Climate Change Science Program
Temperature Trends in the Lower Atmosphere
Understanding and Reconciling Differences

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1   **PREFACE**

2   **Report Motivation and Guidance for Using this Synthesis/Assessment Report**

3

4   A primary objective of the U.S. Climate Change Science Program (CCSP) is to

5   provide the best possible scientific information to support public discussion and

6   government and private sector decision-making on key climate-related issues. To help

7   meet this objective, the CCSP has identified an initial set of 21 synthesis and

8   assessment products that address its highest priority research, observation, and

9   decision-support needs. This Synthesis/Assessment Report, the first of the 21

10   Reports, focuses on understanding the causes of the reported differences between

11   independently produced data sets of atmospheric temperature trends from the surface

12   through the lower stratosphere.

13

14   **Background**

15

16   Measurements of global surface air temperature show substantial increases over the

17   past several decades. In the early 1990s, data from NOAA's polar orbiting satellites

18   were analyzed for multi-decadal trends. These initial analyses indicated that

19   temperatures in the troposphere showed little or no increase, in contrast with surface

20   air measurements from ships, land-based weather stations, and ocean buoys. This

21   result led some to question the reality and/or the cause of the surface temperature

22   increase, on the basis that human influences, thought to be important contributors to

23   observed change, were expected to increase temperatures both at the surface and in

24   the troposphere with larger increases expected in the tropical troposphere. This

25   surprising result led to an intensive effort by climate scientists to better understand

26   the causes of the apparent differences in the rates of temperature increase between the

27   surface and the troposphere.

28

29   Scientists analyzing the data knew that there were complex and unresolved issues

30   related to inadequacies of observing systems that could lead to misinterpretation of

31   the data. There were also uncertainties in our understanding of how the climate might

32   respond to various forcings as is often assessed through the use of climate models. In

33   an attempt to resolve these issues, in 2000 the National Research Council specifically

34   addressed the general issue of troposphere and surface derived temperature trends. In

35   its Report, the NRC concluded that "the warming trend in global-mean surface

36   temperature observations during the past 20 years is undoubtedly real and is

37   substantially greater than the average rate of warming during the twentieth century.

38   The disparity between surface and upper air trends in no way invalidates the

39   conclusion that surface temperature has been rising." The NRC further found that

40   corrections in the Microwave Sounding Unit (MSU) processing algorithms brought

41   the satellite data record into slightly closer alignment with surface temperature trends.

42   They concluded that the substantial disparity that remains probably reflects a less

43   rapid warming of the troposphere than the surface in recent decades due to both

44   natural and human-induced causes.

45

46   In 2001, the Intergovernmental Panel on Climate Change (IPCC) Third Assessment

47   Report devoted additional attention to new analyses of the satellite, weather balloon,

48   and surface data to evaluate the difference in temperature trends between the surface

49   and the troposphere. Similar to the NRC, the IPCC concluded that it was very likely

50   that the surface temperature increases were larger and differed significantly from

51   temperature increases higher in the troposphere. They concluded, "during the past

52   two decades, the surface, most of the troposphere, and the stratosphere have

53   responded differently to climate forcings because different physical processes have

54   dominated in each of the regions during that time." (IPCC; Climate Change 2001 The

55   Scientific Basis, Chapter 2, p. 122-123; Cambridge University Press).

56

57   **Focus of this Synthesis/Assessment Report**

58

59   The efforts of the NRC and IPCC to address uncertainties about the temperature

60   structure of the lower atmosphere (i.e., from the surface through the lower

61   stratosphere) have helped move us closer to a comprehensive understanding of

62   observed trends of temperature. Although these documents provided a great deal of

63   useful information, full resolution of the issue was hampered by the complexities

64   coupled with shortcomings of the available observing systems. To more fully address

65   remaining fundamental questions, a broader examination has been undertaken here to

66   answer the following questions:

67   1)   Why do temperatures vary vertically (from the surface to the stratosphere)
68   and what do we understand about why they might vary and change over
69   time?
70

71   2)   What kinds of atmospheric temperature variations can the current
72   observing systems measure and what are their strengths and limitations,
73   both spatially and temporally?
74
75   3)   What do observations indicate about the changes of temperature in the
76   atmosphere and at the surface since the advent of measuring temperatures
77   vertically?
78
79   4)   What is our understanding of the contribution made by observational or
80   methodological uncertainties to the previously reported vertical differences
81   in temperature trends?
82
83   5)   How well can the observed vertical temperature changes be reconciled with
84   our understanding of the causes of these changes?
85
86   6)   What measures can be taken to improve the understanding of observed
87   changes?
88
89   These questions provide the basis for the six main chapters in this

90   Synthesis/Assessment Report (the chapter numbers correspond to the question

91   numbers above). They highlight several of the fundamental uncertainties and

92   differences between and within the individual components of the existing

93   observational and modeling systems. The responses to the questions are written in a

94   style consistent with major international scientific assessments (e.g., IPCC

95   assessments, and the Global Ozone Research and Monitoring Project of the World

96   Meteorological Organization). The Executive Summary, which presents the key

97   findings from the main body of the Report, is intended to be useful for those involved

98   with the policy-related global climate change issues. The Chapters supporting the

99   Executive Summary are written at a more technical level suitable for non-climate

100   specialists within the scientific community and well-informed lay audiences.

101

102   To help answer the questions posed, climate model simulations of temperature change

103   based on time histories of the forcing factors thought to be important, have been

104   compared with observed temperature changes. If the models replicate the observed

105   temperature changes, this increases confidence in our understanding of the observed

106   temperature record and reduces uncertainties about projected changes. If not, then this

107   implies that the time histories of the important forcings are not adequately known, all

108   of the important forcings are not included, the processes being simulated in the

109   models have serious flaws, the observational record is incorrect, or some combination

110   of these factors.

111

112   This U.S. Climate Change Science Program Assessment/Synthesis Report assesses

113   the uncertainties associated with the data used to determine changes of temperature,

114   and whether such changes are consistent with our understanding of climate processes.

115   This requires a detailed comparison of observations and climate models used to

116   simulate observed changes, including an appreciation of why temperatures might

117   respond differently at the surface compared to higher levels in the atmosphere.

118

119   This CCSP Report addresses the accuracy and consistency of the temperature records

120   and outlines steps necessary to reconcile differences between individual data sets.

121   Understanding exactly how and why there are differences in temperature trends

122   reported by several analysis teams using different observation systems and analysis

123   methods is a necessary step in reducing the uncertainties that underlie current efforts

124   focused on the detection and quantification of surface and tropospheric temperature

125   trends.

126

127   **New observations and analysis since the IPCC and NRC Reports**

128

129   Since the IPCC and NRC assessments, there have been intensive efforts to create new

130   satellite and weather balloon data sets using a range of approaches. Having multiple

131   satellite data sets provides the opportunity for much greater understanding of

132   observed changes and their uncertainty than was possible in the previous assessments.

133   In addition, for the first time a suite of models simulating observed climate since

134   1979 (when satellite data began) has provided us a unique opportunity to inter-

135   compare observed trends from various data sets with model simulations using various

136   scenarios of historical climate forcings. Taken together, these advances lead to a

137   much greater understanding of the issues.

138

139   The science of upper air temperature issues is a rapidly evolving field. During the

140   preparation of this Report, new findings were published and have been included in the

141   current draft, causing numerous changes from draft to draft. The authors certainly

142   expect that new data and discoveries that follow the release of this Report, will

143   further improve our understanding. Some open questions originally discussed in the

144   first drafts of this Report were actually resolved during the deliberations. For

145   example, a recent article cleverly demonstrated a subtle problem in the method used

146   in one of the data sets to correct for satellite orbital drift. Since it was possible for the

147   error to be rectified fairly quickly, a new satellite-derived version of lower

148   tropospheric temperatures was available for this Report. At the same time, another

149   research team produced their first version of satellite-derived lower troposphere

150   temperature, and yet another team updated their tropospheric temperature time series

151   as the final drafts were written. All these results are included in this Report.

152

153   Factors that guided the authors in the selection of the climate records considered

154   extensively in this Report were (a) publication heritage, (b) public availability, (c) use

155   by the community at-large, (d) updated on a monthly basis, and (e) period of record

156   beginning in 1979 or earlier. The three surface analyses that were used have many

157   publications covering their construction methods. These data sets are readily

158   available, and are widely used. Two of the three satellite data sets used, while

159   relatively recent, are based on a heritage of published versions which have

160   incorporated new adjustments as discoveries have been made. Each of these data sets

161   allows ready access to the public and has been used in several research publications.

162   A third, more recently developed, data set has been updated during the preparation of

163   this Report. Two data sets used were based on weather balloon data. One of these

164   data sets publicly appeared in 2005, but the authors had made the preliminary

165   versions and methodology available to scientists as early as 2002 and have built upon

166   the extensive experience acquired from previous versions of these data sets. Another

167   data set has a heritage dating back several decades and was recently updated.

168

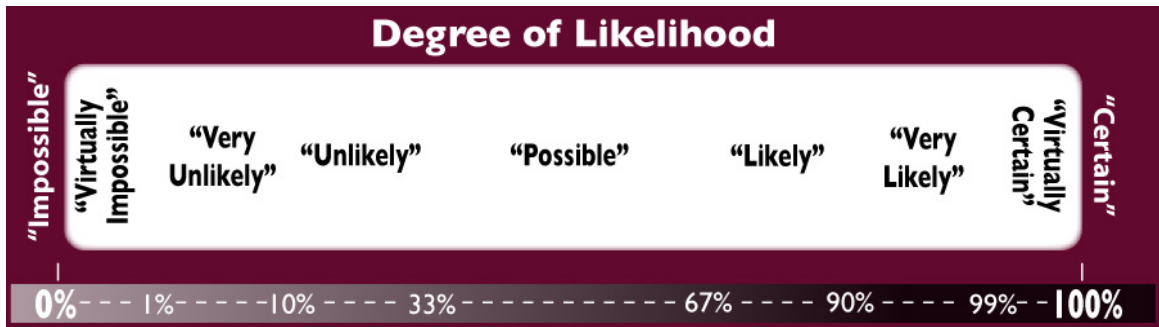169   **How to use this Synthesis/Assessment Report**

170

171   This Report promises to be of significant value to decision-makers, and to the expert

172   scientific and stakeholder communities. Readers of this Report will find that new

173   observations, data sets, analyses, and climate model simulations enabled the Author

174   Team to resolve many of the perplexities noted by the NRC and the IPCC in their

175   earlier Reports. The Synthesis/Assessment Report already has had an important

176   impact on the content of the draft to the Fourth Assessment Report of the

177   Intergovernmental Panel on Climate Change (IPCC), due to be published in 2007. In

178   addition, we expect the information generated here will be used both nationally and

179   internationally e.g., by the Global Climate Observing System Atmospheric

180   Observation Panel to help identify effective ways to reduce observational uncertainty.

181   The findings regarding observations and model-observation comparisons of lower

182   stratospheric temperature trends will be useful for the 2006 WMO/UNEP Ozone

183   Assessment.

184

185   Some terms used in the Report may be unfamiliar to those without training in

186   meteorology; a glossary and list of acronyms is thus included at the end of the Report.

187   Two sets of terms are useful to define at the outset since they are particularly

188   fundamental to this Report. This includes a set of terms related to various levels of

189   agreement or disagreement on key issues and findings among the expert Lead

190   Authors as well as terminology describing their considered judgment about the

191   likelihood of critical key results.

192

193

194   To integrate a wide variety of information, this Report also uses a lexicon of terms to

195   express the team's considered judgment about the likelihood of results. Confidence in

196   results is highest at each end of the spectrum. Unless otherwise noted, all statements

197   are certain.

199   Preface figure 1

200   This illustration shows the layers of the atmosphere of primary interest to this

201   Synthesis/Assessment Report. The multi-colored line on this diagram indicates the

202   variations in temperature with altitude. The chart beneath the diagram defines the

203   terminology used in this Report for the layers of the atmosphere.
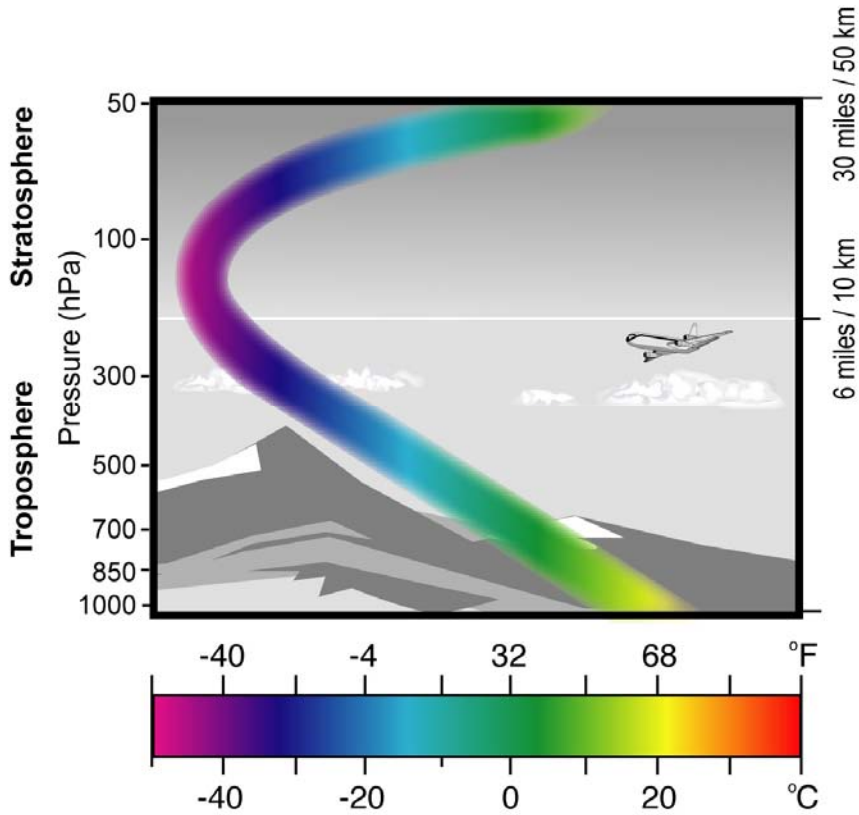
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205   Preface figure 2
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224   **Terms for Layers of the Atmosphere Used in this Report**
225   Preface table 1

Common Term	Abbrev. Term for the temperature of that layer	Main region of Influence	Approximate altitude. (For satellite products: altitude range of bulk (90%) of layer measured.)	Lower and upper pressure level boundaries
Surface	T_S	<u>Air</u>: Just above surface <u>Water</u>: Shallow depth	<u>Surface Air</u>: Land: 1.5 –2.0 m above surface; Ocean: ship deck-height (5 – 25 m) above surface. <u>Surface Water</u>: 1 – 10 m depth in ocean (SSTs)	Surface (or ~1000 hPa at sea level)
Lower Troposphere	T_{2LT}	Low to Mid-Troposphere	Surface – 8 km	Surface to 350 hPa
Troposphere (radiosonde)	$T_{(850\text{-}300)}$	Troposphere	1.5 – 9 km	850 – 300 hPa
Troposphere (satellite)	T^*_G	Troposphere	Surface – 13 km	Surface – 150 hPa
Tropical Troposphere (satellite)	T^*_T	Troposphere (tropics only)	Surface – 16 km	Surface – 100 hPa
Mid Troposphere to Lower Stratosphere	T_2	Mid and upper Troposphere to Low Stratosphere[1]	Surface – 18 km	Surface – 75 hPa
Lower Stratosphere (satellite)	T_4	Low Stratosphere	14 – 29 km	150 – 15 hPa
Lower Stratosphere (radiosonde)	$T_{(100\text{-}50)}$	Low Stratosphere	17 – 21 km	100 – 50 hPa

226
227   *Note:* Abbreviated terms --- Subscript 'S', refers to the <u>S</u>urface. Subscripts '2' and '4'
228   refer to MSU data from channels 2 and 4. Subscript '2LT' refers to a modification of
229   channel 2 data to focus more directly on the <u>L</u>ower <u>T</u>roposphere and reduce the influence
230   of stratospheric temperatures on channel 2 data. Subscripts '850–300' and '100–50' are
231   specific atmospheric layers sampled by radiosondes. Subscript '*G' refers to a
232   combination of channel 2 and channel 4 data derived by Fu and co-workers, applicable to

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[1] Only about 10% of this layer extends into the lower stratosphere

233 global averages, and '$*_T$' refers to applicable tropical averages. For the model-
234 observation comparisons, the observation-based definitions as listed in the Table
235 were employed.
236
237
238

239   **The Authoring Team**

240   A full list of this Reports' authoring team (in addition to a list of lead authors

241   provided at the beginning of each Chapter) is provided in an Appendix at the end of

242   this Report. The focus of this Report follows the Prospectus developed by the Climate

243   Change Science Program and posted on its website at http://www.climatescience.gov.

244

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6   **EXECUTIVE SUMMARY**
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14
15
16   Convening Lead Author: Tom M.L. Wigley
17
18   Lead Authors: V. Ramaswamy, J.R. Christy, J.R. Lanzante, C.A. Mears, B.D. Santer,
19   and C.K. Folland
20
21

22   ***<u>New Results and Findings</u>***

23

24   This Report is concerned with temperature changes in the atmosphere, differences in these

25   changes at various levels in the atmosphere, and our understanding of the causes of these

26   changes and differences. Considerable progress has been made since the production of reports by

27   the National Research Council (NRC) and the Intergovernmental Panel on Climate Change

28   (IPCC) in 2000 and 2001. Data sets for the surface and from satellites and radiosondes

29   (temperature sensors on weather balloons) have been extended and improved, and new satellite

30   and radiosonde data sets have been developed[1]. Many new model simulations of the climate of

31   the 20th century have been carried out using improved climate models[1] and better estimates of

32   past forcing changes, and numerous new and updated model/observed data comparisons have

33   been performed. The present Report reviews this progress. A summary of the main results is

34   presented first. Then, to address the issues in more detail, six questions that provide the basis for

35   the six main chapters in this Synthesis/Assessment Report are posed and answered.

36

37   The important new results presented in this Report include:
38
39
40   <u>Global Average Temperatures</u>
41
42   •   Since the late 1950s, the start of the study period for this Report, the surface and troposphere

43   have warmed[2] substantially, while the stratosphere has cooled[2]. These changes are in accord

44   with our understanding of the effects of radiative forcing agents and with model predictions.

45

46   •   Since the late 1950s, the low and mid troposphere have warmed at a rate slightly faster than

47   the rate of warming at the surface.

48

49   •   During the satellite era (1979 onwards), both the low and mid troposphere have warmed. The

50   majority of data sets show warming at the surface that is greater than in the troposphere.

51   Some data sets, however, show the opposite – tropospheric warming that is greater than that

52   at the surface.

53

54   •   For global-mean temperature changes in the new climate model simulations, some show

55   more warming in the troposphere than at the surface, while a slightly smaller number of

56   simulations show the opposite behavior. Given the range of observed results and the range of

57   model results, there is no inconsistency between models and observations at the global scale.

58

59   •   Studies to detect climate change and attribute its causes using patterns of observed

60   temperature change in space and time (rather than global averages) show clear evidence of

61   human influences on the climate system (due to changes in greenhouse gases, aerosols, and

62   stratospheric ozone).

63

64   •   The observed patterns of change cannot be explained by natural processes alone, nor by the

65   effects of short-live species (such as aerosols and tropospheric ozone) alone.

66

67   <u>Tropical Temperatures (20°S to 20°N)</u>

68   •   The majority of observed data sets show more warming at the surface than in the

69   troposphere, while some newer observed data sets show the opposite behavior. Almost all

70   model simulations show more warming in the troposphere than at the surface.

71

72   These results characterize important changes in our understanding of the details of temperature

73   changes at the surface and higher in the troposphere. In 2000 and 2001, the NRC and the IPCC

74   both concluded that global-mean surface temperature increases were larger and differed

75   significantly from temperature increases in the troposphere. The new and improved observed

76   data sets and new model simulations that have been developed require modifications of these

77   conclusions.

78

79   The crucial issue here is whether changes in the troposphere are greater or less than those at the

80   surface. Greater changes in the troposphere would mean that changes there are "amplified"

81   relative those at the surface. We use the short-hand notation "amplification" to refer to this

82   possibility. Studies of amplification in the tropics have considered changes on month-to-month,

83   year-to-year and decade-to-decade time scales.

84

85   At the global-mean level, observed changes from 1958 through 2004 exhibit amplification: i.e.,

86   they show greater warming trends in the troposphere compared with the surface. Since 1979,

87   however, the situation is different: most data sets show slightly greater warming at the surface.

88

89   Whether or not these results are in accord with expectations based on climate models is a

90   complex issue, one that we have been able to address more comprehensively now using new

91   model results. Over the period since 1979, the range of recent model simulations is almost

92   evenly divided among those that show greater global-mean warming at the surface and others

93   that show greater warming aloft. Given this range of results, there is no conflict between

94   observed changes and the results from climate models.

95

96   In the tropics, the agreement between models and observations depends on the time scale

97   considered. For month-to-month and year-to-year variations, models and observations both show

98   amplification (i.e., the month-to-month and year-to-year variations are larger aloft than at the

99   surface). The magnitude of this amplification is essentially the same in models and observations.

100   On decadal and longer time scales, however, while almost all model simulations show greater

101   warming aloft, most observations show greater warming at the surface.

102

103   These results have at least two possible explanations, which are not mutually exclusive. Either

104   amplification effects on short and long time scales are controlled by different physical

105   mechanisms, and models fail to capture such behavior; and/or remaining errors in some of the

106   observed tropospheric data sets adversely affect their long-term temperature trends. The second

107   explanation is judged more likely.

108

109   **1.** *How do we expect vertical temperature profiles to change?*

110   This Section considers the first question:

111   *Why do temperatures vary vertically (from the surface to the stratosphere) and what do we*
112   *understand about why they might vary and change over time?*
113
114   This is addressed in both Chapter 1 and Chapter 5 of this Report.

115

116   In response to this question, Chapter 1 notes the following …

117

118   (1) Temperatures vary vertically.

119   The effects of solar heating of the surface of the planet combined with the physical properties of

120   the overlying air, lead to the highest temperatures, on average, occurring at the surface. Surface

121   heat is mixed vertically and horizontally, and this mixing, combined with the effects of various

122   physical processes, produces a decrease of temperature with height up to the tropopause

123   (marking the top of the troposphere, i.e., the lower 8 to 16 km of the atmosphere, depending on

124   latitude). Above this, the radiative properties of the air produce a warming with height through

125   the stratosphere (up to about 50 km).

126

127   (2) Temperature trends at the surface can be expected to be different from temperature trends

128   higher in the atmosphere because:

129   • The physical properties of the surface vary substantially according to   location and

130   this produces strong horizontal variations in near-surface temperature. Above the

131   surface, these contrasts are quickly smoothed out so the patterns of change in the

132   troposphere must differ from those at the surface. Temperature trend variations with

133   height must, therefore, vary according to location.

134   • Changes in atmospheric circulation or modes of atmospheric variability (e.g., the El

135   Niño-Southern Oscillation [ENSO]) can produce different temperature trends at the

136   surface and aloft.

137   • Under some circumstances, temperatures may increase with height near the surface or

138   higher in the troposphere, producing a "temperature inversion." Such inversions are

139   more common at night, in winter over continents, and in the trade wind regions. Since

140   the air in inversion layers is resistant to vertical mixing, temperature trends can differ

141   between inversion layers and adjacent layers.

142   •   Forcing factors, either natural or human-induced[3], can result in differing temperature

143   trends at different levels in the atmosphere, and these vertical variations may change

144   over time.

145

146   As noted above, temperatures in the atmosphere vary naturally as a result of internal factors and

147   natural and human-induced perturbations ("forcings"[3]). These factors are expected to have

148   different effects on temperatures near the surface, in the troposphere, and in the stratosphere, as

149   summarized in Table 1. When all forcings are considered, we expect the troposphere to have

150   warmed and the stratosphere to have cooled since the late 1950s (and over the whole 20th

151   century). The relative changes in the troposphere and stratosphere provide information about the

152   causes of observed changes.

153

154 Table 1: Summary of the most important global-scale climate forcing factors and their likely individual effects on
155 global, annual-mean temperatures; based on Figure 1.3 (which gives temperature information) and Table 1.1 (which
156 gives information on radiative forcing) in Chapter 1, and literature cited in Chapter 1. The stated effects are those
157 that would be expected if the change specified in column 1 were to occur. The top two rows are the primary natural
158 forcing factors, while the other rows summarize the main human-induced forcing factors. The relative importance of
159 these different factors varies spatially and over time.
160
161

Forcing Factor	Surface	Low to Mid Troposphere	Stratosphere
Increased solar output	Warming	Warming	Warming
Volcanic eruptions	Cooling	Cooling	Short-term warming
Increased concentrations of well-mixed greenhouse gases (CO_2, CH_4, N_2O, halocarbons)	Warming	Warming	Cooling
Increased tropospheric ozone (O_3)	Warming	Warming	Slight cooling
Decreased stratospheric ozone	Negligible except at high latitudes	Slight cooling	Cooling
Increased loading of sulfate (SO_4) aerosol – sum of direct plus indirect effects	Cooling	Cooling	Negligible
Increased loading of carbonaceous aerosol (black carbon (BC) and organic matter (OM)) – sum of direct plus indirect effects	Regional cooling – possible global-mean cooling	Warming	Uncertain
Land use and land cover changes	Regional cooling or warming – probably slight global-mean cooling	Uncertain	Negligible

162
163

164   Within the troposphere, the relative changes in temperature at different levels are controlled by

165   different processes according to latitude. In the tropics, the primary control is the

166   thermodynamics of moist air (i.e., the effects of evaporation at the surface and the release of

167   latent heat through condensation that occurs in clouds as moist air rises due to convection), and

168   the way these effects are distributed and modified by the atmospheric circulation.

169   Thermodynamic principles require that temperature changes in the tropics will be larger in the

170   troposphere than near the surface ("amplification"), largely independent of the type of forcing. In

171   mid to high latitudes, the processes controlling how temperature changes in the vertical are more

172   complex, and it is possible for the surface to warm more than the troposphere. These issues are

173   addressed in Chapter 1 and Chapter 5.

174

175   ***2. Strengths and limitations of the observational data***

176   The second question is:

177   *What kinds of atmospheric temperature variations can the current observing systems detect and*
178   *what are their strengths and limitations, both spatially and temporally?*
179
180   This is addressed in Chapter 2 of this Report. Chapter 2 draws the following main conclusions …

181

182   (1) The observing systems available for this Report are able to detect small surface and upper air

183   temperature variations from year to year as well as trends[4] in climate since the late 1950s (and

184   over the last century for surface observations), once the raw data are successfully adjusted for

185   changes over time in observing systems and practices, and micro-climate exposure.

186   Measurements from all systems require such adjustments. This Report relies solely on adjusted

187   data sets.

188

189 (2) Independently-performed adjustments to the land surface temperature record have been

190 sufficiently successful that trends given by different data sets are very similar on large (e.g.,

191 continental) scales. This conclusion holds to a lesser extent for the ocean surface record.

192

193 (3) Adjustments for changing instrumentation are most challenging for upper-air datasets. While

194 these show promise for trend analysis, it is not clear that current upper-air climate records have

195 achieved sufficient accuracy to resolve trend-related scientific questions.

196 • Upper-air datasets have been subjected to less scrutiny than surface datasets.

197 • Adjustments are complicated, large compared to the linear trend signal, involve

198 expert judgments, and cannot be stringently evaluated because of lack of traceable

199 standards.

200 • Unlike surface trends, reported upper-air trends vary considerably between research

201 teams beginning with the same raw data owing to their different decisions on how to

202 remove non-climatic factors.

203

204 Many different methods are used to measure temperature changes at the Earth's surface and at

205 various levels in the atmosphere. Near-surface temperatures have been measured for the longest

206 period, over a century, and are measured directly by thermometers. Over land, these data come

207 from fixed meteorological stations. Over the ocean, measurements are of both air temperature

208 and sea-surface (top 10 meters) temperature taken by ships or from buoys.

209

210   The next-longest records are upper-air data measured by radiosondes (temperature sensors

211   carried aloft by weather balloons). These have been collected routinely since 1958. There are still

212   substantial gaps in radiosonde coverage.

213

214   Satellite data have been collected for the upper air since 1979 with almost complete global

215   coverage. The most important satellite records come from Microwave Sounding Units (MSU) on

216   polar orbiting satellites. The microwave data from MSU instruments require calculations and

217   adjustments in order to be interpreted as temperatures. Furthermore, these satellite data do not

218   represent the temperature at a particular level, but, rather, the average temperature over thick

219   atmospheric layers (see Figure 2.2 in Chapter 2). Channel 2 data (mid to upper troposphere, T_2)

220   have a latitudinally-dependent contribution from the stratosphere, while Channel 4 data (lower

221   stratosphere, T_4) have a latitudinally-dependent contribution from the troposphere, factors that

222   complicate their interpretation.

223

224   All measurement systems have inherent uncertainties associated with: the instruments employed;

225   changes in instrumentation; and the way local measurements are combined to produce area

226   averages. All data sets require careful examination for instrument biases and reliability, and

227   adjustments to remove changes that might have arisen for non-climatic reasons. The term

228   "homogenization" is used to describe this adjustment procedure. Recent improvements in and

229   corrections to some of these adjustments have resulted in better agreement between data sets.

230

231

232   ***3. What temperature changes have been observed?***

233   This Section combines information related to questions 3 and 4:

234   *What do observations indicate about the changes of temperature in the atmosphere and at*
235   *the surface since the advent of measuring temperatures vertically?*
236
237   *What is our understanding of the contribution made by observational or methodological*
238   *uncertainties to the previously reported vertical differences in temperature trends?*
239

240   These questions are addressed in Chapters 3 and 4 of this Report. The following conclusions are

241   drawn in these chapters. Supporting information is given in Figure 1 and Figure 2.

242

243   (1) Surface temperatures: For global-mean changes, as well as in the tropics (20°S to 20°N), all

244   data sets show warming at the surface since 1958, with a greater rate of increase since 1979.

245   Differences between the data sets are small.

246

247   • Global-mean temperature increased at about 0.12^{o}C per decade since 1958, and about

248   0.16^{o}C per decade since 1979. In the tropics, temperature increased at about 0.11^{o}C per

249   decade since 1958, and about 0.13^{o}C per decade since 1979.

250

251   • Local biases in surface temperatures may exist due to changes in station exposure and

252   instrumentation over land[5], or changes in measurement techniques by ships and buoys in

253   the ocean. It is likely that these biases are largely random and therefore cancel out over

254   large regions such as the globe or tropics, the regions that are of primary interest to this

255   Report.

256

257   • Errors in observed surface/troposphere trend differences are more likely to come from

258   errors in tropospheric data than from errors in surface data.

259

260   (2) Tropospheric temperatures: All data sets show that the global-mean and tropical troposphere

261   has warmed from 1958 to the present, with the warming in the troposphere being slightly more

262   than at the surface. Since 1979, due to the considerable disagreements between tropospheric data

263   sets, it is not clear whether the troposphere has warmed more than or less than the surface.

264

265   • Global-mean tropospheric temperature increased at about 0.14^{o}C per decade since 1958,

266   and between 0.10^{o}C and 0.20^{o}C per decade since 1979. In the tropics, temperature

267   increased at about 0.13^{o}C per decade since 1958, and between 0.02^{o}C and 0.19^{o}C per

268   decade since 1979.

269

270   • It is very likely that trends in troposphere temperatures are affected by errors that remain

271   in the homogenized radiosonde data sets. Such errors arise because the methods used to

272   produce these data sets are only able to detect and remove the more obvious cases, and

273   involve many subjective decisions. The full consequences of these errors for large-area

274   averages, however, have not yet been fully resolved. Nevertheless, it is likely that a net

275   spurious cooling corrupts the area-averaged homogenized radiosonde data in the tropical

276   troposphere, causing these data to indicate less warming than has actually occurred there.

277

278   • For tropospheric satellite data, a primary cause of trend differences between different

279   versions is differences in how the data from different satellites are merged together.

280   Corrections required to account for drifting measurement times, and diurnal cycle

281   adjustments are also important.

282

283   • Comparisons between satellite and radiosonde temperatures for the mid to upper

284   tropospheric layer (MSU channel 2; T_2) are very likely to be corrupted by excessive

285   stratospheric cooling in the radiosonde data

286

287   (3) Lower stratospheric temperatures: All data sets show that the stratosphere has cooled

288   considerably from 1958 and from 1979 to the present, although there are large differences in the

289   linear trend values from different data sets.

290

291   • The largest differences between data sets are in the stratosphere, particularly between the

292   radiosonde and satellite-based data sets. It is very likely that the satellite/radiosonde

293   discrepancy arises primarily from uncorrected errors in the radiosonde data.

294   Figure 1 shows the various temperature time series examined in this Report.
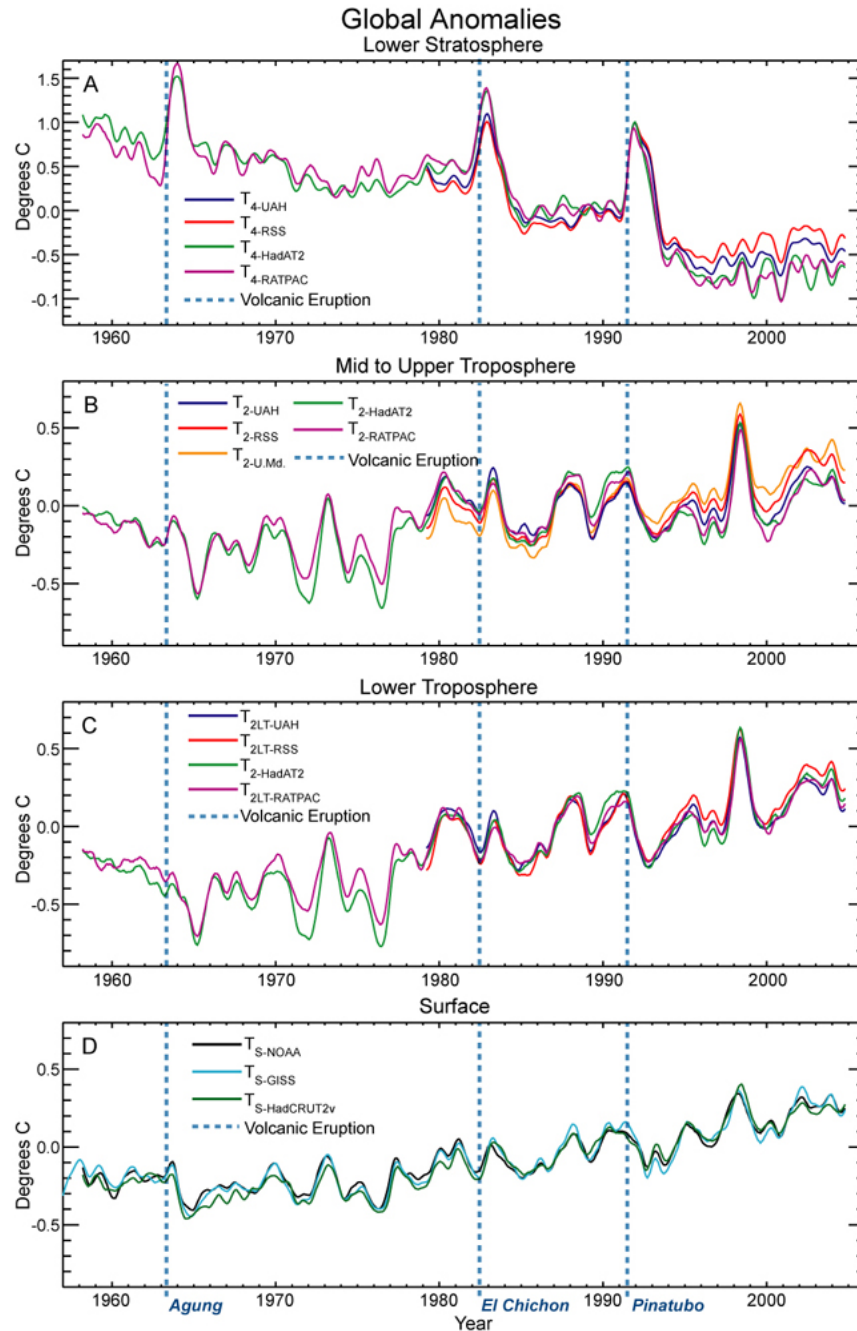
Figure 1: Observed surface and upper air global-mean temperature records. From top to bottom: A, lower stratosphere (denoted T_4) records from two satellite analyses (UAH and RSS) together with equivalently-weighted radiosonde records based on HadAT2 and RATPAC data; B, mid- to upper-troposphere (T_2) records from three satellite analyses (UAH, RSS and U.Md.) together with equivalently-weighted radiosonde records based on HadAT2 and RATPAC; C, lower troposphere (T_{2LT}) records from UAH and RSS (satellite), and from HadAT2 and RATPAC (equivalently-weighted radiosonde); D, surface (T_S). All time series are based on monthly-mean data smoothed with a 7-month running average, expressed as departures from the Jan. 1979 to Dec. 1997 average. Note that the T_2 data (panel B) contain a small contribution (about 10%) from the lower stratosphere. Information here is from Figures 3.1, 3.2 and 3.3 in Chapter 3.

306   For the lower stratosphere, the cooling trend since the late 1950s (which is as expected due to the

307   effects of greenhouse-gas concentration increases and stratospheric ozone depletion) is

308   punctuated by short-term warming events associated with the explosive volcanic eruptions of Mt.

309   Agung (1963), El Chichón (1982) and Mt. Pinatubo (1991).

310

311   Both the troposphere and the surface show warming since the late 1950s. For the surface, most,

312   if not all of the temperature increase since 1958 occurs starting around the mid-1970s, a time

313   coincident with a previously identified climate regime shift. However, there does not appear to

314   be a strong jump up in temperature at this time; rather, the major part of the rise seems to occur

315   in a more gradual fashion. For the balloon-based tropospheric data, the major part of the

316   temperature increase since 1958 appears in the form of a rapid rise in the mid-1970s, apparently

317   in association with the climate regime shift that occurred at this time.

318

319   The dominant shorter time scale fluctuations are those associated with the El Niño-Southern

320   Oscillation phenomenon (ENSO). The major ENSO warming event in 1998 is obvious in all

321   records. Cooling following the eruptions of Mt. Agung and Mt. Pinatubo is also evident, but the

322   cooling effect of El Chichón is masked by an ENSO warming that occurred at the same time.

323   The changes following volcanic eruptions (i.e., surface and tropospheric cooling and

324   stratospheric warming) are consistent with our physical understanding and with model

325   simulations.

326

327   Global-mean temperature changes over the periods 1958 through 2004 and 1979 through 2004

328   are shown in Figure 2 in degrees Celsius and degrees Fahrenheit.

329

330
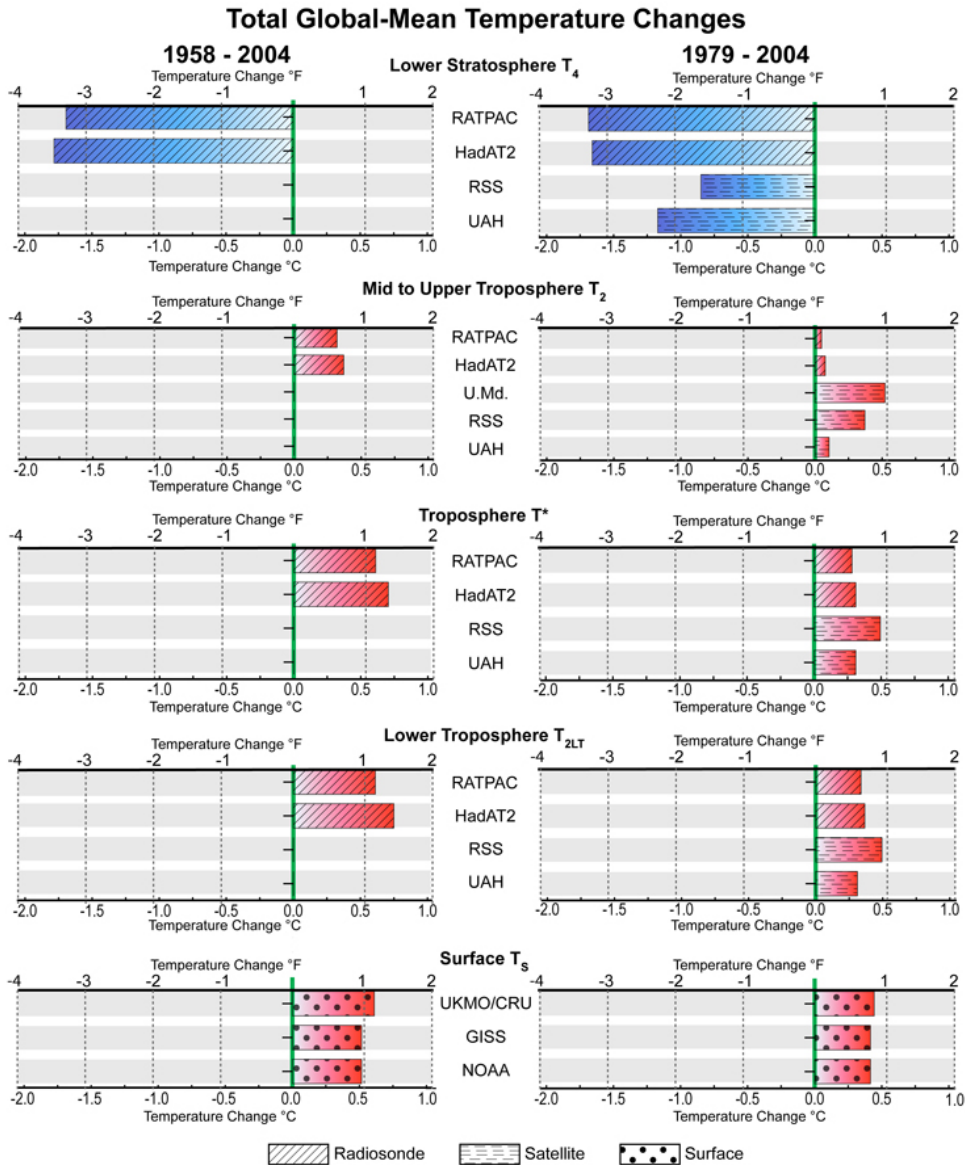
Total Global-Mean Temperature Changes

331
332
333   Figure 2: Total global-mean temperature changes for the surface and different atmospheric layers, from different
334   data sets and over two periods, 1958 to 2004 and 1979 to 2004. The values shown are the total change over the
335   stated period in both degrees Celsius (degC; lower scales) and degrees Fahrenheit (degF; upper scales). All changes
336   are statistically significant at the 5% level except RSS T_4 and RATPAC, HadAT2 and UAH T_2. Total change in
337   degC is the linear trend in degC per decade (see Tables 3.2 and 3.3 in Chapter 3) times the number of decades in the
338   time period considered. Total change in degF is this number times 1.8 to convert to degF. For example, the Table 3.2
339   trend for NOAA surface temperatures over January 1958 through December 2004 is 0.11°C/decade. The total
340   change is therefore 0.11 times 4.7 decades to give a total change of 0.53°C, Multiplying this by 1.8 gives a total
341   change in degrees Fahrenheit of 0.93°F. Warming is shown in red, and cooling in blue.

342

343   ***4. Are model simulations consistent with the observed temperature changes?***

344   Computer-based climate models encapsulate our understanding of the climate system and the

345   driving forces that lead to changes in climate. Such models are the only tools we have for

346   estimating the likely patterns of response of the climate system to different forcing mechanisms.

347   The crucial test of our understanding is to compare model simulations with observed changes.

348   The fifth question therefore is:

349   *How well can the observed vertical temperature changes be reconciled with our*
350   *understanding of the causes of these changes?*
351
352   This question is addressed in Chapter 5 of this Report. Chapter 5 draws the following

353   conclusions …

354

355   PATTERN STUDIES

356

357   (1) Results from many different pattern-based "fingerprint"[5] studies (see Box 5.5 in Chapter 5)

358   provide consistent evidence for human influences on the three-dimensional structure of

359   atmospheric temperature changes over the second half of the 20th century.

360

361   •   Fingerprint studies have identified greenhouse gas and sulfate aerosol signals in observed

362   surface temperature records, a stratospheric ozone depletion signal in stratospheric

363   temperatures, and the combined effects of these forcing agents in the vertical structure of

364   atmospheric temperature changes.

365

366   (2) Natural factors have influenced surface and atmospheric temperatures, but cannot fully

367   explain their changes over the past 50 years.

368

369   LINEAR TREND COMPARISONS[4,6]

370

371   (3) When models are run with natural and human-induced forcings, simulated global-mean

372   temperature trends for individual atmospheric layers are consistent with observations.

373

374   (4) Comparing trend differences between the surface and the troposphere exposes potential

375   model/observed data discrepancies in the tropics.

376

377   • In the tropics, the majority of observational data sets show more warming at the surface

378   than in the troposphere, while almost all model simulations have larger warming aloft

379   than at the surface.

380

381   AMPLIFICATION OF SURFACE WARMING IN THE TROPOSPHERE

382

383   (5) Amplification means that temperatures show larger changes aloft than at the surface.

384   In the tropics, on monthly and inter-annual time scales, both models and observations show

385   amplification of temperature variability in the troposphere relative to the surface. This

386   amplification is of similar magnitude in models and observations. For multi-decadal trends,

387   models show the same amplification that is seen on shorter time scales. A number of observed

388   data sets, however, do not show this amplification.

389

390   • These results have several possible explanations, which are not mutually exclusive. One

391   explanation is that "real world" amplification effects on short and long time scales are

392   controlled by different physical mechanisms, and models fail to capture such behavior. A

393   second explanation is that remaining errors in some of the observed tropospheric data sets

394   adversely affect their long-term temperature trends. The second explanation is more

395   likely in view of the model-to-model consistency of amplification results, the large

396   uncertainties in observed tropospheric temperature trends, and independent physical

397   evidence supporting substantial tropospheric warming.

398

399   OTHER FINDINGS

400

401   (6) Because of differences between different observed data sets, it is important to account for

402   observational uncertainty in comparisons between modeled and observed temperature changes.

403

404   • Large "construction" uncertainties in observed estimates of global-scale atmospheric

405   temperature change can critically influence the outcome of consistency tests between

406   models and observations.

407

408   (7) Inclusion of previously-ignored spatially-heterogeneous forcings in the most recent climate

409   models does not fundamentally alter conclusions about the amplification of warming in the

410   troposphere relative to the surface.

411

412   •   Changes in sulfate aerosols and tropospheric ozone, which have spatially-heterogeneous

413   forcings, have been incorporated routinely in climate model experiments for a number of

414   years. It has been suggested that the spatially-heterogeneous forcing effects of black

415   carbon aerosols and land use/land cover may have had significant effects on regional

416   temperatures that might modify previous conclusions regarding vertical temperature

417   changes. These forcings have been included for the first time in about half of the global

418   model simulations considered here. Within statistical uncertainties, model simulations

419   that include these forcings show the same amplification of warming in the troposphere

420   relative to the surface at very large spatial scales (global and tropical averages) as

421   simulations in which these forcings are neglected.

422

423   Chapter 5 analyses state-of-the-art model simulations from 19 institutions globally, run using

424   combinations of the most important natural and human-induced forcings. The Chapter compares

425   the results of these simulations with a number of different observational data sets for the surface

426   and different atmospheric layers, resulting in a large number of possible model/observed data

427   comparisons.

428

429   Figures 3 and 4 summarize the new model results used in this Report, together with the

430   corresponding observations. Figure 3 gives global-mean results, while Figure 4 gives results for

431   the tropics (20°S to 20°N). Model and observed results are compared in these Figures using

432   linear trends over the period January 1979 through December 1999[7] for the surface, for

433   individual layers, and (right-hand panels) for surface changes relative to the troposphere.

434   Rectangles are used to illustrate the ranges of both model trends (red rectangles) and observed

435   trends (blue rectangles). Individual observed-data trends are also shown.

436

437   Since statistical uncertainties (see Appendix) are not shown in these Figures, the rectangles do

438   not represent the full ranges of uncertainty. However, they allow a meaningful first-order

439   assessment of model/observed similarities and differences. Fully overlapping rectangles indicate

440   consistency, partially overlapping rectangles point to possible discrepancies, while rectangles

441   that either do not overlap or show minimal overlap indicate important model/observed data

442   inconsistencies. At the global-mean level, models and observations generally show fully

443   overlapping rectangles. The only potentially serious inconsistency is in the tropics (Figure 4)

444   where the troposphere warms more rapidly than the surface in all except two of the 49 individual

445   model simulations examined here, while, in the majority of observational data sets, the surface

446   has warmed more rapidly than the troposphere.
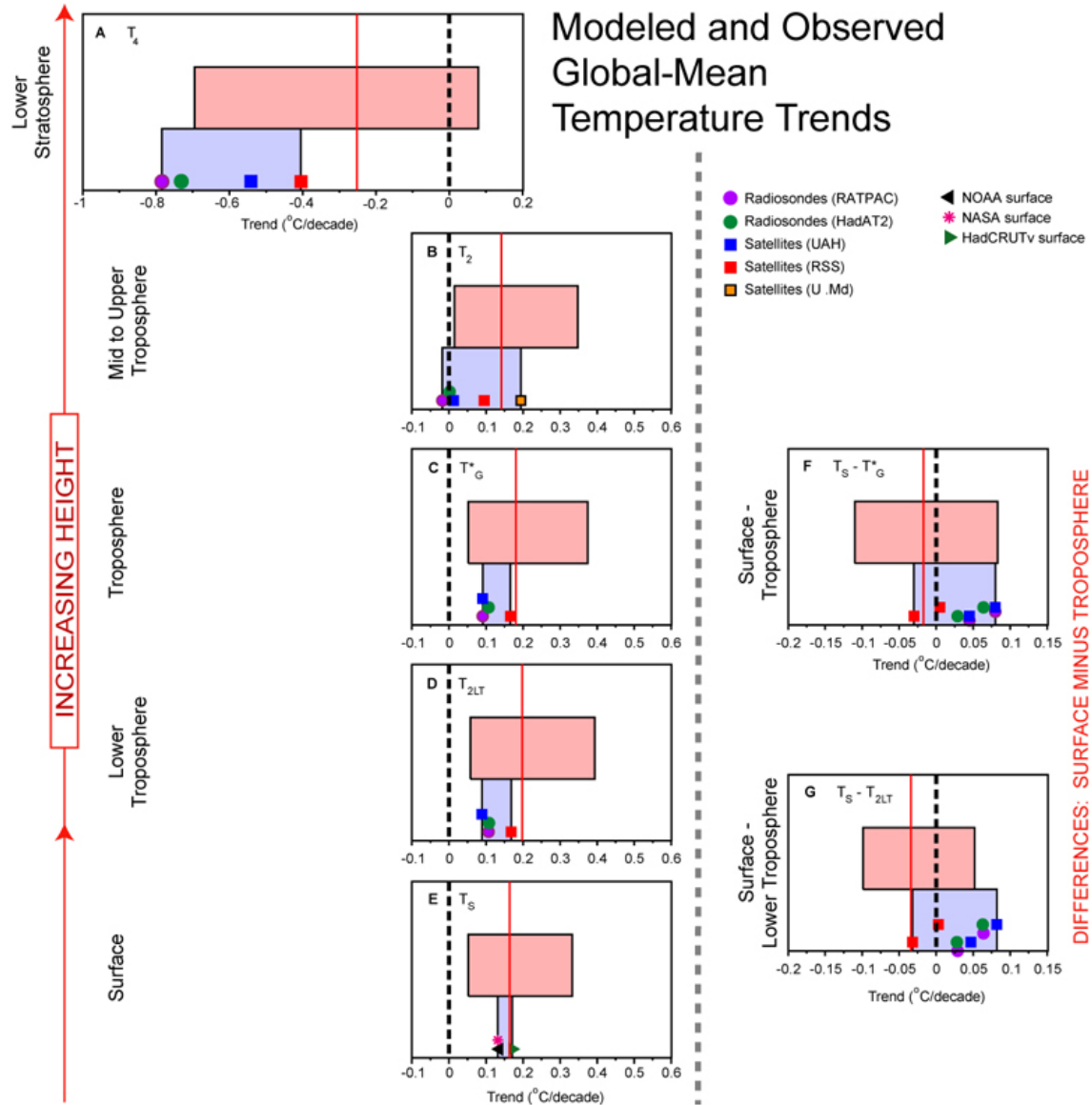
447
448
449 Figure 3: Comparison of observed and model-simulated global-mean temperature trends (left-hand panels) and trend
450 differences (right-hand panels) over January 1979 through December 1999, based on Table 5.4A and Figure 5.3 in
451 Chapter 5. The upper red rectangles in each box show the range of model trends from 49 model simulations. The
452 lower blue rectangles show the range of observed trends, with the individual trends from different data sets indicated
453 by the symbols. From bottom to top, the left-hand panels show trends for the surface (T_S), the lower troposphere
454 (T_{2LT}), the troposphere (T^*), the mid troposphere to lower stratosphere (T_2), and the lower stratosphere (T_4). The
455 right-hand panels show differences in trends between the surface and either the troposphere or the lower
456 troposphere, with a positive value indicating a stronger warming at the surface. The red vertical lines show the
457 average of all model results. The vertical black dashed lines show the zero value. For the observed trend differences,
458 there are eight values corresponding to combinations of the four upper-air data sets (as indicated by the symbols)
459 and either the HadCRUT2v surface data or the NASA/NOAA surface data (which have almost identical trends).
460
461
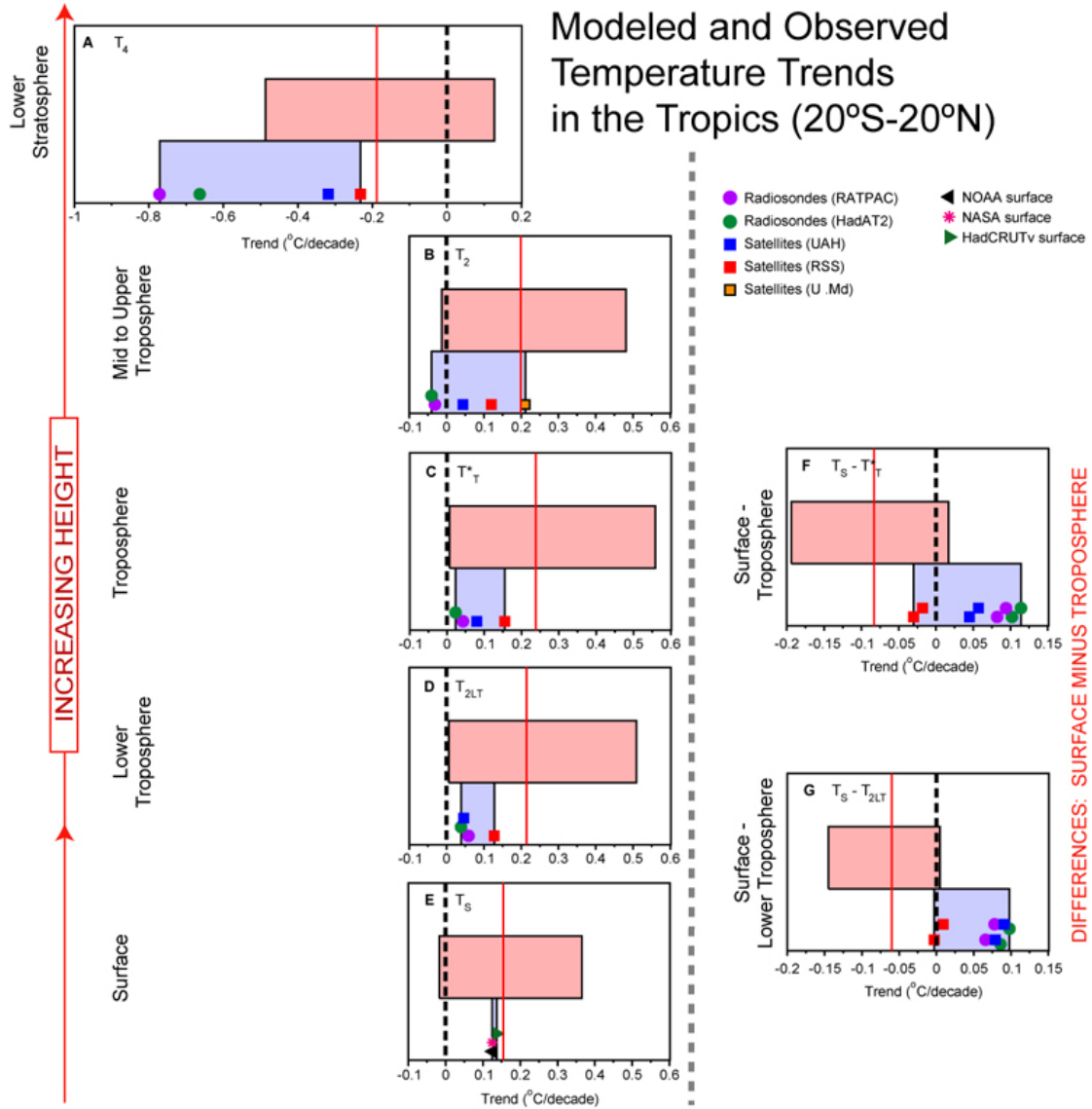
462
463
464   Figure 4: As Figure 3, but for the tropics (20°S to 20°N), based on Table 5.4B and Figure 5.4 in Chapter 5.
465

466
467

468

469

470   ***5. Recommendations***

471   This Section addresses question 6:

472   *What measures can be taken to improve the understanding of observed changes?*

473
474   In answer to this question, drawing on the material presented in the first five chapters of this

475   Report, a set of primary recommendations has been developed and is described in detail in

476   Chapter 6. It should be noted that, rather than invent new proposals or recommendations, the

477   items described in Chapter 6 expand and build upon existing ideas, emphasizing those that are

478   considered to be of highest utility. The seven recommendations are:

479

480   (1) In order to encourage further independent scrutiny, data sets and their full metadata (i.e.,

481   information about instrumentation used, observing practices, the environmental context of

482   observations, and data-processing procedures) should be made openly available. Comprehensive

483   analyses should be carried out to ascertain the causes of remaining differences between data sets

484   and to refine uncertainty estimates.

485

486   (2) Efforts should be made to archive and make openly available surface, balloon-based, and

487   satellite data and metadata that have not previously been exploited. Emphasis should be placed

488   on the tropics.

489

490   (3) Efforts should be made to create climate quality data sets[8] for a range of variables other than

491   temperature. These data sets should subsequently be compared with each other and with

492   temperature data to determine whether they are consistent with our physical understanding.

493

494   (4) Efforts should be made to create several homogeneous atmospheric reanalyses[9]. Particular

495   care needs to be taken to identify and homogenize critical input data. Identification of critical

496   data requires, in turn, observing system experiments where the impacts and relative importance

497   of different observation types from land, radiosonde, and space-based observations are assessed.

498

499   (5) Models that appear to include the same forcings often differ in both the way the forcings are

500   quantified and how these forcings are applied to the model. Hence, efforts are required to

501   separate more formally uncertainties arising from model structure from the effects of forcing

502   uncertainties. This requires running multiple models with standardized forcings, and running the

503   same models individually under a range of plausible scenarios for each forcing.

504

505   (6) The GCOS (Global Climate Observing System) climate monitoring principles should be fully

506   adopted. In particular, when any particular type of instrument is changed or re-sited, there should

507   be a period of overlap between old and new instruments or configurations that is sufficient to

508   allow analysts to adjust for the change with small uncertainties that do not prejudice the analysis

509   of climate trends. The minimum period is a full annual cycle of the climate.

510

511   (7) A small subset (about 5%) of the operational radiosonde network should be developed and

512   implemented as reference sites for all kinds of climate data from the surface to the stratosphere.

513

514

515   **Footnotes**

516
517   [1] For details of new observed data see Table 3.1 in Chapter 3. For details of new models and
518   model simulations see Chapter 5 and http://www-pcmdi.llnl.gov/ipcc/model.documentation.
519
520   [2] We use the words "warming" and "cooling" here to refer to temperature increases or decreases,
521   as is common usage. Technically, these words refer to changes in heat content, which may occur
522   through changes in either the moisture content and/or the temperature of the atmosphere. When
523   we say that the atmosphere has warmed (or cooled) over a given period, this means that there has
524   been an overall positive (or negative) temperature change based on a linear trend analysis.
525
526   [3] The main natural perturbations are changes in solar output and the effects of explosive volcanic
527   eruptions. The main human-induced ("anthropogenic") factors are: the emissions of greenhouse
528   gases (e.g., carbon dioxide [CO_2], methane [CH_4], nitrous oxide [N_2O]); aerosols (tiny droplets
529   or particles such as smoke) and the gases that lead to aerosol formation (most importantly, sulfur
530   dioxide); and changes in land cover and land use. Since these perturbations act to drive or
531   "force" changes in climate, they are referred to as "forcings". Tropospheric ozone [O_3], which is
532   not emitted directly, is also an important greenhouse gas. Tropospheric ozone changes occur
533   through the emissions of gases like carbon monoxide, nitrogen oxides and volatile organic
534   compounds, which are not important directly as greenhouse gases.
535
536   [4] Many of the results in this Report (and here in the Executive Summary) are quantified in terms
537   of linear trends, i.e., by the value of the slope of a straight line that is fitted to the data. A simple
538   straight line is not always the best way to describe temperature data, so a linear trend value may
539   be deceptive if the trend number is given in isolation, removed from the original data.
540   Nevertheless, used appropriately, linear trends provide the simplest and most convenient way to
541   describe the overall change over time in a data set, and are widely used. For a more detailed
542   discussion, see the Appendix.
543
544
545   [5] Some have expressed concern that land temperature data might be biased due to urbanization
546   effects. Recent studies specifically designed to identify systematic problems using a range of
547   approaches have found no detectable urban influence in large-area averages in the data sets that
548   have been "homogenized" (i.e., adjusted to remove non-climatic influences).
549
550   [6] Fingerprint studies use rigorous statistical methods to compare the patterns of observed
551   temperature changes with model expectations and determine whether or not similarities could
552   have occurred by chance. Linear trend comparisons are less powerful than fingerprint analyses
553   for studying cause-effect relationships, but can highlight important differences and similarities
554   between models and observations.
555
556   [7] This is the longest period common to all model simulations.
557
558   [8] Climate quality data sets are those where the best possible efforts have been made to identify
559   and remove non-climatic effects that might produce spurious changes over time.

560
561   [9] Reanalyses are mathematically blended products based upon as many observing systems as
562   practical. Observations are assimilated into a global weather forecasting model to produce
563   globally-comprehensive data sets that are most consistent with both the available data and the
564   assimilation model.
565

1
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5

Chapter 1

6
7
8   Why do temperatures vary vertically (from the surface to the stratosphere) and what do we
9   understand about why they might vary and change over time?
10
11
12
13
14

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45

46   **Summary**

47

48 **Temperatures vary vertically.**

49 The solar heating of the surface of the planet, combined with the physical properties of the

50 overlying air, produce the highest temperatures, on average, at the surface; that heat is mixed

51 vertically and horizontally by various physical processes. Taking into account the distribution of

52 atmospheric moisture and the lower air pressure with progressively increasing altitude, there

53 results a decrease of temperature with height up to the tropopause. The tropopause marks the top

54 of the troposphere, i.e., the lower 8 to 16 km of the atmosphere depending on latitude. Above this

55 altitude, the radiative properties of the air produce a warming with height through the

56 stratosphere (extending from the tropopause to ~50 km).

57

58 **Temperature trends at the surface can be expected to be different from temperature trends**

59 **higher in the atmosphere because:**

60   • Physical properties of the surface vary depending on whether the location has land,

61   sea, snow, or ice. Near the surface, these differing conditions can produce strong

62   horizontal variations in temperature. Above the surface layer, these contrasts are

63   quickly smoothed out, contributing to varying temperature trends with height at

64   different locations.

65   • Changes in atmospheric circulation or modes of atmospheric variability (e.g., El

66   Niño-Southern Oscillation [ENSO]) can produce different temperature trends at the

67   surface and aloft.

68   • Under some circumstances, temperatures may increase with height near the surface or

69   higher in the troposphere, producing a "temperature inversion." Such inversions are

70   more common at night, in winter over continents, and in the trade wind regions. Since

71   the air in inversion layers is resistant to vertical mixing, temperatures trends can differ

72   between inversion layers and adjacent layers.

73   • Forcing factors, either natural (e.g., volcanoes and solar) or human-induced (e.g.,

74   greenhouse gas, aerosols, ozone, and land use) can result in differing temperature

75   trends at different altitudes, and these vertical variations may change over time. This

76   can arise due to spatial changes in the concentrations or properties of the forcing

77   agents.

78

79

80   This Chapter describes the temperature profile of the layers of the atmosphere from the surface

81   through the stratosphere and discusses the basic reasons for this profile. We also use results from

82   global climate model simulations to show how changes in natural and human-induced factors can

83   produce different temperature trends in the various layers of the atmosphere. This discussion

84   provides the background for the presentation of the observed changes (Chapters 2-4), and for the

85   understanding of their causes (Chapter 5). We also describe temperature changes in the

86   stratosphere in recent decades and the influences of these changes on the troposphere. Finally,

87   making use of surface and satellite observations, we examine the physical processes that can

88   result in different temperature trends at the surface and in the troposphere.

89

90   **1.1 The Thermal Structure of the Atmosphere**

91   Surface temperatures are at their warmest in the tropics, where the largest amount of solar

92   radiation is received during the course of the year, and decrease towards the Polar Regions where

93   the annual-mean solar radiation received is at a minimum (Oort and Peixoto, 1992). The

94   temperature contrast between summer and winter increases from the equator to the poles. Since

95   land areas heat up and cool more rapidly than oceans, and because of the preponderance of land

96   in the Northern Hemisphere, there is a larger contrast between summer and winter in the

97   Northern Hemisphere.

98

99   Figure 1.1 illustrates the climatological vertical temperature profiles for December, January,

100 February and June, July, August mean conditions, as obtained from the National Centers for

101 Environmental Prediction (NCEP) reanalyses (Kalnay et al., 1996; updated). It is convenient to

102 think first in terms of climatological conditions upon which spatial and temporal

103 variations/trends are superimposed. The solid line in the plot illustrates the tropopause, which

104 separates the troposphere below from the stratosphere above. The tropopause is at its highest

105 level in the tropics (~20°N-20°S). It descends sharply in altitude from ~16 km at the equator to

106 ~12 km at ~30-40° latitude, and to as low as about 8 km at the poles.
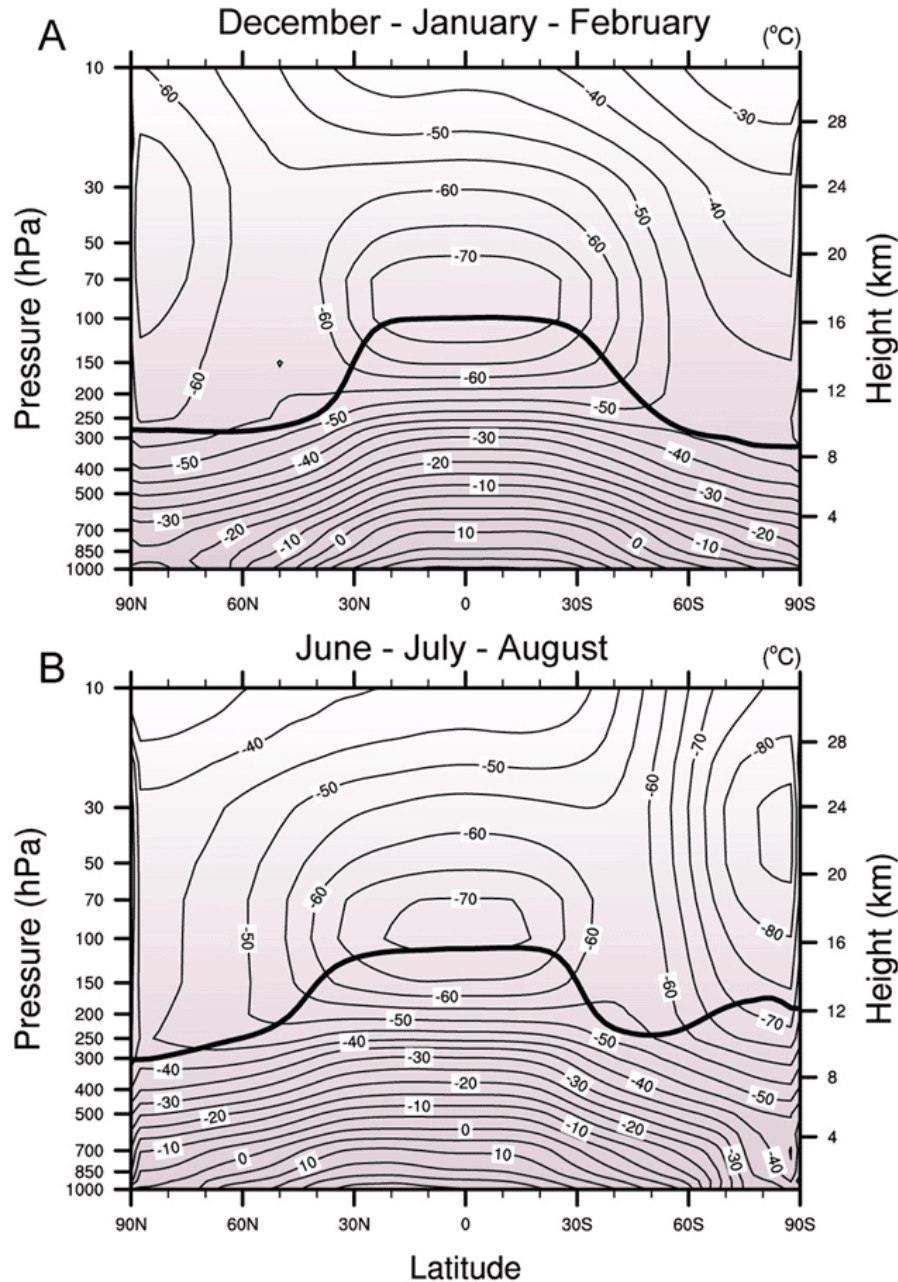
Figure 1.1. Global climatological vertical temperature profiles from surface to troposphere and extending into the
stratosphere for December-January-February and June-July-August mean conditions, as obtained from the National
Centers for Environmental Prediction reanalyses (Kalnay et al., 1996; updated). The solid line denotes the
tropopause which separates the stratosphere from the surface-troposphere system.

Temperatures generally decrease with height from the surface although there are important

exceptions. The rate at which the temperature changes with height is termed the "lapse rate." The

lapse rate can vary with location and season, and its value depends strongly on the atmospheric

116   humidity, e.g., the lapse rate varies from ~5ºC/km near the surface in the moist tropical regions

117   (near the equator) to much larger values (~8-9ºC/km) in the drier subtropics (~20-30°). Important

118   departures from nominal lapse rate values can occur near the surface and in the upper

119   troposphere. In the equatorial tropics, the tropopause region (~16 km) is marked by a smaller

120   value of the lapse rate than in the lower troposphere.

121

122   The thermal structure of the lowest 2-3 km, known as the "planetary boundary layer," can be

123   complicated, even involving inversions (in which temperature increases rather than decreases

124   with height) occurring at some latitudes due to land-sea contrasts, topographic influences,

125   radiative effects and meteorological conditions. Inversions are particularly common during

126   winter over some middle and high latitude land regions and are a climatological feature in the

127   tropical trade wind regions. The presence of inversions acts to decouple surface temperatures

128   from tropospheric temperatures on daily or even weekly timescales.

129

130   Above the tropopause is the stratosphere, which extends to ~50 km and where the temperature

131   increases with height. In the vicinity of the tropical tropopause, (i.e., the upper troposphere and

132   lower stratosphere regions, ~15-18 km), the temperature varies little with height. The

133   extratropical (poleward of 30°) lower stratosphere (at ~8-12 km) also exhibits a similar feature

134   (Holton, 1979). The lapse rate change with altitude in the upper troposphere/lower stratosphere

135   region is less sharp in the extratropical latitudes than in the tropics.

136

137   The global temperature profile of the atmosphere reflects a balance between the radiative and

138   dynamical heating/cooling of the surface-atmosphere system. From a global, annual-average

139   point of view, the thermal profile of the stratosphere is the consequence of a balance between

140   radiative heating and cooling rates due to greenhouse gases, principally carbon dioxide (CO_2),

141   ozone (O_3) and water vapor (H_2O) (Andrews et al., 1987). The vertical profile of the troposphere

142   is the result of a balance of radiative processes involving greenhouse gases, aerosols, and clouds

143   (Stephens and Webster, 1981; Goody and Yung, 1989), along with the strong role of moist

144   convection (Holton, 1979; Kiehl, 1992). An important difference between the troposphere and

145   stratosphere is that the stratosphere is characterized by weak vertical motions, while in the

146   troposphere, the vertical motions are stronger. Most significantly, the moist convective processes

147   that are a characteristic feature of the troposphere include the transfer of large amounts of heat

148   due to evaporation or condensation of water.

149

150   Convective processes are important in determining the temperature profile in the troposphere.

151   This is illustrated by the fact that radiative processes alone would cause the surface to be

152   significantly warmer than it is actually. This would occur because the atmosphere is relatively

153   transparent to the Sun's radiation. By itself, this would lead to a drastic heating of the surface

154   accompanied by a net radiative cooling of the atmosphere (Manabe and Wetherald, 1967).

155   However, the resulting convective motions remove this excess heating from the surface in the

156   form of sensible and latent heat, the latter involving the evaporation of moisture from the surface

157   (Ramanathan and Coakley, 1978). As air parcels rise and expand, they cool due to

158   decompression, leading to a decrease of temperature with height. The lapse rate for a dry

159   atmosphere, when there are no moist processes and the air is rising quickly enough to be

160   unaffected by other heating/cooling sources, is close to 10ºC/km. However, because of moist

161   convection, there is condensation of moisture, formation of clouds and release of latent heat as

162   the air parcels rise, causing the lapse-rate to be much less, as low as 4ºC/km in very humid

163   atmospheres (Houghton, 1977). In a more rigorous sense, the interactions between radiation,

164   convection, cloud physics, and dynamical motions (ranging from large- to meso- and small-

165   scales) govern the actual rate at which temperature decreases with height (the lapse-rate) at any

166   location. Large-scale dynamical mechanisms tend to homogenize temperatures above the

167   boundary layer over horizontal scales (Rossby radius) that vary from planetary scale near the

168   equator to a couple of thousand kilometers at midlatitudes and to a few hundred kilometers near

169   the poles.

170

171   Convective processes and vertical mixing of air can add complexity to the nominal thermal

172   profile in the tropics mentioned above. For example, a more detailed picture in subtropical

173   regions consists of a surface mixed layer (up to about 500 m) and a trade wind boundary layer

174   (up to about 2 km) above which is the free troposphere. Each of the boundary layers is topped by

175   an inversion which tends to isolate the region from the layer above (Sarachik, 1985). This

176   indicates the limitations in assuming nominal lapse rate values from the surface to the tropopause

177   everywhere.

178

179   The radiative-convective picture above is likely of dominant relevance only for the tropics. In the

180   extra-tropics (poleward of 30°), the lapse rate and tropopause height are mostly determined by

181   instabilities associated with the more familiar weather systems ("baroclinic instability"). The

182   rising motions of air parcels in the equatorial moist tropics associated with deep convection

183   descend in the subtropical regions leading to drier environments there (Hadley circulations). In

184   the polar regions (~60-90°), planetary-scale waves forced by the influences of mountains and

185   that of land-sea contrasts upon the flow of air play a significant role in the determination of the

186   wintertime temperatures at the poles.

187

188   Based on these simple ideas, the lapse rate can be expected to decrease with an increase in

189   humidity, and also to depend on the atmospheric circulation. As specific humidity is strongly

190   related to temperature, and is expected to rise with surface warming, the lapse rate (other things

191   being equal) can be expected to decrease with warming such that temperature changes aloft

192   exceed those at the surface.

193

194   The above simple picture of radiative-convective balance, together with the requirement of

195   radiative balance at the top-of-the-atmosphere (namely, equilibrium conditions wherein the net

196   solar energy absorbed by the Earth's climate system must be balanced by the infrared radiation

197   emitted by the Earth), can help illustrate the significance of long-lived infrared absorbing gases

198   in the atmosphere such as carbon dioxide. The presence of strongly infrared-absorbing

199   greenhouse gases (water vapor, carbon dioxide, methane, etc.) causes the characteristic infrared

200   emission level of the planet to be shifted to a higher altitude where temperatures are colder. The

201   re-establishment of thermal equilibrium leads to warming and communication of the added heat

202   input to the troposphere and surface (Goody and Yung, 1989; Lindzen and Emanuel, 2002).

203

204   In the tropical upper troposphere, moisture- and cloud-related features (e.g., upper tropospheric

205   relative humidity, cirrus cloud microphysics, and mesoscale circulations in anvil clouds) are

206   important factors in governing the thermal profile (Ramanathan et al., 1983; Ramaswamy and

207   Ramanathan, 1989; Donner et al., 2001; Sherwood and Dessler, 2003). In the upper troposphere

208   and especially the stratosphere, convective motions become weak enough that radiative (solar

209   and longwave) heating/cooling become important in establishing the lapse rate.

210

211   **1.2 Forcing of climate change**

212

213   Potentially significant variations and trends are superimposed on the basic climatological thermal

214   profile, as revealed by observational data in the subsequent chapters. While the knowledge of

215   the climatological mean structure discussed in the previous section involves considerations of

216   radiative, convective, and dynamical processes, understanding the features and causes of the

217   magnitude of changes involves a study of the perturbations in these processes which then frame

218   the response of the climate system to the forcing. While the understanding of climate variability

219   is primarily based on observations of substantial changes (e.g., sea-surface temperature changes

220   during El Niño), the vertical temperature changes being investigated in this report are changes on

221   the order of a few tenths of degrees on the global-mean scale (local changes could be much

222   greater), as discussed in the subsequent chapters.

223

224   "Unforced" variations, i.e., changes arising from internally generated variations in the

225   atmosphere-ocean-land-ice/snow climate system, can influence surface and atmospheric

226   temperatures substantially, e.g., due to changes in equatorial sea-surface temperatures associated

227   with ENSO. Climate models indicate that global-mean unforced variations on multidecadal

228   timescales are likely to be smaller than, say, the 20th Century global-mean increase in surface

229   temperature (Stouffer et al., 2000). However, for specific regions and/or seasons, this may not be

230   valid and the unforced variability could be substantial. Chapter 5 provides more detail on models

231   and their limitations (see particularly Box 5.1 and 5.2).

232

233   Because of the influence of radiative processes on the thermal structure, anything external to the

234   climate system that perturbs the planet's radiative heating distribution can cause climate changes,

235   and thus is potentially important in accounting for the observed temperature changes (Santer et

236   al., 1996). The radiative (solar plus longwave) heat balance of the planet can be forced by:

237   •   natural factors such as changes in the Sun's irradiance, and episodic, explosive volcanic

238   events (leading to a build-up of particulates in the stratosphere);

239   •   human-induced factors such as changes in the concentrations of radiatively active gases

240   (carbon dioxide, methane, etc.) and aerosols.

241   Potentially important external forcing agents of critical relevance for the surface and atmospheric

242   temperature changes over the 20th Century are summarized in Table 1.1 (for more details, see

243   Ramaswamy et al., 2001; NRC, 2005).

244 Table 1.1. Agents potentially causing an external radiative forcing of climate change in the 20th Century (based on
245 Ramaswamy et al., 2001). See notes below for explanations.
246

Forcing agent	Nat. (N) or Anth. (A)	Solar pert.	Longwave pert.	Surface rad. effect	Tropos. rad. effect	Stratos. rad. effect	Geog. dis. (global G or localized, L)	Level of confidence
Well-mixed greenhouse gases	A	(small)	Y	Y	Y	Y	G	High
Trop. ozone	A	Y	Y	Y	Y	(small)	L	Medium
Strat. ozone	A	Y	Y	(small)	Y	Y	L	Medium
Sulfate aero. (direct)	A	Y	-	Y	(small)	-	L	Low
Black carbon aero. (direct)	A	Y	(small)	Y	Y	-	L	Very low
Organic carbon aero. (direct)	A	Y	-	Y	(small)	-	L	Very low
Biomass burning aero. (direct)	A	Y	-	Y	Y	-	L	Very low
Indirect aerosol effect	A	Y	Y	Y	Y	(small)	L	Very low
Land-use	A	Y	(small)	Y	-	-	L	Very low
Aircraft contrails	A	(small)	(small)	(small)	(small)	-	L	Very low
Sun	N	Y	-	Y	(small)	Y	G	@Very low
Volcanic aero.	N	Y	Y	Y	(small)	Y	#	*

247
248 **Notes**:
249 Natural (N) and Anthropogenic (A) sources of the forcing agents. Direct aerosol forcing is to be contrasted with the
250 indirect effects; the latter comprise the so-called first, second, and semi-direct effects. Y denotes a significant
251 component, "small" indicates considerably less important but not negligible, while no entry denotes a negligible

252 component. Forcings other than well-mixed gases and solar are spatially localized, with the degree of localization
253 having considerable variations amongst the different agents, depending on their respective source locations. In
254 addition, for short-lived species such as ozone and aerosols, the long-range meteorological transport plays an
255 important role in their global distributions. Level of confidence is a subjective measure of the certainty in the
256 quantitative estimate of the global-mean forcing.
257 **#** Typically, the forcing becomes near-global a few months after an intense tropical eruption.
258 * * In the case of volcanic aerosols, the level of confidence in the forcing from the most recent intense eruption, that
259 of Mt. Pinatubo in 1991, is reasonably good because of reliable observing systems in place; for prior explosive
260 eruptions, observations were absent or sparse which affects the reliability of the quantitative estimates for the
261 previous volcanic events.
262 @ Although solar irradiance variations before 1980 have a very low level of confidence, direct observations of the
263 Sun's output from satellite platforms since 1980 are considered to be accurate (Lean et al., 2005). Thus, the forcing
264 due to solar irradiance variations from 1980 to present are known to a much greater degree of confidence than from
265 pre-industrial to present time.
266
267
268 The forcing agents differ in terms of whether their radiative effects are felt primarily in the

269 stratosphere or troposphere or both, and whether the perturbations occur in the solar or longwave

270 spectrum or both. The quantitative estimates of the forcing due to the well-mixed greenhouse

271 gases (comprised of carbon dioxide, methane, nitrous oxide and halocarbons) are known to a

272 higher degree of scientific confidence than the other forcings. Among aerosols, black carbon is

273 distinct because it strongly absorbs solar radiation (see also Box 5.3). In contrast to sulfate

274 aerosols, which cause a perturbation of solar radiation mainly at the surface (causing a cooling

275 effect), black carbon acts to warm the atmosphere while cooling the surface (Chung et al., 2002;

276 Menon et al., 2002). This could have implications for convective activity and precipitation

277 (Ramanathan et al., 2005), and the lapse rate (Chung et al., 2002; Erlick and Ramaswamy, 2003).

278 The response to radiative forcings need not be localized and can manifest in locations remote

279 from the perturbation. This is because atmospheric circulation tends to homogenize the effect of

280 heat perturbations and hence the temperature response. The vertical partitioning of the radiative

281 perturbation determines how the surface heat and moisture budgets respond, how the convective

282 interactions are affected, and hence how the surface temperature and the atmospheric thermal

283 profile are altered. "Indirect" aerosol effects arising from aerosol-cloud interactions can lead to

284 potentially significant changes in cloud characteristics such as cloud lifetimes, frequencies of

285   occurrence, microphysical properties, and albedo (reflectivity) (e.g., Lohmann et al., 2000;

286   Sherwood, 2002; Lohmann and Feichter, 2005). As clouds are important components in both

287   solar and longwave radiative processes and hence significantly influence the planetary radiation

288   budget (Ramanathan et al., 1987; Wielicki et al., 2002), any effect caused by aerosols in

289   perturbing cloud properties is bound to exert a significant effect on the surface-troposphere

290   radiation balance and thermal profile.

291

292   Estimates of forcing from anthropogenic land-use changes have consisted of quantification of the

293   effect of snow-covered surface albedos in the context of deforestation (Ramaswamy et al., 2001).

294   However, there remain considerable uncertainties in these quantitative estimates. There are other

295   possible ways in which land-use change can affect the heat, momentum and moisture budgets at

296   the surface (e.g., changes in transpiration from vegetation) (see also Box 5.4), and thus exert a

297   forcing of the climate (Pielke et al., 2004; NRC, 2005). In addition to the forcings shown in

298   Table 1.1, NRC (2005) has evoked a category of "nonradiative" forcings involving aerosols,

299   land-cover, and biogeochemical changes which may impact the climate system first through

300   nonradiative mechanisms, e.g., modifying the hydrologic cycle and vegetation dynamics.

301   Eventually, radiative impacts could occur, though no metrics for quantifying these nonradiative

302   forcings have been accepted as yet (NRC, 2005).

303

304   Not all the radiative forcings are globally uniform. In fact, even for the increases in well-mixed

305   greenhouse gases, despite their globally uniform mixing ratios, the resulting forcing of the

306   climate system is at a maximum in the tropics due to the temperature contrast between the

307   surface and troposphere there and therefore the increased infrared radiative energy trapping.

308   Owing to the dependence of infrared radiative transfer on clouds and water vapor, which have

309   substantial spatial structure in the low latitudes, the greenhouse gas forcing is non-uniform in the

310   tropics, being greater in the relatively drier tropical domains. For short-lived gases, the

311   concentrations themselves are not globally uniform so there tends to be a distinct spatial

312   character to the resulting forcing, e.g., stratospheric ozone, whose forcing is confined essentially

313   to the mid-to-high latitudes, and tropospheric ozone whose forcing is confined to tropical to

314   midlatitudes. For aerosols, which are even more short-lived than ozone, the forcing has an even

315   more localized structure (see also Box 5.3). However, although tropospheric ozone and aerosol

316   forcing are maximized near the source regions, the contribution to the global forcing from

317   remote regions is not negligible. The natural factors, namely solar irradiance changes and

318   stratospheric aerosols from volcanic eruptions, exert a forcing that is global in scope.

319

320   In terms of the transient changes in the climate system, it is also important to consider the

321   temporal evolution of the forcings. For well-mixed greenhouse gases, the evolution over the past

322   century, and in particular the past four decades, is very well quantified because of reliable and

323   robust observations. However, for the other forcing agents, such as aerosols, there are

324   uncertainties concerning their evolution that can affect the inferences about the resulting surface

325   and atmospheric temperature trends. Stratospheric ozone changes, which have primarily occurred

326   since ~1980, are slightly better known than tropospheric ozone and aerosols. For solar irradiance

327   and land-surface changes, the knowledge of the forcing evolution over the past century is poorly

328   known. Only in the past five years have climate models included time varying estimates of a

329   subset of the forcings that affect the climate system. In particular, current models typically

330   include GHGs, ozone, sulfate aerosol direct effects, solar influences, and volcanoes. Some very

331   recent model simulations also include time-varying effects of black carbon and land use change.

332   Other forcings either lack sufficient physical understanding or adequate global time- and space-

333   dependent datasets to be included in the models at this time. As we gain more knowledge of

334   these other forcings and are better able to quantify their space- and time-evolving characteristics,

335   they will be added to the models used by groups around the world. Experience with these models

336   so far has shown that the addition of more forcings generally tends to improve the realism and

337   details of the simulations of the time evolution of the observed climate system (e.g., Meehl et al.,

338   2004).

339

340   Whether the climate system is responding to internally generated variations in the atmosphere

341   itself, to atmosphere-ocean-land-surface coupling, or to externally applied forcings by natural

342   and/or anthropogenic factors, there are feedbacks that arise which can play a significant role in

343   determining the changes in the vertical and horizontal thermal structure. These include changes

344   in the hydrologic cycle involving water vapor, clouds, sea-ice, and snow, which by virtue of their

345   strong interactions with solar and longwave radiation, amplify the effects of the initial

346   perturbation (Stocker et al., 2001; NRC, 2003) in the heat balance, and thus influence the

347   response of the climate system. Convection and water vapor feedback, and cloud feedback in

348   particular, are areas of active observational studies; they are also being pursued actively in

349   climate modeling investigations to increase our understanding and reduce uncertainties

350   associated with those processes.

351

352   **1.3 Stratospheric forcing and related effects**

353

354   Observed changes in the stratosphere in recent decades have been large and several recent

355   studies have investigated the causes. WMO (2003) and Shine et al. (2003) conclude that the

356   observed vertical profile of cooling in the global, annual-mean stratosphere (from the tropopause

357   up into the upper stratosphere) can, to a substantial extent, be accounted for in terms of the

358   known changes that have taken place in well-mixed greenhouse gases, ozone, and water vapor

359   (Figure 1.2). Even at the zonal, annual-mean scale, the lower stratosphere temperature trend is

360   discernibly influenced by the changes in the stratospheric gases (Ramaswamy and Schwarzkopf,

361   2002; Langematz et al., 2003). In the tropics, there is considerable uncertainty about the

362   magnitude of the lower stratospheric cooling (Ramaswamy et al., 2001). In the high northern

363   latitudes, the lower stratosphere becomes highly variable both in the observations and model

364   simulations, especially during winter, such that causal attribution is difficult to establish. In

365   contrast, the summer lower stratospheric temperature changes in the Arctic and the springtime

366   cooling in the Antarctic can be attributed in large part to the changes in the greenhouse gases

367   (WMO, 2003; Schwarzkopf and Ramaswamy, 2002).
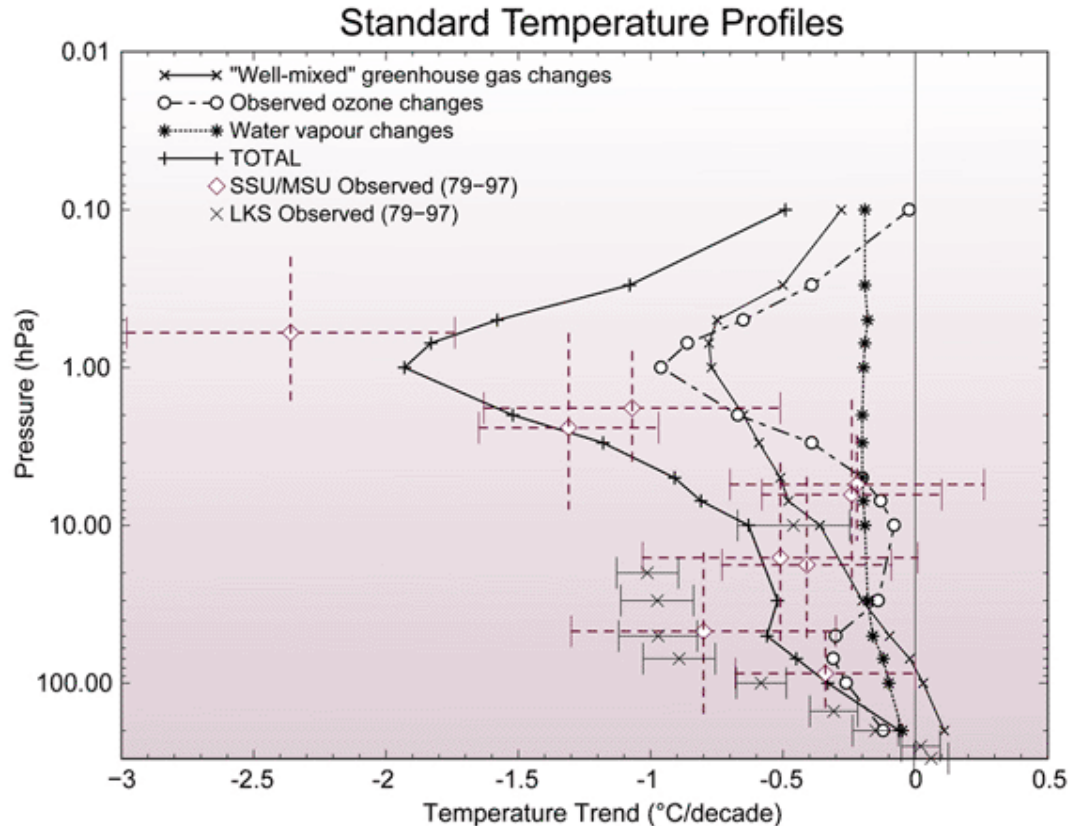
368

369 Figure 1.2. Global- and annual-mean temperature change over the 1979-1997 period in the stratosphere.
370 Observations: LKS (radiosonde), SSU and MSU (satellite) data.
371 Vertical bars on satellite data indicate the approximate span in altitude from where the signals originate, while the
372 horizontal bars are a measure of the uncertainty in the trend. Computed: effects due to increases in well-mixed
373 gases, water vapor, and ozone depletion, and the total effect (Shine et al., 2003).

374

375   Owing to the cooling of the lower stratosphere, there is a decreased infrared emission from the

376   stratosphere into the troposphere (Ramanathan and Dickinson, 1979; WMO, 1999), leading to a

377   radiative heat deficit in the upper troposphere, and a tendency for the upper troposphere to cool.

378   In addition, the depletion of ozone in the lower stratosphere can result in ozone decreases in the

379   upper troposphere due to reduced transport from the stratosphere (Mahlman et al., 1994). This

380   too affects the heat balance in the upper troposphere. Further, lapse rate near the tropopause can

381   be affected by changes in radiatively active trace constituents such as methane (WMO, 1986;

382   WMO/SROC Report, 2004).

383

384   The height of the tropopause (the boundary between the troposphere and stratosphere) is

385   determined by a number of physical processes that make up the integrated heat content of the

386   troposphere and the stratosphere. Changes in the heat balance within the troposphere and/or

387   stratosphere can consequently affect the tropopause height. For example, when a volcanic

388   eruption puts a large aerosol loading into the stratosphere where the particles absorb solar and

389   longwave radiation and produce stratospheric heating and tropospheric cooling, the tropopause

390   height shifts downward. Conversely, a warming of the troposphere moves the tropopause height

391   upward (e.g., Santer et al., 2003). Changes in tropopause height and their potential causes will be

392   discussed further in Chapter 5.

393

394   The episodic presence of volcanic aerosols affects the equator-to-pole heating gradient, both in

395   the stratosphere and troposphere. Temperature gradients in the stratosphere or troposphere can

396   affect the state of the polar vortex in the northern latitudes, the coupling between the stratosphere

397   and troposphere, and the propagation of temperature perturbations into the troposphere and to the

398   surface. This has been shown to be plausible in the case of perturbations due to volcanic aerosols

399   in observational and modeling studies, leading to a likely causal explanation of the observed

400   warming pattern seen in northern Europe and some other high latitude regions in the first winter

401   following a tropical explosive volcanic eruption (Jones et al., 2003; Robock and Oppenheimer,

402   2003; Shindell et al., 2001; Stenchikov et al., 2002). Ozone and well-mixed greenhouse gas

403   changes in recent decades can also affect stratosphere-troposphere coupling (Thompson and

404   Solomon, 2002; Gillett and Thompson, 2003), propagating radiatively-induced temperature

405   perturbations from the stratosphere to the surface in the high latitudes during winter.

406

407   **1.4 Simulated responses in vertical temperature profile to different external forcings**

408   Three-dimensional computer models of the coupled global atmosphere-ocean-land surface

409   climate system have been used to systematically analyze the expected effects of different

410   forcings on the vertical structure of the temperature response and compare these with the

411   observed changes (e.g., Santer et al., 1996; 2003; Hansen et al., 2002). A climate model can be

412   run with time-varying observations of just one forcing over the 20th Century to study the

413   temperature response in the vertical. Then, by running more single forcings, a picture emerges

414   concerning the effects of each one individually. The model can then be run with a combination

415   of forcings to determine the degree to which the simulation resembles the observations made in

416   the 20th Century. Note that a linear additivity of responses, while approximately valid for certain

417   combinations of forcings, need not hold in general (Ramaswamy and Chen, 1997; Hansen et al.,

418   1997; Santer et al., 2003; Shine et al., 2003). To first order, models are able to reproduce the time

419   evolution of globally averaged surface air temperature over the 20th Century, with the warming

420   in the first half of the century generally ascribed to natural forcings (mainly volcanoes and solar)

421   or unforced variations, and the warming in the late 20th Century mostly due to human-induced

422   increases of GHGs (Stott et al., 2000; Mitchell et al., 2001; Meehl et al., 2003; 2004; Broccoli et

423   al., 2003). Such modeling studies used various observed estimates of the forcings, but

424   uncertainties remain regarding details of such factors as solar variability (Frohlich and Lean,

425   2004), historical volcanic forcing (Bradley, 1988), and aerosols (Charlson et al., 1992; Anderson

426   et al., 2003).

427

428   Analyses of model responses to external forcings also require consideration of the internal

429   variability of the climate system for a proper causal interpretation of the observed surface

430 temperature record. For example, the mid-1970s saw the beginning of a significant increase of

431 global temperatures that was associated with an apparent regime shift in the Pacific (Trenberth

432 and Hurrell, 1994). While it has been argued that the warming could have been a delayed

433 response to a regime shift due to the heat capacity of the ocean (Lindzen and Giannitsis, 2002),

434 this increase in temperature starting in the 1970s is also simulated as a response due to changes

435 in external forcing in the models noted above. The relationship between external forcing and

436 internal decadal variability of the climate system (i.e., can the former influence the latter, or are

437 they totally independent) is an intriguing research problem that is being actively studied.

438

439 In addition to the analyses of surface temperatures outlined above, climate models can also show

440 the expected effects of different forcings on temperatures in the vertical. For example using

441 simplified ocean representations for equilibrium $2XCO_2$, Hansen et al. (2002) show that changes

442 of various anthropogenic and natural forcings produce different patterns of temperature change

443 horizontally and vertically. Hansen et al. (2002) also show considerable sensitivity of the

444 simulated vertical temperature response to the choice of ocean representation, particularly for the

445 GHG-only and solar-only cases. For both of these cases, the "Ocean A" configuration (SSTs

446 prescribed according to observations) lacks a clear warming maximum in the upper tropical

447 troposphere, thus illustrating that there could be some uncertainty in our model-based estimates

448 of the upper tropospheric temperature response to GHG forcing (see Chapter 5).

449

450 An illustration of the effects of different forcings on the trends in atmospheric temperatures at

451 different levels from a climate model with time-varying forcings over the latter part of the 20th

452 Century is shown in Figure 1.3. Here, the temperature changes are calculated over the time

453   period of 1979-1999, and are averages of four-member ensembles. As in Hansen et al., this

454   model, the NCAR/DOE Parallel Climate Model (PCM, e.g., Meehl et al., 2004) shows warming

455   in the troposphere and cooling in the stratosphere for an increase of GHGs, warming through

456   most of the stratosphere and a slight cooling in the troposphere for volcanic aerosols, warming in

457   a substantial portion of the atmosphere for an increase in solar forcing, warming in the

458   troposphere from increased tropospheric ozone and cooling in the stratosphere due to the

459   decrease of stratospheric ozone, and cooling in the troposphere and slight warming in the

460   stratosphere from sulfate aerosols. The multiple-forcings run shows the net effects of the

461   combination of these forcings as a warming in the troposphere and a cooling in the stratosphere.

462   Note that these simulations may not provide a full accounting of all factors that could affect the

463   temperature structure, e.g., black carbon aerosols, land use change (Ramaswamy et al., 2001;

464   Hansen et al., 1997; 2002; Pielke, 2001; NRC, 2005; Ramanathan et al., 2005).

PCM Simulations of Zonal-Mean Atmospheric Temperature Charge
Total linear change computed over January 1958 to December 1999
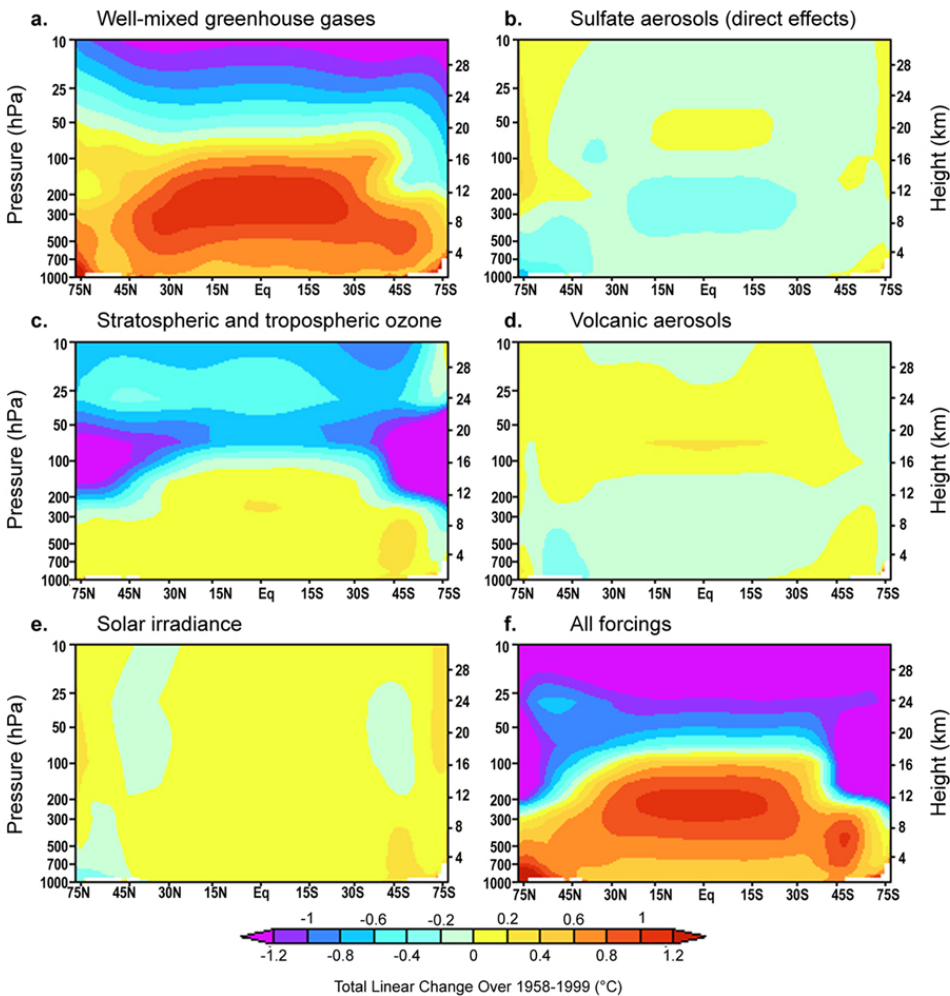
465

466
467 Figure 1.3. PCM simulations of the vertical profile of temperature change due to various forcings, and the effect due
468 to all forcings taken together (after Santer et al., 2000).
469

470 The magnitude of the temperature response for any given model is related to its climate

471 sensitivity. This is usually defined either as the equilibrium warming due to a doubling of CO_2

472 with an atmospheric model coupled to a simple slab ocean, or the transient climate response

473 (warming at time of CO_2 doubling in a 1% per year CO_2 increase experiment in a global coupled

474   model). The climate sensitivity varies among models due to a variety of factors (Cubasch et al.,

475   2001; NRC, 2004).

476

477   The important conclusion here is that representations of the major relevant forcings are important

478   to simulate 20th Century temperature trends since different forcings affect temperature differently

479   at various levels in the atmosphere.

480

481   **1.5 Physical factors, and temperature trends at the surface and in the troposphere**

482

483   Tropospheric and surface temperatures, although linked, are separate physical entities (Trenberth

484   et al., 1992; Hansen et al., 1995; Hurrell and Trenberth, 1996; Mears et al., 2003). Insight into

485   this point comes from an examination of the correlation between anomalies in the monthly-mean

486   surface and tropospheric temperatures over 1979-2003 (Figure 1.4). The correlation coefficients

487   between   monthly   surface   and   tropospheric   temperature   anomalies   (as   represented   by

488   temperatures   derived   from   MSU   satellite   data)   reveal   very   distinctive   patterns,   with   values

489   ranging   from   less   than   zero   (implying   poor   vertical   coherence   of   the   surface   and   tropospheric

490   temperature anomalies) to over 0.9. The highest correlation coefficients (>0.75) are found across

491   the   middle   and   high   latitudes   of   Europe,   Asia,   and   North   America,   indicating   a   strong

492   association between the surface and tropospheric monthly temperature variations. Correlations

493   are   generally   much   less   (~0.5)   over   the   tropical   continents   and   the   North   Atlantic   and   North

494   Pacific Oceans. Correlations less than 0.3 occur over the tropical and southern oceans and are

495   lowest (<0.15) in the tropical western Pacific. Relatively high correlation coefficients (>0.6) are

496   found over the tropical eastern Pacific where the ENSO signal is large and the sea-surface

497   temperature fluctuations influence the atmosphere significantly.
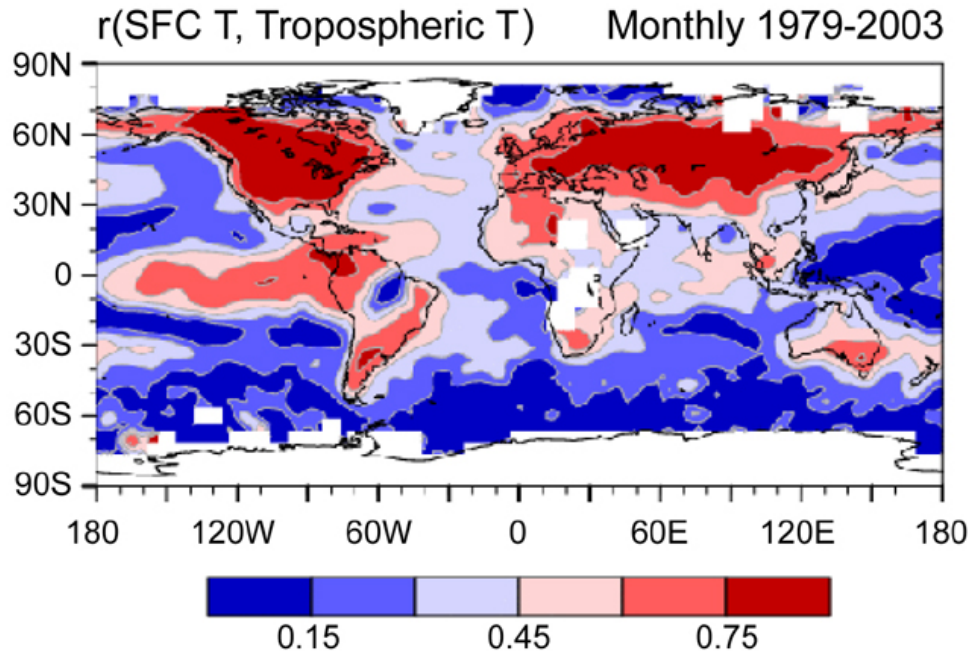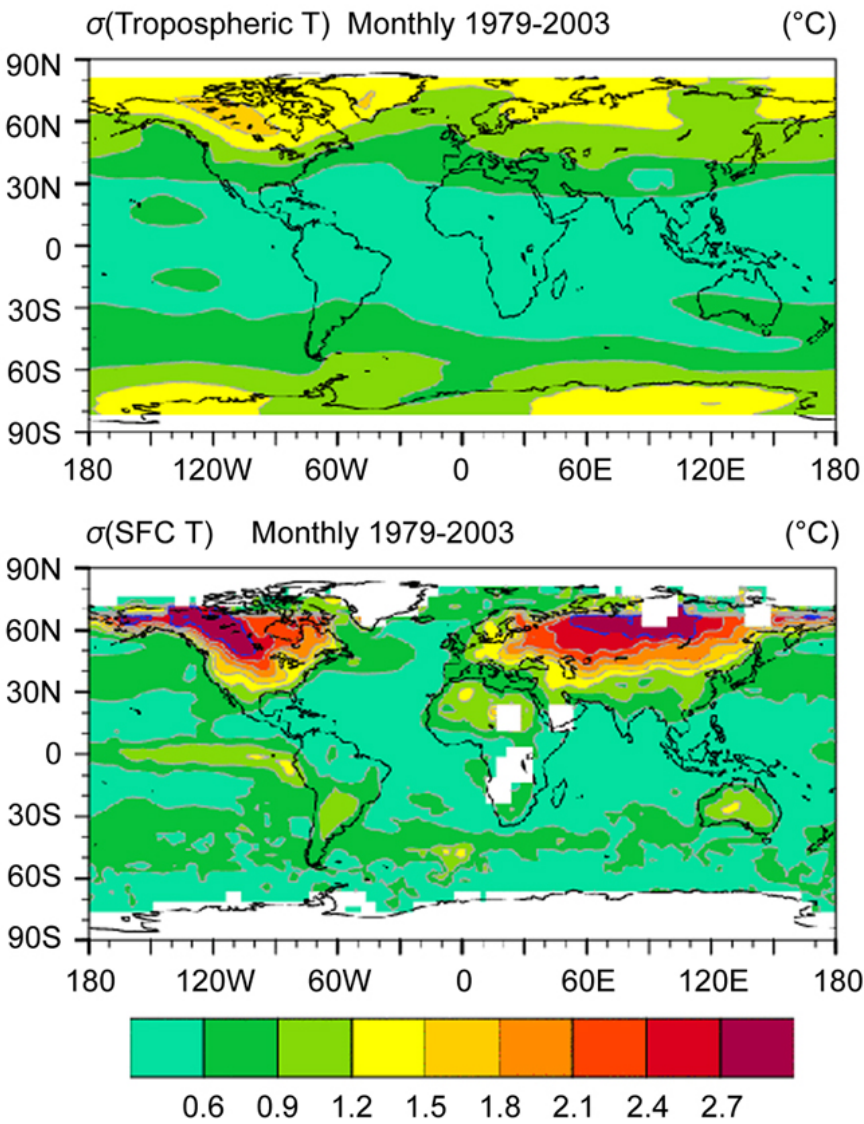

498

499   Figure 1.4. Gridpoint correlation coefficients between monthly surface and tropospheric temperature anomalies over
500   1979-2003. The tropospheric temperatures are derived from MSU satellite data (Christy et al., 2003).

501

502   Differences between the surface and tropospheric temperature records are found where there is

503   some degree of decoupling between the layers of the atmosphere. For instance, as discussed

504   earlier, over portions of the subtropics and tropics, variations in surface temperature are

505   disconnected from those aloft by a persistent trade-wind inversion. Shallow temperature

506   inversions are also commonly found over land in winter, especially in high latitudes on sub-

507   seasonal timescales, so that there are occasional large differences between monthly surface and

508   tropospheric temperature anomalies.

509

510   More important than correlations for trends, however, is the variability of the two temperature

511   records, assessed by computing the standard deviation of the measurement samples of each

512   record (Figure 1.5). The figure exhibits pronounced regional differences in variability between

513   the surface and tropospheric records. Standard deviations also help in accounting for the

514   differences in correlation coefficients, because they yield information on the size and persistence

515   of the climate signal relative to the noise in the data. For instance, large variations in eastern

516   tropical Pacific sea surface temperature associated with ENSO dominate over measurement

517   uncertainties, as do large month-to-month swings in surface temperatures over extratropical

518   continents.

519

520 Figure 1.5. Standard deviations of monthly mean temperature anomalies from the surface and tropospheric
521 temperature records over 1979-2003. The tropospheric temperatures are derived from MSU satellite data (Christy et
522 al., 2003).

523

524 The largest variability in both surface and tropospheric temperature is over the Northern

525 Hemisphere continents. The standard deviation over the oceans in the surface data set is much

526 smaller than over land except where the ENSO phenomenon is prominent. The standard

527 deviations of tropospheric temperature, in contrast, exhibit less zonal variability. Consequently,

528 the standard deviations of the monthly tropospheric temperatures are larger than those of the

529 surface data by more than a factor of two over the North Pacific and North Atlantic. Over land,

530 tropospheric temperatures exhibit slightly less variability than surface temperatures. These

531 differences in variability are indicative of differences in physical processes over the oceans

532 versus the continents. Of particular importance are the roles of the land surface and ocean as the

533 lower boundary for the atmosphere and their very different abilities to store heat, as well as the

534 role of the atmospheric winds that help to reduce regional differences in tropospheric

535 temperature through the movement of heat from one region to another.

536

537 Over land, heat penetration into the surface involves only the upper few meters, and the ability of

538 the land to store heat is low. Therefore, land surface temperatures vary considerably from

539 summer to winter and as cold air masses replace warm air masses and vice versa. The result is

540 that differences in magnitude between surface and lower-atmospheric temperature anomalies are

541 relatively small over the continents: very warm or cold air aloft is usually associated with very

542 warm or cold air at the surface. In contrast, the ability of the ocean to store heat is much greater

543 than that of land, and mixing in the ocean to typical depths of 50 meters or more considerably

544 moderates the sea surface temperature response to cold or warm air. Over the northern oceans,

545 for example, a very cold air mass (reflected by a large negative temperature anomaly in the

546   tropospheric record) will most likely be associated with a relatively small negative temperature

547   anomaly at the sea surface. This is one key to understanding the differences in trends between

548   the two records.

549

550   Long-term changes in the atmospheric circulation can be reflected by trends in indices of

551   patterns (or modes) of natural climate variability such as ENSO, the North Atlantic Oscillation

552   (NAO; also known as the Northern Hemisphere annular mode, or NAM), and the Southern

553   Hemisphere (SH) annular mode (SAM). The exact magnitudes of the index trends depend on the

554   period of time examined. Over the past several decades, for instance, changes in atmospheric

555   circulation (reflected by a strong upward trend in indices of the NAO) have contributed to a Cold

556   Ocean Warm Land (COWL) surface temperature pattern over the Northern Hemisphere (Hurrell,

557   1996; Thompson and Wallace, 2000). In the lower atmosphere, winds blowing from ocean to

558   land to ocean are much stronger than at the surface, and this moderating influence of the winds

559   contributes to less east-west variability in the tropospheric data (Figure 1.5). Thus, the recent

560   warm anomalies over the continents are roughly cancelled by the cold anomalies over the oceans

561   in the tropospheric dataset. This is not the case for the surface temperature record, which is

562   dominated by the warmth over the continents. The result is that the changes in the Northern

563   Hemisphere (NH) atmospheric circulation over the past few decades have produced a significant

564   difference in surface and tropospheric temperature trends (Hurrell and Trenberth, 1996).

565   Similarly, Thompson and Solomon (2002) showed that recent tropospheric temperature trends at

566   high southern latitudes were related to changes in the SAM.

567

568   Physical differences between the two measures of temperature are also evident in their dissimilar

569   responses to volcanic eruptions and ENSO (e.g., Santer et al. 2000). These phenomena have a

570   greater effect on tropospheric than surface temperature, especially over the oceans (Jones, 1994).

571   However, changes in ENSO over the past several decades do not explain long-term changes in

572   tropical tropospheric temperatures (Hegerl and Wallace, 2002). Changes in concentrations of

573   stratospheric ozone could also be important, as the troposphere is cooled more by observed

574   ozone depletion than is the surface (Hansen et al., 1995 ; Ramaswamy et al., 1996). Another

575   contributing factor could be that at the surface, the daily minimum temperature has increased at a

576   faster rate than the daily maximum, resulting in a decrease in the diurnal temperature range over

577   many parts of the world (e.g., Easterling et al., 1997; Dai et al., 1999). Because of nighttime

578   temperature inversions, the increase in the daily minimum temperatures likely involves only a

579   shallow layer of the atmosphere that would not be evident in upper-air temperature records.

580   These physical processes provide indications of why trends in surface temperatures are expected

581   to be different from trends in the troposphere, especially in the presence of strong interannual

582   variability, even if both sets of measurements were perfect. Of course they are not, as described

583   in more detail in Chapter 2, which deals with the strengths and limitations of current observing

584   systems. An important issue implicit in Figure 1.5 is that of spatial sampling, and the

585   accompanying caveats about interpretation of the truly global coverage provided by satellites

586   versus the incomplete space and time coverage offered by radiosondes. These are discussed in

587   depth in Chapters 2 and 3.

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Chapter 2
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9   What kinds of atmospheric temperature variations can the current observing systems
10   detect and what are their strengths and limitations, both spatially and temporally?
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42   Findings and Recommendations

43   •   The observing systems available for this report are able to detect small surface and upper air

44   temperature variations from year to year, for example, those caused by El Niño or volcanic

45   eruptions.

46

47   •   The data from these systems also have the potential to provide accurate trends in climate over

48   the last few decades (and over the last century for surface observations), once the raw data are

49   successfully adjusted for changes over time in observing systems, practices, and micro-climate

50   exposure to produce usable climate records. Measurements from all systems require such

51   adjustments and this report relies on adjusted datasets.

52

53   •   Adjustments to the land surface temperature record have been sufficiently successful that

54   trends are reasonably similar on large (e.g., continental) scales, despite the fact that spatial

55   sampling is uneven and some errors undoubtedly remain. This conclusion holds to a lesser extent

56   for the ocean surface record, which suffers from more serious sampling problems and changes in

57   observing practice.

58

59   •   Adjustments for changing instrumentation are most challenging for upper-air datasets. While

60   these show promise for trend analysis, it is likely that current upper-air climate records give

61   reliable indications of directions of change (e.g. warming of troposphere, cooling of stratosphere)

62   but some questions remain regarding the precision of the measurements.

63   •   Upper-air datasets have been subjected to less scrutiny than surface datasets.

64   • Adjustments are complicated, sometimes as large as the trend itself, involve expert

65   judgments, and cannot be stringently evaluated because of lack of traceable standards.

66   • Unlike surface trends, reported upper-air trends vary considerably between research

67   teams beginning with the same raw data owing to their different decisions on how to

68   remove non-climatic factors.

69   • The diurnal cycle, which must be factored into some adjustments for satellite data, is

70   well observed only by surface observing systems.

71   • No available observing system has reference stations or multi-sensor instrumentation

72   that would provide stable calibration over time.

73   • Most observing systems have not retained complete metadata describing changes in

74   observing practices which could be used to identify and characterize non-climatic

75   influences.

76

77 • Relevant satellite datasets measure broad vertical layers and cannot reveal the detailed vertical

78 structure of temperature changes, nor can they completely isolate the troposphere from the

79 stratosphere. However, retrieval techniques can be used both to approximately isolate these

80 layers and to check for vertical consistency of trend patterns. Consistency between satellite and

81 radiosonde data can be tested by proportionately averaging radiosonde profiles.

82 • Reanalyses and other multi-system products have the potential for addressing issues of

83 surface and atmospheric temperature trends by making better use of available information and

84 allowing analysis of a more comprehensive, internally consistent, and spatially and temporally

85   complete set of climate variables. At present, however, they contain biases, especially in the

86   stratosphere, that affect trends and that cannot be readily removed because of the complexity of

87   the data products.

88

89   •   There are as yet under-exploited data archives with potential to contribute to our

90   understanding of past changes, and new observing systems that may improve estimates of future

91   changes if designed for long-term measurement stability and operated for sufficient periods.

92

93

94   *Recommendation: Current and future observing systems should adhere to the principles for*

95   *climate observations adopted internationally under the Framework Convention on Climate*

96   *Change and documented in "NRC 2000b" and the "Strategic Plan for the U.S. Climate Change*

97   *Science Program (2003)" to significantly mitigate the limitations listed above.*

98

99   *Recommendation: The ability to fully and accurately observe the diurnal cycle should be an*

100 *important consideration in the design and implementation of new observing systems.*

101

102 *Recommendation: When undertaking efforts to retrieve data it is important to also to collect*

103 *detailed metadata which could be used to reduce ambiguity in the timing, sign and magnitude of*

104 *non-climatic influences in the data.*

105

106   *Recommendation: New climate-quality reanalysis efforts should be strongly encouraged and*

107   *specifically designed to minimize small, time-dependent biases arising from imperfections in*

108   *both data and forecast models.*

109

110   *Recommendation: Some largely overlooked satellite datasets should be reexamined to try to*

111   *extend, fortify or corroborate existing microwave-based temperature records for climate*

112   *research, e.g. microwave data from NEMS (1972) and SCAMS (1975), infrared from the HIRS*

113   *suite and radio occultation from GPS.*

114

115

116

117   **1. MAIN OBSERVING SYSTEMS AND SYNTHESIS DATA PRODUCTS**

118

119   Temperature is measured in three main ways; (1) *in situ,* where the sensor is immersed in the

120   substance of interest; (2) by *radiative emission*, where a remote sensor detects the intensity or

121   brightness of the radiation emanating from the substance; and (3) *radiative transmission*, where

122   radiation is modified as it passes through the substance in a manner determined by the

123   substance's temperature. All observations contain some level of random measurement error,

124   which is reduced by averaging; bias, which is not reduced by averaging; and sampling errors (see

125   Appendix).

126

127   a) Surface and near-surface air temperatures

128   Over land, "near-surface" air temperatures are those commonly measured about 1.5 to 2.0 meters

129   above the ground level at official weather stations, at sites run for a variety of scientific purposes,

130   and by volunteer (cooperative) observers (e.g., Jones and Moberg, 2003). These stations often

131   experience relocations, changes in instrumentation and/or exposure and changing observing

132   practices all of which can introduce biases into their long-term records. These changes are often

133   undocumented.

134

135   "Near-surface" air temperatures over the ocean ("Marine Air Temperatures" or MATs) are

136   measured by ships and buoys at various heights from 2 to more than 25m, with poorer coverage

137   than over land (e.g., Rayner et al., 2003). To avoid the contamination of daytime solar heating of

138   the ships' surfaces that may affect the MAT, it is generally preferred to limit these to night MAT

139   (NMAT) readings only. Observations of the water temperature near the ocean surface or "Sea

140   Surface Temperatures" (SSTs) are widely used and are closely tied to MATs; ships and buoys

141   measure SSTs within a few meters below the surface.

142

143   Incomplete geographic sampling, changing measurement methods, and land-use changes all

144   introduce errors into surface temperature compilations. The spatial coverage, indicated in Figure

145   2.1, is far from uniform over either land or ocean areas. The southern oceans, polar regions and

146   interiors of Brazil and Africa are not well sampled by in-situ networks. However, creating

147   global surface temperature analyses involves not only merging land and ocean data but also

148   considering how best to represent areas where there are few or no observations. The most

149   conservative approach is to use only those grid boxes with data, thus avoiding any error

150   associated with interpolation. Unfortunately, the areas without data are not evenly or randomly

151   distributed around the world, leading to considerable uncertainties in the analysis, though it is

152   possible to make an estimate of these uncertainties. Using the conservative approach, the tropical

153   land surface areas would be under-represented, as would the southern ocean. Therefore,

154   techniques have been developed to interpolate data to some extent into surrounding data-void

155   regions. A single group may produce several different such datasets for different purposes. The

156   choice may depend on whether the interest is a particular local region, the entire globe, or use of

157   the dataset with climate models (Chapter 5). Estimates of global and hemispheric scale averages

158   of near-surface temperatures generally begin around 1860 over both land and ocean.

Global Temperature Observations
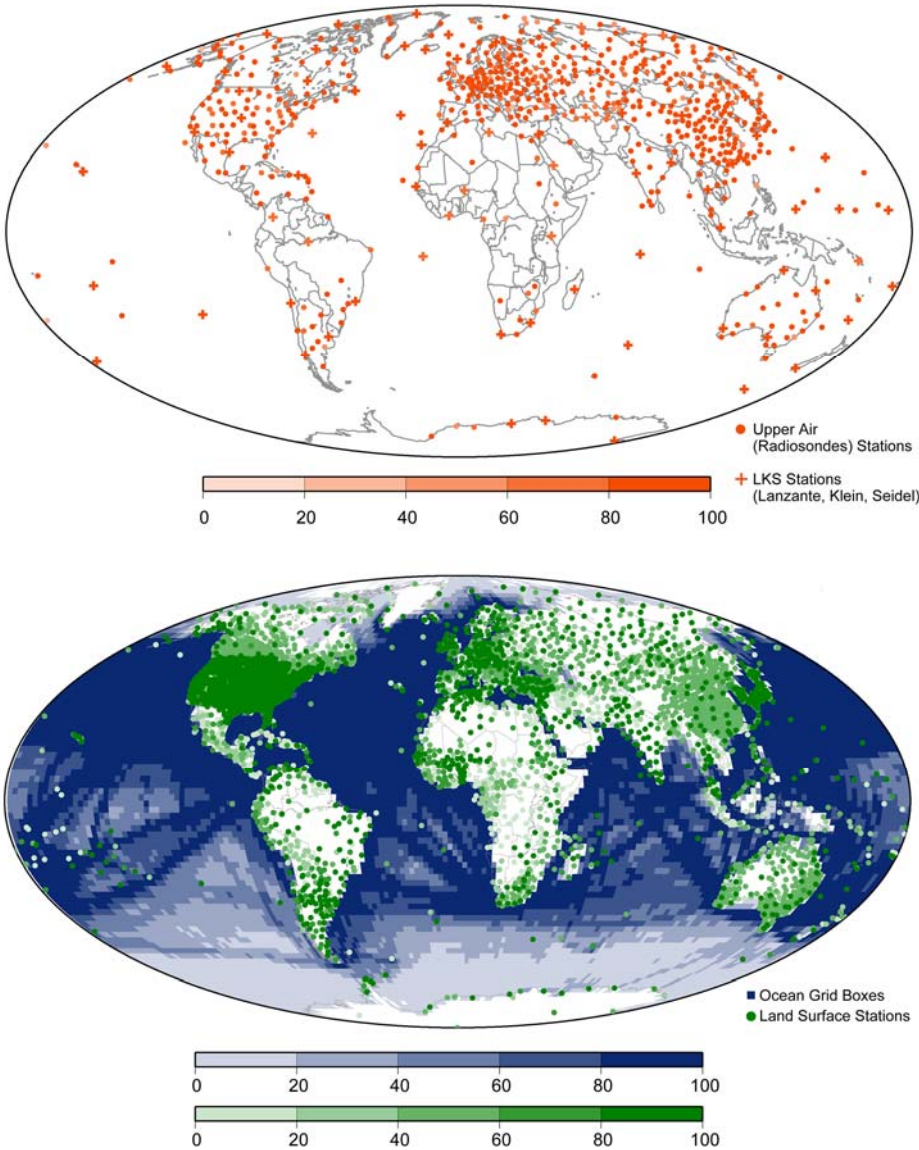
159

160   Figure 2.1 Top: Location of radiosonde stations used in the HadAT upper air dataset with those also in the LKS as
161   crosses. Bottom: Distribution of land stations (green) and SST observations (blue) reporting temperatures used in the
162   surface temperature datasets over the period 1979-2004. Darker colors represent locations for which data were
163   reported with greater frequency.
164   See chapter 3 for definitions of datasets.

165

166   Datasets of near-surface land and ocean temperatures have traditionally been derived from *in-situ*

167   thermometers. With the advent of satellites, some datasets now combine both *in-situ* and

168   remotely sensed data (Reynolds et al., 2002; Smith and Reynolds, 2005), or use exclusively

169   remotely sensed data (Kilpatrick et al., 2001) to produce more geographically complete

170   distributions of surface temperature. Because the satellite sensors measure infrared or microwave

171   emission from the earth's surface (a "skin" typically tens of microns thick that may have a

172   temperature different from either the air above or material at greater depths), calculations are

173   required to convert the skin temperature into the more traditional near-surface air or SST

174   observation (in this context SSTs are called "bulk sea surface temperatures", Chelton, 2005.)

175   Typically, in-situ observations are taken as "truth" and satellite estimates (which may be affected

176   by water vapor, clouds, volcanic aerosols, etc.) are adjusted to agree with them (Reynolds, 1993.)

177   With continued research, datasets with surface temperatures over land, ice, and ocean from

178   infrared and microwave sensors should provide expanded coverage of surface temperature

179   variations (e.g., Aires et al., 2004).

180

181   Sampling errors in ship and buoy SST data typically contribute more to large-scale averages than

182   random measurement errors as shown in Smith and Reynolds (2004), especially as the

183   temperature record extends backward in time. Biases depend on observing method. Most ship

184   observations since the 1950s were made from insulated buckets, hull contact sensors, and engine

185   intake temperatures at depths of one to several meters. Historic correction of ship data prior to

186   1942 is discussed in (Folland and Parker, 1995) and bias and random errors from ships are

187   summarized by (Kent and Taylor, 2004) and (Kent and Challenor, 2004). They report that engine

188   intake temperatures are typically biased 0.1-0.2ºC warmer than insulated buckets. This is

189   primarily due to engine room heating of the water temperatures although there is also

190   evaporative cooling of the water in the insulated buckets. Hull contact sensors are the most

191   accurate though much less common. The bias correction of the ship SST data (Kent and Kaplan,

192   2004) requires information on the type of measurement (e.g. insulated bucket, etc.) which

193   becomes more difficult to determine prior to 1990s due to incomplete documentation. Kent and

194   Kaplan (2005) also found that insulated bucket temperatures may be to cold by 0.12 to 0.16ºC.

195   When the bucket bias is used, engine intake temperatures in the mid-to-late 1970s and the 1980s

196   were found to be smaller than that suggested by previous studies, ranging from 0.09 to 0.18ºC. In

197   addition, their study indicates that engine intake SSTs may have a cold bias of -0.13ºC in the

198   early 1990s. The reliability of these biases are subject to revision due to small sample sizes that

199   sample sizes for these comparisons tend to be small with large random errors. Buoy observations

200   became more plentiful following the start of the Tropical Ocean Global Atmosphere (TOGA)

201   Program (McPhaden, 1995) in 1985. These observations are typically made by an immersed

202   temperature sensor or a hull contact sensor, and are more accurate because they do not have the

203   bias errors of ship injection or insulated bucket temperatures.

204

205   The global surface air temperature data sets used in this report are to a large extent based on data

206   readily exchanged internationally, e.g., through CLIMAT reports and the WMO publication

207   *Monthly Climatic Data for the World*. Commercial and other considerations prevent a fuller

208   exchange, though the United States may be better represented than many other areas. In this

209   report we present three global surface climate records, created from available data by NASA

210   Goddard Institute for Space Studies (GISS), NOAA National Climatic Data Center

211   (NOAA/NCDC) and the cooperative project of the U.K. Hadley Centre and the Climate

212   Research Unit of the University of East Anglia (HadCRUT2v). These will be identified as T_{Sfc-G},

213   T_{Sfc-N} and T_{Sfc-U} respectively.

214

215   <u>b) Atmospheric "upper air" temperatures</u>

216   *1. Radiosondes*

217   Radiosonde or balloon-based observations of atmospheric temperature are *in-situ* measurements

218   as the thermometer (often a thermistor or a capacitance-based sensor), suspended from a balloon,

219   is physically carried through the atmospheric column. Readings are radio-transmitted back to a

220   data recorder. Balloons are released once or twice a day (00 and/or 12 Coordinated Universal

221   Time or UTC) at about 1,000 stations around the globe, many of which began operations in the

222   late 1950s or 1960s. These sites are unevenly distributed, with only the extratropical northern

223   hemisphere land areas and the Western Pacific Ocean/Indonesia/Australia region being well-

224   sampled in space and time. Useful temperature data can be collected from near the surface

225   through the lower and middle stratosphere (though not all balloons survive to these heights).

226   Radiosonde data in the first hundred meters or so above the surface are sometimes erroneous if

227   the sensors have not been allowed to reach equilibrium with the atmosphere before launch, and

228   may not be representative of regional conditions, due to microclimatic and terrain effects.

229

230   Although most radiosonde data are transmitted to meteorological centers around the world and

231   archived, in practice many soundings do not reach this system and are collected later. No

232   definitive archive of radiosonde data exists, but several archives in the U.S. and abroad contain

233   nearly complete collections, though several different schemes have been employed for quality

234   control. To monitor climate, it is desirable to have a long, continuous record of measurements

235   from many well-distributed fixed sites. There are about 700 radiosonde stations that have

236   operated in the same location for at least three decades; many of these are clustered in a few

237   areas, further reducing the effective coverage (Figure 2.1). Thus, a dilemma exists for estimating

238   long-term changes: whether to use a smaller number of stations having long segments of

239   continuous records, or a larger number of stations with shorter records that do not always overlap

240   well. Various analysis groups have approached this differently (see Chapters 3 and 4).

241

242   Typically, radiosonde-based datasets are developed for specific atmospheric pressure surfaces

243   known as "mandatory reporting levels" (Figure 2.2). Such data at discrete vertical levels provide

244   unique information for assessing changes in the structure of the atmosphere. Two such datasets

245   are featured in this report, The Hadley Centre Atmospheric Temperatures from the U.K.

246   (HadAT) and Radiosonde Atmospheric Temperatures Products for Assessing Climate

247   (RATPAC) from NOAA. A product such as $T_{850\text{-}300}$, for example, will be identified as $T_{850\text{-}300\text{-}U}$

248   and $T_{850\text{-}300\text{-}N}$ for HadAT and RATPAC respectively. [1]

---

[1] A third radiosonde dataset was generated by comparing radiosonde observations against the first-guess field of the ERA-40 simulation forecast model (Haimberger, 2004). Adjustments were applied when the relative difference between the radiosonde temperatures and the forecast temperatures changed by a significant amount. The data were not yet in final form for consideration in this report, although the tropospheric values appear to have general agreement with HadAT and RATPAC

249
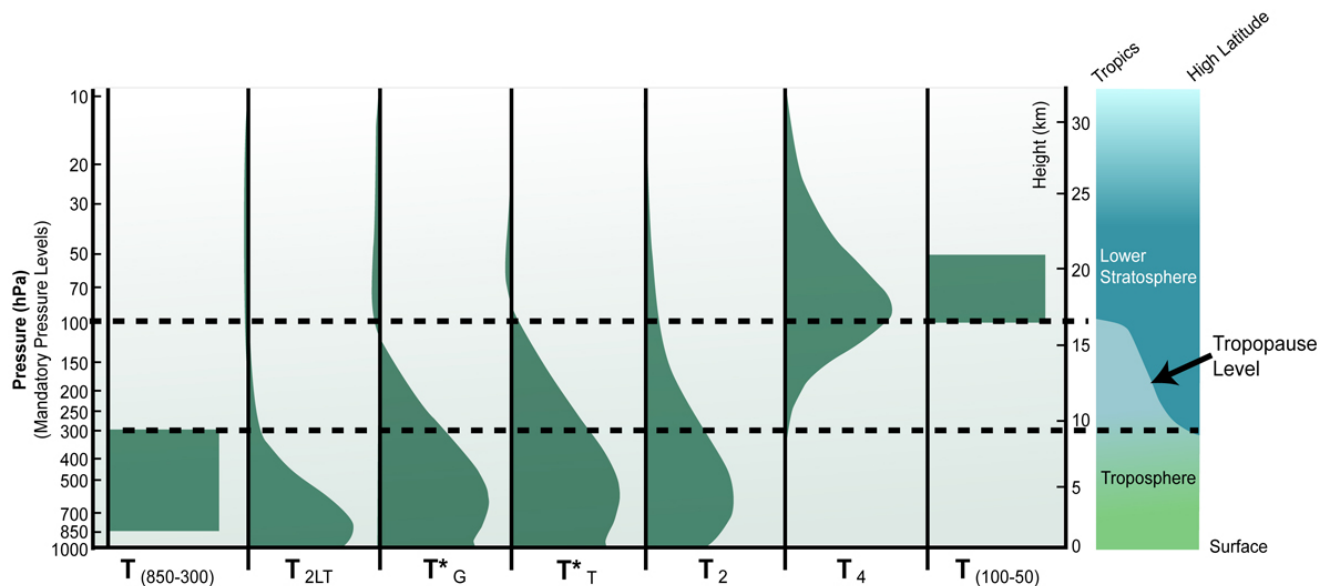
Figure 2.2 Terminology and vertical profiles for the temperature products referred to in this report. Radiosonde-based layer temperatures ($T_{850-300}$, T_{100-50}) are height-weighted averages of the temperature in those layers. Satellite-based temperatures (T_{2LT}, T_2, and T_4) are mass-weighted averages with varying influence in the vertical as depicted by the curved profiles, i.e., the larger the value at a specific level, the more that level contributes to the overall satellite temperature average. The subscript simply indicates the layer where 90% of the information for the satellite average originates.

Notes: (1) because radiosondes measure the temperature at discrete (mandatory) levels, their information may be used to create a temperature value that mimics a satellite temperature (Text Box 2.1), (2) layer temperatures vary from equator to pole so the pressure and altitude relationship here is based on the atmospheric structure over the conterminous U.S., (3) about 10% (5%) of the value of T_{2LT} (T_2) is determined by the surface character and temperature, (4) T^*_T and T^*_G are simple retrievals, being linear combinations of 2 channels, T_2 and T_4.

263   Throughout the radiosonde era there have been numerous changes in stations, types of

264   instrumentation, and data processing methods that can create data discontinuities. Because

265   radiosondes are expendable instruments, instruments are more easily changed than for the more

266   permanent surface sites. The largest discontinuities appear to be related to solar heating of the

267   temperature sensor and changes in design and/or data adjustments intended to deal with this

268   problem. These discontinuities have greatest impact at stratospheric levels (the stratosphere's

269   lower boundary is ~16 km in the tropics, dropping to < 10 km in the high latitudes, Figure 2.2),

270 where direct sunlight can cause radiosonde-measured temperatures to rise several $^{\circ}$C above

271 ambient temperatures. For example, when Australia and U.S. stations changed instrumentation to

272 Vaisala RS-80, processed stratospheric temperatures shifted downward by 1 to 3°C (Parker et al.,

273 1997, Christy et al., 2003). Many other sources of system-dependent bias exist (which often

274 affect the day and night releases differently), including icing of the sensors in regions of super-

275 cooled water, software errors in some radiosonde systems, poor calibration formulae, and

276 operator errors. Documentation of these many changes is limited, especially in the earlier

277 decades.

278

279   *2. Passive Satellite Instrumentation*

280 Unlike radiosondes, passive satellite observations of microwave and infrared brightness

281 temperatures sample thick atmospheric layers (and may include surface emissions), depicted as

282 weighting functions in Figure 2.2. These measurements may be thought of as bulk atmospheric

283 temperatures, as a single value describes the entire layer. Although this bulk measurement is less

284 informative than the detailed information from a radiosonde, horizontal coverage is far superior,

285 and consistency can be checked by comparing the appropriate vertical average from a radiosonde

286 station against nearby satellite observations (see Box 2.2). Furthermore, because there are far

287 fewer instrument systems than in radiosonde datasets, it is potentially easier to isolate and adjust

288 problems in the data.

289

290 The space and time sampling of the satellites varies according to the orbit of the spacecraft,

291 though the longer satellite datasets are based on polar orbiters. These spacecraft circle the globe

292   from pole to pole while maintaining a nominally constant orientation relative to the sun (sun-

293   synchronous). In this configuration, the spacecraft completes about 14 roughly north-south orbits

294   per day as the earth spins eastward beneath it, crosses the equator at a constant local time, and

295   provides essentially global coverage. Microwave measurements utilized in this report begin in

296   late 1978 with the TIROS-N spacecraft using a 4-channel radiometer (Microwave Sounding Unit

297   or "MSU") which was upgraded in 1998 to a 16-channel system (advanced MSU or "AMSU")

298   with better calibration, more stable station-keeping (i.e., the timing and positioning of the

299   satellite in its orbit – see discussion of "Diurnal Sampling" below), and higher spatial and

300   temporal sampling resolution.

301

302   Laboratory estimates of precision (random error) for a single MSU measurement are 0.25 °C.

303   Thus with 30,000 observations per day, this error is inconsequential for global averages. Of far

304   more importance are the time varying biases which arise once the spacecraft is in orbit; diurnal

305   drifting, orbital decay, intersatellite biases and calibration changes due to heating of the

306   instrument in space (see section 3 below.)

307

308   While bulk-layer measurements offer the robustness of a large-volume sample, variations within

309   the observed layer are masked. This is especially true for the layer centered on the mid-

310   troposphere (T_2) for which the temperatures of both lower stratospheric and tropospheric levels,

311   which generally show opposite variations, are merged (Figure 2.2). Three MSU/AMSU-based

312   climate records are presented in this report, prepared by Remote Sensing Systems (RSS) of Santa

313   Rosa, California, The University of Alabama in Huntsville (UAH), and The University of

314   Maryland (UMd). Subscripts identify the team, for example, T_2 will be listed as T_{2-R}, T_{2-A} and

315   T_{2-M} for RSS, UAH and UMd respectively.

316

317   Some polar orbiters also carry the Stratospheric Sounding Unit (SSU), an infrared sensor for

318   monitoring deep layer temperatures above about 15 km. SSU data have been important in

319   documenting temperature variations at higher elevations than observed by MSU instruments on

320   the same spacecraft (Ramaswamy et al., 2001). Generally, the issues that complicate the creation

321   of long-term MSU time series also affect the SSU, with the added difficulty that infrared

322   channels are more sensitive to variations in atmospheric composition (e.g., volcanic aerosols,

323   water vapor, etc.).

324

325   Future observing systems using passive-satellite methods include those planned for the National

326   Polar-orbiting Operational Environmental Satellite System (NPOESS) series: the microwave

327   sensors Conical scanning Microwave Imager/Sounder (CMIS) (which will succeed the Special

328   Sensor Microwave/Imager [SSM/I]), Special Sensor Microwave Imager/Sounder (SSMI/S) and

329   Advanced Technology Microwave Sounder (ATMS) (which will succeed the AMSU), and the

330   infrared sensor Cross-track Infrared Sounder (CrIS) (following the High-resolution Infrared

331   Radiation Sounder [HIRS]). Each of these will follow measuring strategies that are both similar

332   (polar orbit) and dissimilar (e.g., CMIS's conical scanner vs. AMSU's cross-track scanner) but

333   add new spectral and more detailed resolution.

334

335   *3. "Active" satellite instrumentation*

336   A relatively recent addition to temperature monitoring is the use of Global Positioning System

337   (GPS) radio signals, whose time of transmission through the atmosphere is altered by an amount

338   proportional to air density and thus temperature at levels where humidity can be ignored

339   (Kursinski et al., 1997). A key advantage of this technique for climate study is that it is self-

340   calibrating. Current systems are accurate in the upper troposphere and lower to middle

341   stratosphere where moisture is insignificant, but at lower levels, humidity becomes a

342   confounding influence on density. Future versions of this system may overcome this limitation

343   by using shorter wavelengths to measure humidity and temperature independently. Because of

344   the relatively short GPS record and limited spatial coverage to date, its value for long-term

345   climate monitoring cannot yet be definitively demonstrated.

346

347   c) Operational <u>Reanalyses</u>

348   Operational reanalyses (hereafter simply "reanalyses") will be discussed here in chapter 2, but

349   their trends presented only sparingly in the following chapters because of evidence that they are

350   not always reliable, even during the recent period. All authors expressed concern regarding

351   reanalyses trends, a concern that ranged from unanimous agreement that stratospheric trends

352   were likely spurious to mixed levels of confidence regarding tropospheric trends (see chapter 3).

353   Surface temperature trends are a separate issue as reanalyses values are indirectly *estimated*

354   rather than *observed* (see below). However, reanalyses products hold significant potential for

355   addressing many aspects of climate variability and change.

356

357   Reanalyses are not separate observing systems, but are mathematically blended products based

358   upon as many observing systems as practical. Observations are assimilated into a global weather

359   forecasting model to produce analyses that are most consistent with both the available data

360   (given their imperfections) and the assimilation model. The model, which is constrained by

361   known but parameterized atmospheric physics, generates a result that could be more accurate and

362   physically self-consistent than can be obtained from any one observing system. Some data are

363   rejected or adjusted based on detected inconsistencies. Importantly, the operational procedure

364   optimizes only the accuracy of each near-instantaneous ("synoptic") analysis. Time-varying

365   biases of a few hundredths or tenths of a degree, which contribute little to short time scale

366   weather error, present a major problem for climate trends, and these are not minimized (e.g.,

367   Sherwood, 2000). The two main reanalyses available at this time are the National Centers for

368   Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR)

369   reanalysis of data since 1948 (Kalnay et al., 1996) and the European Center for Medium-Range

370   Weather Forecasts Re-Analysis-40 (ECMWF ERA-40) beginning in 1957 (Simmons, 2004).

371

372   Because many observational systems are employed, a change in any one will affect the time

373   series of the final product. Reanalyses would be more accurate than lower-level data products for

374   climate variations only if the above shortcomings were outweighed by the benefits of using a

375   state-of-the-art model to treat unsampled variability. Factors that would make this scenario likely

376   include a relatively skillful forecast model and assimilation system, large sampling errors (which

377   are reduced by reanalysis), and small systematic discrepancies between different instruments.

378   However, current models tend to have significant intrinsic biases that can particularly affect

379   reanalyses when sampling is sparse.

380

381   Reanalysis problems that influence temperature trend calculations arise from changes over time

382   in (a) radiosonde and satellite data coverage, (b) radiosonde biases (or in the corrections applied

383   to compensate for these biases), (c) the effectiveness of the bias corrections applied to satellite

384   data and (d) the propagation of errors due to an imprecise formulation of physical processes in

385   the models. For example, since few data exist for the Southern Hemisphere before 1979,

386   temperatures were determined mainly by model forecasts; a cold model bias (in ERA-40, for

387   example) then produces a spurious warming trend when real data become available. Indirect

388   effects may also arise from changes in the biases of other fields, such as humidity and clouds,

389   which affect the model temperature (Andrae et al., 2004; Simmons et al., 2004.).

390

391   Different reanalyses do not employ the same data. NCEP/NCAR does not include surface

392   temperature observations over land but the analysis still produces estimated near-surface

393   temperatures based on the other data (Kalnay and Cai, 2003). On the other hand, ERA-40 does

394   incorporate these but only indirectly through their modeled impacts on soil temperature and

395   surface humidity (Simmons et al., 2004). Thus, the 2-meter air temperatures of both reanalyses

396   may not track closely with surface observations over time (Kalnay and Cai, 2003). SSTs in both

397   reanalyses are simply those of the climate records used as input.

398

399   For upper air reanalyses temperatures, simultaneous assimilation of radiosonde and satellite data

400   is particularly challenging because the considerably different instrument characteristics and

401   products make it difficult to achieve the consistency possible in theory. Despite data adjustments,

402   artifacts still remain in both radiosonde and satellite analyses; these produce the largest

403   differences in the lower stratosphere in current reanalysis datasets (e.g., Pawson and Fiorino,

404   1999; Santer et al., 1999; Randel, 2004). Some of these differences can now be explained, so that

405   future reanalyses will very likely improve on those currently available. However any calculation

406   of deep-layer temperatures from reanalyses which require stratospheric information are

407   considered in this report to be suspect (see Figure 2.2, T_T, T_2, T_4, and $T_{100\text{-}50}$).

408

409   d.) Simple retrieval techniques

410   A problem in interpreting MSU (i.e., broad-layer) temperature trends is that many channels

411   receive contributions from both the troposphere and stratosphere, yet temperatures tend to

412   change oppositely in these two layers with respect to both natural variability and predicted

413   climate change. In particular, MSU Channel 2 (T_2) receives 10-15% of its emissions from the

414   stratosphere (Spencer and Christy, 1992), which is a significant percentage because stratospheric

415   cooling in recent decades far exceeds tropospheric warming. It is impossible to eliminate all

416   physical stratospheric influences on MSU 2 by simply subtracting out MSU 4 (T_4) influences

417   because any linear combination of these two channels still retains stratospheric influence

418   (Spencer et al., 2005), which will lead to errors. However, it is possible to rely upon radiosonde-

419   measured correlations between tropospheric and stratospheric temperature fluctuations in order

420   to find what linear combination of these two channels leads to a near-cancellation of these errors,

421   i.e., where *y* is determined by regression:

422   *Tropospheric Retrieval = (1+y)•(T₂) - (y)•(T₄)*. The challenge here is that the resulting

423   relationship depends on the training dataset (radiosondes) being globally or tropically

424   representative (i.e., the troposphere/stratosphere boundary varies spatially and thus the

425   relationship between T_2 and T_4 does as well) and free from significant biases.

426

427   Fu et al. (2004) used a radiosonde dataset to estimate values for y (for the globe, tropical region,

428   and Northern and Southern Hemispheres) that most closely reproduced the monthly variability of

429   mean temperature from 850 to 300 hPa, spanning most of the troposphere. From physical

430   arguments, however, it is clear that the true physical contributions to the retrieval come from a

431   broader range of altitudes, which, in the tropics, approximately span the full troposphere (Fu and

432   Johanson, 2004; 2005). Although derived values of y are robust ($\pm 10\%$, Gillett et al., 2004,

433   Johanson and Fu, 2005), the veracity of the retrieval for climate change has been a subject of

434   debate (due to the accuracy and global representativeness issues mentioned above), and will be

435   further addressed in Chapter 4.

436

437   In the following chapters, two simple retrievals will be utilized in comparison studies with the

438   products of the observing systems. The tropospheric retrieval generated from global mean

439   values of T_2 and T_4, is identified as T^*_G where $y = 0.143$ (Johanson and Fu, 2005), and when

440   applied to tropical mean values is identified as T^*_T where $y = 0.100$ (Fu and Johanson, 2005).

441

442   A summary of the sources of biases and uncertainties for the datasets and other products

443   described above is given at the end of this chapter. There are several datasets yet to be generated

444   (or not yet at a stage sufficient for climate analysis) from other sources that have the potential to

445   address the issue of vertical temperature distribution. A generic listing of these datasets with a

446   characterization of their readiness is given in Table 2.1.

447
448
449
450
451   Table 2.1   Dataset types and readiness for high quality climate monitoring related to the vertical temperature
452   structure of the atmosphere. "Usage of Data" indicates the level of application of the dataset to the vertical
453   temperature issue. "Understanding" indicates the level of confidence (or readiness) in the dataset to provide accurate
454   information on this issue.
455

DATA SET SOURCE	Measured Variables	Usage of Data for Vertical Temperature	Understanding	Temporal Sampling	Geographic Completeness
Radiosondes (Balloons)	Upper Air Temperature	🟩	🟨	2x Day	🟨
	Upper Air Humidity	🟨	🟥	2x Day	🟨
	Upper Air Wind	🟥	🟥	2x Day	🟨
Microwave Radiometers Space-based	Upper air Temperature	🟩	🟨	P	🟩
	Sea Surface Temperature	🟥	🟨	P	🟩
	Total Column Vapor (ocean)	🟨	🟩	P	🟩
Surface-based sounders and profilers	Upper air Temperature	🟥	🟨	Hrly	🟥
Infrared Radiometers Space-based	Upper Air Temperature	🟨	🟨	P, G	🟩
	Land Surface Temperature	🟥	🟥	P, G	🟩
	Sea Surface Temperature	🟨	🟨	P, G	🟩
	Upper Air Humidity	🟨	🟥	P, G	🟩
Visible and	Radiative	🟨	🟨	P, G	🟩

Infrared Radiometers	Fluxes				
GPS Satellites	Temperature		🟨	🟨	quasi-P	🟩
Surface Stations Land	Land Surface Air Temperature	🟩	🟩	Hrly	🟨
	Land Surface Air Humidity	🟨	🟨	Hrly	🟨
Surface Instruments Ocean	Sea Surface Temperature	🟩	🟩	Syn	🟨
	Marine Air Temperature	🟨	🟨	Syn	🟨
Reanalyses	All	🟩	🟨	Syn	🟩

456

457   🟩 : Adequate for long-term global climate variations
458   🟨 : Improvements or continued research needed for long-term global climate variations
459   🟥 : Problems exist or a lack of analysis to date inhibit long-term global climate variation studies
460   P: Polar orbiter, twice per day per orbiter per ground location
461   G: Geostationary, many observations per day per ground location
462   2x Day: Twice daily at site
463   Hrly: Up to several times per day, many report hourly
464   Syn: Synoptic or generally up to 8 times per day. (Buoys continuous)

465

466   **2. ANALYSIS OF CLIMATE RECORDS**

467   Two factors can interfere with the accurate assessment of climate variations over multi-year

468   periods and relatively large regions. First, much larger variability (weather or "atmospheric

469   noise") on shorter time or smaller space scales can, if inadequately sampled by the observing

470   network, bias estimates of relatively small climate changes. For example, an extended heat wave

471   in an un-instrumented region accompanied by a compensating cold period in a well-instrumented

472   region may be interpreted as a "global" cold period when it was not. Such biases can result from

473   either spatial or temporal data gaps (Agudelo and Curry, 2004). Second, instrumental errors,

474   particularly biases that change over time, can create erroneous trends. The seriousness of each

475   problem depends not only on the data available but also on how they are analyzed. Finally, even

476   if global climate is known accurately at all times and places, there remains the issue of what

477   measures to use for quantifying climate change; different choices can sometimes create different

478   impressions, e.g., linear trends versus low frequency filtered analyses that retain some

479   information beyond a straight line.

480

481   Upper air layers experience relatively rapid horizontal smoothing of temperature variations, so

482   that on annual mean time scales, the atmosphere is characterized by large, coherent anomaly

483   features, especially in the east-west direction (Wallis, 1998, Thorne et al., 2005b). As a result, a

484   given precision for the global mean value over, say, a year, can be attained with fewer, if

485   properly spaced, upper air measurement locations than at the surface (Hurrell et al., 2000). Thus,

486   knowledge of global, long-term changes in upper-air temperature is limited mainly by

487   instrumental errors. However, for some regional changes (e.g., over sparsely observed ocean

488   areas) sampling problems may compete with or exceed instrumental ones.

489

490   a) Climate Records

491   Various groups have developed long time series of climate records, often referred to as Climate

492   Data Records (CDRs) (NRC, 2000b; 2000c; 2004) from the raw measurements generated by

493   each observing system. Essentially, climate records are time series that include estimates of error

494   characteristics so as to enable the study of climate variation and change on decadal and longer

495   time scales with a known precision.

496

497   Long-term temperature changes occur within the context of shorter-term variations, which are

498   listed in Table 2.2. These shorter changes include: periodic cycles such as day-night and seasonal

499   changes; fairly regular changes due to synoptic weather systems, the Quasi-Biennial Oscillation

500   (QBO), and the El Niño-Southern Oscillation (ENSO); and longer-term variations due to

501   volcanic eruptions or internal climate dynamics. These changes have different vertical

502   temperature signatures, and the magnitude of each signal may be different at the surface, in the

503   troposphere, and in the stratosphere. Details are given in Table 2.2. Some of these signals can

504   complicate the identification of temperature trends in climate records.

505   Table 2.2 Listing of atmospheric temperature variations by time scale and their properties. (Time scales and sources
506   of global temperature variations)
507

Variation	Description	Dominant Period	Approx. Magnitude	Detectability	Effect on Trend Estimates
Diurnal[1]	Warmer days than nights, due to earth's rotation on its axis affecting solar heating.	Daily (outside of polar regions)	Highly variable. Surface skin T changes up to 35K. Boundary layer changes <10K. Free tropospheric changes <1K. Stratospheric changes ~0.1-1 K.	Well detected in surface data. Poorly detected globally in the troposphere and stratosphere due to infrequent sampling (once or twice daily) and potential influence of measurement errors with their own diurnal signal. A few ground-based systems	Satellite data require adjustment of drift in the local equatorial crossing time of spacecraft orbits. Inadequate quantification of the true diurnal cycle hinders this adjustment. Different diurnal adjustments by different groups may

Variation	Description	Dominant Period	Approx. Magnitude	Detectibility	Effect on Trend Estimates
				detect signal well.	partly account for differences in trend estimates.
Synoptic[2]	Temperature changes associated with weather events, such as wave and frontal passages, due to internal atmospheric dynamics.	1-4 days	Up to ~15K or more at middle latitudes, ~3K in Tropics.	Well detected by observing systems designed to observe meteorological variability.	Not significant, but contributes to noise in climate data records.
Intraseasonal[3]	Most notably, an eastward-and vertically-propagating pattern of disturbed weather in the tropical Indo-Pacific ocean region, of unknown cause. Also, atmospheric "blocking" and wet/dry land surface can cause intra-seasonal variations at mid-latitudes.	40-60 days (Tropics), < 180 days (mid-latitudes)	1-2 K at surface, less aloft (tropics), larger in mid-latitudes.	Temperature signals moderately well detected, with tropical atmosphere limited by sparse radiosonde network and IR-based surface temperature limited by cloud. Reanalysis data are useful.	Not significant due to short duration, but may be important if character of the oscillation changes over time.
Annual[4]	Warmer summers than winters, and shift in position of major precipitation	Yearly	~2-30 K; greater over land than sea, greater at high than low latitudes,	Well observed.	Trends are often computed from "anomaly" data, after the

Variation	Description	Dominant Period	Approx. Magnitude	Detectability	Effect on Trend Estimates
	zones, due to tilt of the earth's axis of rotation affecting solar heating.		greater near the surface and tropopause than at other heights.		mean annual cycle has been subtracted. Changes in the nature of the annual cycle could affect annual-average trends.
Quasi-Biennial Oscillation (QBO)[5]	Nearly periodic wind and temperature changes in the equatorial stratosphere, due to internal atmospheric dynamics.	Every 23-28 months (average of 27 months because occasionally periods of up to 36 months occur.)	Up to 10 K locally, ~0.5 K averaged over the tropical stratosphere.	Fairly well observed by equatorial radiosonde stations and satellites.	Like ENSO, can influence trends in short data records, but it is relatively easy to remove this signal.
Interannual[6]	Multiannual variability due to interaction of the atmosphere with dynamic ocean and possibly land surfaces; most notably, ENSO. Can also be caused by volcanic eruptions.	ENSO events occur every 3-7 years and last 6-18 months; major volcanic eruptions, irregular but approximately every 5-20 years with effects lasting ~ 2 years.	Up to 3K in equatorial Pacific (ENSO), smaller elsewhere. Volcanic warming of stratosphere can exceed 5K in tropics cooling of surface <2K.	Fairly well observed, although the vertical structure of ENSO is not as well documented, due to sparseness of the tropical radiosonde network.	ENSO affects surface global mean temperatures by ±0.4K, and more in the tropical troposphere. Large ENSO events near the start or end of a data record can strongly affect computed trends, as was the case for the 1997-98 event. Changes in ENSO

Variation	Description	Dominant Period	Approx. Magnitude	Detectability	Effect on Trend Estimates
					frequency or strength affect (and may be coupled with) long-term trends.
Decadal to interdecadal oscillations and shifts[7]	Like interannual, but longer time scales. Prominent example is the PDO/ Interdecadal Pacific Oscillation. Despite long time scale, changes can occur as abrupt shifts, for example, a warming shift around 1976. Others include regional changes in the North Atlantic, Pacific-North American, Arctic, and the Antarctic oscillations. Some changes also caused by 11-year solar cycle.	Poorly known; 50-year PDO cycle suggested by 20th-century observations; others a decade or two; solar 11-year cycle detectable also.	Not well studied. The 1976-77 shift associated with a sharp warming of at least 0.2K globally, though difficult to distinguish from anthropogenic warming. 11-year cycle leads to stratospheric temperature changes of ~2K, and interacts with the Quasi-Biennial Oscillation (QBO).	Relatively large regional changes are well observed, but global expression is subject to data consistency issues over time and possible real changes.	Can account for a significant fraction of linear trends calculated over periods of a few decades or less regionally. Such trends may differ significantly from one such period to the next.
Sub-centennial 60-80 year fluctuation or "Atlantic	Fluctuates in instrumental and paleo data at least back to c.1600. Seems to	60-80 years	~ ±0.5C in parts of the Atlantic. Apparently detectable	Detectable globally above the noise, clear in North Atlantic SST.	Effects small globally, but probably detectable in last few

Variation	Description	Dominant Period	Approx. Magnitude	Detectibility	Effect on Trend Estimates
Multidecadal Oscillation"[8]	particularly affect Atlantic sector. Possible interhemispheric component.		global mean ~ ±0.1C		decades. Readily detectable over this period in North Atlantic Ocean where it clearly affects surface temperature trends and probably climate generally.
Centennial and longer variations[9]	Warming during 20th Century due to human influences, solar, and internal variability. Earlier changes included the "little ice age" and "medieval warm period."	None confirmed, though 1500 year Bond cycle possible.	20th century warming of ~0.6K globally appears to be as large or larger than other changes during the late Holocene.	Surface warming during 20th century fairly well observed; proxies covering earlier times indicated 20th century warmer than the past 5 centuries	Natural temperature variations occur on the longest time scales accessible in any instrumental record.

508

509

[1] Christy et al., 2003; Mears et al., 2003; Vinnikov and Grody. 2003; Dai and Trenberth, 2004; Jin, 2004; Seidel et al., 2005.

[2] Palmen and Newton, 1969

[3] Duvel et al.,2004.

[4] Wallace and Hobbs, 1977

[5] Christy and Drouilhet, 1994; Randel et al., 1999; Baldwin et al., 2001

[6] Parker and Brownscombe, 1983; Pan and Oort, 1983; Christy and McNider, 1994; Parker et al., 1996; Angell,

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521   2000; Robock, 2000; Michaels and Knappenberger, 2000; Santer et al., 2001; Free and Angell, 2002a; Trenberth et
522   al., 2002; Seidel et al., 2004; Seidel and Lanzante, 2004
523
524   [7] Labitzke, K.,1987; Trenberth and Hurrell, 1994; Lean et al., 1995; Zhang et al., 1997; Thompson et al., 2000;
525   Douglass and Clader, 2002; Seidel and Lanzante, 2004; Hurrell et al., 2003; Folland et al., 1999; Power et al., 1999;
526   Folland et al. 2002.
527
528   [8] Schlesinger and Ramankutty, 1994; Mann et al., 1998; Folland et al., 1999; Andronova and Schlesinger, 2000;
529   Goldenberg et al., 2001; Enfield et al., 2001
530
531   [9] Folland et al., 2001a.

532   Our survey of known atmospheric temperature variations, how well they are measured, and their

533   impact on trend estimates suggests that most observing systems are generally able to quantify

534   well the magnitudes of change associated with shorter time scales. For longer time scale changes,

535   where the magnitudes of change are smaller and the stability requirements more rigorous, the

536   observing systems face significant challenges (Seidel et al., 2004).

537

538   b) Measuring Temperature Change

539

540   Over the last three to five decades, global surface temperature records show increases of almost

541   two tenths of a °C per decade. Explaining atmospheric and surface trends therefore demands

542   relative accuracies of a few hundredths of a degree per decade in global time series of both

543   surface and upper-air observations. As this and subsequent chapters will show, the effects of

544   instrumental biases on the global time series are significantly larger than a few hundredths of a

545   degree for the upper-air data, though the global surface temperature compilations do appear to

546   reach this level of precision in recent decades (Folland et al., 2001b). These biases, especially

547   those of the upper air, must therefore be understood and quantified rather precisely (see section 3

548   below). For this fundamental reason, reliable assessment of lapse rate changes remains a

549   considerable challenge.

550

551   Natural modes of climate variability on regional scales are manifested in decadal fluctuations in

552   (a) the tropical Pacific, e.g., ENSO, and (b) the northern latitudes, e.g., the North Atlantic,

553   Pacific-North American and the Arctic atmospheric oscillations (Table 2.2). Even fluctuations on

554   longer time scales have been proposed, e.g., the Atlantic Multidecadal Oscillation/60-80 year

555   variation (Schlesinger and Ramankutty, 1994; Enfield et al., 2001). Each of these phenomena is

556   associated with regions of both warming and cooling. Distinguishing slow, human-induced

557   changes from such phenomena requires identifying the patterns and separating the influences of

558   such modes from the warming signal (e.g., as attempted by SST by Folland et al., 1999.) In

559   addition, these oscillations could themselves be influenced by human-induced atmospheric

560   changes (Hasselmann, 1999).

561

562   **3. LIMITATIONS**

563

564   A key question addressed in this report is whether climate records built by investigators using

565   various components of the observing system can meet the needs for assessing climate variations

566   and trends with the accuracy and representativeness which allows any human attribution to be

567   reliably identified. Climate record builders have usually underestimated the overall uncertainty in

568   their products by relying on traditional sources of uncertainty that can be quantified using

569   standard statistical methods. For example, published linear trend values exist of the same

570   temperature product from the same observing system whose error estimates do not overlap,

571   indicating serious issues with error determination. Thus, in 2003, three realizations of T_2 (or

572   MSU channel 2) 1979-2002 global trends were published as +0.03 ±0.05 +0.12 ±0.02, and +0.24

573   ±0.02 °C per decade (Christy et al., 2003; Mears et al., 2003; and Vinnikov and Grody, 2003,

574   respectively.) Over 40% of the difference between the first two trends is due to the treatment of a

575   single satellite in the 1984-1986 period, with a combination of lesser differences during later

576   satellite periods. The third dataset has more complex differences, though it is being superseded

577   by a version whose trend is now lower (Grody et al., 2004, Vinnikov et al. 2005).

578

579   This situation illustrates that it is very challenging to determine the true error characteristics of

580   datasets (see Chapter 4), although considerably less attention has been paid to this than to the

581   construction of the datasets themselves. In this report, we refer to systematic errors in the climate

582   data records as "construction errors." Such errors can be thought of as having two fundamentally

583   different sources, *structural* and *parametric* (see Box 2.1). The human decisions that underlie the

584   production of climate records may be thought of as forming a *structure* for separating real and

585   artificial behavior in the raw data. Assumptions made by the experts may not be correct, or

586   important factors may have been ignored; these possibilities lead to *structural uncertainty*

587   (Thorne et al., 2005a) in any trend or other metric obtained from a given the climate record.

588   Experts generally tend to underestimate structural uncertainty (Morgan, 1990). The T_2 example

589   above shows that this type of error can considerably exceed those recognized by the climate

590   record builders. Sorting out which decisions are better than others, given the fact many

591   individual decisions are interdependent and often untestable, is challenging.

592

593   Structural uncertainty is difficult to quantify because this requires considering alternatives to the

594   fundamental assumptions, rather than just to the specific sampling or bias pattern in the available

595   data (the main source of parametric uncertainty). For example, is an apparent diurnal variation

596   due to (a) real atmospheric temperature change, (b) diurnal solar heating of an instrument

597   component, (c) a combination of both, or (d) something else entirely? If the answer is not known

598   *a priori*, different working assumptions may lead to a different result when corrections are

599   determined and applied.

600

601   There may be several ways to identify structural errors. First, it is well known in statistics that

602   one should examine the variability that is left over when known effects are removed in a data

603   analysis, to see whether the residuals appear as small and "random" as implied by the

604   assumptions. Even when the residuals are examined, it is often difficult to identify the cause of

605   any non-randomness. Second, one can compare the results with external or independent data

606   (such as comparing SST and NMAT observations). However, one then encounters the problem

607   of assessing the accuracy of the independent data; because, in the case of global atmospheric

608   temperature data there are no absolute standards for any needed adjustment. Christy et al. (2000)

609   demonstrate the use of internal and external methods for evaluating the error of their upper air

610   time series. They assumed that where agreement of independent measurements exists, there is

611   likely to be increased confidence in the trends. Third, one can try to assess the construction

612   uncertainty by examining the spread of results obtained by multiple experts working

613   independently (e.g., the T_2 example, Thorne et al., 2005a). Unfortunately, though valuable, this

614   does not establish the uncertainties of individual efforts, nor is it necessarily an accurate measure

615   of overall uncertainty. If all investigators make common mistakes, the estimate of construction

616   uncertainty may be too optimistic; but if some investigators are unaware of scientifically sound

617   progress made by others, the estimate can be too pessimistic.

618

619   A general concern regarding all of the datasets used in this analysis - land air temperature, sea

620   surface temperature, radiosonde temperature, and satellite-derived temperature – is the level of

621   information describing the operational characteristics and evolution of the associated observing

622   system. As indicated above, the common factor that creates the biggest differences between

623   analyses of the same source data is the homogeneity adjustments made to account for biases in

624   the raw data. All homogeneity adjustments would improve with better metadata - that is,

625   information (data) about the data (see chapter 6). For satellite-derived temperature, additional

626   metadata such as more data points used in the pre-launch calibration would have been helpful to

627   know, especially if done for differing solar angles to represent the changes experienced on orbit.

628   For the in situ data sets, additional metadata of various sorts likely exist in one form or another

629   somewhere in the world and could be acquired or created. These include the type of instrument,

630   the observing environment, the observing practices and the exact dates for changes in any of the

631   above.

632

633   Below we identify various known issues that led to errors in the datasets examined in this report,

634   and which have generally been addressed by the various dataset builders. Note that reanalyses

635   inherit the errors of their constituent observing systems, though they have the advantage of

636   seeking a degree of consensus among the various observing systems through the constraint of

637   model physics. The complex reanalysis procedure transforms these errors of output data into

638   errors of construction methodology that are hard to quantify.

639

640   **Errors primarily affecting *in situ* observing systems.**

641

642   **Spatial and temporal sampling**: The main source of this error is the poor sampling of oceanic

643   regions, particularly in the Southern Hemisphere, and some tropical and Southern

644   Hemisphere continental regions (see Text Box 2.1). Temporal variations in radiosonde

645   sampling can lead to biases, (e.g., switching from 00 to 12 UTC) but these are generally

646   documented and thus potentially treatable.

647   **Local environmental changes**: Land-use changes, new instrument exposures, etc., create new

648   localized meteorological conditions to which the sensor responds. These issues are most

649   important for land near-surface air temperatures but can also affect the lower elevation

650   radiosonde data. Some changes, e.g., irrigation, can act to increase nighttime minima

651   while decreasing daytime maxima, leaving an ambiguous signal for the daily mean

652   temperature. Such changes are sources of error only if the change in the immediate

653   surroundings of the station is unrepresentative of changes over a larger region.

654   **Changes in methods of observation**: A change in the way in which an instrument is used, as in

655   calibrating a radiosonde before launch, i.e., whether it is compared against a typical

656   outdoor sensor or against a traceable standard.

657   **Changes in data processing algorithms**: A change in the way raw data are converted to

658   atmospheric information can introduce similar problems. For radiosonde data, the raw

659   observations are often not archived and so the effects of these changes are not easily

660   removed.

661

662   **Errors primarily affecting satellite systems**

663

664   **Diurnal sampling**: It is common for polar orbiters to drift slowly away from their "sun-

665   synchronous" initial equatorial crossing times (e.g., 1:30 p.m. to 5 p.m.), introducing

666   spurious trends related to the natural diurnal cycle of daily temperature. The later polar

667   orbiters (since 1998) have more stable station keeping. Diurnal drift adjustments for T_{2LT}

668   and T_2 impact the trend by a few hundredths °C/decade. Changes in local observation

669   time also significantly afflict *in situ* temperature observations, with a lesser impact on the

670   global scale.

671   **Orbit decay**: Variations in solar activity cause expansion and contraction of the thin atmosphere

672   at the altitudes where satellites orbit, which create variable frictional drag on spacecraft.

673   This causes periods of altitude decay, changing the instrument's viewing geometry

674   relative to the earth and therefore the radiation emissions observed. This issue relates

675   most strongly to T_{2LT}, which uses data from multiple view angles, and is of order 0.1

676   °C/decade.

677   **Calibration shifts/changes:**   For satellite instruments, the effects of launch conditions or

678   changes in the within-orbit environment (e.g., varying solar shadowing effects on the

679   spacecraft components as it drifts through the diurnal cycle) may require adjustments to

680   the calibration equations. Adjustment magnitudes vary among the products analyzed in

681   this report but are on the order of 0.1 °C/decade for T_{2LT} and T_2.

682   **Surface emissivity effects**: The intensity of surface emissions in observed satellite radiances

683   can vary over time due to changes in surface properties, e.g. wet vs. dry ground, rough vs.

684   calm seas, etc., and longer-term land cover changes, e.g., deforestation leading to higher

685   daytime skin temperatures and larger diurnal temperature cycles.

686   **Atmospheric effects:** Atmospheric composition can vary over time (e.g., aerosols), affecting

687   satellite radiances, especially the infrared.

688

689   **Errors affecting all observing systems**

690

691   **Instrument Changes**: Systematic variations of calibration between instruments will lead to

692   time-varying biases in absolute temperature. These involve (a) changes in instruments

693   and their related components (e.g., changes in housing can be a problem for *in situ*

694   surface temperatures), (b) changes in instrument design or data processing (e.g.,

695   radiosondes) and (c) copies of the same instrument that are intended to be identical but

696   are not (e.g., satellites).

697

698   **Errors or differences related to analysis or interpretation**

699

700   **Construction Methodology**: As indicated, this is often the source of the largest differences

701   among trends from datasets and is the least quantifiable. When constructing a

702   homogeneous, global climate record from an observing system, different investigators

703   often make a considerable range of assumptions as to how to treat unsampled or

704   undersampled variability and both random and systematic instrument errors. The trends

705   and their uncertainties that are subsequently estimated are sensitive to treatment

706   assumptions (Free et al., 2002b). For example, the trends of the latest versions of T_2 from

707   the three satellite analyses vary from +0.044 to +0.199 °C/decade (chapter 3), reflecting

708   the differences in the combination of individual adjustments determined and applied by

709   each team (structural uncertainty.) Similarly, the T_2 global trends of the radiosonde-

710   based and reanalyses datasets range from –0.036 to +0.067 °C/decade indicating

711   noticeable differences in decisions and methodologies by which each was constructed.

712   Thus the goal of achieving a consensus with an error range of a few hundredths

713   °C/decade is not evidenced in these results.

714   **Trend Methodology**: Differences between analyses can arise from the methods used to

715   determine trends. Trends shown in this report are calculated by least squares linear

716   regression.

717   **Representativeness**: Any given measure reported by climate analysts could under- or overstate

718   underlying climatic behavior. This is not so much a source of error as a problem of

719   interpretation. This is often called statistical error. For example, a trend computed for

720   one time period (say, 1979-2004) is not necessarily representative of either longer or

721   earlier periods (e.g., 1958-1979), so caution is necessary in generalizing such a result. By

722   the same token, large variations during portions of the record might obscure a small but

723   important underlying trend. (See Appendix for Statistical Uncertainties.)

724

725

726   **<u>4. IMPLICATIONS</u>**

727   The observing systems deployed since the late 1950s, and the subsequent climate records derived

728   from their data, have the capability to provide information suitable for the detection of many

729   temperature variations in the climate system. These include temperature changes that occur with

730   regular frequency, e.g., daily and annual cycles of temperature, as well as non-periodic events

731   such as volcanic eruptions or serious heat and cold waves. The data from these systems also have

732   the potential to provide accurate trends in climate over the last few decades (and over the last

733   century for surface observations), once the raw data are successfully adjusted for changes over

734   time in observing systems, practices, and micro-climate exposure to produce usable climate

735   records. Measurements from all systems require such adjustments and this report relies on

736   adjusted datasets. The details of making such adjustments when building climate records from

737   the uncorrected observations are examined in the following chapters.

738

739   **Text Box 2.1: Comparing Radiosonde and Satellite Temperatures**

740   Attempts to compare temperatures from satellite and radiosonde measurements are hindered by a

741   mismatch between the respective raw observations. While radiosondes measure temperatures at

742   specific vertical levels, satellites measure radiances which can be interpreted as the temperature

743   averaged over a deep layer.   To simulate a satellite observation, the different levels of

744   temperature in the radiosonde sounding are proportionally weighted to match the profiles shown

745   in Figure 2.2. This can be done in one of two ways.

746

747   1. Employ a simple set of geographically and seasonally invariant coefficients or weights, called

748   a static weighting function. These coefficients are multiplied by the corresponding set of

749   temperatures at the radiosonde levels and the sum is the simulated satellite temperature. Over

750   land, the surface contributes more to the layer-average than it does over the ocean, and this

751   difference is taken into account by slightly different sets of coefficients applied to land vs. ocean

752   calculations.   This same method may be applied to the temperature level data of global

753   reanalyses. We have applied the "static weighting function" approach in this report.

754

755   2. Take into account the variations in the air mass temperature, surface temperature and pressure,

756   and atmospheric moisture (Spencer et al., 1990).   Here, the complete radiosonde temperature

757   and humidity profiles are ingested into a radiation model to generate the simulated satellite

758   temperature (e.g., Christy and Norris, 2004).   This takes much more computing power to

759   calculate and requires humidity information, which for radiosondes is generally of poorer quality

760   than temperature information or is missing entirely. For climate applications, in which the time

761   series of large-scale anomalies is the essential information, the output from the two methods

762   differs only slightly.

763

764   There are practical difficulties in generating long time series of simulated satellite temperatures

765   under either approach. To produce a completely homogeneous data record, the pressure levels

766   used in the calculation must be consistent throughout time, i.e., always starting at the surface and

767   reaching the same designated altitude. If, for example, soundings achieved higher elevations as

768   time went on, there would likely be a spurious trend due to the effects of having measured

769   observations during the latter period of record, while by necessity, relying on estimates for the

770   missing values in the earlier period. We also note that HadAT utilizes 9 pressure levels for

771   simulating satellite profiles while RATPAC use 15, so differences can arise from these differing

772   inputs.

773

774   An additional complication is that many radiosonde datasets and reanalyses may provide data at

775   mandatory levels beginning with 1000 and/or 850 hPa, i.e., with no identifiable surface. Thus,

776   the location of the material surface, and its temperature, can only be estimated so that an

777   additional source of error to the anomaly time series may occur. There are a number of other

778   processing choices available when producing a time series of simulated satellite data for site-by-

779   site comparisons between actual satellite data and radiosondes (or reanalyses) and these also

780   have the potential to introduce non-negligible biases.

781

782   Averaging of spatially incomplete radiosonde observations for comparison of global and tropical

783   anomalies also introduces some error (Agudelo and Curry, 2004). In this report we have first

784   zonally averaged the data, then generated satellite-equivalent measures from these data and

785   finally calculated global and tropical averages. The spatial coverage differs markedly between

786   the two radiosonde datasets. However, as anomalies are highly correlated in longitude the

787   relative poor longitudinal sampling density of RATPAC (and HadAT outside of the NH mid-

788   latitudes) is not necessarily an impediment (Hurrell et al., 2000). Comparing global averages

789   estimated using only those zonally-averaged grids observed at RATPAC station sites by MSU

790   versus the globally complete fields from MSU, a sampling error of less than ±0.05 °C/decade

791   was inferred for T_{2LT}. Satellite and reanalyses are essentially globally complete and thus do not

792   suffer from spatial subsampling.

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6   **Chapter 3**

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8

9   What do observations indicate about the changes of temperature in the atmosphere and at
10   the surface since the advent of measuring temperatures vertically?

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19

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40   **Key Findings**
41
42 **Observed Changes - Surface**

43   **Globally, as well as in the tropics, the temperature of the air near the Earth's**
44   **surface has increased since 1958, with a greater rate of increase since 1979. All**
45   **three surface temperature datasets are consistent in these conclusions.**

46   • Globally, temperature increased at a rate of about 0.12ºC per decade since
47   1958, and about 0.16ºC per decade since 1979.

48   • In the tropics, temperature increased at a rate of about 0.11ºC per decade
49   since 1958, and about 0.13ºC per decade since 1979.

50   • Most, if not all of the surface temperature increase since 1958 occurs
51   starting around the mid-1970s, a time coincident with a previously identified
52   abrupt climate regime shift. However, there does not appear to be a strong
53   jump up in temperature at this time, rather the major part of the rise seems to
54   occur in a more gradual fashion.

55 **Observed Changes - Troposphere**

56   **Globally, as well as in the tropics, both balloon-based datasets dating back to**
57   **1958 agree that the tropospheric temperature has increased slightly more than**
58   **that of the surface. Since 1979, due to the considerable disagreement among**
59   **tropospheric datasets, it is not clear whether the temperature of the**
60   **troposphere has increased more or less than that of the surface, both globally**
61   **and in the tropics.**

62   • Globally, temperature increased at a rate of about 0.14ºC per decade since
63   1958 according to the two balloon-based datasets. Since 1979, estimates of
64   the increase from the two balloon and three satellite datasets range from
65   about 0.10 to 0.20ºC per decade.

66   • In the tropics, temperature increased at a rate of about 0.13ºC per decade
67   since 1958 according to the two balloon-based datasets. However, since
68   1979, estimates of the increase from the two balloon and three satellite
69   datasets range from about 0.02 to 0.19ºC per decade.

70   • For the balloon-based estimates since 1958, the major part of the
71   temperature increase appears in the form of an abrupt rise in the mid-1970s,
72   apparently in association with a climate regime shift that occurred at this
73   time.

74 **Observed Changes - Lower Stratosphere**

75   **Globally, the temperature of the lower stratosphere has decreased both since**
76   **1958 and since 1979. The two balloon-based datasets yield reasonably**
77   **consistent estimates of the rates of cooling for both time periods. However,**
78   **since 1979 the two balloon datasets estimate a considerably greater rate of**
79   **cooling than the two satellite datasets, which themselves disagree.**

80   • Globally, the rate of cooling since 1958 is about 0.37ºC per decade based on
81   the two balloon datasets. Since 1979, estimates of this decrease are about
82   0.65ºC per decade for the two balloon datasets, and from about 0.33 to 0.45º
83   C per decade for the two satellite datasets.

84   • The bulk of the stratospheric temperature decrease occurred from about the
85   late 1970s to the middle 1990s. It is unclear whether the decrease was
86   gradual or occurred in abrupt steps in the first few years after each major
87   volcanic eruption.

88   **Chapter 3: Recommendations**

89

90 • *Although considerable progress has been made in explaining the causes of*
91   *discrepancies between upper-air datasets, both satellite and balloon-based,*
92   *continuing steps should be taken to thoroughly assess and improve methods used*
93   *to remove time-varying biases that are responsible for these discrepancies.*

94 • *New observations should be made available in order to provide more redundancy*
95   *in climate monitoring. Activities should include both the introduction of new*
96   *observational platforms as well as the necessary processing of data from*
97   *currently under-utilized platforms. For example IR and GPS satellite observations*
98   *have not been used to any great extent, the former owing to complications when*
99   *clouds are present and the latter owing to a short period of record. Additionally,*
100   *the introduction of a network of climate quality, reference stations, that include*
101   *reference radiosondes, would place future climate monitoring on a firmer basis.*

102   **1. Background**

103

104   In this chapter we describe changes in temperature at the surface and in the atmosphere

105   based on four basic types of products derived from observations: surface, radiosonde,

106   satellite and reanalysis. However, we limit our discussion of reanalysis products given

107   their more problematic nature for use in trend analysis (see Chapter 2); only a few trend

108   values are presented for illustrative purposes.

109

110   Each of these four generic types of measurements consists of multiple datasets prepared

111   by different teams of data specialists. The datasets are distinguished from one another by

112   differences in the details of their construction. Each type of measurement system as well

113   as each particular dataset has its own unique strengths and weaknesses. Because it is

114   difficult to declare a particular dataset as being "the best," it is prudent to examine results

115   derived from more than one "credible" dataset of each type. Also, comparing results from

116   more than one dataset provides a better idea of the uncertainties or at least the range of

117   results. In the interest of clarity and conciseness, we have chosen to display and perform

118   calculations for a representative subset of all available datasets. We consider these to be

119   the "state of the art" datasets of their type, based on our collective expert judgment.

120

121   In selecting datasets for use in this report, we limit ourselves to those products that are

122   being actively updated and for which temporal homogeneity is an explicit goal in the

123   construction, as these are important considerations for their use in climate change

124   assessment. By way of a literature review, we discuss additional datasets not used in this

125   report. Since some datasets are derivatives of earlier ones, we mention this where

126   appropriate. One should not misconstrue the exclusion of a dataset from this report as an

127   invalidation of that product. Indeed, some of the excluded datasets have proved to be

128   quite valuable in the past and will continue to be so into the future.

129

130   Most of the analyses that we have performed involve data that were averaged over a large

131   region, such as the entire globe or the tropics. The spatial averaging process is

132   complicated by the fact that the locations (gridpoints or stations) at which data values are

133   available can vary fundamentally by data type (see Chapter 2 for details) and, even for a

134   given type, between data production teams. In an effort towards more consistency, the

135   spatial averages we use represent the weighted average of zonal averages[1] (i.e., averages

136   around an entire latitude line or zone), where the weights are the cosine of latitude[2]. This

137   insures that the different latitude zones are given equal treatment across all datasets.

138

139   This chapter begins with a discussion of the four different data types, introducing some

140   temperature datasets for each type, and then discussing their time histories averaged over

141   the globe. Later we present more detail, concentrating on the analysis of temperature

142   trends for two eras: (1) the period since the widespread availability of radiosonde

143   observations in 1958, and (2) since the introduction of satellite data in 1979. We compare

---

[1] The zonal averages, which were supplied to us by each dataset production team, differ among datasets. We allowed each team to use their judgment as how to best produce these from the available gridpoint or station values in each latitude zone

[2] The cosine factor weights lower latitudes more than higher ones, to account for the fact that lines of longitude converge towards the poles. As a result, a zonal band in lower latitudes encompasses more area than a comparably sized band (in terms of latitude/longitude dimensions) in higher latitudes.

144   overall temperature trends from different measurement systems and then go into more

145   detail on trend variations in the horizontal and vertical.

146

147   **2. Surface Temperatures**

148

149   **2.1   Land-based temperature data**

150   Over land, temperature data come from fixed weather observing stations with

151   thermometers housed in special instrument shelters. Records of temperature from many

152   thousands of such stations exist. Chapter 2 outlines the difficulties in developing reliable

153   surface temperature datasets. One concern is the variety of changes that may affect

154   temperature measurements at an individual station. For example, the thermometer or

155   instrument shelter might change, the time of day when the thermometers are read might

156   change, or the station might move. These problems are addressed through a variety of

157   procedures (see Peterson et al., 1998 for a review) that are generally quite successful at

158   removing the effects of such changes at individual stations (e.g., Vose et al., 2003 and

159   Peterson, 2005) whether the changes are documented in the metadata or detected via

160   statistical analysis using data from neighboring stations as well (Aguilar et al., 2003).

161   Subtle or widespread impacts that might be expected from urbanization or the growth of

162   trees around observing sites might still contaminate a dataset. These problems are

163   addressed either actively in the data processing stage (e.g., Hansen et al., 2001) or

164   through dataset evaluation to ensure as much as possible that the data are not biased[3]

---

[3] Changes in regional land use such as deforestation, aforestation, agricultural practices, and other regional changes in land use are not addressed in the development of these datasets. While modeling studies have

165   (e.g., Jones et al., 1990; Peterson, 2003; Parker, 2004; Peterson and Owen, 2005).

166

167   **2.2   Marine temperature data**

168   Data over the ocean come from moored buoys, drifting buoys, and volunteer observing

169   ships. Historically, ships have provided most of the data, but in recent years an increasing

170   number of buoys have been used, placed primarily in data-sparse areas away from

171   shipping lanes. In addition, satellite data are often used after 1981. Many of the ships and

172   buoys take both air temperature observations and sea surface temperature (SST)

173   observations. Night marine air temperature (NMAT) observations have been used to

174   avoid the problem that the Sun's heating of the ship's deck can make the thermometer

175   reading greater than the actual air temperature. Where there are dense observations of

176   NMAT and SST, over the long term they track each other very well. However, since

177   marine observations in an area may only be taken a few times per month, SST has the

178   advantage over air temperature in that water temperature changes much more slowly than

179   that of air. Also, there are twice as many SST observations as NMAT from the same

180   platforms as SSTs are taken during both the day and night and SST data are

181   supplemented in data sparse areas by drifting buoys which do not take air temperature

182   measurements. Accordingly, only having a few SST observations in a grid box for a

183   month can still provide an accurate measure of the average temperature of the month.

184

suggested over decades to centuries these affects can be important on regional space scales (Oleson et al., 2004), we consider these effects to be those of an external forcing to the climate system and are treated as such by many groups in the simulation of climate using the models described in Chapter 5. To the extent that these effects could be large enough to have a measurable influence on global temperature, these changes will be detected by the land-based surface network.

185   **2.3   Global surface temperature data**

186   Currently, there are three main groups creating global analyses of surface temperature

187   (see Table 3.1), differing in the choice of available data that are utilized as well as the

188   manner in which these data are synthesized. Since the network of surface stations

189   changes over time, it is necessary to assess how well the available observations monitor

190   global or regional temperature. There are three ways in which to make such assessments

191   (Jones, 1995). The first is using "frozen grids" where analysis using only those grid

192   boxes with data present in the sparsest years is used to compare to the full dataset results

193   from other years (e.g., Parker et al., 1994). The results generally indicate very small

194   errors on multi-annual timescales (Jones, 1995). The second technique is subsampling a

195   spatially complete field, such as model output, only where in situ observations are

196   available. Again the errors are small (e.g., the standard errors are less than 0.06°C for the

197   observing period 1880 to 1990; Peterson et al., 1998b). The third technique is comparing

198   optimum averaging, which fills in the spatial field using covariance matrices,

199   eigenfunctions or structure functions, with other analyses. Again, very small differences

200   are found (Smith et al., 2005). The fidelity of the surface temperature record is further

201   supported by work such as Peterson et al. (1999) which found that a rural subset of global

202   land stations had almost the same global trend as the full network and Parker (2004) that

203   found no signs of urban warming over the period covered by this report.

204

205

206

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208

209   **Table 3.1: Temperature datasets utilized in this report.**
210

211   *Our Name   Name given by Producers   Producers*
212   *Web Page*

-- Surface –

215   NOAA   ER-GHCN-ICOADS   NOAA's National Climatic Data Center
216   (NCDC)
217   http://www.ncdc.noaa.gov/oa/climate/monitoring/gcag/gcag.html
218

219   GISS   Land+Ocean Temperature   NASA's Goddard Institute for Space Studies
220   (GISS)
221   http://www.giss.nasa.gov/data/update/gistemp/graphs/
222

223   HadCRUT2v   HadCRUT2v   Climatic Research Unit of the University of East
224   Anglia and the Hadley Centre of the UK Met
225   Office.
226   http://www.cru.uea.ac.uk/cru/data/temperature
227

228   **-- Radiosonde –**
229

230   RATPAC   RATPAC   NOAA's: Air Resources Laboratory (ARL),
231   Geophysical Fluid Dynamics Laboratory
232   (GFDL), and National Climatic Data Center
233   (NCDC)
234   http://www.ncdc.noaa.gov/
235

236

237   HadAT2   HadAT2   Hadley Centre, UK
238   http://www.hadobs.org/
239

240

241   **-- Satellite –**
242

243   ***Temperature of the Lower Troposphere***
244   T_{2LT-A} TLT   University of Alabama in Huntsville (UAH)
245   http://vortex.nsstc.uah.edu/data/msu/t2lt
246

247   T_{2LT-R} TLT   Remote Sensing System, Inc. (RSS)
248   http://www.remss.com/msu/msu_data_description.html
249

250   *Temperature of the Middle Troposphere*
251   T_{2-A} TMT   University of Alabama in Huntsville (UAH)
252   http://vortex.nsstc.uah.edu/data/msu/t2
253
254   T_{2-R} TMT   Remote Sensing System, Inc. (RSS)
255   http://www.remss.com/msu/msu_data_description.html
256
257   T_{2-M} Channel 2   University of Maryland and NOAA/NESDIS
258   (U.Md.)
259
260   *Temperature of the Middle Troposphere minus Stratospheric Influences*
261    $T*_G$ (global) $T_{(850-300)}$ University of Washington, Seattle (UW) and
262    $T*_T$ (tropics)   NOAA's Air Resources Laboratory (ARL)
263
264   *Temperature of the Lower Stratosphere*
265   T_{4-A} TLS   University of Alabama in Huntsville (UAH)
266   http://vortex.nsstc.uah.edu/data/msu/t4
267
268   T_{4-R} TLS   Remote Sensing System, Inc. (RSS)
269   http://www.remss.com/msu/msu_data_description.html
270
271   **-- Reanalysis –**
272
273   US   NCEP50   National Center for Environmental Prediction,
274   NOAA and the National Center for Atmospheric
275   Research
276   http://wesley.ncep.noaa.gov/reanalysis.html
277
278   European   ERA40   European Center for Medium Range
279   Forecasting
280   http://www.ecmwf.int/research/era
281
282

283   **2.3.1 NOAA**

284   The National Oceanic and Atmospheric Administration (NOAA) National Climatic Data

285   Center (NCDC) integrated land and ocean dataset (see Table 3.1) is derived from in situ

286   data. The SSTs come from the International Comprehensive Ocean-Atmosphere Data Set

287   (ICOADS) SST observations release 2 (Slutz et al., 1985; Woodruff et al., 1998; Diaz et

288   al., 2002). Those that pass quality control tests are averaged into monthly 2° grid boxes

289   (Smith and Reynolds, 2003). The land surface air temperature data come from the Global

290   Historical Climatology Network (GHCN) (Peterson and Vose, 1997) and are averaged

291   into 5° grid boxes. A reconstruction approach is used to create complete global coverage

292   by combining together the faster and slower time-varying components of temperature

293   (van den Dool et al., 2000; Smith and Reynolds, 2005).

294

295   **2.3.2 NASA (GISS)**

296   The NASA Goddard Institute for Space Studies (GISS) produces a global air temperature

297   analysis (see Table 3.1) known as GISTEMP using land surface temperature data

298   primarily from GHCN and the U.S. Historical Climatology Network (USHCN;

299   Easterling, et al., 1996). The NASA team modifies the GHCN/USHCN data by

300   combining at each location the time records of the various sources and adjusting the non-

301   rural stations in such a way that their long-term trends are consistent with those from

302   neighboring rural stations (Hansen et al., 2001). These meteorological station

303   measurements over land are combined with in situ sea surface temperatures and Infrared

304   Radiation (IR) satellite measurements for 1982 to the present (Reynolds and Smith, 1994;

305   Smith et al., 1996) to produce a global temperature index (Hansen et al., 1996).

306

307   **2.3.3 UK (HadCRUT2v)**

308   The UK global land and ocean dataset (HadCRUT2v, see Table 3.1) is produced as a

309   joint effort by the Climatic Research Unit of the University of East Anglia and the

310   Hadley Centre of the UK Meteorological (Met) Office. The land surface air temperature

311   data are from Jones and Moberg (2003) of the Climatic Research Unit. The global SST

312   fields are produced by the Hadley Centre using a blend of COADS and Met Office data

313   bank in situ observations (Rayner, et al., 2003). The integrated dataset is known as

314   HadCRUT2v (Jones and Moberg, 2003)[4]. The temperature anomalies were calculated on

315   a 5^ox5^o grid box basis. Within each grid box, the temporal variability of the observations

316   has been adjusted to account for the effect of changing the number of stations or SST

317   observations in individual grid-box temperature time series (Jones et al., 1997, 2001).

318   There is no reconstruction of data gaps because of the problems of introducing biased

319   interpolated values.

320

321   **2.3.4 Synopsis of surface datasets**

322   Since the three chosen datasets utilize many of the same raw observations, there is a

323   degree of interdependence. Nevertheless, there are some differences among them as to

324   which observing sites are utilized. An important advantage of surface data is the fact that

325   at any given time there are thousands of thermometers in use that contribute to a global or

326   other large-scale average. Besides the tendency to cancel random errors, the large number

327   of stations also greatly facilitates temporal homogenization since a given station may

328   have several "near-neighbors" for "buddy-checks." While there are fundamental

329   differences in the methodology used to create the surface datasets, the differing

330   techniques with the same data produce almost the same results (Wuertz et al., 2005). The

---

[4] Although global and hemispheric temperature time series created using a technique known as optimal averaging (Folland et al., 2001a; Parker et al., 2004), which provides estimates of uncertainty in the time series, including the effects of data gaps and uncertainties related to bias corrections or uncorrected biases, are available, we have used the data in their more basic form, for consistency with the other datasets.

331   small differences in deductions about climate change derived from the surface datasets

332   are likely to be due mostly to differences in construction methodology and global

333   averaging procedures.

334

335   **2.4   Global surface temperature variations and differences between the datasets**

336   Examination of the three global temperature anomaly time series (T_{Sfc}) from 1958 to the

337   present shown in Figure 3.1 reveals that the three time series have a very high level of

338   agreement. They all show some temperature decrease from 1958 to around 1976,

339   followed by a strong increase. That most of the temperature change occurs after the mid

340   1970s has been previously documented (Karl et al., 2000; Folland et al., 2001b; Seidel

341   and Lanzante, 2004). The variability of the time series is quite similar as are their trends.

342   The signature of the El Niño-Southern Oscillation (ENSO), whose origin is in the tropics,

343   is responsible for many of the prominent short-term (several year) up and down swings of

344   temperature (Trenberth et al., 2002). The strong El Niño of 1997-98 stands out as an

345   especially large warm event within an overall upward trend.

Global Surface Temperature Anomalies
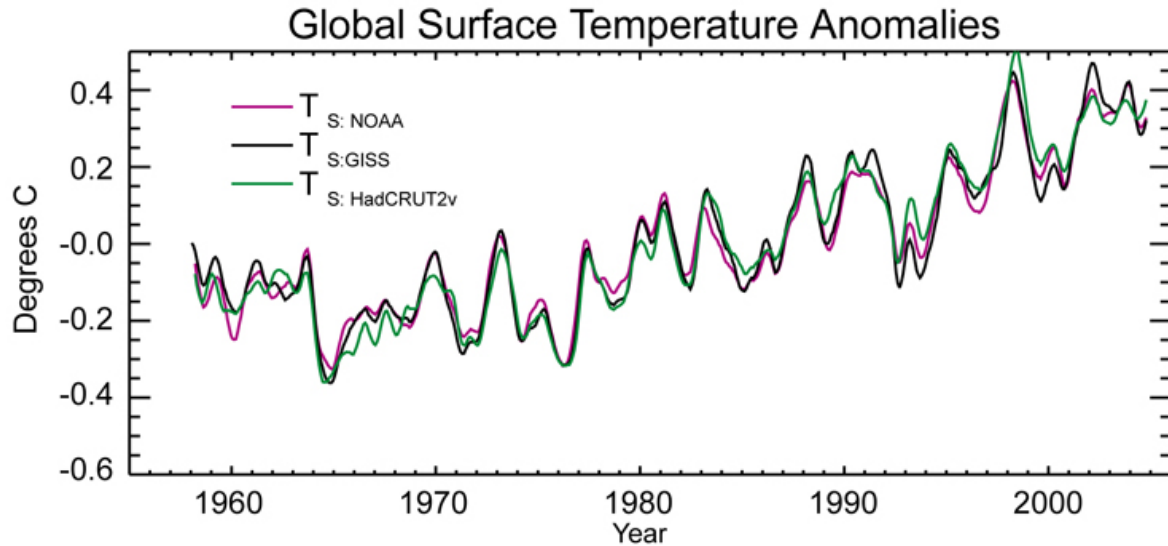
346
347 Figure 3.1 - Time series of globally averaged surface temperature (T_S) for NOAA (violet), GISS (black),
348 and HadCRUT2v (green) datasets. All time series are 7-month running averages (used as a smoother) of
349 original monthly data, which were expressed as a departure (ºC) from the 1979-97 average.
350

351

352 **3 RADIOSONDE TEMPERATURES**

353

354 **3.1   Balloon-borne temperature data**

355 Since the beginning of the radiosonde era, several thousand sites have been used to

356 launch balloons. However, many of these were in operation for only short periods of

357 time. One approach has been to use a fixed station network consisting of a smaller

358 number of stations having long periods of record. A complimentary approach is to grid

359 the data, using many more stations, allowing stations to join or drop out of the network

360 over the course of time. Since each approach has advantages and disadvantages, we

361 utilize both. A further complication is that changes over time in instruments and

362 recording practices have imparted artificial changes onto the temperature records. Some

363 groups have developed methods that try to remove these artificial effects as much as

364   possible. We employ two radiosonde datasets (see Table 3.1), one station-based and one

365   gridded. Both datasets have been constructed using homogeneity adjustments in an

366   attempt to minimize the effects of artificial changes.

367

368   **3.2 Radiosonde temperature datasets**

369

370   **3.2.1 NOAA (RATPAC)**

371   For several decades the 63 station dataset of Angell (Angell and Korshover, 1975) was

372   the most widely used station-based radiosonde temperature dataset for climate

373   monitoring. Recently, due to concerns regarding the effects of inhomogeneities, that

374   network shrank to 54 stations (Angell, 2003). To better address these concerns, LKS

375   (Lanzante, Klein, Seidel) (Lanzante et al., 2003a,b) built on the work of Angell by

376   applying homogeneity adjustment to the time series from many of his stations, as well as

377   several dozen additional stations, to create better regional representation via a network of

378   87 stations. However, because of the labor-intensive nature of the homogenization

379   process on these 87 stations, extension of the LKS dataset beyond 1997 is impractical.

380   Instead, the adjusted LKS dataset is being used as the basis for a new product (see Table

381   3.1), Radiosonde Air Temperature Products for Assessing Climate (RATPAC), that will

382   be updated regularly (Free et al., 2003; Free et al., 2005). A NOAA group (a

383   collaboration between the Air Resources Laboratory, the Geophysical Fluid Dynamics

384   Laboratory, and NCDC) is responsible for the creation of RATPAC.

385

386   The RATPAC product consists of two parts: RATPAC-A and RATPAC-B[5], both of

387   which use the adjusted LKS data, supplemented by an extension up to present using data

388   from the Integrated Global Radiosonde Archive (IGRA). The IGRA data used in

389   RATPAC are based on individual soundings that have been quality controlled and then

390   averaged into monthly station data (Durre, 2005). In this report we use RATPAC-B.

391   Generally speaking, based on data averaged over large regions such as the globe or

392   tropics, trends from RATPAC-A and RATPAC-B are closer to one another than they are

393   to the unadjusted (IGRA) data (Free et al., 2005).

394   **3.2.2 UK (HadAT2)**

395   For several decades the Oort (1983) product was the most widely used gridded

396   radiosonde dataset. With the retirement of Abraham Oort, and cessation of his product,

397   the dataset produced at the Hadley Centre, UK Met Office, HadRT (Parker et al., 1997)

398   became the most widely used gridded product. Because of concern about the effects of

399   artificial changes, this product incorporated homogeneity adjustments, although they

400   were somewhat limited[6]. As a successor to HadRT, the Hadley Centre has created a new

401   product (HadAT2, see Table 3.1) that uses all available digital radiosonde data for a

402   larger network of almost 700 stations having relatively long records[7]. Identification and

---

[5] RATPAC-A uses the adjusted LKS data up through 1997 and provides an extension beyond that using a
different technique to reduce the impact of inhomogeneities (Peterson et al., 1998). However, the
RATPAC-A methodology can only be used to derive homogenized temperature averaged over many
stations, and thus cannot be used to homogenize temperature time series at individual stations. RATPAC-B
consists of the LKS adjusted station time series that have been extended beyond 1997 by appending
(unadjusted) IGRA data.
[6] Adjustments were made to upper levels only (300 hPa and above), and since they were
based on satellite data, only since 1979.

[7] High quality small station subsets, such as Lanzante et al. (2003a) and the Global
Climate Observing System Upper Air Network, were used as a skeletal network from

---

403   adjustment of inhomogeneities was accomplished by way of comparison of neighboring

404   stations.

405

406   **3.2.3 Synopsis of radiosonde datasets**

407   The two chosen datasets differ fundamentally in their selection of stations in that the

408   NOAA dataset uses a relatively small number of highly scrutinized stations, while the

409   UK dataset uses a considerably larger number of stations. Compared to the surface, far

410   fewer thermometers are in use at any given time (hundreds or less) so there is less

411   opportunity for random errors to cancel, but more importantly, there are far fewer

412   suitable "neighbors" to aid in temporal homogenization. While both products incorporate

413   a common building-block dataset (Lanzante et al., 2003a), their methods of construction

414   differ considerably. Any differences in deductions about climate change derived from

415   them could be attributed to both the differing raw inputs as well as differing construction

416   methodologies. Concerns about poor temporal homogeneity are much greater than for

417   surface data.

418

419   **3.3 Global radiosonde temperature variations and differences between the datasets**

420   **3.3.1 Troposphere**

421   Figure 3.2a displays $T_{(850-300)}$ time series for the RATPAC and HadAT2 radiosonde

---

which to define a set of adequately similar station series used in homogenization. The
dataset is designed to impart consistency in both space and time and, by using radiosonde
neighbors rather than satellites or reanalyses, minimizes the chances of introducing
spurious changes related to the introduction of satellite data and their subsequent platform
changes (Thorne et al., 2005).

422   datasets. Several noteworthy features are common to both. First, just as for the surface,

423   ENSO signatures are clearly evident. Second, there is an apparent step-like rise of

424   temperature around 1976-77 associated with the well-documented climate regime shift

425   (Trenberth, 1990). Third, there is a long-term rise in temperatures, although a

426   considerable amount of it may be due to the step-like change (Seidel and Lanzante,

427   2004). To a first approximation, both datasets display these features similarly and there is

428   very little systematic difference between the two. Although a major component of the

429   RATPAC product is used in the construction of the HadAT2 dataset, it should be kept in

430   mind that the former utilizes a much smaller network of stations, although the length of

431   the station records tends to be relatively long. If the good agreement is not fortuitous, this

432   suggests that the reduced RATPAC station network provides representative spatial

433   sampling[8].

---

[8] This result is consistent with the relatively large spatial scales represented by a single radiosonde station
at this level on an annual time scale demonstrated by Wallis (1998) and Thorne et al. (2005).
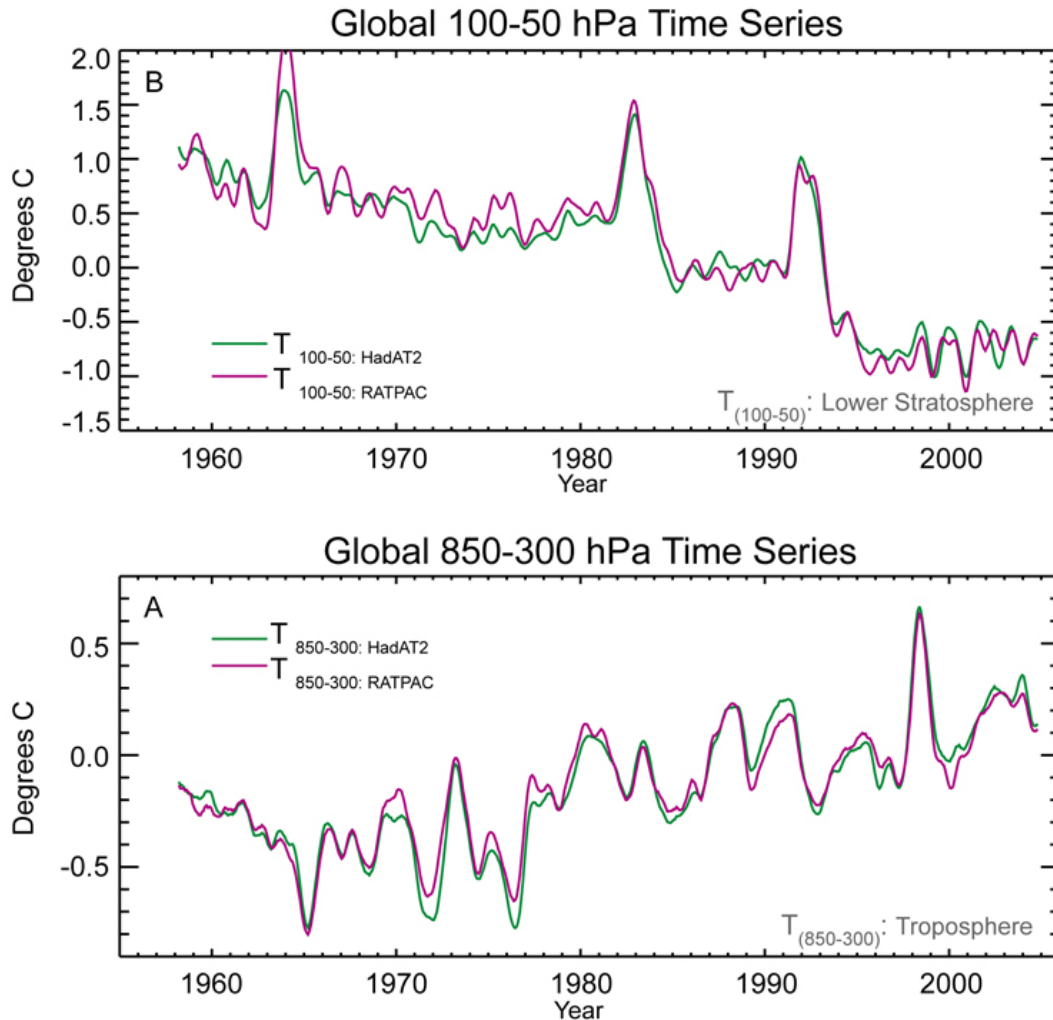
434
435   Figure 3.2a – Bottom: Time series of globally averaged tropospheric temperature (T$_{(850-300)}$) for RATPAC
436   (violet) and HadAT2 (green) radiosonde datasets. All time series are 7-month running averages (used as a
437   smoother) of original monthly data, which were expressed as a departure (ºC) from the 1979-97 average.
438
439   Figure 3.2b – Top: Time series of globally averaged stratospheric temperature (T$_{(100-50)}$) for RATPAC
440   (violet) and HadAT2 (green) radiosonde datasets. All time series are 7-month running averages (used as a
441   smoother) of original monthly data, which were expressed as a departure (ºC) from the 1979-97 average.
442

443

444   **3.3.2 Lower Stratosphere**

445   Figure 3.2b displays global temperature anomaly time series of T$_{(100-50)}$ from the

446   RATPAC and HadAT2 radiosonde datasets. Several noteworthy features are common to

447   both datasets. First is the prominent signature of three climatically important volcanic

448   eruptions: Agung (March 1963), El Chichon (April 1982), and Pinatubo (June 1991).

449   Temperatures rise rapidly as volcanic aerosols are injected into the stratosphere and

450   remain elevated for about 2-3 years before diminishing. There is some ambiguity as to

451   whether the temperatures return to their earlier values or whether they experience step-

452   like falls in the post-volcanic period for the latter two volcanoes, particularly Pinatubo

453   (Pawson et al., 1998; Lanzante et al., 2003a; Seidel and Lanzante, 2004). Second, there

454   are small amplitude variations associated with the tropical quasi-biennial oscillation

455   (QBO) with a period of ~ 2-3 years (Seidel et al., 2004). Third, there is a downward

456   trend, although there is some doubt as to whether the temperature decrease is best

457   described by a linear trend over the period of record. For one thing, the temperature series

458   prior to about 1980 exhibits little or no decrease in temperature. After that, the

459   aforementioned step-like drops represent a viable alternative to a linear decrease (Seidel

460   and Lanzante, 2004).

461

462   In spite of similarities among datasets, closer examination reveals some important

463   differences. There is a rather large difference between RATPAC and HadAT2 time series

464   for the peak volcanic warming associated with Agung in 1963. This may be a reflection

465   of differences in spatial sampling because the horizontal pattern of the response is not

466   uniform (Free and Angell, 2002). More noteworthy for estimates of climate change are

467   some subtle systematic differences between the two datasets that vary over time. A closer

468   examination reveals that the RATPAC product tends to have higher temperatures than the

469   HadAT2 product from approximately 1963-85, with the RATPAC product having lower

470   values before and after this time period[9]. As we will see later, this yields a slightly greater

471   decreasing trend for the RATPAC product. Poorer agreement between the RATPAC and

472   HadAT2 products in the stratosphere compared to the troposphere is not unexpected

473   because of the fact that artificial jumps in temperature induced by changes in radiosonde

474   instruments and measurement systems tend to increase in magnitude from the near-

475   surface upwards (Lanzante et al., 2003b). More details on this issue are given in Chapter

476   4, Section 2.1.

477

478   **4 SATELLITE-DERIVED TEMPERATURES**

479

480   **4.1   Microwave satellite data**

481   Three groups, employing different methodologies, have developed satellite Microwave

482   Sounding Unit (MSU) climate datasets (see Table 3.1). We do not present results from a

483   fourth group (Prabhakara et al., 2000), which developed yet another methodology, since

484   they are not continuing to work on MSU climate analyses and are not updating their time

485   series. One of the main issues that is addressed differently by the groups is the inter-

486   calibration between the series of satellites, and is discussed in Chapters 2 and 4.

487   **4.2 Microwave Satellite Datasets**

488

---

[9] It is worth noting that prominent artificial step-like drops, many of which were associated with the
adoption of a particular type of radiosonde (Vaisala), were found in stratospheric temperatures at Australian
and western tropical Pacific stations in the mid to late 1980s by Parker et al. (1997), Stendel et al. (2000),
and Lanzante et al. (2003a Differences in consequent homogeneity adjustments around this time could
potentially explain a major part of the difference between the NOAA and UK products, although this has
not been demonstrated.

489   **4.2.1 University of Alabama In Huntsville (UAH)**

490   The first group to produce MSU climate products, by adjusting for the differences

491   between satellites and the effects of changing orbits (diurnal drift), was UAH (A). Their

492   approach (Christy et al., 2000; Christy et al., 2003) uses both an offset adjustment to

493   allow for the systematic average differences between satellites and a non-linear hot target

494   temperature[10] calibration to create a homogeneous series. The UAH dataset has products

495   corresponding to three temperature measures: T_{2LT}, T_2, and T_4 (see Chapter 2 for

496   definitions of these measures). In this report we use the most up to date versions available

497   to us at the time, which is version 5.1 of the UAH dataset for T_2 and T_4, and version 5.2

498   for T_{2LT}[11].

499

500   **4.2.2 Remote Sensing Systems (RSS)**

501   After carefully studying the methodology of the UAH team, another group, RSS (R)

502   created their own datasets for T_2 and T_4 using the same input data but with modifications

503   to the adjustment procedure (Mears et al., 2003), two of which are particularly

504   noteworthy: (1) the method of inter-calibration from one satellite to the next and (2) the

505   computation of the needed correction for the daily cycle of temperature. While the second

506   modification has little effect on the overall global trend differences between the two

507   teams, the first is quite important in this regard. Recently the RSS team has created their

508   own version of T_{2LT} (Mears and Wentz, 2005) and in doing so discovered a

---

[10] In fact, two targets are used, both with temperatures that are presumed to be well known. These are *cold* space, pointing away from the Earth, Moon, or Sun, and an onboard *hot* target.
[11] The version number for T_{2LT} differs from that for T_2 and T_4 because an error, which was found to affect the former (and was subsequently corrected), does not affect the latter two measures.

509   methodological error in the corresponding temperature measure of UAH. The UAH T_{2LT}

510   product used in this report is based on their corrected method. In this report we use

511   version 2.1 of the RSS data.

512

513   **4.2.3 University of Maryland (U.Md.)**

514   A very different approach (Vinnikov et al., 2004) was developed by a team involving

515   collaborators from the University of Maryland and the NOAA National Environmental

516   Satellite, Data, and Information Service (NESDIS) and was used to estimate globally

517   averaged temperature trends (Vinnikov and Grody, 2003). After further study, they

518   developed yet another new method (Grody et al., 2004). As done by the other two groups,

519   the U.Md. (M) team's methodology also recalibrates the instruments based on

520   overlapping data between the satellites. However, the manner in which they perform this

521   recalibration differs. Also, they do not adjust for diurnal drift directly, but average the

522   data from ascending and descending orbits. In their second approach, they substantially

523   altered the manner in which target temperatures are used in their recalibration to a

524   scheme more consistent with that of the other two groups (UAH and RSS). The effect of

525   their revision was to reduce the global temperature trends derived from their data from

526   0.22-0.26 to 0.17 ºC/decade. Very recently they have revised their method to produce a

527   third version of their dataset, which we use in this report, whose trends differ only

528   slightly with those from the second version. In this most recent version they apply the

529   nonlinear adjustment of Grody et al. (2004) and estimate the diurnal cycle as described in

530   Vinnikov et al. (2005). The U.Md. group produces only a measure of T_2, hence there is

531   no stratospheric product (T_4) or one corresponding to the lower troposphere (T_{2LT}).

532

533   **4.3 Synopsis of satellite datasets**

534   The relationship among satellite datasets is fundamentally different from that for surface

535   or radiosonde products. For satellites, different datasets use virtually the same raw inputs

536   so that any differences in derived measures are due to construction methodology. The

537   excellent coverage provided by the orbiting sensors, more than half the Earth's surface

538   daily, is a major advantage over in situ observations. The disadvantage is that while in

539   situ observations rely on data from many hundreds or thousands of individual

540   thermometers every day, providing a beneficial redundancy, the satellite data typically

541   come from only one or two instruments at a given time. Therefore, any problem

542   impacting the data from a single satellite can adversely impact the entire climate record.

543   The lack of redundancy, compounded by occasional premature satellite failure that limits

544   the time of overlapping measurements from successive satellites, elevates the issue of

545   temporal homogeneity to the overwhelming explanation for any differences in deductions

546   about climate change derived from the three datasets.

547

548   **4.4   Global satellite temperature variations and differences between the datasets**

549

550   **4.4.1 Temperature of the Troposphere**

551   Two groups (UAH and RSS) produce lower tropospheric temperature datasets, T_{2LT} (see

552   Chapter 2 for definition of this and related temperature measures) directly from satellite

553   measurements. Their time series are shown in Figure 3.3a along with an equivalent

554   measure constructed from the HadAT2 radiosonde dataset (see Box 2.2 for an

555   explanation as to how these equivalent measures were generated).The three temperature

556   series have quite similar behavior, with ENSO-related variations accounting for much of

557   the up and down meanderings, for example the historically prominent 1997-1998 El

558   Niño. But over the full period of record, the amount of increase indicated by the datasets

559   varies considerably. A closer look reveals that as time goes on, the RSS product indicates

560   a noticeably greater increase of temperature than the other two. For comparison purposes,

561   in Figure 3.3b we show an alternate measure of lower tropospheric temperatures, T^*_G ,

562   derived from products produced by the same three groups. From comparison of Figures

563   3.3a and 3.3b we see that both measures of lower tropospheric temperature agree

564   remarkably well, even with regard to the more subtle differences relating to the longer-

565   term changes. We will return to the issue of agreement between T_{2LT} and T^*_G later when

566   we discuss trends (section 6).

567
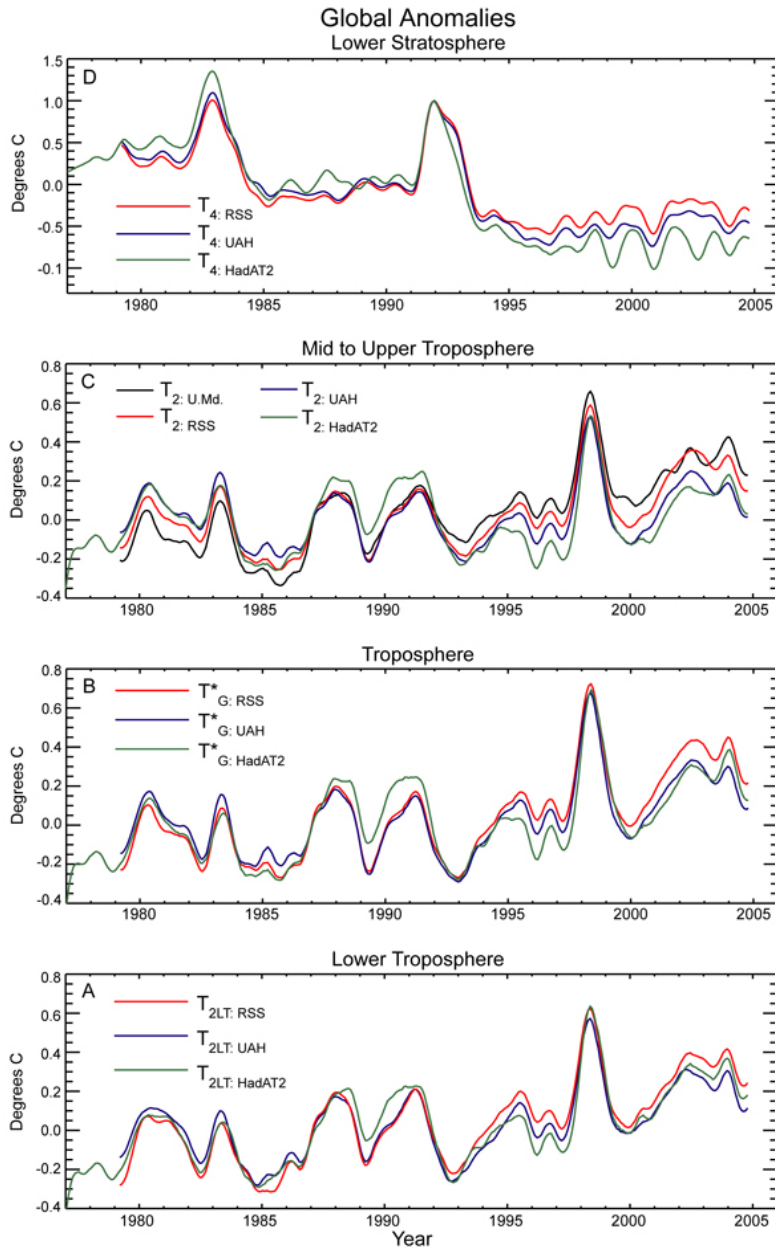
Figure 3.3a– Bottom: Time series of globally averaged lower tropospheric temperature (T_{2LT}) as follows:
UAH (blue) and RSS (red) satellite datasets, and HadAT2 (green) radiosonde data. All time series are 7-
month running averages (used as a smoother) of original monthly data, which were expressed as a
departure (ºC) from the 1979-97 average.

Figure 3.3b– Third: Time series of globally averaged middle tropospheric temperature (T^*_G) as follows:
UAH (blue) and RSS (red) satellite datasets, and HadAT2 (green) radiosonde data. All time series are 7-
month running averages (used as a smoother) of original monthly data, which were expressed as a
departure (ºC) from the 1979-97 average.

Figure 3.3c – Second: Time series of globally averaged upper middle tropospheric temperature (T_2) as

579   follows: UAH) (blue), RSS (red), and U.Md. (black) satellite datasets, and HadAT2 (green) radiosonde
580   data. All time series are 7-month running averages (used as a smoother) of original monthly data, which
581   were expressed as a departure (ºC) from the 1979-97 average.
582
583   Figure 3.3d – Top: Time series of globally averaged lower stratospheric temperature (T_4) as follows: UAH
584   (blue) and RSS (red) satellite datasets, and HadAT2 (green) radiosonde data. All time series are 7-month
585   running averages (used as a smoother) of original monthly data, which were expressed as a departure (ºC)
586   from the 1979-97 average.

587

588   Time series corresponding to the temperature of the upper middle troposphere (T_2) are

589   shown in Figures 3.3c. The products represented in this figure are the same as for the

590   lower troposphere, except that an additional product, that from the U.Md. group is

591   available. Again, all of the time series have similar behavior with regard to the year to

592   year variations. However, closer examination shows that two of the products (U.Md. and

593   RSS satellite data) indicate considerable temperature increase over the period of record,

594   whereas the other two (UAH satellite and HadAT2 radiosonde) indicate slight warming

595   only. A more detailed discussion of the differences between the various products can be

596   found in Chapter 4.

597

598   We note that all of the curves for the various tropospheric temperature series (Figures

599   3.3a-c) exhibit remarkably similar shape over the period of record. For the common time

600   period, the satellite measures are similar to the tropospheric layer-averages computed

601   from radiosonde data. The important differences between the various series are with

602   regard to the more subtle long-term evolution over time, which manifests itself as

603   differences in linear trend, discussed later in more detail.

604

605
606   **4.4.2 Temperature of the Lower Stratosphere**

607   Figure 3.3d shows the temperature of the lower stratosphere (T_4); note that there is no

608   product from the U.Md. team for this layer. The dominant features for this layer are the

609   major volcanic eruptions: El Chichon in 1982 and Pinatubo in 1991. As discussed above,

610   the volcanic aerosols tend to warm the stratosphere for about 2-3 years before

611   diminishing. In contrast, ENSO events have little influence on the stratospheric

612   temperature. Both products show that the stratospheric temperature has decreased

613   considerably since 1979, as compared to the lesser amount of increase that is seen in the

614   troposphere. The $T_{4\text{-R}}$ product shows somewhat less overall decrease than the $T_{4\text{-A}}$

615   product, in large part as a result of the fact that the former increases relative to the latter

616   from about 1992-94. As was the case for the troposphere, the radiosonde series show a

617   greater decrease than the satellite data. Again, the satellite and radiosonde series for the

618   lower-stratosphere exhibit the same general behavior over time.

619

620

621   **5 REANALYSIS TEMPERATURE "DATA"**

622

623   A number of agencies from around the world have produced reanalyses based on

624   different schemes for different time periods. We focus on two of the most widely

625   referenced, which cover a longer time period than the others (see Table 3.1). The U.S.

626   reanalysis represents a collaborative effort between NOAA's National Center for

627   Environmental Prediction (NCEP) and the National Center for Atmospheric Research

628   (NCAR). For U.S. reanalysis, gridded air temperatures at the surface and aloft are

629   available from 1958 to present. Using a completely different system, the European Center

630   for Medium-Range Weather Forecasts (ECMWF) has produced similar gridded data from

631   September 1957 to August 2002 (Europe). Reanalyses are "hybrid products," utilizing

632   raw input data of many types, as well as complex mathematical models to combine these

633   data. For more detailed information on the reanalyses, see Chapter 2. As the reanalysis

634   output does not represent a different observing platform, a separate assessment of

635   reanalysis data will not be made.

636

637   **6. COMPARISONS BETWEEN DIFFERENT LAYERS AND OBSERVING**

638   **PLATFORMS**

639

640   **6.1   During the radiosonde era, 1958 to the present**

641   **6.1.1 Global**

642   As shown in earlier sections, globally averaged temperature time series indicate

643   increasing temperature at the surface and in the troposphere with decreases in the

644   stratosphere over the course of the last several decades. It is desirable to derive some

645   estimates of the magnitude of the rate of these changes. The widely-used, least-squares,

646   linear trend technique is adopted for this purpose with the explicit caveat that long-term

647   changes in temperature are not necessarily linear, as there may be departures in the form

648   of periods of enhanced or diminished change, either linear or nonlinear, as well as abrupt,

649   step-like changes[12]. While it has been shown that such constructs are plausible, it is

650   nevertheless difficult to prove that they provide a better fit to the data, over the time

651   periods addressed in this report, than the simple linear model (Seidel and Lanzante,

652   2004). Additional discussion on this topic is given in the Statistical Appendix.

653

654   Trends computed for the radiosonde era are given in Table 3.2 for the surface as well as

655   various tropospheric and stratospheric layer averages[13]. The surface products are quite

656   consistent with one another, as are the radiosonde products in the troposphere. In the

657   stratosphere, the radiosonde products differ somewhat, although there is an inconsistent

658   relationship involving the two stratospheric measures ($T_{(100-50)}$ and T_4) regarding which

659   product indicates a greater decrease in temperature[14]. The reanalysis products, which are

660   "hybrid-measures," agree better with the "purer" surface and radiosonde measures at and

661   near the surface. Agreement degrades with increasing altitude such that the reanalyses

662   indicate more tropospheric temperature increase and considerably less stratospheric

663   decrease than do the radiosonde products. The disparity between the reanalyses and other

664   products is not surprising given the suspect temporal homogeneity of the reanalyses (see

665   Chapter 2, Section 1c).

---

[12] For example, the tropospheric linear trends in the periods 1958-1979 and 1979-2003 were shown to be much less than the trend for the full period (1958-2003), based on one particular radiosonde dataset (Thorne et al., 2005), due to the abrupt rise in temperature in the mid 1970s.

[13] Note that it is instructive to examine the behavior of radiosonde and reanalysis temperatures averaged in such a way as to correspond to the satellite layers (T_{2LT}, T^*_G, T_2, and T_4) even though there are no comparable satellite measures prior to 1979.

[14] The reason for this inconsistency is that the HadAT2 product records data at fewer vertical levels than the RATPAC product, so the comparison is not one-to-one.

666   Table 3.2 - Global temperature trends in ºC per decade from 1958 through 2004 (except for European
667   which terminates September 2001) calculated for the surface or atmospheric layers by data source. The
668   trend is shown for each, with the approximate 95% confidence interval (2 sigma) below in parentheses. The
669   levels/layers, from left to right, go from the lowest to the highest in the atmosphere. Bold values are
670   estimated to be statistically significantly different from zero (at the 5% level). A Student's t-test, using the
671   lag-1 autocorrelation to account for the non-independence of residual values about the trend line, was used
672   to assess significance (see Appendix for discussion of confidence intervals and significance testing).

673

	T_S	T_{2LT}	$T_{(850-300)}$	T^*_G	T_2	$T_{(100-50)}$	T_4
Surface:							
NOAA	**0.11** (0.017)						
GISS	**0.11** (0.021)						
HadCRUT2v	**0.13** (0.021)						
Radiosonde:							
RATPAC	**0.11** (0.022)	**0.13** (0.026)	**0.13** (0.030)	**0.13** (0.032)	**0.07** (0.030)	**-0.41** (0.093)	**-0.36** (0.082)
HadAT2	**0.12** (0.026)	**0.16** (0.036)	**0.14** (0.039)	**0.15** (0.041)	**0.08** (0.040)	**-0.39** (0.084)	**-0.38** (0.083)
Reanalyses:							
US	**0.12** (0.030)	**0.15** (0.046)	**0.17** (0.052)	**0.17** (0.057)	**0.13** (0.064)	-0.18 (0.232)	-0.18 (0.223)
European	**0.11** (0.027)	**0.15** (0.042)	**0.15** (0.042)	**0.14** (0.044)	**0.10** (0.040)	**-0.21** (0.128)	**-0.17** (0.134)

674

675

676   Perhaps the most important result shown in Table 3.2 is that both the radiosonde and

677   reanalysis trends indicate that the tropospheric temperature has increased as fast as or

678   faster than the surface over the period 1958 to present. For a given dataset, the 3

679   measures (T_{2LT}, $T_{(850-300)}$, and T^*_G) always indicate more increase in the troposphere than

680   at the surface, although this is usually not true when the T_2 measure is considered. The

681   reason for the inconsistency involving T_2 is because of contributions to the layer that it

682   measures from stratospheric cooling, an effect first recognized by Spencer and Christy

683   (1992) (see discussion of this issue in Chapters 2 and 4). The development of T^*_G as a

684   global measure, and its counterpart, T^*_T for the tropics (Fu et al., 2004; Fu and Johanson,

685   2005; Johanson and Fu, 2005) was an attempt to remove the confounding effects of the

686   stratosphere using a statistical approach (see Chapter 2).

687   **6.1.2 Land vs. ocean**

688   Most of the land and ocean surface temperature increased during the radiosonde era, with

689   the exception of parts of the North Atlantic Ocean, the North Pacific Ocean, and a few

690   smaller areas. With a few exceptions, such as the west coast of North America, trends in

691   land air temperature in coastal regions are generally consistent with trends in SST over

692   neighboring ocean areas (Houghton et al., 2001). Because bias adjustments are performed

693   separately for land and ocean areas, before merging to create a global product, it is

694   unlikely that the land-ocean consistency is an artifact of the construction methods used in

695   the various surface analyses. However, land air temperatures did increase somewhat more

696   rapidly than SSTs in some regions during the past two decades. Possibly related to this is

697   the fact that since the mid-1970s, El Niño has frequently been in its "warm" phase, which

698   tends to bring higher than normal temperatures to much of North America, among other

699   regions, which have had strong temperature increases over the past few decades (Hurrell,

700   1996). Also, when global temperatures are rising or falling, the global mean land

701   temperature tends to both rise and fall faster than the ocean, which has a tremendous heat

702   storage capacity (Waple and Lawrimore, 2003).

703

704   **6.1.3 Marine air vs. sea surface temperature**

705   In ocean areas, it is natural to consider whether the temperature of the air and that of the

706   ocean surface (SST) increases or decreases at the same rate. Several studies have

707   examined this question. Overall, on seasonal and longer scales, the SST and marine air

708   temperature generally move at about the same rate globally and in many ocean basin

709   scale regions (Bottomley et al., 1990; Parker et al., 1995; Folland et al., 2001b; Rayner et

710   al., 2003). Differences between SST and marine air temperature in some regions were

711   first noted by Christy et al. (1998) and then examined in more detail by Christy et al.

712   (2001). The latter study found that in the tropics, SST increased more than NMAT from

713   1979 –1999 derived from the Tropical Atmosphere Ocean (TAO) array of tropical buoys

714   and transient marine ship observations. But this difference may be related to changes in

715   surface fluxes associated with ENSO and the interdecadal Pacific oscillation (Folland et

716   al., 2003). Consistent results were found using two datasets, one with more widespread

717   observations from ships, and another, which sampled a more limited number of locations

718   using moored buoys. There were some indications that the accelerated increase of SST

719   compared to air temperature may have been concentrated in two periods: the early 1980s

720   and mid 1990s. So over the satellite era, there are some unexplained differences in these

721   trends that were also noted by Folland et al. (2003) in parts of the tropical south Pacific

722   using the Rayner et al. (2003) NMAT dataset which incorporates new corrections for the

723   effect on NMAT of increasing deck (and hence measurement) heights.

724

725   **6.1.4 Minimum vs. maximum temperatures over land**

726   Daily minimum temperature increased about twice as fast as daily maximum temperature

727   over global land areas during the radiosonde era (Karl et al., 1993; Easterling et al., 1997;

728   Folland et al, 2001b). However, a closer look at recent years has found that during the

729   satellite era, maximum and minimum temperature have been rising at the same rate (Vose

730   et al, 2005). Daily minimum temperature increased in virtually all areas except eastern

731   Canada, Eastern Europe, and other scattered regions often near coasts. Most regions also

732   witnessed an increase in the daily maximum, but over the longer time frame the rate of

733   increase was generally smaller, and decreasing trends were somewhat more common

734   (e.g., in eastern Canada, the southern United States, southern China, eastern Europe, and

735   portions of South America). The causes of this asymmetric warming are still debated, but

736   many of the areas with greater increases of minimum temperatures correspond to those

737   where cloudiness appears to have increased over the period as a whole (Dai et al., 1999;

738   Henderson-Sellers, 1992; Sun and Groisman, 2000; Groisman et al., 2004). This makes

739   physical sense since clouds tend to cool the surface during the day by reflecting incoming

740   solar radiation, and warm the surface at night by absorbing and reradiating infrared

741   radiation back to the surface.

742

6.2   During the satellite era, 1979 to the present

744

6.2.1 Global

746   A comparable set of global trends for the satellite era is given in Table 3.3. Comparison

747   between Tables 3.2 and 3.3 reveals that some of the relationships between levels and

748   layers, as well as among datasets, are different during the two eras. Comparing satellite

749   era trends with the radiosonde era trends for datasets that have both periods in common, it

750   is clear that the surface temperature increase (see Figure 3.1) has accelerated in recent

751   decades while the tropospheric increase (see Figure 3.2a) has decelerated. Since most of

752   the stratospheric decrease has occurred since 1979 (see Figure 3.2b) the rate of

753   temperature decrease there is close to twice as large as during the full radiosonde era.

754   Thus, care must be taken when interpreting results from only the most recent decades.

755   Agreement among different surface and radiosonde datasets is reasonable and about as

756   good as during the longer radiosonde era. The reanalysis datasets show poorer agreement

757   with surface data and especially with stratospheric radiosonde data for the European

758   product.

759   Table 3.3 - Global temperature trends in ℃ per decade from 1979 through 2004 (except for European
760   which terminates September 2001) calculated for the surface or atmospheric layers by data source. The
761   trend is shown for each, with the approximate 95% confidence interval (2 sigma) below in parentheses. The
762   levels/layers, from left to right, go from the lowest to the highest in the atmosphere. Bold values are
763   estimated to be statistically significantly different from zero (at the 5% level). A Student's t-test, using the
764   lag-1 autocorrelation to account for the non-independence of residual values about the trend line, was used
765   to assess significance (see Appendix for discussion of confidence intervals and significance testing).

766

	T_S	T_{2LT}	$T_{(850-300)}$	T^*_G	T_2	$T_{(100-50)}$	T_4
Surface:							
NOAA	**0.16** (0.035)						
GISS	**0.16** (0.043)						
HadCRUT2v	**0.17** (0.037)						
Radiosonde:							
RATPAC	**0.17** (0.050)	**0.13** (0.057)	**0.10** (0.065)	**0.11** (0.075)	0.02 (0.071)	**-0.70** (0.240)	**-0.65** (0.213)
HadAT2	**0.18** (0.050)	**0.14** (0.071)	**0.12** (0.075)	**0.12** (0.084)	0.03 (0.080)	**-0.63** (0.241)	**-0.64** (0.238)
Satellite:							
UAH		**0.12** (0.082)		**0.12** (0.089)	0.04 (0.078)		**-0.45** (0.421)
RSS		**0.19** (0.081)		**0.19** (0.089)	**0.13** (0.077)		-0.33 (0.382)
U.Md.					**0.20** (0.066)		
Reanalyses:							
US	**0.12** (0.074)	**0.12** (0.100)	**0.11** (0.101)	0.06 (0.106)	-0.04 (0.101)	**-0.76** (0.450)	**-0.74** (0.441)
European	**0.11** (0.060)	**0.11** (0.101)	0.10 (0.102)	**0.13** (0.106)	0.07 (0.096)	-0.31 (0.529)	-0.34 (0.493)

767

768

769 Comparisons of trends between different satellite products and between satellite and

770 radiosonde products yields a range of results as indicated by examination of the

771   numerical trend values found in Table 3.3, which are also graphed in Figure 3.4a. While

772   the tropospheric satellite products from the UAH team have trends that are not too

773   dissimilar from the corresponding radiosonde trends, the two other satellite datasets show

774   a considerably greater increase in tropospheric temperature. In the stratosphere, there is a

775   large disagreement between satellite and radiosonde products, with the latter indicating

776   much greater decreases in temperature. Here too, the reanalyses are quite inconsistent,

777   with the European product closer to the satellites and the U.S. product closer to the

778   radiosondes.
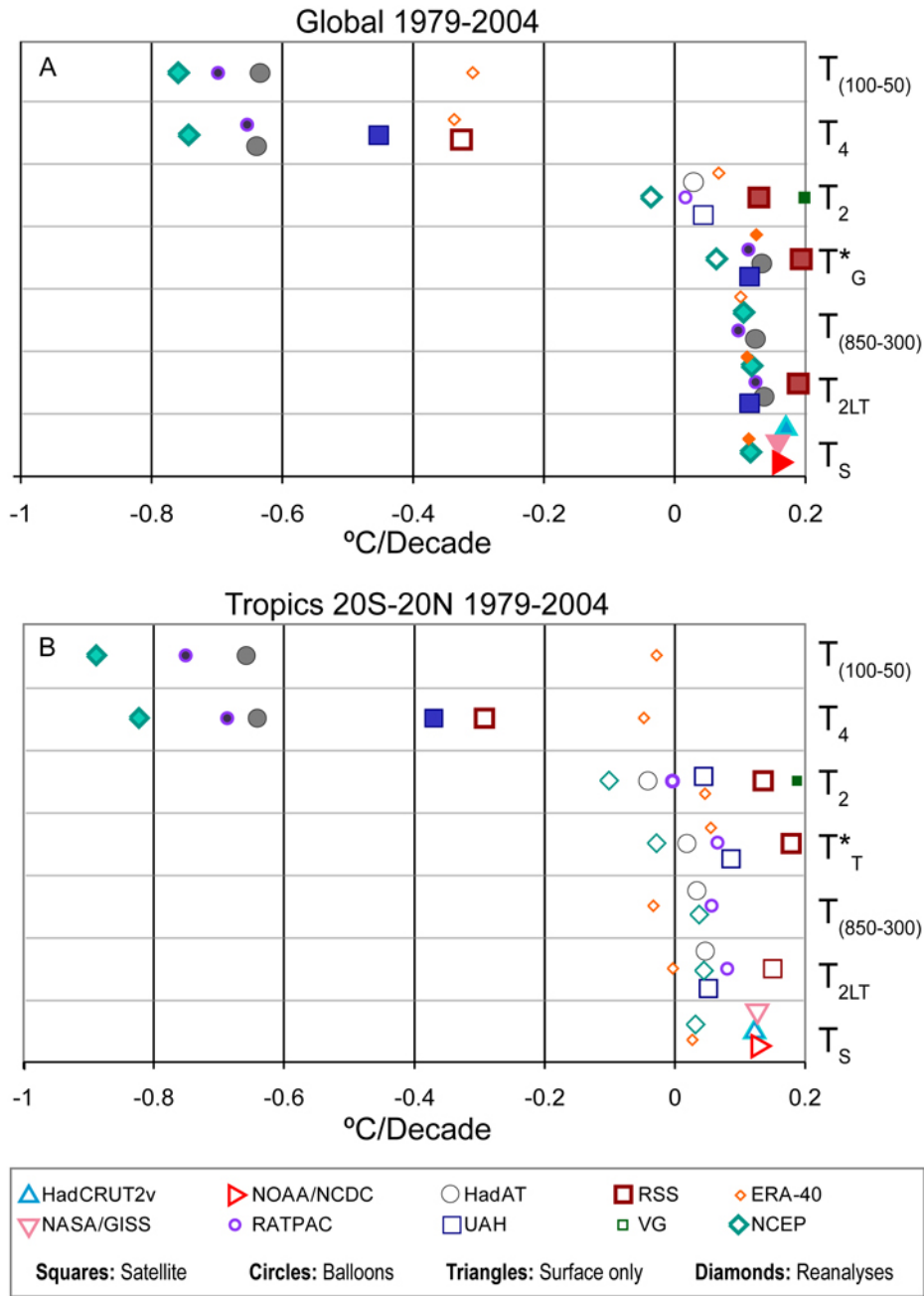
779

780   Figure 3.4a (top) – Global temperature trends (ºC/decade) for 1979-2004 from Table 3.3 plotted as
781   symbols. See figure legend for definition of symbols. Filled symbols denote trends estimated to be
782   statistically significantly different from zero (at the 5% level). A Student's t-test, using the lag-1
783   autocorrelation to account for the non-independence of residual values about the trend line, was used to
784   assess significance (see Appendix for discussion of confidence intervals and significance testing).

785

786   Figure 3.4b (bottom) – Tropical (20°N-20°S) temperature trends (ºC/decade) for 1979-2004 from Table 3.4
787   plotted as symbols. See figure legend for definition of symbols. Filled symbols denote trends estimated to
788   be statistically significantly different from zero (at the 5% level). A Student's t-test, using the lag-1
789   autocorrelation to account for the non-independence of residual values about the trend line, was used to
790   assess significance (see Appendix for discussion of confidence intervals and significance testing).

791

792   Perhaps the most important issue is the relationship between trends at the surface and in

793   the troposphere. As shown in Table 3.3 and Figure 3.4a, both radiosonde datasets as well

794   as the UAH satellite products indicate that, in contrast to the longer radiosonde era,

795   during the satellite era the temperature of the surface has increased more than that of the

796   troposphere. However, tropospheric trends from the RSS satellite dataset, based on both

797   measures of temperature having little or no stratospheric influence (T_{2LT} and T^*_G) yield

798   an opposing conclusion: the tropospheric temperature has increased as much or more than

799   the surface. For the third satellite dataset, comparisons with surface temperature are

800   complicated by the fact that the U.Md. team produces only T_2, which is influenced by

801   stratospheric cooling (see Chapter 2). Nevertheless, we can infer that it too suggests more

802   of a tropospheric temperature increase than that at the surface[15].

803

804   Since global change theory suggests more warming of the troposphere than the surface

805   only in the tropics (see Chapter 1), much of the interest in observed trends has been in

806   this region. Therefore, to compliment the global trends (Figure 3.4a and Table 3.3), we

807   present a similar plot of tropical trends in Figure 3.4b (with corresponding trend values in

---

[15] The difference in trends, T^*_G minus T_2, for the UAH and RSS datasets is about 0.06 to 0.08 ºC/decade.
Adding this amount to the U.Md. T_2 trend (0.20 ºC/decade) yields an estimate of the U.Md. trend in T^*_G of
about 0.26 to 0.28 ºC/decade. In this calculation we are assuming that the effects of the stratospheric
cooling trend on the U.Md. product are the same as from the UAH and RSS datasets.

808   Table 3.4). Compared to the global trends, the tropical trends show even more spread

809   among datasets, particularly in the lower stratosphere[16]. The result of the greater spread is

810   that the range of plausible values for the difference in trends between the surface and

811   troposphere is larger than that for the globe as a whole. Similar to the global case, in the

812   tropics the UAH satellite plus the two radiosonde datasets (RATPAC and HadAT2)

813   suggest more warming at the surface than in the troposphere, while the opposite

814   conclusion is reached based on the other two satellite products (RSS and U.Md.).

815   Resolution of this issue would seem to be of paramount importance in the interpretation

816   of observed climate change central to this synthesis assessment.

817

818

819   Table 3.4 – Tropical (20°N-20°S) temperature trends in ℃ per decade from 1979 through 2004 (except for
820   European which terminates September 2001) calculated for the surface or atmospheric layers by data
821   source. The trend is shown for each, with the approximate 95% confidence interval (2 sigma) below in
822   parentheses. The levels/layers, from left to right, go from the lowest to the highest in the atmosphere. Bold
823   values are estimated to be statistically significantly different from zero (at the 5% level). A Student's t-test,
824   using the lag-1 autocorrelation to account for the non-independence of residual values about the trend line,
825   was used to assess significance (see Appendix for discussion of confidence intervals and significance
826   testing).

827

	T_S	T_{2LT}	$T_{(850-300)}$	T^*_G	T_2	$T_{(100-50)}$	T_4
Surface:							
NOAA	0.13 (0.149)						
GISS	0.13 (0.152)						
HadCRUT2v	0.12 (0.172)						
Radiosonde:							

---

[16] The larger spread may be partially an artifact of the fact that when averaging over a smaller region, there is less cancellation of random variations. In addition, the fact that the networks of in situ observations are much sparser in the tropics than in the extratropics of the Northern Hemisphere may also contribute.

RATPAC	**0.13** (0.068)	0.08 (0.119)	0.06 (0.136)	0.07 (0.153)	0.00 (0.140)	**-0.75** (0.362)	**-0.69** (0.289)
HadAT2	**0.15** (0.115)	0.05 (0.152)	0.03 (0.164)	0.02 (0.176)	-0.04 (0.170)	**-0.66** (0.304)	**-0.64** (0.307)
Satellite:							
UAH		0.05 (0.176)		0.09 (0.191)	0.05 (0.167)		**-0.37** (0.281)
RSS		0.15 (0.192)		0.18 (0.196)	0.14 (0.175)		-0.29 (0.303)
U.Md.					**0.19** (0.159)		
Reanalyses:							
US	0.03 (0.163)	0.05 (0.172)	0.04 (0.173)	-0.03 (0.183)	-0.10 (0.166)	**-0.89** (0.405)	**-0.83** (0.340)
European	0.03 (0.211)	0.00 (0.234)	-0.03 (0.249)	0.06 (0.255)	0.05 (0.232)	-0.03 (0.453)	-0.05 (0.423)

828

6.2.2 Latitude bands

Globally averaged temperatures paint only part of the picture. Different layers of the atmosphere behave differently depending on the latitude. Furthermore, even the processing of the data can make for latitudinal difference in long-term trends. Figure 3.5 shows the trends in temperature for different datasets and levels averaged over latitude bands. Each of these trends was created by making a latitudinally averaged time series of monthly anomalies and then fitting that time series with a standard least-squares linear regression slope.
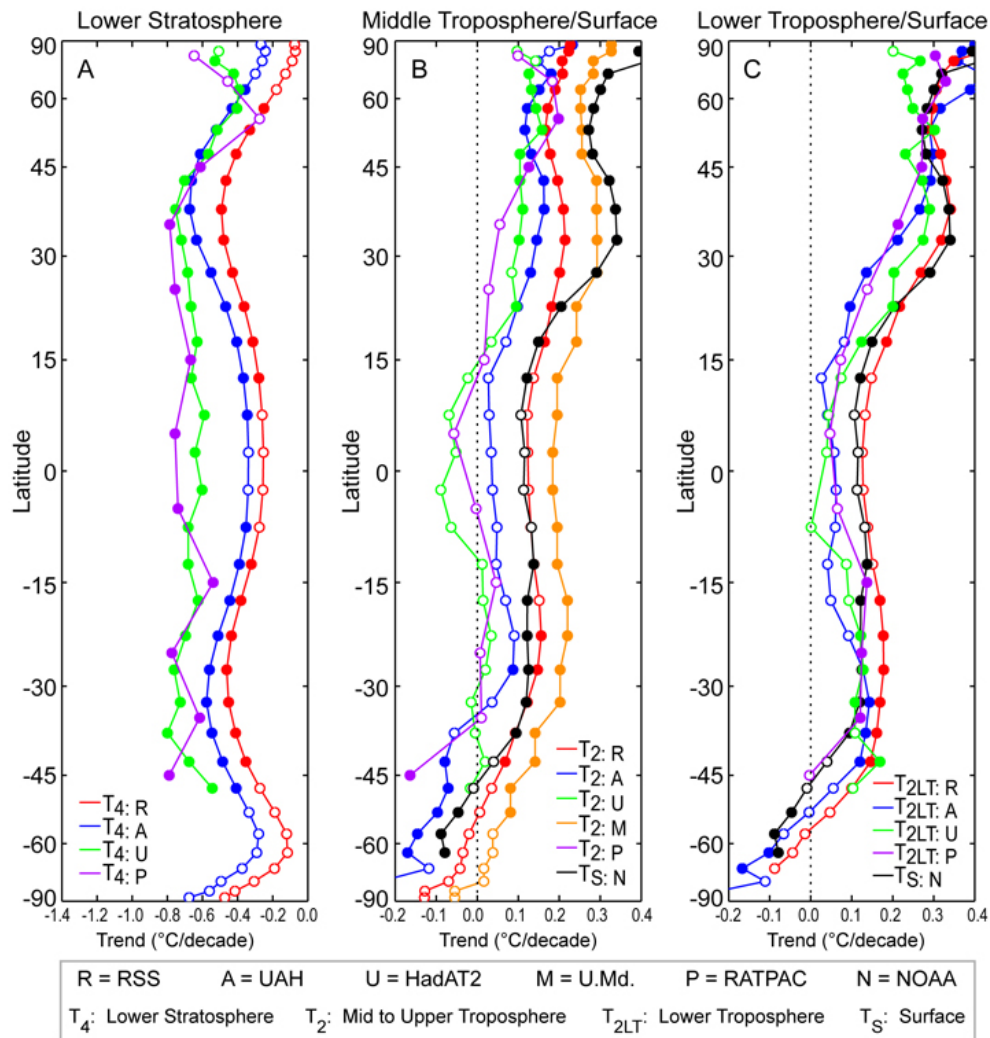
Figure 3.5 -- Temperature trends for 1979-2004 (ºC/decade) by latitude.
Left: stratospheric temperature (T₄) based on RSS (red) and UAH (blue) satellite datasets, and RATPAC (violet) and HadAT2 (green) radiosonde datasets.
Middle: mid-tropospheric temperature (T₂) based on U.Md. (orange), RSS (red) and UAH (blue) satellite datasets, and RATPAC (violet) and HadAT2 (green) radiosonde datasets; and surface temperature (Tₛ) from NOAA data (black).
Right: surface temperature (Tₛ) from NOAA data (black) and lower tropospheric temperature (T₂LT) from RSS (red) and UAH satellite data (blue), and from RATPAC (violet) and HadAT2 (green) radiosonde data.
Filled circles denote trends estimated to be statistically significantly different from zero (at the 5% level). A Student's t-test, using the lag-1 autocorrelation to account for the non-independence of residual values about the trend line, was used to assess significance (see Appendix for discussion of confidence intervals and significance testing).

855   In the stratosphere (left panel of Figure 3.5), trend profiles for the two satellite datasets

856   are fairly similar, with a greater temperature decrease everywhere according to T_{4-A} than

857   T_{4-R}. Some of the largest temperature decrease occurs in the South Polar Region, where

858   ozone depletion is largest. A broad region of weaker decrease occurs in the deep tropics.

859   By contrast, the RATPAC and HadAT2 radiosonde datasets are quite different from the

860   satellite products, with much flatter profiles. It is worth noting that there is a fundamental

861   disagreement between the radiosonde and satellite products. Except for the mid-latitudes

862   of the Northern Hemisphere[17], at most other latitudes the radiosonde products show more

863   of a temperature decrease than the satellite products, with the largest discrepancy in the

864   tropics[18].

865

866   For the middle troposphere (middle panel of Figure 3.5) there is general agreement

867   among the radiosonde and satellite datasets in depicting the same basic structure. The

868   largest temperature increase occurs in the extratropics of the Northern Hemisphere, with

869   a smaller increase or slight decrease in the tropics, and even lesser increase or more

870   decrease in the extratropics of the Southern Hemisphere. At most latitudes, T_{2-M} indicates

871   the most increase (least decrease), followed next by T_{2-R}, then T_{2-A}, and finally the

872   radiosonde products with the least increase (most decrease).

---

[17] The apparently better radiosonde-satellite agreement in the midlatitudes of the Northern Hemisphere may be the result of spurious stratospheric warming at stations located in countries of the former Soviet Union, offsetting the more typical spurious cooling bias of radiosonde temperatures (Lanzante et al., 2003a,b).
[18] We note that in the tropics, where the radiosonde and satellite products differ the most, abrupt artificial drops in temperature appear to be particularly problematic for radiosonde data (Parker et al., 1997; Lanzante et al., 2003a,b). Other studies (Sherwood et al., 2005; Randel and Wu, 2005) also suggest spurious cooling for radiosonde temperatures, especially in the tropics.

873 For the lower troposphere and surface (right panel of Figure 3.5) the profiles are roughly

874 similar in shape to those for the middle troposphere with one major exception: the higher-

875 latitude temperature increase of the Northern Hemisphere is more pronounced compared

876 to the other regions. Comparing the surface temperature trend profile (black) with that

877 from the various tropospheric products in the middle and right panels of Figure 3.5

878 suggests that the sign and magnitude of this difference is highly dependent upon which

879 tropospheric measure is used.

880

881 **6.2.3 Maps**

882 Trend maps represent the finest spatial granularity with which different levels/layers and

883 observing platforms can be compared. However, since maps may not be the optimal way

884 in which to examine trends[19], we present only a limited number of such maps for

885 illustrative purposes. Figure 3.6 presents maps of trends for the surface (bottom), lower

886 troposphere (second from bottom), upper middle troposphere (second from top), and

887 stratosphere (top). The surface map is based on the NOAA dataset[20] while those for the

888 troposphere and stratosphere are based on the RSS satellite dataset[21]. In examining these

---

[19] Averaging over space (e.g., over latitudes, the tropics or the globe, as presented earlier) tends to reduce noise that results from the statistical uncertainties inherent to any observational measurement system. Furthermore, models that are used to study climate change have limited ability to resolve the smallest spatial scales and therefore there is little expectation of detection at the smallest scales (Stott and Tett, 1998). The formal methodology that is used to compare models with observations ("fingerprinting," see Chapter 5) concentrates on the larger-scale signals in both models and observations in order to optimize the comparisons.

[20] Trend maps from other surface datasets (not shown) tend to be fairly similar to that of the NOAA map, differing mostly in their degree of spatial smoothness, which is a function of dataset construction methodology.

[21] A comparison between UAH and RSS trend maps for tropospheric layers is given in Chapter 4.

889   maps it should be kept in mind that based on theory *we expect* the difference in trend

890   between the surface and troposphere to vary by location. For example, as shown in

891   Chapter 1, climate model projections typically indicate that human induced changes

892   should lead to more warming of the troposphere than the surface in the tropics, but the

893   opposite in the Arctic and Antarctic.

894
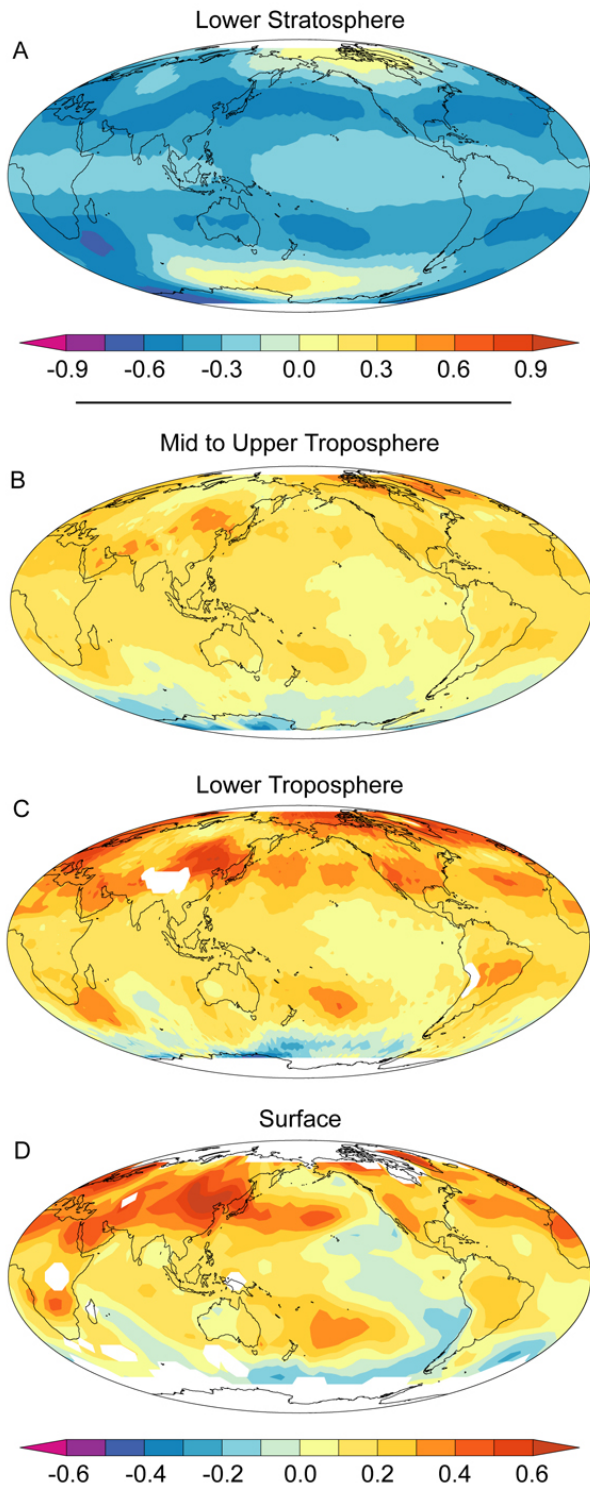
Figure 3.6 – Temperature trends for 1979-2004 (^{o}C /decade).
Bottom (d): NOAA surface temperature (T_{S-N}).
Third (c): RSS lower tropospheric temperature (T_{2LT-R}).
Second (b): RSS upper middle tropospheric temperature (T_{2-R}).

899   Top (a): RSS lower stratospheric temperature (T_{4-R}).

900

901   The trend maps indicate both similarities and differences between the surface and

902   tropospheric trend patterns. There is a rough correspondence in patterns between the two.

903   The largest temperature increase occurs in the extratropics of the Northern Hemisphere,

904   particularly over landmasses. A decreases or smaller increase is found in the high

905   latitudes of the Southern Hemisphere as well as in the eastern tropical Pacific. Note the

906   general correspondence between the above noted features in Figures 3.6c,d and the zonal

907   trend profiles (middle and right panels of Figure 3.5). Note that the upper middle

908   tropospheric temperature is somewhat of a hybrid measure, being affected most strongly

909   by the troposphere, but with a non-negligible influence by the stratosphere.

910

911   In contrast to the surface and troposphere, a temperature decrease is found almost

912   everywhere in the stratosphere (Figure 3.6a). The largest decrease is found in the

913   midlatitudes of the Northern Hemisphere and the South Polar Region, with a smaller

914   decrease in the tropics. Again note the correspondence between the main features of the

915   trend map (Figure 3.6a) and the corresponding zonal trend profiles (left panel of Figure

916   3.5).

917
918   **7. CHANGES IN VERTICAL STRUCTURE**

919

920   **7.1 Vertical profiles of trends**

921   Up to this point, our vertical comparisons have contrasted trends of surface temperature

922   with trends based on different layer-averaged temperatures. Layers are useful because the

923   averaging process tends to reduce noise. The use of layer-averages is also driven by the

924   limitations of satellite measurement systems that are unable to provide much vertical

925   detail. However, as illustrated in Chapter 1, changes in various forcing agents can lead to

926   more complex changes in the vertical. Radiosonde data, because of their greater vertical

927   resolution, are much better suited for this than currently available satellite data.

928

929   Figure 3.7 shows vertical profiles of trends from the RATPAC and HadAT2 radiosonde

930   datasets for temperature averaged over the globe (top) or tropics (bottom) for the

931   radiosonde (left) and satellite (right) eras. The trend values of Figure 3.7 are also given in

932   Table 3.5. Each graph has profiles for the two radiosonde datasets. The tropics are of

933   special interest because many climate models suggest that under global warming

934   scenarios trends should increase from the lower troposphere upwards, maximizing in the

935   upper troposphere (see Chapters 1 and 5).

936   Table 3.5 – Temperature trends in ºC per decade from the RATPAC and HadAT2 radiosonde datasets
937   corresponding to the plots in Figure 3.7 (see figure caption for further details). Global and tropical trends
938   are given for 1958 through 2004 and 1979 through 2004 (except for European which terminates September
939   2001). The HadAT2 dataset does not have temperatures for some of the levels, hence the empty table cells.
940   The trend is shown for each vertical level (hPa), with the approximate 95% confidence interval (2 sigma)
941   below in parentheses. Bold values are estimated to be statistically significantly different from zero (at the
942   5% level). A Student's t-test, using the lag-1 autocorrelation to account for the non-independence of
943   residual values about the trend line, was used to assess significance (see Appendix for discussion of
944   confidence intervals and significance testing).

945

Level (hPa)	1958-RATPAC Global	2004 HadAT2 Global	1958-RATPAC Tropical	2004 HadAT2 Tropical	1979-RATPAC Global	2004 HadAT2 Global	1979-RATPAC Tropical	2004 HadAT2 Tropical
20	**-0.41** (0.078)		**-0.49** (0.143)		**-0.91** (0.141)		**-0.95** (0.319)	
30	**-0.48** (0.091)	**-0.57** (0.100)	**-0.55** (0.179)	**-0.59** (0.204)	**-0.88** (0.234)	**-0.96** (0.249)	**-0.91** (0.522)	**-0.90** (0.586)
50	**-0.53** (0.120)	**-0.55** (0.119)	**-0.63** (0.224)	**-0.52** (0.232)	**-0.89** (0.330)	**-0.88** (0.346)	**-1.01** (0.568)	**-0.83** (0.591)
70	**-0.48** (0.110)		**-0.58** (0.222)		**-0.79** (0.261)		**-0.89** (0.451)	
100	**-0.23** (0.063)	**-0.25** (0.060)	**-0.18** (0.063)	**-0.27** (0.066)	**-0.43** (0.164)	**-0.43** (0.152)	**-0.36** (0.173)	**-0.51** (0.159)
150	-0.05 (0.061)	-0.04 (0.057)	0.05 (0.065)	-0.01 (0.064)	**-0.19** (0159)	-0.13 (0.140)	-0.10 (0.185)	-0.14 (0.158)
200	0.03 (0.047)	**0.05** (0.047)	**0.13** (0.079)	**0.11** (0.089)	-0.08 (0.113)	-0.05 (0.105)	-0.01 (0.204)	-0.02 (0.224)
250	**0.11** (0.037)		**0.15** (0.076)		0.08 (0.096)		0.09 (0.198)	
300	**0.14** (0.038)	**0.14** (0.044)	**0.18** (0.071)	**0.15** (0.084)	**0.12** (0.094)	**0.12** (0.094)	0.13 (0.181)	0.05 (0.208)
400	**0.15** (0.036)		**0.15** (0.063)		**0.13** (0.082)		0.11 (0.147)	
500	**0.14** (0.032)	**0.14** (0.040)	**0.14** (0.057)	**0.11** (0.063)	**0.09** (0.068)	**0.12** (0.074)	0.05 (0.124)	0.01 (0.135)
700	**0.13** (0.026)	**0.15** (0.035)	**0.13** (0.054)	**0.11** (0.064)	**0.09** (0.053)	**0.12** (0.066)	0.05 (0.123)	0.02 (0.129)
850	**0.12** (0.022)	**0.15** (0.029)	**0.08** (0.032)	**0.12** (0.051)	**0.08** (0.047)	**0.13** (0.060)	-0.01 (0.058)	0.06 (0.105)
Surface	**0.11** (0.022)	**0.12** (0.026)	**0.10** (0.031)	**0.11** (0.039)	**0.17** (0.050)	**0.18** (0.050)	**0.13** (0.068)	**0.15** (0.115)

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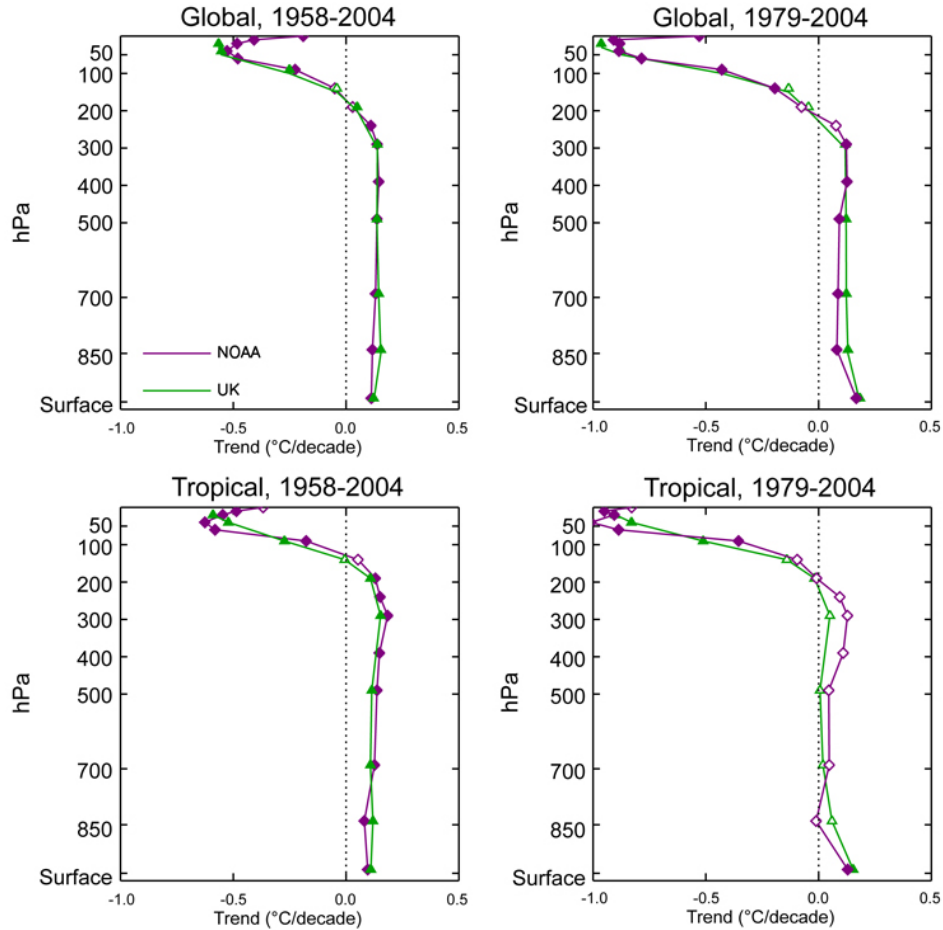
947
948   Figure 3.7 -- Vertical profiles of temperature trend (ºC/decade) as a function of altitude (i.e., pressure in
949   hPa) computed from the RATPAC (violet) and HadAT2 (green) radiosonde datasets. Trends (which are
950   given in Table 3.5) have been computed for 1958-2004 (left) and 1979-2004 (right) based on temperature
951   that has been averaged over the globe (top) or the tropics, 20ºN-20ºS (bottom). Surface data for the
952   HadAT2 product is taken from HadCRUT2v since the HadAT2 dataset does not include values at the
953   surface; the surface values have been averaged so as to match their observing locations with those for the
954   radiosonde data. By contrast, the surface temperatures from the RATPAC product are those from the
955   RATPAC dataset, which are surface station values reported with the radiosonde data. Note that these differ
956   from the NOAA surface dataset values (ER-GHCN-ICOADS) as indicated in Table 3.1. Filled symbols
957   denote trends estimated to be statistically significantly different from zero (at the 5% level). A Student's t-
958   test, using the lag-1 autocorrelation to account for the non-independence of residual values about the trend
959   line, was used to assess significance (see Appendix for discussion of confidence intervals and significance
960   testing).

961

962

963   For the globe, the figure indicates that during the longer period the tropospheric

964   temperature increased slightly more than that of the surface. By contrast, for the globe

965   during the satellite era, the surface temperature increased more than that of the

966   troposphere. Both datasets agree reasonably well in these conclusions. For the tropics, the

967   differences between the two eras are more pronounced. For the longer period there is

968   good agreement between the two datasets in that the temperature increase is smaller at the

969   surface and maximized in the upper troposphere. The largest disagreement between

970   datasets and least amount of tropospheric temperature increase is seen in the tropics

971   during the satellite era. For the RATPAC product, the greatest temperature increase

972   occurs at the surface with a slight increase (or decrease) in the lower and middle

973   troposphere followed by somewhat larger increase in the upper troposphere. The HadAT2

974   product also shows largest increase at the surface, with a small increase in the

975   troposphere, however, it lacks a distinct return to increase in the upper troposphere. In

976   summary, the two datasets have fairly similar profiles in the troposphere with the

977   exception of the tropics during the satellite era[22]. For the stratosphere, the decrease in

978   temperature is noticeably greater for both the globe and the tropics during the satellite

979   than radiosonde era as expected (see Figure 3.2b). Some of the largest discrepancies

980   between datasets are found in the stratosphere.

981

982

983   **7.2 Lapse rates**

984   Temperature usually decreases in the troposphere going upward from the surface. Lapse

---

[22] However, the differences between datasets may not be meaningful since they are small compared to the statistical uncertainty estimates (see Table 3.5 and discussion in the Appendix).

985   rate is defined as the rate of decrease in temperature with increasing altitude and is a

986   measure of the stability of the atmosphere[23]. Most of the observational work to date has

987   not examined lapse rates themselves, but instead has used an approximation in the form

988   of a vertical temperature difference[24]. This difference has taken on the form of the surface

989   temperature minus some tropospheric temperature, either layer-averaged (in the case of

990   satellite data) or at some specific pressure level (in the case of radiosonde data)[25].

991

992   Much of the interest in lapse rate variations has focused on the tropics. Several studies

993   (Brown et al., 2000; Gaffen et al., 2000; Hegerl and Wallace, 2002; Lanzante et al.,

994   2003b) present time series related to tropical lapse rate based on either satellite or

995   radiosonde measures of tropospheric temperature. As examples, we present some such

996   time series in Figure 3.8, based on measures of lower tropospheric temperature from

997   three different datasets. Some essential low-frequency characteristics are common to all.

998   A considerable proportion of the variability of the tropical lapse rate is associated with

999   ENSO[26], a manifestation of which are the up and down swings of about 3-7 years in the

---

[23] A larger lapse rate implies more unstable conditions and a greater tendency towards vertical mixing of air.

[24] The reasons for this are two-fold: (1) satellite measurement systems are only able to resolve temperatures in deep layers rather than at specific levels, and (2) radiosonde measurements are consistently recorded at a fixed number of constant pressure rather than height levels.

[25] When constant pressure level data from radiosondes are used, the resulting lapse rate quantity may be influenced by changes in the thickness (i.e., average temperature) of the layer. However, some calculations by Gaffen et al. (2000) suggest that thickness changes do not have very much influence. Therefore, we consider vertical temperature differences to be a suitable approximation of lapse rate

[26] Lapse rate changes occur about five to six months after a particular change in ENSO (Hegerl and Wallace, 2002; Lanzante et al., 2003b). During a tropical warming event (El Niño) the tropical troposphere warms relative to the surface; the opposite is true during a tropical cooling event (La Niña).

1000   series shown in Figure 3.8. Another feature evident in the four studies cited above, and

1001   seen in Figure 3.8 as well, is an apparent strong association with the climate regime shift

1002   that occurred ~1976-77 (Trenberth and Hurrell, 1994). There is a rather sharp drop in

1003   tropical lapse rate at this time[27], coincident with an abrupt change in a measure of

1004   convective stability (Gettelman et al. 2002). Overall, the variation in tropical lapse rate

1005   can be characterized as highly complex, with rapid swings over a few years,

1006   superimposed upon persistent periods of a decade or more, as well as longer-term drifts

1007   or trends evident during some time periods.
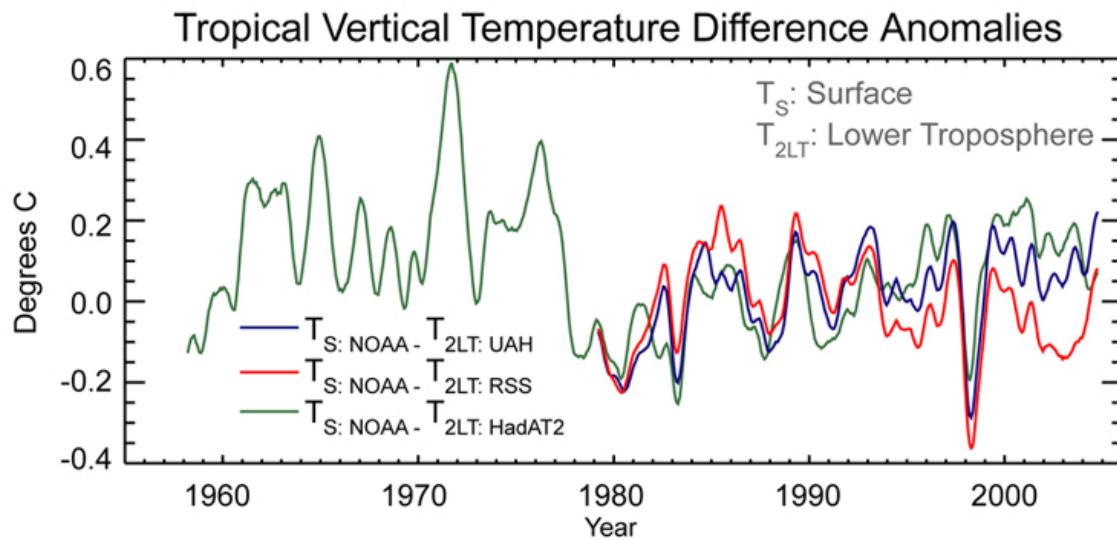
1008
1009
1010

1011   Figure 3.8 - Time series of vertical temperature difference (surface minus lower troposphere) for the tropics
1012   (20°N-20°S). NOAA surface temperatures (T_{S-N}) are used in each case to compute differences with lower
1013   tropospheric temperature (T_{2LT}) from three different groups: HadAT2 radiosonde (green), RSS satellite
1014   (red), and UAH satellite (blue). All time series are 7-month running averages (used as a smoother) of
1015   original monthly data, which were expressed as a departure (°C) from the 1979-97 average.

1016

[27] Lanzante et al. (2003b) also noted an apparent decrease in the amplitude of ENSO-related tropical lapse
rate variations after the ~1976-77 regime shift.

1017   The feature of the tropical lapse rate series that has drawn the most interest is the linear

1018   trend component during the satellite era. From a long historical perspective (see also

1019   Figure 3.8), this trend is a rather subtle feature, being overshadowed by both the ENSO-

1020   related variations as well as the regime shift of the late 1970s. Several studies (Brown et

1021   al., 2000; Gaffen et al., 2000; Hegerl and Wallace, 2002; Lanzante et al., 2003b) have

1022   estimated trends in lower tropospheric lapse rate while another (Christy et al., 2001) has

1023   estimated trends in the difference between SST and surface air temperature.

1024

1025   The different trend estimates vary considerably among the above-cited studies, being

1026   dependent upon the details of the calculations[28]. From the cited studies, satellite-era

1027   trends in lapse rate based on temperatures averaged over the tropics range from nearly

1028   zero (no change) to about 0.20ºC/decade (surface warms more than the troposphere). The

1029   time series of Figure 3.8 also exhibit a wide range of satellite-era trends[29]. During the

1030   longer radiosonde era, the various studies found trends of opposite sign (i.e., air

1031   temperature at the surface increases more slowly than that of air aloft) and show less

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[28]These details include: time period, latitude zone, datasets utilized, station network vs. grid, time of day of observations, use of homogeneity adjustment, and whether or not measurements in the troposphere and surface were taken from the same locations. Particularly noteworthy is the fact that Lanzante et al. (2003b) found that during the satellite era, use of homogenized data could, depending the other details of the analysis, either halve or eliminate the positive tropical lapse rate trend found using the unadjusted data.

[29] Trends from 1979 to 2004 (ºC /decade) for the three time series in Figure 3.8 are: 0.11 (HadAT2 radiosonde), 0.08 (UAH satellite), and -0.02 (RSS satellite). While the first two of these trends are statistically significant at the 5% level, the third is not (see Appendix for discussion of significance testing).

1032   sensitivity, with a range of values of near-zero to about -0.05ºC/decade[30].

1033

1034   Spatial variations in lapse rate trends have also been examined. During the satellite era,

1035   some have found predominantly increasing trends in the tropics (Gaffen et al., 2000;

1036   Brown et al., 2000) while others have found a greater mixture, with more areas of

1037   negative trends (Hegerl and Wallace, 2002; Lanzante et al., 2003b). Outside of the

1038   tropics, both Hegerl and Wallace (2002) and Lanzante et al. (2003b) found complex

1039   spatial patterns of trend. Lanzante et al. (2003b) also found considerable local sensitivity

1040   to homogeneity adjustment in the tropics and even more so over the extratropics of the

1041   Southern Hemisphere, which is quite sparsely sampled.

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[30] The trend from 1958 to 2004 for the HadAT2 radiosonde series shown in Figure 3.8 is -0.02 ºC/decade. This trend is not statistically significant at the 5% level (see Appendix for discussion of significance testing).

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6   **Chapter 4**
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9   What is our understanding of the contribution made by observational or
10   methodological uncertainties to the previously reported vertical differences in
11   temperature trends?
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16   Convening Lead Author: Carl Mears
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18   Lead Authors: Chris Forest, Roy Spencer, Russell Vose, and Dick Reynolds
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20   Contributing Authors
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22   Peter Thorne and John Christy
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46   **Chapter 4: Key Findings**

47

48   *Surface*

49

50   **It is likely that errors in the homogenized surface air temperature data do not**

51   **contribute substantially to the large-scale differences between trends for**

52   **different levels because these errors are very likely to be smaller than those for**

53   **the upper air data.**

54   - Systematic local biases in surface trends may exist due to changes in station

55   exposure or instrumentation over land, and due to the small number of

56   measurements over a number of regions of the earth, including parts of the

57   oceans, sea ice areas, and some land areas. Such biases have been

58   documented at the local and regional scale, but no such effect has been

59   identified in the zonal and global averages presented in this report. On large

60   spatial scales, sampling studies suggest that these local biases in trends are

61   likely to mostly cancel through the use of many observations with differing

62   instrumentation.

63   - Since all known bias adjustments have not yet been applied to sea surface

64   temperature data, it is likely that errors remain in these data, though it is

65   generally agreed that these errors are likely to be small compared to errors in

66   radiosonde and satellite measurements of the upper air, especially for the

67   satellite era.

68

69   *Troposphere*

70   **While all datasets indicate that the troposphere has warmed over both the**

71   **radiosonde era and the satellite era, uncertainties in the tropospheric data**

72   **make it difficult to determine whether the troposphere has warmed more than**

73   **or less than the surface. Some tropospheric datasets indicate that the**

74   **troposphere has warmed more than the surface, while others indicate the**

75   **opposite.**

76   • It is very likely that errors remain in the homogenized radiosonde datasets in

77   the troposphere since the methods used to produce them are only able to

78   detect and remove the more obvious errors, and involve many subjective

79   decisions. It is likely that a net spurious cooling corrupts the area-averaged

80   homogenized radiosonde data in the tropical troposphere in at least one and

81   probably both of the datasets, causing the data to indicate less warming than

82   has actually occurred.

83   • For tropospheric satellite data (T_2 and T_{2LT}), the primary cause of trend

84   discrepancies between different versions of the datasets is differences in

85   how the data from the different satellites are merged together.

86   • A secondary contribution to the differences between these datasets is the

87   difference between the diurnal adjustments that are used to account for

88   drifting measurement times. These differences in the diurnal adjustment are

89   more important for regional trends than for global trends, though regional

90   trend differences are also partly influenced by differences in merging

91   methods.

92   • Each tropospheric satellite dataset has strengths and weaknesses that are

93   coming into better focus. Improvements have occurred in several datasets

94   even during the drafting of this report, each moving it closer to the others,

95   suggesting that further convergence in the not-too-distant future is a strong

96   possibility.

97   • Comparisons between radiosonde data and satellite data for T_2 are very

98   likely to be corrupted by the excessive cooling in the radiosonde data from

99   the stratosphere which are used to help construct the radiosonde-derived T_2

100   data. Trend discrepancies between radiosonde and satellite datasets are

101   reduced by considering a multi-channel retrieval that estimates and removes

102   the stratospheric influence (T^*_G).

103

104   ***Stratosphere***

105   **Despite their large discrepancies, all datasets indicate that the stratosphere has**

106   **cooled considerably over both the radiosonde era and the satellite era.**

107   • The largest discrepancies between datasets are in the stratosphere,

108   particularly between the radiosonde and satellite-based datasets. It is very

109   likely that the satellite-sonde discrepancy arises primarily from uncorrected

110   errors in the radiosonde data.

111   • There are also substantial discrepancies between the satellite datasets in the

112   stratosphere, indicating that there remain unresolved issues with these

113   datasets as well.

114

115   **<u>Chapter 4 recommendations</u>**

116

117   *All of the surface and atmospheric temperature datasets used in this report require*

118   *ongoing assessment to further quantify uncertainty and to identify and remove any*

119   *possible systematic biases that remain after the appropriate homogenization methods*

120   *have been applied.*

121

122   • *The diurnal cycles in both atmospheric and surface temperature need to be*

123   *accurately determined and validated to reduce uncertainties in the satellite data*

124   *due to the diurnal adjustment. Possible approaches include examining more*

125   *model or reanalysis data to check the diurnal adjustments currently in use,*

126   *concerted in situ measurement campaigns at a number of representative*

127   *locations, or operating a satellite-borne sounder in a non sun-synchronous orbit.*

128   *Information about the surface skin temperature diurnal cycle may be obtained by*

129   *studying data from existing satellites, or the upcoming Global Precipitation*

130   *Mission.*

131   • *The relative merits of different merging methods for satellite data for all relevant*

132   *layers need to be diagnosed in detail. Possible approaches include comparison*

133   *with other temperature data sources (radiosondes or IR satellites) over limited*

134   *time periods where the discrepancies between the satellite results are the greatest,*

135   *comparison with other ancillary data sources such as winds and integrated water*

136   *vapor, and comparison of trends on regional spatial scales, particularly in*

137   *regions where trends are large or well characterized by radiosonde data.*

138   •   *The methods used to remove radiosonde inhomogeneities and their effects on*

139   *trends need to be rigorously studied. The detailed intercomparisons of the*

140   *methods used by different groups to construct satellite based climate records has*

141   *been beneficial to our understanding of these products, and similar parallel*

142   *efforts to create climate records from radiosonde data would be likely to provide*

143   *similar benefits.*

144   •   *Possible errors in trends in spatially averaged surface temperature need to be*

145   *assessed further. On land these errors may arise from local errors due to*

146   *changes in instrumentation or local environment that do not completely cancel*

147   *when spatial averaging is performed. Over the ocean, these errors may arise*

148   *from the small number of samples available in many regions, and long-term*

149   *changes in measurement methods. For historical data, these assessments may*

150   *benefit from the recovery of additional metadata to better characterize possible*

151   *non-climatic signals.*

152   •   *Tools and methods need to be developed to help reduce structural uncertainty by*

153   *providing methods to objectively differentiate between different datasets and*

154   *construction methods. To the extent possible, such tools should be based on*

155   *generally accepted physical principles, such as consistency of the temperature*

156   *changes at adjacent levels in the atmosphere, include physically-based*

157   *comparisons with external ancillary data, and take account of the consistency of*

158   *intermediate data generated while producing the datasets.*

159

160   **1. Background**

161

162   In the previous chapter, we have discussed a number of estimates of vertically resolved

163   global temperature trends. Different sources of data (e.g., surface measurements, vertical

164   profiles from radiosondes, and data from satellite borne sounding radiometers), as well as

165   different analysis methods applied to the same data, can yield long term (multi-decadal)

166   temperature trends that differ by as much as several tenths of a ℃ per decade. This is of

167   comparable magnitude to the actual climate change signal being searched for. In this

168   chapter we discuss these discrepancies in light of the observing system capabilities and

169   limitations described in Chapter 2. We note the degree to which estimates of uncertainty

170   can account for the differences in reported values for the temperature trends in given

171   layers, and differences in the trends of adjacent layers. Most of the time our focus will be

172   on the period from 1979-2004, during which atmospheric temperatures were observed

173   using multiple observing systems.

174

175   We begin our discussion in the stratosphere, and move to successively lower layers until

176   we reach the Earth's surface. We proceed in this order because the largest discrepancies

177   in trends between data sources occur in upper atmospheric layers, especially the

178   stratosphere. As mentioned in Box 2.2, when satellite-equivalent measures are made from

179   vertically resolved radiosonde data to facilitate comparisons between the two systems,

180   large stratospheric errors can significantly influence measures centered much lower in the

181   atmosphere.

182

183   **2. Uncertainty in stratospheric temperature trends**

184

185   Long-term observations of the stratosphere have been made by two observing systems:

186   radiosondes and satellite-borne sounders. On both the global and the zonally averaged

187   scale, there is considerably less variation between datasets derived from the same type of

188   observing system for this layer than between those from different observing systems.

189   This can be seen in the leftmost panel of Figure 3.5, which shows the zonally averaged

190   trends over the satellite era (1979-2004) for two radiosonde-based datasets, and two

191   satellite-based datasets. The radiosonde data ($T_{4-HadAT2}$ and T_{4-NOAA}) show more cooling

192   than datasets based on satellite data (T_{4-UAH} and T_{4-RSS}), and also do not show the reduced

193   cooling in the tropics relative to the mid-latitudes that is seen in the satellite data.

194

195   *2.1 Radiosonde Uncertainty*

196   As discussed in Chapter 2, radiosonde data are plagued by numerous spurious

197   discontinuities in measured temperature that must be detected and removed in order to

198   construct a homogenized long-term record of atmospheric temperature, a task that is

199   particularly difficult in the absence of reliable metadata describing changes in

200   instrumentation or observing practice. A number of physical sources of such

201   discontinuities have larger effects in the stratosphere. The lower atmospheric pressure in

202   the stratosphere leads to reduced thermal contact between the air and the temperature

203   sensor in the radiosonde package. This in turn leads to increased errors due to daytime

204   solar heating and lags between the real atmospheric temperature and the sensor response

205   as the instrument rises through atmospheric layers with rapidly varying temperatures.

206   Such systematic errors are not important for trend studies provided that they do not

207   change over the time period being studied. In practice, as noted in Chapter 2, radiosonde

208   design, observing practices, and procedures used to attempt to correct for radiation and

209   lag errors have all changed over time.

210

211   Past attempts to make adjustments to radiosonde data using detailed physical models of

212   the instruments (Luers and Eskridge, 1998) improved data homogeneity in the

213   stratosphere, but not in the troposphere (Durre et al., 2002). Since it is important to use

214   the same methods for all radiosonde levels for consistency, scientists have tended to

215   instead use empirical methods to deduce the presence and magnitude of any suspected

216   discontinuity. Both of the homogenized radiosonde datasets used in this report make

217   these estimates using retrospective statistical analyses of the radiosonde data without

218   input from other measurements. The investigators who constructed these datasets have

219   attempted to identify and to adjust for the effects of suspected change points, either by

220   examination of station time series in isolation (NOAA), or by comparison with nearby

221   stations (UK). Both approaches can most successfully identify changes that are large and

222   step-like. While based in statistics, both these methods also include significant subjective

223   components. As a result, different investigators with nominally the same sets of

224   radiosonde data can calculate different trend estimates because of differences in

225   adjustment procedures (Free et al., 2002). The lack of sensitivity to small or gradual

226   changes may bias the resulting homogenized products if such changes are numerous and

227   predominantly of one sign or the other[1]. The relative frequency of large step-like changes

228   and smaller changes that may be statistically indistinguishable from natural variability

229   remains an open question.

230

231   Since the adjustments needed to remove the resulting discontinuities tend to be larger for

232   the stratosphere than for lower levels (Parker et al., 1997; Christy et al., 2003; Lanzante

233   et al., 2003), the uncertainty associated with the homogenization procedures is very likely

234   to be larger in the stratosphere than at lower levels, as has been shown for the UK

235   radiosonde dataset (Thorne et al., 2005). The best estimate of the size of this source of

236   uncertainty is obtained by comparing the statistics (e.g., the trends) from the two adjusted

237   radiosonde datasets that are currently available. However, the UK group analysis is partly

238   based upon the NOAA dataset, so we may be under-estimating the uncertainty. Only

239   through increasing the number of independently produced datasets under different

240   working assumptions can we truly constrain the uncertainty (Thorne et al., 2005).

241

242   Differences in trends between daytime and nighttime observations in the uncorrected

243   radiosonde data used in constructing the NOAA and UK radiosonde datasets, suggest that

244   the biases caused by solar heating[2] have been reduced over time, leading to a spurious

245   cooling trend in the raw daytime data (Sherwood et al., 2005). Many of the changes in

---

[1] It is speculated that gradual changes could result from the same changes in instrumentation or practices that cause the step like changes, provided that these changes are implemented gradually (Lanzante et al., 2003).

[2] For some types of radiosondes, radiation adjustments based on information provided by the manufacturer are made as part of routine processing of radiosonde data by the observing station. The findings cited here refer to data that has already had these corrections performed. The reduction in daytime biases is likely to be due to a combination of improvements in instrument design, and improvements in the radiation adjustment procedure.

246   observing practice will affect both day and night time observations; e.g., a change in

247   practice may yield a spurious 0.5ºC daytime cooling and 0.4ºC night time cooling, so

248   day-night differences cannot be used in isolation to correct the observations. Whether the

249   NOAA and UK methods have successfully removed day-night and other effects, or if

250   sufficiently targeted are capable of doing so, is a matter for ongoing research. Randel and

251   Wu (2005) have shown for a subset of tropical stations in the NOAA dataset, there is

252   strong evidence for step-like residual cooling biases following homogenization, which

253   will cause a spurious cooling in the tropical area-averaged NOAA time series considered

254   here. They find that the effect is not limited to daytime launches, as would be expected

255   from discussions above, and that it is likely to affect at least the upper-troposphere as

256   well as the stratosphere. Finally, the balloons that carry the instruments aloft have

257   improved over time, so they are less likely to burst at high altitudes or in extreme cold.

258   This could also lead to a warm sampling bias within the stratosphere in early radiosondes

259   which has gradually ameliorated with time, introducing a spurious stratospheric cooling

260   signal (Parker et al., 1997). Taken together these results imply that any residual

261   systematic errors in the homogenized radiosonde products will likely lead to a spurious

262   cooling bias.

263

264   Since the radiosonde stations selected for inclusion in the homogenized datasets do not

265   cover the entire globe, there can be a bias introduced in to the global mean trend

266   depending on the locations of the chosen stations. On a global scale, this bias has been

267   estimated to be less than 0.02ºC/decade for T_4 by sub-sampling globally complete

268   satellite or reanalysis datasets at the station locations[3], and thus it is not an important

269   cause of the differences between the datasets on large spatial scales (Free and Seidel,

270   2004). Though they have not been explicitly calculated, sampling errors are likely to be

271   more important for the zonal radiosonde trends plotted in Figure 3.5, and may account for

272   some of the zone-to-zone variability seen in the radiosonde data in that figure that is not

273   duplicated in the smoother satellite data. The sampling effects also permeate in the

274   vertical – above 100hPa there is a significant reduction in the number of valid

275   measurements whereas below this level the number of measurements is relatively stable.

276   Because the trends vary with height, this can lead to errors, particularly when calculating

277   satellite-equivalent measures.

278

279   *2.2 Satellite Uncertainty*

280
281   The two satellite-based stratospheric datasets ($T_{4\text{-UAH}}$ and $T_{4\text{-RSS}}$) have received

282   considerably less attention than their tropospheric counterparts (see section 4.3 below),

283   though they differ in estimated trend by roughly the same absolute amount

284   (~0.1ºC/decade) as the corresponding tropospheric datasets produced by the same

285   institutions. However the importance of the differences is perceived to be much less

286   because the trend is much larger (a cooling over 1979-2004 of approximately 0.8ºC). A

287   detailed comparison of the methods used to construct the two datasets has not yet been

288   performed. Despite the lack of such a study, it is very likely that in the stratosphere, like

289   the troposphere (discussed in section 4.3), structural uncertainty is the most important

---

[3] This estimate is valid for the NOAA dataset and a previous version of the UK dataset. The estimated bias increases to about 0.05K for a tropical average. In the cited work the tropics were defined to be 30S to 30N – we would expect the sampled error to be a few hundredths of a degree per decade larger for the 20S to 20N definition of the tropics used in this report.

290   source of uncertainty. Two important types of structural uncertainty are likely to

291   dominate: those associated with the method of correcting for drifts in diurnal sampling

292   time, and those associated with the method of correcting calibration drifts associated with

293   the temperature of the hot calibration target. Section 3 discusses how these uncertainty

294   sources are treated in the troposphere.

295

296   Despite unresolved problems in the satellite datasets, the similarity of the satellite

297   measurement and homogenization methods suggest that the satellite measurements of the

298   stratosphere are no more uncertain than those of the mid-troposphere, where satellites and

299   radiosondes are in much closer agreement. This assessment, coupled with the evidence

300   presented above that residual artificial cooling is likely to exist in the stratospheric

301   radiosonde data, particularly in the tropics, implies that the discrepancy between

302   radiosondes and satellite estimates of stratospheric trends (see Table 3.3) during the

303   satellite era is very likely to be mostly due to uncorrected biases in the radiosonde

304   measurements.

305

306   **3. Uncertainty in tropospheric trends**

307   In contrast to the stratosphere, differences in reported tropospheric trends from the same

308   type of measurement are as large as or larger than differences in trends reported from

309   different data sources. This can be seen in Figure 3.5 and Tables 3.3 and 3.4. Also note

310   that the radiosonde data for the two tropospheric layers show the same general north-

311   south pattern (i.e. more temperature increase in the mid-latitudes than at the poles or in

312   the tropics) as the satellite data, in contrast to the stratospheric results.

313

314   *3.1 Radiosonde uncertainty*

315   The main sources of error in tropospheric radiosonde trends are similar to those

316   encountered in the stratosphere. The challenge is to assess to what extent these types of

317   errors, which in the stratosphere likely result in artificial cooling even in homogenized

318   datasets, extend down into the troposphere. Another important issue is that when

319   performing calculations to directly compare radiosonde data with satellite trends for the

320   T_2 layer, the contribution of errors in the stratospheric trends to the results for this layer

321   become important, since 10% to 15% of the weight for this layer comes from the

322   stratosphere.

323

324   *3.1.1 Removing non-climatic influences.*

325   There are several pieces of evidence that suggest that any residual bias in tropospheric

326   radiosonde data will be towards a cooling. First, the more obvious step-like

327   inhomogeneities that have been found tend to predominantly introduce spurious cooling

328   into the raw time series, especially in the tropics. This suggests that any undetected

329   change points may also favor spurious cooling (Lanzante et al., 2003). Second, solar-

330   heating-induced errors, while largest in the stratosphere have been found to bias daytime

331   measurements to higher temperatures at all levels, particularly in the tropics. Periodic

332   radiosonde intercomparisons (most recently at Mauritius in Feb. 2005) undertaken under

333   the auspices of WMO imply that the magnitude of these errors has been reduced over

334   time, and that radiosondes from independent manufacturers have become increasingly

335   similar (and presumably more accurate) over time[4] (Silveira et al., 2003; Pathack et al.,

336   2005). If these effects have on average been uncorrected by the statistical procedures

337   used to construct the homogenized radiosonde datasets discussed in this report, they

338   would introduce an artificial cooling signal into the radiosonde records. Of course on an

339   individual station basis the picture is likely to be much more ambiguous and many

340   stations records, even following homogenization efforts, are likely to retain large residual

341   warm or cold biases. But on average, the evidence outlined above suggests that if there is

342   a preferred sign it is likely to be towards a residual cooling. It is important to stress that to

343   date the quantitative evidence to support such an argument, at least away from a small

344   number of tropical stations (Randel and Wu, 2005), is at best ambiguous.

345

346   *3.1.2 Sampling uncertainty*

347   The fact that most radiosonde data are primarily collected over Northern Hemispheric

348   land areas naturally leads to uncertainties about whether or not averages constructed from

349   radiosonde data can faithfully represent global trends. However, (Wallis, 1998) and

350   (Thorne et al., 2005) show that stations can be representative of much larger scale

351   averages above the boundary layer, particularly within the deep tropics. Spatial and

352   temporal sampling errors for the radiosonde datasets have been assessed by sub-sampling

353   trends in reanalyses or satellite data at the locations of radiosonde stations used in the

354   production of global datasets, and comparing the results to the full global average of the

355   reanalysis or satellite data (Free and Seidel, 2004). Typically, errors of a few hundredths

356   of a ℃ per decade have been estimated for global averages, too small to fully account for

---

[4] These intercomparisons provide a source data about the differences between different type of sondes that have not yet been used to homogenize sonde data.

357   the differences between radiosonde and satellite trends, though it has been suggested that

358   the existing sampling could lead to a warm bias in the radiosonde record (Agudelo and

359   Curry, 2004). As is the case for the stratosphere, sampling errors may be part of the cause

360   for the zone to zone variability seen in the radiosonde data. Residual differences between

361   two radiosonde dataset global means are assessed to be approximately equally caused by

362   sampling error, choice of raw data, and choice of adjustments made[5].

363

364   *3.1.3 The influence of uncertainty in stratospheric measurements*

365   To compare data that represent identical layers in the atmosphere, "satellite-equivalent"

366   radiosonde data products are constructed using a weighted average of radiosonde

367   temperatures at a range of levels (see Box 2.2). The T_2 radiosonde datasets are

368   constructed to match the weighting function for Microwave Sounding Unit (MSU)

369   channel 2. Since 10% to 15% of the weight for this channel comes from the stratosphere

370   (see Figure 2.1), it is important to keep in mind the suspected relatively large errors in the

371   stratospheric measurements made by radiosondes. It is possible that stratospheric errors

372   could cause the trends in the radiosonde-derived T_2 to be as much as 0.05ºC/decade too

373   cool, particularly in the tropics, where the suspected stratospheric errors are the largest

374   (Randel and Wu, 2005) and therefore have a large impact on area-weighted averages.

375   This error source may be partly eliminated by considering the multi-channel tropospheric

376   retrievals discussed in section 5 below.

377

---

[5] This comparison was made using a previous version of the UK dataset (HadRT), which uses a different set of stations than the current version. This difference is very unlikely to substantially alter these conclusions.

378   *3.2. Satellite uncertainty*

379   Satellite-derived temperature trends in the middle and upper troposphere have received

380   considerable attention. In particular, the causes of the differences between T_{2-UAH} and T_{2-}

381    $_{RSS}$ have been examined in detail; less work has been done concerning $T_{2-U.Md.}$ because

382   this dataset is newer. There are two potentially important contributions to the residual

383   uncertainty in satellite estimates of global trends for the satellite-based datasets: (1)

384   corrections for drifts in diurnal sampling, and (2) different methods of merging data from

385   the series of different satellites.

386

387   *3.2.1 Diurnal Sampling Corrections*

388   During the lifetime of each satellite, the orbital parameters tend to drift slowly with time.

389   This includes both a slow change of the local equator crossing time (LECT), and a decay

390   of orbital height over time due to drag by the upper atmosphere. The LECT is the time at

391   which the satellite passes over the equator in a northward direction. Changes in LECT

392   indicate corresponding changes in local observation time for the entire orbit. Because the

393   temperature changes with the time of day (e.g., the cycle of daytime heating and

394   nighttime cooling), slow changes in observation time can cause a spurious long-term

395   trend. These diurnal sampling effects must be estimated and removed in order to produce

396   a climate-quality data record.

397

398   The three research groups that are actively analyzing data from microwave satellite

399   sounders first average together the ascending and descending orbits, which has the effect

400   of removing most of the first harmonic of the diurnal cycle. For the purposes of this

401   report, "diurnal correction" means the removal of the second and higher harmonics. Each

402   group uses a different method to perform the diurnal correction.

403

404   The UAH group calculates mean differences by subtracting the temperature

405   measurements on one side of the satellite track from the other (Christy et al., 2000). This

406   produces an estimate of how much, on average, the temperature changes due to the

407   difference in local observation times from one side of the satellite swath to another,

408   typically about 40 minutes. This method has the advantage of not relying on data from

409   other sources to determine the diurnal cycle, but it has been shown to be sensitive to

410   satellite attitude errors (Mears and Wentz, 2005), and is too noisy to produce a diurnal

411   adjustment useable on small spatial scales.

412

413   The RSS group uses hourly output from a climate model in a microwave radiative

414   transfer model to estimate the diurnal cycle in brightness temperature at each grid point in

415   the satellite dataset (Mears et al., 2003). This method has the advantage that a diurnal

416   adjustment can be made at the data resolution. However, it is likely that the climate

417   model-based adjustment contains errors, both because models are often unable to

418   accurately represent the diurnal cycle (Dai and Trenberth, 2004), and because the

419   parameterization of the ocean surface temperature used as a lower boundary for the

420   atmospheric model used does not include diurnal variability. The model has been shown

421   to represent the first harmonic of the diurnal cycle for MSU channel 2 with less than 10%

422   error, but less is known about the accuracy of the second and higher harmonics that are

423   more important for adjusting for the diurnal sampling errors (Mears et al., 2003).

424

425   Both groups use their diurnal cycle techniques to adjust the satellite data before merging

426   the data from the different satellites. In contrast, the Maryland group averaged the

427   ascending and descending satellite data to remove only the first harmonic in the diurnal

428   cycle before merging, and used a fitting procedure to account for both the first and

429   second harmonic diurnal components when performing the trend analysis *after* merging

430   the data from different satellites (Vinnikov and Grody, 2003; Vinnikov et al., 2005).

431   Since they only accounted for the first harmonic diurnal component during the merging

432   of satellite data, errors in the diurnal cycle can cause errors in the data analysis following

433   the merging procedure. However, the removal of the diurnal cycle before merging may

434   also introduce some error into UAH and RSS merging procedures if the assumed diurnal

435   cycle is inaccurate, but physically, the removal of the diurnal harmonics before merging

436   seems to be a more logical approach as the diurnal harmonics will tend to add noise

437   unless removed.

438

439   On a global scale, the total impact of the diurnal correction applied by the RSS and UAH

440   groups to the microwave sounding data for the RSS data is to increase the decadal trend

441   by about 0.03ºC/decade for T_2 (Christy et al., 2003; Mears et al., 2003). The impact of the

442   Maryland group's adjustment is almost negligible. For the RSS T_2 data, when a diurnal

443   correction is applied that is 50% or 150% as large as the best estimate, these adjustments

444   significantly worsen the magnitude of the intersatellite differences. Changes of this

445   magnitude in the diurnal cycle lead to temperature trends that differ by 0.015ºC; so we

446   estimate that the uncertainty in trends due to uncertainty in the diurnal correction is about

447   0.015ºC/decade for T_2. The UAH group estimates that the diurnal correction for T_2 is

448   known to 0.01ºC/decade (Christy et al., 2000). These estimates of residual uncertainty are

449   relatively small, and are considerably less than the structural uncertainties associated with

450   the satellite merging methodology described in the next section. Despite the global

451   agreement for the diurnal adjustment for the RSS and UAH results, significant

452   differences in the adjustments exist as a function of location (Mears and Wentz, 2005),

453   which may explain some of the difference on smaller spatial scales between these two

454   datasets[6].

455

456   *3.2.2 Satellite merging methodology*

457

458   It is very likely that the most important source of uncertainty in microwave sounding

459   temperature trends is due to inter-satellite calibration offsets, and calibration drifts that

460   are correlated with the temperature of the calibration target (Christy et al., 2000; Mears et

461   al., 2003). When results from supposedly identical co-orbiting satellites are compared,

462   intersatellite offsets are immediately apparent. These offsets, typically a few tenths of a

463   ºC, must be identified and removed or they will produce errors in long-term trends of

464   several tenths of a ºC per decade. When constant offsets are used to remove the inter-

465   satellite differences, the UAH group found that significant differences still remain that

466   are strongly correlated with the temperature of the calibration target[7] (Christy *et al.,*

467   *2000*). This effect has since been confirmed by the RSS group (Mears *et al.,* 2003). Both

---

[6] See for example Figure 3.5 versus Figure 4.3.
[7] The calibration target can change temperature by tens of ºC over the course of the life of the satellite due to orbit- and season-dependent solar heating.

468   the UAH and RSS groups remove the calibration target temperature effect using a model

469   that includes a constant offset for each satellite, and an additional empirical "target

470   factor" multiplied by the calibration target temperature.

471

472   Despite the similarity in methods, the RSS and UAH groups obtain significantly different

473   values for the global temperature trends (see Table 3.3). In particular, the difference

474   between the trends for T_2 has received considerable attention. A close examination of the

475   procedures suggests that about 50% of the discrepancy in trends is accounted for by a

476   difference between the target factor for the NOAA-09 instrument deduced by the two

477   groups. This difference mainly arises from the subsets of data used by the two groups

478   when determining the satellite merging parameters (i.e., offsets and target factors). The

479   UAH group emphasizes pairs of satellites that have long periods of overlap, and thus uses

480   data from six pairs of satellites, while RSS uses all available (12) overlapping pairs of

481   satellites. Most of the remainder of the difference is due to a smaller difference in the

482   calibration target temperature proportionality constant for NOAA-11, and to small

483   differences in the diurnal correction. Both these differences primarily affect the

484   measurements made by NOAA-11 and NOAA-14, due to their large drifts in local

485   measurement time.

486

487   In Figure 4.1a, we plot the difference ($T_{2\text{-RSS}} - T_{2\text{-UAH}}$) between the RSS and UAH time

488   series. There is an obvious step that occurs in 1986, near the end of the NOAA-09

489   observation period, and a gradual slope that occurs during the observation periods of

490   NOAA-11 and NOAA-14. Note that the trend difference between these two datasets is

491   statistically significant at the 1% level, even though the error ranges quoted in Table 3.3

492   overlap, due to the presence of nearly identical short term fluctuations in the two datasets

493   (see Appendix A for more details).

494
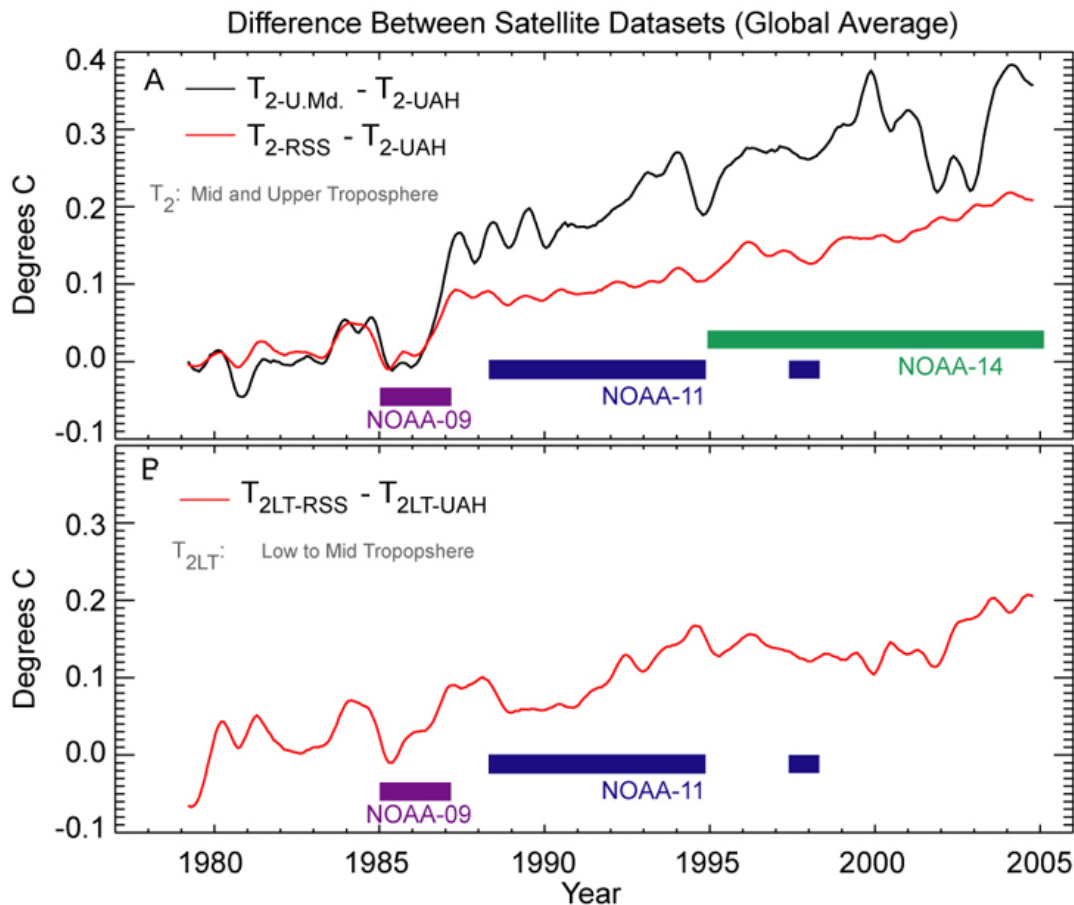
495

496   Figure 4.1 (a) Time series of the difference between global averages of satellite-derived T_2 datasets. Both
497   the RSS and UMD datasets show a step-like feature relative to the UAH dataset during the lifetime of
498   NOAA-09. The difference between the RSS and the UAH datasets shows a slow drift during the NOAA-11
499   and NOAA-14 lifetimes. Both these satellites drifted more than 4 hours in observations time. (b) Time
500   series difference between global averages of satellite derived T_{2LT} datasets. A slow drift is apparent during
501   the lifetime of NOAA-11, but the analysis during the NOAA-14 lifetime is complicated because the $T_{2LT\text{-}RSS}$
502   dataset does not include data from the AMSU instruments on NOAA-15 and NOAA-16, while the $T_{2LT\text{-}UAH}$
503   dataset does. All time series have been smoothed using a Gaussian filter with width = 7 months.

504

505   The Maryland group data set ($T_{2\text{-U.Md.}}$), in its most recent version (Grody et al., 2004;

506   Vinnikov et al., 2005), implemented a more detailed, physically based error model to

507   describe the errors that correlated with a nonlinear combination of the observed

508   brightness temperature measurements and the warm target temperature used for

509   calibration[8]. They use a substantially different merging procedure to deduce values of the

510   parameters that describe the intersatellite differences. First, as noted above, only the first

511   harmonic diurnal component is accounted for during the satellite merging, possibly

512   causing errors in the retrieved parameters. Second, they only use the spatial variation

513   seen by the different MSU instruments to derive the calibration adjustments and perform

514   long-time-scale temporal averaging of the measured temperatures to reduce the noise in

515   the overlapping satellite measurements. This averaging procedure may attenuate the time

516   dependent signal that the UAH empirical error model was introduced to explain. The

517   large step in the $T_{2\text{-U.Md.}} - T_{2\text{-UAH}}$ difference time series that occurs in 1986 (see Figure

518   4.1a) suggests that uncertainty in the parameters for the NOAA-09 satellite are also

519   important for this dataset[9]. The cause of the large fluctuations in the difference during the

520   2000-2004 time period is not known, but may be related to the absence of AMSU data in

521   the $T_{2\text{-U.Md.}}$ dataset. Due to its relatively recent appearance, considerably less is known

522   about the reasons for the differences between the Maryland dataset and the RSS and

523   UAH datasets, thus the comments about these differences should be viewed as more

524   speculative than the statements about the RSS-UAH differences.

---

[8] The Maryland group accounted for uncertainties in the radiometers non-linearity parameter as well as errors in the warm target radiation temperature (due to uncertainties in its emissivity and physical temperature) and errors in the cold space radiation temperature (due to uncertain antenna side lobe contributions for example). However, while all of these error sources are accounted for, they are assumed to be constant during the lifetime of a given instrument and thus do not take into account the possibility of contributions to the side lobe response from the earth or warm parts of the satellites whose temperature varies with time. These error sources lead, when globally averaged and linearized, to an expression where the target temperature is the most important factor. Thus while the exact physical cause of the observed effect is not known precisely, it is possible to accurately model and remove it on a global scale from the data using either method

[9] The trend in this difference time series is statistically significant at the 1% level.

525

526   These differences are an excellent example of structural uncertainty, where identical

527   input data and three seemingly reasonable methodologies lead to trends that differ

528   significantly more than the amount expected given their reported internal uncertainties.

529   Since methodological differences yield data products showing differences in trends in T_2

530   of about 0.1 °C per decade, it is clear that the most important source of uncertainty for

531   satellite data are structural uncertainties and that these need to be included in any overall

532   uncertainties assessed for tropospheric temperature trends and lapse rates.

533

534   *3.2.3 Differences in spatial pattern.*

535   Only $T_{2\text{-UAH}}$ and $T_{2\text{-RSS}}$ have provided gridded results. Maps of gridded trends for these

536   products are shown in Fig 4.2, along with a map of the difference between the trends. The

537   overall pattern in the trends is very similar between the two datasets, aside from

538   difference in the globally averaged trends. Differences in the latitude dependence are due

539   to the use of zonally varying intersatellite offsets in the construction of $T_{2\text{-UAH}}$ (in contrast

540   to the constant offsets in $T_{2\text{-RSS}}$) and to differences in the applied diurnal adjustment as a

541   function of latitude. Other differences may be caused by the spatial smoothing applied to

542   the $T_{2\text{-UAH}}$ during the construction of the data set, and to differences in spatial averaging

543   performed on the diurnal adjustment before it was applied. This last difference will be

544   discussed in more detail in section 4.4 below because the effects are more obvious for the

545   T_{2LT} layer.

T~Mid-Trop~ Temperature Trend 1979-2003 (°C/Decade)
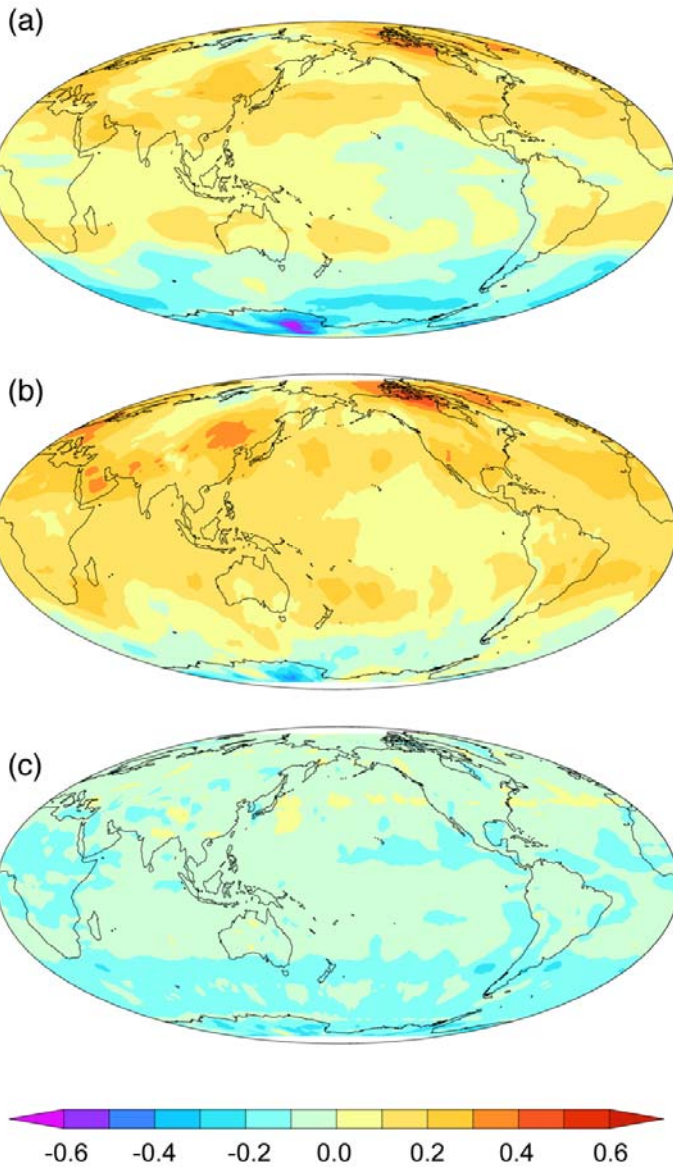
Figure 4.2 Global maps of trends from 1979-2004 for (a) $T_{2\text{-UAH}}$ and (b) $T_{2\text{-RSS}}$. Except for an overall difference between the two results, the spatial patterns are very similar. A map of the difference $T_{2\text{-UAH}}$ − $T_{2\text{-RSS}}$ between trends for the two products shown in (c) reveals more subtle differences in the trend.

551
552

4 Uncertainty in Lower Tropospheric Trends

554

555   *4.1 Radiosonde Uncertainty*

556   Uncertainties in lower tropospheric trends measured by radiosondes are very similar to

557   those discussed above for the middle-upper troposphere. The most important difference is

558   that when comparing to the T_{2LT} satellite product, the contribution of the stratospheric

559   radiosonde trends, which is suspected to be erroneous to some extent, is substantially less

560   than for the T_2 data records. This decreases the likelihood that T_{2LT} data products

561   constructed from radiosonde data are biased toward excess cooling. However, it is

562   possible that undetected negative trend bias remains in all tropospheric levels (see section

563   3.1 above for more details), so radiosonde trends may still be biased cold.

564

565   *4.2 Satellite Uncertainty*

566   Currently, there are two lower tropospheric satellite data records, $T_{2LT\text{-}UAH}$ and $T_{2LT\text{-}RSS}$.

567   As discussed in the Preface, both datasets are relatively recent, thus little is known about

568   the specific reasons for their differences. Because of the noise amplification effects of the

569   differencing procedure[10] used to construct the data record, the merging parameters tend to

570   be more sensitive to the methods used to deduce them. A number of different methods

571   were explored in the creation of $T_{2LT\text{-}RSS}$, leading to an estimate of the structural

572   uncertainty of 0.08ºC/decade for global trends. When combined with internal uncertainty,

573   the estimated total global trend uncertainty for this dataset is 0.09ºC/decade (Mears and

574   Wentz, 2005). Note that the difference between the global trends for $T_{2LT\text{-}RSS}$

---

[10] The T_{2LT} datasets are constructed by subtracting 3 times the average temperature measured by the outermost 4 (near-limb) views from 4 times the average temperature measured by the 4 adjacent views, which are closer to nadir. This has the effect of removing most of the stratospheric signal, and moving the effective weighting function lower in the troposphere (Spencer and Christy 1992). Assuming that the errors is each measurement are uncorrelated, this have the effect of amplifying these errors by a factor of about 5 relative to T_2 (Mears and Wentz 2005). Even if some of the error is correlated between view, this argument still applies to the uncorrelated portion of the error.

575   (0.189ºC/decade)   and $T_{2LT-UAH}$ (0.115ºC/decade)   shown   in

576   Table 3.3 is less than this estimated uncertainty. The estimated global trends in the

577   radiosonde datasets are also within the $T_{2LT-RSS}$ error range. In Figure 4.1b we plot the

578   difference ($T_{2LT-RSS} - T_{2LT-UAH}$) between the RSS and UAH time series. This time series

579   shows more variability than the corresponding T_2 difference time series, making it more

580   difficult to speculate about the underlying causes of the differences between them. The

581   step-like feature during the 1985-1987 period is less obvious, and while there appears to

582   be a slow drift during the NOAA-11 lifetime, a corresponding drift during the NOAA-14

583   lifetime is less obvious, perhaps because the RSS data do not yet include data from the

584   more recent AMSU satellites. We speculate that the drift during NOAA-11 is in part due

585   to differences in the diurnal correction applied. The UAH diurnal correction is based on

586   a parameterization of the diurnal cycle which is constrained by measurements made

587   during a time period with 3 co-orbiting satellites, , while RSS uses a model-based diurnal

588   correction analogous to that used for TMT.

589

590   In Figure 4.3, we show global maps of the gridded trends for $T_{2LT-UAH}$ and $T_{2LT-RSS}$, along

591   with a map of the trend differences. The spatial variability in the trend differences

592   between the two datasets is much larger than the variability for T_2, though both datasets

593   show similar patterns in general, with the greatest temperature increase occurring in the

594   Northern Hemisphere, particularly over Eastern Asia, Europe, and Northern Canada. The

595   two datasets are in relatively good agreement north of 45°N latitude. In the tropics and

596   subtropics, the largest differences occur over land, particularly over arid regions.

(a)

(b)

(c)

597
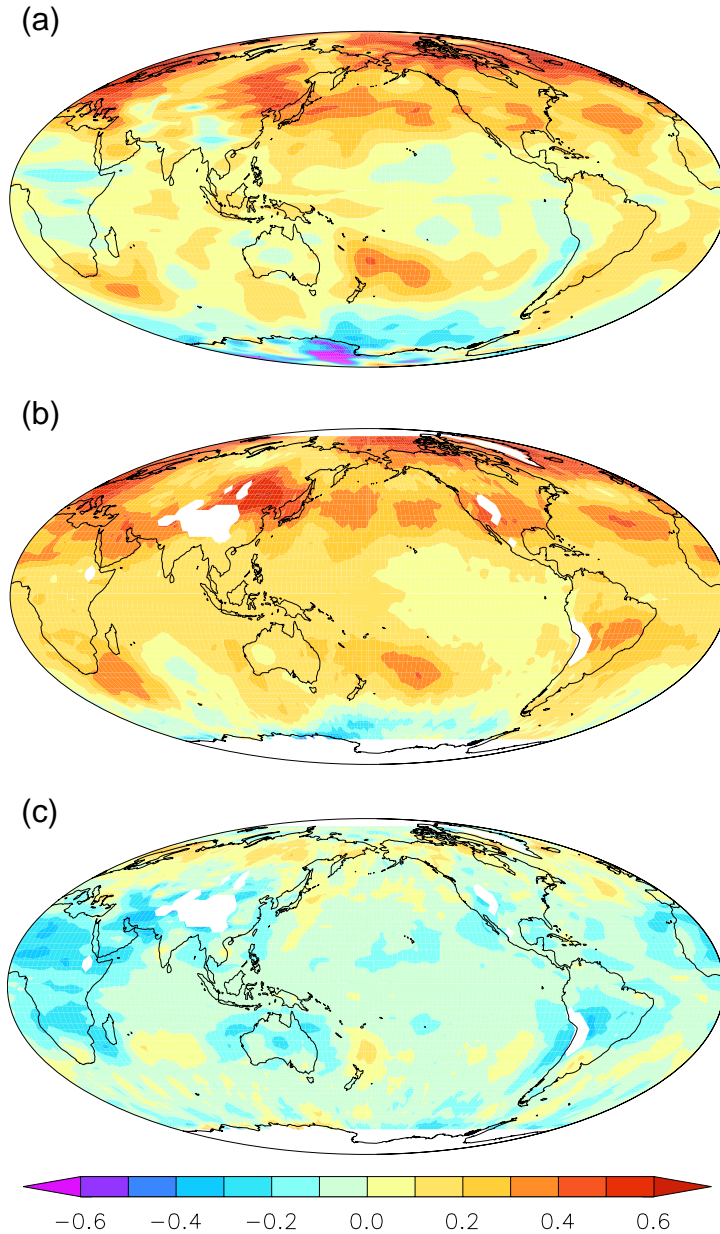
Figure 4.3 Global maps of trends from 1979-2004 for (a) $T_{2LT-UAH}$ and (b) $T_{2LT-RSS}$. Except for an overall
difference between the two results, the spatial patterns are similar. A map of the difference $T_{2LT-UAH} - T_{2LT-RSS}$ between trends for the two products shown in (c) shows that the largest differences are over tropical and subtropical land areas. Data from land areas with elevation higher than 2000m are excluded from the $T_{2LT-RSS}$ dataset and shown in white.

We speculate that this may be in part due to differences in how the diurnal adjustment is

done by the two groups. The UAH group applies an averaged diurnal adjustment for each

607   zonal band, based on different adjustments used for land and ocean. The RSS group uses

608   a grid-point resolution diurnal correction. The UAH method may lead to errors for

609   latitudes where the diurnal cycle varies strongly with longitude. More arid regions (e.g.,

610   subtropical Africa), which typically have much larger surface diurnal cycles, may be

611   under-adjusted when the zonally averaged correction is applied, leading to long-term

612   trends that are too low. Correspondingly more humid regions and oceans may be over-

613   adjusted, in some cases making up for the overall difference between the two datasets,

614   perhaps accounting for the good agreement in regions such as Southeast Asia, Southern

615   India, and Northern South America. Further analysis is required using a range of

616   alternative diurnal correction estimation techniques for definitive conclusions to be

617   reached. Other differences, such the north-south streaking seen in the RSS data, may be

618   caused by differences in spatial smoothing, and by the inclusion of AMSU data in $T_{2LT\text{-}}$

619    $_{UAH}$, but not in $T_{2LT\text{-}RSS}$.

620

621   The decay of orbital height over each satellite's lifetime can cause substantial errors in

622   satellite-derived T_{TLT} because changes in height lead to changes in the earth incidence

623   angles for the near-limb observations used to construct the data record Wentz and

624   Schabel, 1998). Both the RSS and UAH groups correct for this error by calculating the

625   expected change in observed temperature as a function of incidence angle, and then using

626   this estimate to remove the effect of orbital decay. The straight-forward method used to

627   make these corrections, combined with its insensitivity to assumptions about the vertical

628   structure of the atmosphere, leads to the conclusion that errors due to orbital decay have

629   been accurately removed from both datasets and are not an important cause of any

630   differences between them.

631

632   *4.3 Comparison between satellite and well characterized radiosonde stations*

633   Point-by-point comparisons between radiosonde and satellite data eliminate many

634   sources of sampling error normally present in radiosonde data. Also, since uniform global

635   coverage is less important when using radiosondes to validate satellite data locally,

636   stations can be chosen to minimize the contribution due to undocumented changes in

637   radiosonde instrumentation or observing practice. For instance, if one restricts

638   comparisons of the satellite and radiosonde data to 29 Northern Hemisphere radiosonde

639   stations that have consistently used a single type of instrumentation (the Viz sonde) since

640   1979, the average difference between these radiosonde trends and $T_{2LT-UAH}$ trends since

641   1979-2004 is only 0.03ºC/decade (Christy et al., 2003; Christy et al., 2005). Similarly,

642   when this set of radiosondes is extended to include a set of Southern Hemisphere stations

643   where instrument changes were well documented, agreement between $T_{2LT-UAH}$ and

644   radiosonde trends is almost as good (Christy and Norris, 2004; Christy et al., 2005). This

645   suggests that, for the T_{2LT} layer, where the stratospheric problems with radiosonde data

646   are minimized, some level of corroboration can be attained from these two diverse

647   measurement systems.

648

649   **<u>5 Multi-channel retrievals of tropospheric temperature.</u>**

650   As mentioned above, the single channel statellite measurements commonly identified as

651   tropospheric temperature (T_2) are impossible to interpret as solely tropospheric

652   temperatures because 10% to 15% (seasonally and latitudinally varying) of the signal

653   measured by MSU channel 2 arises from the stratosphere. In principle, it is possible to

654   reduce the stratospheric contribution to Channel 2 by subtracting out a portion of the

655   stratospheric Channel 4, though the exact values of the weights used in this procedure are

656   controversial (see Chapter 2 for more details). Despite this controversy, there is little

657   doubt that the resulting trends are more representative of the troposphere than the T_2

658   datasets. The reduction in stratospheric signal also reduces the difference between trends

659   in the satellite data and the radiosonde data (see Table 3.3), because the error-prone

660   stratospheric levels in the stratosphere have reduced (but still non-zero) weight.

661

662   The existence of a stratosphere-corrected tropospheric retrieval allows tests for

663   consistency of temperature trends among the different datasets constructed by a research

664   group for different atmospheric layers. One test, when applied to an earlier version (v5.1)

665   of the UAH global average trends, did not prove inconsistency on the global scale,

666   because the difference between the $T_{2LT-UAH}$ trend and the retrieval-calculated T_{2LT} trend

667   was well within the published margin of error. However, a clearer inconsistency was

668   found for the tropics (Fu and Johanson, 2005; Johanson and Fu, 2005). In this case, the

669   difference between the retrieval-calculated trend and $T_{2LT-UAH}$ trend was larger than its

670   estimated error range, an indication of uncharacterized error in at least one of the UAH

671   products, or more generally that $T_{2LT-UAH}$, T_{2-UAH} and T_{4-UAH} were not strictly self-

672   consistent as a set. This inconsistency is now resolved (within error estimates) with the

673   introduction of a new version of the $T_{2LT-UAH}$ dataset. The RSS versions of the T_2, T_4 and

674   T* datasets were found to be consistent for both global and tropical averages (Fu and

675   Johanson, 2005). The trends in the RSS version of the TLT dataset (produced after Fu

676   and Johanson was submitted) is also consistent with the other RSS based datasets.

677

678   **6 Uncertainty in Surface Trends.**

679

680   *6.1 Sea surface temperature uncertainty*

681   Temperature analyses over the ocean are produced from sea surface temperatures (SST)

682   instead of marine air temperatures. This is because marine air temperatures are biased

683   from daytime ship deck heating (Folland and Parker, 1995; Rayner et al., 2003) and

684   because satellite observations are available for SST beginning in November 1981 to

685   augment *in situ* data (Reynolds and Smith, 1994). Spatially complete analyses of SSTs

686   can be produced by combining satellite and *in situ* data (from ships and buoys) (Reynolds

687   et al., 2002; Rayner et al., 2003), from *in situ* data alone (Smith and Reynolds, 2004), or

688   from satellite data alone (Kilpatrick et al., 2001).

689

690   *6.1.1 Satellite SST uncertainties*

691   Climate comparison analyses based on infra-red satellite data alone are not useful

692   because of possible large time-dependent biases. These biases have typically occurred

693   near the end of a satellite's life time when the instrument no longer works properly, or

694   during periods when assumptions made about the atmospheric profile in the satellite

695   algorithm are no longer valid, e.g., during periods immediately following volcanic

696   eruptions, when a large amount of dust from the eruption is present in the stratosphere

697   (Reynolds, 1993; Reynolds et al., 2004). These problems may be partially mitigated in

698   the future by use of the microwave SST sensors that became available starting with the

699   launch of the Tropical Rainfall Measuring Mission (TRMM) in 1987 (Wentz et al.,

700   2000), but these microwave SST data have not been available long enough to derive

701   meaningful trends, and are difficult to calibrate absolutely due to various instrument

702   related problems (Wentz et al., 2001; Gentemann et al., 2004). Thus, analyses now use

703   multiple satellite instruments blended with or anchored to *in situ* data that reduce the

704   overall analysis errors (e.g., Reynolds et al., 2002, Rayner et al., 2003).

705

706   *6.1.2 In Situ SST uncertainties*

707   As discussed in Chapter 2, the primary sources of uncertainty in *in situ* SST

708   measurements are non-climatic signals caused by changes in the mix of instrumentation

709   over time and sampling errors. Over time the measurements have typically evolved from

710   insulated bucket measurements to engine intake, through hull, and buoy mounted sensors

711   – these changes are not necessarily accurately recorded in the metadata Both non-climatic

712   signals and sampling error are thought to be largest in sparsely sampled regions, such as

713   the southern oceans, where a single erroneous or unrepresentative measurement could

714   bias the average for an entire measurement cell for the month in question. Both types of

715   errors have been calculated for the ERSST dataset and included in the quoted error range

716   (see Figure 4.4).
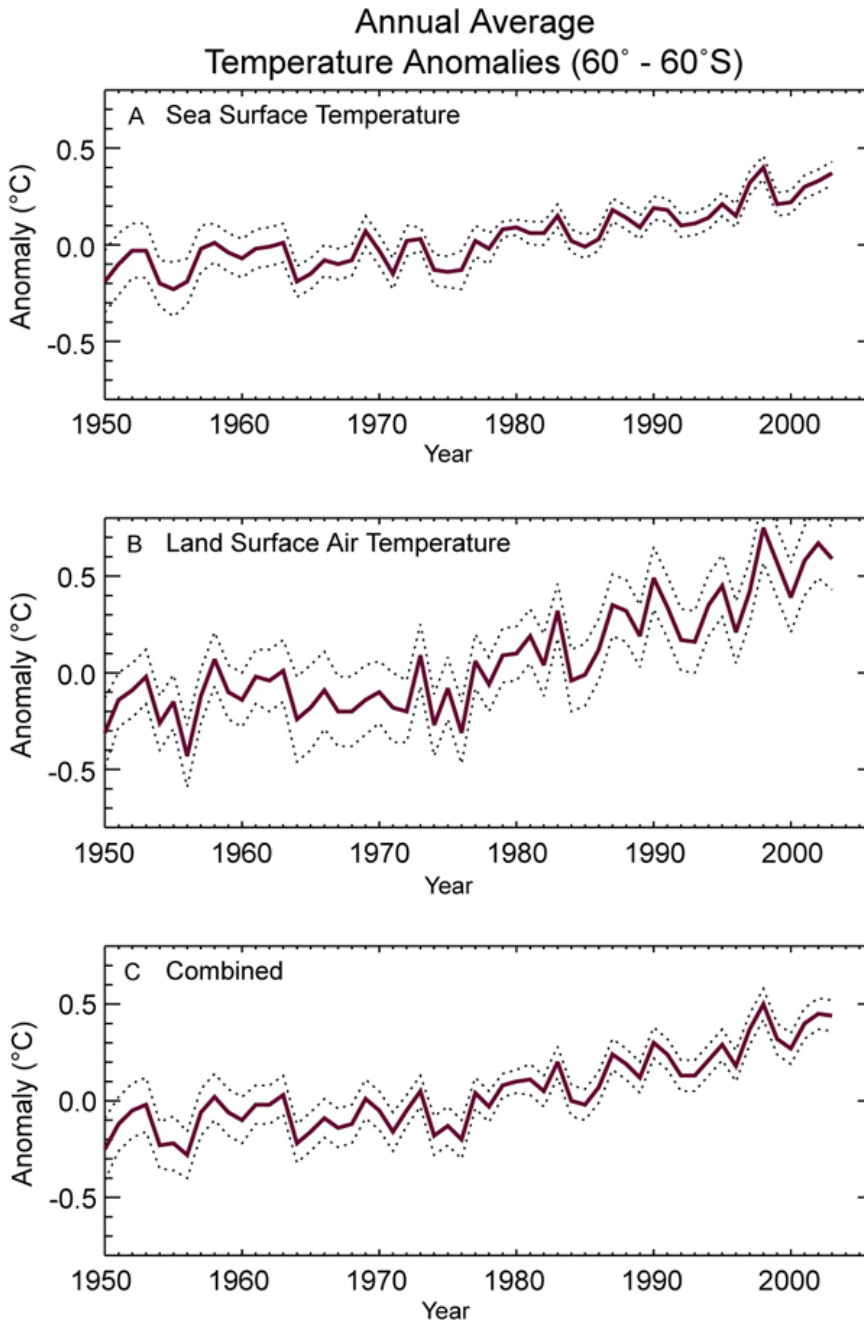
717

718   Figure 4.4. SST, Land Surface Air Temperature, and the Combined Temperature Data Record anomaly
719 averaged annually and between 60°S and 60°N (purple), with its estimated 95% confidence intervals
720 (dashed). Data are from the NOAA GHCN-ERSST dataset (Smith and Reynolds 2005).
721

722 *6.2 Land surface air temperature uncertainty*

723 The three surface temperature analyses exhibit similar warming rates since 1958. As the

724 surface data sets have many stations in common, they are not totally independent.

725   However, the MSU series take identical input, and radiosonde datasets have common

726   data also, so this problem is not unique to the surface records. The fact that the range in

727   trends is much smaller for the surface datasets than for these other datasets implies that

728   the structural uncertainty arising from dataset construction choices is much smaller at the

729   surface, in agreement with the arguments made in Thorne et al. (2005a). Also, a number

730   of studies e.g., (Peterson et al., 1999; Vose et al., 2004) suggest that long-term, large-

731   scale trends are not particularly sensitive to variations in choice of station networks. But

732   because most land networks were not designed for climate monitoring, the data contain

733   biases that dataset creators address with different detailed methods of analysis. The

734   primary sources of uncertainty from a land-surface perspective are (a) the construction

735   methods used in the analyses and (b) local environmental changes around individual

736   observing stations that may not have been addressed by the homogeneity assessments.

737

738   Because the stations are not fully representative of varying-within-area land surface,

739   coastal, and topographical effects, global data sets are produced by analyzing deviations

740   of temperature from station averages (anomalies) as these deviations vary more slowly

741   with a change in location than the temperatures themselves (Jones et al., 1997). Random

742   errors in inhomogeneity detection and adjustments may result in biased trend analyses on

743   a grid box level. However, on the relatively large space scales of greatest importance to

744   this Report, such problems are unlikely to be significant in current data sets in the period

745   since 1958 except where data gaps are still serious, e.g., in parts of central Africa, central

746   South America, and over parts of Antarctica. Note that for the contiguous United States,

747   the period 1958-2004 uses the greatest number of stations per grid box anywhere on the

748   Earth's land surface, generally upwards of 20 stations per grid box. For regions with

749   either poor coverage or data gaps, trends in surface air temperature should be regarded

750   with considerable caution, but do not have serious effects on the largest of scales as most

751   of the spatial variability is well sampled.

752

753   Local micro-climatological environmental changes around observing stations may be

754   problematic, particularly if a similar change occurred near many observing stations (e.g.,

755   Davey and Pielke, 2005). For instance, urbanization may have increased temperatures in

756   many locations. Numerous investigators have used a variety of approaches to study these

757   effects and most have shown that any bias is likely to be small in comparison to the

758   warming signal for large-scale means (e.g., Peterson, 2003; Parker, 2004). To insure that

759   potential urbanization effects do not impact analyses, NASA adjusts the data from all

760   urban stations so that their long-term trends are consistent with those from neighbouring

761   rural stations (Hansen et al., 2001). It is generally accepted that local biases in trends

762   mostly cancel through the use of many stations or ocean observations. Because such a

763   cancellation has not been rigorously proved, partly due to the lack of adequate metadata,

764   it is conceivable that systematic changes in many station exposures of a similar kind may

765   exist over the land during the last few decades, which may give biases in trends of one

766   sign over large land regions.

767
768   *6.3 Combined land-ocean analyses uncertainty.*

769   Global combined surface temperature products are computed by combining ocean and

770   land gridded datasets. The latest version of the UK surface dataset, HadCRUT2v, (Jones

771   and Moberg, 2003) has been optimally averaged with uncertainties for the globe and

772 hemispheres. The NOAA surface temperature dataset produced by (Smith and Reynolds,

773 2005), uses Global Historical Climatology Network (GHCN), merged with the *in situ*

774 Extended Reconstruction SST (ERSST) analysis of (Smith and Reynolds, 2004). The

775 analyses are done separately over the ocean and the land following the ERSST methods.

776 Error estimates include the bias, random and sampling errors.

777

778 As an example of uncertainties in a combined land-ocean analysis, near-global time

779 series (60°S to 60°N) are shown in Figure 4.4 for SST, land-surface air temperature, and

780 the combined SST and land-surface air temperature (Smith and Reynolds, 2005). (The

781 combined product is the GHCN-ERSST product used in Chapter 3). The SST has the

782 tightest (95%) uncertainty limits (upper panel). The land-surface air temperature (middle

783 panel) has a larger trend over the period since 1958, but its uncertainty limits are also

784 larger than for SST. Land surface air temperature uncertainty is larger than the

785 uncertainty for SST because of higher variability of surface air temperature over land (see

786 Chapter 1), persistently un-sampled regions, including central Africa and interior South

787 America, and because the calculations include an increasing urbanization bias-error

788 estimate. Merged temperature anomalies and their uncertainty (lower panel) closely

789 resemble the SST result, since oceans cover most of the surface area. Similar uncertainty

790 was found by (Folland et al., 2001) using different methods.

791

792

793 **7. Interlayer comparisons.**

794

795   *7.1 Troposphere/Stratosphere*

796

797   All data sources agree that on a global scale, the stratosphere has cooled substantially

798   while the troposphere has warmed over both the 1958-2004 and the 1979-2004 time

799   periods (note that this is not true for all 25-year time periods within the longer 1958-2004

800   time period). We suspect that the stratospheric cooling trends estimated from radiosondes

801   are larger in magnitude than the actual trend. Despite the uncertainty in the exact

802   magnitude of stratospheric cooling, we have very high confidence that the lower

803   stratosphere has cooled relative to the troposphere by several tenths of a ℃ per decade

804   over the past 5 decades.

805

806   *7.2 Lower Troposphere/Mid-Upper Troposphere*

807

808   The difference in trend between the lower troposphere and mid-upper troposphere is not

809   well characterized by the existing data. On a global scale, all data sets suggest that T_{2LT} is

810   warming relative to T_2, but it is important to note that the T_2 data records have significant

811   stratospheric contributions that reduce their warming trends. Radiosonde measurements

812   suggest that the $T_{(850-300)}$ layer (which does not include the stratosphere) is warming at

813   about the same rate as T_{2LT}, while satellite data suggest that T^*_G is warming more rapidly

814   than T_{2LT}. The magnitude of these inter-dataset differences are typically less than their

815   individual estimates of uncertainty, substantially reducing confidence in our ability to

816   deduce even the sign of the lower troposphere-mid-upper troposphere trend difference.

817

818   *7.3 Surface/Lower Troposphere*

819

820   On a global scale, both radiosonde datasets and one of the satellite datasets ($T_{2LT-UAH}$)

821   indicate that the surface warmed more than the lower troposphere between 1979 and

822   2004, while one satellite dataset ($T_{2LT-RSS}$) suggests the opposite. The magnitude of these

823   differences is less than the uncertainty estimates for any one data record. The situation is

824   similar in the tropics. However, in some regions, such as North America and Europe

825   (regions where the most reliable radiosonde stations are located), the warming in the

826   surface and lower troposphere appears to be very similar in all datasets.

827

828   *7.4 Surface/Mid Troposphere*

829

830   It is also interesting to consider the trend differences between the surface and mid

831   troposphere since more satellite datasets are available for T_2. Here, mostly due to the

832   large structural uncertainty in the trends in T_2, the various datasets are unable to agree on

833   the sign of the trend difference over the 1979-2004 period. On a global scale, the two

834   radiosonde datasets and two of the satellite datasets suggest that T_2 has warmed less than

835   the surface, but the other satellite dataset suggests that the opposite is true. The situation

836   is similar in the tropics. For the longer 1958-2004 period, all available datasets agree that

837   T_2 warms more than the surface. When T^*_G is considered, the difference between the

838   surface and tropospheric trends is reduced, with two satellite datasets indicating more

839   warming than at the surface.

840

841   **8. Resolution of Uncertainty**

842

843   In almost all of the tropospheric and stratospheric data records considered, our

844   uncertainty is dominated by structural uncertainty arising through dataset construction

845   choices (Thorne et al., 2005). Differences arising as a result of different, seemingly

846   plausible correction models applied by different groups to create a climate-quality data

847   record are significantly larger than the uncertainties internal to each method, in the raw

848   data measurements, or in the sampling uncertainties. These structural uncertainties are

849   difficult to assess in an absolute sense. The best estimates we can currently make come

850   from examining the spread of results obtained by different groups analyzing the same

851   type of data. This "all datasets are equal" approach has been employed in our present

852   analysis. As outlined in Chapter 2, this estimate of uncertainty can either be too small or

853   too large, depending on the situation. Given this caveat, it is always better to have

854   multiple (preferably at least three) data records that purport to measure the same aspects

855   of climate with the same data, so we can get some idea of the structural uncertainty.

856

857   In reality, all datasets are not equally plausible realizations of the true climate system

858   evolution. The climate system has evolved in a single way, and some datasets will be

859   closer to this truth than others. Given that the importance of structural uncertainty,

860   particularly for trends aloft, has only recently been recognized, it is perhaps unsurprising

861   that we are unable to quantify this at present. We could make value-based judgments to

862   imply increased confidence in certain datasets, but these would not be unambiguous, may

863   eventually be proven wrong, and are not a tenable approach in the longer term from a

864   scientific perspective. Therefore tools need to be developed to objectively discriminate

865   between datasets. These may include (1) measures of the internal consistency of the

866   construction methods, (2) assessment of the physical plausibility of the merged products,

867   including consistency of vertically resolved trends, and (3) comparisons with vicarious

868   data – for example, changes in temperature need to be compared with changes in water

869   vapor, winds, clouds, and various measures of radiation to assess consistency with the

870   expected physical relationships between these variables. Taken together such a suite of

871   indicators can be used to provide an objectively based way of highlighting residual

872   problems in the datasets and gaining a closer estimate of the truth. Such an audit of

873   current datasets should be seen as very high priority and preferably undertaken

874   independently of the dataset builders in a similar manner to the model intercomparisons

875   performed at Lawrence Livermore National Laboratory. In addition to an agreed set of

876   objective analysis tools, such an effort requires full and open access to all of the datasets

877   including a full audit trail.

878

879   Some specific suggestions for resolving some of the issues brought forward in this

880   chapter are mentioned here, but these are not exhaustive and further investigation is

881   required.

882

883   *8.1 Radiosondes.*

884

885   A significant contribution to the long-term inhomogeneity of the radiosonde record

886   appears to be related to changes in radiative heating of the temperature sensor for various

887   radiosonde models, and changes in the adjustments made to attempt to correct for these

888   changes. Recent work suggests that such problems may account for much of the tropical

889   cooling shown in unadjusted data. Other recent work suggests that step-like changes in

890   bias may still remain, even in adjusted datasets. Suitable tests on radiosonde products

891   may therefore include: stability of day-night differences, spatial consistency, internal

892   consistency (perhaps including wind data that to date have not been incorporated), and

893   consistency with MSU-derived and other independent estimates.

894

895   *8.2 Satellites.*

896

897   The most important contributions to satellite uncertainty are merging methodology and

898   the diurnal adjustment. The satellite data are simple enough that considerable

899   understanding can result from examination of intermediate results in the merging process,

900   including intersatellite differences that remain after the merging adjustments are

901   complete. Consistent reporting of such results can help differentiate between methods. It

902   appears that the differences in merging methodology often result in sharp step-like

903   features in difference time series between datasets. Other datasets, such as spatially

904   averaged adjusted radiosonde data, might be expected to show more slowly changing

905   errors, since their errors are due to the overlap of many different, potentially step-like

906   errors that occur at different times. So comparisons of satellite data with radiosonde data

907   over short time periods may help differentiate between satellite datasets. The diurnal

908   adjustment can be improved by a more rigorous validation of model-derived diurnal

909   cycles, or by further characterization of the diurnal cycle using the TRMM satellite or

910   concerted radiosonde observing programs designed to characterize the diurnal cycle at a

911   number of representative locations.

912

913   *8.3 Surface.*

914

915   The uncertainty in the historical near-surface temperature data is dominated by sampling

916   uncertainty, systematic changes in the local environment of surface observing stations,

917   and by difficult-to-characterize biases due to changes in SST measurement methods. The

918   relative maturity of the surface datasets suggest that to a large degree, these problems

919   have been addressed to the extent possible for the historical data, due to the absence of

920   the required metadata (for the bias-induced uncertainties) or the existence of any

921   observations at all. However, it is likely that much of the relatively recent SST data can

922   be adjusted for measurement type as some of the needed metadata is available or can be

923   estimated.

924

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1

2

3

4

Chapter 5

6

How well can the observed vertical temperature changes be reconciled with our

understanding of the causes of these changes?

9

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27 **KEY FINDINGS**

28

29   PATTERN STUDIES

30

31   Fingerprint studies use rigorous statistical methods to compare spatial and temporal patterns

32   of climate change in computer models and observations.

33

34 *1. Both human and natural factors have affected Earth's climate. Computer models are the*

35   *only tools we have for estimating the likely climate response patterns ("fingerprints")*

36   *associated with different forcing mechanisms.*

37

38   To date, most formal fingerprint studies have focused on a relatively small number of

39   climate forcings. Our best scientific understanding is that:

40

41   • Increases in well-mixed greenhouse gases (which are primarily due to fossil fuel

42   burning) result in large-scale warming of the Earth's surface and troposphere and

43   cooling of the stratosphere.

44   • Human-induced changes in the atmospheric burdens of sulfate aerosol particles

45   cause regional-scale cooling of the surface and troposphere.

46   • Depletion of stratospheric ozone cools the lower stratosphere and upper troposphere.

47   • Large volcanic eruptions cool the surface and troposphere (over 3 to 5 years) and

48   warm the stratosphere (over 1 to 2 years).

49   • Increases in solar irradiance warm throughout the atmospheric column (from the

50   surface to the stratosphere).

51

52   *2.   Results from many different fingerprint studies provide consistent evidence for a human*

53   *influence on the three-dimensional structure of atmospheric temperature over the second*

54   *half of the 20th century.*

55

56   Robust results are:

57

58   • Detection of greenhouse-gas and sulfate aerosol signals in observed surface

59   temperature records.

60   • Detection of an ozone depletion signal in stratospheric temperatures.

61   • Detection of the combined effects of greenhouse gases, sulfate aerosols, and ozone

62   in the vertical structure of atmospheric temperature changes (from the surface to the

63   stratosphere).

64

65   *3.   Natural factors have influenced surface and atmospheric temperatures, but cannot fully*

66   *explain their changes over the past 50 years.*

67

68   • The multi-decadal climatic effects of volcanic eruptions and solar irradiance

69   changes are identifiable in some fingerprint studies, but results are sensitive to

70   analysis details.

71

72

73   TREND COMPARISONS

74

75   Linear trend comparisons are less powerful than "fingerprinting" for studying cause-effect

76   relationships, but can highlight important differences (and similarities) between models and

77   observations.

78

79   *4.   When run with natural and human-caused forcings, model global-mean temperature*

80   *trends for individual atmospheric layers are consistent with observations.*

81

82   *5.   Comparing trend differences between the surface and the troposphere exposes potential*

83   *model-data discrepancies in the tropics.*

84

85   •   Differencing surface and tropospheric temperature time series (a simple measure of

86   the temperature lapse rate) removes much of the common variability between these

87   layers. This makes it easier to identify discrepancies between modeled and observed

88   lapse-rate changes.

89   •   For globally-averaged temperatures, model-predicted trends in tropospheric lapse

90   rates are consistent with observed results.

91   •   In the tropics, most observational datasets show more warming at the surface than in

92   the troposphere, while most model runs have larger warming aloft than at the

93   surface.

94

95

96   AMPLIFICATION OF SURFACE WARMING IN THE TROPOSPHERE

97

98   *6.*   *In the tropics, surface temperature changes are amplified in the free troposphere. Models*

99   *and observations show similar amplification behavior for monthly- and interannual*

100   *temperature variations, but not for decadal temperature changes.*

101

102   • Tropospheric amplification of surface temperature anomalies is due to the release of

103   latent heat by moist, rising air in regions experiencing convection.

104   • Despite large inter-model differences in variability and forcings, the size of this

105   amplification effect is remarkably similar in the models considered here, even across

106   a range of timescales (from monthly to decadal).

107   • On monthly and annual timescales, amplification is also a ubiquitous feature of

108   observations, and is very similar to values obtained from models and basic theory.

109   • For longer-timescale temperature changes over 1979 to 1999, only one of four

110   observed upper-air datasets has larger tropical warming aloft than in the surface

111   records. All model runs with surface warming over this period show amplified

112   warming aloft.

113   • These results have several possible explanations, which are not mutually exclusive.

114   One explanation is that "real world" amplification effects on short and long time

115   scales are controlled by different physical mechanisms, and models fail to capture

116   such behavior. A second explanation is that remaining errors in some of the

117   observed tropospheric data sets adversely affect their long-term temperature trends.

118   The second explanation is more likely in view of the model-to-model consistency of

119   amplification results, the large uncertainties in observed tropospheric temperature

120   trends, and independent physical evidence supporting substantial tropospheric

121   warming.

122

123   OTHER FINDINGS

124

125   *7.*   *It is important to account for observational uncertainty in comparisons between modeled*

126   *and observed temperature changes.*

127

128   •   There are large "construction uncertainties" in the process of generating climate data

129   records from raw observations. These uncertainties can critically influence the outcome

130   of consistency tests between models and observations.

131

132   *8.*   *Inclusion of spatially-heterogeneous forcings in the most recent climate models does not*

133   *fundamentally alter simulated lapse-rate changes at the largest spatial scales.*

134

135   •   Changes in black carbon aerosols and land use/land cover (LULC) may have had

136   significant influences on regional temperatures, but these influences have not been

137   quantified in formal fingerprint studies.

138   • These forcings were included for the first time in about half the global model

139   simulations considered here. Their incorporation did not significantly affect simulations

140   of lapse-rate changes at very large spatial scales (global and tropical averages).

141

142   **RECOMMENDATIONS**

143

144   *1.   Separate the uncertainties in climate forcings from uncertainties in the climate response*

145   *to forcings.*

146

147   The simulations of Twentieth Century (20CEN) climate analyzed here show climate

148   responses that differ because of differences in:

149

150   • Model physics and resolution;

151   • The forcings incorporated in the 20CEN experiment;

152   • The chosen forcing history, and the manner in which a specific forcing was applied.

153

154   We consider it a priority to partition the uncertainties in climate forcings and model

155   responses, and thus improve our ability to interpret differences between models and

156   observations. This could be achieved by better coordination of experimental design,

157   particularly for the 20CEN simulations that are most relevant for direct comparison with

158   observations.

159

160   ***2.   Quantify the contributions of changes in black carbon aerosols and land use/land cover***

161   ***to recent large-scale temperature changes.***

162

163   We currently lack experiments in which the effects of black carbon aerosols and LULC are

164   varied individually (while holding other forcings constant). Such "single forcing" runs will

165   help to quantify the contributions of these forcings to global-scale changes in lapse-rates.

166

167   ***3.   Explicitly consider model and observational uncertainty.***

168

169   Efforts to evaluate model performance or identify human-induced climate change should

170   always account for uncertainties in both observations and in model simulations of historical

171   and future climate. This is particularly important for comparisons involving long-term

172   changes in upper-air temperatures. It is here that current observational uncertainties are

173   largest and require better quantification.

174

175   ***4.   Perform the "next generation" of detection and attribution studies.***

176

177   Formal detection and attribution studies utilizing the new generation of model and

178   observational datasets detailed herein should be undertaken as a matter of priority.

179   1   ***Introduction***

180

181   A key scientific question addressed in this report is whether the Earth's surface has warmed more

182   rapidly than the troposphere over the past 25 years (*NRC*, 2000). Chapter 1 noted that there are

183   good physical reasons why we do not expect surface and tropospheric temperatures to evolve in

184   unison at all places and on all timescales. Chapters 2, 3, and 4 summarized our current

185   understanding of observed changes in surface and atmospheric temperatures. These chapters

186   identified important differences between surface and tropospheric temperatures, some of which

187   may be due to remaining problems with the observational data, and some of which are likely to be

188   real.

189

190   In Chapter 5, we seek to explain and reconcile the apparently disparate estimates of observed

191   changes in surface and tropospheric temperatures. We make extensive use of computer models of

192   the climate system. In the real world, multiple "climate forcings" vary simultaneously, and it is

193   difficult to identify and separate the climate effects of individual factors. Furthermore, the

194   experiment that we are performing with the Earth's atmosphere lacks a suitable control – we do not

195   have a convenient "parallel Earth" on which there are no human-induced changes in greenhouse

196   gases, aerosols, or other climate forcings. Climate models can be used to perform such controlled

197   experiments, or to simulate the response to changes in a single forcing or combination of forcings,

198   and thus have real advantages for studying cause-effect relationships. However, models also have

199   systematic errors that can diminish their usefulness as a tool for interpretation of observations

200   (*Gates et al.*, 1999; *McAvaney et al.*, 2001).

201

202   We evaluate published research that has made rigorous quantitative comparisons of modeled and

203   observed temperature changes, primarily over the satellite and radiosonde eras. Some new model

204   experiments (performed in support of the IPCC Fourth Assessment Report) involve simultaneous

205   changes in a wide range of natural and human-induced climate forcings. These experiments are

206   highly relevant for direct comparison with satellite-, radiosonde-, and surface-based temperature

207   observations. We review their key results here.

208

209   2   ***Model Simulations of Recent Temperature Change***

210

211   Many different types of computer model are used for studying climate change issues (*Meehl*, 1984;

212   *Trenberth*, 1992; see Box 5.1). Models span a large range of complexity, from the one- or two-

213   dimensional energy-balance models (EBMs) through Earth System models of intermediate

214   complexity (EMICs) to full three-dimensional atmospheric General Circulation Models (AGCMs)

215   and coupled atmosphere-ocean GCMs (CGCMs). Each type has advantages and disadvantages for

216   specific applications. The more complex AGCMs and CGCMs are most appropriate for

217   understanding problems related to the atmosphere's vertical temperature structure, since they

218   explicitly resolve that structure, and incorporate many of the physical processes (*e.g.*, convection,

219   interactions between clouds and radiation) thought to be important in maintaining atmospheric

220   temperature profiles. They are also capable of representing the horizontal and vertical structure of

221   unevenly-distributed climate forcings that may contribute to differential warming of the surface and

222   troposphere. Examples include volcanic aerosols (*Robock*, 2000) or the sulfate and soot aerosols

223   arising from fossil fuel or biomass burning (*Penner et al.*, 2001; *Ramaswamy et al.*, 2001a,b).

224

225 BOX 5.1:   Climate Models
226
227 Climate models provide us with estimates of how the real world's climate system behaves and is
228 likely to respond to changing natural and human-caused forcings. Because of limitations in our
229 physical understanding and computational capabilities, models are simplified and idealized
230 representations of a very complex reality. The most sophisticated climate models are direct
231 descendants of the computer models used for weather forecasting. While weather forecast models
232 seek to predict the specific timing of weather events over a period of days to several weeks, climate
233 models attempt to simulate future changes in the *average distribution* of weather events.
234 Simulations of 21st Century climate are typically based on "scenarios" of future emissions of
235 GHGs, aerosols and aerosol precursors, which in turn derive from scenarios of population changes,
236 economic growth, energy usage, developments in energy production technology, *etc*.
237

238

239 Climate models are also used to "hindcast" the climate changes that we have observed over the 20th

240 Century. When run in "hindcast" mode, a climate model is not constrained by *actual* weather

241 observations from satellites or radiosondes. Instead, it is driven by our best estimates of changes in

242 some (but probably not all) of the major "forcings", such as GHG concentrations, the Sun's energy

243 output, and the amount of volcanic dust in the atmosphere. In "hindcast" experiments, a climate

244 model is free to simulate the full four-dimensional (latitude, longitude, height/depth and time)

245 distributions of temperature, moisture, *etc*. Comparing the results of such an experiment with long

246 observational records constitutes a valuable test of model performance.

247

248 AGCM experiments typically rely on an atmospheric model driven by observed time-varying

249 changes in sea-surface temperatures (SSTs) and sea-ice coverage. This is a standard reference

250 experiment that many AGCMs have performed as part of the Atmospheric Model Intercomparison

251 Project ("AMIP"; *Gates et al.*, 1999). The AMIP-style experiments discussed here also include

252 specified changes in a variety of natural and human-caused forcing factors (*Hansen et al.*, 1997,

253 2002; *Folland et al.*, 1998; *Tett and Thorne*, 2004).

254

255   In both observations and climate models, variations in the El Niño/Southern Oscillation (ENSO)

256   have pronounced effects on surface and tropospheric temperatures (*Yulaeva and Wallace*, 1994;

257   *Wigley*, 2000; *Santer et al.*, 2001; *Hegerl and Wallace*, 2002; *Hurrell et al.*, 2003). When run in an

258   AMIP configuration, an atmospheric model "sees" the same changes in ocean surface temperature

259   that the real world's atmosphere experienced. The time evolution of ENSO effects on atmospheric

260   temperature is therefore very similar in the model and observations. This facilitates the direct

261   comparison of modeled and observed temperature changes.[1] Furthermore, AMIP experiments

262   reduce climate noise by focusing on the random variability arising from the atmosphere rather than

263   on the variability of the coupled atmosphere-ocean system (which is larger in amplitude). This

264   "noise reduction" aspect of AMIP runs has been exploited in efforts to identify human effects on

265   year-to-year changes in atmospheric temperatures (*Folland et al.*, 1998; *Sexton et al.*, 2001)

266

---

[1]This does not mean, however, that the atmospheric model will necessarily capture the correct amplitude and horizontal and vertical structure of the tropospheric temperature response to the specified SST ice changes. Note also that even with the specification of ocean boundary conditions, the time evolution of modes of variability that are forced by both the ocean and the atmosphere (such as the North Atlantic Oscillation; see *Rodwell et al.*, 1999) will not be the same in the model and in the real world (except by chance).

[1]Volcanic forcing provides an example of the signal estimation problem. The aerosols injected into the stratosphere during a massive volcanic eruption are typically removed within 2-3 years (*Sato et al.*, 1993; *Hansen et al.*, 2002; *Ammann et al.*, 2003). Because the large thermal inertia of the oceans cause a lag in response to this forcing, the cooling effect of the aerosols on the troposphere and surface persists for much longer than 2-3 years (*Santer et al.*, 2001; *Free and Angell*, 2002; *Wigley et al.*, 2005a). In the real world and in "AMIP-style" experiments, this slow, volcanically-induced cooling of the troposphere and surface is sometimes masked by the warming effects of El Niño events (*Christy and McNider*, 1994; *Wigley*, 2000; *Santer et al.*, 2001), thus hampering volcanic signal estimation.

[1]There are a variety of different spin-up strategies.

[1]In most of the experiments reported on here, *n* is between 3 and 5.
 and sea-ice changes. Note also that even with the specification of ocean boundary conditions, the time evolution of modes of variability that are forced by both the ocean and the atmosphere (such as the North Atlantic Oscillation; see *Rodwell et al.*, 1999) will not be the same in the model and in the real world (except by chance).

---

267   One disadvantage of the AMIP experimental set-up is that significant errors in one or more of the

268   applied forcing factors (or omission of key forcings) are not "felt" by the prescribed SSTs. Such

269   errors are more obvious in a CGCM experiment, where the ocean surface is free to respond to

270   imposed forcings. The lack of an ocean response and the masking effects of natural variability

271   make it difficult to use an AMIP-style experiment to estimate the slow response of the climate

272   system to an imposed forcing change.[2] CGCM experiments are more useful for this specific

273   purpose (see Chapter 1, Figure 1.3).

274

275   The CGCM experiments of interest here involve a model that has been "spun-up" until it reaches

276   some quasi-steady climate state[3]. The CGCM is then run with estimates of how a variety of natural

277   and human-caused climate forcings have changed over the 20th century. We refer to these

278   subsequently as "20CEN" experiments. Since the true state of the climate system is never fully

279   known, the same forcing changes are applied n times,[4] each time starting from a slightly different

280   initial climate state. This procedure yields n different realizations of climate change. All of these

281   realizations contain some underlying "signal" (the climate response to the imposed forcing

282   changes) upon which are superimposed n different manifestations of "noise" (natural internal

283   climate variability). Taking averages over these n realizations yields less noisy estimates of the

284   signal (*Wigley et al.*, 2005a).

---

[2]Volcanic forcing provides an example of the signal estimation problem. The aerosols injected into the stratosphere during a massive volcanic eruption are typically removed within 2-3 years (*Sato et al.*, 1993; *Hansen et al.*, 2002; *Ammann et al.*, 2003). Because the large thermal inertia of the oceans cause a lag in response to this forcing, the cooling effect of the aerosols on the troposphere and surface persists for much longer than 2-3 years (*Santer et al.*, 2001; *Free and Angell*, 2002; *Wigley et al.*, 2005a). In the real world and in "AMIP-style" experiments, this slow, volcanically-induced cooling of the troposphere and surface is sometimes masked by the warming effects of El Niño events (*Christy and McNider*, 1994; *Wigley*, 2000; *Santer et al.*, 2001), thus hampering volcanic signal estimation.

[3]There are a variety of different spin-up strategies.

[4]In most of the experiments reported on here, n is between 3 and 5.

285

286   In a CGCM, ocean temperatures are fully predicted rather than prescribed. This means that even a

287   (hypothetical) CGCM which perfectly captured all important aspects of ENSO physics would not

288   have the same timing of El Niño and La Niña events as the real world (except by chance). The fact

289   that ENSO variability – and its effects on surface and atmospheric temperatures – does not "line up

290   in time" in observations and CGCM experiments hampers direct comparisons between the two.[5]

291   This problem can be ameliorated by statistical removal of ENSO effects (*Santer et al.*, 2001;

292   *Hegerl and Wallace*, 2002; *Wigley et al.*, 2005a).[6]

293

294   The bottom line is that AMIP-style experiments and CGCM runs are both useful tools for exploring

295   the possible causes of differential warming.[7] We note that even if these two experimental

296   configurations employ the same atmospheric model and the same climate forcings, they can yield

297   noticeably different simulations of changes in atmospheric temperature profiles. These differences

298   arise for a variety of reasons, such as AGCM-versus-CGCM differences in sea-ice coverage, SST

299   distributions, and cloud feedbacks, and hence in climate sensitivity (*Sun and Hansen*, 2003).[8]

300

301   Most models undergo some form of "tuning". This involves changing poorly-known parameters

302   which directly affect key physical processes, such as convection and rainfall. Parameters are varied

---

[5]If *n* is large enough to adequately sample the (simulated) effects of natural variability on surface and tropospheric temperatures, it is not necessarily a disadvantage that the simulated and observed variability does not line up in time. In fact, this type of experimental set-up allows one to determine whether the single realization of the observations is contained within the "envelope" of possible climate solutions that the CGCM simulates.

[6]Residual effects of these modes of variability may remain in the data.

[7]Provided that comparisons with observations account for the specific advantages and disadvantages noted above.

[8]See, for example, the Ocean A and Ocean E results in Figure 3 of *Sun and Hansen* (2003).

303   within plausible ranges, which are generally derived from direct observations. The aim of tuning is

304   to reduce the size of systematic model errors and improve simulations of present-day climate.

305   Tuning does *not* involve varying uncertain model parameters over the course of a 20CEN

306   experiment, in order to improve a given model's simulation of observed climate change over the

307   20th Century.[9]

308

309   Several groups are now beginning to explore model "parameter space", and are investigating the

310   possible impact of parameter uncertainties on simulations of mean present-day climate and future

311   climate change (*Allen*, 1999; *Forest et al.*, 2002; *Murphy et al.*, 2004; *Stainforth et al.*, 2005). Such

312   work will help to quantify one component of model uncertainty. Another component of model

313   uncertainty arises from differences in the basic structure of models.[10] Section 5 considers results

314   from a range of state-of-the-art CGCMs, and thus samples some of the "structural uncertainty" in

315   model simulations of 20th Century climate change (Table 5.1). A further component of the "spread"

316   in simulations of 20th Century climate is introduced by uncertainties in the climate forcings with

317   which models are run (Table 5.2). These are discussed in the following Section.

318   Table 5.1: Acronyms of climate models referenced in this Chapter. All 19 models performed simulations of 20th
319   century climate change ("20CEN") in support of the IPCC Fourth Assessment Report. The ensemble size "ES" is the
320   number of independent realizations of the 20CEN experiment that were analyzed here.

---

[9]Potentially, highly uncertain climate forcings (particularly those associated with the indirect effects of aerosol particles on clouds) could be adjusted to improve the correspondence between modeled and observed global-mean surface temperature changes over the 20th Century. Such tuning does not occur *per se* and would be an unacceptable procedure, quite different from the parameter adjustments that are made when improving AGCM and CGCM simulations of mean climate.

[10]The computer models constructed by different research groups can have quite different "structures" in terms of their horizontal and vertical resolution, atmospheric dynamics (so-called "dynamical cores"), numerical implementation (*e.g.*, spectral versus grid-point), and physical parameterizations. They do, however, share many common assumptions.

	MODEL ACRONYM	COUNTRY	INSTITUTION	ES
1	CCCma-CGCM3.1(T47)	Canada	Canadian Centre for Climate Modelling and Analysis	1
2	CCSM3	United States	National Center for Atmospheric Research	5
3	CNRM-CM3	France	Météo-France/Centre National de Recherches Météorologiques	1
4	CSIRO-Mk3.0	Australia	CSIRO[1] Marine and Atmospheric Research	1
5	ECHAM5/MPI-OM	Germany	Max-Planck Institute for Meteorology	3
6	FGOALS-g1.0	China	Institute for Atmospheric Physics	1
7	GFDL-CM2.0	United States	Geophysical Fluid Dynamics Laboratory	3
8	GFDL-CM2.1	United States	Geophysical Fluid Dynamics Laboratory	3
9	GISS-AOM	United States	Goddard Institute for Space Studies	2
10	GISS-EH	United States	Goddard Institute for Space Studies	5
11	GISS-ER	United States	Goddard Institute for Space Studies	5
12	INM-CM3.0	Russia	Institute for Numerical Mathematics	1
13	IPSL-CM4	France	Institute Pierre Simon Laplace	1
14	MIROC3.2(medres)	Japan	Center for Climate System Research / NIES[2] / JAMSTEC[3]	3
15	MIROC3.2(hires)	Japan	Center for Climate System Research / NIES[2] / JAMSTEC[3]	1
16	MRI-CGCM2.3.2	Japan	Meteorological Research Institute	5
17	PCM	United States	National Center for Atmospheric Research	4
18	UKMO-HadCM3	United Kingdom	Hadley Centre for Climate Prediction and Research	1
19	UKMO-HadGEM1	United Kingdom	Hadley Centre for Climate Prediction and Research	1

321   [1]CSIRO is the Commonwealth Scientific and Industrial Research Organization.
322   [2]NIES is the National Institute for Environmental Studies.
323   [3]JAMSTEC is the Frontier Research Center for Global Change in Japan.
324

325

326   Table 5.2: Forcings used in IPCC simulations of 20th century climate change. This Table was compiled using
327   information   provided   by   the   participating   modeling   centers   (see   http://www-
328   pcmdi.llnl.gov/ipcc/model.documentation). Eleven different forcings are listed: well-mixed greenhouse gases (G),
329   tropospheric and stratospheric ozone (O), sulfate aerosol direct (SD) and indirect effects (SI), black carbon (BC) and
330   organic carbon aerosols (OC), mineral dust (MD), sea salt (SS), land use/land cover (LU), solar irradiance (SO), and
331   volcanic aerosols (V). Shading denotes inclusion of a specific forcing. As used here, "inclusion" means specification of

332 a time-varying forcing, with changes on interannual and longer timescales. Forcings that were varied over the seasonal
333 cycle only are not shaded.

	MODEL	G	O	SD	SI	BC	OC	MD	SS	LU	SO	V
1	CCCma-CGCM3.1(T47)	green		blue								
2	CCSM3	green	magenta	blue		black	gray				yellow	cyan
3	CNRM-CM3	green	magenta	blue		black						
4	CSIRO-Mk3.0	green		blue								
5	ECHAM5/MPI-OM	green	magenta	blue	dark blue							
6	FGOALS-g1.0	green		blue								
7	GFDL-CM2.0	green	magenta	blue		black	gray			olive	yellow	cyan
8	GFDL-CM2.1	green	magenta	blue		black	gray			olive	yellow	cyan
9	GISS-AOM	green		blue					dark green			
10	GISS-EH	green	magenta	blue	dark blue	black	gray	red	dark green	olive	yellow	cyan
11	GISS-ER	green	magenta	blue	dark blue	black	gray	red	dark green	olive	yellow	cyan
12	INM-CM3.0	green		blue							yellow	
13	IPSL-CM4	green		blue	dark blue							
14	MIROC3.2(medres)	green	magenta	blue		black	gray	red	dark green	olive	yellow	cyan
15	MIROC3.2(hires)	green	magenta	blue		black	gray	red	dark green	olive	yellow	cyan
16	MRI-CGCM2.3.2	green		blue							yellow	
17	PCM	green	magenta	blue							yellow	cyan
18	UKMO-HadCM3	green	magenta	blue	dark blue							
19	UKMO-HadGEM1	green	magenta	blue	dark blue	black	gray			olive	yellow	cyan

334

335

336

337   3   *Forcings in Simulations of Recent Climate Change*

338

339   In an ideal world, there would be reliable quantitative estimates of all climate forcings – both

340   natural and human-induced – that have made significant contributions to differential warming of

341   the surface and troposphere. We would have detailed knowledge of spatial and temporal changes in

342   these forcings. Finally, we would have used standard forcings to perform climate-change

343   experiments with a whole suite of numerical models, thus isolating uncertainties arising from

344   structural differences in the models themselves (see Box 5.2).

345

346

347

348 | Box 5.2:   Uncertainties in Simulated Temperature Changes
349 |
350 | In discussing the major sources of uncertainty in observational estimates of temperature change,
351 | Chapter 2 partitioned uncertainties into three distinct categories: "structural," "parametric," and
352 | "statistical." Uncertainties in simulated temperature changes fall into similar categories. In the
353 | modeling context, "structural" uncertainties can be thought of as the uncertainties resulting from
354 | the choice of a particular climate model, model configuration (Section 2), or forcing dataset
355 | (Section 3).
356 |
357 | Within a given model, there are small-scale physical processes (such as convection, cloud
358 | formation, precipitation, *etc.*) which cannot be simulated explicitly. Instead, so-called
359 | "parameterizations" represent the large-scale effects of these unresolved processes. Each of these
360 | has uncertainties in the values of key parameters.[11] Varying these parameters within plausible
361 | ranges introduces "parametric" uncertainty in climate change simulations (*Allen*, 1999; *Forest et
362 | al*., 2002; *Murphy et al*., 2004). Finally (analogous to the observational case), there is statistical
363 | uncertainty that arises from the unpredictable "noise" of internal climate variability, from the
364 | choice of a particular statistical metric to describe climate change, or from the application of a
365 | selected metric to noisy data.

366

---

[11]Note that some of these parameters influence not only the climate response, but also the portrayal of the forcing itself. Examples include parameters related to the size of sulfate aerosols, and how aerosol particles scatter incoming sunlight.

367

368   Unfortunately, this ideal situation does not exist. As part of the IPCC Third Assessment Report,

369   *Ramaswamy et al*. (2001b) assigned subjective confidence levels to our current "level of scientific

370   understanding" (LOSU) of the changes in a dozen different climate forcings. Only in the case of

371   well-mixed greenhouse gases ("GHGs"; carbon dioxide [CO_2], methane, nitrous oxide, and

372   halocarbons) was the LOSU characterized as "high." The LOSU of changes in stratospheric and

373   tropospheric ozone was judged to be "medium." For all other forcings (various aerosols, mineral

374   dust, land use-induced albedo changes, solar, *etc*.), the LOSU was estimated to be "low" or "very

375   low" (see Chapter 1, Table 1).[12]

376

377

378   In selecting the forcings for simulating the climate of the 20th Century (20CEN), there are at least

379   three strategies that modeling groups can adopt. The first strategy is to incorporate only those

380   forcings whose changes and effects are thought to be better understood, and for which time- and

381   space-resolved datasets suitable for performing 20CEN experiments are readily available. The

382   second strategy is to include a large number of different forcings, even those for which the LOSU

383   is "very low." A third strategy is to vary the size of poorly-known 20CEN forcings. This yields a

384   range of simulated climate responses, which are then used to estimate the levels of the forcings that

385   are consistent with observations (*e.g*., *Forest et al*., 2002).

386

387   The pragmatic focus of Chapter 5 is on climate forcings that have been incorporated in many

388   CGCM simulations of 20th century climate. The primary forcings that we consider are changes in

---

[12]We note that there is no direct relationship between the LOSU of a given forcing and the contribution of that forcing to 20th Century climate change. Forcings with "low" or "very low" LOSU may have had significant climatic impacts at regional and even global scales.

389   well-mixed GHGs, the direct effects of sulfate aerosol particles, tropospheric and stratospheric

390   ozone, volcanic aerosols, and solar irradiance. These are forcings whose effects on surface and

391   atmospheric temperatures have been quantified in rigorous fingerprint studies (see Section 4). This

392   does not diminish the importance of other climate forcings, whose global-scale contribution to

393   "differential warming" has not been reliably quantified to date.

394

395   Examples of these "other forcings" include carbon-containing aerosols produced during fossil fuel

396   or biomass combustion, human-induced changes in land surface properties, and the indirect effects

397   of tropospheric aerosols on cloud properties. There is emerging scientific evidence that such

398   spatially-variable forcings may have had important impacts on regional and even on global climate

399   (*NRC*, 2005). Some of this evidence is summarized in Box 5.3 and Box 5.4 for the specific cases of

400   carbonaceous aerosols and land use change. These and other previously-neglected forcings have

401   been included in many of the new CGCM simulations of 20th century climate described in Section

402   5 (see Tables 5.1 and 5.2).

403

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412	Box 5.3: Example of a Spatially-Heterogeneous Forcing: Black Carbon Aerosols
413	
414	Carbon-containing aerosols (also known as "carbonaceous" aerosols) exist in a variety of chemical
415	forms (*Penner et al.*, 2001). Two main classes of carbonaceous aerosol are generally distinguished:
416	"black carbon" (BC) and "organic carbon" (OC). Both types of aerosol are emitted during fossil
417	fuel and biomass burning. Most previous modeling work has focused on BC aerosols rather than
418	OC aerosols. Some of the new model experiments described in Section 5 have now incorporated
419	both types of aerosol in CGCM simulations of 20th century climate changes (see Tables 5.2 and
420	5.3).
421	
422	Black carbon aerosols absorb sunlight and augment the GHG-induced warming of the troposphere
423	(*Hansen et al.*, 2000; *Satheesh and Ramanathan*, 2000; *Penner et al.*, 2001; *Hansen*, 2002; *Penner*
424	*et al.*, 2003).[13] Their effects on atmospheric temperature profiles are complex, and depend on such
425	factors as the chemical composition, particle size, and height distribution of the aerosols (*e.g.*,
426	*Penner et al.*, 2003).
427	
428	*Menon et al.* (2002) showed that the inclusion of fossil fuel and biomass aerosols over China and
429	India[14] directly affected simulated vertical temperature profiles by heating the lower troposphere
430	and cooling the surface. In turn, this change in atmospheric heating influenced regional circulation
431	patterns and the hydrological cycle. *Krishnan and Ramanathan* (2002) found that an increase in
432	black carbon aerosols has reduced the surface solar insolation (exposure to sunlight) over the
433	Indian subcontinent. Model experiments performed by *Penner et al.* (2003) suggest that the net
434	effect of carbonaceous aerosols on global-scale surface temperature changes depends critically on
435	how aerosols affect the vertical distribution of clouds. On regional scales, the surface temperature
436	effects of these aerosols are complex, and vary in sign (*Penner et al.*, 2005).
437	
438	
439	
440	
441	
442	
443	
444	

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[13]Note that soot particles are sometimes transported long distances by winds, and can also have a "far field" effect on climate by reducing the reflectivity of snow in areas remote from pollution sources (*Hansen and Nazarenko*, 2003; *Jacobson*, 2004).

[14]During winter and spring, black carbon aerosols contribute to a persistent haze over large areas of Southern Asian and the Northern Indian Ocean (*Ramanathan et al.*, 2001).

445 | Box 5.4:   Example of a Spatially-Heterogeneous Forcing: Land Use Change
446 | Humans have transformed the surface of the planet through such activities as conversion of forest
447 | to cropland, urbanization, irrigation, and large water diversion projects (see Chapter 4). These
448 | changes can affect a variety of physical properties of the land surface, such as the albedo
449 | (reflectivity), the release of water by plants (transpiration), the moisture-holding capacity of soil,
450 | and the surface "roughness." Alterations in these physical properties may in turn affect runoff, heat
451 | and moisture exchanges between the land surface and atmospheric boundary layer, wind patterns,
452 | and even rainfall (*e.g.*, *Pitman et al.*, 2004). Depending on the nature of the change, either warming
453 | or cooling of the land surface may occur (*Myhre and Myhre*, 2003).
454 |
455 | At the regional level, modeling studies of the Florida peninsula (*Marshall et al.*, 2004) and
456 | southwest Western Australia (*Pitman et al.*, 2004) have linked regional-scale changes in
457 | atmospheric circulation and rainfall to human transformation of the natural vegetation. Modeling
458 | work focusing on North America suggests that the conversion of natural forest and grassland to
459 | agricultural production has led to a cooling in summertime (*Oleson et al.*, 2004). The global-scale
460 | signal of land use/land cover (LULC) changes from pre-industrial times to the present is estimated
461 | to be a small net cooling of surface temperature (*Matthews et al.*, 2003, 2004; *Brovkin et al.*, 2004;
462 | *Hansen et al.*, 2005a; *Feddema et al.*, 2005). Larger regional trends of either sign are likely to be
463 | evident (*e.g.*, *Hansen et al.*, 2005a).[15]

464

465   Clearly, we will *never* have complete and reliable information on all forcings that are thought to

466   have influenced climate over the late 20th century. A key question is whether those forcings most

467   important for understanding the differential warming problem are reliably represented. This is

468   currently difficult to answer. What we *can* say, with some certainty, is that the expected

469   atmospheric temperature signal due to forcing by well-mixed GHGs alone is distinctly different

470   from the signal due to the combined effects of multiple natural and human forcing factors (Chapter

471   1; *Santer et al.*, 1996; *Tett et al.*, 1996; *Hansen et al.*, 1997, 2002; *Bengtsson et al.*, 1999; *Santer et*

472   *al.*, 2003a).

473

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[15]Note that larger regional trends do not necessarily translate to enhanced detectability. Although the signals of LULC and other spatially-heterogeneous forcings are likely to be larger regionally than globally, the "noise" of natural climate variability is also larger at smaller spatial scales. It is not obvious *a priori*, therefore, how signal-to-noise relationships (and detectability of a given forcing's climate effects) behave as one moves from global to continental to regional scales.

474   This is illustrated by the 20CEN and "single forcing" experiments performed with the Parallel

475   Climate Model (PCM; *Washington et al.*, 2000). In PCM, changes in the vertical profile of

476   atmospheric temperature over 1979 to 1999 are primarily forced by changes in well-mixed GHGs,

477   ozone, and volcanic aerosols (Figure 5.1). Changes in solar irradiance and the scattering effects of

478   sulfate aerosols are of secondary importance over this period. Even without performing formal

479   statistical tests, it is visually obvious from Figure 5.1 that radiosonde-based estimates of observed

480   stratospheric and tropospheric temperature changes are in better agreement with the PCM 20CEN

481   experiment than with the PCM "GHG only" run.

482

483
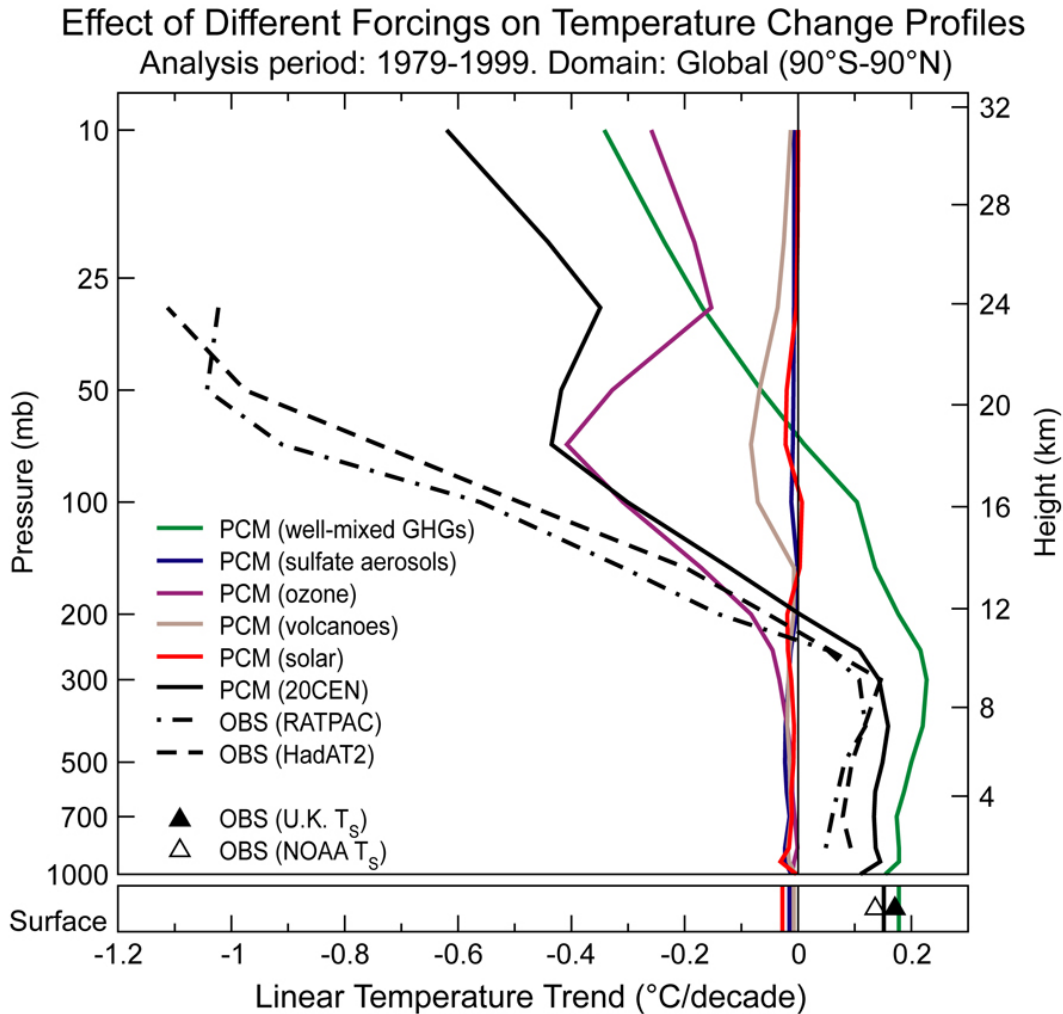
Figure 5.1: Vertical profiles of global-mean atmospheric temperature change over 1979 to 1999. Surface temperature changes are also shown. Results are from two different radiosonde data sets (HadAT2 and RATPAC; see Chapter 3) and from single forcing and combined forcing experiments performed with the Parallel Climate Model (PCM; *Washington et al.*, 2000). PCM results for each forcing experiment are averages over four different realizations of that experiment. All trends were calculated with monthly mean anomaly data.

This illustrates the need for caution in comparisons of modeled and observed atmospheric

temperature change. The differences evident in such comparisons have multiple interpretations.

They may be due to real errors in the models,[16] errors in the forcings used to drive the models, the

neglect of important forcings, and residual inhomogeneities in the observations themselves. They

---

[16]These may lie in the physics, parameterizations, inadequate horizontal or vertical resolution, *etc*.

495   may also be due to different manifestations of natural variability noise in the observations and a

496   given CGCM realization. All of these factors may be important in model evaluation work.

497

498   4.   *Published Comparisons of Modeled and Observed Temperature Changes*

499

500   A number of observational and modeling studies have attempted to shed light on the possible

501   causes of "differential warming".[17] We have attempted to organize the discussion of results so that

---

[17]We do not discuss studies which provide empirical estimates of "equilibrium climate sensitivity" – the steady-state warming of the Earth's surface that would eventually be reached after the climate system equilibrated to a doubling of pre-industrial atmosp tropospheric temperatures to massive volcanic eruptions (*Hansen et al.*, 1993; *Lindzen and Giannitsis*, 1998; *Douglass and Knox*, 2005; *Wigley et al.*, 2005a,b; *Robock*, 2005); the "intermediate" (100- to 150-year) response of surface temperatures to natural and human-caused forcing changes over the 19th and 20th centuries (*Andronova and Schlesinger*, 2001; *Forest et al.*, 2002; *Gregory et al.*, 2002; *Harvey and Kaufmann*, 2002) or to solar and volcanic forcing changes over the past 1-2 millennia (*Crowley*, 2000), and the slow (100,000-year) response of Earth's temperature to orbital changes between glacial and interglacial conditions (*Hoffert and Covey*, 1992; *Hansen et al.*, 1993). These investigations are not directly relevant to elucidation of the causes of changes in the vertical structure of atmospheric temperatures, which is the focus of our Chapter.

[17]It is useful to mention one technical issue relevant to model-data comparisons. As noted in Chapter 2, the satellite-based Microwave Sounding Unit (MSU) monitors the temperature of very broad atmospheric layers. To facilitate comparisons with observed MSU datasets, many of the studies reported on here calculate "synthetic" MSU temperatures from climate model experiments. Technical aspects of these calculations are discussed in Chapter 2, Box 2.

[17]The studies by *Jones* (1994) and *Christy and McNider* (1994) remove volcano and ENSO effects from T_{2LT}, and estimate residual trends of 0.093 and 0.090°C/decade over 1979 to 1993. A similar investigation by *Michaels and Knappenberger* (2000) obtained a residual trend of 0.041°C/decade over 1979 to 1999. The error bars on these residual trend estimates are either not given, or claimed to be very small (*e.g.*, ± 0.005°C/decade in *Christy and McNider*). A fourth study removed combined ENSO, volcano, and solar effects from T_{2LT}, and estimated a residual trend of 0.065 ± 0.012°C/decade over 1979 to 2000 (*Douglass and Clader*, 2000).
heric CO_2 levels. This is often referred to as $\Delta T_{2 \times CO2}$. Estimates of $\Delta T_{2 \times CO2}$ have been obtained by studying Earth's temperature response to "fast", "intermediate", and "slow" forcing of the climate system. Examples include the "fast" (<10-year) response of surface and tropospheric temperatures to massive volcanic eruptions (*Hansen et al.*, 1993; *Lindzen and Giannitsis*, 1998; *Douglass and Knox*, 2005; *Wigley et al.*, 2005a,b; *Robock*, 2005); the "intermediate" (100- to 150-year) response of surface temperatures to natural and human-caused forcing changes over the 19th and 20th centuries (*Andronova and Schlesinger*, 2001; *Forest et al.*, 2002; *Gregory et al.*, 2002; *Harvey and Kaufmann*, 2002) or to solar and volcanic forcing changes over the past 1-2 millennia (*Crowley*, 2000), and the slow (100,000-year) response of Earth's temperature to orbital changes between glacial and interglacial conditions (*Hoffert and Covey*, 1992; *Hansen et al.*, 1993). These investigations are not directly relevant to elucidation of the causes of changes in the vertical structure of atmospheric temperatures, which is the focus of our Chapter.

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502   investigations with similar analysis methods are grouped together.[18] Our discussion proceeds from

503   simple to more complex and statistically rigorous analyses.

504

505   4.1   *Regression studies using observed global-mean temperature data*

506

507   One class of study that has attempted to address the causes of recent tropospheric temperature

508   change relies on global-mean observational data only (*Jones*, 1994; *Christy and McNider*, 1994;

509   *Michaels and Knappenberger*, 2000; *Douglass and Clader*, 2002). Such work uses a multiple

510   regression model to quantify the statistical relationships between various "predictor variables"

511   (typically time series of ENSO variability, volcanic aerosol loadings, and solar irradiance) and a

512   single "predictand" (typically T_{2LT}). The aim is to remove the effects of the selected predictors on

513   tropospheric temperature, and to estimate the residual trend that may arise from human-induced

514   forcings. The quoted values for this residual trend in T_{2LT} range from 0.04 to 0.09°C/decade.[19]

515

516   These studies make the unrealistic assumption that the uncertainties inherent in such statistical

517   signal separation exercises are very small. They do not explore the sensitivity of regression results

518   to uncertainties in the predictor variables, and generally use solar and volcanic *forcings* as

519   predictors rather than the climate *responses* to those forcings. Distinctions between forcing and

---

[18]It is useful to mention one technical issue relevant to model-data comparisons. As noted in Chapter 2, the satellite-based Microwave Sounding Unit (MSU) monitors the temperature of very broad atmospheric layers. To facilitate comparisons with observed MSU datasets, many of the studies reported on here calculate "synthetic" MSU temperatures from climate model experiments. Technical aspects of these calculations are discussed in Chapter 2, Box 2.

[19]The studies by *Jones* (1994) and *Christy and McNider* (1994) remove volcano and ENSO effects from T_{2LT}, and estimate residual trends of 0.093 and 0.090°C/decade over 1979 to 1993. A similar investigation by *Michaels and Knappenberger* (2000) obtained a residual trend of 0.041°C/decade over 1979 to 1999. The error bars on these residual trend estimates are either not given, or claimed to be very small (*e.g.*, ± 0.005°C/decade in *Christy and McNider*). A fourth study removed combined ENSO, volcano, and solar effects from T_{2LT}, and estimated a residual trend of 0.065 ± 0.012°C/decade over 1979 to 2000 (*Douglass and Clader*, 2000).

520   response are important (*Wigley et al.*, 2005a). Accounting for uncertainties in predictor variables

521   (and use of responses rather than forcings as predictors) expands the range of uncertainties in

522   estimates of residual T_{2LT} trends (*Santer et al.*, 2001).[20]

523

524   Regression methods have also been used to estimate the net effects of ENSO and volcanoes on

525   trends in global-mean surface and tropospheric temperatures. For T_{2LT}, both *Jones* (1994) and

526   *Christy and McNider* (1994) found that ENSO effects induced a small net warming of 0.03 to

527   0.05°C/decade over 1979 to 1993, while volcanoes caused a cooling of 0.18°C/decade over the

528   same period. *Michaels and Knappenberger* (2000) also reported a relatively small ENSO influence

529   on T_{2LT} trends.[21] *Santer et al.* (2001) noted that over 1979 to 1997, volcanoes had likely cooled the

530   troposphere by more than the surface. Removing the combined volcano and ENSO effects from

531   surface and UAH T_{2LT} data helped to explain some of the observed differential warming: the "raw"

532   T_S-minus-T_{2LT} trend over 1979 to 1997 decreased from roughly 0.15°C/decade to 0.05°-

533   0.13°C/decade.[22] Removal of volcano and ENSO influences also brought observed lapse rate trends

534   closer to model results, but could not fully reconcile modeled and observed lapse rate trends.[23]

535

---

[20]*Santer et al.* (2001) obtain residual T_{2LT} trends ranging from 0.06 to 0.16°C/decade over 1979 to 1999. Their regression model is iterative, and involves removal of ENSO and volcano effects only.

[21]Their T_{2LT} trends were 0.04°C/decade over 1979 to 1998 and 0.01°C/decade over 1979 to 1999. This difference in the net ENSO influence on T_{2LT} (with the addition of only a single year of record) arises from the El Niño event in 1997/98, and illustrates the sensitivity of this kind of analysis to so-called "end effects".

[22]The latter results were obtained with the HadCRUTv surface data (*Jones et al.*, 2001) and version d03 of the UAH T_{2LT} data. The range of residual lapse-rate trends arises from parametric uncertainty, *i.e.*, from the different choices of ENSO predictor variable and volcano parameters.

[23]*Santer et al.* (2001) analyzed model experiments performed with the ECHAM4/OPYC model developed at the Max-Planck Institute for Meteorology in Hamburg (*Roeckner et al.*, 1999). The experiments included forcing by well-mixed greenhouse gases, direct and indirect sulfate aerosol effects, tropospheric and stratospheric ozone, and volcanic aerosols (Pinatubo only).

536   4.2   *Regression studies using spatially-resolved temperature data*

537

538   Other regression studies have attempted to remove natural variability influences using spatially-

539   resolved temperature data. Regression is performed "locally" at individual grid-points and/or

540   atmospheric levels. To obtain a clearer picture of volcanic effects on atmospheric temperatures,

541   *Free and Angell* (2002) removed the effects of variability in ENSO and the Quasi-Biennial

542   Oscillation (QBO) from Hadley Centre radiosonde data[24]. Their work clearly shows that the

543   cooling effect of massive volcanic eruptions has been larger in the upper troposphere than in the

544   lower troposphere. The implication is that volcanic effects probably contribute to slow changes in

545   observed lapse rates.

546

547   *Hegerl and Wallace* (2002) used regression methods to identify and remove different components

548   of natural climate variability from gridded fields of surface temperature data, the UAH T_{2LT}, and

549   "synthetic" T_{2LT} calculated from radiosonde data. They focused on the variability associated with

550   ENSO and the so-called "cold ocean warm land" (COWL) pattern (*Wallace et al.*, 1995). While

551   ENSO and COWL variability made significant contributions to the month-to-month and year-to-

552   year variability of temperature differences between the surface and T_{2LT}, it had very little impact on

553   decadal fluctuations in lapse rate. The authors concluded that natural variability alone was unlikely

554   to explain these slow lapse-rate changes. However, the removal of ENSO and COWL effects more

555   clearly revealed a volcanic contribution, consistent with the findings of *Santer et al.* (2001) and

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[24]The HadRT2.1 dataset of Parker *et al.* (1997). Like *Santer et al.* (2001), *Free and Angell* (2002) also found some sensitivity of the estimated volcanic signals to "parametric" uncertainty.

556   *Free and Angell* (2002). A climate model control run (with no changes in forcings) and a 20CEN

557   experiment were unable to replicate the observed decadal changes in lapse rate.[25]

558

559   4.3   *Other studies of global and tropical lapse-rate trends*

560

561   Several studies have investigated lapse-rate trends without attempting to remove volcano effects or

562   natural climate noise. *Brown et al*. (2000) used surface, radiosonde, and satellite data to identify

563   slow, tropic-wide changes in the lower tropospheric lapse rate.[26] In their analysis, the surface

564   warmed relative to the troposphere between the early 1960s and mid-1970s and after the early

565   1990s. Between these two periods, the tropical troposphere warmed relative to the surface. The

566   spatial coherence of these variations (and independent evidence of concurrent variations in the

567   tropical general circulation) led *Brown et al*. (2000) to conclude that tropical lapse rate changes

568   were unlikely to be an artifact of residual errors in the observations.

569

570   Very similar decadal changes in lower tropospheric lapse rate were reported by *Gaffen et al*.

571   (2000).[27] Their study analyzed radiosonde-derived temperature and lapse rate changes over two

572   periods: 1960 to 1997 and 1979 to 1997. Tropical lapse rates decreased over the longer period[28]

---

[25]The model was the ECHAM4/OPYC CGCM used by *Bengtsson et al*. (1999). The 20CEN experiment analyzed by *Hegerl and Wallace* (2002) involved combined changes in well-mixed greenhouse gases, the direct and indirect effects of sulfate aerosols, and tropospheric ozone. Forcing by volcanoes and stratospheric ozone depletion was not included.

[26]The *Brown et al*. (2000) study employed UKMO surface data (HadCRUT), version d of the UAH T_{2LT}, and an early version of the Hadley Centre radiosonde dataset (HadRT2.0) that was uncorrected for instrumental biases.

[27]*Gaffen et al*. (2000) used a different radiosonde dataset from that employed by *Brown et al*. (2000). The two groups also analyzed different surface temperature datasets.

[28]Corresponding to a tendency towards a more stable atmosphere.

---

573   and increased over the satellite era.[29] To evaluate whether natural climate variability could explain

574   these slow variations, *Gaffen et al.* (2000) computed lapse rates from the control runs performed

575   with three different CGCMs. Each control run was 300 years in length. These long runs provided

576   estimates of the "sampling variability" of modeled lapse rate changes on timescales relevant to the

577   two observational periods (38 and 19 years).[30] Model-based estimates of natural climate variability

578   could not explain the observed tropical lapse rate changes over 1979 to 1997. Similar conclusions

579   were reached by *Hansen et al.* (1995) and *Santer et al.* (2000). Including natural and anthropogenic

580   forcings in the latter study narrowed the gap between modeled and observed estimates of recent

581   lapse-rate changes, although a significant discrepancy between the two still remained.

582

583   It should be emphasized that *all* of the studies reported on to date in Section 4 relied on satellite

584   data from one group only (UAH), on early versions of the radiosonde data[31], and on experiments

585   performed with earlier model "vintages." It is likely, therefore, that this work may have

586   underestimated the structural uncertainties in observed and simulated estimates of lapse rate

587   changes. We will consider in Section 5 whether modeled and observed lapse rate changes can be

588   better reconciled by the availability of more recent 20CEN runs and more comprehensive estimates

589   of structural uncertainties in observations.

590

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[29]These lapse-rate changes were accompanied by increases and decreases in tropical freezing heights (which were inferred from the same radiosonde data).

[30]This was done by generating, for each control run, distributions of 38-year and 19-year lapse rate trends. For example, a 300-year control run can be split up into 15 different "segments" that are each of length 19 years (assuming there is no overlap between segments). From these segments, one obtains 15 different estimates of how the lapse rate might vary in the absence of any forcing changes. The observed lapse rate change over 1979 to 1997 is then compared with the model trend distribution to determine whether the observed result could be explained by natural variability alone.

[31]These radiosonde datasets were either unadjusted for inhomogeneities, or had not been subjected to the rigorous adjustment procedures used in more recent work (*Lanzante et al.*, 2003; *Thorne et al.*, 2005).

591   4.4   *Pattern-based "fingerprint" detection studies*

592

593   Fingerprint detection studies rely on *patterns* of temperature change (Box 5.5). The patterns are

594   typically either latitude-longitude "maps" (*e.g.*, for T_4, T_2, T_S, *etc.*) or latitude-height cross-sections

595   through the atmosphere.[32] The basic premise in fingerprinting is that different climate forcings have

596   different characteristic patterns of temperature response ("fingerprints"), particularly in the free

597   atmosphere (Chapter 1, Figure 1.3; *Hansen et al.*, 1997, 2002, 2005a; *Bengtsson et al.*, 1999;

598   *Santer et al.*, 1996; *Tett et al.*, 1996).

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[32]In constructing these cross-sections, the temperature changes are generally averaged along individual bands of latitude. Zonal averages are then displayed at individual pressure levels, starting at the lowest model or radiosonde level and ending at the top of the model atmosphere or highest reported radiosonde level (see, *e.g.*, Chapter 1, Figure 3).

612	Box 5.5: Fingerprint Studies
613	
614	Detection and attribution ("D&A") studies attempt to represent an observed climate dataset as a
615	linear combination of the climate signals ("fingerprints") arising from different forcing factors and
616	the noise of natural internal climate variability (Section 4.4). A number of different fingerprint
617	methods have been applied to the problem of identifying human-induced climate change. Initial
618	studies used relatively simple pattern correlation methods (*Barnett and Schlesinger, 1987; Santer*
619	*et al.*, 1996; *Tett et al.*, 1996). Later work involved variants of the "optimal detection" approach
620	suggested by *Hasselmann* (1979, 1993, 1997).[33] These are essentially regression-based techniques
621	that seek to estimate the strength of a given fingerprint pattern in observational data (*i.e.*, how
622	much a given fingerprint pattern has to be scaled up or down in order to best match observations).
623	For example, if the regression coefficient for a GHG-induced T_S fingerprint is significantly
624	different from zero, GHG effects are deemed to be "detected" in observed surface temperature
625	records. Attribution tests address the question of whether these regression coefficients are also
626	consistent with unity – in other words, whether the size of the model fingerprint is consistent with
627	its amplitude in observations (*e.g.*, *Allen and Tett*, 1999; *Mitchell et al.*, 2001).
628	
629	There are two broad classes of regression-based D&A methods (*Mitchell et al.*, 2001). One class
630	assumes that although the fingerprint's amplitude changes over time, its spatial pattern does not
631	(*Hegerl et al.*, 1996, 1997; *Santer et al.*, 2003a,b, 2004). The second class explicitly considers both
632	the spatial structure <u>and</u> time evolution of the fingerprint (*Allen and Tett*, 1999; *Allen et al.*, 2005;
633	*Stott and Tett*, 1998; *Stott et al.*, 2000; *Tett et al.*, 1999, 2002; *Barnett et al.*, 2001, 2005). This is
634	particularly useful if the time evolution of the fingerprint contains specific information (such as a
635	periodic 11-year solar cycle) that may help to distinguish it from natural internal climate variability
636	(*North et al.*, 1995; *North and Stevens*, 1998).
637	A number of choices must be made in applying D&A methods to real-world problems. One of the
638	most important decisions relates to "reduction of dimensionality". D&A methods require some
639	knowledge of the correlation structure of natural climate variability.[34] This structure is difficult to
640	estimate reliably, even from long model control runs, because the number of time samples available
641	to estimate correlation behavior is typically much smaller than the number of spatial points in the
642	field. In practice, the total amount of spatial information (the "dimensionality") must be reduced.
643	This is often done by using a mathematical tool (Empirical Orthogonal Functions) to reduce a
644	complex space-time dataset to a very small number of spatial patterns ("EOFs") that capture most
645	of the information content of the dataset.[35] Different analysts use different procedures to determine
646	the number of patterns to retain. Further decisions relate to the choice of data used for estimating

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[33]*Hasselmann* (1979) noted that the engineering field had extensive familiarity with the problem of identifying coherent signals embedded in noisy data, and that many of the techniques routinely used in signal processing were transferable to the problem of detecting a human-induced climate change signal.

[34]The relationship between variability at different points in a spatial field.

[35]The number of patterns retained is often referred to as the "truncation dimension". How the truncation dimension should be determined is a key decision in optimal detection studies (*Hegerl et al.*, 1996; *Allen and Tett*, 1999).

647 fingerprint and noise, the number of fingerprints considered, the selection of observational data, the
648 treatment of missing data, *etc*.[36]
649

650 D&A methods have some limitations. They do not work well if fingerprints are highly uncertain, or
651 if the fingerprints arising from two different forcings are similar.[37] They make at least two
652 important assumptions: that model-based estimates of natural climate variability are a reliable
653 representation of "real-world" variability, and that the sum of climate responses to individual
654 forcing mechanisms is equivalent to the response obtained when these factors are varied in concert.
655 Testing the validity of both assumptions remains an important research activity (*Allen and Tett*,
656 1999; *Santer et al*., 2003a; *Gillett et al*., 2004a).

657

658 Most analysts rely on a climate model to provide physically-based estimates of each fingerprint's

659 structure, size, and evolution. The model fingerprints are searched for in observational climate

660 records, using rigorous statistical methods to quantify the degree of correspondence with observed

661 patterns of climate change.[38] Fingerprints are also compared with patterns of climate change in

662 model control runs. This helps to determine whether the correspondence between the fingerprint

663 and observations is truly significant, or could arise through internal variability alone (Box 5.5).

664 Model errors in internal variability[39] can bias detection results, although most detection work tries

665 to guard against this possibility by performing "consistency checks" on modeled and observed

666 variability (*Allen and Tett*, 1999), and by using variability estimates from multiple models (*Hegerl*

667 *et al*., 1997; *Santer et al*., 2003a,b).

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[36]Another important choice determines whether global-mean changes are included or removed from the detection
analysis. Removal of global means focuses attention on smaller-scale features of modeled and observed climate-change
patterns, and provides a more stringent test of model performance.

[37]This problem is known as "degeneracy". Formal tests of fingerprint degeneracy are sometimes applied (*e.g*., *Tett et
al*., 2002).

[38]The fingerprint can be either the response to an individual forcing or a combination of forcings. One strategy, for
example, is to search for the climate fingerprint in response to combined changes in a suite of different human-caused
forcings.

[39]For example, current CGCMs fail to simulate the stratospheric temperature variability associated with the QBO or
with solar-induced changes in stratospheric ozone (*Haigh*, 1994). Such errors may help to explain why one particular
CGCM underestimated observed temperature variability in the equatorial stratosphere (*Gillett*, 2000). In the same
model, however, the variability of temperatures and lapse rates in the tropical troposphere was in reasonable agreement
with observations.

668

669   The application of fingerprint methods involves a variety of decisions, which introduce uncertainty

670   in detection results (Box 5.5). Our confidence in fingerprint detection results is increased if they are

671   shown to be consistent across a range of plausible choices of statistical method, processing options,

672   and model and observational datasets.

673

674   *Surface temperature changes*

675

676   Most fingerprint detection studies have focused on surface temperature changes. The common

677   denominator in this work is that the model fingerprints resulting from forcing by well-mixed GHGs

678   and sulfate aerosols[40] are statistically identifiable in observed surface temperature records (*Hegerl*

679   *et al.*, 1996, 1997; *North and Stevens*, 1998; *Tett et al.*, 1999, 2002; *Stott et al.*, 2000). These results

680   are robust to a wide range of uncertainties (*Allen et al.*, 2005).[41] In summarizing this body of work,

681   the IPCC concluded that "There is new and stronger evidence that most of the warming observed

682   over the last 50 years is attributable to human activities" (*Houghton et al.*, 2001, page 4). The

683   causes of surface temperature change over the first half of the 20th Century are more ambiguous

684   (*IDAG*, 2005).

685

686   Most of the early fingerprint detection work dealt with global-scale patterns of surface temperature

687   change. The positive detection results obtained for "GHG-only" fingerprints were driven by model-

---

[40]Most of this work considers only the direct scattering effects of sulfate aerosols on incoming sunlight, and not indirect aerosol effects on clouds.

[41]For example, to uncertainties in the applied greenhouse-gas and sulfate aerosol forcings, the model responses to those forcings, and model-based estimates of natural internal climate variability.

688   data pattern similarities at very large spatial scales (*e.g.*, at the scale of individual hemispheres, or

689   land-versus-ocean behavior). Fingerprint detection of GHG effects becomes more challenging at

690   continental or sub-continental scales.[42] It is at these smaller scales that spatially heterogeneous

691   forcings, such as those arising from changes in aerosol loadings and land use patterns, may have

692   large impacts on regional climate (see Box 5.3 and 5.4). This is illustrated by the work of *Stott and*

693   *Tett* (1998), who found that a combined GHG and sulfate aerosol signal was identifiable at smaller

694   spatial scales than a "GHG-only" signal.

695

696   Recently, *Stott* (2003) and *Zwiers and Zhang* (2003) have claimed positive identification of the

697   continental- or even sub-continental features of combined GHG and sulfate aerosol fingerprints in

698   observed surface temperature records.[43] Using a variant of "classical" fingerprint methods,[44] *Min et*

699   *al*. (2005) identified a GHG signal in observed records of surface temperature change over East

700   Asia. *Karoly and Wu* (2005) suggest that GHG and sulfate aerosol effects are identifiable at even

701   smaller spatial scales ("of order 500 km in many regions of the globe"). These preliminary

702   investigations raise the intriguing possibility of formal detection of anthropogenic effects at

703   regional scales that are of direct relevance to policymakers.

704

705   *Changes in latitude/longitude patterns of atmospheric temperature or lapse rate*

706

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[42]This is partly due to the fact that natural climate noise is larger (and models are less skillful) on smaller spatial scales.

[43]Another relevant "sub-global" detection study is that by *Karoly et al*. (2003). This showed that observed trends in a variety of area-averaged "indices" of North American climate (*e.g*., surface temperature, daily temperature range, and the amplitude of the seasonal cycle) were consistent with model-predicted trends in response to anthropogenic forcing, but were inconsistent with model estimates of natural climate variability.

[44]Involving Bayesian statistics.

707   Fingerprint methods have also been applied to spatial "maps" of changes in layer-averaged

708   atmospheric temperatures (*Santer et al.*, 2003b; *Thorne et al.*, 2003) and lapse rate (*Thorne et al.*,

709   2003). The study by *Santer et al.* (2003b) compared modeled and observed changes in T_2 and T_4.

710   Model fingerprints were estimated from 20CEN experiments performed with PCM (see Table 5.1),

711   while observations were taken from two different satellite datasets (UAH and RSS; see *Christy et*

712   *al.*, 2003, and *Mears et al.,* 2003). The aim of this work was to assess the sensitivity of detection

713   results to structural uncertainties in observed MSU data.

714

715   For the T_4 layer, the model fingerprint of combined human and natural effects was consistently

716   detectable in both satellite datasets. In contrast, PCM's T_2 fingerprint was identifiable in RSS data

717   (which show net warming over the satellite era), but not in UAH data (which show little overall

718   change in T_2; see Chapter 3). Encouragingly, once the global-mean differences between RSS and

719   UAH data were removed, the PCM T_2 fingerprint was detectable in *both* observed datasets. This

720   suggests that the structural uncertainties in RSS and UAH T_2 data are most prominent at the global-

721   mean level, and that this global-mean difference masks underlying similarities in smaller-scale

722   pattern structure (Chapter 4; *Santer et al.*, 2004).

723

724   *Thorne et al.* (2003) applied a "space-time" fingerprint method to six individual climate variables.

725   These variables contained information on patterns[45] of temperature change at the surface, in broad

726   atmospheric layers (the upper and lower troposphere), and in the lapse rates between these layers.[46]

---

[45]The "patterns" are in the form of temperature averages calculated over large areas rather than temperatures on a regular latitude/longitude grid.

[46]*Thorne et al.* calculated the lapse rate changes between the surface and lower troposphere, the surface and upper troposphere, and the lower and upper troposphere.

727   *Thorne et al.* explicitly considered uncertainties in the searched-for fingerprints, the observed

728   radiosonde data[47], and in various data processing/fingerprinting options. They also assessed the

729   detectability of fingerprints arising from multiple forcings.[48] The "bottom-line" conclusion of

730   *Thorne et al.* is that two human-caused fingerprints – one arising from changes in well-mixed

731   GHGs alone, and the other due to combined GHG and sulfate aerosol effects – were robustly

732   identifiable in the observed surface, lower tropospheric, and upper tropospheric temperatures.

733   Evidence for the existence of a detectable volcanic signal was more equivocal. Volcanic and

734   human-caused fingerprints were not consistently identifiable in observed patterns of lapse rate

735   change.[49]

736

737   *Changes in latitude/height profiles of atmospheric temperature*

738

739   Initial detection work with zonal-mean profiles of atmospheric temperature change used pattern

740   correlations to compare model fingerprints with radiosonde data (*Karoly et al.*, 1994; *Santer et al.*,

741   1996; *Tett et al.*, 1996; *Folland et al.*, 1998; *Sexton et al.*, 2001). These early investigations found

742   that model fingerprints of the stratospheric cooling and tropospheric warming in response to

743   increases in atmospheric CO_2 were identifiable in observations (Chapter 1, Figure 1.3a). The

744   pattern similarity between modeled and observed changes generally increased over the period of

745   the radiosonde record.

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[47]The model fingerprint was estimated from 20CEN runs performed with two different versions of the Hadley Centre CGCM (HadCM2 and HadCM3). Observational data were taken from two early compilations of the Hadley Centre radiosonde data (HadRT2.1 and HadRT2.1s).

[48]Well-mixed greenhouse gases, the direct effects of sulfate aerosols, combined greenhouse-gas and sulfate aerosol effects, volcanic aerosols, and solar irradiance changes.

[49]The failure to detect volcanic signals is probably due to the coarse time resolution of the input data (five-year averages) and the masking effects of ENSO variability in the radiosonde observations. Note that the two models employed in this work yielded different estimates of the size of the natural and human-caused fingerprints.

746

747   The inclusion of other human-induced forcings in 20CEN experiments – particularly the effects of

748   stratospheric ozone depletion and sulfate aerosols – tended to improve agreement with observations

749   (*Santer et al.*, 1996a; *Tett et al.*, 1996; *Sexton et al.*, 2001). The addition of ozone depletion cooled

750   the lower stratosphere and upper troposphere. This brought the height of the "transition level"

751   between stratospheric cooling and tropospheric warming lower down in the atmosphere, and in

752   better accord with observations (Chapter 1, Figure 1.3F). It also improved the agreement between

753   simulated and observed patterns of T_4 (*Ramaswamy et al.*, 1996), and decreased the size of the

754   "warming maximum" in the upper tropical troposphere, a prominent feature of CO_2-only

755   experiments (compare Figures 1.3A and 1.3F in Chapter 1).

756

757   Early work on the direct scattering effects of sulfate aerosols suggested that this forcing was

758   generally stronger in the Northern Hemisphere (NH) than in the Southern Hemisphere (SH), due to

759   the larger emissions of sulfur dioxide in industrialized regions of the NH. This asymmetry in the

760   distribution of anthropogenic sulfur dioxide sources should yield greater aerosol-induced

761   tropospheric cooling in the NH (*Santer et al.*, 1996a,b). Other forcings can lead to different

762   hemispheric temperature responses. Increases in atmospheric CO_2, for example, tend to warm land

763   more rapidly than ocean (Chapter 1). Since there is more land in the NH than in the SH, the

764   expected signal due to CO_2 increases is greater *warming* in the NH than in the SH. Because the

765   relative importance of CO_2 and sulfate aerosol forcings evolves in a complex way over time (*Tett et*

766   *al.*, 2002; *Hansen et al.*, 2002),[50] the "imprints" of these two forcings on NH and SH temperatures

767   must also vary with time (*Santer et al.*, 1996b; Stott *et al.*, 2005).

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[50]See, for example, Figure 1a in *Tett et al.* (2002) and Figure 8b in *Hansen et al.* (2002).

769   Initial attempts to detect sulfate aerosol effects on atmospheric temperatures did not account for

770   such slow changes in the hemispheric-scale features of the aerosol fingerprint. They searched for a

771   *time-invariant* fingerprint pattern in observed radiosonde data (*Santer et al.*, 1996a). This yielded

772   periods of agreement and periods of disagreement between the (fixed) aerosol fingerprint and the

773   time-varying effect of aerosols on atmospheric temperatures. Some have interpreted the periods of

774   disagreement as 'evidence of absence' of a sulfate aerosol signal (*Michaels and Knappenberger*,

775   1996). However, subsequent studies (see below) illustrate that such behavior is expected if one uses

776   a fixed sulfate aerosol fingerprint, and that it is important for detection studies to account for large

777   temporal changes in the fingerprint.

778

779   "Space-time" optimal detection schemes explicitly account for time variations in the signal pattern

780   and in observational data (Box 5.5). Results from recent "space-time" detection studies support

781   previous claims of an identifiable sulfate aerosol effect on surface temperature (*Stott et al.*, 2005)

782   and on zonal-mean profiles of atmospheric temperature (*Allen and Tett*, 1999; *Forest et al.*, 2001,

783   2002; *Thorne et al.*, 2002; *Tett et al.*, 2002; *Jones et al.*, 2003). This work also illustrates that the

784   identification of human effects on atmospheric temperatures can be achieved using tropospheric

785   temperatures alone (*Thorne et al.*, 2002). Positive detection results are not solely driven by the

786   inclusion of strong stratospheric cooling in the vertical pattern of temperature change (as has been

787   claimed by *Weber*, 1996).

788

789   In summary, fingerprint detection studies provide strong and consistent evidence that human-

790   induced changes in greenhouse gases and sulfate aerosols are identifiable in radiosonde records of

791   free atmospheric temperature change. The fingerprint evidence is much more equivocal in the case

792   of solar and volcanic signals in the troposphere. These natural signals have been detected in some

793   studies (*Jones et al.*, 2003) but not in others (*Tett et al.* 2002), and their identification appears to be

794   more sensitive to specific processing choices that are made in applying fingerprint methods (*Leroy*,

795   1998; *Thorne et al.*, 2002, 2003).

796

797   5   ***New Comparisons of Modeled and Observed Temperature Changes***

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799   In this section, we evaluate selected results from recently-completed CGCM 20CEN experiments

800   that have been performed in support of the IPCC Fourth Assessment Report (AR4). The runs

801   analyzed here were performed with 19 different models, and involve modeling groups in 10

802   different countries (Table 5.1). They use new model versions, and incorporate historical changes in

803   many (but not all) of the natural and human forcings that are thought to have influenced

804   atmospheric temperatures over the past 50 years[51] (Table 5.2). These new experiments provide our

805   current best estimates of the expected climate change due to combined human and natural effects.

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809   The new 20CEN runs constitute an "ensemble of opportunity" (*Allen and Stainforth*, 2002). The

810   selection and application of natural and anthropogenic forcings was not coordinated across

811   modeling groups.[52] For example, only seven of the 19 modeling groups applied time-varying

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[51]This was not the case in previous model intercomparison exercises, such as AMIP (*Gates et al.*, 1999) and CMIP2 (*Meehl et al.*, 2000).

[52]In practice, experimental coordination is very difficult across a range of models of varying complexity and sophistication. Aerosols are a case in point. Some modeling groups that contributed 20CEN simulations to the IPCC AR4 do not have the technical capability to explicitly include aerosols, and instead attempt to represent their net radiative effects by adjusting the surface albedo.

812   changes in LULC (Table 5.2). Groups that included LULC effects did not always use the same

813   observational dataset for specifying this forcing, or apply it in the same way (Table 5.3). Only six

814   models included some representation of the indirect effects of anthropogenic aerosols, which are

815   thought to have had a net cooling influence on surface temperatures through their effects on cloud

816   properties (*Ramaswamy et al*., 2001b).

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829   Table 5.3: Forcings used in 20CEN experiments performed with the PCM, CCSM3.0, GFDL CM2.1, and GISS-EH
830   models. Grey shading denotes a forcing that was included in the experimental design. Shading indicates a forcing that
831   was not incorporated or that did not vary over the course of the experiment.

	PCM	CCSM3.0	GFDL CM2.1	GISS-EH
Well-mixed greenhouse gases	IPCC Third Assessment Report.	IPCC Third Assessment Report.	IPCC Third Assessment Report and *World Meteorological Organization* (2003).	CH_4, N_2O and CFC spatial distributions are fit to *Minschwaner et al.* (1998).
Sulfate aerosols (direct effects)	Spatial patterns of sulfur dioxide [SO_2] emissions prescribed over seasonal cycle. Year-to-year changes scaled by estimates of historical changes in SO_2 emissions.[1]	Sulfur cycle model using time and space-varying SO_2 emissions (*Smith et al.*, 2001, 2005).[2]	Computed from an atmospheric chemistry transport model.[3]	Based on simulations of *Koch et al.* (1999) and *Koch* (2001).[4]
Sulfate aerosols (indirect effects)	Not included.	Not included.	Not included.	Parameterization of aerosol indirect effects on cloud albedo and cloud cover.[4]
Stratospheric ozone	Assumed to be constant up to 1970. After 1970 prescribed from a NOAA dataset.[1]	Assumed to be constant up to 1970. After 1970 prescribed from a NOAA dataset.[2]	Specified using data from *Randel and Wu* (1999).	Specified using data from *Randel and Wu* (1999).[4]
Tropospheric ozone	Computed from an atmospheric chemistry transport model. Held constant after 1990.[1]	Computed from an atmospheric chemistry transport model. Held constant after 1990.[2]	Computed from an atmospheric chemistry transport model.[3]	Computed from an atmospheric chemistry transport model (*Shindell et al.*, 2003).[4]
Black carbon aerosols	Not included.	Present-day estimate of distribution and amount of black carbon, scaled by population changes over 20th Century.[2]	Computed from an atmospheric chemistry transport model.[3]	Based on simulations of *Koch et al.* (1999) and *Koch* (2001).[4]
Organic aerosols	Not included.	Not included.	Computed from an atmospheric chemistry transport model.[3]	Based on simulations of *Koch et al.* (1999) and *Koch* (2001).[4]
Sea salt	Not included.	Distributions held fixed in 20th Century at year 2000 values.[2]	Distributions held fixed at 1990 values.	
Dust	Not included.	Distributions held fixed in 20th Century at year 2000 values.[2]	Distributions held fixed at 1990 values.	
Land use change	Distributions held fixed at present-day values.	Distributions held fixed at present-day values.	*Hurtt et al.* (2006) global land use reconstruction history. Includes effect on surface albedo, surface roughness, stomatal resistance, and effective water capacity.	Uses *Ramankutty and Foley* (1999) and *Klein Goldewijk* (2001) time-dependent datasets. Effects on albedo and evapotranspiration included, but no irrigation effects.[4]

Volcanic stratospheric aerosols	*Ammann et al.* (2003).	*Ammann et al.* (2003).	"Blend" between *Sato et al.* (1993) and *Ramachandran et al. (2000).*	Update of *Sato et al.* (1993).
Solar irradiance	*Hoyt and Schatten* (1993).	*Lean et al.* (1995).	*Lean et al.* (1995).	Uses solar spectral changes of *Lean* (2000).

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833   [1]See *Dai et al.* (2001) for further details.
834   [2]See *Meehl et al.* (2005) for further details.
835   [3]The chemistry transport model (MOZART; see *Horowitz et al.*, 2003; *Tie et al.*, 2005) was driven by meteorology from the Middle
836   Atmosphere version of the Community Climate Model ("MACCM"; version 3). "1990" weather from MACCM3 was used for all
837   years between 1860 and 2000.
838   [4]See *Hansen et al.* (2005a) for further det

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842   One important implication of Table 5.3 is that model-to-model differences in the applied forcings

843   are intertwined with model-to-model differences in the climate *responses* to those forcings. This

844   makes it more difficult to isolate systematic errors that are common to a number of models, or to

845   identify problems with a specific forcing dataset. Note, however, that the lack of a coordinated

846   experimental design is also an advantage, since the "ensemble of opportunity" spans a wide range

847   of uncertainty in current estimates of climate forcings.

848

849   In addition to model forcing and response uncertainty, the 20CEN ensemble also encompasses

850   uncertainties arising from inherently unpredictable climate variability. Roughly half of the

851   modeling groups that submitted 20CEN data performed multiple realizations of their historical

852   forcing experiment (see Section 2 and Table 5.1). For example, the five-member ensemble of

853   CCSM3.0 20CEN runs contains an underlying signal (which one might define as the ensemble-

854   average climate response to the forcings varied in CCSM3.0) plus five different sequences of

855   climate noise. Such multi-member ensembles provide valuable information on the relative sizes of

856   signal and noise. In all, a total of 49 20CEN realizations were examined here.

857

858   The following Section presents preliminary results from analyses of these 20CEN runs and the new

859   observational datasets described in Chapters 2-4. Our primary focus is on the tropics, since

860   previous work by *Gaffen et al.* (2000) and *Hegerl and Wallace* (2002) suggests that this is where

861   climate models have significant problems in simulating observed lapse rates changes. We also

862   discuss comparisons of global-mean changes in atmospheric temperatures and lapse rates. We do

863   not discount the importance of comparing models and data at much smaller scales (particularly in

864   view of the incorporation of regional-scale forcing changes in many of the runs analyzed here), but

865   comprehensive regional-scale comparisons were not feasible given the limited time available for

866   completion of this report.

867

868   In order to facilitate "like with like" comparisons between modeled and observed atmospheric

869   temperature changes, we calculate synthetic MSU T_4, T_2, and T_{2LT} from the model 20CEN results

870   (see Chapter, Box 2). Both observed and synthetic MSU T_2 data include a contribution from the

871   cooling stratosphere (*Fu et al.*, 2004a,b), and hence complicate the interpretation of slow changes

872   in T_2. To provide a less ambiguous measure of "bulk" tropospheric temperature changes, we use

873   the statistical approach of *Fu et al.* (2004a, 2005) to remove stratospheric influences, thereby

874   obtaining T^*_G and T^*_T in addition to T_{2LT}.[53] As a simple measure of lapse-rate changes, we

875   consider temperature differences between the surface and three different atmospheric layers (T_{2LT},

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[53]There is still some debate over the reliability of T^*_G trends estimated with the *Fu et al.* (2004a) statistical approach (*Tett and Thorne*, 2004, *Gillett et al.*, 2004; *Kiehl et al.*, 2005; *Fu et al.*, 2004b; Chapter 4). T^*_T is derived mathematically (from the overlap between the T_4 and T_2 weighting functions) rather than statistically, and is now generally accepted as a reasonable measure of temperature change in the tropical troposphere.

876   T*$_G$, and T*$_T$). Each of these layers samples slightly different portions of the troposphere (Figure

877   2.2).

878

879   The trend comparisons shown in Sections 5.1 and 5.2 do not involve any formal statistical

880   significance tests (see Statistical Appendix). While such tests are entirely appropriate for

881   comparisons of <u>individual</u> model and observational trends,[54] they are less relevant here, where we

882   compare a 49-member ensemble of model trends with a relatively small number of observationally-

883   based estimates. The model ensemble encapsulates uncertainties in climate forcings and model

884   responses, as well as the effects of climate noise on trends. The observational range characterizes

885   current structural uncertainties in historical changes. We simply assess whether the simulated trend

886   distributions do or do not overlap with these observations. Our goal here is to determine where

887   model results are qualitatively consistent with observations, and where serious inconsistencies are

888   likely to exist. This does not obviate the need for the more rigorous statistical comparisons

889   described in Box 5.5, which should be a high priority (see Recommendations).

890

891   5.1   *Global-Mean Temperature and Lapse-Rate Trends*

892

893   In all but two of the 49 20CEN realizations, the global-mean temperature of the lower stratosphere

894   experiences a net cooling over 1979 to 1999 (Figures 5.2A, 5.3A).[55] The model average T$_4$ trend is

895   –0.25°C/decade (Table 5.4A). Most of this cooling is due to the combined effects of stratospheric

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[54]For example, such tests have been performed by *Santer et al*. (2003b) in comparisons between observed MSU trends (in RSS and UAH) and synthetic MSU trends in four PCM 20CEN realizations.

[55]In the following, all inter-model and model-data comparisons are over January 1979 to December 1999. This is the longest period of overlap (at least during the satellite era) between the model experiments (which generally end in 1999) and the satellite data (which start in 1979).

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896 ozone depletion and increases in well-mixed GHGs (*Ramaswamy et al.*, 2001a,b), with the former

897 the dominant influence on T_4 changes over the satellite era (*Ramaswamy et al.*, 1996; *Santer et al.*,

898 2003a). The model average cooling is larger (–0.35°C/decade) and closer to the satellite-based

899 estimates if it is calculated from the subset of 20CEN realizations that include forcing by ozone

900 depletion. The range of model T_4 trends encompasses the trends derived from satellites, but not the

901 larger trends estimated from radiosondes. The most likely explanation for this discrepancy is a

902 residual cooling trend in the radiosonde data (Chapter 4).[56] The neglect of stratospheric water vapor

903 increases in most of the 20CEN runs considered here (*Shine et al.*, 2003) may be another

904 contributory factor.[57]

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[56]Recent work suggests that this residual trend is largest in the lower stratosphere and upper troposphere, and is largely related to temporal changes in the solar heating of the temperature sensors carried by radiosondes (and failure to properly correct for this effect; see *Sherwood et al.*, 2005; *Randel and Wu*, 2005).

[57]Recent stratospheric water vapor increases are thought to be partly due to the oxidation of methane, and are expected to have a net cooling effect on T_4. To our knowledge, CH_4-induced stratospheric water vapor increases were explicitly incorporated in only two of the 19 models considered here (GISS-EH and GISS-ER; *Hansen et al.*, 2005a).
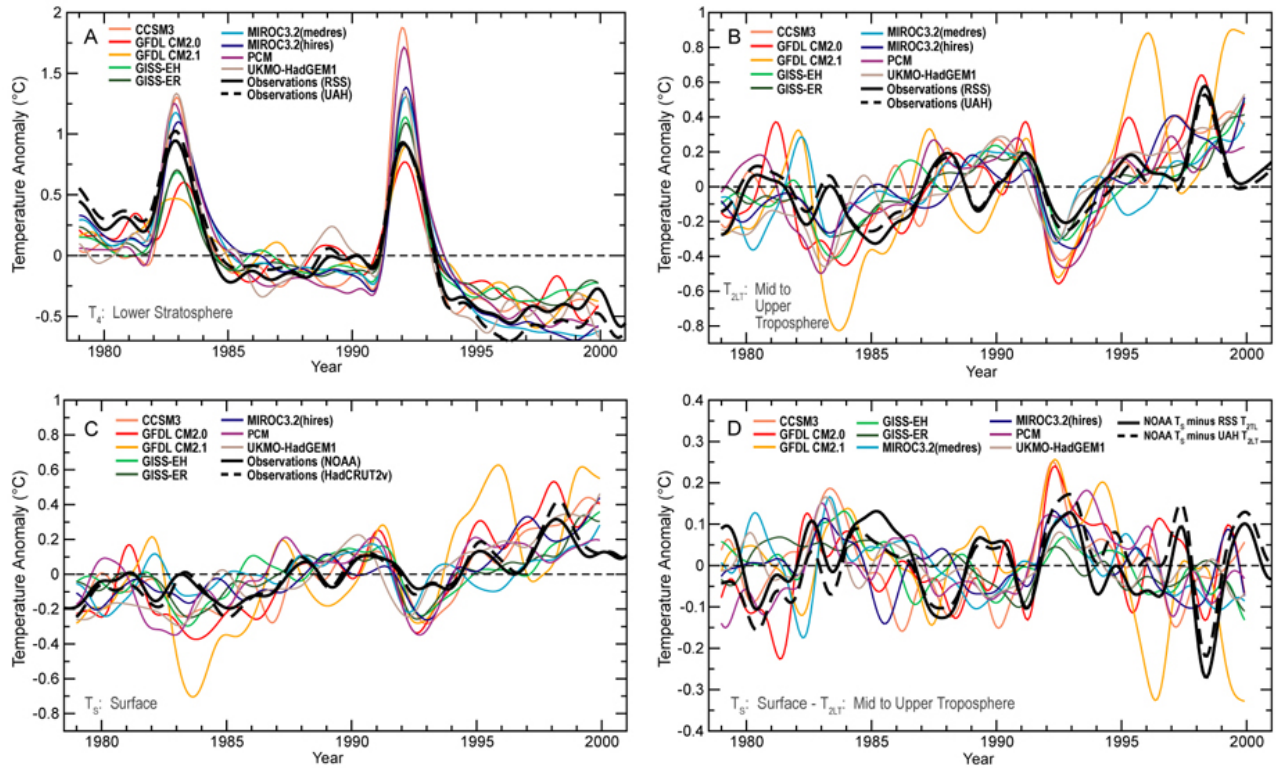
914

915 Figure 5.2A: Modeled and observed changes in global-mean monthly-mean lower stratospheric temperature (T_4). A
916 simple weighting function approach (Box 2.2) was used to calculate a "synthetic" T_4 (equivalent to the MSU T_4
917 monitored by satellites) from model temperature data. Synthetic T_4 results are from "20CEN" experiments performed
918 with nine different models (see Table 5.1). These models were chosen because they satisfy certain minimum
919 requirements in terms of the forcings applied in the 20CEN run: all nine were driven by changes in well-mixed GHGs,
920 sulfate aerosol direct effects, tropospheric and stratospheric ozone, volcanic aerosols, and solar irradiance (in addition
921 to other forcings; see Table 5.2). Observed satellite-based estimates of T_4 changes were obtained from both RSS and
922 UAH (see Chapter 3). All T_4 changes are expressed as departures from a 1979 to 1999 reference period average, and
923 were smoothed with the same filter. To make it easier to compare the variability of T_4 in models with different
924 ensemble sizes (see Table 5.1), only the first 20CEN realization is plotted from each model. This also facilitates
925 comparisons of modeled and observed variability.
926

927 Figure 5.2B: As for Figure 5.2A, but for time series of global-mean, monthly-mean lower tropospheric temperature
928 anomalies (T_{2LT}).
929

930 Figure 5.2C: As for Figure 5.2A, but for time series of global-mean, monthly-mean surface temperature anomalies (T_S).
931

932 Figure 5.2D: As for Figure 5.2A, but for time series of global-mean, monthly-mean temperature differences between
933 the surface and T_{2LT}.
934

Modeled and Observed Global-Mean Temperature Trends
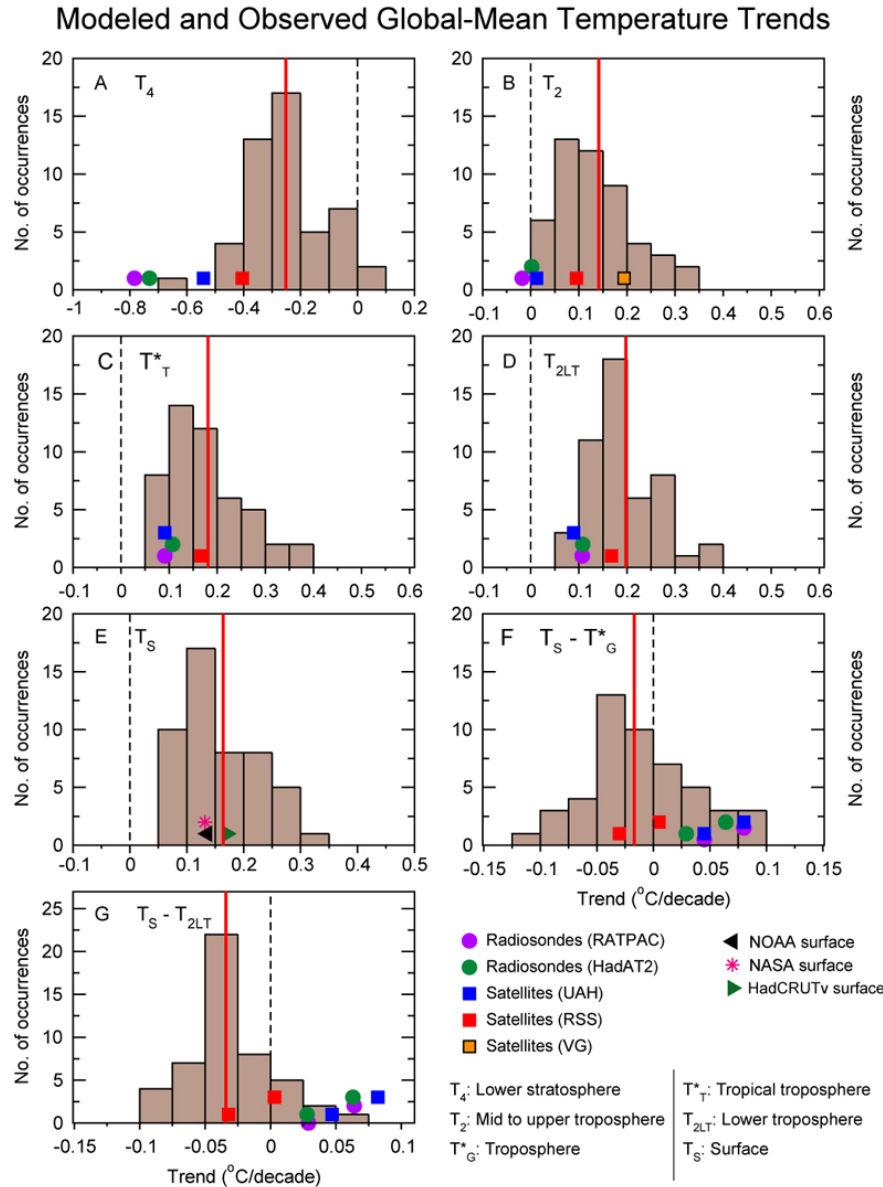
935

936   Figure 5.3: Modeled and observed trends in time series of global-mean T_4 (panel A), T_2 (panel B),
937   T^*_G (panel C), T_{2LT} (panel D), T_S (panel E), T_S minus T^*_G (panel F), and T_S minus T_{2LT}. All trends
938   were calculated using monthly-mean anomaly data. The analysis period is 1979 to 1999. Model
939   results are displayed in the form of histograms. Each histogram is based on results from 49
940   individual realizations of the 20CEN experiment, performed with 19 different models (Table 5.1).
941   The applied forcings are listed in Table 5.2. The vertical red line in each panel is the mean of the
942   model trends, calculated with a sample size of $n = 19$ (see Table 5.4A). Observed trends are
943   estimated from two radiosonde and three satellite datasets (T_2), two radiosonde and two satellite
944   datasets (T_4, T^*_G and T_{2LT}), and three different surface datasets (T_S) (see Chapter 3). The bottom
945   "rows" of the observed difference trends in panels F and G were calculated with NOAA T_S data.
946   The top "rows" of observed results in F and G were computed with HadCRUT2v T_S data. The
947   vertical offsetting of observed results in these panels (and also in panels B-E) is purely for the

948   purpose of simplifying the visual display – observed trends bear no relation to the y-axis scale. To
949   simplify the display, the Figure does not show the statistical uncertainties arising from the fitting of
950   linear trends to noisy data. GISS T_S trends (not shown) are very close to those estimated with
951   NOAA T_S data (see Chapter 3).
952
953

954   Table 5.4A: Summary statistics for global-mean temperature trends calculated from 49 different realizations of 20CEN
955   experiments performed with 19 different coupled models. Results are for four different atmospheric layers (T_4, T_2, $T*_G$,
956   and T_{2LT}), the surface (T_S), and differences between the surface and the troposphere (T_S minus $T*_G$ and T_S minus T_{2LT}).
957   All trends were calculated over the 252-month period from January 1979 to December 1999 using global-mean
958   monthly-mean anomaly data. Results are in °C/decade. The values in the "Mean" column correspond to the locations of
959   the red lines in the seven panels of Figure 5.3A. For each layer, means, medians and standard deviations were
960   calculated from a sample size of *n* = 19, *i.e.*, from ensemble means (if available) and individual realizations (if
961   ensembles were not performed). This avoids placing too much weight on results from a single model with a large
962   number of realizations. Maximum and minimum values were calculated from all available realizations (*i.e.*, from a
963   sample size of *n* = 49).
964   Superimposed on the overall cooling of T_4 are the large stratospheric warming signals in response

Layer	Mean	Median	Std. Dev. (1σ)	Minimum	Maximum
T_4	-0.252	-0.281	0.194	-0.695	0.079
T_2	0.142	0.122	0.079	0.015	0.348
$T*_G$	0.181	0.167	0.077	0.052	0.375
T_{2LT}	0.198	0.186	0.070	0.058	0.394
T_S	0.164	0.156	0.062	0.052	0.333
$T_S - T*_G$	-0.017	-0.017	0.046	-0.110	0.083
$T_S - T_{2LT}$	-0.034	-0.031	0.030	-0.099	0.052

965

966   to the eruptions of El Chichón (in April 1982) and Pinatubo (in June 1991).[58] Nine of the 19 IPCC

967   models explicitly included volcanic aerosols (Figure 5.2A and Table 5.2).[59] Seven of these nine

968   models overestimate the observed stratospheric warming after Pinatubo. GFDL CM2.1 simulates

969   the Pinatubo response reasonably well, but underestimates the response to El Chichón. Differences

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[58]These warming signals occur because volcanic aerosols absorb both incoming solar radiation and outgoing thermal
radiation (*Ramaswamy et al.*, 2001a).

[59]The documentation for the Russian INM-CM3.0 model claims that volcanic aerosols were incorporated in the 20CEN
run, but does not show evidence of stratospheric warming signatures after massive volcanic eruptions. This suggests
that volcanic cooling effects on surface temperature were implicitly incorporated by changing the surface albedo (a
procedure that would not yield volcanically-induced stratospheric warming signals).

970   in the magnitude of the applied volcanic aerosol forcings must account for some of the inter-model

971   differences in the T_4 warming signals (Table 5.3).[60]

972

973   Over 1979 to 1999, the global-mean troposphere warms in all 49 20CEN simulations considered

974   here (Figures 5.2B, 5.3B-D). The shorter-term cooling signals of the El Chichón and Pinatubo

975   eruptions are superimposed on this gradual warming.[61] Because of the influence of stratospheric

976   cooling on T_2, the model average trend is smaller for this layer than for either T_{2LT} or T^*_G, which

977   are more representative of temperature changes in the bulk of the troposphere (Table 5.4A).[62] All of

978   the satellite- and radiosonde-based trends in T_{2LT} and T^*_G are contained within the spread of model

979   results. This illustrates that there is no fundamental discrepancy between modeled and observed

980   trends in global-mean tropospheric temperature.

981

982   In contrast, the T_2 trends in both radiosonde datasets are either slightly negative or close to zero,

983   and are smaller than all of the model results. This difference is most likely due to contamination

984   from residual stratospheric and upper-tropospheric cooling biases in the radiosonde data (Chapter

985   4; *Sherwood et al.*, 2005; *Randel and Wu*, 2005). The satellite-based T_2 trends are either close to

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[60]More subtle details of the forcing are also relevant to interpretation of inter-model T_4 differences, such as different assumptions regarding the aerosol size distribution, the vertical distribution of the volcanic aerosol relative to the model tropopause, *etc*. Note that observed T_4 changes over the satellite era are not well-described by a simple linear trend, and show evidence of a step-like decline in stratospheric temperatures after the El Chichón and Pinatubo eruptions (*Pawson et al.*, 1998; *Seidel and Lanzante*, 2004). Model-model differences in the applied ozone forcings and solar forcings may help to explain why the GFDL, GISS, and HadGEM1 models appear to reproduce some of this step-like behavior, particularly after El Chichón, while T_4 decreases in PCM are much more linear (*Dameris et al.*, 2005; *Ramaswamy et al.*, 2006).

[61]Because of differences in the timing of modeled and observed ENSO events (Section 5.2), the tropospheric and surface cooling caused by El Chichón is more noticeable in all models than in observations (where it was partially masked by the large 1982/83 El Niño; Figures 5.2B,C).

[62]Because of ozone-induced cooling of the lower stratosphere, the model-average T_2 trend is slightly smaller (0.12°C/decade) and closer to the RSS result if it is estimated from the subset of 20CEN runs that include stratospheric ozone depletion. Subsetting in this way has little impact on the model-average T_{2LT} and T^*_G trends.

986   the model average (RSS and VG) or just within the model range (UAH; Fig. 5.3B). Even without

987   formal statistical tests, it is clear that observational uncertainty is an important factor in assessing

988   the consistency between modeled and observed changes in mid- to upper tropospheric temperature

989   (*Santer et al.*, 2003b).

990

991   Observed T_S trends closely bracket the model average (Figures 5.2C, 5.3E). There is no evidence of

992   a serious inconsistency between modeled and observed surface temperature changes. Structural

993   uncertainties in observed T_S trends are much smaller than for trends in T_4 or tropospheric layer-

994   average temperatures (see Chapter 4).

995

996   The model-simulated ranges of lapse-rate trends also encompass virtually all observational results

997   (Figures 5.3F,G).[63] Closer inspection reveals that the model-average trends in tropospheric lapse

998   rate are slightly negative,[64] indicating larger warming aloft than at the surface. Most combinations

999   of observed T_S, T^*_G, and T_{2LT} datasets yield the converse result, and show <u>smaller</u> warming aloft

1000   than at the surface. As in the case of global-mean T^*_G and T_{2LT} trends, RSS-based lapse-rate trends

1001   are invariably closest to the model average results. Both models and observations show a tendency

1002   towards positive values of T_S minus T_{2LT} for several years after the El Chichón and Pinatubo

1003   eruptions, indicative of larger cooling aloft than at the surface (Figure 5.2D; Section 5.4).

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1005   5.2   *Tropical Temperature and Lapse-Rate Trends*

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[63]Note that the subtraction of temperature variability common to surface and troposphere decreases (by about a factor of two) the large range of model trends in T_S, T^*_G, and T_{2LT} (Table 5.4A).

[64]Values are –0.02°C/decade in the case of T_S minus T^*_G and –0.03°C/decade for T_S minus T_{2LT}.

1007   The previous section examined whether simulated global-mean temperature trends were contained

1008   within current estimates of structural uncertainty in observations. Since ENSO is primarily a

1009   tropical phenomenon, its influence on surface and tropospheric temperature is more pronounced in

1010   the tropics than in global averages. Observations contain only one specific sequence of ENSO

1011   fluctuations from 1979 to present, and only one sequence of ENSO effects on tropical

1012   temperatures. The model 20CEN runs examined here provide many different sequences of ENSO

1013   variability. We therefore expect – and find – that these runs yield a wide range of trends in tropical

1014   surface and tropospheric temperature (Figure 5.4)[65]. It is of interest whether this large model range

1015   encompasses the observed trends.

1016

1017   At the surface, results from the multi-model ensemble include all observational estimates of

1018   tropical temperature trends (Figure 5.4E; Table 5.4B). Observed results are close to the model

1019   average T_S trend of +0.16°C/decade. There is no evidence that the models significantly over- or

1020   underestimate the observed surface warming. In the troposphere, all observational results are still

1021   within the range of possible model solutions, but the majority of model results show tropospheric

1022   warming that is larger than observed (Figures 5.4B-D). As in the case of the global-mean T_4 trends,

1023   the cooling of the tropical stratosphere in both radiosonde datasets is larger than in any of the

---

[65]This would be true even for a hypothetical "perfect" climate model run with "perfect" forcings. This large model range of tropical temperature trends is not solely due to the effects of ENSO and other modes of internal variability. It also arises from uncertainties in the models and forcings (see Box 5.2 and Table 5.2). Note that the trends discussed here are calculated over a relatively short period of time (several decades).

1024   satellite datasets or model results (Figure 5.4A).[66] The UAH and RSS T_4 trends are close to the

1025   model average.[67]

---

[66]This supports recent findings of a residual cooling bias in tropical radiosonde data (*Sherwood et al.*, 2005; *Randel and Wu*, 2005).

[67]The model average is –0.27°C/decade when estimated from the subset of 20CEN runs that include stratospheric ozone depletion.

1026

1027   Figure 5.4: As for Figure 5.3, but for trends in the tropics (20°N-20°S).

1028

1029

1030   Table 5.4B: As for Table 5.4A, but for tropical temperature trends (calculated from spatial averages over 20°N-20°S).

1031

Layer	Mean	Median	Std. Dev. (1σ)	Minimum	Maximum
T_4	-0.188	-0.189	0.152	-0.487	0.127
T_2	0.199	0.188	0.098	-0.013	0.481
T^*_T	0.238	0.213	0.105	0.007	0.558
T_{2LT}	0.215	0.194	0.092	0.006	0.509
T_S	0.155	0.144	0.067	-0.017	0.365
$T_S - T^*_T$	-0.083	-0.079	0.040	-0.194	0.017
$T_S - T_{2LT}$	-0.060	-0.053	0.028	-0.145	0.005

1032

1033   In the model results, trends in the two measures of tropical lapse-rate (T_S minus T_{2LT} and T_S minus

1034 T^*_T) are almost invariably negative, indicating larger warming aloft than at the surface (Figure

1035   5.4F,G). Similar behavior is evident in only one of the four upper-air datasets examined here

1036   (RSS).[68] The RSS trends are just within the range of model solutions.[69] Tropical lapse-rate trends in

1037   both radiosonde datasets and in the UAH satellite data are always positive (larger warming at the

1038   surface than aloft), and lie outside the range of model results.

1039

1040   This comparison suggests that discrepancies between our current best estimates of simulated and

1041   observed lapse-rate changes may be larger and more serious in the tropics than in globally-

1042   averaged data. Large structural uncertainties in the observations (even in the <u>sign</u> of the trend in

---

[68]Note that the VG group do not provide either a stratospheric or lower-tropospheric temperature retrieval, and so could not be included in the comparison of modeled and observed trends in T_S minus T^*_T or T_S minus T_{2LT}.

[69]Three of the four RSS-based results in Figures 5.4F and 5.4G are within two standard deviations of the model average values (see Table 5.4B).

1043   tropical lapse-rate changes) make it difficult to reach more definitive conclusions regarding the

1044   significance and importance of model-data discrepancies (see Section 5.4).

1045

1046   5.3   *Spatial Patterns of Lapse-Rate Trends*

1047

1048   Maps of the trends in lower tropospheric lapse rate help to identify geographical regions where the

1049   model-data discrepancies in Figures 5.4F and 5.4G are most pronounced. We focus first on four

1050   U.S. models: CCSM3.0, PCM, GFDL CM2.1, and GISS-EH (Table 5.3). These show qualitatively

1051   similar patterns of trends in T_S minus T_{2LT} (Figures 5.5A-D). Over most of the tropical ocean, the

1052   simulated warming is larger in the troposphere than at the surface. All models have some tropical

1053   land areas where the surface warms relative to the troposphere. The largest relative warming of the

1054   surface occurs at high latitudes in both hemispheres.
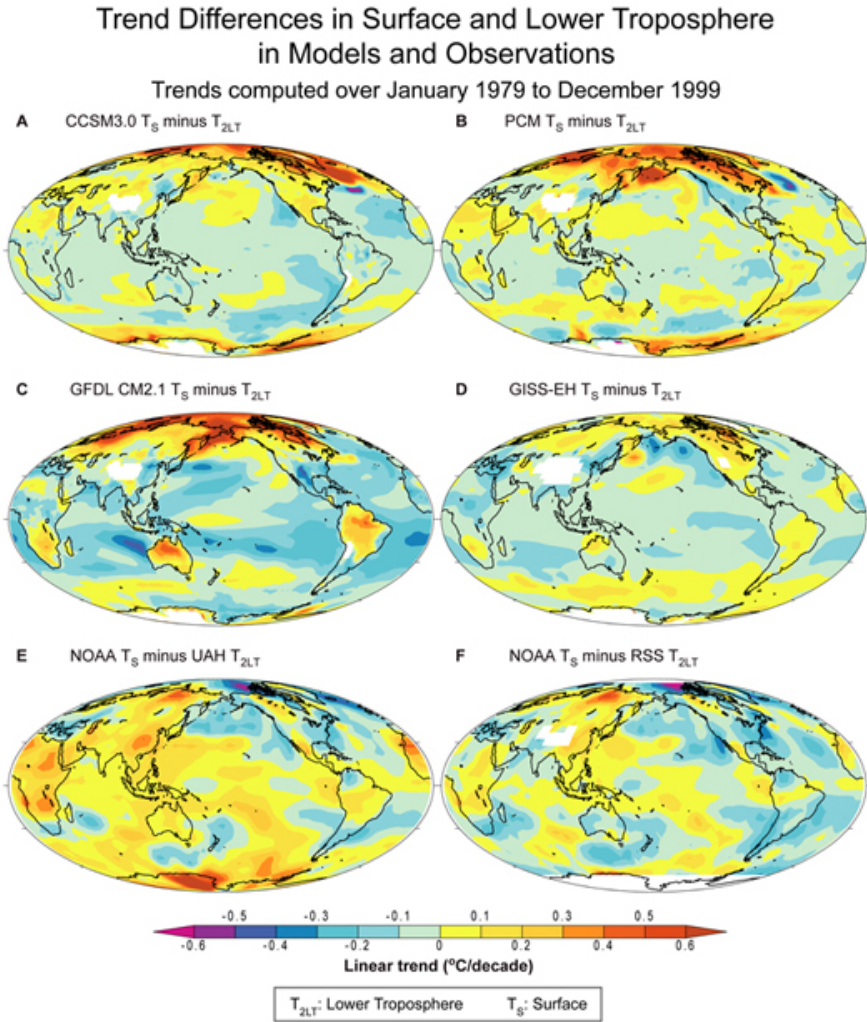
Figure 5.5: Modeled and observed maps of the differences between trends in T_S and T_{2LT}. All trends in T_S and T_{2LT} were calculated over the 252-month period from January 1979 to December 1999. Model results are ensemble means from 20CEN experiments performed with CCSM3.0 (panel A), PCM (panel B), GFDL CM2.1 (panel C), and GISS-EH (panel D). Observed results rely on NOAA T_S trends and on two different satellite estimates of trends in T_{2LT}, obtained from UAH (panel E) and RSS (panel F). White denotes high elevation areas where it is not meaningful to calculate synthetic T_{2LT} (panels A-D). Note that RSS mask T_{2LT} values in such regions, while UAH do not (*c.f.* panels F, E).

To illustrate structural uncertainties in the observed data, we show two different patterns of trends

in T_S minus T_{2LT}. Both rely on the same NOAA surface data, but use either UAH (Figure 5.5E) or

RSS (Figure 5.5F) as their source of T_{2LT} results. The "NOAA minus UAH" combination provides

a picture that is very different from the model results, with coherent warming of the surface relative

to the troposphere over much of the world's tropical oceans. While "NOAA minus RSS" also has

1068   relative warming of the surface in the Western and tropical Pacific, it shows relative warming of

1069   the *troposphere* in the eastern tropical Pacific and Atlantic Oceans. This helps to clarify why

1070   simulated lapse-rate trends in Figures 5.4F and 5.4G are closer to NOAA minus RSS results than to

1071   NOAA minus UAH results.

1072

1073   As pointed out by *Santer et al.* (2003b) and *Christy and Spencer* (2003), we cannot use such

1074   model-data comparisons alone to determine whether the UAH or RSS T_{2LT} dataset is closer to (an

1075   unknown) "reality." As the next section will show, however, models and basic theory can be used

1076   to identify aspects of observational behavior that require further investigation, and may help to

1077   constrain observational uncertainty.

1078

1079   5.4   *Tropospheric Amplification of Surface Temperature Changes*

1080

1081   When surface and lower tropospheric temperature changes are spatially averaged over the deep

1082   tropics, and when day-to-day tropical temperature changes are averaged over months, seasons, or

1083   years, it is evident that temperature changes aloft are larger than at the surface. This "amplification"

1084   behavior has been described in many observational and modeling studies, and is a consequence of

1085   the release of latent heat by moist convecting air (*e.g.*, *Manabe and Stouffer*, 1980; *Horel and*

1086   *Wallace*, 1981; *Pan and Oort*, 1983; *Yulaeva and Wallace*, 1994; *Hurrell and Trenberth*, 1998;

1087   *Soden*, 2000; *Wentz and Schabel*, 2000; *Hegerl and Wallace*, 2002; *Knutson and Tuleya*, 2004).[70]

---

[70]The essence of tropical atmospheric dynamics is that the tropics cannot support large temperature gradients, so waves (Kelvin, Rossby, gravity) even out the temperature field between convecting and non-convective regions. The temperature field throughout the tropical troposphere is more or less on the moist adiabat lapse rate set by convection over the warmest waters. This is why there is a trade wind inversion where this profile finds itself inconsistent with boundary layer temperatures in the colder regions.

1088

1089   A recent study by *Santer et al*. (2005) examined this amplification behavior in the same 20CEN

1090   runs and observational datasets considered in the present report. The sole difference (relative to the

1091   data used here) was that *Santer et al*. analyzed a version of the UAH T_{2LT} data that had not yet been

1092   adjusted for a recently-discovered error (*Mears and Wentz,* 2005).[71] The amplification of tropical

1093   surface temperature changes was assessed on different timescales (monthly, annual, and multi-

1094   decadal) and in different atmospheric layers ($T*_T$ and T_{2LT}).

1095

1096   On short timescales (month-to-month and year-to-year variations in temperature), the estimated

1097   tropospheric amplification of surface temperature changes was in good agreement in all model and

1098   observational datasets considered, and was in accord with basic theory. This is illustrated in Figure

1099   5.6, which shows the standard deviations of monthly-mean T_S anomalies plotted against the

1100   standard deviations of monthly-mean anomalies of T_{2LT} (panel A) and $T*_T$ (panel B). All model

1101   and observational results lie above the black line indicating equal temperature variability aloft and

1102   at the surface. All have similar "amplification factors" between their surface and tropospheric

1103   variability.[72] In the models, these similarities occur despite differences in physics, resolution, and

1104   forcings, and despite a large range (roughly a factor of 5) in the size of simulated temperature

1105   variability. In observations, the scaling ratios estimated from monthly temperature variability are

1106   relatively unaffected by the structural uncertainties discussed in Chapter 4.

---

[71] The error was related to the UAH group's treatment of systematic drifts in the time of day at which satellites sample Earth's diurnal temperature cycle (see Chapter 4).

[72] Note that the slope of the red regression lines that has been fitted to the model results is slightly steeper for $T*_T$ than for T_{2LT} (*c.f.* panels 5.6A and 5.6B). This is because $T*_T$ samples more of the mid-troposphere than T_{2LT} (see Prospectus). Amplification is expected to be larger in the mid-troposphere than in the lower troposphere.
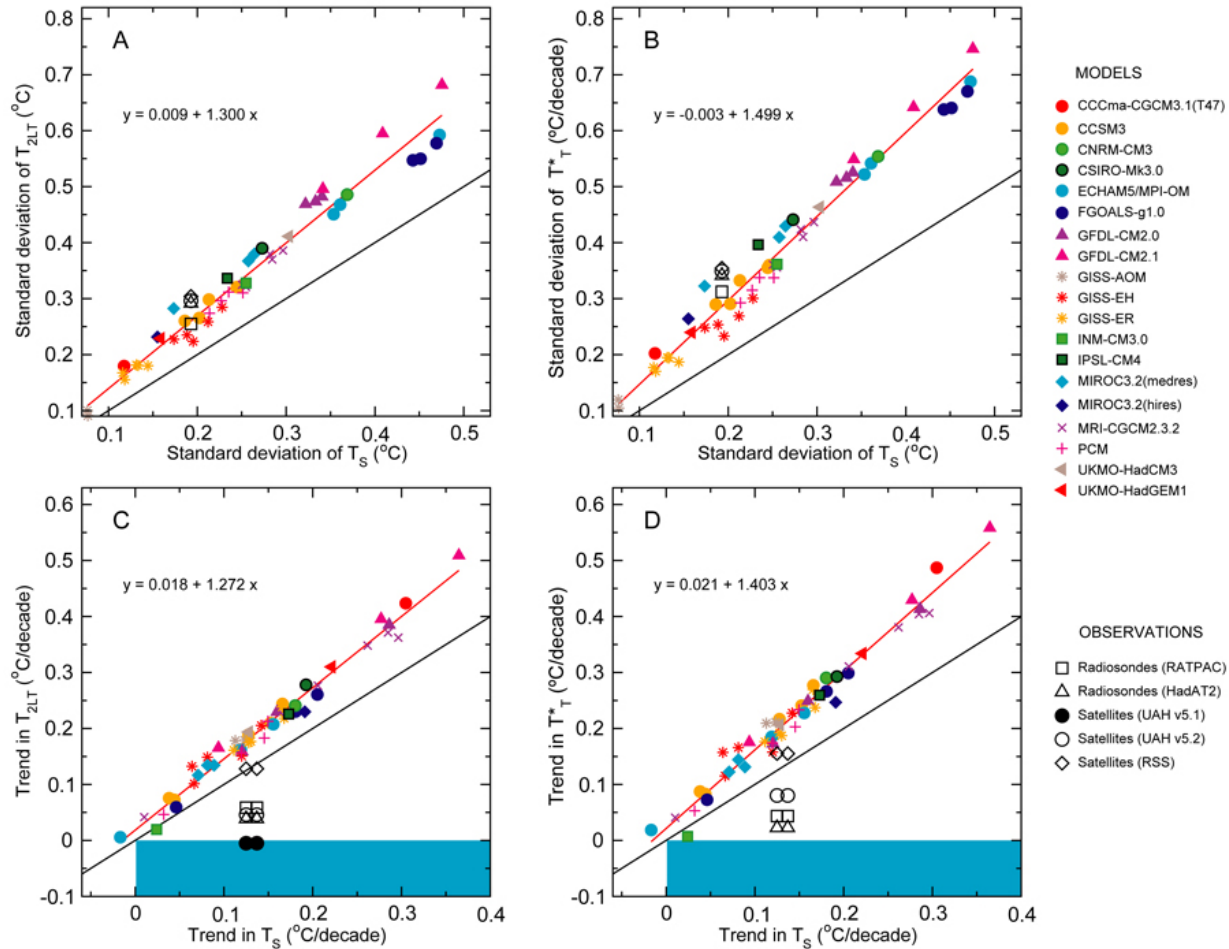
1107

Figure 5.6: Scatter plots showing the relationships between tropical temperature changes at Earth's surface and in two different layers of the troposphere. All results rely on temperature data that have been spatially-averaged over the deep tropics (20°N-20°S). Model data are from 49 realizations of 20CEN runs performed with 19 different models (Table 5.1). Observational results were taken from four different upper-air datasets (two from satellites, and two from radiosondes) and two different surface temperature datasets (see Chapter 3). The two upper panels provide information on the month-to-month variability in T_S and T_{2LT} (panel A) and in T_S and T^*_T (panel B). The two bottom panels consider temperature changes on multi-decadal timescales, and show the trends (over 1979 to 1999) in T_S and T_{2LT} (panel C) and in T_S and T^*_T (panel D). The red line in each panel is the regression line through the model points. Its slope provides information on the amplification of surface temperature variability and trends in the free troposphere. The black line in each panel is given for reference purposes, and has a slope of 1. Values above (below) the black lines indicate tropospheric amplification (damping) of surface temperature changes. There are two columns of observational results in C and D. These are based on the NOAA and HadCRUT2v TS (0.12 and 0.14°C/decade, respectively). Note that panel C show results from published and recently-revised versions of the UAH T_{2LT} data (versions 5.1 and 5.2). Since the standard deviations calculated from NOAA and HadCRUT2v monthly T_S anomalies are very similar, observed results in A and B use NOAA standard deviations only. The blue shading in the bottom two panels defines the region of simultaneous surface warming and tropospheric cooling.

1124
1125

1126   A different picture emerges if amplification behavior is estimated from decadal changes in tropical

1127   temperatures. Figures 5.6C and 5.6D show multi-decadal trends in T_S plotted against trends in T_{2LT}

1128   and T*$_T$. The 20CEN runs exhibit amplification factors that are consistent with those estimated

1129   from month-to-month and year-to-year temperature variability.[73] Only one observational upper-air

1130   dataset (RSS) shows amplified warming aloft, and similar amplification relationships on short and

1131   on long timescales. The other observational datasets have scaling ratios less than 1, indicating

1132   tropospheric <u>damping</u> of surface warming (*Fu et al*., 2005; *Santer et al*., 2005.[74]

1133

1134   These analyses shed further light on the differences between modeled and observed changes in

1135   tropical lapse rates described in Section 5.2. They illustrate the usefulness of comparing models

1136   and data on different timescales. On short timescales, it is evident that models successfully capture

1137   the basic physics that controls "real world" amplification behavior. On long timescales, model-data

1138   consistency is sensitive to structural uncertainties in the observations. One possible interpretation

1139   of these results is that in the real world, different physical mechanisms govern amplification

1140   processes on short and on long timescales, and models have some common deficiency in

1141   simulating such behavior. If so, these "different physical mechanisms" need to be identified and

1142   understood.

1143

1144   Another interpretation is that the same physical mechanisms control short- and long-term

1145   amplification behavior. Under this interpretation, residual errors in one or more of the observed

---

[73]As in the case of amplification factors inferred from short-timescale variability, the factors estimated from multi-decadal temperature changes are relatively insensitive to inter-model differences in physics and the applied forcings (see Table 5.3). At first glance, this appears to be a somewhat surprising result in view of the large spatial and temporal heterogeneity of certain forcings (see Section 3). Black carbon aerosols, for example, are thought to cause localized heating of the troposphere relative to the surface (Box 5.3), a potential mechanism for altering amplification behavior. The fact that amplification factors are similar in experiments that include and exclude black carbon aerosols suggests that aerosol-induced tropospheric heating is not destroying the connection of large areas of the tropical ocean to a moist adiabatic lapse rate. Single-forcing experiments (see Recommendations) will be required to improve our understanding of the physical effects of black carbon aerosols and other spatially-heterogeneous forcings on tropical temperature-change profiles.

[74]The previous version of the UAH T$_{2LT}$ data yielded a negative amplification factor for multi-decadal changes in tropical temperatures.

1146   datasets must affect their representation of long-term trends, and must lead to different scaling

1147   ratios on short and long timescales. This explanation appears to be the more likely one in view of

1148   the large structural uncertainties in observed upper-air datasets (Chapter 4) and the complementary

1149   physical evidence supporting recent tropospheric warming (see Section 6).

1150

1151   "Model error" and "observational error" are not mutually exclusive explanations for the

1152   amplification results shown in Figures 5.6C and D. Although a definitive resolution of this issue

1153   has not yet been achieved, the path towards such resolution is now more obvious. We have learned

1154   that models show considerable consistency in terms of what they tell us about tropospheric

1155   amplification of surface warming. This consistency holds on a range of different timescales.

1156   Observations display consistent amplification behavior on short timescales, but radically different

1157   behavior on long timescales. Clearly, not all of the observed lapse-rate trends can be equally

1158   probable. Intelligent use of "complementary evidence" – from the behavior of other climate

1159   variables, from remote sensing systems other than MSU, and from more systematic exploration of

1160   the impacts of different data adjustment choices – should ultimately help us to constrain

1161   observational uncertainty, and reach more definitive conclusions regarding the true significance of

1162   modeled and observed lapse-rate differences.

1163

1164   5.5   *Vertical Profiles of Atmospheric Temperature Change*

1165

1166   Although formal fingerprint studies have not yet been completed with atmospheric temperature-

1167   change patterns estimated from the new 20CEN runs, it is instructive to make a brief qualitative

1168   comparison of these patterns. This helps to address the question of whether the inclusion of

1169   previously-neglected forcings (like carbonaceous aerosols and land use/land cover changes; see

1170   Section 2) has fundamentally modified the "fingerprint" of human-induced atmospheric

1171   temperature changes searched for in previous detection studies.

1172

1173   We examine the zonal-mean profiles of atmospheric temperature change in 20CEN runs performed

1174   with four U.S. models (CCSM3, PCM, GFDL CM2.1, and GISS-EH). All four show a common

1175   large-scale fingerprint of stratospheric cooling and tropospheric warming over 1979 to 1999

1176   (Figures 5.7A-D). The pattern of temperature change estimated from HadAT2 radiosonde data is

1177   broadly similar, although the transition height between stratospheric cooling and tropospheric

1178   warming is noticeably lower than in the model simulations (Figure 5.7E). Another noticeable

1179   difference is that the HadAT2 data show a relative lack of warming in the tropical troposphere,[75]

1180   where all four models simulate maximum warming. This particular aspect of the observed

1181   temperature-change pattern is very sensitive to data adjustments (*Sherwood et al.*, 2005; *Randel*

1182   *and Wu*, 2005). Tropospheric warming in the observations is most obvious in the NH extratropics,

1183   where our confidence in the reliability of radiosonde records is greatest.

---

[75]Despite the "end point" effect of the large El Niño event in 1997-1998 (see Chapter 3).

1184

Zonal-Mean Atmospheric Temperature Change in Models and Data
Trends computed over January 1979 to December 1999
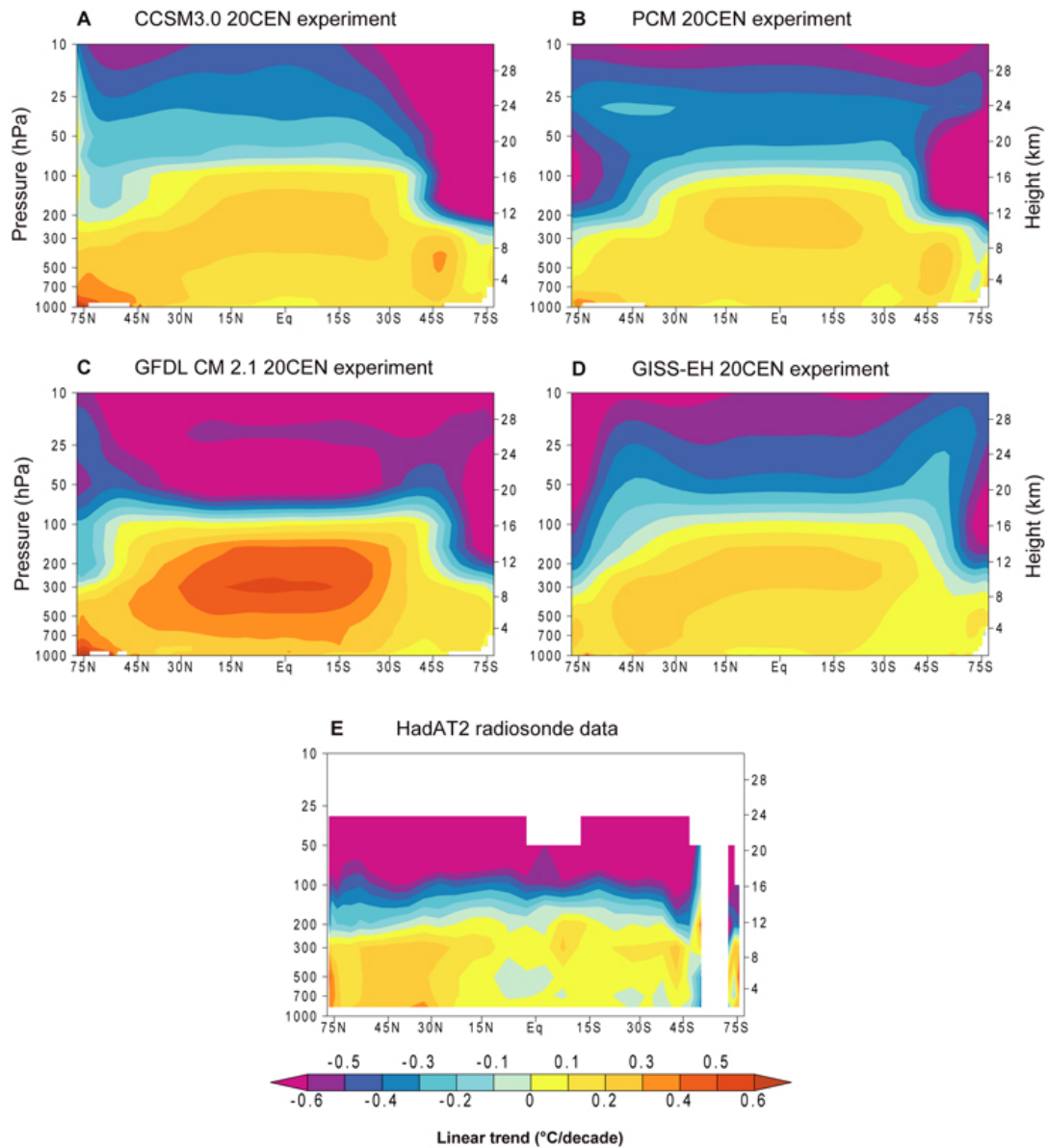
1185

Figure 5.7: Zonal-mean patterns of atmospheric temperature change in "20CEN" experiments performed with four different climate models and in observational radiosonde data. Model results are for CCSM3.0 (panel A), PCM (panel B), GFDL CM 2.1 (panel C), and GISS-EH (panel D). The model experiments are ensemble means. There are differences between the sets of climate forcings that the four models used in their 20CEN runs (Table 5.3). Observed changes (panel E) were estimated with HadAT2 radiosonde data (*Thorne et al.*, 2005, and Chapter 3). The HadAT2 temperature data do not extend above 30 hPa, and have inadequate coverage at high latitudes in the Southern Hemisphere. All temperature changes were calculated from monthly-mean data and are expressed as linear trends (in °C/decade) over 1979 to 1999.

1194

1195

1196   Note that some of the details of the model fingerprint pattern are quite different. For example,

1197   GFDL's cooling maximum immediately above the tropical tropopause is not evident in any of the

1198   other models. Its maximum warming in the upper tropical troposphere is noticeably larger than in

1199   CCSM3.0, PCM, or GISS-EH. While CCSM and GFDL CM2.1 have pronounced hemispheric

1200   asymmetry in their stratospheric cooling patterns, with largest cooling at high latitudes in the SH,[76]

1201   this asymmetry is less apparent in PCM and GISS-EH.

1202

1203   Future work should consider whether the conclusions of detection studies are robust to such

1204   fingerprint differences. This preliminary analysis suggests that the large-scale "fingerprint" of

1205   stratospheric cooling and tropospheric warming over the satellite era – a robust feature of previous

1206   detection work – has not been fundamentally altered by the inclusion of hitherto-neglected forcings

1207   like carbonaceous aerosols and LULC changes (see Table 5.3). This does not diminish the need to

1208   quantify the individual contributions of these forcings in appropriate "single forcing" experiments.

1209

1210   6   ***Changes in "Complementary" Climate Variables***

1211

1212   Body temperature is a simple metric of our physical well-being. A temperature of 40°C (104°F) is

1213   indicative of an illness, but does not by itself identify the cause of the illness. In medicine,

1214   investigation of causality typically requires the analysis of many different lines of evidence.

1215   Similarly, analyses of temperature alone provide incomplete information on the causes of climate

1216   change. For example, there is evidence that major volcanic eruptions affect not only the Earth's

---

[76]This may be related to an asymmetry in the pattern of stratospheric ozone depletion: the largest ozone decreases over the past 2-3 decades have occurred at high latitudes in the SH.

1217   radiation budget (*Wielicki et al.*, 2002; *Soden et al.*, 2002) and atmospheric temperatures (*Hansen*

1218   *et al.*, 1997, 2002; *Free and Angell*, 2002; *Wigley et al.*, 2005a), but also water vapor (*Soden et al.*,

1219   2002), precipitation (*Gillett et al.*, 2004c), atmospheric circulation patterns (see, *e.g.*, *Robock*, 2000,

1220   and *Ramaswamy et al.*, 2001a; *Robock and Oppenheimer*, 2003), ocean heat content and sea level

1221   (*Church et al.*, 2005), and even global-mean surface pressure (*Trenberth and Smith*, 2005). These

1222   responses are physically interpretable and internally consistent.[77] The combined evidence from

1223   changes in all of these variables makes a stronger case for an identifiable volcanic effect on climate

1224   than evidence from a single variable only.

1225

1226   A "multi-variable" perspective may also be beneficial in understanding the possible causes of

1227   differential warming. The value of "complementary" climate datasets for studying this specific

1228   problem has been recognized by *Pielke* (2004) and by *Wentz and Schabel* (2000). The latter found

1229   internally-consistent increases in SST, T_{2LT}, and marine total column water vapor over the 12-year

1230   period from 1987 to 1998.[78] Multi-decadal increases in surface and lower tropospheric water vapor

1231   were also reported in the IPCC Second Assessment Report (*Folland et al.*, 2001).[79] More recently,

---

[77]The physical consistency between the temperature and water vapor changes after the Pinatubo eruption has been clearly demonstrated by *Soden et al.* (2002). The surface and tropospheric cooling induced by Pinatubo caused a global-scale reduction in total column water vapor. Since water vapor is a strong GHG, the reduction in water vapor led to less trapping of outgoing thermal radiation by Earth's atmosphere, thus *amplifying* the volcanic cooling. This is referred to as a "positive feedback." *Soden et al.* "disabled" this feedback in a climate model experiment, and found that the "no water vapor feedback" model was incapable of simulating the observed tropospheric cooling after Pinatubo. Inclusion of the water vapor feedback yielded close agreement between the simulated and observed T_{2LT} responses to Pinatubo. This suggests that the model used by *Soden et al.* captures important aspects of the physics linking the real world's temperature and moisture changes.

[78]The Wentz and Schabel study used NOAA optimally-interpolated SST data, a version of the UAH T_{2LT} data that had been corrected for orbital decay effects, and information on total column water vapor from the satellite-based Special Sensor Microwave Imager (SSM/I).

[79]More specifically, *Folland et al.* (2001) concluded that "Changes in water vapour mixing ratio have been analysed for selected regions using *in situ* surface observations as well as lower-tropospheric measurements based on satellites and weather balloons. A pattern of overall surface and lower-tropospheric water vapour mixing ratio increases over the past few decades is emerging, although there are likely to be some time-dependent biases in these data and regional variations in trends. The more reliable data sets show that it is likely that total atmospheric water vapour has increased

---

1232   *Trenberth et al.* (2005) found significant increases in total column water vapor over the global

1233   ocean.[80] At constant relative humidity, water vapor increases nonlinearly with increasing

1234   temperature (*Hess*, 1959). Slow increases in tropospheric water vapor therefore provide

1235   circumstantial evidence in support of tropospheric warming. However, water vapor measurements

1236   are affected by many of the same data quality and temporal homogeneity problems that influence

1237   temperature measurements (*Elliott*, 1995; *Trenberth et al.*, 2005), so the strength of this

1238   circumstantial evidence is still questionable.[81]

1239

1240   Other climate variables also corroborate the warming of Earth's surface over the second half of the

1241   20th Century. Examples include increases in ocean heat content (*Levitus et al.*, 2000, 2005; *Willis* et

1242   al., 2004), sea-level rise (*Cabanes et al.*, 2001), thinning of major ice sheets and ice shelves

1243   (*Krabill et al.*, 1999; *Rignot and Thomas*, 2002; *Domack et al.*, 2005), and widespread glacial

1244   retreat, with accelerated rates of glacial retreat over the last several decades (*Arendt et al.*, 2002;

1245   *Paul et al.,* 2004).[82]

1246

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several per cent per decade over many regions of the Northern Hemisphere since the early 1970s. Changes over the Southern Hemisphere cannot yet be assessed".

[80]*Trenberth et al.* (2005) reported an increase in total column water vapor over 1988 to 2001 of "1.3 ± 0.3% per decade for the ocean as a whole, where the error bars are 95% confidence intervals." This estimate was obtained with an updated version of the SSM/I dataset analyzed by *Wentz and Schabel* (2000).

[81]Note, however, that SSM/I-derived water vapor measurements may have some advantages relative to temperature measurements obtained from MSU. *Wentz and Schabel* (2000) point out that (under a constant relative humidity assumption), the 22 GHz water vapor radiance observed by SSM/I is three times more sensitive to changes in air temperature than the MSU T_2 54 GHz radiance. Furthermore, while drift in sampling the diurnal cycle influences MSU-derived tropospheric temperatures (Chapter 4), it has a much smaller impact on SSM/I water vapor measurements.

[82]*Folland et al.* (2001) note that "Long-term monitoring of glacier extent provides abundant evidence that tropical glaciers are receding at an increasing rate in all tropical mountain areas". Accelerated retreat of high-elevation tropical glaciers is occurring within the tropical lower tropospheric layer that is a primary focus of this report, and provides circumstantial support for warming of this layer over the satellite era".

1247   Changes in some of these "complementary" variables have been used in detection and attribution

1248   studies. Much of this work has focused on ocean heat content. When driven by anthropogenic

1249   forcing, a number of different CGCMs capture the overall increase in observed ocean heat content

1250   estimated by *Levitus et al.* (2000; 2005), but not the large decadal variability in heat content

1251   (*Barnett et al.*, 2001; *Levitus et al.*, 2001; *Reichert et al.*, 2002; *Sun and Hansen*, 2003; *Pielke*,

1252   2003; *Gregory et al.*, 2004; *Hansen et al.*, 2005b).[83] It is still unclear whether this discrepancy

1253   between simulated and observed variability is primarily due to model deficiencies or is an artifact

1254   of how *Levitus et al.* (2000; 2005) "infilled" data-sparse ocean regions (*Gregory et al.*, 2004;

1255   *AchutaRao et al.*, 2005).

1256

1257   In summary, the behavior of complementary variables enhances our confidence in the reality of

1258   large-scale warming of the Earth's surface, and tells us that the signature of this warming is

1259   manifest in many different aspects of the climate system. Pattern-based fingerprint detection work

1260   performed with ocean heat content (*Barnett et al.*, 2001; *Reichert et al.*, 2002; *Barnett et al.*, 2005;

1261   *Pierce et al.*, 2005), sea-level pressure (*Gillett et al.,* 2003), and tropopause height (*Santer et al.*,

1262   2003a, 2004) suggests that anthropogenic forcing is necessary in order to explain observed changes

1263   in these variables. This supports the findings of the surface- and atmospheric temperature studies

1264   described in Section 4.4. To date, however, investigations of complementary variables have not

1265   enabled us to narrow uncertainties in satellite- and radiosonde-based estimates of tropospheric

1266   temperature change over the past two-and-a-half decades.[84] Formal detection and attribution studies

---

[83]Model control runs cannot generate such large multi-decadal increases in the heat content of the global ocean.

[84]The tropopause is the transition zone between the turbulently-mixed troposphere, where most weather occurs, and the more stably-stratified stratosphere (see Preface and Chapter 1). Increases in tropopause height over the past 3-4 decades represent an integrated response to temperature changes above and below the tropopause (*Highwood et al.*, 2000; *Santer et al.*, 2004), and are evident in both radiosonde data (*Highwood et al.*, 2000; *Seidel et al.*, 2001) and reanalyses (*Randel et al.*, 2000). In model 20CEN simulations, recent increases in tropopause height are driven by the

1267   involving water vapor changes may be helpful in this regard, since observations suggest a recent

1268   moistening of the troposphere, consistent with tropospheric warming.

1269

1270   7   ***Summary***

1271

1272   This chapter has evaluated a wide range of scientific literature dealing with the possible causes of

1273   recent temperature changes, both at the Earth's surface and in the free atmosphere. It shows that

1274   many factors – both natural and human-related – have probably contributed to these changes.

1275   Quantifying the relative importance of these different climate forcings is a difficult task. Analyses

1276   of observations alone cannot provide us with definitive answers. This is because there are

1277   important uncertainties in the observations and in the climate forcings that have affected them.

1278   Although computer models of the climate system are useful in studying cause-effect relationships,

1279   they, too, have limitations. Advancing our understanding of the causes of recent lapse-rate changes

1280   will best be achieved by comprehensive comparisons of observations, models, and theory – it is

1281   unlikely to arise from analysis of a single model or observational dataset.

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combined effects of GHG-induced tropospheric warming and ozone-induced stratospheric cooling (*Santer et al.*, 2003a). Available reanalysis products do not provide a consistent picture of the relative contributions of stratospheric and tropospheric temperature changes to recent tropopause height increases (*Pielke and Chase*, 2004; *Santer et al.*, 2004).

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5   **Chapter 6**
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9   What measures can be taken to improve the understanding of observed changes?
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15   Convening Lead Author:
16   Chris Folland
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18   Lead Authors:
19   David Parker, Richard Reynolds, Steve Sherwood, and Peter Thorne
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46
47   **Background**

48

49   There remain differences between independently estimated temperature trends for the surface,

50   troposphere and lower stratosphere, and differences between the observed changes and model

51   simulations, that are, as yet, not fully understood, although recent progress is reported in

52   previous chapters. This Chapter makes recommendations that address these specific problems

53   rather than more general climate research aims, building on the discussions, key findings, and

54   recommendations of the previous chapters. Because the previous chapters fully discuss the

55   many issues, we only provide a summary here. Furthermore, we only list key references to

56   the peer reviewed literature. To ensure traceability and to enable easy cross-referencing we

57   refer to the chapters by e.g., **(C5)** for Chapter 5. We do not specifically refer to sub-sections

58   of chapters.

59

60   Much previous work has been done to address, or plan to address, most of the problems

61   discussed in this Report. Rather than invent brand new proposals and recommendations, we

62   have tried to expand and build upon existing ideas emphasizing those we believe to be of

63   highest utility. Key documents in this regard are: the Global Climate Observing System

64   (GCOS) Implementation Plan for the Global Observing System (GCOS, 2004), the wider

65   Global Earth System of Systems (GEOSS) 10 year Implementation Plan Reference

66   Document (GEO, 2005) which explicitly includes the GCOS Implementation Plan as its

67   climate component; and the over-arching Climate Change Science Program plan (CCSP,

68   2004).

69

70   The remainder of this Chapter is split into six sections. Each section discusses requirements

71   under a particular theme, aiming to encapsulate the key findings and recommendations of the

72   earlier chapters and culminating in one main recommendation in each of Sections 1 to 5 and

73   two recommendations in Section 6. Sections 1 to 5 focus on key actions that should be

74   carried out in the near future, making use of existing historical data and current climate

75   models. Section 6 discusses future climate monitoring in relation to the vertical profile of

76   temperature trends in the atmosphere.

77

78   **1. Constraining observational uncertainty**

79

80   An important advance since recent in-depth reviews of the subject of this Report (NRC,

81   2000a, IPCC, 2001) has been a better appreciation of the uncertainties in our estimates of

82   recent temperature changes, particularly above the surface **(C2, C3, C4)**. Many observations

83   that are used in climate studies are taken primarily for the purposes of operational weather

84   forecasting **(C2).** Not surprisingly, there have been numerous changes in instrumentation,

85   observing practices, and the processing of data over time. While these changes have

86   undoubtedly led to improved forecasts of weather, they add significant complexity to

87   attempts to reconstruct past climate trends, **(C2, C4).** The main problem is that such an

88   evolution tends to introduce artificial (non-climatic) changes into the data **(C2)**.

89

90   Above the Earth's surface, the spread in independently-derived estimates of climate change,

91   representing what is referred to in this report as "construction" uncertainty **(C2, C4,**

92   **Appendix)** (Thorne et al., 2005), is of similar magnitude to the expected climate signal itself

93   **(C3, C4, C5)**. Changes in observing practices have been particularly pervasive aloft, where

94   the technical challenges in maintaining robust, consistent measurements of climate variables

95   are considerably greater than at the surface **(C2, C4, C5)**. This does not imply that there are

96   no problems in estimating temperature trends at the surface. Such problems include

97   remaining uncertainties in corrections that must be made to sea surface temperatures (SSTs)

98   in recent decades **(C2, C4),** and uncertainties in accounting for changes in micro-climate

99   exposure for some individual land stations **(C2, C4)** or simply allowing for genuinely bad

100   stations (Davey and Pielke, 2005). Differences between surface data sets purporting to

101   measure the same variable become larger as the spatial resolution being considered decreases.

102   This implies that many problems tend to have random effects on climate analyses at the large

103   spatial scales, that are the focus of this Report, but can be systematic at much smaller scales

104   **(C2, C3, C4)**.

105

106   The climate system has evolved in a unique way, and, by definition the best analysis is that

107   which most closely approaches this actual evolution. However, because we do not know the

108   evolution of the climate system exactly, we have generally had to treat apparently well

109   constructed but divergent data sets, of atmospheric temperature changes in particular, as

110   equally valid **(C3, C4, C5)**. Clearly, this approach is untenable in the longer-term. Thus, it is

111   imperative that we reduce the uncertainty in our knowledge of how the three-dimensional

112   structure of atmospheric temperature has evolved **(C4)**.

113

114   To ascertain unambiguously the causes of differences in datasets generally requires extensive

115   metadata[1] for each data set **(C4;** NRC, 2000b**)**. Appropriate metadata, whether obtained from

116   the peer-reviewed literature or from data made available on-line, should include, for data on

117   all relevant spatial and temporal scales:

118   • Documentation of the raw data and the data sources used in the data set construction

119   to enable quantification of the extent to which the raw data overlap with other similar

120   datasets;

121   • Details of instrumentation used, the observing practices and environments and their

122   changes over time to help assessments of, or corrections for, the changing accuracy of

123   the data;

124   • Supporting information such as any adjustments made to the data and the numbers

125   and locations of the data through time;

126   • An audit trail of decisions about the adjustments made, including supporting evidence

127   that identifies non-climatic influences on the data and justifies any consequent

128   adjustments to the data that have been made; and

129   • Uncertainty estimates and their derivation.

130   This information should be made openly available to the research community.

131

132   There is evidence, discussed in earlier chapters, for a number of unresolved issues in existing

133   data sets that should be addressed:

134   • Systematic, historically-varying biases in day-time relative to night-time radiosonde

135   temperature data are important, particularly in the tropics **(C4)**. These are likely to

---

[1] Metadata are literally "data about data" and are typically records of instrumentation used, observing practices, the environmental context of observations, and data-processing procedures.

136   have been poorly accounted for by present approaches to quality controlling such data

137   (Sherwood et al., 2005) and may seriously affect trends.

138   • Radiosonde stratospheric records are strongly suspected of retaining a spurious long-

139   term cooling bias, especially in the tropics **(C4)**.

140   • Diurnal adjustment techniques for satellite temperature data are uncertain **(C2, C4)**.

141   This effect is particularly important for the 2LT retrieval **(C4)**. Further efforts are

142   required to refine our quantification of the diurnal cycle, perhaps through use of

143   reanalyses, in-situ observations, or measurements from non-sun-synchronous orbiters

144   **(C4)**.

145   • Different methods of making inter-satellite bias adjustments, particularly for satellites

146   with short periods of overlap, can lead to large discrepancies in trends **(C4)** (see also

147   Section 6).

148   • Variable biases in modern SST data remain that have not been adequately addressed

149   **(C4)**. Some historical metadata are now available for the first time, but are yet to be

150   fully exploited (Rayner et al., 2005). Better metadata, better use of existing metadata,

151   and use of recently bias-adjusted day-time marine air temperature data are needed to

152   assess remaining artifacts **(C4)**.

153   • Land stations may have had undocumented changes in the local environment that

154   could lead to their records being unrepresentative of regional- or larger-scale changes

155   **(C2, C4)**.

156

157   In addition to making data sets and associated metadata openly available and addressing the

158   issues discussed above, it would be useful to develop a set of guidelines that can be used to

159   help assess the quality of data sets **(C4)**. It is important that numerous tests be applied to

160   reduce ambiguity. There are three types of check that may be used:

161

162   1. *Internal consistency checks*

163   For example, we expect only relatively small real changes in the diurnal cycle of

164   temperature above the atmospheric boundary layer **(C1)** (Sherwood et al., 2005), so

165   an apparently homogenized data set that shows large changes in the diurnal cycle in

166   these regions should be closely scrutinized.

167   2. *Inter-dataset comparisons*

168   For example, comparisons are needed between radiosonde and MSU temperature

169   measures representing the same regions (Christy and Norris, 2004).

170   3. *Consistency with changes in other climate variables and parameters*

171   This is a potentially powerful but much under-utilized approach and is discussed

172   further in Section 3.

173

174

175   **RECOMMENDATION** *In order to encourage further independent scrutiny, data sets and*

176   *their full metadata (i.e., information about instrumentation used, observing practices, the*

177   *environmental context of observations, and data-processing procedures) should be made*

178   *openly available. Comprehensive analyses should be carried out to ascertain the causes of*

179   *remaining differences between data sets and to refine uncertainty estimates.*

180

181   **2.   Making better use of existing observational data**

182

183   There is a considerable body of observational data that have either been under-utilized or not

184   used at all when constructing the data sets of historical temperature changes discussed in this

185   Report **(C2, Table 2.1)**. Estimates of temperature changes can potentially be made from

186   several satellite instruments beside the (Advanced) Microwave Sounding Unit data

187   considered here **(C2, C3).** In particular, largely overlooked satellite datasets should be re-

188   examined to try to extend, fortify or corroborate existing microwave-based temperature

189   records for climate research, e.g., microwave data from other instruments such as the Nimbus

190   5 (Nimbus E) Microwave Spectrometer (NEMS) (1972) and the Nimbus 6 Scanning

191   Microwave Spectrometer (SCAMS) (1975), infra-red data from the High Resolution Infrared

192   Radiation Sounder (HIRS) suite, and radio occultation data from Global Positioning System

193   (GPS) satellites **(C2)**. Some of these instruments may allow us to extend the records back to

194   the early 1970s. Many unused radiosonde measurements of a relatively short length exist in

195   regions of relatively sparse coverage and, with some effort, could be advantageously used to

196   fill gaps. Many additional surface temperature data exist, mainly over land over the period

197   considered in this Report, but are either not digitized or not openly available. This latter

198   problem is particularly common in many tropical regions where much of the interest in this

199   Report resides. Given the needed level of international cooperation, we could significantly

200   improve our current estimates of tropical temperature changes over land and derive better

201   estimates of the changing temperature structure of the lower atmosphere **(C2)**.

202

203   In addition to the recovery and use of such existing data, we need to improve the access to

204   metadata for existing raw observations **(C2)**. Additional information on when and how

205   changes occurred in observing practices, the local environment, etc., is potentially available

206   in national meteorological and hydrometeorological services. Such metadata would help

207   reduce current uncertainties in estimates of observed climate change. In the absence of

208   comprehensive metadata, investigators have to make decisions regarding the presence of

209   heterogeneities (non-climatic jumps or trends) using statistical methods alone. Statistical

210   methods of adjusting data for inhomogeneities have a very useful role, but are much more

211   valuable in the presence of good and frequent metadata that can be used to confirm the

212   presence, type, and timing of non-climatic influences. Metadata requirements will vary

213   according to observing system, but, if in doubt, all potentially important information should

214   be included. For example, surface temperature metadata may include:

215   • Current and historical photographs and site sketches to ascertain changes in micro-

216   climate exposure and their timing, collected during the routine site inspections made

217   by most meteorological services;

218   • The history of instrumentation changes;

219   • Changes in the way stations are maintained and in their immediate environment;

220   • Changes in observers; and

221   • Changes in observing and reporting practices.

222   For other instrument types, e.g., for humidity measurements, the detailed metadata

223   requirements will vary. A further discussion on the challenges of collecting climate data can

224   be found in Folland et al. (2000).

225

226   **RECOMMENDATION** *Efforts should be made to archive and make openly available*

227   *surface, balloon-based, and satellite data and metadata that have not previously been*

228   *exploited. Emphasis should be placed on the tropics and on recovery and inclusion of*

229   *satellite data before 1979 which may allow better characterization of the climate regime shift*

230   *in the mid-1970's*

231

232

233   3. **Multivariate analyses**

234

235   Temperature changes alone are a necessary, but insufficient, constraint on understanding the

236   evolution of the climate system. Even with a perfect knowledge of temperature changes,

237   knowledge about changes in the climate system would be incomplete. Consequently,

238   understanding temperature trends also requires knowledge about changes in other measures

239   of the climate system. For example, changes in atmospheric circulation and accompanying

240   dynamical effects, and also in latent heat transport, have significant implications for vertical

241   profiles of temperature trends **(C1)**.

242

243   Changes in variables other than temperature may be used to confirm the attribution of climate

244   change to given causes **(C5)** and to test the physical plausibility of reported temperature

245   changes **(C3, C4)**. It is likely that to fully understand changes in atmospheric temperature, it

246   will be necessary to consider changes in at least some of the following physical parameters

247   and properties of the climate system beside its temperature:

248   • Water vapor content **(C1, C5)**

249   • Ocean heat content **(C5)**

250   • The height of the tropopause **(C5)**

251   • Wind fields

252   • Cloud cover and the characteristics of clouds

253   • Radiative fluxes

254   • Aerosols and trace gases

255   • Changes in glacial mass, sea ice volume, permafrost and snow cover **(C5)**

256

257   Our current ability to undertake such multivariate analyses of climate changes is constrained

258   by the relative paucity of accurate climate datasets for variables other than temperature.

259   Furthermore, since our analysis of temperature datasets has highlighted the importance of

260   construction uncertainty in determining trends **(C2, C4, Appendix),** it is very likely that

261   similar considerations will pertain to these other data types. It is therefore necessary to

262   construct further independent estimates of the changes in these variables even where datasets

263   already exist. Similar considerations to those discussed in Section 1 are also important for

264   these additional data.

265

266   **RECOMMENDATION** *Efforts should be made to create climate quality[2] data sets for a*

267   *range of variables other than temperature. These should subsequently be compared with each*

268   *other and with temperature data to determine whether they are consistent with our physical*

269   *understanding. It is important to create several independent estimates for each parameter in*

270   *order to assess the magnitude of construction uncertainties.*

271

272   4. **Climate quality reanalyses**

273

274   Reanalyses are derived from Numerical Weather Prediction (NWP) (forecast) models run

275   retrospectively with historical observations to produce physically consistent, fully global

276   fields with high temporal and spatial resolution. As in NWP, reanalyses employ all available

277   observations to produce their analysis and minimize the instantaneous differences between

---

[2] "Climate quality" refers to a record for which the best possible efforts have been made to identify and remove non-climatic effects that produce spurious changes over time.

278   the available observations and a background forecast field initiated a number of hours earlier.

279   Reanalyses also use the same NWP model throughout the reanalysis period. However, as for

280   observed climate datasets, pervasive changes in the raw observations lead to discontinuities

281   and spurious drifts **(C2)**. Because such discontinuities and drifts have been identified in the

282   temperature fields of the current generation of reanalyses, these have been deemed

283   inappropriate for the purpose of long-term temperature trend characterisation by this Report's

284   authors **(C2, C3). However,** it is recognised that some progress has been made (e.g.,

285   Simmons et al., 2004, **C2**). This does not preclude the usefulness of reanalyses for

286   characterizing seasonal to interannual timescale variability and processes, or trends in other,

287   related, variables such as tropopause height **(C5)**. Indeed, they have proven to be a very

288   important tool for the climate research community.

289

290   A more homogeneous reanalysis that minimized time-dependent biases arising from changes

291   in the observational network would be of enormous benefit for multivariate analyses of

292   climate change **(C2, C3)**. Advances in NWP systems, which will continue to happen

293   regardless of climate requirements, will inevitably lead to better future reanalyses of

294   interannual climate variability. Some advances, such as so-called 'feedback files'[3] from the

295   data assimilation of reanalyses, could be uniquely helpful for climate reanalysis and should

296   be encouraged for this reason if no other. However, to determine trends accurately from

297   reanalyses will also require intensive efforts by the reanalysis community to understand

298   which observations are critical for trend characterization and to homogenize these data

299   insofar as possible to eliminate non-climatic changes before input to the reanalysis system.

300   This in turn requires observing system experiments (OSEs) where the impact on trends of

---

[3] "Feedback files" are diagnostic summaries of adjustments applied to data during assimilation.

301   different observation types from land, radiosonde, and space-based observations are assessed.

302   A few examples (far from an exhaustive list) are:

303   •   Run a short period (e.g., a year) of reanalysis with and without radiosonde

304   temperatures.

305   •   Carry out experiments incorporating long radiosonde data records only, then repeat

306   with additional less certain but spatially more complete short radiosonde records.

307   •   Successively include or remove specific satellite retrievals (e.g., MSU Channel 2).

308   •   Carry out test reanalyses for one or more decades with different corrections to the

309   observed data for inhomogeneities within their construction uncertainty estimates.

310

311   Progress would depend on reanalyses and data construction experts from all the key groups

312   working closely together.

313

314

315   **RECOMMENDATION** *Consistent with Key Action 24 of GCOS (2004)[4] and a 10 Year*

316   *Climate Target of GEOSS (2005), efforts should be made to create several homogeneous*

317   *atmospheric reanalyses. Particular care needs to be taken to identify and homogenize critical*

318   *input climate data, and to more effectively manage large-scale changes in the global*

319   *observing system to avoid non-climatic influences.[5]*

320

321   5.   **Better understanding of uncertainties in model estimates**

---

[4] Parties are urged to give high priority to establishing a sustained capacity for global climate reanalysis, to develop improved methods for such reanalysis, and to ensure coordination and collaboration among Centers conducting reanalyses.

[5] A focal point for planning of future U.S. reanalysis efforts is the CCSP Synthesis and Assessment Product 1.3: "Re-analyses of historical climate data for key atmospheric features. Implications for attribution of causes of observed change". Ongoing progress in the planning of future U. S. reanalysis efforts can be found at: http://www.joss.ucar.edu/joss_psg/meetings/climatesystem/

322

323   New state-of-the-art global climate models have simulated the influences of natural and

324   anthropogenic climate forcings on tropospheric and surface temperature. The simulations

325   generally cover the period since the late nineteenth century, but results are only reported over

326   the period of primary interest to this Report, 1979-1999 (the satellite era), in Chapter 5.

327   Taken together, these models, for the first time, consider most of the recognized first-order

328   climate forcings and feedbacks as identified in IPCC (2001), NRC (2003), and NRC (2005).

329   This is an important step forward **(C5)**.

330

331   However, most individual models considered in this Report still do not make use of all likely

332   important climatic forcings **(C5, Table 5.2)**. In addition, many of the forcings are not yet well

333   quantified. Models that appear to include the same forcings often differ in both the way the

334   forcings are quantified and how these forcings are applied to the model. This makes it

335   difficult to separate intrinsic differences between models from the effects of different forcings

336   on predicted temperature trends. Thus, within the "ensemble of opportunity" considered in

337   this Report **(C5),** it is difficult to separate differences in:

338   • Model physics and resolution;

339   • The details of the way the forcings are applied in the experiments;

340   • The chosen history of the changes in the forcing.

341

342   To better quantify the impacts of the various forcings on vertical temperature trends, a further

343   suite of experiments is needed along the following lines:

344   • Runs with one forcing applied in a single experiment with a given model; these are

345   already required in some detection and attribution studies **(C5)**. They have been

346   performed for a small number of models already. This approach is particularly

347   important for the recently developed and spatially heterogeneous land use / land cover

348   change and black carbon aerosol forcings **(C5)**.

349   •   Apply the same forcing in exactly the same manner to a suite of models so that the

350   differences that result are due unambiguously to model differences **(C5)**.

351   •   Apply the full range of important forcings, with their uncertainties explicitly sampled

352   to a small subset of the most advanced models to gain an overall estimate of the

353   effects on temperature trends of the uncertainties in these forcings.

354

355   It is recognized that there are many problems in achieving this, so a considerable effort will

356   be needed over a number of years. In addition, these model runs should be compared to the

357   full range of observational estimates to avoid ambiguity **(C5)**. Finally, detection and

358   attribution studies should be undertaken using this new range of observations and model-

359   based estimates to reassess previous results; which have consistently identified a human-

360   induced influence **(C5).**

361

362

363   **RECOMMENDATION** *Models that appear to include the same forcings often differ in both*

364   *the way the forcings are quantified and how these forcings are applied to the model. Hence,*

365   *efforts are required to more formally separate uncertainties arising from model structure*

366   *from the effects of forcing uncertainties. This requires running multiple models with*

367   *standardized forcings, and running the same models individually under a range of plausible*

368   *scenarios for each forcing."*

369

370   6. **Future monitoring of climate**

371

372   Much of this Report hitherto has concerned historical climate measurements. However, over

373   the coming decades new, mainly space-based, observations will yield very large increases in

374   the volume and types of data available. These will come from many different instruments

375   making measurements with greater accuracy and detail, especially in the vertical direction,

376   and with greater precision **(C2, C3)**. In fact, new types of more accurate data such as

377   temperature and moisture profiles from GPS radio-occultation measurements are already

378   available, although, as yet, few efforts have been made to analyse them **(C2, C3)**. Current and

379   planned multi-spectral infra-red satellite sounders such as the Atmospheric InfraRed Sounder

380   (AIRS) and the Infrared Atmospheric Sounding Interferometer (IASI) have much finer

381   vertical resolution than earlier satellite sounders used in the Report. They have the potential

382   to resolve quite fine vertical and horizontal details of temperature and humidity through the

383   depth of much of the atmosphere. These higher spectral resolution data should also permit a

384   continuation of records equivalent to earlier coarser infrared satellite data (e.g., from the

385   HIRS satellite instruments). The new suite of satellite data will not only prove useful for

386   sensing changes aloft. For example, satellite data to remotely sense sea-surface temperatures

387   now include microwave products that can sense surface temperatures even in cloudy

388   conditions (C4). The Global Ocean Data Assimilation Experiment (GODAE) High-

389   resolution Sea Surface Temperature (SST) Pilot Project (GHRSST- PP) has been established

390   to give international focus and coordination to the development of a new generation of global,

391   multi-sensor, high-resolution SST products (Donlon et al., 2005).

392

393 Many other agencies and bodies (e.g., NRC, 2000b; GCOS, 2004; GEOSS, 2005; CCSP,

394 2004) have already made recommendations for managing such new data developments.

395 These include such subjects as:

396 • Adherence to the GCOS Climate Monitoring Principles, needed to create and

397   maintain homogenous data sets of climate quality and for which there is a special set

398   for satellites (GCOS, 2004, Appendix 3)

399 • Continuation of records equivalent to current monitoring abilities: e.g., use new and

400   more detailed satellite data to create equivalent MSU measures of temperature to

401   allow the indefinite extension of the historical records used in this Report.

402 • Full implementation of national and international climate monitoring networks such

403   as the GCOS Upper-Air Network and the GCOS Surface Network.

404 • Overlap of measurement systems as they evolve in time.

405

406 This last point is of primary importance. It was given prominence by NRC (2000b) and is

407 emphasized in the GCOS Climate Monitoring Principles and leads to the following

408 recommendation. If this recommendation had been followed in the past, one of the major

409 problems in producing a homogeneous record of MSU temperatures would have been largely

410 removed **(C4)**:

411

412 **RECOMMENDATION**: *The GCOS Climate Monitoring principles should be fully adopted.*

413 *In particular when any type of instrument for measuring climate is changed or re-sited, the*

414 *period of overlap between the old and new instruments or configurations should be sufficient*

415 *to allow analysts to adjust for the change with small uncertainties that do not prejudice the*

416 *analysis of climate trends. The minimum period is a full annual cycle of the climate. Thus,*

417   *replacement satellite launches should be planned to take place at least a year prior to the*

418   *expected time of failure of a satellite.*

419

420

421   Finally, we expand on a recommendation made in GCOS (2004) that is imperative for

422   successful future monitoring of temperatures at and above the Earth's surface. The main

423   lesson learned from this Report is that great difficulties in identifying and removing non-

424   climatic influences from upper-air observations have led to a very large spread in trend

425   estimates **(C2, C3, C4)**. These differences can lead to fundamentally different interpretations

426   both of the extent of any discrepancies in trends between the surface and the troposphere

427   **(C3,C4)**; and of the skill of climate models **(C5)**. The problem has arisen because there has

428   been no high quality reference or "ground truth" data, however restricted in scope, against

429   which routine observations can be compared to facilitate rigorous removal of non-climatic

430   influences.

431

432   Our key recommendation in this regard is a set of widely distributed (perhaps about 5% of the

433   operational radiosonde network) reference sites that will provide high quality data for

434   anchoring more globally-extensive monitoring efforts (satellites, reanalyses, etc.). At such

435   reference sites (which could coincide with selected GCOS Upper Air Network (GUAN),

436   GCOS Surface Network (GSN) or Global Atmospheric Watch (GAW) sites) there would be

437   full, high-quality measurements of atmospheric column properties, both physical and

438   chemical. This requires a large suite of instrumentation and redundancy in measurements[6].

439   These globally distributed reference sites should incorporate upward looking instruments

---

[6] Measurement of the same parameter by two or more independent instruments

440   (radar, lidar, GPS-related data, microwave sensors, wind profilers, etc.) along with high-

441   quality temperature, relative humidity and wind measurements on balloons regularly

442   penetrating well into the stratosphere[7] A key requirement is an end-to-end management

443   system including archiving of coincident observations made from over-flying satellites. The

444   data would be made openly available. The development of such a reference network is

445   recommended in outline by GCOS (2004). The ideas are currently being discussed in more

446   detail as part of an on-going process led by NOAA and WMO. Further details can be found at

447   http://www.oco.noaa.gov/workshop/.

448

449   **RECOMMENDATION** *Following Key Action 12[8] of the GCOS Implementation Plan*

450   *(GCOS, 2004), develop and implement a subset of about 5% of the operational radiosonde*

451   *network as reference network sites for all kinds of climate data from the surface to the*

452   *stratosphere.*

453
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[7] Recent inter-comparisons under the auspices of WMO suggest that new operational sondes are as accurate as
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6   **<u>APPENDIX: STATISTICAL ISSUES REGARDING TRENDS</u>**

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8   Tom M.L. Wigley

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20   **<u>Abstract:</u>**

21

22   The purpose of this Appendix is to explain the statistical terms and methods used in this Report.

23   We begin by introducing a number of terms: mean, standard deviation, variance, linear trend,

24   sample, population, signal, and noise. Examples are given of linear trends in surface,

25   tropospheric, and stratospheric temperatures. The least squares method for calculating a best fit

26   linear trend is described. The method for quantifying the statistical uncertainty in a linear trend is

27   explained, introducing the concepts of standard error, confidence intervals, and significance

28   testing. A method to account for the effects of temporal autocorrelation on confidence intervals

29   and significance tests is described. The issue of comparing two data sets to decide whether

30   differences in their trends could have occurred by chance is discussed. The analysis of trends in

31   state-of-the-art climate model results is a special case because we frequently have an ensemble of

32   simulations for a particular forcing case. The effect of ensemble averaging on confidence

33   intervals is illustrated. Finally, the issue of practical versus statistical significance is discussed. In

34   practice, it is important to consider construction uncertainties as well as statistical uncertainties.

35   An example is given showing that these two sources of trend uncertainty can be of comparable

36   magnitude.

37

38   **(1) Why do we need statistics?**

39

40   Statistical methods are required to ensure that data are interpreted correctly and that apparent

41   relationships are meaningful (or "significant") and not simply chance occurrences.

42

43   A "statistic" is a numerical value that describes some property of a data set. The most commonly

44   used statistics are the average (or "mean") value, and the "standard deviation", which is a

45   measure of the variability within a data set around the mean value. The "variance" is the square

46   of the standard deviation. The linear trend is another example of a data "statistic".

47

48   Two important concepts in statistics are the "population" and the "sample". The population is a

49   theoretical concept, an idealized representation of the set of all possible values of some measured

50   quantity. An example would be if we were able to measure temperatures continuously at a single

51   site for all time – the set of all values (which would be infinite in size in this case) would be the

52   population of temperatures for that site. A sample is what we actually see and can measure: i.e.,

53   what we have available for statistical analysis, and a necessarily limited subset of the population.

54   In the real world, all we ever have is limited samples, from which we try to estimate the

55   properties of the population.

56

57   As an analogy, the population might be an infinite jar of marbles, a certain proportion of which

58   (say 60%) is blue and the rest (40%) are red. We can only draw off a finite number of these

59   marbles (a sample) at a time; and, when we measure the numbers of blue and red marbles in the

60   sample, they need not be in the precise ratio 60:40. The ratio we measure is called a "sample

61   statistic". It is an estimate of some hypothetical underlying population value (the corresponding

62   "population parameter"). The techniques of statistical science allow us to make optimum use of

63   the sample statistic and obtain a best estimate of the population parameter. Statistical science

64   also allows us to quantify the uncertainty in this estimate.

65

66   **(2) Definition of a linear trend**

67

68   If data show underlying smooth changes with time, we refer to these changes as a trend. The

69   simplest type of change is a linear (or straight line) trend, a continuous increase or decrease over

70   time. For example, the net effect of increasing greenhouse-gas concentrations and other human-

71   induced factors is expected to cause warming at the surface and in the troposphere and cooling in

72   the stratosphere (see Figure 1). Warming corresponds to a positive (or increasing) linear trend,

73   while cooling corresponds to a negative (or decreasing) trend. These changes are not expected to

74   be strictly linear, but the linear trend provides a simple way of characterizing the change and of

75   quantifying its magnitude.

76

77
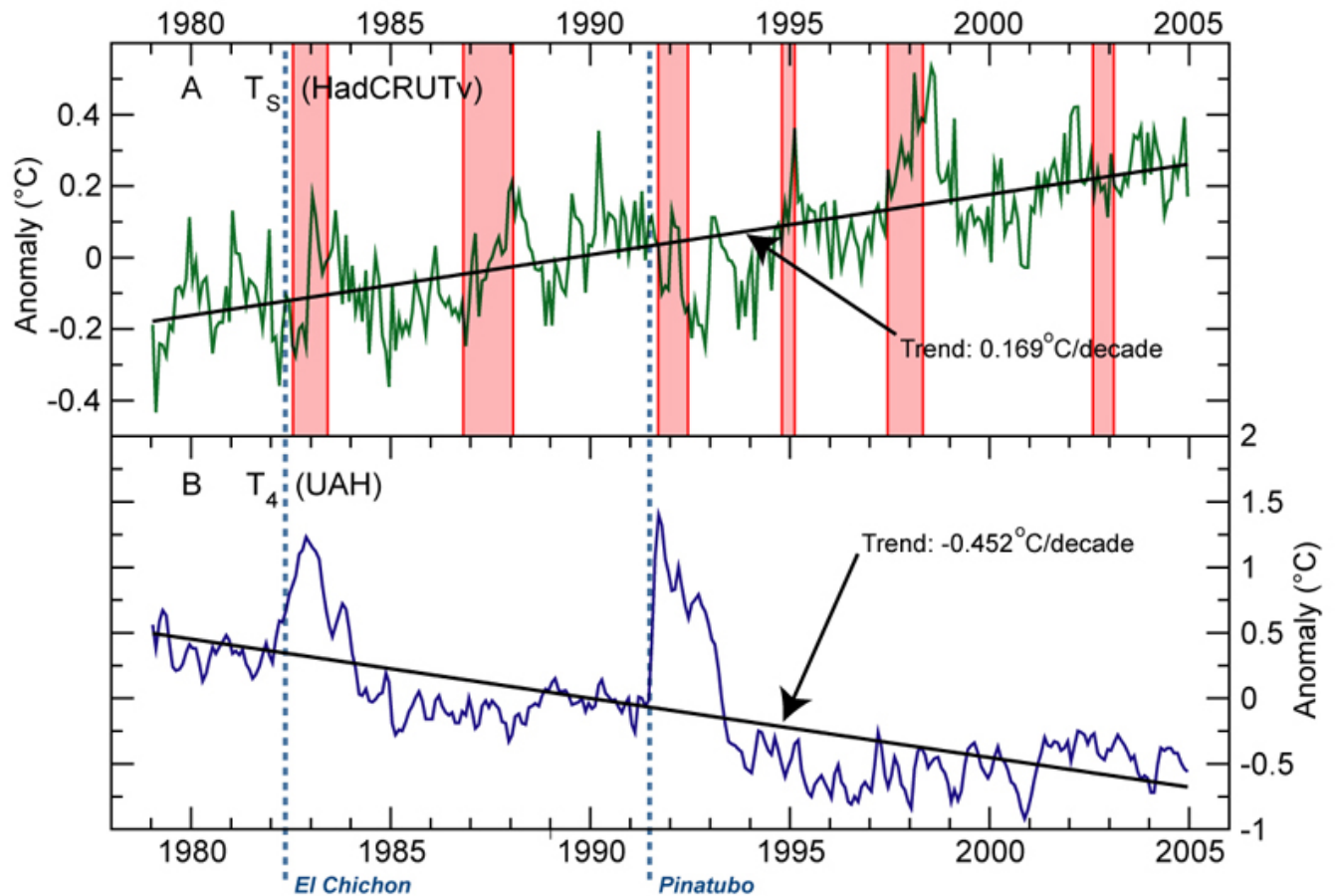
Figure 1: Examples of temperature time series with best-fit (least squares) linear trends: A, global-mean surface temperature from the UKMO Hadley Centre/Climatic Research Unit data set (HadCRUT2v); and B, MSU channel 4 data (T₄) for the lower stratosphere from the University of Alabama at Huntsville (UAH). Note the much larger temperature scale on the lower panel. Temperature changes are expressed as anomalies relative to the 1979 to 1999 mean (252 months). Dates for the eruptions of El Chichón and Mt Pinatubo are shown by vertical lines. El Niños are shown by the shaded areas.

Alternatively, there may be some physical process that causes a rapid switch or change from one mode of behavior to another. In such a case the overall behavior might best be described as a linear trend to the changepoint, a step change at this point, followed by a second linear trend portion. Many temperature data sets show this type of behavior, arising from a change in the pattern of variability in the Pacific that occurred around 1976 (a switch in a mode of climate variability called the Pacific Decadal Oscillation).

93

94   Step changes can lead to apparently contradictory results. For example, a data set that shows an

95   initial cooling trend, followed by a large upward step, followed by a renewed cooling trend could

96   have an overall warming trend. To state simply that the data showed overall warming would

97   misrepresent the true underlying behavior.

98

99   A linear trend may therefore be deceptive if the trend number is given in isolation, removed from

100   the original data. Nevertheless, used appropriately, linear trends provide the simplest and most

101   convenient way to describe the overall change over time in a data set, and are widely used.

102

103   Linear temperature trends are usually quantified as the temperature change per year or per

104   decade (even when the data are available on a month by month basis). For example, the trend for

105   the surface temperature data shown below in Figure 1 is 0.169^{o}C per decade. This is a more

106   convenient representation than the trend per month, which would be $0.169/120 = 0.00141^{o}$C per

107   month, a very small number. An alternative method is to use the "total trend" over the full data

108   period – i.e., the total change for the fitted line from the start to the end of the record (see Figure

109   2 in the Executive Summary). In Figure 1, the data shown span January 1979 through December

110   2004 (312 months or 2.6 decades). The total change is therefore $0.169 \times 2.6 = 0.439^{o}$C.

111

112 **(3) Expected temperature changes: signal and noise**

113

114 Different physical processes generally cause different spatial and temporal patterns of change.

115 For example, anthropogenic emissions of halocarbons at the surface have led to a reduction in

116 stratospheric ozone and a contribution to stratospheric cooling over the past three or four

117 decades. Now that these chemicals are controlled under the Montreal Protocol, the

118 concentrations of the controlled species are decreasing and there is a trend towards a recovery of

119 the ozone layer. The eventual long-term effect on stratospheric temperatures is expected to be

120 non-linear: a cooling up until the late 1990s followed by a warming as the ozone layer recovers.

121

122 This is not the only process affecting stratospheric temperatures. Increasing concentrations of

123 greenhouse gases lead to stratospheric cooling; and explosive volcanic eruptions cause sharp, but

124 relatively short-lived stratospheric warmings (see Figure 1)[1]. There are also natural variations,

125 most notably those associated with the Quasi-Bienniel Oscillation (QBO)[2]. Stratospheric

126 temperature changes (indeed, changes at all levels of the atmosphere) are therefore the combined

127 results of a number of different processes acting across all space and time scales.

128

129 In climate science, a primary goal is to identify changes associated with specific physical

130 processes (causal factors) or combinations of processes. Such changes are referred to as

131 "signals". Identification of signals in the climate record is referred to as the "detection and

132 attribution" (D&A) problem. "Detection" is the identification of an unusual change, through the

133 use of statistical techniques like significance testing (see below); while "attribution" is the

134 association of a specific cause or causes with the detected changes in a statistically rigorous way.

135

136   The reason why D&A is a difficult and challenging statistical problem is because climate signals

137   do not occur in isolation. In addition to these signals, temperature fluctuations in all parts of the

138   atmosphere occur even in the absence of external driving forces. These internally-driven

139   fluctuations represent the "noise" against which we seek to identify specific externally-forced

140   signals. All climate records, therefore, are "noisy", with the noise of this natural variability

141   tending to obscure the externally-driven changes. Figure 1 illustrates this. At the surface, a

142   primary noise component is the variability associated with ENSO (the El Niño/Southern

143   Oscillation phenomenon)[1], while, in the stratosphere, if our concern is to identify anthropogenic

144   influences, the warmings after the eruptions of El Chichón and Mt Pinatubo constitute noise.

145

146   If the underlying response to external forcing is small relative to the noise, then, by chance, we

147   may see a trend in the data due to random fluctuations purely as a result of the noise. The science

148   of statistics provides methods through which we can decide whether the trend we observe is

149   "real" (i.e., a signal associated with some causal factor) or simply a random fluctuation (i.e.,

150   noise).

151

152   **(4) Deriving trend statistics**

153

154   There are a number of different ways to quantify linear trends. Before doing anything, however,

155   we should always inspect the data visually to see whether a linear trend model is appropriate. For

156   example, in Fig. 1, the linear warming trend appears to be a reasonable description for the

157   surface data (top panel), but it is clear that a linear cooling model for the lower stratosphere

158   (lower panel) fails to capture some of the more complex changes that are evident in these data.

159   Nevertheless, the cooling trend line does give a good idea of the magnitude of the overall

160   change.

161

162   There are different ways to fit a straight line to the data. Most frequently, a "best fit" straight line

163   is defined by finding the particular line that minimizes the sum, over all data points, of the

164   squares of deviations about the line (these deviations are generally referred to as "residuals" or

165   "errors"). This is an example of a more general procedure called least squares regression.

166

167   In linear regression analysis, a predictand (Y) is expressed as a linear combination of one or

168   more predictors (X_i):

169

170        $Y_{est} = b_0 + b_1 X_1 + b_2 X_2 + \ldots$ ….. (1)

171

172   where the subscript 'est' is used to indicate that this is the estimate of Y that is given by the fitted

173   relationship. Differences between the actual and estimated values of Y, the residuals, are defined

174   by

175

176        $e = Y - Y_{est}$ ….. (2)

177

178   For linear trend analysis of temperature data (T) there is a single predictor, time (t; t = 1,2,3, …).

179   The time points are almost always evenly spaced, month by month, year by year, etc. – but this is

180   not a necessary restriction. In the linear trend case, the regression equation becomes:

181

182        $T_{est} = a + b\,t$ ….. (3)

183

184   In equ. (3), 'b' is the slope of the fitted line – i.e., the linear trend value. This is a sample statistic,

185   i.e., it is an estimate of the corresponding underlying population parameter. To distinguish the

186   population parameter from the sample value, the population trend value is denoted   .

187

188   The formula for b is:

189

190        $b = [\ \ ((t - <t>)T_t)]/[\ _.(t - <t>)^2)]$ ….. (4)

191

192   where <…> denotes the mean value, and the summation is over t = 1,2,3, … n (i.e., the sample

193   size is n). T_t denotes the value of temperature, T, at time 't'. Equation (4) produces an unbiased

194   estimate[3] of population trend,   .

195

196   For the usual case of evenly spaced time points, <t> = (n+1)/2, and

197

198        $(t - <t>)^2) = n(n^2 - 1)/12$ ….. (5)

199

200   When we are examining deviations from the fitted line the sign of the deviation is not important.

201   This is why we consider the squares of the residuals in least squares regression. An important

202   and desirable characteristic of the least squares method is that the average of the residuals is

203   zero.

204

205   Estimates of the linear trend are sensitive to points at the start or end of the data set. For

206   example, if the last point, by chance, happened to be unusually high, then the fitted trend might

207   place undue weight on this single value and lead to an estimate of the trend that was too high.

208   This is more of a problem with small sample sizes (i.e., for trends over short time periods). For

209   example, if we considered tropospheric data over 1979 through 1998, because of the unusual

210   warmth in 1998 (associated with the strong 1997/98 El Niño; see Figure 1), the calculated trend

211   may be an overestimate of the true underlying trend.

212

213   There are alternative ways to estimate the linear trend that are less sensitive to endpoints.

214   Although we recognize this problem, for the data used in this Report tests using different trend

215   estimators give results that are virtually the same as those based on the standard least-squares

216   trend estimator.

217

218

218   **(5) Trend uncertainties**

219

220   Some examples of fitted linear trend lines are shown in Figure 1. This Figure shows monthly

221   temperature data for the surface and for the lower stratosphere (MSU channel 4) over 1979

222   through 2004 (312 months). In both cases there is a clear trend, but the fit is better for the surface

223   data. The trend values (i.e., the slopes of the best fit straight lines that are shown superimposed

224   on monthly data) are $+0.169^{\circ}$C/decade for the surface and -0.452°C/decade for the stratosphere.

225   For the stratosphere, although there is a pronounced overall cooling trend, as noted above

226   describing the change simply as a linear cooling considerably oversimplifies the behavior of the

227   data[1].

228

229   A measure of how well the straight line fits the data (i.e., the "goodness of fit") is the average

230   value of the squares of the residuals. The smaller this is, the better is the fit. The simplest way to

231   define this average would be to divide the sum of the squares of the residuals by the sample size

232   (i.e., the number of data points, n). In fact, it is usually considered more correct to divide by n – 2

233   rather than n, because some information is lost as a result of the fitting process and this loss of

234   information must be accounted for. Dividing by n – 2 is required in order to produce an unbiased

235   estimator.

236

237   The population parameter we are trying to estimate here is the standard deviation of the trend

238   estimate, or its square, the variance of the distribution of b, which we denote Var(b). The larger

239   the value of Var(b), the more uncertain is b as an estimate of the population value,   .

240

241   The formula for Var(b) is …

242

243        $Var(b) = [\sigma^2]/[\sum (t - <t>)^2)]$ ….. (6)

244

245   where σ^2 is the population value for the variance of the residuals. Unfortunately, we do not in

246   general know what σ^2 is, so we must use an unbiased sample estimate of σ^2. This estimate is

247   known as the Mean Square Error (MSE), defined by …

248

249        $MSE = [\sum (e^2)]/(n - 2)$ ….. (7)

250

251   Hence, equ. (6) becomes

252

253        $Var(b) = (SE)^2 = MSE/[\sum (t - <t>)^2)]$ ….. (8)

254

255   where SE, the square root of Var(b), is called is called the "standard error" of the trend estimate.

256   The smaller the value of the standard error, the better the fit of the data to the linear change

257   description and the smaller the uncertainty in the sample trend as an estimate of the underlying

258   population trend value. The standard error is the primary measure of trend uncertainty. The

259   standard error will be large if the MSE is large, and the MSE will be large if the data points show

260   large scatter about the fitted line.

261

262   There are assumptions made in going from equ. (6) to (8): viz. that the residuals have mean zero

263   and common variance, that they are Normally (or "Gaussian") distributed[4], and that they are

264   uncorrelated or statistically independent. In climatological applications, the first two are

265   generally valid. The third assumption, however, is often not justified. We return to this below.

266

267   **(6) Confidence intervals and significance testing**

268

269   In statistics we try to decide whether a trend is an indication of some underlying cause, or merely

270   a chance fluctuation. Even purely random data may show periods of noticeable upward or

271   downward trends, so how do we identify these cases?

272

273   There are two common approaches to this problem, through significance testing and by defining

274   confidence intervals. The basis of both methods is the determination of the "sampling

275   distribution" of the trend, i.e., the distribution of trend estimates that would occur if we analyzed

276   data that were randomly scattered about a given straight line with slope   . This distribution is

277   approximately Gaussian with a mean value equal to   and a variance (standard deviation

278   squared) given by equ. (8). More correctly, the distribution to use is Student's 't' distribution,

279   named after the pseudonym 'Student' used by the statistician William Gosset. For large samples,

280   however (n more than about 30), the distribution is very nearly Gaussian.

281

282   *Confidence intervals*

283

284   The larger the standard error of the trend, the more uncertain is the slope of the fitted line. We

285   express this uncertainty probabilistically by defining confidence intervals for the trend associated

286   with different probabilities. If the distribution of trend values were strictly Gaussian, then the

287   range b – SE to b + SE would represent the 68% confidence interval (C.I.) because the

288   probability of a value lying in that range for a Gaussian distribution is 0.68. The range b –

289   1.645(SE)to b + 1.645(SE) would give the 90% C.I.; the range b – 1.96(SE)to b + 1.96(SE)

290   would give the 95% C.I.; and so on. Quite often, for simplicity, we use b – 2(SE)to b + 2(SE) to

291   represent (to a good approximation) the 95% confidence interval.

292

293   Because of the way C.I.s are usually represented graphically, as a bar centered on the best-fit

294   estimate, they are often referred to as "error bars". Confidence intervals may be expressed in two

295   ways, either (as above) as a range, or as a signed error magnitude. The approximate 95%

296   confidence interval, therefore, may be expressed as b ± 2(SE), with appropriate numerical values

297   inserted for b and SE.

298

299   As will be explained further below, showing confidence interval for linear trends may be

300   deceptive, because the purely statistical uncertainties that they represent are not the only sources

301   of uncertainty. Such confidence intervals quantify only one aspect of trend uncertainty, that

302   arising from statistical noise in the data set. There are many other sources of uncertainty within

303   any given data set and these may be as or more important than statistical uncertainty. Showing

304   just the statistical uncertainty may therefore provide a false sense of accuracy in the calculated

305   trend.

306

307   ***Significance testing***

308

309   An alternative method for assessing trends is hypothesis testing. In practice, it is much easier to

310   disprove rather than prove a hypothesis. Thus, the standard statistical procedure in significance

311   testing is to set up a hypothesis that we would like to disprove. This is called a "null hypothesis".

312   In the linear trend case, we are often interested in trying to decide whether an observed data trend

313   that is noticeably different from zero is sufficiently different that it could not have occurred by

314   chance – or, at least, that the probability that it could have occurred by chance is very small. The

315   appropriate null hypothesis in this case would be that there was no underlying trend (= 0). If

316   we disprove (i.e., "reject") the null hypothesis, then we say that the observed trend is

317   "statistically significant" at some level of confidence and we must accept some alternate

318   hypothesis. The usual alternate hypothesis in temperature analyses is that the data show a real,

319   externally-forced warming (or cooling) trend. (In cases like this, the statistical analysis is

320   predicated on the assumption that the observed data are reliable. If a trend were found to be

321   statistically significant, then an alternative possibility might be that the observed data were

322   flawed.)

323

324   An alternative null hypothesis that often arises is when we are comparing an observed trend with

325   some model expectation. Here, the null hypothesis is that the observed trend is equal to the

326   model value. If our results led us to reject this null hypothesis, then (assuming again that the

327   observed data are reliable) we would have to infer that the model result was flawed – either

328   because the external forcing applied to the model was incorrect and/or because of deficiencies in

329   the model itself.

330

331   An important factor in significance testing is whether we are concerned about deviations from

332   some hypothesized value in any direction or only in one direction. This leads to two types of

333   significance test, referred to as "one-tailed" (or "one-sided") and "two-tailed" tests. A one-tailed

334   test arises when we expect a trend in a specific direction (such as warming in the troposphere due

335   to increasing greenhouse-gas concentrations). Two-tailed tests arise when we are concerned only

336   with whether the trend is different from zero, with no specification of whether the trend should

337   be positive or negative. In temperature trend analyses we generally know the sign of the expected

338   trend, so one-tailed tests are more common.

339

340   The approach we use in significance testing is to determine the probability that the observed

341   trend could have occurred by chance. As with the calculation of confidence intervals, this

342   involves calculating the uncertainty in the fitted trend arising from the scatter of points about the

343   trend line, determined by the standard error of the trend estimate (equ. (8)). It is the ratio of the

344   trend to the standard error (b/SE) that determines the probability that a null hypothesis is true or

345   false. A large ratio (greater than 2, for example) would mean that (except for very small samples)

346   the 95% C.I. did not include the zero trend value. In this case, the null hypothesis is unlikely to

347   be true, because the zero trend value, the value assumed under the null hypothesis, lies outside

348   the range of trend values that are likely to have occurred purely by chance.

349

350   If the probability that the null hypothesis is true is small, and less than a predetermined threshold

351   level such as 0.05 (5%) or 0.01 (1%), then the null hypothesis is unlikely to be correct. Such a

352   low probability would mean that the observed trend could only have occurred by chance one

353   time in 20 (or one time in 100), a highly unusual and therefore "significant" result. In technical

354   terms we would say that "the null hypothesis is rejected at the prescribed significance level", and

355   declare the result "significant at the 5% (or 1%) level". We would then accept the alternate

356   hypothesis that there was a real deterministic trend and, hence, some underlying causal factor.

357

358   Even with rigorous statistical testing, there is always a small probability that we might be wrong

359   in rejecting a null hypothesis. The reverse is also true – we might accept a null hypothesis of no

360   trend even when there is a real trend in the data. This is more likely to happen when the sample

361   size is small. If the real trend is small and the magnitude of variability about the trend is large, it

362   may require a very large sample in order to identify the trend above the background noise.

363

364   For the null hypothesis of zero trend, the distribution of trend values has mean zero and standard

365   deviation equal to the standard error. Knowing this, we can calculate the probability that the

366   actual trend value could have exceeded the observed value by chance if the null hypotheses were

367   true (or, if we were using a two-tailed test, the probability that the magnitude of the actual trend

368   value exceeded the magnitude of the observed value). This probability is called the 'p-value'. For

369   example, a p-value of 0.03 would be judged significant at the 5% level (since 0.03<0.05), but not

370   at the 1% level (since 0.03>0.01).

371

372   Since both the calculation of confidence intervals and significance testing employ information

373   about the distribution of trend values, there is a clear link between confidence intervals and

374   significance testing.

375

376   *A complication; the effect of autocorrelation*

377

378   The significance of a trend, and its confidence intervals, depend on the standard error of the trend

379   estimate. The formula given above for this standard error (equ. (8)) is, however, only correct if

380   the individual data points are unrelated, or statistically independent. This is not the case for most

381   temperature data, where a value at a particular time usually depends on values at previous times;

382   i.e., if it is warm today, then, on average, it is more likely to be warm tomorrow than cold. This

383   dependence is referred to as "temporal autocorrelation" or "serial correlation". When data are

384   autocorrelated (i.e., when successive values are not independent of each other), many statistics

385   behave as if the sample size was less than the number of data points, n.

386

387   One way to deal with this is to determine an "effective sample size", which is less than n, and

388   use it instead of n in statistical formulae and calculations. The extent of this reduction from n to

389   an effective sample size depends on how strong the autocorrelation is. Strong autocorrelation

390   means that individual values in the sample are far from being independent, so the effective

391   number of independent values must be much smaller than the sample size. Strong autocorrelation

392   is common in temperature time series. This is accounted for by reducing the divisor 'n – 2' in the

393   mean square error term (equ. (7)) that is crucial in determining the standard error of the trend

394   (equ. (8)).

395

396   There are a number of ways that this autocorrelation effect may be quantified. A common and

397   relatively simple method is described in Santer et al. (2000). This method makes the assumption

398   that the autocorrelation structure of the temperature data may be adequately described by a "first-

399   order autoregressive" process, an assumption that is a good approximation for most climate data.

400   The lag-1 autocorrelation coefficient (r_1) is calculated from the observed data[5], and the effective

401   sample size is determined by

402

403        $n_{eff} = n (1 - r_1)/(1 + r_1)$ ….. (9)

404

405   There are more sophisticated methods than this, but testing on observed data shows that this

406   method gives results that are very similar to those obtained by more sophisticated methods.

407

408   If the effective sample size is noticeably smaller than n, then, from equs. (7) and (8) it can be

409   seen that the standard error of the trend estimate may be much larger than one would otherwise

410   expect. Since the width of any confidence interval depends directly on this standard error (larger

411   SE leading to wider confidence intervals), then the effect of autocorrelation is to produce wider

412   confidence intervals and greater uncertainty in the trend estimate. A corollary of this is that

413   results that may show a significant trend if autocorrelation is ignored are frequently found to be

414   non-significant when autocorrelation is accounted for.

415

416   **(7) Comparing trends in two data sets**

417

418   Assessing the magnitude and confidence interval for the linear trend in a given data set is

419   standard procedure in climate data analysis. Frequently, however, we want to compare two data

420   sets and decide whether differences in their trends could have occurred by chance. Some

421   examples are:

422

423   (a) comparing data sets that purport to represent the same variable (such as two versions of a

424   satellite data set) – an example is given in Figure 2;
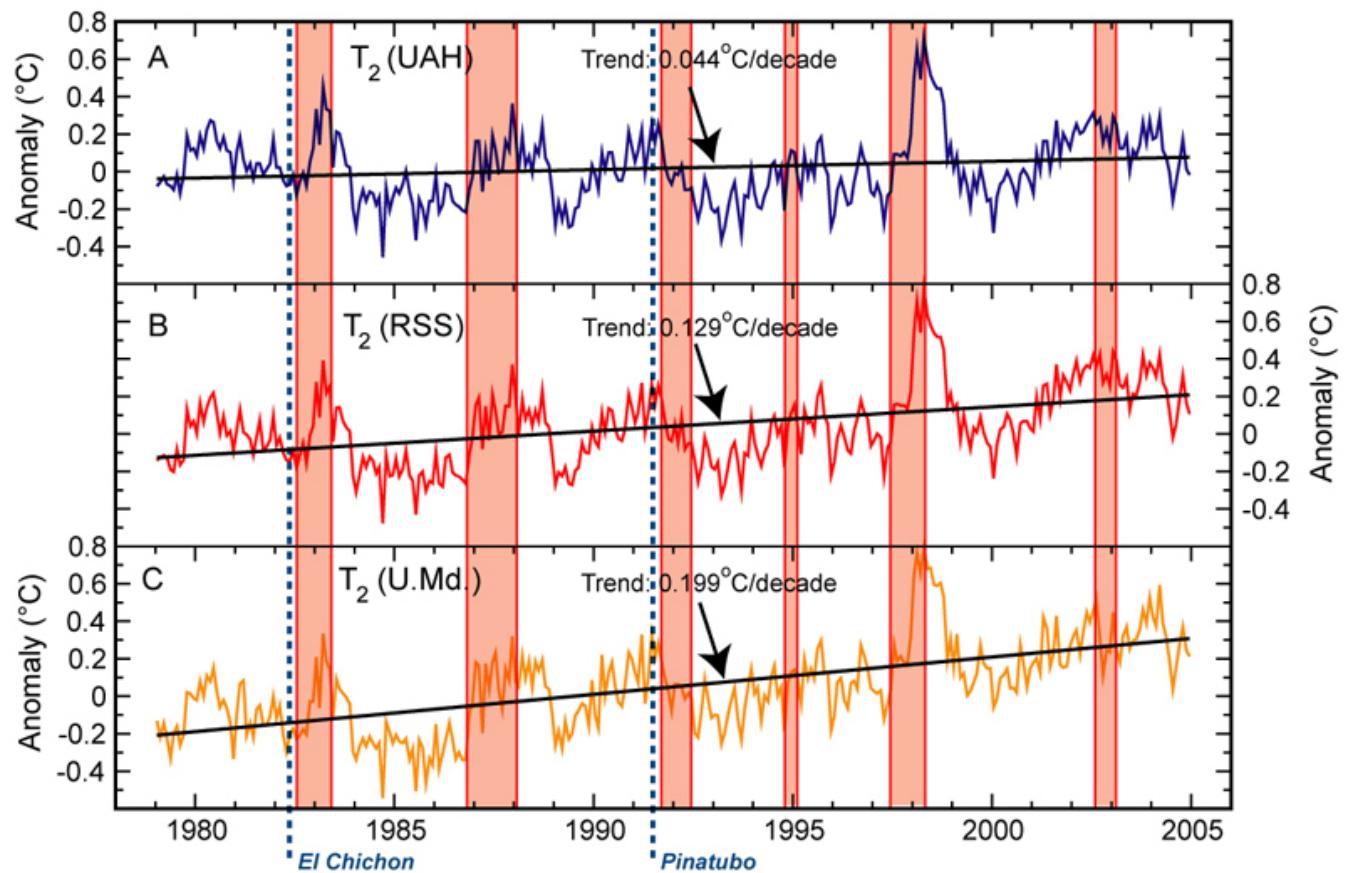
425   (b) comparing the same variable at different levels in the atmosphere (such as surface and

426   tropospheric data); or

427   (c) comparing models and observations.

428

429

430

Figure 2: Three estimates of temperature changes for MSU channel 2 (T_2), expressed as anomalies relative to the 1979 to 1999 mean. Data are from: A, the University of Alabama at Huntsville (UAH); B, Remote Sensing Systems (RSS); and C, the University of Maryland (U.Md.) The estimates employ the same 'raw' satellite data, but make different choices for the adjustments required to merge the various satellite records and to correct for instrument biases. The statistical uncertainty is virtually the same for all three series. Differences between the series give some idea of the magnitude of structural uncertainties. Volcano eruption and El Niño information are as in Figure 1.

In the first case (Figure 2), we know that the data sets being compared are attempts to measure

precisely the same thing, so that differences can arise only as a result of differences in the

methods used to create the final data sets from the same 'raw' original data. Here, there is a

pitfall that some practitioners fall prey to by using what, at first thought, seems to be a

reasonable approach. In this naïve method, one would first construct C.I.s for the individual trend

estimates by applying the single sample methods described above. If the two C.I.s overlapped,

446   then we would conclude that there was no significant difference between the two trends. This

447   approach, however, is seriously flawed.

448

449   An analogous problem, comparing two means rather than two trends, discussed by Lanzante

450   (2005), gives some insights. In this case, it is necessary to determine the standard error for the

451   difference between two means. If this standard error is denoted 's', and the individual standard

452   errors are s_1 and s_2, then

453

454        $s^2 = (s_1)^2 + (s_2)^2$ …..(10)

455

456   The new standard error is often called the pooled standard error, and the pooling method is

457   sometimes called "combining standard errors in quadrature". In some cases, when the trends

458   come from data series that are unrelated (as in the model/observed data comparison case; (c)

459   above) a similar method may be applied to trends. If the data series are correlated with each

460   other, however (cases (a) and (b)), this procedure is not correct. Here, the correct method is to

461   produce a difference time series by subtracting the first data point in series 1 from the first data

462   point in series 2, the second data points, the third data points, etc. The result of doing this with

463   the microwave sounding unit channel 2 (MSU T_2) data shown in Figure 2 is shown in Figure 3.

464   To assess the significance of trend differences we then apply the same methods used for trend

465   assessment in a single data series to the difference series.

466

467
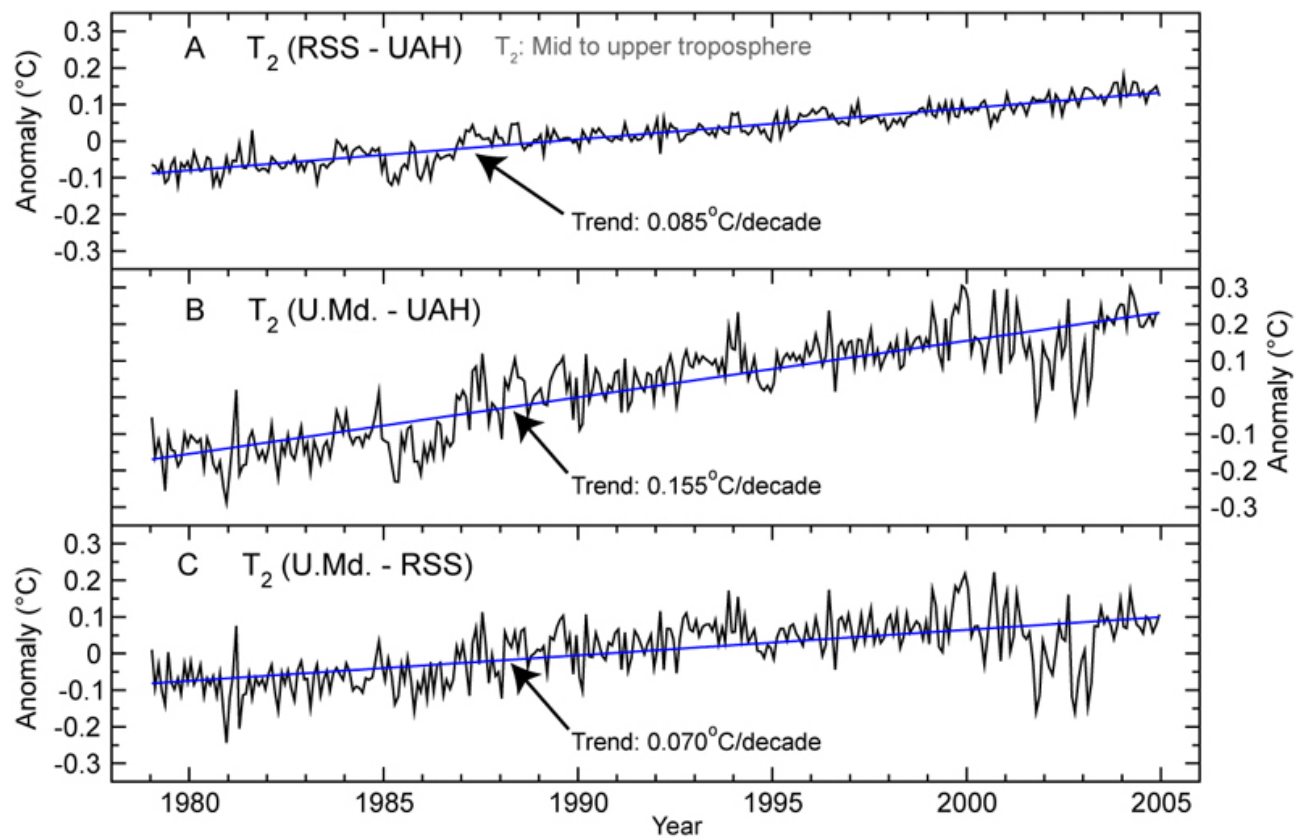
469 Figure 3: Difference series for the MSU T_2 series shown in Figure 2. Variability about the trend line is least for the
470 UAH minus RSS series indicating closer correspondence between these two series than between U.Md. and either
471 UAH or RSS.

472

473

474 Analyzing differences removes the variability that is common to both data sets and isolates those

475 differences that may be due to differences in data set production methods, temperature

476 measurement methods (as in comparing satellite and radiosonde data), differences in spatial

477 coverage, etc.

478

479 Figures 2 and 3 provide a striking example of this. Here, the three series in Figure 2 have very

480 similar volcanic and ENSO signatures. In the individual series, these aspects are noise that

481 obscures the underlying linear trend and inflates the standard error and the trend uncertainty.

482   Since this noise is common to each series, differencing has the effect of canceling out a large

483   fraction of the noise. This is clear from Figure 3, where the variability about the trend lines is

484   substantially reduced. Figure 4 shows the effects on the trend confidence intervals (taking due

485   account of autocorrelation effects). Even though the individual series look very similar in Figure

486   2, this is largely an artifact of similarities in the noise. It is clear from Figures 3 and 4 that there

487   are, in fact, very significant differences in the trends, reflecting differences in their methods of

488   construction.

489

490
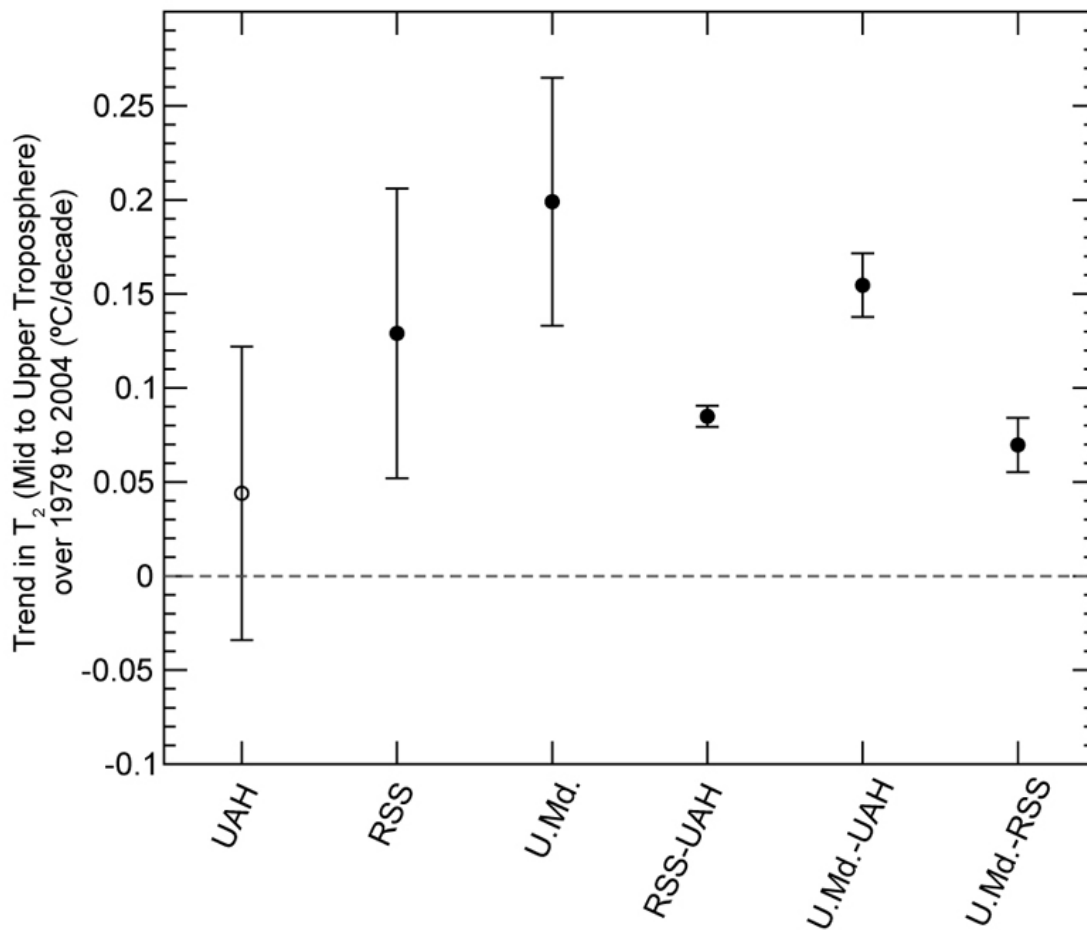
491

492   Figure 4: 95% confidence intervals for the three MSU T_2 series shown in Figure 2 (see Table 3.3 in Chapter 3), and
493   for the three difference series shown in Figure 3.
494

495

496   Comparing model and observed data for a single variable, such as surface temperature,

497   tropospheric temperature, etc., is a different problem. Here, when using data from a state-of-the-

498   art climate model (a coupled Atmosphere/Ocean General Circulation Model[6], or "AOGCM"),

499   there is no reason to expect the background variability to be common to both the model and

500   observations. AOGCMs generate their own internal variability entirely independently of what is

501   going on in the real world. In this case, standard errors for the individual trends can be combined

502   in quadrature (equ. (10). (There are some model/observed data comparison cases where an

503   examination of the difference series may still be appropriate, such as in experiments where an

504   atmospheric GCM is forced by observed sea surface temperature variations so that ocean-related

505   variability should be common to both the observations and the model.)

506

507   For other comparisons, the appropriate test will depend on the degree of similarity between the

508   data sets expected for perfect data. For example, a comparison between MSU T_2 and MSU T_{2LT}

509   produced by a single group should use the difference test – although interpretation of the results

510   may be tricky because differences may arise either from construction methods or may represent

511   real physical differences arising from the different vertical weighting profiles, or both.

512

513   There is an important implication of this comparison issue. While it may be common practice to

514   use error bars to illustrate C.I.s for trends of individual time series, when the primary concern (as

515   it is in many parts of this Report) is the comparison of trends, individual C.I.s can be quite

516   misleading. In some cases in this Report, therefore, where it might seem that error bars should be

517   given, we consider the disadvantage of their possible misinterpretation to outweigh their

518   potential usefulness. Instead, we have chosen to express individual trend uncertainties through

519   the use of significance levels, which can be represented by a less obtrusive symbol. As noted in

520   Section (9) below, there are other reasons why error bars can be misleading.

521

522   **(8) Multiple AOGCM simulations**

523

524   Both models and the real world show weather variability and other sources of internal variability

525   that are manifest on all time scales, from daily up to multi-decadal. With AOGCM simulations

526   driven by historical forcing spanning the late-19th and 20th Centuries, therefore, a single run with

527   a particular model will show not only the externally-forced signal, but also, superimposed on

528   this, underlying internally-generated variability that is similar to the variability we see in the real

529   world. In contrast to the real world, however, in the model world we can perturb the model's

530   initial conditions and re-run the same forcing experiment. This will give an entirely different

531   realization of the model's internal variability. In each case, the output from the model is a

532   combination of signal (the response to the forcing) and noise (the internally-generated

533   component). Since the noise parts of each run are unrelated, averaging over a number of

534   realizations will tend to cancel out the noise and, hence, enhance the visibility of the signal. It is

535   common practice, therefore, for any particular forcing experiment with an AOGCM, to run

536   multiple realizations of the experiment (i.e., an ensemble of realizations). An example is given

537   in Figure 5, which shows four separate realizations and their ensemble average for a simulation

538   using realistic 20th Century forcing (both natural and anthropogenic).

539

540

541
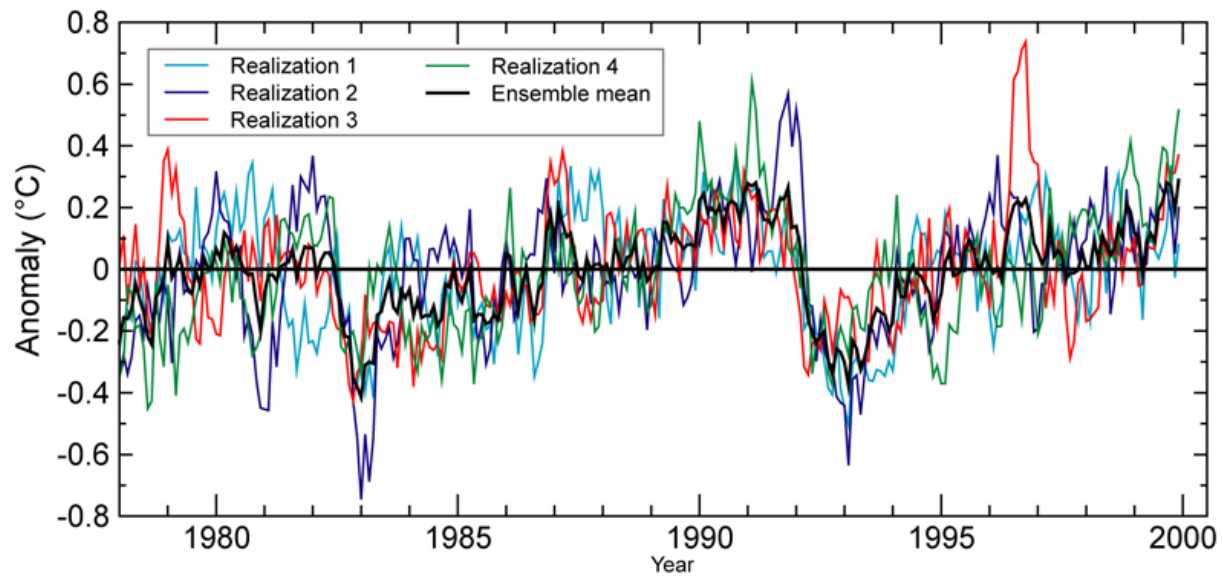
Figure 5: Four separate realizations of model realizations of global-mean MSU channel 2 (T_2) temperature changes,
and their ensemble average, for a simulation using realistic 20th Century forcing (both natural and anthropogenic)
carried out with one of the National Centre for Atmospheric Research's AOGCMs, the Parallel Climare Model
(PCM). The cooling events around 1982/3 and 1991/2 are the result of imposed forcing from the eruptions of El
Chichón (1982) and Mt. Pinatubo (1991). Note that the El Chichón cooling is more obvious than in the observed
data shown in Fig. 1, because, in the model simulations, the ENSO sequences differed from the real world, and from
each other.

This provides us with two different ways to assess the uncertainties in model results, such as in

the model-simulated temperature trend over recent decades. One method is to express

uncertainties using the spread of trends across the ensemble members (see, e.g., Figures 3 and 4

in the Executive Summary). Alternatively, the temperature series from the individual ensemble

members may be averaged and the trend and its uncertainty calculated using these average data.

Ensemble averaging, however, need not reduce the width of the trend confidence interval

compared with an individual realization. This is because of compensating factors: the time series

variability will be reduced by the averaging process (as is clear in Figure 5), but, because

560   averaging can inflate the level of autocorrelation, there may be a compensating increase in

561   uncertainty due to a reduction in the effective sample size. This is illustrated in Figure 6.

562
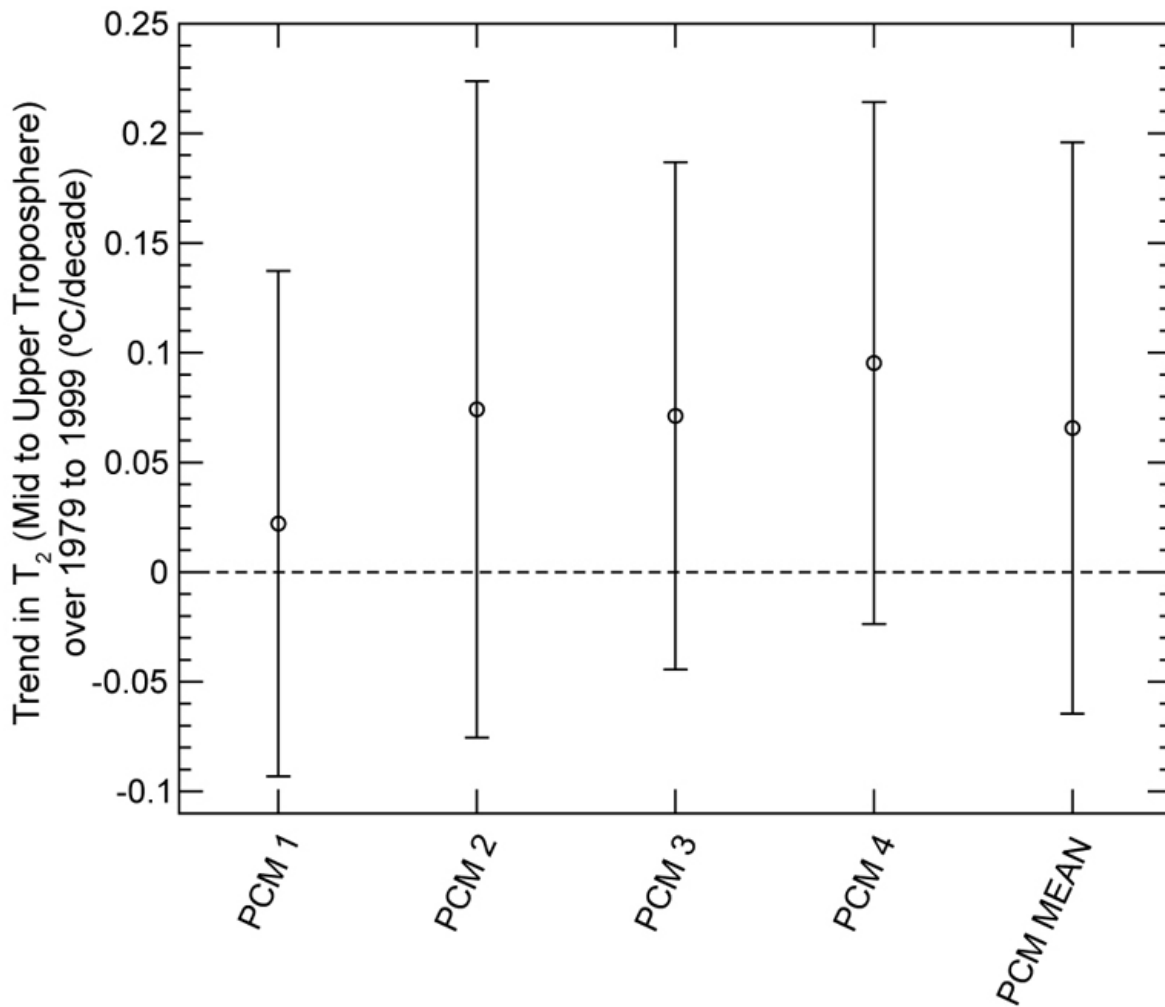
563

564   Figure 6: 95% confidence intervals for individual model realizations of MSU T_2 temperature changes (as shown in
565   Fig. 5), compared with the 95% confidence interval for the ensemble (n=4) average.
566

567   Averaging across ensemble members, however, does produce a net gain. Although the width of

568   the C.I. about the mean trend may not be reduced relative to individual trend C.I.s, averaging

569   leaves just a single best-fit trend rather than a spread of best-fit trend values.

570   **(9) Practical versus statistical significance**

571

572   The Sections above have been concerned primarily with statistical uncertainty, uncertainty

573   arising from random noise in climatological time series – i.e., the uncertainty in how well a data

574   set fits a particular 'model' (a straight line in the linear trend case). Statistical noise, however, is

575   not the only source of uncertainty in assessing trends. Indeed, as amply illustrated in this Report,

576   other sources of uncertainty may be more important.

577

578   The other sources of uncertainty are the influences of non-climatic factors. These are referred to

579   in this Report as "construction uncertainties". When we construct climate data records that are

580   going to be used for trend analyses, we attempt to minimize construction uncertainties by

581   removing, as far as possible, non-climatic biases that might vary over time and so impart a

582   spurious trend or trend component – a process referred to as "homogenization".

583

584   The need for homogenization arises in part because most observations are made to serve the

585   short-term needs of weather forecasting (where the long-term stability of the observing system is

586   rarely an important consideration). Most records therefore contain the effects of changes in

587   instrumentation, instrument exposure, and observing practices made for a variety of reasons.

588   Such changes generally introduce spurious non-climatic changes into data records that, if not

589   accounted for, can mask (or possibly be mistaken for) an underlying climate signal.

590

591   An added problem arises because temperatures are not always measured directly, but through

592   some quantity related to temperature. Adjustments must therefore be made to obtain temperature

593   information. The satellite-based microwave sounding unit (MSU) data sets provide an important

594   example. For MSU temperature records, the quantity actually measured is the upwelling

595   emission of microwave radiation from oxygen atoms in the atmosphere. MSU data are also

596   affected by numerous changes in instrumentation and instrument exposure associated with the

597   progression of satellites used to make these measurements.

598

599   Thorne et al. (2005) divide construction uncertainty into two components: "structural

600   uncertainty" and "parametric uncertainty". Structural uncertainty arises because there is no *a*

601   *priori* knowledge of the correct way to homogenize a given raw data set. Independent

602   investigators given the same raw data will make different seemingly sensible and defensible

603   adjustment choices based on their training, technological options at their disposal, and their

604   understanding of the raw data, amongst other factors. Differences in the choice of adjustment

605   pathway and its structure lead to structural uncertainties. Parametric uncertainty arises because,

606   once an adjustment approach or pathway has been chosen, additional choices may have to be

607   made with regard to specific correction factors or parameters.

608

609   Sensitivity studies using different parameter choices may allow us to quantify parametric

610   uncertainty, but this is not always done. Quantifying structural uncertainty is very difficult

611   because it involves consideration of a number of fundamentally different (but all plausible)

612   approaches to data set homogenization, rather than simple parameter "tweaking". Differences

613   between results from different investigators give us some idea of the magnitude of structural

614   uncertainty, but this is a relatively weak constraint. There are a large number of conceivable

615   approaches to homogenization of any particular data set, from which we are able only to consider

616   a small sample – and this may lead to an under-estimation of structural uncertainty. Equally, if

617   some current homogenization techniques are flawed then the resulting uncertainty estimate will

618   be too large.

619

620   An example is given above in Figure 2, showing three different MSU T_2 records with trends of

621   0.044^{o}C/decade, 0.129^{o}C/decade, and 0.199^{o}C/decade over 1979 through 2004. These

622   differences, ranging from 0.070^{o}C/decade to 0.155^{o}C/decade, represent a considerable degree of

623   construction uncertainty. For comparison, the statistical uncertainty, which is very similar for

624   each series and which can be quantified by the 95% confidence interval, is ±0.066 to

625   ±0.078^{o}C/decade.

626

627   An important implication of this comparison is that statistical and construction uncertainties may

628   be of similar magnitude. For this reason, showing, through confidence intervals, information

629   about statistical uncertainty alone, without giving any information about construction

630   uncertainty, can be misleading.

631   **Footnotes**

632

633   [1] Figure 1 shows a number of interesting features. In the stratosphere, the warmings following

634   the eruptions of El Chichón (April 1982) and Mt Pinatubo (June 1991) are pronounced. For El

635   Chichón, the warming appears to start before the eruption, but this is just a chance natural

636   fluctuation. The overall cooling trend is what is expected to occur due to anthropogenic

637   influences. At the surface, on short time scales, there is a complex combination of effects. There

638   is no clear cooling after El Chichón, primarily because this was offset by the very strong 1982/83

639   El Niño. Cooling after Pinatubo is more apparent, but this was also partly offset by the El Niño

640   around 1992/93 (which was much weaker than that of 1982/83). El Niño events, characterized by

641   warm temperatures in the tropical Pacific, have a noticeable effect on global-mean temperature,

642   but the effect lags behind the Pacific warming by 3-7 months. This is very clear in the surface

643   temperature changes at and immediately after the 1986/87 and 1997/98 El Niños, also very large

644   events. The most recent El Niños were weak and have no clear signature in the surface

645   temperatures.

646

647   [2] The QBO is a quasi-periodic reversal in winds in the tropical stratosphere that leads to

648   alternating warm and cold tropical stratospheric temperatures with a periodicity of 18 to 30

649   months.

650

651   [3] An unbiased estimator is one where, if the same experiment were to be performed over and

652   over again under identical conditions, then the long-run average of the estimator will be equal to

653   the parameter that we are trying to estimate. In contrast, in a biased estimator, there will always

654   be some slight difference between the long-run average and the true parameter value that does

655   not tend to zero no matter how many times the experiment is repeated. Since our goal is to

656   estimate population parameters, it is clear that unbiased estimators are preferred.

657

658   [4] The "Gaussian" distribution (often called the "Normal" distribution) is the most well-known

659   probability distribution. This has a characteristic symmetrical "bell" shape, and has the property

660   that values near the center (or mean value) of the distribution are much more likely than values

661   far from the center.

662

663   [5] From the time series of residuals about the fitted line.

664

665   [6] An AOGCM interactively couples together a three-dimensional ocean General Circulation

666   Model (GCM) and an atmospheric GCM (AGCM). The components are free to interact with one

667   another and they are able to generate their own internal variability in much the same way that the

668   real-world climate system generates its internal variability (internal variability is variability that

669   is unrelated to external forcing). This differs from some other types of model (e.g, an AGCM)

670   where there can be no component of variability arising from the ocean. An AGCM, therefore,

671   cannot generate variability arising from ENSO, which depends on interactions between the

672   atmosphere and ocean.

673

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686

Appendix B

Climate Change Science Program
Temperature Trends in the Lower Atmosphere
Steps for Understanding and Reconciling Differences

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Chapters:

1. Why do temperatures vary vertically (from the surface to the stratosphere) and what do we understand about why they might vary and change over time?
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2. What kinds of atmospheric temperature variations can the current observing systems measure and what are their strengths and limitations, both spatially and temporally?
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3. What do observations indicate about the changes of temperature in the atmosphere and at the surface since the advent of measuring temperatures vertically?
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4. What is our understanding of the contribution made by observational or methodological uncertainties to the previously reported vertical differences in temperature trends?
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5. How well can the observed vertical temperature changes be reconciled with our understanding of the causes of these temperature changes?
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6. What measures can be taken to improve the understanding of observed changes?
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Appendix
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Glossary & Acronyms, Symbols, Abbreviations

Glossary

Aerosols   tiny particles suspended in the air

Adjusted   refers to time series data that has been homogenized for time
                            dependent biases; owing to uncertainties inherent in data bias
                            removal, the term "adjusted" is often used instead of "corrected"

Albedo   the fraction of incident light that is reflected

Anthropogenic   human-induced

Black carbon   soot particles primarily from fossil fuel burning

Climate sensitivity: the equilibrium change in global-average surface air temperature
                            following a change in radiative forcing; in current usage, this term
                            generally refers to the warming that would result if atmospheric
                            carbon dioxide concentrations were to double from their pre-
                            industrial levels

Contrails   condensation trails from aircraft

Convection   motions in a fluid or the air that are predominantly vertical and
                            driven by buoyancy forces; a principal means of vertical energy
                            transfer

Diurnal   occurring daily; varying within the course of a day

Dewpoint   temperature at which water vapor condenses into liquid water
temperature   when cooled at constant pressure

Error   the difference between an estimated or observed value and the true
                            value

Forcing   a factor that influences climate

Greenhouse gases   gases including water vapor, carbon dioxide, methane, nitrous
                            oxide, and halocarbons that trap infrared heat, warming the air near
                            the surface and lower levels of the atmosphere

Homogenization   Removing changes in time series data that might have arisen for
                            non-climatic reasons

Internal variability   natural cycles and variations in climate

47 **Temperature inversion**
48   a condition in which the air temperature increases with height, in
49   contrast to the more common situation in which temperature
50   decreases with altitude
51
52 **Isothermal**   constant temperature; often refers to a temperature profile meaning
53   constant temperature with height
54
55 **Lapse rate**   the rate at which temperature decreases with increasing elevation
56
57 **Latent heat**   the heat required to change the phase of a substance (solid to
58   liquid, liquid to vapor or solid to vapor); the temperature does not
59   change during this process; the heat is released for the reverse
60   process
61
62 **Latent heat of water** the energy released or gained by water during phase changes
63
64 **Metadata**   supplemental records used to interpret measurements, such as how
65   and where measurements were collected and processed
66
67 **Moist enthalpy**   sensible and latent heat content of the air; includes contributions
68   from temperature and from the amount of water vapor expressed as
69   specific humidity
70
71 **Parameterization**   a mathematical representation of a process that cannot be explicitly
72   resolved in a climate model
73
74 **Radiosonde**   a balloon carrying a thermometer or other sensing device that takes
75   measurements in the atmosphere and transmits them by radio to a
76   data recorder on the surface
77
78 **Reanalysis**   a mathematically blended record that incorporates a variety of
79   observational data sets (with adjustments) in an assimilation model
80
81 **Reference networks** a small subset of sites consisting of multiple instruments that
82   independently measure the same variable which if well
83   coordinated could provide full characterization of instrument
84   errors and biases, significantly reducing uncertainty in observed
85   climate change
86
87 **Relative humidity**   the percentage of water vapor in the air relative to what is required
88   for saturation to occur at a given temperature
89
90 **Sensible heat**   heat that can be measured by a thermometer
91

92	**Specific humidity**	the amount of water vapor in the air in units of kilograms of water
93		vapor per kilogram of air
94		
95	**Trend**	a systematic change over time
96		
97		
98		

99 **Acronyms, Symbols, Abbreviations**

100		
101	AATSR	Advanced Along-Track Scanning Radiometer
102	AGCM	Atmospheric General Circulation Model
103	AIRS	Atmospheric InfraRed Sounder
104	AMIP	Atmospheric Model Intercomparison Project
105	AMSR	Advanced Microwave Scanning Radiometer
106	AMSU	Advanced Microwave Sounding Unit
107	AOGCM	Atmosphere-Ocean General Circulation Model
108	ATMS	Advanced Technology Microwave Sounder
109	ATSR	Along-Track Scanning Radiometer
110	AVHRR	Advanced Very High Resolution Radiometer
111	C20C	Climate of the 20th Century
112	CCSM	Community Climate System Model
113	CDR	Climate Data Record
114	CGCM	Coupled Atmosphere-Ocean General Circulation Model
115	CH_4	Methane
116	CLIVAR	Climate Variability and Prediction
117	CMIP2	Coupled Model Intercomparison Project (version 2)
118	CMIS	Conical scanning Microwave Imager/Sounder
119	CO_2	Carbon Dioxide
120	COADS	Comprehensive Ocean-Atmosphere Data Set
121	COWL	Cold Ocean Warm Land
122	CrIS	Cross-track Infrared Sounder
123	CRN	Climate Reference Network
124	EBM	Energy Balance Model
125	ECMWF	European Center for Medium-Range Weather Forecasts
126	ENSO	El Niño-Southern Oscillation
127	EOF	Empirical Orthogonal Function
128	ERA	ECMWF Re-Analysis
129	ERSST	Extended Reconstruction Sea Surface Temperature
130	GCM	General Circulation Model
131	GCOS	Global Climate Observing System
132	GCSM	Global Climate System Model (includes atmosphere, land, oceans,
133		glaciers)
134	GEOSS	Global Earth Observation System of Systems
135	GFDL	Geophysical Fluid Dynamics Laboratory
136	GHCN	Global Historical Climatology Network
137	GHGs	Greenhouse Gases

138	GHRSST-PP	GODAE High-Resolution Sea Surface Temperature Pilot Project
139	GISS	Goddard Institute for Space Studies
140	GODAE	Global Ocean Data Assimilation Experiment
141	GPS	Global Positioning System
142	GSN	GCOS Surface Network
143	GUAN	GCOS Upper Air Network
144	HadCM	Hadley Centre Climate Model
145	HadRT	Hadley Centre Radiosonde Temperatures
146	hPa	hectoPascals, a measure of pressure
147	HIRS	High-resolution Infrared Radiation Sounder
148	IASI	Infrared Atmospheric Sounding Interferometer
149	ICOADS	International Comprehensive Ocean-Atmosphere Data Set
150	IGRA	Integrated Global Radiosonde Archive
151	IPCC	Intergovernmental Panel on Climate Change
152	IR	Infrared Radiation
153	ITCZ	Inter Tropical Convergence Zone
154	LECT	Local Equator Crossing Time
155	LKS	Lanzante, Klein, Seidel
156	LOSU	Level of Scientific Understanding
157	LTER	Long Term Ecological Research
158	LR	Linear Regression
159	MAT	Marine Air Temperatures
160	MPS	Median of Pair-wise Slopes
161	MSU	Microwave Sounding Unit
162	NAM	Northern Hemisphere Annual Mode
163	NAO	North Atlantic Oscillation
164	NASA	National Aeronautics and Space Administration
165	NCAR	National Center for Atmospheric Research
166	NCDC	National Climatic Data Center
167	NCEP	National Centers for Environmental Prediction
168	NEMS	Nimbus E Microwave Spectrometer
169	NESDIS	National Environmental Satellite, Data, and Information Service
170	NH	Northern Hemisphere
171	NMAT	Night Marine Air Temperatures
172	N_2O	Nitrous Oxide
173	NOAA	National Oceanic and Atmospheric Administration
174	NPOESS	National Polar-orbiting Operational Environmental Satellite System
175	NRC	National Research Council
176	O_3	Ozone
177	PCM	Parallel Climate Model
178	PDO	Pacific Decadal Oscillation
179	QBO	Quasi-Biennial Oscillation
180	RATPAC	Radiosonde Atmospheric Temperature Products for Assessing Climate
181	RSS	Remote Sensing Systems
182	SAM	Southern Hemisphere Annual Mode
183	SCAMS	SCAnning Microwave Spectrometer

184	SH	Southern Hemisphere
185	SMMR	Scanning Multi-spectral Microwave Radiometer
186	SO_4	Sulfate
187	SRN	Surface Reference Network
188	SSM/I	Special Sensor Microwave Imager
189	SST	Sea Surface Temperature
190	SSU	Stratospheric Sounding Unit
191	TAO	Tropical Atmosphere Ocean
192	TIROS	Television InfraRed Observation Satellite
193	TLT	Temperature of the Lower Troposphere
194	TOGA	Tropical Ocean Global Atmosphere
195	TOVS	TIROS Operational Vertical Sounder
196	TRMM	Tropical Rainfall Measuring Mission
197	UAH	University of Alabama in Huntsville
198	UNFCCC	United Nations Framework Convention on Climate Change
199	USHCN	United States Historical Climatology Network
200	UTC	Coordinated Universal Time
201	VOS	Voluntary Observing Ships
202	VTPR	Vertical Temperature Profile Radiometer
203	WMO	World Meteorological Organization
204		