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5   **Chapter 1**
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7
8   Why do temperatures vary vertically (from the surface to the stratosphere) and what do we
9   understand about why they might vary and change over time?
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46   **Summary**

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48   **Temperatures vary vertically.**

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

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

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

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

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

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

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

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

57

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

59   **higher in the atmosphere because:**

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

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

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

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

64   different locations.

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

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

67   surface and aloft.

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

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

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

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

72   between inversion layers and adjacent layers.

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

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

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

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

77   agents.

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79

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

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

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

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

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

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

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

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

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

89

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

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

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

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

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

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

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

97   Northern Hemisphere.

98

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

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

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

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

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

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

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

106   ~12 km at ~30-40° latitude, and to as low as about 8 km at the poles.

107

108 Figure 1.1. Global climatological vertical temperature profiles from surface to troposphere and extending into the
109 stratosphere for December-January-February and June-July-August mean conditions, as obtained from the National
110 Centers for Environmental Prediction reanalyses (Kalnay et al., 1996; updated). The solid line denotes the
111 tropopause which separates the stratosphere from the surface-troposphere system.

112
113 Temperatures generally decrease with height from the surface although there are important

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

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

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

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

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

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

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

121

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

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

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

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

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

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

128   from tropospheric temperatures on daily or even weekly timescales.

129

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

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

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

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

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

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

136

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

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

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

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

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

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

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

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

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

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

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

148   due to evaporation or condensation of water.

149

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

169   the poles.

170

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

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

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

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

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

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

177   everywhere.

178

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

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

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

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

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

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

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

186   wintertime temperatures at the poles.

187

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

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

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

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

192   exceed those at the surface.

193

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

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

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

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

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

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

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

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

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

203

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

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

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

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

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

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

210

211   **1.2 Forcing of climate change**

212

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

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

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

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

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

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

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

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

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

222   greater), as discussed in the subsequent chapters.

223

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

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

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

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

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

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

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

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

232

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

290   radiation balance and thermal profile.

291

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

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

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

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

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

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

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

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

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

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

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

303

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

319

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

338   2004).

339

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

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

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

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

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

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

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

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

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

349   climate modeling investigations to increase our understanding and reduce uncertainties

350   associated with those processes.

351

352   **1.3 Stratospheric forcing and related effects**

353

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

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

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

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

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

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

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

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

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

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

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

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

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

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

368

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

374

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

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

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

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

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

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

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

382 WMO/SROC Report, 2004).

383

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

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

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

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

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

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

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

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

392   discussed further in Chapter 5.

393

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

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

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

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

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

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

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

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

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

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

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

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

406

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

426   et al., 2003).

427

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

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

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

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

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

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

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

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

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

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

438

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

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

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

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

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

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

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

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

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

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

449

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

465

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

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

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

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

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

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

475   2001; NRC, 2004).

476

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

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

479   at various levels in the atmosphere.

480

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

482

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

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

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

486   surface and tropospheric temperatures over 1979-2003 (Figure 1.4). The correlation coefficients

487   between monthly surface and tropospheric temperature anomalies (as represented by

488   temperatures derived from MSU satellite data) reveal very distinctive patterns, with values

489   ranging from less than zero (implying poor vertical coherence of the surface and tropospheric

490   temperature anomalies) to over 0.9. The highest correlation coefficients (>0.75) are found across

491   the middle and high latitudes of Europe, Asia, and North America, indicating a strong

492   association between the surface and tropospheric monthly temperature variations. Correlations

493   are generally much less (~0.5) over the tropical continents and the North Atlantic and North

494   Pacific Oceans. Correlations less than 0.3 occur over the tropical and southern oceans and are

495   lowest (<0.15) in the tropical western Pacific. Relatively high correlation coefficients (>0.6) are

496   found over the tropical eastern Pacific where the ENSO signal is large and the sea-surface

497   temperature fluctuations influence the atmosphere significantly.

498

499   Figure 1.4. Gridpoint correlation coefficients between monthly surface and tropospheric temperature anomalies over
500   1979-2003. The tropospheric temperatures are derived from MSU satellite data (Christy et al., 2003).

501

502   Differences between the surface and tropospheric temperature records are found where there is

503   some degree of decoupling between the layers of the atmosphere. For instance, as discussed

504   earlier, over portions of the subtropics and tropics, variations in surface temperature are

505   disconnected from those aloft by a persistent trade-wind inversion. Shallow temperature

506   inversions are also commonly found over land in winter, especially in high latitudes on sub-

507   seasonal timescales, so that there are occasional large differences between monthly surface and

508   tropospheric temperature anomalies.

509

510   More important than correlations for trends, however, is the variability of the two temperature

511   records, assessed by computing the standard deviation of the measurement samples of each

512   record (Figure 1.5). The figure exhibits pronounced regional differences in variability between

513   the surface and tropospheric records. Standard deviations also help in accounting for the

514   differences in correlation coefficients, because they yield information on the size and persistence

515   of the climate signal relative to the noise in the data. For instance, large variations in eastern

516   tropical Pacific sea surface temperature associated with ENSO dominate over measurement

517   uncertainties, as do large month-to-month swings in surface temperatures over extratropical

518   continents.

519

520   Figure 1.5. Standard deviations of monthly mean temperature anomalies from the surface and tropospheric
521   temperature records over 1979-2003. The tropospheric temperatures are derived from MSU satellite data (Christy et
522   al., 2003).

523

524   The largest variability in both surface and tropospheric temperature is over the Northern

525   Hemisphere continents. The standard deviation over the oceans in the surface data set is much

526   smaller than over land except where the ENSO phenomenon is prominent. The standard

527   deviations of tropospheric temperature, in contrast, exhibit less zonal variability. Consequently,

528   the standard deviations of the monthly tropospheric temperatures are larger than those of the

529   surface data by more than a factor of two over the North Pacific and North Atlantic. Over land,

530   tropospheric temperatures exhibit slightly less variability than surface temperatures. These

531   differences in variability are indicative of differences in physical processes over the oceans

532   versus the continents. Of particular importance are the roles of the land surface and ocean as the

533   lower boundary for the atmosphere and their very different abilities to store heat, as well as the

534   role of the atmospheric winds that help to reduce regional differences in tropospheric

535   temperature through the movement of heat from one region to another.

536

537   Over land, heat penetration into the surface involves only the upper few meters, and the ability of

538   the land to store heat is low. Therefore, land surface temperatures vary considerably from

539   summer to winter and as cold air masses replace warm air masses and vice versa. The result is

540   that differences in magnitude between surface and lower-atmospheric temperature anomalies are

541   relatively small over the continents: very warm or cold air aloft is usually associated with very

542   warm or cold air at the surface. In contrast, the ability of the ocean to store heat is much greater

543   than that of land, and mixing in the ocean to typical depths of 50 meters or more considerably

544   moderates the sea surface temperature response to cold or warm air. Over the northern oceans,

545   for example, a very cold air mass (reflected by a large negative temperature anomaly in the

546 tropospheric record) will most likely be associated with a relatively small negative temperature

547 anomaly at the sea surface. This is one key to understanding the differences in trends between

548 the two records.

549

550 Long-term changes in the atmospheric circulation can be reflected by trends in indices of

551 patterns (or modes) of natural climate variability such as ENSO, the North Atlantic Oscillation

552 (NAO; also known as the Northern Hemisphere annular mode, or NAM), and the Southern

553 Hemisphere (SH) annular mode (SAM). The exact magnitudes of the index trends depend on the

554 period of time examined. Over the past several decades, for instance, changes in atmospheric

555 circulation (reflected by a strong upward trend in indices of the NAO) have contributed to a Cold

556 Ocean Warm Land (COWL) surface temperature pattern over the Northern Hemisphere (Hurrell,

557 1996; Thompson and Wallace, 2000). In the lower atmosphere, winds blowing from ocean to

558 land to ocean are much stronger than at the surface, and this moderating influence of the winds

559 contributes to less east-west variability in the tropospheric data (Figure 1.5). Thus, the recent

560 warm anomalies over the continents are roughly cancelled by the cold anomalies over the oceans

561 in the tropospheric dataset. This is not the case for the surface temperature record, which is

562 dominated by the warmth over the continents. The result is that the changes in the Northern

563 Hemisphere (NH) atmospheric circulation over the past few decades have produced a significant

564 difference in surface and tropospheric temperature trends (Hurrell and Trenberth, 1996).

565 Similarly, Thompson and Solomon (2002) showed that recent tropospheric temperature trends at

566 high southern latitudes were related to changes in the SAM.

567

568   Physical differences between the two measures of temperature are also evident in their dissimilar

569   responses to volcanic eruptions and ENSO (e.g., Santer et al. 2000). These phenomena have a

570   greater effect on tropospheric than surface temperature, especially over the oceans (Jones, 1994).

571   However, changes in ENSO over the past several decades do not explain long-term changes in

572   tropical tropospheric temperatures (Hegerl and Wallace, 2002). Changes in concentrations of

573   stratospheric ozone could also be important, as the troposphere is cooled more by observed

574   ozone depletion than is the surface (Hansen et al., 1995 ; Ramaswamy et al., 1996). Another

575   contributing factor could be that at the surface, the daily minimum temperature has increased at a

576   faster rate than the daily maximum, resulting in a decrease in the diurnal temperature range over

577   many parts of the world (e.g., Easterling et al., 1997; Dai et al., 1999). Because of nighttime

578   temperature inversions, the increase in the daily minimum temperatures likely involves only a

579   shallow layer of the atmosphere that would not be evident in upper-air temperature records.

580   These physical processes provide indications of why trends in surface temperatures are expected

581   to be different from trends in the troposphere, especially in the presence of strong interannual

582   variability, even if both sets of measurements were perfect. Of course they are not, as described

583   in more detail in Chapter 2, which deals with the strengths and limitations of current observing

584   systems. An important issue implicit in Figure 1.5 is that of spatial sampling, and the

585   accompanying caveats about interpretation of the truly global coverage provided by satellites

586   versus the incomplete space and time coverage offered by radiosondes. These are discussed in

587   depth in Chapters 2 and 3.

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