenhouse gas increases (see Chapter 9).

3.6.6 Other Indices

As noted earlier, many "modes of variability" 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 two climate signals that have recently drawn the attention of scientific community: the Atlantic Multi-decadal Oscillation and the Antarctic Circumpolar Wave.

3.6.6.1 Atlantic multi-decadal osscillation

Over the instrumental period (since the 1850s) North Atlantic SSTs show a 65–70 year variation (0.4 K range), with 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 Atlantic Multidecadal Oscillation (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. 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 dating back the last four centuries (Delworth and Mann, 2000; Gray et al., 2004). The AMO signal seems to extend to other regions, such as the North Pacific. The existence of a pan-oceanic interaction between the tropical Pacific and the Atlantic Ocean at multi-decadal time scales through an "atmospheric bridge" involving anomalous fresh water fluxes has been suggested (Latif, 2001). 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).

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, 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, when influences from the tropical Pacific region, become important, especially for the SST in the western tropical North Atlantic. Chelliah and Bell (2004) defined a tropical multi-decadal mode based on 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, and which incorporates aspects related to the AMO, the PDO and wintertime NAO.

[INSERT FIGURE 3.6.8]

3.6.6.2 Antarctic circumpolar wave

The Antarctic circumpolar wave (ACW) is 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 and description based on filtered data (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) and even its very existence (Park et al., 2004).

To account for the changes in time of the ACW, Venegas (2003) suggested that the ACW 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 year 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, 2000; White and Cherry, 1999).

3.6.7 Summary

ENSO is the dominant mode of variability on interannual time scales and is a truly coupled mode involving the atmosphere and ocean in the tropical Pacific. Although ENSO has a preferred period, its strength and frequency exhibit significant low frequency variations. Of particular note is the 1976–1977 climate shift, related to a phase change in the PDO/IPO. The extent to which Pacific decadal variability is independent of ENSO is not yet clear. Indeed, nonlinearity of precipitation and related processes in their relation to SST suggests that it cannot be fully independent. The nature of the variability becomes richer from interactions with the tropical Atlantic and Indian Oceans, and also from the extratropical North and South Pacific. Responses of the extratropical ocean become more dynamic as the time scale is extended, and processes such as subduction, gyre changes, and the thermohaline circulation come into play.

Such decadal variations in the teleconnections discussed here considerably complicate the interpretation of climate change. 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. Hence accounting for these patterns is vital for understanding regional climate change in Pacific rim countries and ocean changes that affect fisheries and other economic interests.

Since the TAR, it has become clear that a small number of teleconnection patterns account for much of the seasonal-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 where evidence suggests the circumpolar westerlies are increasing in both hemispheres simultaneously. In the NH, this is a major part of 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. 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. Multi-decadal variability is also evident in the Atlantic, and appears to be related to the thermohaline circulation. All of these patterns or perhaps modes of variability have profound influences on the ocean and its food web and marine life, as well as on ecosystems on land.

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, analysis of observed long-term changes have mainly relied on land-based rain gauge data.

The global monsoon system embraces an overturning circulation (Trenberth et al., 2000) that is intimately associated with the seasonal variation of monsoon precipitation over all major continents and adjacent oceans. 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 Maritime Continent, the equatorial parts of western and central Africa, Central America, Southeast Asia, and eastern Australia have been found for 1948–2003 (Chen et al., 2004), while increasing trends are evident over the United States and northwestern Australia (see also Section 3.3.2.2). These results are consistent with those of Dai et al. (1997b). The pattern of trend (Chen et al., 2004) has a large equatorially symmetric component. 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., 2004; GHCN, see Section 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 Chen et al. 2002), despite some differences in details and discrepancies in northwest India. Significant decreases in the annual range are observed over the NH tropical monsoon regions (e.g., Southeast Asia, North Africa, and Central America). The decreasing NH monsoon precipitation is accompanied by an equatorward increase and poleward decrease of monsoon rainfall, which is an indication of weakening of the NH monsoon circulation (Figure 3.7.2a). Over the East Asian monsoon region, the epochal 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 (20ºN–50ºN, 105ºE–140ºE) shows no appreciable change (Figure 3.7.2). In the Indonesian-Australian monsoon region, the change between the two periods it is characterized by an increase in northwest Australia and Java but a decrease in northeast Australia and the SPCZ region (Figure 3.7.1, see also Smith, 2004 who has shown statistically significant increases in monsoon rainfall over northern, western, and central Australia). However, the average rainfall over a large region (5ºS–20ºS, 110ºE–150ºE) shows no long-term trend but significant interannual and interdecadal variations (Figure 3.7.2b), and the same applies over South America for area-averaged summer precipitation (Figures 3.7.1, 3.7.2b). 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 slight decreasing trend in area-averaged precipitation over the South African monsoon region (Figure 3.7.2b), although the statistical significance of this trend has not been established.

Although the relationship of observed convective available potential energy (CAPE) changes to the intensity of the large-scale monsoon is far from clear (e.g., McBride and Frank, 1999; DeMott and Randall, 2004), the predominant use of CAPE as a metric on moist convection potential raises its importance (see also Section 3.8.3.1 for a discussion of changes in CAPE).

Monsoon systems span multiscale variability from days to decades and even centuries, with spatial scales from a few to thousands of kilometres. As a result, 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, highly 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 atmosphere and the land surface (see also Box 3.1).

[INSERT FIGURE 3.7.1]

[INSERT FIGURE 3.7.2]

3.7.1.1 Asia

The Asian monsoon has a profound influence on the social and economic condition of over 60% of the global population. The Asian monsoon can be divided into two systems: the East Asian and the South Asian 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 3.7.3). 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 3.7.3, based on the newly developed HadSLP2 data set (Allan and Ansell, 2005), 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 3.7.3]

Gong and Ho (2002) suggested the change in summer rainfall over the Yangtze River valley was due to a southward shift in the late 1970s and Ho et al. (2003) also noted a sudden change in Korea in the late 1970s. The notable regime shift of summer rainfall over the middle-lower valley of the Yangtze River and over eastern China, which happened in the late 1970s, occurs at the same time as the step change in the 500 hPa geopotential height over the northern Pacific (Gong et al., 2002) (see Section 3.6.3). The change is related to the enlargement, intensification and southwestward extension of the northwest Pacific subtropical high. The change in SSTs over the eastern tropical Pacific and tropical Indian Ocean is primarily responsible for the shift of the summer rainfall over the Yangtze River through changes in the subtropical high. Such changes could be linked to multi-decadal variation of the monsoon circulation in East Asia (Qian et al., 2003) and Pacific decadal variability (Huang et al., 2003). 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 295-year reconstruction of monsoon rainfall in the central Himalayas shows evidence of substantial multi-decadal variability with a weakening monsoon in the 18th century, a strengthening monsoon from the early 19th into the early 20th century, and a weakening monsoon from 1920 to the present (Duan et al., 2004).

The Indian monsoon season occurs from June to September, when about 70% of annual rainfall is received. The Indian summer monsoon rainfall exhibits decadal variability and 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 of SST differences across the equatorial Indian Ocean 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). Changes in SSTs other than ENSO are also important on decadal time scales.

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 of 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). Trends in CAPE in northern Australia are weak and not significant (Gettelman et al., 2002; DeMott and Randall, 2004). Using Australian 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 2003/04 (Figure 3.7.4) (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, as discussed further below. 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 3.7.4). Using northeastern Australian rainfall, Latif et al. (1997) has shown that rainfall was much increased during decades when the 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 3.7.4) 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 the decadal time scale.

[INSERT FIGURE 3.7.4]

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., 2005). 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., 2005). It responds to seasonal changes in the thermal contrast between the continent and the adjacent Atlantic Ocean and it is a key factor for the warm season precipitation regime. In northern Brazil, different precipitation trends have been observed over northern and southern Amazonia, showing a dipole structure suggesting a southward shift of the SAMS (Marengo, 2004). 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 source 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. The seasonal average of this pattern is well correlated (r=0.74) with the Atlantic Multidecadal Oscillation (Section 3.6.6), suggesting that the Atlantic Ocean is a likely source of important climate variability for eastern Africa. Hence 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 1970–1990 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 the rainfall deficit 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 9ºN. The decreasing rainfall and devastating droughts in the Sahel region during the last three decades of the 20th century (Figure 3.7.5) are among the most undisputed and largest climate changes anywhere recognized by the climate research community. Dai et al. (2004a) provided an updated analysis of the normalised Sahel rainfall index based on the years 1920–2003 (Figure 3.7.5). Following the major El Niño event of 1983–1984, 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. (2005) report a decrease in average rainfall intensity and an increase in dry spell length (consecutive dry day length) for 1961–2000.

[INSERT FIGURE 3.7.5]

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 season, 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 SST 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 poleward energy transport mechanism 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. There is also a tendency for a weaker Indian monsoon and drought over Southern Africa (Diaz and Markgraf, 2000). Decadal SST changes may also impact local mass circulations around tropical continents. 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) found 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) associated with numerous climate variables throughout the tropical Indo-Pacific region, including rainfall, cloudiness, SST, and sea level pressure. 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 He (2002) indicated a significant decadal shift of the Northwest Pacific subtropical high about 1979/1980, since 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).

In summary, large interannual variability associated with ENSO dominates the HC and monsoons, and there is good evidence for decadal changes that are associated with monsoonal rainfall changes, especially across the 1976–1977 climate shift, but data uncertainties compromise evidence for trends.

3.8 Changes in Extreme Events

3.8.1 Background

There is increasing concern that climate extremes may be changing in frequency and intensity as a result of human influences on climate. Extremes generally refer to rare events based on a statistical model of particular weather elements. Indeed climate change is characterized by changes in variability and extremes, as well as changes in the mean. As illustrated in the TAR, changes in extremes may relate to changes in the mean and variance in complicated ways. For any change in the mean, there will always be larger changes in extremes of one or both signs on a percentage basis. Climate change may be perceived most in 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.

In this section 'weather extremes' and 'climate extremes' are used almost interchangeably, as weather and climate are part of the same continuum. However, for climate extremes we need to look for a pattern of behaviour over several synoptic weather events; examples are given in Box 3.5. Changes in extremes are assessed, therefore, at all time and space scales (e.g., from extremely warm years globally to peak rainfall intensities locally). However, the availability of observational data restricts the type of extremes that can be analysed. Also in the TAR, extremes generally refer to rare events based on a statistical model of particular weather elements. 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., 2004b). 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). Data are generally required at a daily (or less) time scale.

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.A.2 and 3.A.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 indices of extremes provides an unprecedented global picture of changes in daily temperature and precipitation extremes (Alexander et al., 2005a 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.A.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 for the era 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 satellite information over the oceans, which is expected to significantly affect storm activity.

In this section observational evidence for changes in extremes is assessed for temperature, precipitation, wet spells, 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 and 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 nighttime temperatures across much of the globe.

Since the TAR, a number of regional studies confirm the picture of a gradual reduction in recent decades of frost days over North America (Kunkel et al., 2004; Vincent and Mekis, 2005), Western Europe and East Asia (Kiktev et al., 2003) and China (Zhai and Pan, 2003), whereas an analysis over a 150 year period shows that this decrease of cold extremes for winter (DJF) in Europe and China persists throughout the whole period (Yan et al., 2002). Consistent with this trend, a significant increase in the number of days with thaw is found in Fennoscandia during the second half of the 20th century (Groisman et al., 2003) as well as intense warming of the lowest daily minimum temperatures over North America (Robeson, 2004). In addition, Groisman et al. (2003, 2005a) found a significant decrease (twofold in MAM) in very cold nights (i.e., more than two standard deviations above/below the mean monthly values of T_{min} for each month) over the Arctic in all seasons except autumn (SON) over the last 50 years and similar changes during daytime but with smaller amplitude. This warming was accompanied by a reduction of the daily temperature variability in all seasons. Even in tropical regions such as the Caribbean, the percent of days with very cold temperatures (lowest 10%) has decreased since the late 1950s (Peterson et al., 2002). Large parts of Africa show coherent increases in warm nighttime temperatures (upper 10%) (Easterling et al., 2003).

A number of recent regional studies have been completed for southern South America (Vincent et al., 2005), Central America and northern South America (Aguilar et al., 2005), southern and western Africa (New et al., 2005), the Middle East (Zhang et al., 2005), Australasia and southeast Asia (Griffiths et al., 2005), and central and southern Asia (Klein Tank et al., 2005). 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 nighttime and daytime 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, both minimum temperature extremes and maximum temperature extremes have increased. Such warming of both the nighttime and daytime extremes was also found for the other regions where data have been analysed. Only for Australasia and southeast Asia are the results not directly comparable, because the extremes considered are further towards the tail of the distribution (at 1% and 99% rather than 10% and 90%). For stations considered to be non-urban in this region, the dominant distribution change 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 mean, impacting on either one or both distribution tails, as well as a significant change in 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 (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) find 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. (2005) analysed 22 Spanish records for the period 1894–2003 and find 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. Vincent and Mekis (2005) find progressively fewer extreme cold nights (cold days) and conversely more extreme warm nights (warm days) for Canada between 1900–2003. 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. (2005a) has brought all these and other regional results together, gridding the common indices for the period 1951–2003. Over 76% of the global land area sampled showed a significant decrease in the annual occurrence of cold nights and a significant increase in the annual occurrence of warm nights tok place over 72% of the regions (Table 3.6, Figure 3.8.1 and Question 3.3). Some regions experienced a more than doubling of these indices. 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 stronger increase in minimum temperature than maximum temperature leading to a reduction in DTR (see Section 3.2.2.1). The change in the four extremes indices (Question 3.3, Figure 1, 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 (Figure 3.8.1). The fact that the number of cold and warm events represented by the indices have changed almost equally, 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. Trends, ±2 standard error ranges and significances (**bold: <1%**; *italic, 1%–5%*) were estimated by Restricted Maximum Likelihood (Diggle et al., 1999, see Appendix 3.A.1.2) 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. 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]

3.8.2.2 *Precipitation*

For precipitation, which is inherently intermittent, 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. Due to warming, observed increases in water vapour (Section 3.4.2) imply increases in intensity, but reduced frequency or duration in some sense. 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 not coherent, 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% quantiles 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; Brunettti 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, northern 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., 2005a).

A number of recent regional studies have been completed for southern South America (Haylock et al., 2005), Central America and northern South America (Aguilar et al., 2005), southern and western Africa (New et al., 2005), the Middle East (Zhang et al., 2005), and central and southern Asia (Klein Tank et al., 2005). 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., 2005). 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 (leading to a southward shift in storm tracks) has had an important effect on the observed rainfall trends. No significant increases in the total amounts are found over Central America and northern South America, 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 spatial 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. (2005a) 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 annual precipitation does show widespread increases since 1950 (see also Section 3.3). 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 (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 for many extratropical regions (Groisman et al., 2005a) (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 in 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 (2005) 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. (2005a) 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]

Since the TAR, several regional analyses have been undertaken for statistics with return periods much higher 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., 2005a), 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 no evidence that the changes at the extreme tail of the distribution had opposite sign from changes inferred for more robust statistics based on 75th and 95th percentiles.

[START OF BOX 3.4]

Box 3.4: Tropical Cyclones and Climate Change

As human-induced climate change increases global warming, the environment in which tropical storms form changes. Hurricanes and typhoons generally form only where SSTs exceed about 26ºC and SSTs are expected to rise, thereby potentially expanding the areas over which such storms can form. Further, surface atmospheric temperatures and water vapour increase and thereby increase the moist static energy that fuels convection and thunderstorms. However, whether or not Convective Available Potential Energy (CAPE) increases depends on changes in atmospheric circulation, especially subsidence, and the static stability of the atmosphere. The potential intensity, defined as the maximum wind speed achievable in a given thermodynamic environment (e.g., Emanuel, 2005), similarly depends critically on SSTs and atmospheric structure. While changes in atmospheric composition increase radiative heating aloft, thus stabilizing the atmosphere, the tropospheric lapse rate is maintained mostly by convective transports of heat upwards (either in convection, thunderstorms, or thunderstorm complexes, including mesoscale disturbances, various waves, and tropical storms) and radiative cooling (not heating). In models, the parameterization of sub-grid scale convection plays a critical role in determining large-scale CAPE and whether it is released or not. In addition, changes in winds and wind shear can affect large-scale CAPE. All of these factors, in addition to SSTs determine whether convective complexes become organized as rotating storms and form a vortex. Increases in low level moist static energy suggest that there will be a corresponding increase in convection and thunderstorms in the tropics, but whether or not they become organized as tropical storms, hurricanes or typhoons is not clear. Once such storms are formed, however, the potential for greater intensity and duration is clearly present.

From an observational perspective then, key issues are the tropical storm formation regions, the frequency, intensity 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.

[END OF BOX 3.4]

3.8.3 Evidence for Changes in Tropical and Extratropical Storms and Extreme Events

3.8.3.1 Tropical cyclones

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 3.4. 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.

An important measure of the total seasonal activity is NOAA's Accumulated Cyclone Energy (ACE) index (Levinson and Waple, 2004), which accounts for the collective intensity and duration of Atlantic tropical storms and hurricanes during a given hurricane season (see Figure 3.8.3). The ACE index 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 from the discontinuities inherent in more widely used measures of activity such as the number of tropical storms, hurricanes, or major hurricanes. The ACE index is also used to define above-, near-, and below-normal hurricane seasons. 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). Prior to about the late 1960s, when satellite imagery became used to help estimate the intensity and strength of tropical storms, the estimates of ACE values 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.

In addition, empirical rules have been developed for the Potential Intensity (PI) of tropical cyclones (Emanuel, 2003) that can be computed from observational data based primarily on vertical profiles of temperature and humidity, CAPE, and SSTs. In analysing CAPE from selected tropical radiosonde stations around the globe 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); see Section 3.2 and Appendix 3.A.5. Hence, homogeneity of radiosonde data and reanalyses compromise results, although it seems clear that trends in PI are small and statistically insignificant at a scattering of stations in the tropics (Free et al., 2004a).

Of more direct relevance is the destructiveness of tropical storms, which relates to the total power dissipation within each storm (Emanuel, 2005), as the main dissipation is from surface friction and wind stress effects. Consequently it is proportional to the wind speed cubed. An index of the total power dissipation (Emanuel, 2005) shows substantial upward trends beginning in the mid 1970s, roughly doubling since then. It comes about because of longer storm lifetimes and greater storm intensity, and 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 even as total number of cyclones and cyclone days has decreased in most basins. The largest increase is in the North Pacific, Indian and Southwest Pacific oceans.

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, by far the highest ACE year is 1997, when a major El Niño event occurred and surface temperatures were subsequently the highest on record (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 and 2004 the fourth highest globally in ACE values. Emanuel's (2005) 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]

3.8.3.1.1 Western North Pacific

In the western North Pacific, the annual total number of typhoons decreased from the 1960s to the 1970s, and then increased from 1980 to the mid-1990s before decreasing again (Chan and Liu, 2004), although with further increases in the last few years after that study was completed (Figure 3.8.3). For tropical cyclones making landfall in Guangdong Province in southern China between 1949 and 2000, no obvious long-term trend can be discerned (He et al., 2003; Liu and Chan, 2003). The annual number of tropical cyclones making landfall along the south China coast within 300 km of Hong Kong between 1961 and 2002 decreased slightly but not significantly (Leung et al., 2004). A strong interdecadal variation overwhelms any trends (Chan and Liu, 2004).

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; Leung et al., 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, 140ºE–180ºE) and decrease in the northwestern quadrant (17ºN–30ºN, 120ºE–140ºE) (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 21 typhoons, 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)

3.8.3.1.2 North Atlantic

Beginning with 1995 all but two Atlantic hurricane seasons have been above normal. The exceptions are the two El Niño years of 1997 and 2002. ENSO significantly impacts seasonal Atlantic hurricane activity (Gray 1984; Bove et al., 1998), with El Niño acting to reduce activity and La Niña acting to increase activity. 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 active phase of the AMO (see Section 3.6.6.1) has also been a primary contributing factor to the increased hurricane activity since 1995 (Goldenberg et al., 2001) and is well depicted in Atlantic SSTs (Figure 3.6.8), including those in the tropics.

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. NOAA classifies all but two of these nine seasons (the exceptions being the El Niño years of 1997 and 2002) as above normal. 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). The active phase of the Atlantic multi-decadal signal 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, very high SSTs and favourable atmospheric conditions have already (July) led to a very active season with two tropical storms in June and two major hurricanes in July.

Key factors in the recent increase in 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 (MDR) where most Atlantic hurricanes form (Aiyyer and Thorncroft, 2005) 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.

The most recent decade has the highest SSTs on record in the tropical North Atlantic (Figure 3.6.8), apparently as part of a trend and a favourable phase of the AMO. Higher SSTs are generally accompanied by higher water vapour in the lower troposphere (Section 3.4.2.2 and Figure 3.4.5) and higher low-level moist static energy. This also acts to increase CAPE, so that it provides an environment favouring enhanced convection. These factors also increase Potential Intensity of hurricanes (Emanuel, 2003; 2005), which is thought to be the most relevant thermodynamic variable. However in the mid-troposphere, CAPE also depends a great deal on large-scale subsidence, that is often associated with wind shear (e.g., see Chan and Liu, 2004). During El Niño events, the main action shifts to the Pacific and large-scale subsidence in the Atlantic sector adversely affects CAPE and the environment for tropical storms to develop, while the reverse is true during La Niña. However, generally in the Atlantic, the changing environmental conditions have been more favourable for tropical storms to develop in the past decade. In 2004 the Power Dissipation Index (Emanuel, 2005) is by far the highest on record since 1930 for the North Atlantic.

[INSERT FIGURE 3.8.3]

3.8.3.1.3 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 Quasi-Biennial Oscillation (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), however values are unreliable prior to 1970. 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 trends are not apparent.

3.8.3.1.4 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 (SIO) is normally active from December through April (note, in Figure 3.8.3 the SIO tropical cyclone data are summarized by calendar year, 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. ENSO is a factor in tracks of tropical cyclones and risk of tropical cyclone landfall on Mozambique (see 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.

3.8.3.1.5 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). The period of reliable tropical cyclone data is too short for reliable trends to be determined.

3.8.3.1.6 South Atlantic

In late March 2004 in the South Atlantic, off the coast of Brazil, the first and only 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 prior to 1970 or so. 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 (Section 3.6).

3.8.3.2 Extratropical cyclones

Extratropical cyclones are low pressure systems that occur throughout the mid-latitudes of both hemispheres as baroclinic instabilities that reduce horizontal temperature gradients. Intense extratropical cyclones are generally accompanied by severe windstorms. Significant increases in the number (or intensity) of intense extra-tropical cyclone systems have been documented in a number of studies (e.g., Lambert, 1996; Gustafsson, 1997; McCabe et al., 2001) with associated changes in the preferred tracks of storms as described further 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. In some cases earlier results have been questioned since the TAR, 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 the North Atlantic (Graham and Diaz, 2001; Gulev et al., 2001; Simmonds and Keay, 2002), but only the North Pacific trend is found statistically significant (Simmonds and Keay, 2002; Wang et al., 2005). 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 the decadal variability in the NH, linked to changes in the 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/DOE Reanalysis-2 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 and followed again by a decline, see Figure 3.8.4 (Alexandersson et al., 2000). A positive correlation (0.4-0.6) was found between storminess and the NAO. 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 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., 2005b). Thus the pressure station data for the North Atlantic region show a modest increase 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 (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 NCEP reanalyses 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. Hence local differences across northwestern Europe 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]

3.8.3.3 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. 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 mainly concerns weak tornadoes (F0 and F1 on the Fujita scale), thus paralleling the evolution of tornado reports in the United States after 1950 and making it likely that this increase in reports in Europe is at least dominated (if not solely caused) by enhanced detection and reporting efficiency.

Meehl et al. (2000) and Easterling et al. (2000) 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 (SCSWP), the density of surface meteorological observing stations is too coarse to measure all such events. Moreover, homogeneity of existing station series is questionable.

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), and appears linked to the reduced cyclone frequency and increasing winter (DJF) temperatures (Qian et al., 2002) along with vegetation degradation and drought, plus increased surface wind speed. (Zou and Zhai, 2004; Wang and Zhai, 2004).

[START OF BOX 3.5]

Box 3.5: 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 recent (post-TAR) extreme climate events where the odds may have shifted to make them more likely than in an unchanging climate. A lack of long and homogeneous observational data makes it difficult to place some of these events in a longer context. Attribution of the change in odds typically requires extensive model experiments; a topic taken up in Chapter 9. Thus, in the absence of such extensive testing, the most that can be said is that individual events are consistent with expectations arising from climate change.

Box 3.5.1 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 3.5.2 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. As well, 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 and were much higher than previously during droughts. The mean annual maximum temperature for 2002 was 0.5 K warmer than during the 1994 drought and 0.9 K 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 3.5.3 Drought in the 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 (USGS, 2004; 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). 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.

During the dry and unusually warm summers in the recent drought millions of hectares in the forested region of the southwestern United States suffered forest dieback and very extensive summer wildfire seasons (e.g., McKenzie et al., 2004). Warmer winters allowed pine beetles and other insects to take hold, weakening trees. In historic climate regimes, western United States pine beetle populations were mostly limited to lower elevation pines; but the warmer winters have allowed this insect to expand its range into high elevation, whitebark pine ecosystems with devastating consequences.

Accompanying the dryness and high temperatures was a widespread tendency toward early snowmelts. 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).

Global climate anomalies were involved in provoking these drought patterns. 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. Further warming may foster extensive regional drought (Cook et al., 2004). 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]

Box 3.5.4 Floods in Europe, Summer 2002

A catastrophic flood occurred along several central European rivers in August 2002. The floods resulting from this 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 Erz 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 Erz

Mountains and western parts of the Czech Republic. A slowly moving low pressure system moved 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 Erz 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 of the 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 3.5.5 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 and the United Kingdom according to the information supplied by national weather agencies (WMO, 2003). 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 K above the 1961–1990 average) and 1.4 K warmer than any other summer in this period (next warmest 1807). Luterbacher et al. (2004) estimate that 2003 is likely to have been the warmest summer since 1500 based on earlier documentary records. The 2003 heat wave was associated with a very robust and persistent blocking high pressure system that some weather services suggested may be a manifestation of an exceptional northward extension of the Hadley Cell. 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]

[END OF BOX 3.5]

3.8.4 Summary

Even though our archived datasets are not very good for determining long-term trends in extremes, new findings on observed changes arise for different types of extremes. 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 extremes has also increased, and heat waves have increased in duration. So far, no consistent answer is available to the question of whether the reduction of cold extremes is systematically higher than the increase in warm extremes, leading to decreased variability in the daily temperature distribution. In some regions the opposite change has been found.

Analysis of updated trends and results for regions that were missing at the time of the TAR show increases in heavy precipitation 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 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, and the AMO also has a strong influence on multi-decadal variability especially in the Atlantic and eastern North Pacific. Limitations on quality of data in most basins compromise evaluations of trends, but it is evident that any trends are small compared with the variability. The records are probably most complete in the Atlantic, where higher decadal mean SSTs accompany 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 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 occurred, 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 3.3.3, the PDSI shows a large drying trend from the middle of the 20th century over northern hemispheric land 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.

[START OF QUESTION 3.3]

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

Heat waves have increased and widespread increases have occurred in warm nights. Droughts also have increased as precipitation over land has decreased somewhat while evaporation has been enhanced by warmer conditions. Generally, heavy daily precipitation events have increased but not everywhere. Tropical storm and hurricane frequencies have varied and while overall trends are fairly small, evidence suggests 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. 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 show warming as well, but generally less marked. The distributions of minimum temperature and maximum temperatures have not only shifted, but also changed in shape such the cold extremes have warmed considerably more than the warm extremes over the last 50 years (Question 3.3, Figure 1). Nevertheless, this does imply increases in 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. For more rare precipitation events, increasing trends are reported as well but results for such extremes are available only for few areas.

Drought is by far the simplest extreme to measure and, due to its long duration, it is not dependent 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 northern hemispheric land since the mid-1950s, with widespread drying over much of Eurasia, northern Africa, Canada and Alaska. In the southern hemisphere, 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 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. Based on the PDSI data, one study shows that 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 El Niño/Southern Oscillation (ENSO)-induced precipitation decrease 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 stroms around the world. Increases may have occurred in recent years, but apart from the North Atlantic basin, most measures only begin in the 1950s or 1960s and have likely missed some events in the early decades. Numbers of hurricanes in the North Atlantic have been above normal in 8 of the last 10 years, but levels were about as high in the 1950s and 1960s. However, estimates of potential destructiveness of hurricanes show a substantial upward trend since the 1970s, with a trend toward longer lifetimes and greater storm intensity. The most active hurricane/typhoon season was 1997 in association with the warmest 12-month period on record and the major 1997-98 El Niño event. There is some evidence for a long-term change in extratropical storminess, but in the best-observed region (the northeastern Atlantic) storminess estimated using simple counts, estimated winds and deepness of the centres during the 1990s only slightly exceeded that in the 1880s and 1890s. The variations that have occurred are related to the North Atlantic Oscillation. 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 the phenomenon.

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 held by members of the public make good pictures for news bulletins, fostering a belief that weather-related extremes are increasing in frequency. An extreme weather event becomes a weather-related 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. However, the net effect is to transform more extreme events into climatic disasters, with major adverse impacts on human welfare.

[INSERT QUESTION 3.3, FIGURE 1]

[END OF QUESTION 3.3]

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 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 focus 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. The main sections where more detailed information can be found are given in square brackets following each bullet. Figure 3.9.1 presents the synthesis schematically for firstly the temperature-related variables, and secondly for the moisture-related variables. [Currently these are from the TAR: to be updated].

- Global-mean surface temperatures show overall warming of 0.75 K over the 1901–2004 period (using 20-year smoothed data) although rates of temperature rise are much greater after 1979. 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 mass and thermal inertia of the oceans. Some areas have not warmed in recent decades, and a few have cooled although not significantly. [3.2.2; 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. The increase in the number of daily warm extremes and reduction in cold extremes across over 70% of land regions studied have been most marked at night over the 1951–2003 period. The greater increase in extreme nighttime as opposed to daytime temperatures has continued. [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 increase of DTR overall has likely reduced to negligible values when considered over the 1979–2004 period. [3.2.2; 3.4.3]
- The temperature increases are consistent with the observed nearly worldwide reduction in glacier and ice cap mass and extent in the 20th century. Glaciers and ice caps respond not only to temperatures but also changes in precipitation, and both global mean winter accumulation and summer melting have increased over the last half century in association with temperature increases. Precipitation anomalies are also important before 1900 in glacier fluctuations. 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 as 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. [4.5]
- Snow cover has decreased in many NH regions, particularly in the spring season, 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 up to 3 K 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 Eurasia continent. [3.3.2.3; 4.2.4, 4.8]
- Sea-ice extents have decreased in the Arctic, particularly in the spring and summer seasons, and patterns of the changes are consistent with regions showing a temperature increase, although changes in winds are also a major factor. 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 evolution of surface temperature south of 65°S, which also shows no trend over that period. Decreases are found in the length of the freeze season of river and lake ice. [3.2.2.3; 4.3, 4.4, 5.3.3]
- Surface temperature variability and trends since 1979 are 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, different satellites and stratospheric influence on the T2 records, and also with ERA-40 estimates of lower-tropospheric temperatures. The range from different datasets of global surface warming since 1979 is 0.15 to 0.18 compared to 0.12 to 0.19 K decade^{-1} for MSU estimates of lower tropospheric temperatures. Over extratropical land, the larger warming at night is associated with larger surface temperature changes. [3.2.2; 3.4.1; Question 3.1]
- Stratospheric temperature estimates from radiosondes, satellites (T4) and reanalyses are in qualitative agreement recording a cooling of between 0.3 and 0.8 K decade^{-1} since 1979. Increasing evidence suggests increasing warming with altitude from 1979 to 2004 from the surface through much of the troposphere in the tropics, cooling in the stratosphere, and a higher tropopause, consistent with expectations from observed increased greenhouse gases and changes in stratospheric ozone. [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. [3.4.3; 3.4.4]
- 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. [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 evaporation. Although

records are sparse, continental-scale estimates of pan evaporation show decreases, due to decreases in surface radiation associated with increases in clouds, changes in cloud properties, and increases in air pollution in different regions from 1970 to 1990. There is tentative evidence to suggest that this has reversed in recent years. The inferred enhanced evaporation and reduced temperature increase is physically consistent with enhanced latent versus sensible heat fluxes from the surface in wetter conditions. [3.3.5; Question 3.2; 3.4.4.2]

- 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. [3.4.3]
- Consistent with rising amounts of water vapour in the atmosphere, increases in the numbers of heavy precipitation events (e.g., 90/95th 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. [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. [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 evaporation and drying associated with warming and decreased precipitation are important factors in increases in drought. 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. [3.3.4, Question 3.2; 3.8.3, 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 increasing precipitation over the northern oceans and in adjacent land areas (as observed). [3.3.2, 3.3.3, 5.3.2, 5.5.3]
- Changes in the large-scale atmospheric circulation are apparent. Increasing mid-latitude westerlies have been evident in both hemispheres as enhanced annular modes. In the NH, the NAM and NAO change the flow from oceans to continents and are a major part of the wintertime observed change in storm tracks, 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. [3.5, 3.6, 3.8.3]
- The 1976–1977 climate shift toward more El Niños has affected Pacific Ocean islands and Pacific rim countries and monsoons throughout the tropics. 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. [3.6, 3.7]
- Variations in extratropical storminess are mostly strongly associated with changes in mean atmospheric circulation, such as changes and variations in ENSO, NAO, PDO, and SAM. Wind and significant wave height analysis support the reanalysis-based evidence for an increase in extratropical storm activity in the NH in recent decades. After the mid 1990s, however, some of these variations have likely changed sign. [3.5, 3.6, 3.8.3.2]
- 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. Increases in intensity and lifetimes of tropical storms since the 1970s are consistent with increases in SSTs and atmospheric water vapour. [3.8.3.1]
- Sea level likely rose about 18 ± 3 cm during the 20th century, but increased 3.0 ± 0.4 mm/year after 1992, when confidence increases from global altimetry measurements. During this period, glacier melt has increased ocean mass by approximately 1.0 mm/year, increases in ocean heat content and associated ocean expansion are estimated to contribute 1.6 mm/year, while changes in land water

storage are uncertain but may have taken water out of 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.[4.5, 4.7, 4.9.8, 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 not 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 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 winds, 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 just as much from increased evaporation 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, and sea level rise.

References

Abakumova, G.M., et al., 1996: Evaluation of long-term changes in radiation, cloudiness, and surface temperature on the territory of the former Soviet Union. *J. Climate*, **9,** 1319–1327.

Ackerman, A.S., et al., 2000: Reduction of tropical cloudiness by soot. *Science*, **288**, 1042–1047.

Adam, J.C. and D.P. Lettenmaier, 2003: Adjustment of global gridded precipitation for systematic bias, *J. Geophys. Res*., **108,** 4257, doi:10.1029/2002JD002499.

Adler, R.F., et al., 2001: Intercomparison of global precipitation products: The Third Precipitation Intercomparison Project (PIP-3). *Bull. Amer. Meteor. Soc.*, **82**, 1377–1396.

Adler, R.F., et al., 2003: The version 2 Global Precipitation Climatology Project (GPCP) monthly precipitation analysis (1979–present). *J. Hydrometeor*., **4**, 1147–1167.

Agudelo, P.A. and J.A. Curry, 2004: Analysis of spatial distribution in tropospheric temperature trends. *Geophys. Res. Lett.,* **31**, L22207, doi:10.1029/2004GL02818.

Aguilar E., et al., 2003: Guidelines on Climate Metadata and Homogenization. WCDMP-No. 53, WMO-TD No. 1186, World Meteorological Organization, Geneva, Switzerland, 55 pp.

Aguilar, E., et al., 2005: Changes in precipitation and temperature extremes in Central America and northern South America, 1961–2003, *J. Geophys. Res*., submitted.

Aiyyer A.R. and C. Thorncroft, 2005: Climatology of vertical wind shear over the tropical Atlantic. *J. Climate*, revised submission.

Aksoy, B., 1997: Variations and trends in global solar radiation for Turkey. *Theor. Appl. Climatol*., **58**, 71–77.

Alexander, L., et al., 2005a: Global observed changes in daily climate extremes of temperature and precipitation. *J. Geophys. Res*., submitted.

Alexander, L., S. Tett, and T. Jónsson, 2005b: Recent observed changes in severe storms over the United Kingdom and Iceland. *Geophys. Res. Lett*., **32,** L13704, doi:10.1029/2005GL022371.

Alexandersson, H., et al., 1998: Long-term variations of the storm climate over NW Europe. *The Global Atmosphere and Ocean System*, **6**, 97–120.

Alexandersson H., et al., 2000: Trends of storms in NW Europe derived from an updated pressure data set. *Climate Res*., **14**, 71–73.

Allan, R. and T. Ansell, 2005: A new globally complete monthly historical gridded mean sea level pressure data set (HadSLP2); 1850–2003. *J. Climate*, submitted.

Allan, R.P. and A. Slingo, 2002: Can current climate model forcings explain the spatial and temporal signatures of decadal OLR variations? *Geophys. Res. Lett*., **29**, 1141, doi:10.1029/2001GL014620.

Allan, R.P., A. Slingo, and V. Ramaswamy, 2002: Analysis of moisture variability in the European Centre for Medium-Range Weather Forecasts 15-year reanalysis over tropical oceans. *J. Geophys. Res.*, **107**, doi:10.1029/2001JD001132.

Allan, R.P, M.A. Ringer, and A. Slingo, 2003: Evaluation of moisture in the Hadley Centre Climate Model using simulations of HIRS water vapour channel radiances. *Quart. J. Roy. Meteor. Soc*., **128**, 1–18.

Allan, R.P, et al., 2004: Simulation of the Earth's radiation budget by the European Centre for Medium-Range Weather Forecasts 40-year reanalysis (ERA40). *J Geophys. Res*., **109**, D18107, doi:10.1029/2004JD004816.

Allan, J., and P. Komar, 2000: Are ocean wave heights increasing in the eastern North Pacific? *Eos,Trans.Amer.Geophys.U.*, **81**, 561–567.

Allen, M.R. and W.J. Ingram, 2002: Constraints on future changes in climate and the hydrological cycle. *Nature*, 419, 2224–2232.

Alley, W.M., 1984: The Palmer Drought Severity Index: limitation and assumptions. *J. Climate Appl. Meteor*., **23**, 1100–1109.

Alpert, P., et al., 2002: The paradoxical increase of Mediterranean extreme daily rainfall in spite of decrease in total values. *Geophys. Res. Lett*., **29**(11), doi:10.1029/2001GL013554.

Ambrizzi T., B.J. Hoskins, and H.-H. Hsu, 1995: Rossby wave propagation and teleconnection patterns in the austral winter. *J. Atmos. Sci*., **52,** 3661–3672.

Andrea F., et al., 1998: Northern Hemisphere atmospheric blocking as simulated by 15 atmospheric general circulation models in the period 1979–1988. *Climate Dyn*., **14**, 385–407.

Andreae, M.O., et al., 2004: Smoking rain clouds over the Amazon, *Science*, **303**, 1337–1342.

Andrews, D.G., J.R. Holton, and C.B. Leovy, 1987: *Middle Atmosphere Dynamics*. Orlando: Academic Press, San Diego, 489 pp.

Angell, J.K., 2003: Effect of exclusion of anomalous tropical stations on temperature trends from a 63-station radiosonde network, and comparison with other analyses. *J. Climate*, **16**, 2288–2295.
Angell, J. K., 2005: Variation in size and position of the 300 mb north circumpolar vortex, 1963–2001. *J. Climate*, submitted.
Aoki, S., 2002: Coherent sea level response to the Antarctic Oscillation. *Geophys. Res. Lett.*, **29**, 1950, doi:10.1029/2002GL15733.
Ashok, K., Z. Guan, and T. Yamagata, 2001: Impact of the Indian Ocean Dipole on the decadal relationship between the Indian monsoon rainfall and ENSO. *Geophys. Res. Lett*., **28**, 4499–4502.
Assell, R., K. Cronk, and D. Norton, 2003: Recent trends in Laurentian Great Lakes ice cover, *Climatic Change*, **57**, 185–204.
Auer, I., et al., 2005: A new instrumental precipitation dataset in the greater alpine region for the period 1800–2002. *Int. J. Climatol.*, **24**, 139–166.
Baldwin, M. and T. Dunkerton, 1999: Downward propagation of the Arctic Oscillation from the stratosphere to the troposphere, *J. Geophys. Res*., **104**, 30,937–30,946.
Baldwin, M.P., 2001: Annular modes in global daily surface pressure. *Geophys. Res. Lett.*, **28**, 4115–4118.
Baldwin, M.P., and T.J. Dunkerton, 2001: Stratospheric harbingers of anomalous weather regimes. *Science*, **244**, 581–584.
Baldwin, M.P., et al., 2003: Stratospheric memory and extended-range weather forecasts. *Science*, **301**, 636–640.
Balling, R., and S. Brazel, 1987: Recent changes in Phoenix summertime diurnal precipitation patterns. *Theor. Appl. Climatol*., **38**, 50–54.
Balling, R.C., Jr., et al., 1998: Impacts of land degradation on historical temperature records from the Sonoran Desert. *Climatic Change*, **40**, 669–681.
Balling, R.C., Jr. and R.S. Cerveny, 2003: Vertical dimensions of seasonal trends in the diurnal temperature range across the central United States. *Geophys. Res. Lett.*, **30**, 1878, doi:10.1029/2003GL017776.
Barlow, M., H. Cullen, and B. Lyon, 2002: Drought in central and southwest Asia: La Niña, the warm pool, and Indian Ocean precipitation. *J. Climate*, **15**, 697–700.
Barnett, T..P., et al., 1999: Interdecadal interactions between the tropics and midlatitudes in the Pacific basin. *Geophys. Res. Lett*., **26**, 615–618.
Barnston, A.G. and R.E. Livezey, 1987: Classification, seasonality and persistence of low frequency atmospheric circulation patterns. *Mon. Wea. Rev.*, **115**, 1083–1126.
Bärring, L. and H. von Storch, 2004: Scandinavian storminess since about 1800. *Geophys. Res. Lett.*, **31**, L20202, doi:10.1029/2004GL020441.
Barriopedro, D., et al., 2005: A climatology of Northern Hemisphere blocking. *J. Climate*, accepted.
Barros, V., et al., 2000: Influence of the South Atlantic Convergence Zone and South Atlantic Sea Surface Temperature on interannual summer rainfall variability in southeastern South America. *Theor. and Appl. Meteor*., **67**, 123–133.
Barros, V.R., et al., 2002: Climate Variability over South America and the South America Monsoon: A Review. *Meteorologica*, **27**, 33–57.
Barros, V.R., M.E. Castañeda, and M.E. Doyle, 2000: Recent precipitation trends in southern South America east of the Andes: An indication of climatic variability. In: *Southern Hemisphere Paleo- and Neoclimates. Key Sites, Methods, Data and Models.* [Smolka, P.P and W. Volkheimer (eds.)]. Springer Verlag, Berlin, pp. 187–206.
Barros, V.R., et al. 2004: The major discharge events in the Paraguay River: magnitudes, source regions and climate forcings. *J. Hydrometeorol.*, **6**, 1161–1170.
Basher, R. E. and X. Zheng, 1995: Tropical cyclones in the southwest Pacific: Spatial patterns and relationships to Southern Oscillation and sea surface temperature. *J. Climate*, **8**, 1249-1260.
Bates, J.J. and D.L. Jackson, 2001: Trends in upper-tropospheric humidity. *Geophys. Res. Lett*., **28**, 1695–1698.
Bengtsson, L., S. Hagemann, and K.L. Hodges, 2004: Can climate trends be calculated from reanalysis data? *J. Geophys. Res*., **109**, D11111, doi:10.1029/2004JD004536.
Beniston, M. and H. F. Diaz, 2004: The 2003 heat wave as an example of summers in a greenhouse climate? Observations and climate model simulations for Basel, Switzerland. *Global and Planetary Change*, **44**, 73–81.
Beniston, M., 2004: The 2003 heat wave in Europe. A shape of things to come? *Geophys. Res. Lett.*, **31**, L02022, doi:10.1029/2003GL018857.

Berg, A.A., et al., 2003: Impact of bias correction to reanalysis products on simulation of North American soil moisture and hydrologic fluxes. *J. Geophys. Res.*, **108**, 4490, doi:10.1029/2002JD003334.
Berri, G.S., M.A. Ghietto, and N.O. García, 2002: The influence of ENSO in the flows of the Upper Paraná River of South America over the past 100 years. *J. Hydrometeorol.*, **3**, 57–65.
Berry, D.I., E.C. Kent, and P.K. Taylor, 2004: An analytical model of heating errors in marine air temperatures from ships. *J. Atmos. Oceanic Technol.*, **21**, 1198–1215.
Betts, A.K., J.H. Ball, and P. Viterbo, 2003: Evaluation of the ERA-40 surface water budget and surface temperature for the Mackenzie River Basin. *J. Hydrometeorol.*, **4,** 1194–1211.
Biondi, F., A. Gershunov, and D.R. Cayan, 2001: North Pacific decadal climate variability since AD 1661. *J. Climate*, **14**, 5–10.
Bischoff, S.A., et al., 2000: Climate variability and Uruguay River flows. *Water International*, **25**, 446–456.
Bogdanova, E.G, B.M. Ilyin, and I.V. Dragomilova, 2002a: Application of a comprehensive bias correction model to precipitation measured at Russian North Pole drifting stations. *J. Hydrometeorol*., **3**, 700–713.
Bogdanova, E.G., et al., 2002b: A new model for bias correction of precipitation measurements, and its application to polar regions of Russia. *Russian Meteorol. and Hydrol.,* **10**, 68–94.
Böhm, R., 1998: Urban bias in temperature series – a case study for the city of Vienna. *Climatic Change*, **38**, 113–128.
Bojariu, R. and L. Gimeno, 2003: The influence of snow cover fluctuations on multiannual NAO persistence. *Geophys. Res. Lett.*, **30**, 1156, doi:10.1029/2002GL015651.
Bonan, G.B., 2001: Observational evidence for reduction of daily maximum temperature by croplands in the midwest United States. *J. Climate,* **14**, 2430–2442.
Bonnazola, M., and P.H. Haynes, 2004: A trajectory-based study of the tropical tropopause region. *J. Geophys. Res*., **109**, D020112, doi:10.1029/2003JD004356.
Bonsal, B. R., et al., 2001: Characteristics of daily and extreme temperatures over Canada. *J. Climate*, **14**, 1959–1976.
Bornstein, R., and Q. Lin, 2000: Urban heat islands and summertime convective thunderstorms in Atlanta: three cases studies. *Atmos. Environ*., **34**, 507–516.
Bottomley, M., et al., 1990: *Global Ocean Surface Temperature Atlas "GOSTA"*. Joint project of the U.K. Met Office and Massachusetts Institute of Technology supported by U.S. Dept. of Energy, U.S. National Science Foundation and U.S. Office of Naval Research, HMSO, London, 20 pp.+iv, 313 plates.
Boucher, O., 1999: Air traffic may increase cirrus cloudiness. *Nature*, **397**, 30–31.
Bove, M.C., et al., 1998: Effect of El Niño on U.S. landfalling hurricanes, revisited. *Bull. Amer. Meteor. Soc.*, **79**, 2477–2482.
Branstator, G., 2002: Circumglobal teleconnections, the jetstream waveguide, and the North Atlantic Oscillation. *J. Climate*, **15**, 1893–1910.
Brest, C.L., W.B. Rossow, and M. Roiter, 1997: Update of radiance calibrations for ISCCP. *J. Atmos. Ocean Tech.*, **14**, 1091–1109.
Brohan, P. et al., 2005: Uncertainty estimates in regional and global observed temperature changes , J. Geophys. Res. (submitted).
Bromirski, P.D., R.E. Flick, and D.R. Cayan, 2003: Storminess variability along the California coast: 1858–2000. *J. Climate*, **16**, 982–993.
Bromwich, D.H. and R.L. Fogt, 2004: Strong trends in the skill of the ERA-40 and NCEP–NCAR reanalyses in the high and middle latitudes of the Southern Hemisphere, 1958–2001. *J. Climate*, **17**, 4603–4619.
Brönnimann, S., et al., 2004: Extreme climate of the global troposphere and stratosphere in 1940–42 related to El Niño. *Nature*, **431,** doi:10.1038/nature02982.
Brunet, M., et al., 2005: Spatial and temporal temperature variability and change over Spain during 1850–2003. *J. Geophys. Res*. Submitted.
Brunetti, M., et al., 2004: Changes in daily precipitation frequency and distribution in Italy over the last 120 years. *J. Geophys. Res*., **109**, D05102, doi:10.1029/2003JD004296.
Brutsaert, W. and M.B. Parlange, 1998: Hydrologic cycle explains the evaporation paradox. *Nature*, **396,** 30.
Bryazgin, N.N. and A.A. Dement'ev, 1996: *Dangerous meteorological events in Russian Arctic* (in Russian), Gidrometeoizdat, St. Petersburg, 156 pp.
Burian, S.J. and J.M. Shepherd, 2005: Effects of urbanization on the diurnal rainfall pattern in Houston. *Hydrological Processes*: *Special Issue on Rainfall and Hydrological Processes*, **19**, 1089–1103.
Burnett, A.W., et al., 2003: Increasing Great Lake-effect snowfall during the Twentieth Century: A regional response to global warming? *J. Climate*, **16**, 3535–3541.

Cai, W. and P.G. Baines, 2001: Forcing of the Antarctic Circumpolar Wave by El Niño-Southern Oscillation teleconnections. *J. Geophys. Res.*, **106,** 9019–9038.
Cai, W., P.H. Whetton, and D.J. Karoly, 2003: The response of the Antarctic Oscillation to increasing and stabilized atmospheric CO_2. *J. Climate*, **16**, 1525–1538.
Caires, S. and A. Sterl, 2005: 100-year return value estimates for wind speed and significant wave height from the ERA-40 data. *J. Climate*, **18**, 1032–1048.
Camargo, S.J. and A.H. Sobel, 2004: *Western North Pacific tropical cyclone intensity and ENSO*. International Research Institute for Climate Prediction Technical Report No. 04-03, Palisades, NY, 25 pp.
Camilloni, I.A. and V.R. Barros, 2000: The Paraná River Response to El Niño 1982–83 and 1997–98 Events. *J. Hydrometeorol.*, **1**, 412–430.
Camilloni, I.A. and V.R. Barros, 2003: Extreme discharge events in the Paraná River and their climate forcing. *J. Hydrol.*, **278**, 94–106.
Camilloni, I.A. and R.M. Caffera, 2005: The largest floods in the Uruguay River and their climate forcing. *J. Hydrometeorol.* submitted.
Campbell, G., and T.H. vonder Haar, 2005: Global cloudiness: Nearly constant in time from ISCCP observations. *Geophys. Res. Lett*. submitted.
Camuffo, D. and Jones, P.D. (eds.), 2002: *Improved Understanding of Past Climatic Variability from Early Daily European Instrumental Sources*. Kluwer Academic Publishers, Dordrecht, 392 pp.
Cardone, V.J., J.G. Greenwood, and M.A. Cane, 1990: On trends in historical marine wind data. *J. Climate*, **3**, 113–127.
Carril, A.F., and A. Navarra, 2001: Low-frequency variability of the Antarctic Circumpolar Wave. *Geophys. Res. Lett*., **28**, 4623–4626.
Carter, D.J.T., 1999: Variability and trends in the wave climate of the North Atlantic: a review. In: *Proc. 9th ISOPE Conf.*, Vol. III, Brest, pp. 12–18.
Cassou, C. and L. Terray, 2001: Dual influence of Atlantic and Pacific SST anomalies on the North Atlantic/Europe winter climate. *Geophys. Res. Lett.*, **28**, 3195–3198.
Cassou, C., et al., 2004: North Atlantic winter climate regimes: spatial asymmetry, stationarity with time and oceanic forcing. *J. Climate*, **17**, 1055–1068.
Cayan, D.R., et al., 2001: Changes in the onset of spring in the western United States. *Bull. Amer. Meteor. Soc.*, **82**, 399–415.
CCSP, 2005: Synthesis product report on the vertical profiles of temperature trends. National Oceanic and Atmospheric Administration and National Research Council, U.S. Report. To be published 2005.
Cess, R.D. and P.M. Udelhofen, 2003: Climate change during 1985–1999: Cloud interactions determined from satellite measurements. *Geophys. Res. Lett*., **30**, 1019, doi:10.1029/2002GL016128.
Chagnon, F.J.F. and R. L. Bras, 2005: Contemporary climate change in the Amazon. *Geophys. Res. Lett*., **32**, L13703, doi:10.1029/2005GL022722.
Chan, J.C.L. and K.S. Liu, 2004: Global warming and Western North Pacific typhoon activity from an observational perspective. *J. Climate*, **17**, 4590–4602.
Chang, C.P., Y. Zhang, and T. Li, 1999: Interannual and interdecadal variations of the East Asian summer monsoon and tropical Pacific SSTs. Part I: Roles of the subtropical ridge. *J. Climate*, **13**, 4310–4325.
Chang, E.K.M., 2003: Midwinter suppression of the Pacific storm track activity as seen in aircraft observations. *J. Atmos. Sci*., **60**, 1345–1358.
Chang, E.K.M. and Y. Fu, 2002: Interdecadal variations in Northern Hemisphere winter storm track intensity. *J. Climate*, **15**, 642–658.
Chang, E.K.M. and Y. Fu, 2003: Using mean flow change as a proxy to infer interdecadal storm track variability. *J. Climate*, **16**, 2178–2196.
Changnon, D., M. Sandstrom, and C. Schaffer, 2003: Relating changes in agricultural practices to increasing dew points in extreme Chicago heat waves. *Clim. Res*. **24**, 243–254.
Changnon, S.A., et al., 1981: *METROMEX: A Review and Summary*. Amer. Meteor. Soc. Monogr., **18**, Boston, 81 pp.
Changnon, S.A. and N.E. Westcott, 2002: Heavy rainstorms in Chicago: Increasing frequency, altered impacts, and future implications. *J. Am. Water Res. Assoc*., **38**, 1467–1475.
Chao, Y., M. Ghil, and J.C. McWilliams, 2000: Pacific Interdecadal variability in this century's sea surface temperatures. *Geophys. Res. Lett.*, **27**, 2261–2264.
Chase, T.N., J.A. Knaff, and P.A. Pielke Sr., and E. Kalnay, 2003: Changes in global monsoon circulation since 1950. *Natural Hazards*, **29**, 229–254.

Chattopadhyay, N. and M. Hulme, 1997: Evaporation and potential evapotranspiration in India under conditions of recent and future climate change. *Agr. Forest Meteorol*., **87**, 55–73.
Chelliah, M., and Bell, G.D. 2004: Tropical multidecadal and interannual climate variability in the NCEP–NCAR Reanalysis. *J. Climate*: **17**, 1777–1803.
Chen, M., P. Xie, and J.E. Janowiak, 2002: Global land precipitation: a 50-yr monthly analysis based on gauge observations. *J. Hydrometeorol.*, **3**, 249–266.
Chen, T.-C., et al., 2004: Variation of the East Asian summer monsoon rainfall. *J. Climate*, **17**, 744–762.
Chenoweth, M., 2000: A new methodology for homogenisation of 19th century marine air temperature data. *J. Geophys. Res.*, **105**, 29145–29154.
Christy, J.R. and W.B. Norris, 2004: What may we conclude about tropospheric temperature trends? *Geophys. Res. Lett*., **31**, L0621, doi:10.1029/2003GL019361.
Christy, J.R., R.W. Spencer, and E.S. Lobl, 1998, Analysis of the merging procedure for the MSU daily temperature time series. *J. Clim*ate, **11**, 2016–2041.
Christy, J.R., R.W. Spencer, and W.D. Braswell, 2000: MSU tropospheric temperatures: Dataset construction and radiosonde comparisons. *J. Atmos. Oceanic Tech*. **17**, 1153–1170.
Christy, J.R., et al., 2001: Differential trends in tropical sea surface and atmospheric temperature since 1979. *Geophys. Res. Lett.*, **28**, 183–186.
Christy, J.R., et al., 2003: Error estimates of version 5.0 of MSU/AMSU bulk atmospheric temperatures. *J. Atmos. Oceanic Tech.*, **20,** 613–629.
Clark, C.O., P.J. Webster, and J.E. Cole, 2003: Interdecadal variability of the relationship between the Indian Ocean zonal mode and east African coastal rainfall anomalies. *J. Climate*, **16**, 548–554.
Cluis, D. and C. Laberge, 2001: Climate change and trend detection in selected rivers within the Asia-Pacific region. *Water International*, **26**, 411–424.
Cohen, S., B. Liepert, and G. Stanhill, 2004: Global dimming comes of age. *Eos, Trans. Amer. Geophys. Union*, **85**, 362.
Cole, J.E. and E.R. Cook, 1998: The changing relationship between ENSO variability and moisture balance in the continental United States. *Geophys. Res. Lett.*, **25**, 4529–4532.
Comiso, J.C., 2000: Variability and trends in Antarctic surface temperatures from in situ and satellite infrared measurements. *J. Climate*: **13**, 1674–1696.
Compo, G.P. and P.D. Sardeshmukh, 2004: Storm track predictability on seasonal and decadal scales. *J. Climate*, **17,** 3701–3720.
Connolley, W.M., 2003: Long-term variation of the Antarctic Circumpolar Wave. *J. Geophys. Res*., **108**, 8076, doi:10.1029/2000JC000380.
Considine, D.B., et al., 2001: An interactive model study of the influence of the Mount Pinatubo aerosol on stratospheric methane and water trends. *J. Geophys. Res*., **106**, 27711–27728.
Cook C., C.J.C. Reason, and B.C. Hewitson, 2004: Wet and dry spells within particularly wet and dry summers in the South African summer rainfall region. *Climate Res.*, **26**, 17–31.
Cook, E.R., et al., 1999: Drought reconstructions for the continental United States. *J. Climate*, **12,** 1145–1162.
Cook, E.R., R.D. D'Arrigo, and M.E. Mann, 2002: A well-verified, multiproxy reconstruction of the winter North Atlantic Oscillation index since A.D. 1400. *J. Climate*, **15**, 1754–1764.
Cook, E.R., et al., 2004: Long-term aridity changes in the western United States. *Science*, **306**, 1015–1018.
Corti, S., F. Molteni, and T.N. Palmer, 1999: Signature of recent climate change in frequencies of natural atmospheric circulation regimes. *Nature*, **398**, 799–802.
Cowell, C.M. and R.T. Stoudt, 2002: Dam-induced modifications to upper Allegheny River streamflow patterns and their biodiversity implications. *J. Amer. Water Res. Asso.* **38,** 187–196.
Crutzen P.J., 2004: New Directions: The growing urban heat and pollution "island" effect—impact on chemistry and climate. *Atmos. Environ.*, **38**, 3539–3540.
Cullen, H. and P.B. deMenocal, 2000: North Atlantic influence on Tigris-Euphrates streamflow, *Int. J. Climatol*., **20**, 853–863.
Curtis, S. and R.F. Adler, 2003: The evolution of El Niño-precipitation relationships from satellites and gauges. *J. Geophys. Res.–Atmospheres*, **108**, 4153, doi:10.1029/2002JD002690.
Czaja, A., A.W. Robertson, and T. Huck: 2003: The role of Atlantic ocean-atmosphere coupling in affecting North Atlantic Oscillation variability. In: *The North Atlantic Oscillation: Climatic Significance and Environmental Impact* [J.W. Hurrell, et al. (eds.)]. *Geophys. Monogr.*, **134**, Amer. Geophys. U., Washington, DC, pp. 147–172.
Dai, A., 2005: Recent climatology, variability and trends in global surface humidity. *J. Climate*. submitted.

Dai, A. and K.E. Trenberth, 2002: Estimates of freshwater discharge from continents: Latitudinal and seasonal variations. *J. Hydrometeorol*., **3**, 660–687.
Dai, A. and K.E. Trenberth, 2004: The diurnal cycle and its depiction in the Community Climate System Model. *J. Climate*, **17**, 930–995.
Dai, A., A.D. DelGenio, and I.Y. Fung, 1997a: Clouds, precipitation and temperature range. *Nature*, **386**, 665–666.
Dai, A., I.Y. Fung, and A.D. Del Genio, , 1997b: Surface observed global land precipitation during 1900–1988. *J. Climate*, **10**, 2943–2962.
Dai, A., K.E. Trenberth, and T.R. Karl, 1999: Effects of clouds, soil moisture, precipitation and water vapor on diurnal temperature range. *J. Climate*, **12**, 2451–2473.
Dai, A., et al., 2004a: Comment: The recent Sahel drought is real. *Int. J. Climatol*., **24**, 1323–1331.
Dai A., K.E. Trenberth, and T. Qian, 2004b: A global data set of Palmer Drought Severity Index for 1870–2002: Relationship with soil moisture and effects of surface warming. *J. Hydrometeorol.*, **5**, 1117–1130.
Dai, A., et al., 2005: Recent trends in cloudiness over the United States: A tale of monitoring inadequacies. *Bull. Amer. Meteor. Soc.* Accepted.
Davis, C.H., et al., 2005: Snowfall-driven growth in east Antarctic ice sheet mitigates recent sea-level rise*, Science**, *308**, 1898–1901, doi: 10.1126/science.1110662.
DeGaetano, A.T., and R.J. Allen, 2002: Trends in twentieth-century extremes across the United States. *J. Climate*, **15**, 3188–3205.
Delworth, T.L. ,and M.E. Mann, 2000: Observed and simulated multidecadal variability in the Northern Hemisphere. *Climate Dyn.*, **16**, 661–676.
DeMott, C.A., and D.A. Randall, 2004: Observed variations of tropical convective available potential energy. *J. Geophys. Res*. **109**, D02102, doi:10.129/2003JD003784.
Déry, S.J., and E.F. Wood, 2005: Decreasing river discharge in northern Canada, *Geophys. Res. Lett*., **32**, L10401, doi:10.1029/2005GL022845.
Deser, C., 2000: On the teleconnectivity of the Arctic Oscillation. *Geophys. Res. Lett.*, **27**, 779–782.
Deser, C., M.A. Alexander, and M.S. Timlin, 1996: Upper-ocean thermal variations in the North Pacific during 1970–1991. *J. Climate*, **9**, 1840–1855.
Deser, C., M.A. Alexander, and M.S. Timlin, 1999: Evidence for a wind-driven intensification of the Kuroshio Current Extension from the 1970s to the 1980s. *J. Climate*, **12**, 1697–1706.
Deser, C., J.E. Walsh, and M.S. Timlin, 2000: Arctic sea ice variability in the context of recent atmospheric circulation trends. *J. Climate*, **13**, 617–633.
Deser, C., M.A. Alexander, and M. S. Timlin, 2003: Understanding the persistence of sea surface temperature anomalies in midlatitudes. *J. Climate*, **16**, 57–72.
Deser, C., A.S. Phillips, and J.W. Hurrell, 2004: Pacific interdecadal climate variability: Linkages between the tropics and the north Pacific during boreal winter since 1900. *J. Climate*, **17**, 3109–3124.
Dessler, A.E. and S.C. Sherwood, 2004: Effect of convection on the summertime extratropical lower stratosphere, *J. Geophys. Res*., **109**, D23301, doi:10.1029/2004JD005209.
DeWeaver E. and S. Nigam, 2000: Zonal-eddy dynamics of the North Atlantic Oscillation. *J. Climate*, **13**, 3893–3914.
Diaz, H.F. and V. Markgraf, 2000: *El Niño and the Southern Oscillation: Multiscale Variability and Global and Regional Impacts.* Cambridge University Press, Cambridge, 496 pp.
Diaz, H.F., et al., 2002: Workshop on Advances in the Use of Historical Marine Climate Data, 29th Jan – 1st Feb 2002, Boulder, CO., U.S. *WMO Bulletin*, **51**, 377–380.
Dickson, R.R., et al., 2000: The Arctic Ocean response to the North Atlantic Oscillation. *J. Climate*, **13**, 2671–2696.
Diem, J.E. and D.P. Brown, 2003: Anthropogenic impacts on summer precipitation in central Arizona, U.S.A. *The Professional Geographer*, **55,** 343–355.
Diggle, P.J., K.Y. Liang, and S.L. Zeger, 1999: *Analysis of longitudinal data*. Clarendon Press, Oxford, 253 pp.
Ding, Y.H, C.Y. Li, and Y.J. Liu, 2004, Overview of the South China Seas monsoon experiment. *Adv. Atmos. Sci.*, **21**, 343–360.
Dirmeyer, P.A., A.J. Dolman, and N. Sato, 1999: The Global Soil Wetness Project: A pilot project for global land surface modeling and validation. *Bull. Amer. Meteor. Soc*., **80**, 851–878.
Dixon, P.G. and T.L. Mote, 2003: Patterns and causes of Atlanta's urban heat island-initiated precipitation. *J. Appl. Meteor.*, **42**, 1273–1284.

Donlon, C.J., et al., 2002: Toward improved validation of satellite sea surface temperature measurements for climate research. *J. Climate*, **15**, 353–369.
Dotzek, N., 2003: An updated estimate of tornado occurrence in Europe. *Atmos. Res*., **67–68**, 153–161.
Dow, C.L. and D.R. DeWalle, 2000: Trends in evaporation and Bowen ratio on urbanizing watersheds in eastern United States. *Water Resour. Res.*, **36**, 1835–1843.
Doyle M.E. and V.R. Barros, 2002: Midsummer low-level circulation and precipitation in subtropical South America and related sea surface temperature anomalies in the South Atlantic, *J. Climate*, **15**, 3394–3410.
Duan, K., T. Yao, and L. G. Thompson, 2004: Low-frequency of southern Asian monsoon variability using a 295-year record from the Dasuopu ice core in the central Himalayas. *Geophys. Res. Lett*., **31**, L16209, doi:10.1029/2004GL020015.
Durre, I. and J.M. Wallace, 2001: Factors influencing the cold-season diurnal temperature range in the United States. *J. Climate*, **14**, 3263–3278.
Durre, I., T. Peterson, and R. Vose, 2002: Evaluation of the effect of the Luers-Eskridge radiation adjustments on radiosonde temperature homogeneity. *J. Climate*, **15**, 1335–1347.
Durre, I., R.S. Vose, and D.B. Wuertz, 2005, Overview of the integrated global radiosonde archive. *J. Climate. Submitted*
Easterling, D.R., 2002: Recent changes in frost days and the frost-free season in the United States. *Bull. Amer. Meteor. Soc*., **83**, 1327–1332.
Easterling, D.R. and T.C. Peterson, 1995: A new method of detecting undocumented discontinuities in climatological time series. *Int. J. Climatol*., **15**, 369–377.
Easterling, D.R.. T.C. Peterson, and T.R. Karl, 1996: Notes and correspondence: On the development and use of homogenized climate datasets. *J. Climate*, **9**, 1429–1434.
Easterling, D.R., et al., 1997: Maximum and minimum temperature trends for the globe. *Science*, **277**, 364–367.
Easterling, D.R., et al., 2000: Observed variability and trends in extreme climate events: A brief review. *Bull. Amer. Meteorol. Soc*., **81**, 417–425.
Easterling, D.R, et al., 2003: CCl/CLIVAR workshop to develop priority climate indices. *Bull. Amer. Meteor. Soc.*, **84**, 1403–1407.
Easterling, D.R., et al., 2005: A comparison of local and regional trends in surface and lower tropospheric temperatures in western North Carolina. *Earth Interactions*, in press.
Elliott, W.P., R.J. Ross, and W.H. Blackmore, 2002: Recent changes in NWS upper-air observations with emphasis on changes from VIZ to Vaisala radiosondes. *Bull. Amer. Meteor. Soc*., **83**, 1003–1017.
Ellis, A.W. and J.J. Johnson, 2004: Hydroclimatic analysis of snowfall trends associated with the North American Great Lakes. *J. Hydrometeorol.*, **5**, 471–486.
Emanuel, K., 2003: Tropical cyclones. *Ann. Rev. Earth. Planet. Sci*., **31**, 75–104.
Emanuel, K., 2005: Increasing destructiveness of tropical cyclones over the past 30 years. *Nature*, **436**, 686–688.
Enfield, D.B., A.M. Mestas-Nuñez, and P.J Trimble, 2001: The Atlantic Multidecadal Oscillation and its relation to rainfall and river flows in the continental US. *Geophys. Res. Lett.*, **28,** 2077–2080.
Evans, S.J., et al., 1998: Trends in stratospheric humidity and the sensitivity of ozone to these trends, *J. Geophys. Res*., **103**, 8715-8726.
Evans, M.N., et al., 2001: Support for tropically-driven Pacific decadal variability based on paleoproxy evidence. *Geophys. Res. Lett*., **28,** 3689–3692.
Falarz, M., 2002: Long-term variability in reconstructed and observed snow cover over the last 100 winter seasons in Cracow and Zakopane (southern Poland). *Climate Res*., **29**, 247–256.
Fan, Y., et al., 2003: A 51-year reanalysis of the U.S. land-surface hydrology. *GEWEX*, **13**, 6–10.
Fauchereau, N., et al., 2003: Rainfall variability and changes in Southern Africa during the 20th century in the global warming context. *Natural Hazards*, **29,** 139–154.
Feldstein, S.B., 2002: The recent trend and variance increase of the Annular Mode. *J. Climate*, **15**, 88–94.
Fioletov, V.E., and T.G. Shepherd, 2003: Seasonal persistence of midlatitude total ozone anomalies. *Geophys. Res. Lett.*, **30**, 1417, doi:10.1029/2002GL016739.
Fogt, R.L. and D.H. Bromwich, 2005: Negative winter trends in the Southern Annular Mode. *Geophys. Res. Lett*.**,** submitted.
Folland, C.K., 1988: Numerical models of the raingauge exposure problem, field experiments and an improved collector design. *Quart. J. Roy. Meteorol. Soc.*, **114,** 1485–1516.

Folland, C.K., 2005: Tests of bias corrections to sea surface temperature using a climate model. *Int. J. Climatol.*, **25**, 895–911.
Folland, C.K. and D.E. Parker, 1995: Correction of instrumental biases in historical sea surface temperature data. *Quart. J. R. Meteor. Soc.*, **121**, 319–367.
Folland, C.K., et al., 1993: A study of six operational sea surface temperature analyses. *J. Climate*, **6**, 96–113**.**
Folland, C.K., et al., 1999: Large scale modes of ocean surface temperature since the late nineteenth century. In: *Beyond El Niño: Decadal and Interdecadal Climate Variability* [Navarra, A. (ed.)]. Springer-Verlag, Berlin, pp. 73–102.
Folland, C.K., et al., 2001: Global temperature change and its uncertainties since 1861. *Geophys. Res. Lett.*, **28**, 2621–2624.
Folland, C.K., et al., 2002: Relative influences of the interdecadal Pacific oscillation and ENSO on the South Pacific convergence zone. *Geophys. Res. Lett.*, **29**(13), doi:10.1029/2001GL014201.
Folland C.K., et al., 2003: Trends and variations in South Pacific island and ocean surface temperature. *J. Climate*, **16,** 2859–2874.
Førland, E.J., and I. Hanssen-Bauer, 2000: Increased precipitation in the Norwegian Arctic: True or false? *Climatic Change*, **46**, 485–509.
Forster, P.M.D., and K.P. Shine, 1999: Stratospheric water vapour changes as a possible contributor to observed stratospheric cooling. *Geophys. Res. Lett*, **26**, 3309–3312.
Forster, P.M.D., and K.P. Shine, 2002: Assessing the climate impact of trends in stratospheric water vapor. *Geophys. Res. Lett*, **29,** 1086, doi:10.1029/2001GL013909.
Forster, P.M.D., and S. Solomon, 2003: Observations of a “weekend effect” in diurnal temperature range. *Proc. Natl. Acad. Sci.*, **100**, 11225–11230.
Fowler, H.J. and C.G. Kilsby, 2003a: Implications of changes in seasonal and annual extreme rainfall. *Geophys. Res. Lett*., **30**, 1720, doi:10.1029/2003GL017327.
Fowler, H.J., and C.G. Kilsby, 2003b: A regional frequency analysis of United Kingdom extreme rainfall from 1961 to 2000. *Int. J. Climatol*., **23**, 1313–1334.
Frauenfeld, O.W., and R.E. Davis, 2003: Northern Hemisphere circumpolar vortex trends and climate change implications. *J. Geophys. Res*., **108,** 4423, doi:10.1029/2002JD002958.
Free, M., et al., 2002: Creating climate reference datasets: CARDS Workshop on Adjusting Radiosonde Temperature Data for Climate Monitoring. *Bull. Am. Met. Soc*., **83**, 891–899.
Free, M., M. Blister, and K. Emanuel, 2004a: Potential intensity of tropical cyclones: comparison of results from radiosonde and reanalysis data. *J. Climate*, **17**, 1722–1727.
Free, M., et al., 2004b: Using first differences to reduce inhomogeneity in radiosonde temperature datasets. *J. Climate*, **17**, 4171–4179.
Free M., et al., 2005: Radiosonde atmospheric temperature products for assessing climate (RATPAC): A new dataset of large-area anomaly time series. *J. Geophys. Res*., submitted.
Free, M., and D. Seidel, 2005: Causes of differing temperature trends in radiosonde upper-air datasets. *J. Geophys. Res.*, **110**, D07101, doi:10.1029/2004JD005481.
Frei, C., and C. Schär, 2001: Detection of probability of trends in rare events: Theory and application to heavy precipitation in the Alpine region. *J. Climate*, **14**, 1568–1584.
Frich, P., et al., 2002: Observed coherent changes in climatic extremes during the second half of the twentieth century. *Climate Res.*, **19**, 193–212.
Fu, Q., and C.M. Johanson, 2004: Stratospheric influence on MSU-derived tropospheric temperature trends: A direct error analysis. *J. Climate*, **17**, 4636–4640.
Fu, Q., et al., 2004a: Contribution of stratospheric cooling to satellite-inferred tropospheric temperature trends. *Nature*, **429,** 55–58.
Fu, Q., et al**.**, 2004b: Stratospheric cooling and the troposphere (reply). *Nature*, **432**, doi:10.1038/nature03210.
Fu, Q., and C.M. Johanson, 2005: Satellite-derived vertical dependence of tropical tropospheric temperature trends. *Geophys. Res. Lett.*, **32**, L10703, doi:10.1029/2004GL022266.
Fueglistaler, S., H. Wernli, and T. Peter, 2004: Tropical troposphere-to-stratosphere transport inferred from trajectory calculations. *J. Geophys. Res*., **109**, D03108, doi:10.1029/2003JD004069.
Fueglistaler, S. and P.H. Haynes, 2005: Control of interannual and longer-term variability of stratospheric water vapor, *J. Geophys. Res*. submitted.
Fujibe, F., 2003: Long-term surface wind changes in the Tokyo metropolitan area in the afternoon of sunny days in the warm season. *J. Meteor. Soc. Japan*, **81**,141–149.

Fusco, A.C. and M.L. Salby, 1999: Interannual variations of total ozone and their relationship to variations of planetary wave activity. *J. Climate*, **12**, 1619–1629.
Fye, F.K., D.W. Stahle, and E.R. Cook, 2003: Paleoclimatic analogs to Twentieth-Century moisture regimes across the United States. *Bull. Amer. Meteor. Soc*., **84**, 901–909.
Gaffen, D.J., 1996: *A digitized metadata set of global upper-air station histories*. NOAA Technical Memorandum ERL ARL-211.
Gallo, K.P., et al., 1999: Temperature trends of the U.S. Historical Climatology Network based on satellite-designated land use/land cover. *J. Climate*, **12**, 1344–1348.
García, N.O. and W.M. Vargas, 1998: The temporal climatic variability in the 'Río de la Plata' basin displayed by the river discharges. *Climatic Change*, **38**, 359–379.
García-Herrera, R., et al., 2005: CLIWOC: A climatological database for the world's oceans 1750–1854. *Climatic Change*, in press.
GCOS, 2004: *GCOS Implementation Plan for Global Observing System for Climate in support of UNFCC*. GCOS-92, WMO/TD 1219, 136 pp.
Gedalof, Z., N.J. Mantua, and D.L. Peterson, 2002: A multi-century perspective of variability in the Pacific Decadal Oscillation: new insights from tree rings and coral. *Geophys. Res. Lett*., **29,** 2204, doi:10.1029/2002GL015824.
Geng, Q. and M. Sugi, 2001: Variability of the North Atlantic cyclone activity in winter analyzed from NCEP-NCAR reanalysis data. *J. Climate*, **14**, 3863–3873.
Genta, J.L., G. Perez-Iribarren, and C.R. Mechoso, 1998: A recent increasing trend in the streamflow of rivers in southeastern South America. *J. Climate*, **11,** 2858–2862.
Genthon, C., G. Krinner, and M. Sacchettini, 2003: Interannual Antarctic tropospheric circulation and precipitation variability. *Climate Dyn.*, **21**, 289–307.
Gershunov, A. and T.P. Barnett, 1998: Interdecadal modulation of ENSO teleconnections. *Bull. Amer. Meteor. Soc*., **79**, 2715–2725.
Gettelman, A., et al., 2002: Multi-decadal trends in tropical convective available potential energy. *J. Geophys. Res*., **107**, 4606, doi:10.1029/2001JD001082.
Giannini, A., R. Saravannan, and P. Chang, 2003: Ocean forcing of Sahel rainfall on interannual to interdecadal time scales. *Science*, **302**, 1027–1030.
Gilgen, H., M. Wild, and A. Ohmura, 1998: Means and trends of shortwave irradiance at the surface estimated from global energy balance archive data. *J. Climate*, **11**, 2042–2061.
Gille, S.T., 2002: Warming of the Southern Ocean since the 1950s. *Science*, **295**, 1275–1277.
Gillett, N. and D. Thompson, 2003: Simulation of recent Southern Hemisphere climate change. *Science*, **302**, 273–275.
Gillett, N.P., et al. 2003: Detection of human influence on sea-level pressure. *Nature*, **422**, 292–294.
Gillett, N.P., B.D. Santer, and A.J. Weaver, 2004: Stratospheric cooling and the troposphere. *Nature*, **432**, doi:10.1038/nature03209.
Goddard L. and N.E. Graham, 1999: Importance of the Indian Ocean for simulating rainfall anomalies over eastern and southern Africa. *J. Geophys. Res.*, **104**, 19099–19116.
Goldenberg, S.B, et al., 2001: The recent increase in Atlantic hurricane activity: causes and implications. *Science*, **293**, 474–479.
Golubev, V.S., and E.G. Bogdanova, 1996: Accounting of blowing snow events in precipitation measurement in Russia. In: *Proc. of the ACSYS Solid Precipitation Climatology Project Workshop*, Reston, VA, U.S., 12–15 September 1995, World Meteorol. Org., WCRP No. 93, WMO/TD No. 739 .
Golubev, V.S., V.V. Koknayeva, and A.Yu. Simonenko, 1995: Results of the atmospheric precipitation measurements with national standard instruments of Canada, Russia, and USA. *Meteozologiya i Gidzologiya [Meteorology and Hydrology*, **2**, 102–110.
Golubev, V.S., et al., 1999: Precipitation measurement correction and quality of corrected data from Valdai hydrological station. *Russian Meteorol. and Hydrol.*, **1**, 65–74.
Golubev, V.S., et al., 2001: Evaporation changes over the contiguous United States and the former USSR: A reassessment. *Geophys. Res. Lett.*, **28,** 2665–2668.
Gong, D. and S. Wang, 1999: Definition of Antarctic oscillation index. *Geophys. Res. Lett.*, **26**, 459–462.
Gong D.Y. and S.W. Wang, 2000: Severe summer rainfall in China associated with enhanced global warming. *Climate Res.*, **16**, 51–59.
Gong D.Y. and C.-H. Ho, 2002: Shift in the summer rainfall over the Yangtze River valley in the late 1970s. *Geophys. Res. Lett.*, **29**(3), doi:10.1029/2001GL014523.

Gong D.Y. and X.Z He., 2002: Interdecadal change in western Pacific subtropical high and climate effects. *Acta Geographica Sinica*, **2**, 185–193.

Gong, D.Y., S.W. Wang, and J.H. Zhu, 2001: East Asian winter monsoon and Arctic Oscillation. *Geophys. Res. Lett*., **28**, 2073–2076.

Gong, G., D. Entekhabi, and J. Cohen, 2002: A large-ensemble model study of the wintertime AO-NAO and the role of interannual snow perturbations. *J. Climate*, **15**, 3488–3499.

Gong, G., D. Entekhabi, and J. Cohen, 2003: Modeled Northern Hemisphere winter climate response to realistic Siberian snow anomalies. *J. Climate*, **16**, 3917–3931.

Goodison, B.E., P.Y.T. Louie, and D. Yang, 1998: *WMO solid precipitation intercomparison, Final Report*. World Meteorol. Org., Instruments and Observing Methods Rep. 67, WMO/TD 872, 87 pp. + Annexes.

Goodwin, I.D., et al., 2004: Mid latitude winter climate variability in the south Indian and southwest Pacific regions since 1300 AD. *Climate Dyn.*, **22**, 783–794.

Gower, J.F.R., 2002: Temperature, wind and wave climatologies, and trends from marine meteorological buoys in the northeast Pacific. *J. Climate*, **15**, 3709–3718.

Graham, N.E. and H.F. Diaz, 2001: Evidence for intensification of North Pacific winter cyclones since 1948. *Bull. Amer. Meteor. Soc*., **82**, 1869–1893.

Gray, W.M., 1984: Atlantic seasonal hurricane frequency: Part I: El Niño and 30-mb quasi-biennial oscillation influences. *Mon. Wea. Rev.*, **112**, 1649–1668.

Gray, S.T., et al., 2004: A tree-ring based reconstruction of the Atlantic Multidecadal Oscillation since 1567 A.D. *Geophys. Res. Lett*., **31**, L12205, doi:10.1029/2004GL019932.

Griffiths, G.M., M.J. Salinger, and I. Leleu, 2003: Trends in extreme daily rainfall in the South Pacific and relations to the South Pacific Convergence Zone. *Int. J. Climatol*., **23**, 847–869.

Griffiths, G.M., et al., 2005: Change in mean temperature as a predictor of extreme temperature change in the Asia-Pacific region. *Int. J. Climatol*, **25**, 1301-1330.

Grist, J.P. and S.A. Josey, 2003: Inverse analysis of the SOC air-sea flux climatology using ocean heat transport constraints. *J. Climate*, **16**, 3274–3295.

Grody, N.C., et al., 2004: Calibration of multi-satellite observations for climatic studies: Microwave sounding unit (MSU). *J. Geophys. Res*., **109**, D24104, doi:10.1029/2004JD005079.

Groisman, P.Ya. and E.Ya. Rankova, 2001: Precipitation trends over the Russian permafrost-free zone: removing the artifacts of pre-processing. *Int. J. Climatol.*, **21**, 657–678.

Groisman, P.Ya., R.W. Knight, and T.R. Karl, 2001: Heavy precipitation and high streamflow in the contiguous United States: Trends in the 20th century. *Bull. Amer. Meteor. Soc*., **82**, 219–246.

Groisman, P.Ya., et al., 2003. Contemporary climate changes in high latitudes of the Northern Hemisphere: Daily time resolution. In: *Proc. Intl Symp. Climate Change*, Beijing, China, 31 March–3 April, 2003, World Meteorol. Org. Publ. 1172, pp. 51–55.

Groisman, P.Ya., et al., 2004: Contemporary changes of the hydrological cycle over the contiguous United States: Trends derived from *in situ* observations. *J. Hydrometeorol.*, **5**, 64–85.

Groisman, P.Ya., et al. 2005a: Trends in intense precipitation in the climate record. *J. Climate*, 18, 1326–1350.

Groisman, P.Ya, et al., 2005b: Potential forest fire danger over northern Eurasia: Changes during the 20th century. *Earth and Planetary Change*, submitted.

Grollmann, T. and S. Simon, 2002: Flutkatastrophen – Boten des Klimawandels. *Z. Versicher*., **53**, 682–689.

Gruza, G.V., et al., 1999: Indicators of climatic change for the Russian Federation. *Climatic Change*, **42**, 219–242.

Gu, D.F. and S.G.H. Philander, 1997: Interdecadal climate fluctuations that depend on exchanges between the tropics and extratropics. *Science*, **275**, 805–807.

Guetter, A.K. and K.P. Georgakakos, 1993: River outflow of the conterminous United States, 1939–1988. *Bull. Amer. Meteor. Soc.*, **74**, 1873–1891.

Guichard, F., D. Parsons, and E. Miller, 2000: Thermodynamic and radiative impact of the correction of sounding humidity bias in the tropics. *J. Climate*, **13**, 3611–3624.

Gulev, S.K. and L. Hasse, 1999: Changes of wind waves in the North Atlantic over the last 30 years. *Int. J. Climatol*., **19**, 1091–1117.

Gulev, S.K. and V. Grigorieva, 2004: Last century changes in ocean wind wave height from global visual wave data. *Geophys. Res. Lett*., 31, L24302, doi:10.1029/2004GL021040.

Gulev, S. K., O. Zolina, and S. Grigoriev, 2001: Extratropical cyclone variability in the Northern Hemisphere winter from the NCEP/NCAR reanalysis data. *Climate Dyn.*, **17**, 795–809.

Gulev, S.K., T. Jung, and E. Ruprecht, 2005: Estimation of the impact of sampling errors in the VOS observations on air-sea fluxes. Part II. Impact on trends and interannual variability. *J. Climate*, submitted.
Guo, Q.Y., et al., 2003: Interdecadal variability of East-Asian summer monsoon and its impact on the climate of China. *Acta Geographica Sinica*, **4**, 569–576.
Gustafsson, M.E.R., 1997: Raised levels of marine aerosol deposition owing to increased storm frequency: A cause of forest decline in southern Sweden? *Agric. and Forest Meteor.*, **84**, 169–177.
Hahn, C.J., and S.G. Warren, 2003: *Cloud Climatology for Land Stations Worldwide, 1971–1996*. Report NDP-026D, 35 pp. Carbon Dioxide Information Analysis Center, Oak Ridge, Tennessee, U.S.. http://cdiac.ornl.gov/ftp/ndp026d/.
Haimberger, L. 2005: *Homogenization of radiosonde temperature time series using ERA-40 analysis feedback information*. ERA-40 Project Report Series 23, ECMWF, Reading, U.K., 68 pp.
Hansen, J., et al., 2001: A closer look at United States and global surface temperature change. *J. Geophys. Res.*, **106**, 23947–23963.
Harnik N. and E.K.M. Chang, 2003: Storm track variations as seen in radiosonde observations and reanalysis data. *J. Climate*, **16**, 480–495.
Harris, B.A. and G.A Kelly, 2001: A satellite radiance bias correction scheme for data assimilation. *Quart. J. Roy. Meteor. Soc*., **127**, 1453–1468.
Hartmann, D.L. and F. Lo, 1998: Wave-driven flow vacillation in the Southern Hemisphere. *J. Atmos. Sci.*, **55**, 1303–1315.
Hartmann, D.L., et al., 2000: Can ozone depletion and global warming interact to produce rapid climate change? In: *Proc. National Academy of Sciences of the USA*, **97**, 1412–1417.
Haylock, M.R. and C.M. Goodess, 2004: Interannual variability of extreme European winter rainfall and links with mean large-scale circulation. *Int. J. Climatol*., **24**, 759–776.
Haylock M.R., et al., 2005. Trends in total and extreme South American rainfall 1960–2000. *J. Climate*, In revision.
He, H., et al., 2003: Some climatic features of the tropical cyclones landed onto Guangdong Province during the recent 50 years. *Scientia Meteorologica Sinica*, **23**, 401–409. (In Chinese with English Abstract).
Held, I.M. and B.J. Soden, 2000: Water vapor feedback and global warming. *Ann. Rev. Energy and the Environment*, **25**, 441–475.
Henderson-Sellers, A., 1992: Continental cloudiness changes this century. *Geo Journal*, **27**, 255–262.
Hennessy, K.J., R. Suppiah, and C.M. Page, 1999: Australian rainfall changes, 1910–1995. *Austr. Meteor. Mag.*, **48**, 1–13.
Herath, S. and U. Ratnayake, 2004: Monitoring rainfall trends to predict adverse impacts – a case study from Sri Lanka (1964–1993). *Glob. Env. Change*, **14**, 71–79.
Higgins, R.W. and W. Shi, 2000: Dominant factors responsible for interannual variability of the summer monsoon in the southwestern United States. *J. Climate*, **13,** 759–776.
Highwood, E.J., B.J. Hoskins and P. Berrisford, 2000: Properties of the Arctic tropopause. *Quart. J. Roy. Meteor. Soc*., **126**, 1515–1532.
Hines, K.M., D.H. Bromwich, and G.J. Marshall, 2000: Artificial surface pressure trends in the NCEP-NCAR reanalysis over the Southern Ocean and Antarctica. *J. Climate*, **13**, 3940–3952.
Hirabayashi, Y., S. Kanae, and T. Oki, 2005: A 100-year (1901–2000) global retrospective estimation of terrestrial water cycle, *J. Geophys. Res.*, in press.
Hobbins, M.T., J.A. Ramirez, and T.C. Brown, 2004: Trends in pan evaporation and actual evaportranspiration across the conterminous U.S.: Paradoxical or complementary? *Geophys. Res. Lett*., **31**, L13503, doi:10/10029/2004GL019846.
Ho, C.-H., et al., 2003: A sudden change summer rainfall characteristics in Korea during the late 1970s. *Int. J. Climatol* , **23**, 117–128.
Hodgkins, G.A., R.W. Dudley, and T.G. Huntington, 2003: Changes in the timing of high river flows in New England over the 20th century. *J. Hydrol.*, **278,** 244–252.
Hoerling, M. and A. Kumar, 2003: The perfect ocean for drought. *Science*, **299**, 691–694.
Horel, J.D. and J.M. Wallace, 1981: Planetary-scale atmospheric phenomena associated with the Southern Oscillation. *Mon. Wea. Rev.*, **109,** 813–829.
Hoskins, B.J. and D.J. Karoly, 1981: Steady linear response of a spherical atmosphere to thermal and orographic forcing. *J. Atmos. Sci.*, **38,** 1179–1196.
Hu, Q. and S. Feng, 2001: A southward migration of centennial-scale variations of drought/flood in eastern China and the western United States. *J. Climate*, **14**, 1323–1328.

Hu, Q., Y. Tawaye, and S. Feng, 2004: Variations of the Northern Hemisphere atmospheric energetics: 1948–2000. *J. Climate*, **17**, 1975–1986.
Huang, R.H., L. Zhou, and W. Chen, 2003: The progresses of recent studies on the variabilities of the East Asian monsoon and their causes. *Advances Atmos. Sci.,* **1**, 55–69.
Huffman, G., et al., 1997: The Global Precipitation Climatology Project (GPCP): combined precipitation dataset. *Bull Amer. Meteor. Soc*., **78**, 5–20.
Hughes, C.W., et al., 2003: Coherence of Antarctic sea levels, Southern Hemisphere Annular Mode, and flow through the Drake Passage. *Geophys. Res. Lett.*, **30**, 1464, doi:10.1029/2003GL017240.
Huntington, T.G., et al., 2004: Changes in the proportion of precipitation occurring as snow in New England (1949–2000). *J. Climate*, **17**, 2626–2636.
Hurrell, J.W., 1995: Decadal trends in the North Atlantic Oscillation and relationships to regional temperature and precipitation. *Science,* **269**, 676–679.
Hurrell, J.W., 1996: Influence of variations in extratropical wintertime teleconnections on Northern Hemisphere temperature. *Geophys. Res. Lett.*, **23,** 665–668.
Hurrell, J.W. and H. van Loon, 1994: A modulation of the atmospheric annual cycle in the Southern Hemisphere. *Tellus*, **46A**, 325–338.
Hurrell, J.W. and H. van Loon, 1997: Decadal variations associated with the North Atlantic Oscillation. *Climatic Change,* **36,** 301–326.
Hurrell, J.W. and K.E. Trenberth, 1999: Global sea surface temperature analyses: multiple problems and their implications for climate analysis, modeling and reanalysis, *Bull. Amer. Meteor. Soc.*, **80**, 2661–2678.
Hurrell, J.W., et al., 2000: Comparison of tropospheric temperatures from radiosondes and satellites: 1979–98. *Bull. Amer. Meteor. Soc*., **81**, 2165–2177.
Hurrell, J.W., M.P. Hoerling, and C.K. Folland, 2001: Climatic variability over the North Atlantic. In: *Meteorology at the Millennium* [Pearce, R. (ed.)]. Academic Press, London, pp. 143–151.
Hurrell, J.W., et al., 2002: The relationship between tropical Atlantic rainfall and the summer circulation over the North Atlantic. In: Proc. U. S. CLIVAR Atlantic meeting, [Legler, D. (ed.)]. Boulder, CO., pp. 108–110.
Hurrell, J.W., et al., 2003: An overview of the North Atlantic Oscillation. In: *The North Atlantic Oscillation: Climatic Significance and Environmental Impact* [Hurrell, J.W., et al. (eds.)]. *Geophys. Monogr*., **134**, Amer. Geophys. U., Washington, DC, pp. 1–35.
Hurrell, J.W., et al., 2004: Twentieth century North Atlantic climate change. Part I: Assessing determinism. *Climate Dyn.*, **23**, 371–389.
Indeje, M., H.F.M. Semazzi, and L.J.Ogallo, 2000. ENSO signals in East African rainfall seasons. *Int. J. Climatol.*, **20**, 19–46.
Intergovernmental Panel on Climate Change (IPCC), 1999: *Aviation and the global atmosphere*, [J. E. Penner, et al. (eds.)], Cambridge University Press, Cambridge, United Kingdom, 384pp.
IPCC: 2001, *Climate Change 2001: The Scientific Basis. Contribution of Working Group 1 to the Third IPCC Scientific Assessment.* [Houghton, J.T., et al. (eds.)]. Cambridge University Press , Cambridge, United Kingdom, 881 pp.
Inoue, T. and F. Kimura, 2004: Urban effects on low-level clouds around the Tokyo metropolitan area on clear summer days. *Geophys. Res. Lett.*, **31**, L05103, doi:10.1029/2003GL018908.
Ishii, M., et al., 2005: Objective analysis of SST and marine meteorological variables for the 20th Century using ICOADS and the Kobe Collection. *Int. J. Climatol.*, **25**, 865–879.
Iskenderian, H. and R. Rosen, 2000: Low-frequency signals in mid-tropospheric submonthly temperature variance. *J. Climate*, **13**, 2323–2333.
Jacobowitz, H., et al., 2003: The Advanced Very High Resolution Radiometer Pathfinder Atmosphere (PATMOS) climate dataset: A resource for climate research. *Bull. of the Amer. Meteor. Soc*. **84**, 785–793.
Jacobs, G.A. and J.L. Mitchell, 1996: Ocean circulation variations associated with the Antarctic Circumpolar Wave. *Geophys. Res. Lett.*, **23**, 2947–2950.
Janicot, S., Trzaska, S., and Poccard, I., 2001: Summer Sahel-ENSO teleconnection and decadal time scale SST variations. *Climate Dyn.* **18**, 303–320.
Jauregui, E. and E. Romales, 1996: Urban effects on convective precipitation in Mexico City. *Atm. Env*., **30**, 3383–3389.
Jin, M. and R.E. Dickinson, 2002: New observational evidence for global warming from satellite. *Geophys. Res. Lett.*, **29**(10)**,** doi:10.1029/2001GL013833.

Jones, D., et al., 2004. A new tool for tracking Australia's climate variability and change. *Bull. Australian Meteor. and Oceanographic Soc.*, **17**, 65–69.

Jones, J.M. and M. Widmann, 2003: Instrument- and tree-ring based estimates of the Antarctic Oscillation. *J. Climate*, **16**, 3511–3524.

Jones, J.M. and M. Widmann, 2004: Variability of the Antarctic Oscillation during the 20th century. *Nature*, **432**, 290–291.

Jones, P.D. and A. Moberg, 2003: Hemispheric and large-scale surface air temperature variations: An extensive revision and update to 2001. *J. Climate*, **16**, 206–223.

Jones, P.D., et al., 1990: Assessment of urbanization effects in time series of surface air temperature over land. *Nature*, **347,** 169–172.

Jones, P.D., T. Jónsson, and D. Wheeler, 1997: Extension to the North Atlantic Oscillation using early instrumental pressure observations from Gibraltar and south-west Iceland, *Int. J. Climatol.*, **17**, 1433–1450.

Jones, P.D., et al., 1999: Surface air temperature and its changes over the past 150 years. *Rev. Geophys.*, **37**, 173–199.

Jones, P.D., T.J. Osborn, and K.R. Briffa, 2003: Pressure-based measures of the North Atlantic Oscillation (NAO): A comparison and an assessment of changes in the strength of the NAO and in its influence on surface climate parameters. In: *The North Atlantic Oscillation: Climatic Significance and Environmental Impact* [Hurrell, J.W., et al. (eds.)]. *Geophysical Monograph*, **134**, American Geophysical Union, Washington, DC, pp. 1–35.

Jones, P.D., et al., 2001: Adjusting for sampling density in grid-box land and ocean surface temperature time series. *J. Geophys. Res.*, **106**, 3371–3380.

Josey, S.A. and R. Marsh, 2005: Surface freshwater flux variability and recent freshening of the North Atlantic in the eastern subpolar gyre. *J. Geophys. Res.*, 110, C05008, doi:10.1029/2004JC002521.

Joshi, M. M. and K.P. Shine, 2003: A GCM study of volcanic eruptions as a cause of increased stratospheric water vapor, *J. Climate*, **16**, 3525–3534.

Jung, H.-S., et al., 2002: Recent trends in temperature and precipitation over South Korea. *Int. J. Climatol.,* **22**, 1327–1337.

Jury, M.R., 2003: The coherent variability of African river flows: composite climate structure and the Atlantic Circulation. *Water SA*, **29**, 1, 1–10.

Kaiser, D.P., 1998: Analysis of total cloud amount over China, 1951–1994. *Geophys. Res. Lett*, **25**, 3599–3602.

Kaiser, D.P. and Y. Qian, 2002: Decreasing trends in sunshine duration over China for 1954–1998: Indication of increased haze pollution? *Geophys. Res. Lett*., **29**, 2042, doi:10.1029/2002GL016057.

Kalnay, E., et al., 1996: The NCEP/NCAR Reanalysis Project. *Bull. Amer. Meteor. Soc*., **77**, 437–471.

Kalnay, E. and M. Cai, 2003: Impact of urbanization and land-use change on climate. *Nature*, **423,** 528–531.

Kanamitsu, M., et al., 2002: An overview of NCEP-DOE AMIP-II Reanalysis (R-2). *Bull. Amer. Meteor. Soc*., **83**, 1631–1643.

Kaplan, A., et al., 1997: Reduced space optimal analysis for historical data sets: 136 years of Atlantic sea surface temperatures. *J. Geophys. Res.*, **102**, 27835–27860.

Karl, T.R., H.F. Diaz, and G. Kukla, 1988: Urbanization: its detection and effect in the United States climate record. *J. Climate*, **1**, 1099–1123.

Karl, T.R., et al., 1986: A model to estimate the time of observation bias associated with monthly mean maximum, minimum, and mean temperature for the United States. *J. Climate Appl. Meteor*., **25**, 145–160.

Karoly, D.J., 2003: Ozone and climate change. *Science*, **302**, 236–237.

Karoly, D., J. Risbey, and A. Reynolds, 2003. *Global warming contributes to Australia's worst drought*, WWF Australia, 8 pp., available at www.wwf.org.au.

Karoly, D.J. and Q. Wu, 2005: Detection of regional surface temperature trends. *J. Climate*, Revised submission.

Kaufman, Y.J., D. Tanré, and O. Boucher, 2002: A satellite view of aerosols in the climate system. *Nature*, **419**, 215–223.

Keable, M., I. Simmonds, and K. Keay, 2002: Distribution and temporal variability of 500 hPa cyclone characteristics in the Southern Hemisphere. *Int. J. Climatol.*, **22,** 131–150.

Kent, E.C. and P.G. Challenor, 2005: Towards estimating climatic trends in SST data, part 2: Random errors. *J. Atmos. Oceanic Technol.*, in press.

Kent, E.C. and A. Kaplan, 2005: Towards estimating climatic trends in SST data, part 3: Systematic biases. *J. Atmos. Oceanic Technol.*, in press.

Kernthaler, S.C., R. Toumi, and J.D. Haigh, 1999: Some doubts concerning a link between cosmic ray fluxes and global cloudiness. *Geophys. Res. Lett.*, **26**, 863–865.

Kerr, R., 2000: A North Atlantic climate pacemaker for the centuries. *Science*, **288**, doi:10.1126/science.288.5473.1984.

Kharin, V.V. and F.W. Zwiers, 2000: Changes in extremes in an ensemble of transient climate simulations with a coupled atmosphere-ocean GCM. *J. Climate*, **13**, 3760–3780.

Kidson, J.W., 1999: Principal modes of Southern Hemisphere low frequency variability obtained from NCEP-NCAR reanalyses. *J. Climate*, **12**, 2808–2830.

Kiehl, J.T. and K.E. Trenberth, 1997: Earth's annual global mean energy budget. *Bull. Amer. Meteor. Soc.*, **78**, 197–208.

Kiktev, D., et al., 2003: Comparison of modeled and observed trends in indices of daily climate extremes. *J. Climate*, **16**, 3560–3571.

Kiladis, G.N. and K.C. Mo, 1998: Interannual and intraseasonal variability in the Southern Hemisphere. In *Meteorology of the Southern Hemisphere* [Karoly, D.J., and Vincent, D.G. (eds.)]. *Meteorological Monographs*, **49**, Amer. Meteorol. Soc., Boston, MA, pp. 1–46.

Kilpatrick, K.A., G.P. Podesta, and R. Evans, 2001: Overview of the NOAA/NASA advanced very high resolution radiometer Pathfinder algorithm for sea surface temperature and associated matchup database. *J. Geophys. Res.*, **106**, 9179–9198.

Kinter III, J.L., et al., 2004: An evaluation of the apparent interdecadal shift in the tropical divergent circulation in the NCEP–NCAR Reanalysis *J. Climate* , **17**, 349–361.

Kinter III, J.L., K. Miyakoda, and S. Yang, 2002: Recent change in the connection from the Asia monsoon to ENSO. *J. Climate,* **15**, 1203–1215.

Kistler, R., et al., 2001: The NCEP-NCAR 50-year reanalysis: Month means CD-ROM and documentation. *Bull. Amer. Meteor. Soc*., **82**, 247–268.

Klein Tank, A.M.G. and G.P. Können, 2003: Trends in indices of daily temperature and precipitation extremes in Europe, 1946–1999. *J. Climate*, **16**, 3665–3680.

Klein Tank, A.M.G., et al., 2005: Changes in daily temperature and precipitation extremes in Central and South Asia. *J. Geophys. Res*., submitted.

Kley, D., J.M. Russell, and C. Phillips, 2000: *SPARC Assessment of upper tropospheric and stratospheric water vapour*. WCRP, Geneva, WCRP No. 113, WMO/TD No. 1043, 325 pp.

Kodera, K., and H. Koide, 1997: Spatial and seasonal characteristics of recent decadal trends in the northern hemispheric troposphere and stratosphere, *J. Geophys. Res*., **102**, 19433–19447.

Kodera, K., Y. Kuroda, and S. Pawson, 2000: Stratospheric sudden warmings and slowly propagating zonal-mean zonal wind anomalies. *J. Geophys. Res*., **105**, 12351–12359.

Können, G. P., et al., 1998: Pre-1866 extensions of the Southern Oscillation index using early Indonesian and Tahitian meteorological readings. *J. Climate*, **11,** 2325–2339.

Kostopoulou, E. and P.D. Jones, 2005: Assessment of climate extremes in the eastern Mediterranean, *Meteor. and Atmospheric Physics*, **89**, 69–85.

Kothawale, D.R. and K. Rupa Kumar, 2005: On the recent changes in surface temperature trends over India. *Geophys. Res. Lett.* submitted.

Krepper, C.M., N.O. García, and P.D. Jones, 2003: Interannual variability in the Uruguay River basin. *Int. J. Climatol.*, **23**, 103–115.

Kripalani, R.H. and A. Kulkarni, 1997a: Climatic impact of El Niño / La Niña on the Indian monsoon: A new perspective. *Weather*, **52**, 39–46.

Kripalani, R.H., and A. Kulkarni, 1997b: Rainfall variability over Southeast Asia: Connections with Indian monsoon and ENSO extremes: New perspectives. *Int. J. Climatol.*, **17**, 1155–1168.

Kripalani, R.H., and A. Kulkarni, 2001: Monsoon rainfall variations and teleconnections over South and East Asia. *Int. J. Climatol.,* **21**, 603–616.

Kripalani, R.H., A. Kulkarni, and S.S. Sabade, 2001: El Niño Southern Oscillation, Eurasian snow cover and the Indian monsoon rainfall. *Proc. Indian Nat. Sci. Academy*, **67A**, 361–368.

Kripalani, R.H., et al., 2003: Indian monsoon variability in a global warming scenario. *Natural Hazards*, **29**, 189–206.

Krishnamurthy, V. and B.N. Goswami, 2000: Indian monsoon-ENSO relationship on interdecadal timescale, *J. Clim*., **13**, 579–595.

Kristjánsson, J.E. and J. Kristiansen, 2000: Is there a cosmic ray signal in recent variations in global cloudiness and cloud radiative forcing? *J. Geophys. Res*., **105**, 11851–11863.
Kristjánsson, J.E., et al., 2002: A new look at possible connections between solar activity, clouds and climate. *Geophys. Res. Lett*., **29**, 2107, doi:10.1029/2002GL015646.
Krüger, K., B. Naujokat, and K. Labitzke, 2005: The unusual midwinter warming in the southern hemisphere stratosphere in 2002: A comparison to northern hemisphere phenomena. *J. Atmos. Sci*., **62**, 602–613.
Kuang, Z., Y. Jiang, and Y.L. Yung, 1998: Cloud optical thickness variations during 1983–1991: solar cycle or ENSO. *Geophys. Res. Lett*., **25**, 1415–1417.
Kuleshov, Y. and G. de Hoedt, 2003: Tropical cyclone activity in the Southern Hemisphere. *Bull. Australian Meteor. Oceanog. Soc*., **16**, 135–137.
Kumar, K.K., B. Rajagopalan, and A.M. Cane, 1999: On the weakening relationship between the Indian monsoon and ENSO, *Science*, **284**, 2156–2159.
Kunkel, K.E., et al., 2003: Temporal variations of extreme precipitation events in the United States: 1895–2000, *Geophys. Res. Lett*., **30**, 1900, doi:10.1029/2003GL018052.
Kunkel, K.E., et al., 2004: Temporal variations in frost-free season in the United States: 1895–2000. *Geophys. Res. Lett.*, **31**, L03201, doi:10.1029/2003GL018624.
Kuroda, Y. and K. Kodera, 2001: Variability of the polar-night jet in the northern and southern hemispheres. *J. Geophys. Res*., **106**, 20,703–20,713.
Kwok, R. and J.C. Comiso, 2002a: Southern ocean climate and sea ice anomalies associated with the Southern Oscillation. *J. Climate*, **15,** 487–501.
Kwok, R. and J. Comiso, 2002b: Spatial patterns of variability in Antarctic surface temperature: Connections to the Southern Hemisphere Annular Mode and the Southern Oscillation. *Geophys. Res. Lett.*, **29**, 1705, doi:10.1029/2002GL015415.
Labat, D., et al., 2004: Evidence for global runoff increase related to climate warming. *Advances in Water Resources*, **27,** 631–642.
Labitzke, K., and M. Kunze, 2005: Stratospheric temperature over the Arctic: Comparison of three data sets, *Meteorol. Zeitschrif*, **14**, 65–74.
Lambert, S.J., 1996: Intense extratropical Northern Hemisphere winter cyclone events: 1899–1991. *J. Geophys. Res.*, **101**, 21319–21325.
Lammers, R.B., et al., 2001: Assessment of contemporary Arctic river runoff based on observational discharge records. *J. Geophys. Res.-Atm.*, **106,** 3321–3334.
Lander, M., 1994: An exploratory analysis of the relationship between tropical storm formation in the Western North Pacific and ENSO. *Mon. Wea. Rev.*, **122**, 636–651.
Landsea, C.W., et al., 1998: The extremely active 1995 Atlantic hurricane season: Environmental conditions and verification of seasonal forecasts. *Mon. Wea. Rev.*,**126**, 1174–1193.
Langematz, U., et al. 2003: Thermal and dynamical changes of the stratosphere since 1979 and their link to ozone and CO_2 changes. *J. Geophys. Res*., **108**, 4027, doi:10.1029/2002JD002069.
Lanzante, J.R., S.A. Klein, and D.J. Seidel, 2003a: Temporal homogenization of monthly radiosonde temperature data. Pt I: Methodology. *J. Climate*, **16**, 224–240.
Lanzante, J.R., S.A. Klein, and D.J. Seidel, 2003b: Temporal homogenization of monthly radiosonde temperature data. Pt II: Trends, sensitivities, and MSU comparison. *J. Climate*, **16**, 241–262.
Laternser, M. and M. Schneebeli, 2003: Long-term snowclimate trends in the Swiss Alps (1931–99). *Int. J. Climatol*., **23**, 733–750.
Latif, M., 2001: Tropical Pacific/Atlantic ocean interactions at multi-decadal time scales. *Geophys. Res. Lett*., **28**, 539–542.
Latif, M., R. Kleeman, and C. Eckert, 1997: Greenhouse warming, decadal variability, or El Nino? An attempt to understand the anomalous 1990s. *J. Climate*, **10**, 2221–2239.
Lawrence, S.P., D.T. Llewellyn Jones, and S.J. Smith, 2004: The measurement of climate change using data from the Advanced Very High Resolution and Along Track Scanning Radiometers. *J. Geophys. Res.,* **109,** C08017, doi:10.1029/2003JC002104.
Lawrimore, J., et al., 2002: Beginning a new era of drought monitoring across North America. *Bull. Amer. Meteor. Soc.*, **83**, 1191–1192.
Le Barbe, L., Lebel, T., and Tapsoba, D., 2002: Rainfall variability in West Africa during the years 1950–1990. *J. Clim*., **15**, 187–202.
Lee, R. B., et al., 2004: On-orbit calibrations of the ERBE active cavity radiometers on the Earth Radiation Budget Satellite (ERBS): 1984–2003. *Proc. of SPIE*, **5234**, 433–444.

Lefebvre, W., et al., 2004: Influence of the Southern Annular Mode on the sea ice-ocean system. *J. Geophys. Res.*, **109**, C09005, doi:10.1029/2004JC002403.
Legates, D.R., and C.J. Willmott, 1990: Mean seasonal and spatial variability in gauge-corrected, global precipitation, *Int. J. Climatol.*, **10**, 111–127.
Leung, Y.K., et al., 2004: Climate Change in Hong Kong. Hong Kong Observatory. Hong Kong Observatory Technical Note, 107, 41 pp.
Levinson, D. H. (ed), 2005: State of the climate in 2004. *Bull. Amer. Meteor. Soc*., **86**, June 2005, S1-S84.
Levinson, D.H. and A.M. Waple (eds.), 2004: State of the climate in 2003. *Bull. Amer. Meteor. Soc*., **85**, June 2004, S1-S72.
Li, W. and P.M. Zhai, 2003: Variability in occurrence of China's spring dust storm and its relationship with atmospheric general circulation. *Acta Meteorologica Sinica*, ***17*(4)**, 396–405.
Li, Q., et al., 2004: Urban heat island effect on annual mean temperature during the last 50 years in China. *Theor.Appl.Climatol..*,**79**,165–174.
Liebmann, B., et al., 1999: Submonthly Convective Variability over South America and the South Atlantic Convergence Zone. *J. Climate,* **12,** 1877–1891.
Liebmann, B., et al., 2004: An observed trend in Central South American precipitation. *J. Climate*, **22**, 4357–4367.
Liepert, B.G., 2002: Observed reductions of surface solar radiation at sites in the United States and worldwide from 1961 to 1990. *Geophys. Res. Lett.*, **29,** 1421, 10.1029/2002GL014910.
Liepert, B.G., et al., 2004: Can Aerosols spin down the water cycle in a warmer and moister world? *Geophys. Res. Lett*. **31**, doi:10.1029/2003GL019060.
Lim, E.-P. and I. Simmonds, 2002: Explosive cyclone development in the Southern Hemisphere and a comparison with Northern Hemisphere events. *Mon. Wea. Rev.*, **130**, 2188–2209.
Limpasuvan, V. and D.L. Hartmann, 2000: Wave-maintained annular modes of climate variability. *J. Climate*, **13**, 4414–4429.
Limpasuvan, V., D. Thompson and D. Hartmann, 2004: The life cycle of northern hemispheric sudden stratospheric warmings. *J. Climate*, **17**, 2584–2596.
Lins, H.F. and J.R. Slack, 1999: Streamflow trends in the United States. *Geophys. Res. Lett.*, **26,** 227–230.
Linsley, B.K., et al., 2004: Geochemical evidence from corals for changes in the amplitude and spatial pattern of South Pacific interdecadal climate variability over the last 300 years. *Climate Dyn.*, **22**, doi:10.1007/s00382-003-0364-y.
Liu, B.H., et al., 2004a: A spatial analysis of pan evaporation trends in China, 1955–2000. *J. Geophys. Res.-Atmospheres*, **109,** D15102, doi:10.1029/2004JD004511.
Liu, J., J.A. Curry, and D.G. Martinson, 2004b: Interpretation of recent Antarctic sea ice variability. *Geophys. Res. Lett.*, **31**, L02205, doi:10.1029/2003GL018732.
Liu, K.S. and J.C.L. Chan, 2003: Climatological characteristics and seasonal forecasting of tropical cyclones making landfall along the south China coast. *Mon Wea. Rev*., **131**, 1650–1662.
Livezey, R.E. and T.M. Smith, 1999: Covariability of aspects of North American climate with global sea surface temperatures on interannual to interdecadal timescales. *J. Climate*, **12,** 289–302.
Loose, T. and R.D. Bornstein, 1977: Observations of Mesoscale Effects on Frontal Movement Through an Urban Area. *Mon. Wea. Rev*., **105**, 563–571.
Lorenz, D.J. and D.L. Hartmann. 2001: Eddy–Zonal Flow Feedback in the Southern Hemisphere. *J. Atmos. Sci.*, **58**, 3312–3327.
Lorenz, D.J. and D.L. Hartmann, 2003: Eddy-zonal flow feedback in the Northern Hemisphere winter. *J. Climate*, **16**, 1212–1227.
Lucarini, V. and G. L. Russell, 2002: Comparison of mean climate trends in the Northern Hemisphere between National Centers for Environmental Prediction and two atmosphere-ocean model forced runs. *J. Geophys. Res*., **107**, 4269, doi:10.1029/2001JD001247.
Luo, Y., et al., 2001: Characteristics of spatial distribution of yearly variation of aerosol optical depth over China in the last 30 years. *J. Geophys. Res*. **106**(D13), 14501–14513.
Luterbacher, J., et al., 2004: European seasonal and annual temperature variability, trends, and extremes since 1500. *Science*, **303**, 1499–1503.
Ma, Z.G. and C.B. Fu, 2003: Interannual characteristics of the surface hydrological variables over the arid and semi-arid areas of northern China. *Global and Planetary Change*, **37**, doi:10.1016/S0921-8181(02)00203-5.
Madden, R.A. and J. Williams, 1978: The correlation between temperature and precipitation in the United States and Europe. *Mon. Wea. Rev*., **106**, 142–147.

Maheras, P., et al., 2004: On the relationships between circulation types and changes in rainfall variability in Greece. *Int. J. Climatol*., **24**, 1695–1712.
Maistrova, V.V., et al., 2003: Long-term trends in temperature and specific humidity of free atmosphere in the Northern Polar region. *Doklady Earth Sci*. **391**, 755–759.
Manney, G., et al., 2005: The remarkable 2003–2004 winter and other recent warm winters in the Arctic stratosphere since the late 1990s, *J. Geophys. Res*., **110**, D04107, doi:10.1029/2004JD005367.
Manton, M.J., et al., 2001: Trends in extreme daily rainfall and temperature in Southeast Asia and the South Pacific: 1961–1998. *Int. J. Climatol*., **21**, 269–284.
Mantua, N.J., et al. 1997: A Pacific interdecadal climate oscillation with impacts on salmon production. *Bull. Am. Met. Soc*., **78**, 1069–1079.
Mantua, N. and S.J. Hare, 2002: The Pacific Decadal Oscillation. *J. Oceanogr*., **58**, 35–44.
Marengo, J., 2004: Interdecadal variability and trends of rainfall across the Amazon Basin. *Theor. Appl. Climatol*. **78**, 79–96.
Marengo, J.A., et al., 2004: Climatology of the low-level jet east of the Andes as derived from the NCEP–NCAR Reanalyses: Characteristics and temporal variability. *J. Climate,* **17**, 2261–2280.
Marquart, S., et al., 2003: Future development of contrail cover, optical depth, and radiative forcing: Impacts of increasing air traffic and climate change. *J. Climate*, **16**, 2890–2904.
Marsh, N.D. and H. Svensmark, 2000a: Low cloud properties influenced by cosmic rays. *Phys. Rev. Lett*., **85**, 5004–5007.
Marsh, N.D. and H. Svensmark, 2000b: Cosmic rays, clouds, and climate. *Space Science Rev*., **94**, 215–230.
Marsh, N. and H. Svensmark, 2003: Galactic cosmic ray and El Nino-Southern Oscillation trends in International Satellite Cloud Climatology Project D2 low-cloud properties. *J. Geophys. Res*., **108**, 4195, doi:10.1029/2001JD001264.
Marsh, N.D. and H. Svensmark, 2004: Comment by N. D. Marsh and H. Svensmark on "Solar influences on cosmic rays and cloud formation: a re-assessment.", *J. Geophys. Res*., **109**, D14205, doi:10.1029/2003JD004063.
Marshall, G.J., 2002: Analysis of recent circulation and thermal advection change on the northern Antarctic Peninsula. *Int. J. Climatol*., **22**, 1557–1567.
Marshall, G.J., 2003: Trends in the Southern Annular Mode from observations and reanalyses. *J. Climate*, **16**, 4134–4143.
Marshall, G.J., et al., 2004: Causes of exceptional atmospheric circulation changes in the Southern Hemisphere. *Geophys. Res. Lett.*, **31**, L14205, doi:10.1029/2004GL019952.
Marshall, J., H. Johnson, and J. Goodman, 2001: A study of the interaction of the North Atlantic Oscillation with the ocean circulation. *J. Climate*, **14**, 1399–1421.
Martyn, D., 1992: Climates of the world. Elsevier, 436 pp.
Mason, P.J., et al., 2003: The Second Report on the Adequacy of the Global Observing Systems for Climate in support of the UNFCCC. Global Climate Observing System **GCOS-82**, WMO/TD No. 1143, 74 pp.
Mauget, S.A., 2003a: Intra- to multidecadal climate variability over the continental United States: 1932–99. *J. Climate*, **16**, 2215–2231.
Mauget, S.A., 2003b: Multidecadal regime shifts in US streamflow, precipitation, and temperature at the end of the twentieth century. *J. Climate*, **16,** 3905–3916.
Maurer, E.P., et al., 2002: A long-term hydrologically based dataset of land surface fluxes and states for the conterminous United States. *J. Climate*, **15**, 3237–3251.
McBride, J.L., 1998: Indonesia, Papua New Guinea, and tropical Australia: The southern hemisphere monsoon. In: *Meteorology of the Southern Hemisphere*, [Karoly, D. and D. Vincent (eds.)]. American Meteorological Society Boston, U.S., 89–99.
McBride, J.L. and W.M. Frank, 1999: Relationships between stability and monsoon convection. *J. Atmos. Sci.*, **56**, 24–36.
McCabe, G.J., M.P. Clark, and M.C. Serreze, 2001: Trends in Northern Hemisphere surface cyclone frequency and intensity. *J. Climate*, **14**, 2763–2768.
McCabe G.J. and D.M. Wolock, 2002: Trends and temperature sensitivity of moisture conditions in the conterminous United States. *Climate Res.*, **20**, 19–29.
McCabe, G., M. Palecki, and J.L. Betancourt, 2004: Pacific and Atlantic Ocean influences on multi-decadal drought frequency in the United States. *Proc. Natl. Acad. Sci*., **101**, 4136–4141.
McCarthy, M.P. and R. Toumi, 2004: Observed interannual variability of tropical troposphere relative humidity. *J. Climate,* **17**, 3181–3191.
McKenzie, D., et al., 2004: Climatic change, wildfire, and conservation. *Conservation Biology*, **18**, 890–902.

Mears, C.A., M.C. Schabel, and F.J. Wentz, 2003: A reanalysis of the MSU channel 2 tropospheric temperature record. *J. Climate*, **16**, 3650–3664.
Mears, C.A. and F.J. Wentz, 2005: The effect of diurnal correction on satellite-derived lower tropospheric temperature. *Science*, **309**, doi: 10.1126/science.1114772 (online).
Meehl, G. A., et al., 2000: An Introduction to trends in extreme weather and climate events: Observations, socioeconomic impacts, terrestrial ecological impacts, and model projections. *Bull. Amer. Meteor. Soc.*, **81**, 413–416.
Mehta, A. and J. Susskind, 1999: Outgoing Longwave Radiation from the TOVS Pathfinder Path A data set. *J. Geophys. Res*., **104**, 12,193–12,212.
Mekis, E. and W.D. Hogg, 1999: Rehabilitation and analysis of Canadian daily precipitation time series, *Atmosphere-Ocean*, **37**, 53-85.
Menne, M.J. and C.N. Williams, Jr., 2005: Detection of undocumented change points: On the use of multiple test statistics and composite reference series. *J. Climate*, in press.
Menon, S., et al., 2002: Climate effects of black carbon aerosols in China and India , *Science*, **297**, 2250–2253.
Meredith, M.P., et al., 2004: Changes in the ocean transport through Drake Passage during the 1980s and 1990s, forced by changes in the Southern Annular Mode. *Geophys Res. Lett.* , **31**, L21305, doi:10.1029/2004GL021169.
Milly, P.C.D. and K.A. Dunne, 2001: Trends in evaporation and surface cooling in the Mississippi River basin. *Geophys. Res. Lett.*, **28,** 1219–1222.
Milly, P.C.D., et al., 2002: Increasing risk of great floods in a changing climate. *Nature*, **415,** 514–517.
Miloshevich, L.M., et al., 2004 Development and validation of a time-lag correction for Vaisala radiosonde humidity measurements. *J. Atmos. Oceanic Technol.*, **21**, 1305–1327.
Minnis, P., et al., 2004: Contrails, cirrus trends, and climate. *J. Climate*, **17**, 1671–1685.
Minschwaner. K. and A.E. Dessler, 2004: Water vapor feedback in the tropical upper troposphere: Model results and observations, *J. Climate*, **17**, 1272–1282.
Minobe, S., 1997: A 50–70 year climatic oscillation over the North Pacific and North America. *Geophys. Res. Lett.*, **24,** 683–686.
Minobe, S., 1999: Resonance in bidecadal and pentadecadal oscillations over the North Pacific: Role in climate regime shifts. *Geophys. Res. Lett*., **26**, 855–858.
Minobe, S. and T. Nakanowatari, 2002: Global structure of bidecadal precipitation variability in boreal winter. *Geophys. Res. Lett*., **29**, 1396, doi:10.1029/2001GL014447.
Mitas, C.M. and A. Clement, 2005: Has the Hadley cell been strengthening in recent decades? *Geophys. Res. Lett*., **32**, L03809, doi:10.1029/2004GL021765.
Mitchell, K.E., et al., 2004: The multi-institution North American Land Data Assimilation System (NLDAS): Utilizing multiple GCIP products and partners in a continental distributed hydrological modeling system. *J. Geophys. Res.*, **109,** D07S90, doi:10.1029/2003JD003823.
Mitchell, T.D., and P.D. Jones, 2005: An improved method of constructing a database of monthly climate observations and associated high-resolution grids. *Int. J. Climatol*., **25**, 693–712
Mo, K.C., 2000: Relationships between low-frequency variability in the Southern Hemisphere and sea surface temperature anomalies. *J. Climate*, **13,** 3599–3610.
Mo, K.C. and R.W. Higgins, 1998: The Pacific-South American modes and tropical convection during the Southern Hemisphere winter. *Mon. Wea. Rev.*, **126,** 1581–1596.
Moberg, A. and P.D. Jones, 2005: Trends in indices for extremes of daily temperature and precipitation in central and western Europe 1901–1999. *Int. J. Climatol.*, **25**, 1173–1188.
Molders, N. and M.A. Olson, 2004: Impact of urban effects on precipitation in high latitudes. *J. Hydromet*., **5**, 409–429.
Monahan, A.H., L. Pandolfo, and J.C. Fyfe, 2001: The preferred structure of variability of the Northern Hemisphere atmospheric circulation. *Geophys. Res. Lett.*, **28**, 1019–1022.
Mu, Q.Z., et al., 2002: Simulation Study on Variation of Western Pacific Subtropical High during the Last Hundred Years. *Chinese Science Bulletin*, **7**, 550–553.
Mudelsee, M., et al., 2003: No upward trends in the occurrence of extreme floods in central Europe. *Nature*, **425**, 166–169.
Nakamura, H. and T. Sampe, 2002: Trapping of synoptic-scale disturbances into the North-Pacific subtropical jet core in midwinter. *Geophys. Res. Lett*., **29**(16), doi:10.1029/2002GL015535.
Nakamura, H., T. Izumi, and T. Sampe, 2002: Interannual and decadal modulations recently observed in the Pacific storm track activity and East Asia winter monsoon. *J. Climate*, **15**, 1855–1874.

Nakamura, H., et al., 2004: Observed associations among storm tracks, jet streams and midlatitude oceanic fronts. *Earth's Climate: The Ocean-Atmosphere Interaction*, C. Wang, S.-P. Xie and J. A. Carton, Eds., *Geophys. Monogr.*, **147**, 329–346.
Naujokat, B., et al., 2002: The early major warming in December 2001—Exceptional? *Geophys. Res. Lett*., **29**, 2023, doi:10.1029/2002GL015316.
Nazemosadat, M.J. and I. Cordery, 2000: On the relationships between NSO and autumn rainfall in Iran. *Int. J. Climatol*., **20**, 47–61.
Nazemosadat M.J. and A.R. Ghasemi, 2004: Quantifying the ENSO-related shifts in the intensity and probability of drought and wet periods in Iran. *J. Climate*, **17**, 4005–4018.
Nedoluha, G.E., et al., 2003: An evaluation of trends in middle atmospheric water vapor as measured by HALOE, WVMS, and POAM, *J. Geophys. Res*., **108**, 4391, doi:10.1029/2002JD003332.
Neelin, J.D., et al., 1998: ENSO theory. *J. Geophys. Res.*, **103**, 14261–14290.
New, M., M. Hulme, and P.D. Jones, 1999: Representing twentieth-century space-time climate variability, Part 1: Development of a 1961–90 mean monthly terrestrial climatology. *J. Climate*, **12**, 829–856.
New, M.B., et al., 2005: Evidence of trends in daily climate extremes over Southern and West Africa.. *Geophys. Res. Lett.* submitted.
Newman, P.A. and E.R. Nash, 2000: Quantifying the wave drinking of the stratosphere. *J. Geophys. Res*., **105**, 12,485–12,497.
Newman, P.A. and E. R. Nash, 2005: The unusual Southern Hemisphere stratosphere winter of 2002. *J. Atmos. Sci.*, **62**, doi:10.1175/JAS-3323.1.
Newman, M., Compo, G., and M.A. Alexander, 2003: ENSO-forced variability of the Pacific Decadal Oscillation. *J. Climate*, **23**, 3853–3857.
Nicholls, N., 2004: The changing nature of Australian droughts. *Climatic Change*, **63**, 323–336.
Nobre P., and J. Shukla, 1996: Variations of sea surface temperature, wind stress, and rainfall over the tropical Atlantic and South America, *J. Climate*, **9**, 2464-2479.
Nogués-Paegle, J., et al., 2002: Progress in Pan American CLIVAR research: Understanding the South American monsoon. *Meteorologica*, **27**, 3–32.
Noone, D. and I. Simmonds, 2002: Annular variations in moisture transport mechanisms and the abundance of δ^{18}O in Antarctic snow. *J. Geophys. Res*., **107**, 4742, doi:10.1029/2002JD002262.
Norris, J.R., 1999: On trends and possible artifacts in global ocean cloud cover between 1952 and 1995. *J. Climate*, **12**, 1864–1870.
Norris, J.R., 2000: What can cloud observations tell us about climate variability. *Space Sci. Rev*., **94**, 375–380.
Norris, J.R., 2001: Has northern Indian Ocean cloud cover changed due to increasing anthropogenic aerosol? *Geophys. Res. Lett*., **28**, 3271–3274.
Norris, J.R., 2005a: Multidecadal changes in near-global cloud cover and estimated cloud cover radiative forcing. *J. Geophys. Res.*, accepted.
Norris, J. R., 2005b: Trends in upper-level cloud cover and atmospheric circulation over the Indo-Pacific region between 1952 and 1997. *J. Geophys. Res.* submitted.
Notholt, J. et al., 2005: Influence of tropospheric SO_2 emissions on particle formation and the stratospheric humidity, *Geophys. Res. Lett*., **32**, L07810, doi:10.1029/2004GL022159.
O'Carroll, A.G., R.W. Saunders, and J.G. Watts, 2005: The measurement of the sea surface temperature climatology by satellites from 1991 to 2005. *J. Climate*, submitted.
Ohmura, A. and M. Wild, 2002: Is the hydrological cycle accelerating? *Science*, **298**, 1345–1346.
Oinas, V., et al., 2001: Radiative cooling by stratospheric water vapor: big differences in GCM results, *Geophys. Res. Lett*. **28**, 2791–2794.
Oke, P.R. and M.H. England, 2004: Oceanic response to changes in the latitude of the Southern Hemisphere subpolar westerly winds. *J. Climate*, **17**, 1040–1054.
Oki, T., T. Nishimura, and P. Dirmeyer, 1999: Assessment of annual runoff from land surface models using Total Runoff Integrating Pathways (TRIP). *J. Meteor. Soc. Japan*, **77**, 235–255.
Oltmans, S.J., et al., 2000: The increase in stratospheric water vapor from balloon borne, frostpoint hygrometer measurements at Washington, DC, and Boulder, Colorado. *Geophys. Res. Lett*, **27**, 3453–3456.
Omran, M.A., 2000: Analysis of solar radiation over Egypt. *Theor. Appl. Climatol*. **67**, 225–240.
Oort, A.H. and Yienger, J.J., 1996: Observed interannual variability in the Hadley circulation and its connection to ENSO. *J. Climate*, **9**, 2751–2767.

Orr, A., et al., 2004: A 'low-level' explanation for the recent large warming trend over the western Antarctic Peninsula involving blocked winds and changes in zonal circulation. *Geophys. Res. Lett.*, **31**, L06204, doi:10.1029/2003GL019160.

Osborn, T.J. and M. Hulme, 2002: Evidence for trends in heavy rainfall events over the U.K. *Phil. Trans. Roy. Soc., Series A*, **360**, 1313–1325.

Osborn, T.J., et al., 2000: Observed trends in the daily intensity of United Kingdom precipitation. *Int. J. Climatol.*, **20**, 347–364.

Ostermeier, G.M. and J.M. Wallace, 2003: Trends in the North Atlantic Oscillation – Northern Hemisphere annular mode during the twentieth century. *J. Climate*, **16**, 336–341.

Paciorek, C.J., et al., 2002: Multiple indices of Northern Hemisphere cyclone activity, winters 1949–99. *J. Climate*, **15**, 1573–1590.

Pagano, T., et al., 2004: Water year 2004: Western water managers feel the heat. *Eos, Trans. Amer. Geophys. Union*, **85**, 392–393.

Pallé, E., et al., 2004: Changes in Earth's reflectance over the past two decades. *Science*, **304**, 1299–1301.

Palmer, T.N., 1999: A nonlinear dynamical perspective on climate prediction. *J. Climate*, **12**, 575–591.

Park, Y., F. Roquet, and F. Vivier, 2004: Quasi-stationary ENSO wave signals versus the Antarctic Circumpolar Wave scenario. *Geophys. Res. Lett*., **31**, L09315, doi:10.1029/2004GL019806.

Parker, D.E., 2004: Large-scale warming is not urban. *Nature*, **432**, 290–290.

Parker, D.E. and D.I. Cox, 1995: Towards a consistent global climatological rawinsonde data-base. *Int. J. Climatol.*, **15**, 473–496.

Parker, D.E., M. Gordon, D.P.N. Cullum, D.M.H. Sexton, C.K. Folland and N. Rayner, 1997: A new global gridded radiosonde temperature data base and recent temperature trends. *Geophys. Res. Lett.*, **24**, 1499–1502.

Parker, D.E., L.V. Alexander, and J. Kennedy, 2004: Global and regional climate in 2003. *Weather*, **59,** 145–152.

Pawson, S. and B. Naujokat, 1999: The cold winters of the middle 1990s in the northern lower stratosphere. *J. Geophys. Res*., **104**, 14209–14222.

Pekárová, P., P. Miklánek, and J. Pekár, 2003: Spatial and temporal runoff oscillation analysis of the main rivers of the world during the 19th–20th centuries. *J. Hydrol.*, **274,** 62–79.

Pepin, N.C. and D.J. Seidel, 2005: A global comparison of surface and free-air temperatures at high elevations. *J. Geophys. Res*., **110**, D03104, doi:10.1029/2004JD005047.

Peterson, T.C., 2003: Assessment of urban versus rural *in situ* surface temperatures in the contiguous United States: no difference found. *J. Climate*, **16,** 2941–2959.

Peterson, T.C. and T.W. Owen, 2005: Urban heat island assessment: Metadata are important. *J. Climate*, **18**, 2637–2646.

Peterson, T.R. and R.S. Vose, 1997: An overview of the Global Historical Climatology Network Temperature Database, *Bull. Am. Met. Soc*., **78**, 2837–2848.

Peterson, T. C., V. S. Golubev, and P. Y. Groisman, 1995: Evaporation losing its strength. *Nature*, **377,** 687–688.

Peterson, T.C., et al., 1998: Global Historical Climatology Network (GHCN) quality control of monthly temperature data. *Int. J. Climatol*., **18**, 1169-1179.

Peterson, T.C., et al., 2000: A blended satellite – *in situ* near-global surface temperature dataset. *Bull. Amer. Meteorol. Soc.*, **81,** 2157–2164.

Peterson, T.C., et al., 2002, "Recent Changes in Climate Extremes in the Caribbean Region". *J. Geophys. Res. - Atmospheres*, **107**, 4601, doi:10.1029/2002JD002251.

Pezza, A.B. and I. Simmonds, 2005: The first South Atlantic hurricane: unprecedented blocking, low shear and climate change. *Geophys. Res. Lett.*, **32**, L15712, doi:10.1029/2005GL023390.

Philipona, R., and B. Dürr, 2004: Radiative forcing - measured at Earth's surface - corroborate the increasing greenhouse effect. *Geophys. Res. Lett*., **31**, L03202, doi:1029/2003GL018765.

Picon. L. et al., 2003: A new METEOSAT "water vapor" archive for climate studies. *J. Geophys. Res*., **108**, 4301, doi:10.1029/2002JD002640.

Piechota, T., et al., 2004: The western drought: How bad is it? *Eos, Trans. Amer. Geophys. Union*, **85**(32), 301.

Pinker, R.T., B. Zhang, and E.G. Dutton, 2005: Do satellites detect trends in surface solar radiation?, *Science*, **308**, 850–854.

Plummer, N., et al., 1999. Changes in climate extremes over the Australian region and New Zealand during the Twentieth Century. *Climatic Change*, 42(1), 183–202.

Polyakov, I.V., et al., 2003: Variability and trends of air temperature in the Maritime Arctic, 1875–2000. *J. Climate*, **16,** 2067–2077.
Poncelet, L., 1959: Comparison of raingauges. *WMO Bull.* **8(4)**, 186–190.
Pottier, C., et al., 2004: Dominant propagating signals in sea level anomalies in the Southern Ocean. *Geophys. Res. Lett*., **31**, L11305, doi:10.1029/2004GL019565.
Power, H.C. and D.M. Mills, 2005: Solar radiation climate change over South Africa and an assessment of the radiative impact of volcanic eruptions. *Int. J. Climatol*., **25**, 295–318.
Power, S., et al., 1998: Australian temperature, Australian rainfall and the Southern Oscillation, 1910–1992: coherent variability and recent changes. *Aust. Met. Mag*., **47**, 85–101.
Power, S., et al., 1999a: Decadal climate variability in Australia during the twentieth century. *Int. J. Climatol*., **19**, 169–184.
Power, S., et al., 1999b: Inter-decadal modulation of the impact of ENSO on Australia. *Climate Dyn.*, **15**. 319–324.
Probst, J.L. and Y. Tardy, 1987: Long-range streamflow and world continental runoff fluctuations since the beginning of this century. *J. Hydrol*., **94**, 289–311.
Probst, J.L. and Y. Tardy, 1989: Global runoff fluctuations during the last 80 years in relation to world temperature-change. *American J. Science*, **289,** 267–285.
Przybylak, R., 2000: Diurnal temperature range in the Arctic and its relation to hemispheric and Arctic circulation patterns. *Int. J. Climatol.*, **20**, 231–253.
Qian, W.H., L.S. Quan, and S. Y. Shi, 2002: Variations of dust storm in China and its climatic control. *J. Climate*, **15**, 1216–1229.
Qian, W. H., et al., 2003: Centennial-scale dry-wet variations in East Asia. *Climate Dyn.*, **21**, 77–89.
Qian, T., et al., 2005: Simulation of global land surface conditions from 1948–2002. Part I: Forcing data and evaluation. *J. Hydrometeorol.*, submitted.
Quadrelli, R., Pavan, V., and Molteni, F., 2001: Wintertime variability of Mediterranean precipitation and its links with large-scale circulation anomalies. *Climate Dyn*., **17**, 457–466.
Quadrelli, R. and J.M. Wallace, 2004: A simplified linear framework for interpreting patterns of Northern Hemisphere wintertime climate variability. *J. Climate*, **17**, 3728–3744.
Ramanathan, V., et al., 2001: Aerosols, climate and the hydrological cycle. *Science*, **294**, 2119–2124.
Ramaswamy, V., et al., 2001: Stratospheric temperature changes: observations and model simulations. *Rev. Geophys*., **39**, 71–122.
Randel D.L, et al., 1996: A new global water vapor dataset. *Bull. Amer. Meteor. Soc*., **77**, 1233–1246.
Randel, W.J. and F. Wu, 1999: Cooling of the Arctic and Antarctic polar stratospheres due to ozone depletion. *J. Climate*, **12**, 1467–1479.
Randel, W.J. and F. Wu, 2005: Biases in stratospheric temperature trends derived from historical radiosonde data. *J. Climate* Accepted .
Randel, W.J., F. Wu, and D.J. Gaffen, 2000: Interannual variability of the tropical tropopause derived from radiosonde data and NCEP reanalyses. *J. Geophys. Res.*, **105**, 15,509–15,524.
Randel, W., F. Wu, and R. Stolarski, 2002: Changes in column ozone correlated with the stratospheric EP flux. *J. Meteorol. Soc. Japan*, **80**, 849–862.
Randel, W.J., et al., 2004: Interannual changes of stratospheric water vapor and correlations with tropical tropopause temperatures, *J. Atmos. Sci.*, **61**, 2133–2148.
Raphael, M.N., 2003: Impact of observed sea-ice concentration on the Southern Hemisphere extratropical atmospheric circulation in summer. *J. Geophys. Res.*, **108**, 4687, doi:10.1029/2002JD003308.
Rayner, N.A., et al., 2003: Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J. Geophys. Res.*, **108**, 4407, doi:10.1029/2002JD002670.
Rayner, N.A., et al., 2005: Improved analyses of changes and uncertainties in marine temperature measured *in situ* since the mid-nineteenth century. *J. Climate*, Submitted.
Read, W.G., et al., 2004: Dehydration in the tropical tropopause layer: Implications from the UARS Microwave Limb Sounder. *J. Geophys. Res*., **109**, D06110, doi:10.1029/2003JD004056.
Regonda, S.K., et al., 2005: Seasonal cycle shifts in hydroclimatology over the Western U.S. *J. Climate.*, **18**, 372–384.
Renwick, J.A., 1998: ENSO-related variability in the frequency of South Pacific blocking. *Mon. Wea. Rev.*, **126,** 3117–3123.
Renwick, J.A., 2002: Southern Hemisphere circulation and relations with sea ice and sea surface temperature. *J. Climate*, **15,** 3058–3068.

Renwick, J.A., 2004: Trends in the Southern Hemisphere polar vortex in NCEP and ECMWF reanalyses. *Geophys. Res. Lett.*, **31**, L07209, doi:10.1029/2003GL019302.
Renwick, J.A., 2005: Persistent positive anomalies in the Southern Hemisphere circulation. *Mon. Wea. Rev.***,** **133**, 977–988.
Renwick, J. A. and M. J. Revell, 1999: Blocking over the South Pacific and Rossby wave propagation. *Mon. Wea. Rev.*, **127,** 2233–2247.
Revercomb, H.E., et al., 2003: The ARM program's water vapor intensive observation periods: Overview, initial accomplishments, and future challenges. *Bull. Amer. Meteor. Soc.* **84**, 217-236.
Rex, M., et al., 2004: Arctic ozone loss and climate change. *Geophys. Res. Lett.*, **31**, L04116, doi:10.1029/2003GL018844.
Reynolds, R.W., et al., 2002: An improved in situ and satellite SST analysis for climate. *J. Climate,* **15,** 1609–1625.
Reynolds, R.W., C.L. Gentemann, and F. Wentz, 2004: Impact of TRMM SSTs on a climate-scale SST analysis. *J. Climate*, **17,** 2938–2952.
Robertson, A.W., C.R. Mechoso, and N.O. García, 2001a: Interannual prediction of the Paraná River. *Geophys. Res. Lett*., **28**, 4235–4238.
Robertson, F.R., R.W. Spencer, and D.E. Fitzjarrald, 2001b: A new satellite deep convective ice index for tropical climate monitoring: possible implications for existing oceanic precipitation data sets, *Geophys. Res. Lett*., **28**, 251–254.
Robertson F.R., D.E. Fitzjarrald, and C.D. Kummerow, 2003: Effects of uncertainty in TRMM precipitation radar path integrated attenuation on interannual variations of tropical oceanic rainfall. *Geophys. Res. Lett*., **30**, 1180, doi:10.1029/2002GL016416.
Robeson, S., 2004: Trends in time-varying percentiles of daily minimum and maximum temperature over North America. *Geophys. Res. Lett*, **31**, L04203, doi:10.1029/2003GL019019.
Robinson, P.J., 2000: Temporal trends in United States dew point temperatures. *Int. J. Climatol.,* **20**, 985–1002.
Robinson, G., 2005: Rigorous quality control of the ISCCP B3 10μm dataset: Automated identification and treatment of radiometric noise, navigation and rectification errors. *J. Atmos. Oceanic Tech.*, submitted.
Robock, A., et al., 2000: The global soil moisture data bank. *Bull. Amer. Meteor. Soc.*, **81,** 1281–1299.
Robock, A., et al., 2005: Forty five years of observed soil moisture in Ukraine: No summer desiccation (yet). *Geophys. Res. Lett.***,** **32**, L03401, doi:10.0129/2004GL021914.
Röckmann, T. et al., 2004: The impact of anthropogenic chlorine emissions, stratospheric ozone change and chemical feedbacks on stratospheric water. *Atmos. Chem. Phys*., **4**, 693-699.
Roderick, M.L. and G.D. Farquhar, 2002: The cause of decreased pan evaporation over the past 50 years. *Science*, **298**, 1410–1411.
Roderick, M.L. and G.D. Farquhar, 2004: Changes in Australian pan evaporation from 1970–2002. *Int. J. Climatol.* **24**, 1077–1090.
Rodwell, M.J., 2003: On the predictability of North Atlantic climate. In: *The North Atlantic Oscillation: Climatic significance and environmental impact* [Hurrell, J.W., et al. (eds.)]. *Geophysical Monograph*, **134**, American Geophysical Union, Washington, DC, pp. 173–192.
Roscoe, H.K., 2004: A review of stratospheric H_2O and NO_2, *Adv. Space Res.*, **34**, 1747–1754.
Rosenfeld, D., 2000: Suppression of rain and snow by urban and industrial air pollution. *Science*, **287**, 1793–1796.
Rosenlof, K.H., et al. 2001: Stratospheric water vapor increases over the past half-century. *Geophys. Res. Lett*, **28**, 1195–1198.
Rosenlof, K.H., 2002: Transport changes inferred from HALOE water and methane measurements, *J. Meteor. Soc. Japan*, **80**, 831–848.
Rossow, W.B., and R.A. Schiffer, 1991: ISCCP cloud data products. *Bull. Amer. Meteor. Soc*., **72**, 2–20.
Rossow, W.B. and E.N. Dueñas, 2004: The International Satellite Cloud Climatology Project (ISCCP) web site. *Bull. Am. Meteorol. Soc.*, **85**, doi:10.1175/BAMS-85-2-167.
Rudolf, B., et al., 1998: Precipitation Data for Verification of NWP Model Re-Analyses: The Accuracy of Observational Results. Proc. First WCRP Intl Conf. Reanalyses, Washington, DC, U.S., 27-31 Oct 1997, WMO/TD-No. 876, pp. 215–218.
Rudolf, B. and J. Rapp, 2003: The Century Flood of the River Elbe in August 2002: Synoptic Weather Development and Climatological Aspects. Quarterly Report German NWP-System of the Deutscher Wetterdienst, 2, Pt 1, 8-23.

Rudolf, B., et al., 1994: Terrestrial precipitation analysis: Operational method and required density of point measurements. In: *Global Precipitations and Climate Change* [Bubois, M. and F. Désalmand (eds.)]. NATO ASI Series, Vol. **I 26**, Springer Verlage, pp. 173–186.
Ruiz-Barradas, A. and S. Nigam, 2005: Warm-season rainfall variability over the US Great Plains in observations, NCEP and ERA-40 reanalyses, and NCAR and NASA atmospheric model simulations: Intercomparisons for NAME. *J. Climate*, **18**, 1808–1830.
Russak, V., 1990: Trends of solar radiation, cloudiness and atmospheric transparency during recent decades in Estonia. *Tellus* **42B**, 206–210.
Rusticucci, M. and O. Penalba, 2000: Precipitation seasonal cycle over southern South America. *Climate Res*., **16,** 1–15, 2000.
Rusticucci, M. and M. Barrucand, 2004: Observed trends and changes in temperature extremes over Argentina. *J. Climate,* **17**, 4099–4107.
Rutllant, J. and H. Fuenzalida, 1991: Synoptic aspects of the central Chile rainfall variability associated with the Southern Oscillation. *Int. J. Climatol.*, **11,** 63–76.
Salinger, M.J., J.A. Renwick, and A.B. Mullan, 2001: Interdecadal Pacific Oscillation and South Pacific climate. *Int. J. Climatol.*, **21**, 1705–1721.
Salinger, M.J., G.M. Griffiths, and A. Gosai, 2005: Extreme pressure differences at 0900 NZST and winds across New Zealand. *Int. J. Climatol*.**, 25**, 1203–1222.
Santer B.D., et al. 1999: Uncertainties in observationally based estimates of temperature change in the free atmosphere. *J. Geophys. Res*., **104**, 6305–6333.
Santer, B.D., et al., 2004: Identification of anthropogenic climate change using a second-generation reanalysis. *J. Geophys. Res.*, **109,** D21104, doi:1029/2004/JD005075.
Santer, B.D., et al., 2005: Amplification of surface temperature trends and variability in the tropical atmosphere. *Science*, **309**, doi: 10.1126/science.1114867 (online).
Sarkar, S., R.P. Singh, and M. Kafatos, 2004: Further evidences for the weakening relationship of Indian rainfall and ENSO over India. *Geophys. Res. Lett*., **31**, L13209, doi:10.1029/2004GL020259.
Schär, C., et al., 2004: The role of increasing temperature variability in European summer heat waves. *Nature*, **427**, 332–336.
Scherrer, S.C., et al., 2005: Two dimensional indices of atmospheric blocking and their statistical relationship with winter climate patterns in the Euro-Atlantic region. *Int. J. Climatol.*, accepted.
Schlesinger, M.E. and N. Ramankutty, 1994: An oscillation in the global climate system of period 65–70 years. *Nature,* **367**, 723–726.
Schmidli, J., et al., 2002: Mesoscale precipitation in the Alps during the 20th century. *Int. J. Climatol.*, **22**, 1049–1074.
Schmidli, J. and C. Frei, 2005: Trends of heavy precipitation and wet and dry spells in Switzerland during the 20th century. *Int. J. Climatol.*, **25**, 753–771.
Schmith, T., Kaas, E., and T.-S. Li, 1998: Northeast Atlantic winter storminess 1875–1995 re-analysed. *Climate Dyn.*, **14**, 529–536.
Schneider, D.P., E.J. Steig, and J.C. Comiso, 2004: Recent climate variability in Antarctica from satellite-derived temperature data. *J. Climate*, **17**, 1569–1583.
Schönwiese, C.-D. and J. Rapp, 1997: Climate trend atlas of Europe based on observations 1891–1990. Dordrecht, Kluwer Academic Press, 228 pp.
Schreck, C.J. III and F.H.M. Semazzi, 2004: Variability of the recent climate of Eastern Africa. *Int. J. Climatol.*, **24**, 681–701.
Schroder, T., et al., 2003: Validating the microwave sounding unit stratospheric record using GPS occultation. *Geophys. Res. Lett*., **30**, 1734, doi:10.1029/2003GL017588.
Seidel, D.J., et al., 2001: Climatological characteristics of the tropical tropopause as revealed by radiosondes. *J. Geophys. Res.*, **106**, 7857–7878.
Seidel, D.J., et al., 2004: Uncertainty in signals of large-scale climate variations in radiosonde and satellite upper-air temperature datasets. *J. Climate,* **17,** 2225–2240.
Seidel, D. and J. Lanzante, 2004: An assessment of three alternatives to linear trends for characterizing global atmospheric temperature changes. *J. Geophys. Res.*, **109**, D14108, doi:10.1029/2003JD004414.
Sen Roy, S. and R.C. Balling, 2004: Trends in extreme daily rainfall indices in India. *Internat. J. Climatol*., **24**, 457–466.
Sen Roy, S. and R.C. Balling, 2005: Analysis of trends in maximum and minimum temperature, diurnal temperature range, and cloud cover over India. *Geophys. Res. Lett.*, **32**, L12702, doi:10.1029/2004GL022201.

Serreze, M.C., et al., 1997: Icelandic low cyclone activity: climatological features, linkages with the NAO, and relationships with recent changes in the Northern Hemisphere circulation. *J. Climate*, **10,** 453–464.
Sevruk, B., 1982: Methods of correction for systematic error in point precipitation measurement for operational use. *WMO Oper. Hydrol. Rep*., **21**, Publ. 589, World Meteorological Organization, Geneva, Switzerland, 91 pp.
Sevruk, B. and Hamon, W.R 1984: International comparison of national precipitation gages with a reference pit gage. *WMO Instruments and Observing Methods Report No. 17*, World Meteorological Organization, Geneva, Switzerland, 111 pp.
Sexton, D.M.H., 2001: The effect of stratospheric ozone depletion on the phase of the Antarctic Oscillation. *Geophys. Res. Lett.*, **28**, 3697–3700.
Shabbar, A. and W. Skinner, 2004: Summer drought patterns in Canada and the relationship to global sea surface temperatures. *J. Climate*, **17**, 2866–2880.
Shepherd, J. M., H. Pierce, and A. J. Negri, 2002: Rainfall modification by major urban areas: Observations from spaceborne rain radar on the TRMM satellite. *J. Appl. Meteor*., **41**, 689–701.
Shepherd, J.M. and S.J. Burian, 2003: Detection of urban-induced rainfall anomalies in a major coastal city. *Earth Interactions,* **7**, 1–17.
Shepherd, J.M., L. Taylor, and C. Garza, 2004: A dynamic multi-criteria technique for siting NASA-Clark Atlanta rain gauge network. *J. Atmos. Oceanic Technol.*, **21**, 1346–1363.
Shepherd, J.M., 2005: Evidence of urban-induced precipitation variability in arid climate regimes. *J. Arid Environments* (in review).
Shepherd, T.G. and T.A. Shaw, 2004: The angular momentum constraint on climate sensitivity and downward influence in the middle atmosphere. *J. Atmos. Sci*., **61**, 2899–2908.
Sherwood, S.C., 2002: A microphysical connection among biomass burning, cumulus clouds, and stratospheric moisture. *Science*, **295**, 1271–1275.
Sherwood, S., J. Lanzante, and C. Meyer, 2005: Radiosonde daytime biases and late 20th century warming. *Science*, **309**, doi: 10.1126/science.1115640 (online).
Shine, K. P., et al., 2003: A comparison of model-simulated trends in stratospheric temperatures. *Quart. J. Roy. Meteor. Soc.*, **129**, 1565–1588.
Silvestri, G.E. and C.S. Vera, 2003: Antarctic Oscillation signal on precipitation anomalies over southeastern South America. *Geophys. Res. Lett.*, **30**, 2115, doi:10.1029/2003GL018277.
Simmonds, I., 2003: Modes of atmospheric variability over the Southern Ocean. *J. Geophys. Res.*, **108,** 8078, doi:10.1029/2000JC000542.
Simmonds, I. and K. Keay, 2000: Variability of Southern Hemisphere extratropical cyclone behavior 1958–97. *J. Climate*, **13**, 550–561.
Simmonds, I. and K. Keay, 2002: Surface fluxes of momentum and mechanical energy over the North Pacific and North Atlantic Oceans. *Meteor. Atmos. Phys.*, **80**, 1–18.
Simmonds, I., K. Keay, and E.-P. Lim, 2003: Synoptic activity in the seas around Antarctica. *Mon. Wea. Rev.*, **131,** 272–288.
Simmons, A.J., et al., 2004: Comparison of trends and low-frequency variability in CRU, ERA-40 and NCEP/NCAR analyses of surface air temperature. *J. Geophys. Res*., **109**, D24115, doi:10.1029/2004JD006306.
Simmons, A.J., et al., 2005: ECMWF analyses and forecasts of stratospheric winter polar vortex breakup: September 2002 in the Southern Hemisphere and related events. *J. Atmos. Sci*., **62**, 668–689.
Sinclair, M.R., J.A. Renwick, and J.W. Kidson, 1997: Low-frequency variability of Southern Hemisphere sea level pressure and weather system activity. *Mon. Wea. Rev*., **125**, 2531–2543.
Slutz, R. J., et al., 1985: *COADS: Comprehensive Ocean-Atmosphere Data Set. Release 1*. NOAA Climate Research Program, Environmental Research Laboratories, Boulder, CO, 262 pp.
Small, E.E., L.C. Sloan, and R. Nychka, 2001: Changes in surface air temperature caused by desiccation of the Aral Sea. *J. Climate*, **14**, 284–299.
Smith, C.A., J.D. Haigh, and R. Toumi, 2001: Radiative forcing due to trends in stratospheric water vapour. *Geophys. Res. Lett*, **28**, 179–182.
Smith, I., 2004: An assessment of recent trends in Australian rainfall. *Austr. Meteor. Mag.*, **53**, 163–173.
Smith, L. C., 2000: Trends in Russian Arctic river-ice formation and breakup, 1917 to 1994. *Physical Geography*, **21**, 46–56.
Smith, T.M. and R.W. Reynolds, 2002: Bias corrections for historical sea surface temperatures based on marine air temperatures. *J. Climate*, **15,** 73–87.

Smith, T.M. and R.W. Reynolds, 2004: Improved extended reconstruction of SST (1854–1997). *J. Climate,* **17,** 2466–2477.
Smith, T.M. and R.W. Reynolds, 2005: A global merged land and sea surface temperature reconstruction based on historical observations (1880–1997). *J. Climate*, **18**, 2021–2036.
Smith, T.M., et al., 2005: New surface temperature analyses for climate monitoring. *Geophys. Res. Lett.* **32**, L14712, doi:10.1029/2005GL023402.
Smits, A., A.M.G. Klein Tank, and G.P. Können, 2005: Trends in storminess over the Netherlands, 1962–2002*. Int. J. Climatol*., **25**, 1331-1344.
Snow, J.T. (ed.), 2003: Special Issue: European Conference on Severe Storms. *Atmos. Res.*, **67–68**, 703 pp.
Soden, B.J. and S.R. Schroeder, 2000: Decadal variations in tropical water vapor: A comparison of observations and a model simulation. *J. Climate*, **13**, 3337–3340.
Soden, B. J., et al., 2002: Global cooling after the eruption of Mount Pintubo: A test of climate feedback by water vapor. *Science*, **296**, 727–730.
Soden, B.J., et al., 2004: An analysis of satellite, radiosonde, and lidar observations of upper tropospheric water vapor from the Atmospheric Radiation Measurement Program. *J. Geophys. Res*., **109**, D04105, doi:10.1029/2003JD003828.
Soden, B.J., et al., 2005: The radiative signature of upper tropospheric moistening, *Science*, submitted.
Sohn, B.-J., and E. A. Smith, 2003: Explaining sources of discrepancy in SSM/I water vapor algorithms. *J. Climate*, **16**, 3229–3255.
Song, Y. and W.A. Robinson, 2004: Dynamical mechanisms for stratospheric influences on the troposphere. *J. Atmos. Sci.*, **61**, 1711–1725.
Sparks, J., D. Changnon, and J. Starke, 2002: Changes in the frequency of extreme warm-season surface dewpoints in northeastern Illinois: Implications for cooling-system design and operation. *J. Appl. Meteor.* **41**, 890–898.
Stafford, J.M., G. Wendler, and J. Curtis, 2000: Temperature and precipitation of Alaska: 50 year trend analysis. *Theoret. Appl. Climatol.*, **67**, 33–44.
Stanhill, G. and S. Cohen, 2001: Global dimming, a review of the evidence for a widespread and significant reduction in global radiation with a discussion of its probable causes and possible agricultural consequences. *Agric. For. Meteorol*. **107**, 255–278.
Sterl, A., 2001: On the impact of gap-filling algorithms on variability patterns of reconstructed oceanic surface fields. *J. Geophys. Res*., **28**, 2473–2476.
Sterl, A., 2004: On the (in)homogeneity of reanalysis products. *J. Climate*, **17**, 3866–3873.
Stewart, I.T., D.R. Cayan, and M.D. Dettinger, 2005: Changes towards earlier streamflow timing across western North America. *J. Climate*, **18**, 1136–1155.
Stone, D.A., A.J. Weaver, and F.W. Zwiers, 2000: Trends in Canadian precipitation intensity. *Atmos. Ocean*, **38**(2), 321–347.
Stone, R.S., et al., 2002: Earlier spring snowmelt in northern Alaska as an indicator of climate change, *J. Geophys. Res.*, **107**, doi:10.1029/2000JD000286.
Straus, D.M. and J. Shukla, 2002: Does ENSO force the PNA? *J. Climate*, **15,** 2340–2358.
Stuber, N., et al., 2001: Is the climate sensitivity to ozone perturbations enhanced by stratospheric water vapor feedback? *Geophys. Res. Lett*., **28**, 2887–2890.
Sturaro, G., 2003: A closer look at the climatological discontinuities present in the NCEP/NCAR reanalysis temperature due to the introduction of satellite data. *Climate Dyn.*, **21**, 309–316.
Sun, B.M., 2003: Cloudiness over the contiguous United States: Contemporary changes observed using ground-based and ISCCP D2 data. *Geophys. Res. Lett.*, **30**, doi:10.1029/2002GL015887.
Sun, B., and R. S. Bradley, 2002: Solar influences on cosmic rays and cloud formation: a re-assessment. *J. Geophys. Res*., **107**(D14), doi:10.1029/2001JD000560.
Sun, B. and R.S. Bradley, 2004: Reply to comment by N.D. Marsh and H. Svensmark on "Solar influences on cosmic raya and cloud formation: A reassessment". *J. Geophys. Res.* **109**, D14206, doi:10.1029/2003JD004479.
Sun, B. and P.Ya. Groisman, 2000: Cloudiness variations over the former Soviet Union. *Int. J. Climatol.*, **20**, 1097–1111.
Sun, B.M., P.Ya. Groisman, and I.I. Mokhov, 2001: Recent changes in cloud-type frequency and inferred increases in convection over the United States and the former USSR. *J. Climate*, **14**, 1864–1880.
Sun, B.M. and P.Ya. Groisman, 2004: Variations in low cloud cover over the United States during the second half of the twentieth century. *J. Climate*, **17**, 1883–1888.

Sutton, R.T. and Hodson, D.L.R., 2003: Influence of the ocean on North Atlantic climate variability 1871–1999. *J. Climate*, **16**, 3296–3313.
Svensmark, H., 1998: Influence of cosmic rays on Earth's climate. *Phys. Rev. Lett*., **81**, 5027–5030.
Svensmark, H. and E. Friis-Christensen, 1997: Variation of cosmic ray flux and global cloud coverage – a missing link in solar-climate relationships. *J. Atmos. Solar-Terr. Phys*., **59**, 1225–1232.
Swanson, R. E., 2003: Evidence of possible sea-ice influence on Microwave Sounding Unit tropospheric temperature trends in polar regions. *Geophys. Res., Lett.* **30**, 2040, doi:10.1029/2003GL017938.
Tett, S.F.B. and P.W. Thorne, 2004: Tropospheric temperature series from satellites. *Nature*, doi:10.1038/nature03208.
Thielen, J., et al., 2000: The possible influence of urban surfaces on rainfall development: a sensitivity study in 2D in the meso-gamma scale. *Atmos. Res*., **54**, 15–39.
Thompson, D.W. and S. Solomon, 2002: Interpretation of recent Southern Hemisphere climate change. *Science*, **296**, 895–899.
Thompson. D.W.J. and S. Solomon, 2005: Recent stratospheric climate trends: Global structure and tropospheric linkages, *J. Climate*, (accepted).
Thompson, D.W.J and J. M. Wallace, 1998: The Arctic Oscillation signature in the wintertime geopotential height and temperature fields. *Geophys. Res. Lett.*, **25**, 1297–1300.
Thompson, D.W.J. and J. M. Wallace, 2000: Annular modes in the extratropical circulation, Part I: Month-to-month variability. *J. Climate*, **13**, 1000–1016.
Thompson, D.W.J., J.M. Wallace, and G.C. Hegerl, 2000: Annular modes in the extratropical circulation. Part II: Trends. *J. Climate*, **13**, 1018–1036.
Thompson, D.W.J., M.P. Baldwin, and J.M. Wallace, 2002: Stratospheric connection to Northern Hemisphere wintertime weather: Implications for prediction. *J. Climate*, **15**, 1421–1428.
Thompson, D.W.J., S. Lee, and M.P. Baldwin, 2003: Atmospheric processes governing the Northern Hemisphere Annular Mode/North Atlantic Oscillation. In: *The North Atlantic Oscillation: Climatic Significance and Environmental Impact* [Hurrell, J.W., et al. (eds.)] *Geophysical Monograph*, **134**, pp. 81–112.
Thompson, D.W.J., M.P. Baldwin, and S. Solomon, 2005: Stratosphere/troposphere coupling in the Southern Hemisphere. *J. Atmos. Sci*., **62**, doi:10.1175/JAS-3321.1.
Thorne, P.W., et al., 2005: A new realisation of upper-air temperature evolution since the mid-twentieth century and its implications. *J. Geophys. Res*. (accepted).
Tibaldi, S., et al., 1994: Northern and Southern Hemisphere seasonal variability of blocking frequency and predictability. *Mon. Weather Rev*., **122**, 1971–2003.
Trenberth, K.E., 1984: Signal versus noise in the Southern Oscillation. *Mon. Wea. Rev*., **112,** 326–332.
Trenberth, K.E., 1990: Recent observed interdecadal climate changes in the Northern Hemisphere. *Bull. Amer. Meteor. Soc.*, **71**, 988–993.
Trenberth, K.E., 2002: Changes in tropical clouds and radiation. *Science*, **296**, 2095a (online), http://www.sciencemag.org/cgi/content/full/296/5576/2095a.
Trenberth, K.E., 2004: Rural land-use change and climate. *Nature*, **427**, 213.
Trenberth, K.E. and J.W. Hurrell, 1994: Decadal atmosphere–ocean variations in the Pacific. *Climate Dyn.,* **9**, 303–319.
Trenberth, K.E. and J.M. Caron, 2000: The Southern Oscillation revisited: Sea level pressures, surface temperatures and precipitation. *J. Climate*, 13, 4358–4365.
Trenberth, K. and J. Caron, 2001: Estimates of meridional atmosphere and ocean heat transports. *J. Climate*, **14**, 3433–3443.
Trenberth, K.E. and D.J. Shea, 2005: Relationships between precipitation and surface temperature. *Geophys. Res. Lett.*, **32**, L14703, doi:10.1029/2005GL022760.
Trenberth, K.E., and L. Smith, 2005: The mass of the atmosphere: A constraint on global analyses. *J. Climate*, **18**, 864–875.
Trenberth, K.E. and D.P. Stepaniak, 2003a: Co-variability of components of poleward atmospheric energy transports on seasonal and interannual timescales. *J. Climate*, **16**, 3690–3704.
Trenberth, K.E. and D.P. Stepaniak, 2003b: Seamless poleward atmospheric energy transports and implications for the Hadley circulation. *J. Climate*, **16**, 3705–3721.
Trenberth, K.E., D.P. Stepaniak, and J.M. Caron, 2000: The global monsoon as seen through the divergent atmospheric circulation. *J. Climate*, **13**, 3969–3993.
Trenberth, K.E., et al., 2002a: The evolution of ENSO and global atmospheric temperatures. *J. Geophys. Res*., **107**, 4066, doi:10.1029/2000JD000298.

Trenberth, K.E., D.P. Stepaniak, and J.M. Caron, 2002b: Interannual variations in the atmospheric heat budget. *J. Geophys. Res*., **107**(4066), 10.1029/2000JD000297.
Trenberth, K.E., et al., 2003: The changing character of precipitation. *Bull. Amer. Meteor. Soc.*, **84**, 1205–1217.
Trenberth, K.E., J. Fasullo, and L. Smith, 2005a: Trends and variability in column integrated atmospheric water vapor. *Climate Dyn.*, **24**, 741–758. doi:10.1007/s00382-005-0017-4.
Trenberth, K.E., D.P. Stepaniak, and L. Smith, 2005b: Interannual variability of the patterns of atmospheric mass distribution. *J. Climate*, **18**, 2812–2825.
Trepte, S. and P. Winkler, 2004: Reconstruction of erythrmal UV irradiance and dose at Hohenpeissenberg (1968–2001) considering trends of total ozone, cloudiness and turbidity. *Theor. Appl. Climatol*., 77, 159–171, doi:10.1007/s00704-004-00340-y.
Trigo, R.M., et al., 2004: Climate impact of the European winter blocking episodes from the NCEP/NCAR Reanalyses. *Climate Dyn.*, **23**, 17–28.
Trömel, S. and C.-D. Schönwiese, 2005: A generalized method of time series decomposition into significant components including probability assessments of extreme events and application to observed German precipitation data. *Meteorol. Z.*, **14**, 1–11.
Troup, A.J., 1965: The Southern Oscillation. *Quart. J. Roy. Meteor. Soc.*, **91,** 490–506.
Tuller, S.E., 2004: Measured wind speed trends on the west coast of Canada. *Int. J. Climatol*., **24**, 1359–1374.
Tuomenvirta, R.H., et al., 2000: Trends in Nordic and Arctic temperature extremes. *J. Climate*, **13**, 977–990.
Turner, D.D., et al., 2003: Dry bias and variability in Vaisala RS80-H radiosondes: The ARM experience. *J. Atmos. Oceanic Technol.*, **20**, 117–132.
Turner, J., et al., 2005: Antarctic climate change during the last 50 years, *Int. J. Climatol.*, **25**, 279–294.
Tyrrell, J., 2003: A tornado climatology for Ireland. *Atmos. Res*., **67-68**, pp. 671–684.
Ulbrich, U., et al., 2003a: The central European floods of August 2002: Part 1 – Rainfall periods and flood development. *Weather*, **58**, 371–377.
Ulbrich, U., et al., 2003b: The central European floods of August 2002: Part 2 – Synoptic causes and considerations with respect to climatic change. *Weather*, **58**, 434–442.
Uppala, S.M., et al., 2005: The ERA-40 reanalysis. *Quart. J. Roy. Meteor. Soc.*, (in press).
U.S. Geological Survey (USGS), 2004: Climatic Fluctuations, Drought, and Flow in the Colorado River Basin. U.S. Department of the Interior, U.S. Geological Survey, USGS Fact Sheet 2004–3062.
van den Broeke, M.R. and N.P.M. van Lipzig, 2003: Response of wintertime Antarctic temperatures to the Antarctic Oscillation: Results of a regional climate model. In: *Antarctic Peninsula Climate Variability: Historical and Paleoenvironmental Perspectives* [Domack, E., et al. (eds.)]. Antarctic Research Series, **79**, Amer. Geophys. U., Washington, D.C., pp. 43–58.
van den Dool, H., J. Huang, and Y. Fan, 2003: Performance and analysis of the constructed analogue method applied to U.S. soil moisture over 1981–2001. *J. Geophys. Res*., **108**, 8617, doi:10.1029/2002JD003114.
van der Schrier, G., et al., 2005: Summer moisture variability across Europe. *J. Climate* (submitted).
Venegas, S.A. 2003: The Antarctic Circumpolar Wave: A combination of two signals? *J. Climate*, **16**, 2509–2525.
Venegas, S.A. and Mysak, L.A. 2000: Is there a dominant timescale of natural climate variability in the Arctic? *J. Climate*, **13**, 3412–3434.
Vera, C., et al., 2005: A unified view of the American monsoon systems. *J. Climate*, (in press).
Vimont, D.J., D.S. Battisti, and A.C. Hirst, 2001: Footprinting: A seasonal connection between the Tropics and midlatitudes. *Geophys. Res. Lett*., **28**, 3923–2936.
Vincent, L.A. 1998: A technique for the identification of inhomogeneities in Canadian temperature series. *J. Climate* **11**, 1094–1104.
Vincent, L.A. and E. Mekis, 2005, Changes in daily and extreme temperature and precipitation indices for Canada over the 20th century. *Atmos. Ocean*, (submitted).
Vincent, L.A., et al., 2002: Homogenization of daily temperatures over Canada. *J. Climate*, **15**, 1322–1334.
Vincent, L.A., et al., 2005: Observed trends in indices of daily temperature extremes in South America 1960–2000. *J. Climate*, Revised submission.
Vinnikov, K.Y. and N.C. Grody, 2003, Global warming trend of mean tropospheric temperature observed by satellites. *Science*. **302**, 269–272.
Visbeck, M., et al., 2003: The Ocean's Response to North Atlantic Oscillation variability. In: *The North Atlantic Oscillation: Climatic Significance and Environmental Impact* [Hurrell, J.W., et al. (eds.)]. *Geophys. Monogr*, **134**, Amer. Geophys. U., Washington, DC, pp. 113–145.

Vose, R.S., et al., 1992: The Global Historical Climatology Network: Long-term monthly temperature, precipitation, sea level pressure, and station pressure data. ORNL/CDIAC-53, NDP-041, Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, Oak Ridge, TN, U.S., 325 pp.
Vose, R.S., et al., 2004: Impact of land-use change on climate. *Nature*, **427**, 213–214.
Vose, R.S., D.R. Easterling, and B. Gleason, 2005a: Maximum and minimum temperature trends for the globe: An update through 2004. *Geophys. Res. Lett.* (submitted).
Vose, R.S., et al., 2005b: An intercomparison of surface air temperature analyses at the global, hemispheric and grid-box scale. *Geophys. Res. Lett*. (submitted).
Wallace, J.M. and D.S. Gutzler, 1981: Teleconnections in the geopotential height field during the Northern Hemisphere winter. *Mon. Wea. Rev*., **109,** 784–812.
Walter K. and H.-F. Graf, 2002: On the changing nature of the regional connection between the North Atlantic Oscillation and sea surface temperature. *J. Geophys. Res*., **107,** 4338, doi:10.1029/2001JD000850.
Walter, M.T., et al., 2004: Increasing evapotranspiration from the conterminous United States. *J. Hydrometeor.*, **5**, 405–408.
Wang, B., 1994: Climatic regimes of tropical convection and rainfall. *J. Climate*, **7**, 1109–11118.
Wang, B. and Z. Fan, 1999: Choice of South Asian summer monsoon indices**.** *Bull. Amer. Meteor. Sci.*, **80**, 629–638.
Wang B. and J.C.L. Chan, 2002: How strong ENSO events affect tropical storm activity over the western North Pacific. *J. Climate*, **15**, 1643–1658.
Wang, J.H., H.L. Cole, and D.J. Carlson, 2001: Water vapor variability in the tropical western Pacific from 20-year radiosonde data, *Adv. Atmos. Sci.*, **18**, 752–766.
Wang, J.H., et al., 2002a: Corrections of humidity measurement errors from the Vaisala RS80 radiosonde – Application to TOGA COARE data. *J. Atmos. Oceanic Technol.*, **19**, 981–1002.
Wang, P.H., et al., 2002b: Satellite observations of long-term changes in tropical cloud and outgoing longwave radiation from 1985 to 1998. *Geophys. Res. Lett.,* **29**(10), doi:10.1029/2001GL014264.
Wang, J., et al., 2003: Performance of operational radiosonde humidity sensors in direct comparison with a chilled mirror dew-point hygrometer and its climate implication. *Geophys. Res. Lett*. , **30,** (1860), 10.129/2003GL016985.
Wang, J.X.L. and D.J. Gaffen, 2001: Trends in extremes of surface humidity, temperatures and summertime heat stress in China. *Adv. Atmos. Sci.*, **18**, 742–751.
Wang, X. and J.R. Key, 2003: Recent trends in Arctic surface, cloud, and radiation properties from space. *Science*, **299**, 1725–1728.
Wang, X.L. and P.M. Zhai, 2004: Variation of spring dust storms in China and its association with surface winds and sea level pressures. *Acta Meteorologica Sinica*, **62**, 96–103. (in Chinese).
Wang X.L., V.R. Swail, and F.W. Zwiers, 2005: Climatology and changes of extra-tropical storm tracks and cyclone activity: Comparison of ERA-40 with NCEP/NCAR Reanalysis for 1958–2001. *J. Climate*, (submitted).
Wang, Z.W. and P.M. Zhai, 2003: Climate change in drought over northern China during 1950–2000. *Acta Geographica Sinica*, **58**(supplement), 61–68. (in Chinese).
Waple, A.M. and J.H. Lawrimore, 2003: State of the climate in 2002. *Bull. Amer. Meteor. Soc*., **84**, S1–S68.
Waple, A.M., et al., 2002: Climate assessment for 2001. *Bull. Amer. Meteor. Soc*., **83**, S1–S62.
Ward, N.M. and B.J. Hoskins, 1996: Near surface wind over the global ocean 1949–1988. *J. Climate*, **9**, 1877–1895.
Ward, N.M.,1998: Diagnosis and short-lead time prediction of summer rainfall in tropical North Africa at interannual and multidecadal timescales. *J. Climate*, **11**, 3167-3191.
Wardle, R. and I. Smith, 2004: Modeled response of the Australian monsoon to changes in land surface temperatures. *Geophys. Res. Lett.*, **31**, L16205, doi:10.1029/2004GL020157.
Watkins, A., 2002: 2002 Australian climate summary: Dry and warm conditions dominate. *Bulletin of the Australian Meteorological and Oceanographic Society*, **15**, 109–114.
Waugh, D., et al., 1999: Persistence of the lower stratospheric polar vortices. *J. Geophys. Res*., **104**, 27191–27201.
Webster, P.J., and S. Yang, 1992: Monsoon and ENSO: selective interactive systems. *Quart. J. Roy. Meteor. Soc*,. **118**, 877–926.
Webster, P.J., et al., 1998: Monsoons: Processes, predictability, and the prospects for prediction. *J. Geophys. Res*., **103**, 14451–14510.

Webster, P. J., et al., 2005: Changes in tropical cyclone number, duration and intensity in a warming environment. *Science*, accepted.
Wells N., S. Goddard, and M. J. Hayes, 2004: A self-calibrating Palmer Drought Severity Index. *J. Climate*, **17**, 2335–2351.
Wentz, F.J., et al., 2000: Satellite measurements of sea surface temperature through clouds. *Science*, **288**, 847–850.
Wentz, F.J. and M. Schabel, 1998: Effects of satellite orbital decay on MSU lower tropospheric temperature trends. *Nature*, **394**, 661–664.
Wentz, F.J. and R.W. Spencer, 1998: SSM/I rain retrievals within a unified all-weather ocean algorithm, *J. Atmos. Sci*., **55**, 1613–1627.
Wettstein, J.J. and L.O. Mearns, 2002: The influence of the North Atlantic-Arctic Oscillation on mean, variance, and extremes of temperature in the northeastern United States and Canada. *J. Climate*, **15,** 3586–3600.
Wheeler, M C. and J.L. McBride, 2005: Australian-Indonesian monsoon. In: *Intraseasonal Variability of the Atmosphere-Ocean Climate System* [Lau, W.K.M. and D.E. Waliser, (eds.)], Praxis Publishing, Chichester U.K., pp. 125–173.
White, W.B., 2000: Influence of the Antarctic Circumpolar Wave on Australian precipitation from 1958 to 1997. *J. Climate*, **13**, 2125–2141.
White, W.B. and R.G. Peterson, 1996: An Antarctic Circumpolar Wave in surface pressure, wind, temperature and sea-ice extent. *Nature*, **380**, 699–702.
White, W.B. and N.J. Cherry, 1999: Influence of the Antarctic Circumpolar Wave upon New Zealand temperature and precipitation during Autumn–Winter. *J. Climate*, **12**, 960–976.
White, W.B. and J. Annis, 2004: Influence of the Antarctic Circumpolar Wave on El Niño and its multidecadal changes from 1950 to 2001. *J. Geophys. Res.*, **109**, C06019, doi:10.1029/2002JC001666.
White, W.B., P. Gloersen, and I. Simmonds, 2004: Tropospheric response in the Antarctic Circumpolar Wave along the sea ice edge around Antarctica. *J. Climate* **17**, 2765–2779.
Wibig, J. and B. Glowicki, 2002: Trends of minimum and maximum temperature in Poland. *Climate Res.,* **20**, 123–133.
Wielicki, B.A., et al., 2002a: Evidence for large decadal variability in the tropical mean radiative energy budget. *Science*, **295**, 841–844.
Wielicki, B.A., et al., 2002b: Response. *Science*, **296**, http://www/sciencemag.org/cgi/content/full/296/5576/2095a.
Wielicki, B.A., et al., 2005: Change in Earth's albedo measured by satellite. *Science*, **308**, 825.
Wiedenmann, J.M., et al., 2002: The climatology of blocking anticyclones for the Northern and Southern Hemispheres: Block intensity as a diagnostic. *J. Climate*, **15,** 3459–3473.
Wigley, T.M.L., P.D. Jones, and S.C.B. Raper, 1997: The observed global warming record: What does it tell us?, *Proc. Natl. Acad. Sci*., **94**, 8314–8320.
Wijngaard, J.B., Klein Tank, A.M.G., and Können, G.P., 2003: Homogeneity of 20th century European daily temperature and precipitation series. *Int. J. Climatol*, **23**, 679–692.
Wild, M., et al., 2004: On the consistency of trends in radiation and temperature records and implications for the global hydrological cycle. *Geophys. Res. Lett*., **31**, L11201, doi:10.1029/2003GL019188.
Wild, M.A., et al., 2005: From dimming to brightening: Decadal changes in solar radiation at Earth's surface, *Science*, **308**, 847–850.
Willis, J.K., D. Roemmich, and B. Cornuelle, 2004: Interannual variability in upper-ocean heat content, temperature and thermosteric expansion on global scales. *J. Geophys. Res.*, **109**, C12036, doi:10.1029/2003JC002260.
Wittman, M.A.H., et al., 2004: Stratospheric influence on baroclinic lifecycles and its connection to the Arctic Oscillation. *Geophys. Res. Lett*., **31**, L16113, doi:10.1029/2004GL020503.
WMO, 2003: *World Meteorological Organization statement on the status of global climate in 2003*. WMO publications, Geneva, 12 pp.
Wong, T.D.F., M.H. Young, and S. Weckmann, 2000: Validation of the CERES/TRMM ERBE-like monthly mean clear-sky longwave dataset and the effects of the 1998 ENSO event. *J. Climate*, **13**, 4256–4267.
Wong, T., et al., 2005: Re-examination of the observed decadal variability of Earth Radiation Budget using altitude-corrected ERBE/ERBS nonscanner WFOV data. *J. Climate, submitted*.
Woodruff, S.D., et al., 1998: COADS Release 2 data and metadata enhancements for improvements of marine surface flux fields. *Phys. Chem. Earth*, **23**, 517–527.

Woolf, D.K., P.G. Challenor, and P.D. Cotton, 2002: The variability and predictability of North Atlantic wave climate. *J. Geophys. Res*., **107**, 3145, doi:10.1029/2001JC001124.
Worley, S.J., et al., 2005: ICOADS release 2.1 data and products. *Int. J. Climatol*., **25**, 823–842.
Wu, M.C., W.L. Chang, and W.M. Leung, 2004: Impacts of El Niño-Southern Oscillation events on tropical cyclone landfalling activity in the western North Pacific. *J. Climate*, **17**, 1419–1428.
Wylie D.P. and W.P. Menzel, 1999: Eight years of high cloud statistics using HIRS. *J. Climate*, **12**, 170–184.
Wylie, D., et al., 2005: Trends in global cloud cover in two decades of HIRS observations. *J. Climate* (accepted).
Xie, P. and Arkin, P.A. 1997: Global precipitation: A 17-year monthly analysis based on gauge observations, satellite estimates and numerical model outputs. *Bull. Amer. Meteor. Soc.*, **78**, 2539–2558.
Yan, Z, et al., 2002: Trends of extreme temperatures in Europe and China based on daily observations. *Clim Change*, **53**, 355–392.
Yang, D, 1999: An improved precipitation climatology for the arctic ocean. *Geophys. Res. Lett.,* **26**, 1625–1628.
Yang, D. and T. Ohata, 2001: A bias corrected Siberian regional precipitation climatology. *J. Hydrometeorol*., **2**, 122–139.
Yang, D., et al., 1999: Bias correction of precipitation data for Greenland. *J. Geophys. Res.-Atmospheres*, **105**, 6171–6182.
Yang, D., et al., 2002: Siberian Lean River hydrologic regime and recent change. *J. Geophys. Res.*, **107**, 4694, doi:10.1029/2002JD002542.
Yang, D., B. Ye, and D. L. Kane, 2004: Streamflow changes over Siberian Yenisei River Basin. *J. Hydrol.*, **296,** 59–80.
Ye, H. and M. Ellison, 2003: Changes in transitional snowfall season length in northern Eurasia, *Geophys. Res. Lett.*, **30**,(5), 1252, doi:10.1029/2003GL016873.
Ye, B.S., D.Q. Yang, and D.L. Kane, 2003: Changes in Lena River streamflow hydrology: Human impacts versus natural variations. *Water Resour. Res.*, **39,** 1200, doi:10.1029/2003WR001991.
Yin, X., A. Gruber, and P. Arkin, 2004: Comparison of the GPCP and CMAP merged gauge-satellite monthly precipitation products for the period 1979–2001. *J. Hydrometeor*, **5**, 1207–1222.
Yu, L.S. and Rienecker, M.M. 1999: Mechanisms for the Indian Ocean warming during the 1997–1998 El Niño. *Geophysical Res. Lett*., **26**, 735–738.
Yu, R., B. Wang, and T. Zhou, 2004: Climate effects of the deep continental stratus clouds generated by the Tibetan Plateau. *J. Climate*, **17**, 2702–2713.
Yuan, X. and D.G. Martinson, 2001: The Antarctic Dipole and its predictability. *Geophys. Res. Lett.*, **28,** 3609–3612.
Zhai, P.M. and Pan X.H., 2003: Trends in temperature extremes during 1951–1999 in China. *Geophys. Res. Lett.*, **30**, 1913, doi:10.1029/203GL018004.
Zhai, P.M., et al., 2004: Trends in total precipitation and frequency of daily precipitation extremes over China. *J. Climate*, **18**, 1096–1108.
Zhang, X., W.D. Hogg, and E. Mekis, 2001a: Spatial and temporal characteristics of heavy precipitation events over Canada. *J. Climate*, **14**, 1923–1936.
Zhang, X.B., et al., 2001b: Trends in Canadian streamflow. *Water Resour. Res.*, **37,** 987–998.
Zhang , X., et al., 2004a: Climatology and interannual variability of Arctic cyclone activity: 1948–2002. *J. Climate*, **17**, 2300–2317.
Zhang, X., F.W. Zwiers, and G. Li, 2004b: Monte Carlo experiments on the detection of trends in extreme values. *J. Climate*, **17**, 1945–1952.
Zhang, Y., et al., 2004c: Calculation of radiative fluxes from the surface to top of atmosphere based on ISCCP and other global data sets: refinements of the radiative transfer model and the input data. *J. Geophys. Res*., **109**, D19105, doi:10.1029/2003JD004457.
Zhang, X., et al., 2005: Trends in Middle East climate extremes indices during 1930–2003. *J. Geophys. Res.*, Submitted.
Zhou, Z. J. and G. C. Zhang, 2003: Typical severe dust storms in northern China (1954–2002). *Chinese Science Bulletin*, ***48***, 1224–1228. (in Chinese.)
Zhou, S., et al., 2000: An inter-hemisphere comparison of the persistent stratospheric polar vortex, *Geophys. Res. Lett*., **27**, 1123–1126.
Zhou, X.L., M.A. Geller, and M.H. Zhang, 2001: Cooling trend of the tropical cold point tropopause temperatures and its implications, *J. Geophys. Res*., **106**, 1511–1522.

Zou, X.K. and P.M. Zhai, 2004: Relationship between vegetation coverage and spring dust storms over northern China. *J. Geophys. Res.*, **109**, D03104, doi:10.1029/2003JD003913.
Zou X.K., P.M. Zhai, and Q. Zhang, 2005: Variations in droughts over China: 1951–2003. *Geophys. Res. Lett.*, **32,** L04707, doi:10.1029/2004GL021853.
Zveryaev, I.I. and P.S. Chu , 2003: Recent climate changes in precipitable water in the global tropics as revealed in National Centers for Environmental Prediction/National Center for Atmospheric Research reanalysis. *J. Geophys. Res.* **108**, 4311, doi:10.129/2002JD002476.

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

3.A.1 Methods of Temperature Analysis

3.A.1.1 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., 2005; Smith and Reynolds, 2005; Brohan et al., 2005). 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. Absolute 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., 2005; Brohan et al., 2005).

The effects of changes in coverage over the instrumental period (since 1861 for global scales) were first assessed by the “frozen grid” approach (see Jones et al., 1997, 1999). Subsequently, reduced-space optimal interpolation (RSOI) has been used to complete incomplete and noisy fields and to provide local error estimates (Kaplan et al., 1997; Rayner et al., 2003; Brohan et al., 2005). Reduced space optimal averaging (OA) yields large-area averages with error-bars (Folland et al., 2001; Brohan et al., 2005). 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 or 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.

In addition to errors from changing coverage and from random measurement and sampling errors, errors arise from biases (see Section 3.A.2 on homogenizing records). Major efforts have been made to adjust for known systematic biases, but some adjustments nonetheless are quite uncertain, and such errors are not included in the assigned error bars. For example, for SSTs, the transition from taking temperatures from water samples from uninsulated or partially insulated buckets to engine intakes is adjusted for but details are quite uncertain near the time of major change during World War II.

3.A.1.2 Linear trend estimations

Some of the linear trends for global fields and averages in this Chapter have been estimated using Restricted Maximum Likelihood (REML, Diggle et al., 1999). REML estimates yields error bars which take account both of the estimated errors of the input data and of the autocorrelation of the residuals from the fitted trend. The error bars are, therefore, wider and more realistic than those provided by the Ordinary Least Squares (OLS) technique. If, for example, a century-long series has multidecadal 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 period of record when used, for example, to estimate trends from MSU data, which began in 1979. So, the errors estimated by REML may still be too small for short records.

3.A.2 Adjustments to Homogenize Land Temperature Observations

Long-term temperature data from individual climate stations almost always suffers from inhomogeneities, owing to non-climatic factors such as changes in station location, instrumentation, observer methods, or the environment surrounding the station. There are two kinds of inhomogeneities that impact temperature data, gradual changes due to urban development, land-use changes, and instrument drift; and discontinuities due to abrupt changes such as a station relocation. Most abrupt changes tend to produce random effects on regional and global trends. 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. Urbanization usually produces warming, although examples exist of cooling in arid areas where irrigation effects dominate.

When dates for discontinuities are known, the data for a target station are compared 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) (e.g., Easterling and Peterson, 1995; Vincent, 1998; Menne and Williams, 2005). An extensive review is given by Aguilar et al. (2003).

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). However, 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).

Adjustments for urban impacts on temperature data have been limited to approaches such as linear regression against population (Karl et al., 1988). Urbanization impacts on global and hemispheric temperature trends (Karl et al., 1988; Jones et al., 1990; Easterling et al., 1997; Peterson, 2003; Parker, 2004) have been found to be small. Furthermore, once the landscape around a station becomes urbanized, long-term trends for that station are consistent with near-by rural stations (Böhm, 1998; Easterling et al., 2005, Peterson and Owen, 2005).

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. Assessment of potential homogeneity problems in a network of 60 daily temperature series (maxima and minima), for Europe for the 20th century by Wijngaard et al. (2003), suggests than 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.A.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., 2005) 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 K 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 (Folland et al., 2001), 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, 2005) and between overall ship and buoy observations (Rayner et al., 2005). 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, 2005; Kent and Kaplan, 2005), which may have caused an underestimate of recent warming (Rayner et al., 2005). 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 an absolute 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 Along Track Scanning Radiometer (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 1979–2004 from *in situ* data (0.14 K 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., 2005). 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 accordingly (Rayner et al., 2005). Owing to biases arising from solar heating of ships' fabric, marine air temperature analyses have so far been based on nighttime 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 the 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 (Diaz et al., 2002), 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.A.4 Solid/Liquid Precipitation: Undercatch and Adjustments for Homogeneity

3.A.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 that 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.A.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 beyond 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.

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

3.A.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. 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 outlier checks (e.g., Durre et al., 2005). 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. (2005), in contrast, created a global database, HadAT, 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. Moisture data from radiosondes generally contain even more complex problems, and no climate quality homogenised databases have yet been produced.

3.A.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. 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.A.5.3 Reanalysis and climate trends

Comprehensive reanalyses from NCEP/NCAR (Kalnay et al., 1996; Kistler et al., 2001), NCEP-2 reanalysis (Kanamitsu et al., 2002) and ERA15/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 hour 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). 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.A.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.A.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 1979 (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 (Christy et al., 2003). 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) are 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 (~75%) 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 K 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), 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.