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Chapter 5

How well can the observed vertical temperature changes be reconciled with our understanding of the causes of these changes?

Convening Lead Author:

Ben D. Santer

Lead Authors:

Joyce E. Penner, and Peter W. Thorne

Contributors:

W. Collins, K. Dixon, T.L. Delworth, C. Doutriaux, Chris K. Folland, Chris E. Forest, I. Held, John Lanzante, Gerald A. Meehl, V. Ramaswamy, Dian J. Seidel, M.F. Wehner, and Tom M.L. Wigley

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 and observations show similar amplification behavior for monthly- and interannual temperature variations, but not for decadal temperature changes.*

99

100

101

102

- Tropospheric amplification of surface temperature anomalies is due to the release of latent heat by moist, rising air in regions experiencing convection.

103

104

- Despite large inter-model differences in variability and forcings, the size of this amplification effect is remarkably similar in the models considered here, even across a range of timescales (from monthly to decadal).

105

106

107

- On monthly and annual timescales, amplification is also a ubiquitous feature of observations, and is very similar to values obtained from models and basic theory.

108

109

- For longer-timescale temperature changes over 1979 to 1999, only one of four observed upper-air datasets has larger tropical warming aloft than in the surface records. All model runs with surface warming over this period show amplified warming aloft.

110

111

112

113

- These results have several possible explanations, which are not mutually exclusive. One explanation is that “real world” amplification effects on short and long time scales are controlled by different physical mechanisms, and models fail to capture such behavior. A second explanation is that remaining errors in some of the

114

115

116

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

¹This 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).

¹Volcanic 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.

¹There are a variety of different spin-up strategies.

¹In 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 state³. 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).

²Volcanic 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.

³There are a variety of different spin-up strategies.

⁴In 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

⁵If 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.

⁶Residual effects of these modes of variability may remain in the data.

⁷Provided that comparisons with observations account for the specific advantages and disadvantages noted above.

⁸See, 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.

⁹Potentially, 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.

¹⁰The 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.

321 ¹CSIRO is the Commonwealth Scientific and Industrial Research Organization.

322 ²NIES is the National Institute for Environmental Studies.

323 ³JAMSTEC 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-](http://www-pcmdi.llnl.gov/ipcc/model.documentation)
328 [pcmdi.llnl.gov/ipcc/model.documentation](http://www-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.

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

¹¹Note 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 [CO₂], 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

¹²We 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).

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412 Box 5.3: Example of a Spatially-Heterogeneous Forcing: Black Carbon Aerosols

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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 India¹⁴ 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).

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¹³Note 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).

¹⁴During 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

¹⁵Note 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

484

485 Figure 5.1: Vertical profiles of global-mean atmospheric temperature change over 1979 to 1999. Surface temperature
486 changes are also shown. Results are from two different radiosonde data sets (HadAT2 and RATPAC; see Chapter 3)
487 and from single forcing and combined forcing experiments performed with the Parallel Climate Model (PCM;
488 *Washington et al., 2000*). PCM results for each forcing experiment are averages over four different realizations of that
489 experiment. All trends were calculated with monthly mean anomaly data.
490

491 This illustrates the need for caution in comparisons of modeled and observed atmospheric
492 temperature change. The differences evident in such comparisons have multiple interpretations.
493 They may be due to real errors in the models,¹⁶ errors in the forcings used to drive the models, the
494 neglect of important forcings, and residual inhomogeneities in the observations themselves. They

¹⁶These 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

¹⁷We 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 atmo 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.

¹⁷It 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.

¹⁷The 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.*, $\pm 0.005^\circ\text{C}/\text{decade}$ in *Christy and McNider*). A fourth study removed combined ENSO, volcano, and solar effects from T_{2LT} , and estimated a residual trend of $0.065 \pm 0.012^\circ\text{C}/\text{decade}$ over 1979 to 2000 (*Douglass and Clader*, 2000).

heric CO₂ levels. This is often referred to as $\Delta T_{2\times\text{CO}_2}$. Estimates of $\Delta T_{2\times\text{CO}_2}$ 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.

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

¹⁸It 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.

¹⁹The 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.*, $\pm 0.005^\circ\text{C}/\text{decade}$ in *Christy and McNider*). A fourth study removed combined ENSO, volcano, and solar effects from T_{2LT} , and estimated a residual trend of $0.065 \pm 0.012^\circ\text{C}/\text{decade}$ 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.

²¹Their 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”.

²²The 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 data²⁴. 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

²⁴The 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 period²⁸

²⁵The 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.

²⁶The *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.

²⁸Corresponding 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 data³¹, 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

²⁹These lapse-rate changes were accompanied by increases and decreases in tropical freezing heights (which were inferred from the same radiosonde data).

³⁰This 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.

³¹These 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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³²In 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 and 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

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.

³⁴The relationship between variability at different points in a spatial field.

³⁵The 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 variability³⁹ 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*).

³⁶Another 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.

³⁷This problem is known as “degeneracy”. Formal tests of fingerprint degeneracy are sometimes applied (*e.g., Tett et al., 2002*).

³⁸The 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.

³⁹For 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 aerosols⁴⁰ 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-

⁴⁰Most of this work considers only the direct scattering effects of sulfate aerosols on incoming sunlight, and not indirect aerosol effects on clouds.

⁴¹For 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

⁴²This is partly due to the fact that natural climate noise is larger (and models are less skillful) on smaller spatial scales.

⁴³Another 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.

⁴⁴Involving 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 patterns⁴⁵ 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

⁴⁵The “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 data⁴⁷, 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 CO₂ 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.

⁴⁷The 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).

⁴⁸Well-mixed greenhouse gases, the direct effects of sulfate aerosols, combined greenhouse-gas and sulfate aerosol effects, volcanic aerosols, and solar irradiance changes.

⁴⁹The 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 CO₂-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 CO₂, 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 CO₂ increases is greater *warming* in the NH than in the SH. Because the
765 relative importance of CO₂ 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).

768

⁵⁰See, 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*

798

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 years⁵¹ (Table 5.2). These new experiments provide our
805 current best estimates of the expected climate change due to combined human and natural effects.

806

807

808

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

⁵¹This was not the case in previous model intercomparison exercises, such as AMIP (*Gates et al.*, 1999) and CMIP2 (*Meehl et al.*, 2000).

⁵²In 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.

832

833 ¹See *Dai et al.* (2001) for further details.834 ²See *Meehl et al.* (2005) for further details.835 ³The 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 ⁴See *Hansen et al.* (2005a) for further det

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841

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} ,

⁵³There 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 individual 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 $-0.25^\circ\text{C}/\text{decade}$ (Table 5.4A). Most of this cooling is due to the combined effects of stratospheric

⁵⁴For 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.

⁵⁵In 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).

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 ($-0.35^\circ\text{C}/\text{decade}$) 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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⁵⁶Recent 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).

⁵⁷Recent 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

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).

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 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

⁵⁸These warming signals occur because volcanic aerosols absorb both incoming solar radiation and outgoing thermal radiation (*Ramaswamy et al.*, 2001a).

⁵⁹The 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

⁶⁰More 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).

⁶¹Because 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).

⁶²Because 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 smaller 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).

1004

1005 5.2 *Tropical Temperature and Lapse-Rate Trends*

1006

⁶³Note 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).

⁶⁴Values are $-0.02^\circ\text{C}/\text{decade}$ in the case of T_S minus T^*_G and $-0.03^\circ\text{C}/\text{decade}$ 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 $+0.16^\circ\text{C}/\text{decade}$. 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

⁶⁵This 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 T₄ trends are close to the
1025 model average.⁶⁷

⁶⁶This supports recent findings of a residual cooling bias in tropical radiosonde data (*Sherwood et al., 2005; Randel and Wu, 2005*).

⁶⁷The model average is $-0.27^{\circ}\text{C}/\text{decade}$ 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

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 sign of the trend in

⁶⁸Note 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} .

⁶⁹Three 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.

1055

1056 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}
1057 were calculated over the 252-month period from January 1979 to December 1999. Model results are ensemble means
1058 from 20CEN experiments performed with CCSM3.0 (panel A), PCM (panel B), GFDL CM2.1 (panel C), and GISS-EH
1059 (panel D). Observed results rely on NOAA T_S trends and on two different satellite estimates of trends in T_{2LT} , obtained
1060 from UAH (panel E) and RSS (panel F). White denotes high elevation areas where it is not meaningful to calculate
1061 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).
1062

1063 To illustrate structural uncertainties in the observed data, we show two different patterns of trends
1064 in T_S minus T_{2LT} . Both rely on the same NOAA surface data, but use either UAH (Figure 5.5E) or
1065 RSS (Figure 5.5F) as their source of T_{2LT} results. The “NOAA minus UAH” combination provides
1066 a picture that is very different from the model results, with coherent warming of the surface relative
1067 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

⁷⁰The 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.

⁷¹The 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).

⁷²Note 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

1108 Figure 5.6: Scatter plots showing the relationships between tropical temperature changes at Earth's surface and in two
1109 different layers of the troposphere. All results rely on temperature data that have been spatially-averaged over the deep
1110 tropics (20°N-20°S). Model data are from 49 realizations of 20CEN runs performed with 19 different models (Table
1111 5.1). Observational results were taken from four different upper-air datasets (two from satellites, and two from
1112 radiosondes) and two different surface temperature datasets (see Chapter 3). The two upper panels provide information
1113 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
1114 consider temperature changes on multi-decadal timescales, and show the trends (over 1979 to 1999) in T_S and T_{2LT}
1115 (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
1116 slope provides information on the amplification of surface temperature variability and trends in the free troposphere.
1117 The black line in each panel is given for reference purposes, and has a slope of 1. Values above (below) the black lines
1118 indicate tropospheric amplification (damping) of surface temperature changes. There are two columns of observational
1119 results in C and D. These are based on the NOAA and HadCRUT2v TS (0.12 and 0.14°C/decade, respectively). Note
1120 that panel C show results from published and recently-revised versions of the UAH T_{2LT} data (versions 5.1 and 5.2).
1121 Since the standard deviations calculated from NOAA and HadCRUT2v monthly T_S anomalies are very similar,
1122 observed results in A and B use NOAA standard deviations only. The blue shading in the bottom two panels defines
1123 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 damping 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

⁷³As 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.

⁷⁴The 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.

⁷⁵Despite the “end point” effect of the large El Niño event in 1997-1998 (see Chapter 3).

1184

1185

1186 Figure 5.7: Zonal-mean patterns of atmospheric temperature change in “20CEN” experiments performed with four
1187 different climate models and in observational radiosonde data. Model results are for CCSM3.0 (panel A), PCM (panel
1188 B), GFDL CM 2.1 (panel C), and GISS-EH (panel D). The model experiments are ensemble means. There are
1189 differences between the sets of climate forcings that the four models used in their 20CEN runs (Table 5.3). Observed
1190 changes (panel E) were estimated with HadAT2 radiosonde data (*Thorne et al.*, 2005, and Chapter 3). The HadAT2
1191 temperature data do not extend above 30 hPa, and have inadequate coverage at high latitudes in the Southern
1192 Hemisphere. All temperature changes were calculated from monthly-mean data and are expressed as linear trends (in
1193 °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

⁷⁶This 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,

⁷⁷The 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.

⁷⁸The 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).

⁷⁹More 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

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 “ $1.3 \pm 0.3\%$ 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).

⁸¹Note, 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 T₂ 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

⁸³Model control runs cannot generate such large multi-decadal increases in the heat content of the global ocean.

⁸⁴The 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.

1282

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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