Chapter 1

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

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Summary

Temperatures vary vertically.

The solar heating of the surface of the planet, combined with the physical properties of the overlying air, produce the highest temperatures, on average, at the surface; that heat is mixed vertically and horizontally by various physical processes. Taking into account the distribution of atmospheric moisture and the lower air pressure with progressively increasing altitude, there results a decrease of temperature with height up to the tropopause. The tropopause marks the top of the troposphere, i.e., the lower 8 to 16 km of the atmosphere depending on latitude. Above this altitude, the radiative properties of the air produce a warming with height through the stratosphere (extending from the tropopause to ~50 km).

Temperature trends at the surface can be expected to be different from temperature trends higher in the atmosphere because:

- Physical properties of the surface vary depending on whether the location has land, sea, snow, or ice. Near the surface, these differing conditions can produce strong horizontal variations in temperature. Above the surface layer, these contrasts are quickly smoothed out, contributing to varying temperature trends with height at different locations.
- Changes in atmospheric circulation or modes of atmospheric variability (e.g., El Niño-Southern Oscillation [ENSO]) can produce different temperature trends at the surface and aloft.

- Under some circumstances, temperatures may increase with height near the surface or higher in the troposphere, producing a "temperature inversion." Such inversions are more common at night, in winter over continents, and in the trade wind regions. Since the air in inversion layers is resistant to vertical mixing, temperatures trends can differ between inversion layers and adjacent layers.
- Forcing factors, either natural (e.g., volcanoes and solar) or human-induced (e.g., greenhouse gas, aerosols, ozone, and land use) can result in differing temperature trends at different altitudes, and these vertical variations may change over time. This can arise due to spatial changes in the concentrations or properties of the forcing agents.

This Chapter describes the temperature profile of the layers of the atmosphere from the surface through the stratosphere and discusses the basic reasons for this profile. We also use results from global climate model simulations to show how changes in natural and human-induced factors can produce different temperature trends in the various layers of the atmosphere. This discussion provides the background for the presentation of the observed changes (Chapters 2-4), and for the understanding of their causes (Chapter 5). We also describe temperature changes in the stratosphere in recent decades and the influences of these changes on the troposphere. Finally, making use of surface and satellite observations, we examine the physical processes that can result in different temperature trends at the surface and in the troposphere.

1.1 The Thermal Structure of the Atmosphere

Surface temperatures are at their warmest in the tropics, where the largest amount of solar radiation is received during the course of the year, and decrease towards the Polar Regions where the annual-mean solar radiation received is at a minimum (Oort and Peixoto, 1992). The temperature contrast between summer and winter increases from the equator to the poles. Since land areas heat up and cool more rapidly than oceans, and because of the preponderance of land in the Northern Hemisphere, there is a larger contrast between summer and winter in the Northern Hemisphere.

Figure 1.1 illustrates the climatological vertical temperature profiles for December, January, February and June, July, August mean conditions, as obtained from the National Centers for Environmental Prediction (NCEP) reanalyses (Kalnay et al., 1996; updated). It is convenient to think first in terms of climatological conditions upon which spatial and temporal variations/trends are superimposed. The solid line in the plot illustrates the tropopause, which separates the troposphere below from the stratosphere above. The tropopause is at its highest level in the tropics (~20°N-20°S). It descends sharply in altitude from ~16 km at the equator to ~12 km at ~30-40° latitude, and to as low as about 8 km at the poles.

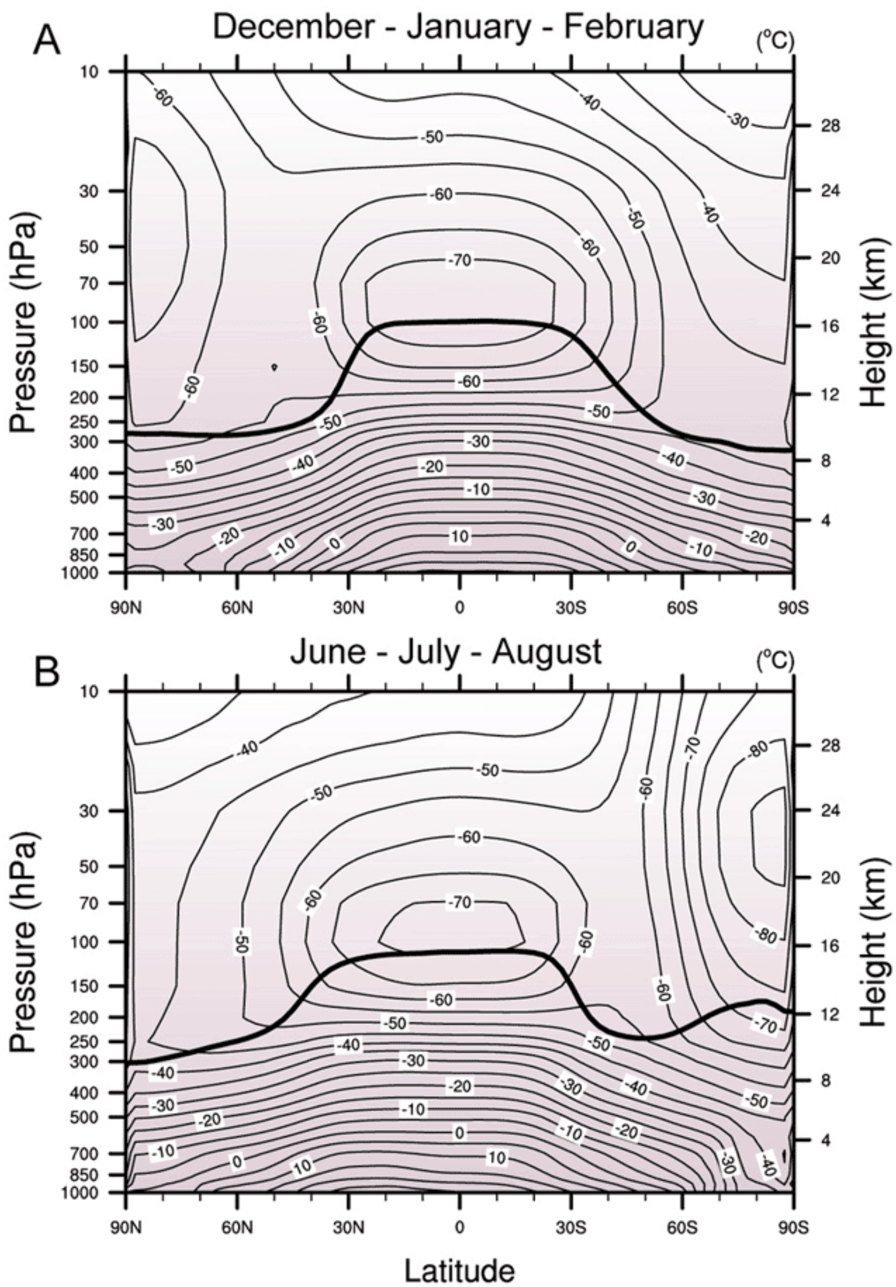

Figure 1.1. Global climatological vertical temperature profiles from surface to troposphere and extending into the stratosphere for December-January-February and June-July-August mean conditions, as obtained from the National Centers for Environmental Prediction reanalyses (Kalnay et al., 1996; updated).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

Temperatures generally decrease with height from the surface although there are important exceptions. The rate at which the temperature changes with height is termed the "lapse rate." The lapse rate can vary with location and season, and its value depends strongly on the atmospheric humidity, e.g., the lapse rate varies from ~5ºC/km near the surface in the moist tropical regions (near the equator) to much larger values (~8-9ºC/km) in the drier subtropics (~20-30°). Important departures from nominal lapse rate values can occur near the surface and in the upper troposphere. In the equatorial tropics, the tropopause region (~16 km) is marked by a smaller value of the lapse rate than in the lower troposphere.

The thermal structure of the lowest 2-3 km, known as the "planetary boundary layer," can be complicated, even involving inversions (in which temperature increases rather than decreases with height) occurring at some latitudes due to land-sea contrasts, topographic influences, radiative effects and meteorological conditions. Inversions are particularly common during winter over some middle and high latitude land regions and are a climatological feature in the tropical trade wind regions. The presence of inversions acts to decouple surface temperatures from tropospheric temperatures on daily or even weekly timescales.

Above the tropopause is the stratosphere, which extends to ~50 km and where the temperature increases with height. In the vicinity of the tropical tropopause, (i.e., the upper troposphere and lower stratosphere regions, ~15-18 km), the temperature varies little with height. The extratropical (poleward of 30°) lower stratosphere (at ~8-12 km) also exhibits a similar feature (Holton, 1979). The lapse rate change with altitude in the upper troposphere/lower stratosphere region is less sharp in the extratropical latitudes than in the tropics.

The global temperature profile of the atmosphere reflects a balance between the radiative and dynamical heating/cooling of the surface-atmosphere system. From a global, annual-average point of view, the thermal profile of the stratosphere is the consequence of a balance between radiative heating and cooling rates due to greenhouse gases, principally carbon dioxide (CO_2), ozone (O_3) and water vapor (H_2O) (Andrews et al., 1987). The vertical profile of the troposphere is the result of a balance of radiative processes involving greenhouse gases, aerosols, and clouds (Stephens and Webster, 1981; Goody and Yung, 1989), along with the strong role of moist convection (Holton, 1979; Kiehl, 1992). An important difference between the troposphere and stratosphere is that the stratosphere is characterized by weak vertical motions, while in the troposphere, the vertical motions are stronger. Most significantly, the moist convective processes that are a characteristic feature of the troposphere include the transfer of large amounts of heat due to evaporation or condensation of water.

Convective processes are important in determining the temperature profile in the troposphere. This is illustrated by the fact that radiative processes alone would cause the surface to be significantly warmer than it is actually. This would occur because the atmosphere is relatively transparent to the Sun’s radiation. By itself, this would lead to a drastic heating of the surface accompanied by a net radiative cooling of the atmosphere (Manabe and Wetherald, 1967). However, the resulting convective motions remove this excess heating from the surface in the form of sensible and latent heat, the latter involving the evaporation of moisture from the surface (Ramanathan and Coakley, 1978). As air parcels rise and expand, they cool due to decompression, leading to a decrease of temperature with height. The lapse rate for a dry atmosphere, when there are no moist processes and the air is rising quickly enough to be unaffected by other heating/cooling sources, is close to 10ºC/km. However, because of moist convection, there is condensation of moisture, formation of clouds and release of latent heat as the air parcels rise, causing the lapse-rate to be much less, as low as 4ºC/km in very humid atmospheres (Houghton, 1977). In a more rigorous sense, the interactions between radiation, convection, cloud physics, and dynamical motions (ranging from large- to meso- and small-scales) govern the actual rate at which temperature decreases with height (the lapse-rate) at any location. Large-scale dynamical mechanisms tend to homogenize temperatures above the boundary layer over horizontal scales (Rossby radius) that vary from planetary scale near the equator to a couple of thousand kilometers at midlatitudes and to a few hundred kilometers near the poles.

Convective processes and vertical mixing of air can add complexity to the nominal thermal profile in the tropics mentioned above. For example, a more detailed picture in subtropical regions consists of a surface mixed layer (up to about 500 m) and a trade wind boundary layer (up to about 2 km) above which is the free troposphere. Each of the boundary layers is topped by an inversion which tends to isolate the region from the layer above (Sarachik, 1985). This indicates the limitations in assuming nominal lapse rate values from the surface to the tropopause everywhere.

The radiative-convective picture above is likely of dominant relevance only for the tropics. In the extra-tropics (poleward of 30°), the lapse rate and tropopause height are mostly determined by instabilities associated with the more familiar weather systems ("baroclinic instability"). The rising motions of air parcels in the equatorial moist tropics associated with deep convection descend in the subtropical regions leading to drier environments there (Hadley circulations). In the polar regions (~60-90°), planetary-scale waves forced by the influences of mountains and that of land-sea contrasts upon the flow of air play a significant role in the determination of the wintertime temperatures at the poles.

Based on these simple ideas, the lapse rate can be expected to decrease with an increase in humidity, and also to depend on the atmospheric circulation. As specific humidity is strongly related to temperature, and is expected to rise with surface warming, the lapse rate (other things being equal) can be expected to decrease with warming such that temperature changes aloft exceed those at the surface.

The above simple picture of radiative-convective balance, together with the requirement of radiative balance at the top-of-the-atmosphere (namely, equilibrium conditions wherein the net solar energy absorbed by the Earth's climate system must be balanced by the infrared radiation emitted by the Earth), can help illustrate the significance of long-lived infrared absorbing gases in the atmosphere such as carbon dioxide. The presence of strongly infrared-absorbing greenhouse gases (water vapor, carbon dioxide, methane, etc.) causes the characteristic infrared emission level of the planet to be shifted to a higher altitude where temperatures are colder. The re-establishment of thermal equilibrium leads to warming and communication of the added heat input to the troposphere and surface (Goody and Yung, 1989; Lindzen and Emanuel, 2002).

In the tropical upper troposphere, moisture- and cloud-related features (e.g., upper tropospheric relative humidity, cirrus cloud microphysics, and mesoscale circulations in anvil clouds) are important factors in governing the thermal profile (Ramanathan et al., 1983; Ramaswamy and Ramanathan, 1989; Donner et al., 2001; Sherwood and Dessler, 2003). In the upper troposphere and especially the stratosphere, convective motions become weak enough that radiative (solar and longwave) heating/cooling become important in establishing the lapse rate.

1.2 Forcing of climate change

Potentially significant variations and trends are superimposed on the basic climatological thermal profile, as revealed by observational data in the subsequent chapters. While the knowledge of the climatological mean structure discussed in the previous section involves considerations of radiative, convective, and dynamical processes, understanding the features and causes of the magnitude of changes involves a study of the perturbations in these processes which then frame the response of the climate system to the forcing. While the understanding of climate variability is primarily based on observations of substantial changes (e.g., sea-surface temperature changes during El Niño), the vertical temperature changes being investigated in this report are changes on the order of a few tenths of degrees on the global-mean scale (local changes could be much greater), as discussed in the subsequent chapters.

"Unforced" variations, i.e., changes arising from internally generated variations in the atmosphere-ocean-land-ice/snow climate system, can influence surface and atmospheric temperatures substantially, e.g., due to changes in equatorial sea-surface temperatures associated with ENSO. Climate models indicate that global-mean unforced variations on multidecadal timescales are likely to be smaller than, say, the 20th Century global-mean increase in surface temperature (Stouffer et al., 2000). However, for specific regions and/or seasons, this may not be valid and the unforced variability could be substantial. Chapter 5 provides more detail on models and their limitations (see particularly Box 5.1 and 5.2).

Because of the influence of radiative processes on the thermal structure, anything external to the climate system that perturbs the planet's radiative heating distribution can cause climate changes, and thus is potentially important in accounting for the observed temperature changes (Santer et al., 1996). The radiative (solar plus longwave) heat balance of the planet can be forced by:

- natural factors such as changes in the Sun's irradiance, and episodic, explosive volcanic events (leading to a build-up of particulates in the stratosphere);
- human-induced factors such as changes in the concentrations of radiatively active gases (carbon dioxide, methane, etc.) and aerosols.

Potentially important external forcing agents of critical relevance for the surface and atmospheric temperature changes over the 20th Century are summarized in Table 1.1 (for more details, see Ramaswamy et al., 2001; NRC, 2005).

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

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

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

component. Forcings other than well-mixed gases and solar are spatially localized, with the degree of localization having considerable variations amongst the different agents, depending on their respective source locations. In addition, for short-lived species such as ozone and aerosols, the long-range meteorological transport plays an important role in their global distributions. Level of confidence is a subjective measure of the certainty in the quantitative estimate of the global-mean forcing.

Typically, the forcing becomes near-global a few months after an intense tropical eruption.

* * In the case of volcanic aerosols, the level of confidence in the forcing from the most recent intense eruption, that of Mt. Pinatubo in 1991, is reasonably good because of reliable observing systems in place; for prior explosive eruptions, observations were absent or sparse which affects the reliability of the quantitative estimates for the previous volcanic events.

@ Although solar irradiance variations before 1980 have a very low level of confidence, direct observations of the Sun's output from satellite platforms since 1980 are considered to be accurate (Lean et al., 2005). Thus, the forcing due to solar irradiance variations from 1980 to present are known to a much greater degree of confidence than from pre-industrial to present time.

The forcing agents differ in terms of whether their radiative effects are felt primarily in the stratosphere or troposphere or both, and whether the perturbations occur in the solar or longwave spectrum or both. The quantitative estimates of the forcing due to the well-mixed greenhouse gases (comprised of carbon dioxide, methane, nitrous oxide and halocarbons) are known to a higher degree of scientific confidence than the other forcings. Among aerosols, black carbon is distinct because it strongly absorbs solar radiation (see also Box 5.3). In contrast to sulfate aerosols, which cause a perturbation of solar radiation mainly at the surface (causing a cooling effect), black carbon acts to warm the atmosphere while cooling the surface (Chung et al., 2002; Menon et al., 2002). This could have implications for convective activity and precipitation (Ramanathan et al., 2005), and the lapse rate (Chung et al., 2002; Erlick and Ramaswamy, 2003). The response to radiative forcings need not be localized and can manifest in locations remote from the perturbation. This is because atmospheric circulation tends to homogenize the effect of heat perturbations and hence the temperature response. The vertical partitioning of the radiative perturbation determines how the surface heat and moisture budgets respond, how the convective interactions are affected, and hence how the surface temperature and the atmospheric thermal profile are altered. "Indirect" aerosol effects arising from aerosol-cloud interactions can lead to potentially significant changes in cloud characteristics such as cloud lifetimes, frequencies of occurrence, microphysical properties, and albedo (reflectivity) (e.g., Lohmann et al., 2000; Sherwood, 2002; Lohmann and Feichter, 2005). As clouds are important components in both solar and longwave radiative processes and hence significantly influence the planetary radiation budget (Ramanathan et al., 1987; Wielicki et al., 2002), any effect caused by aerosols in perturbing cloud properties is bound to exert a significant effect on the surface-troposphere radiation balance and thermal profile.

Estimates of forcing from anthropogenic land-use changes have consisted of quantification of the effect of snow-covered surface albedos in the context of deforestation (Ramaswamy et al., 2001). However, there remain considerable uncertainties in these quantitative estimates. There are other possible ways in which land-use change can affect the heat, momentum and moisture budgets at the surface (e.g., changes in transpiration from vegetation) (see also Box 5.4), and thus exert a forcing of the climate (Pielke et al., 2004; NRC, 2005). In addition to the forcings shown in Table 1.1, NRC (2005) has evoked a category of “nonradiative” forcings involving aerosols, land-cover, and biogeochemical changes which may impact the climate system first through nonradiative mechanisms, e.g., modifying the hydrologic cycle and vegetation dynamics. Eventually, radiative impacts could occur, though no metrics for quantifying these nonradiative forcings have been accepted as yet (NRC, 2005).

Not all the radiative forcings are globally uniform. In fact, even for the increases in well-mixed greenhouse gases, despite their globally uniform mixing ratios, the resulting forcing of the climate system is at a maximum in the tropics due to the temperature contrast between the surface and troposphere there and therefore the increased infrared radiative energy trapping.

Owing to the dependence of infrared radiative transfer on clouds and water vapor, which have substantial spatial structure in the low latitudes, the greenhouse gas forcing is non-uniform in the tropics, being greater in the relatively drier tropical domains. For short-lived gases, the concentrations themselves are not globally uniform so there tends to be a distinct spatial character to the resulting forcing, e.g., stratospheric ozone, whose forcing is confined essentially to the mid-to-high latitudes, and tropospheric ozone whose forcing is confined to tropical to midlatitudes. For aerosols, which are even more short-lived than ozone, the forcing has an even more localized structure (see also Box 5.3). However, although tropospheric ozone and aerosol forcing are maximized near the source regions, the contribution to the global forcing from remote regions is not negligible. The natural factors, namely solar irradiance changes and stratospheric aerosols from volcanic eruptions, exert a forcing that is global in scope.

In terms of the transient changes in the climate system, it is also important to consider the temporal evolution of the forcings. For well-mixed greenhouse gases, the evolution over the past century, and in particular the past four decades, is very well quantified because of reliable and robust observations. However, for the other forcing agents, such as aerosols, there are uncertainties concerning their evolution that can affect the inferences about the resulting surface and atmospheric temperature trends. Stratospheric ozone changes, which have primarily occurred since ~1980, are slightly better known than tropospheric ozone and aerosols. For solar irradiance and land-surface changes, the knowledge of the forcing evolution over the past century is poorly known. Only in the past five years have climate models included time varying estimates of a subset of the forcings that affect the climate system. In particular, current models typically include GHGs, ozone, sulfate aerosol direct effects, solar influences, and volcanoes. Some very recent model simulations also include time-varying effects of black carbon and land use change. Other forcings either lack sufficient physical understanding or adequate global time- and space-dependent datasets to be included in the models at this time. As we gain more knowledge of these other forcings and are better able to quantify their space- and time-evolving characteristics, they will be added to the models used by groups around the world. Experience with these models so far has shown that the addition of more forcings generally tends to improve the realism and details of the simulations of the time evolution of the observed climate system (e.g., Meehl et al., 2004).

Whether the climate system is responding to internally generated variations in the atmosphere itself, to atmosphere-ocean-land-surface coupling, or to externally applied forcings by natural and/or anthropogenic factors, there are feedbacks that arise which can play a significant role in determining the changes in the vertical and horizontal thermal structure. These include changes in the hydrologic cycle involving water vapor, clouds, sea-ice, and snow, which by virtue of their strong interactions with solar and longwave radiation, amplify the effects of the initial perturbation (Stocker et al., 2001; NRC, 2003) in the heat balance, and thus influence the response of the climate system. Convection and water vapor feedback, and cloud feedback in particular, are areas of active observational studies; they are also being pursued actively in climate modeling investigations to increase our understanding and reduce uncertainties associated with those processes.

1.3 Stratospheric forcing and related effects

Observed changes in the stratosphere in recent decades have been large and several recent studies have investigated the causes. WMO (2003) and Shine et al. (2003) conclude that the observed vertical profile of cooling in the global, annual-mean stratosphere (from the tropopause up into the upper stratosphere) can, to a substantial extent, be accounted for in terms of the known changes that have taken place in well-mixed greenhouse gases, ozone, and water vapor (Figure 1.2). Even at the zonal, annual-mean scale, the lower stratosphere temperature trend is discernibly influenced by the changes in the stratospheric gases (Ramaswamy and Schwarzkopf, 2002; Langematz et al., 2003). In the tropics, there is considerable uncertainty about the magnitude of the lower stratospheric cooling (Ramaswamy et al., 2001). In the high northern latitudes, the lower stratosphere becomes highly variable both in the observations and model simulations, especially during winter, such that causal attribution is difficult to establish. In contrast, the summer lower stratospheric temperature changes in the Arctic and the springtime cooling in the Antarctic can be attributed in large part to the changes in the greenhouse gases (WMO, 2003; Schwarzkopf and Ramaswamy, 2002).

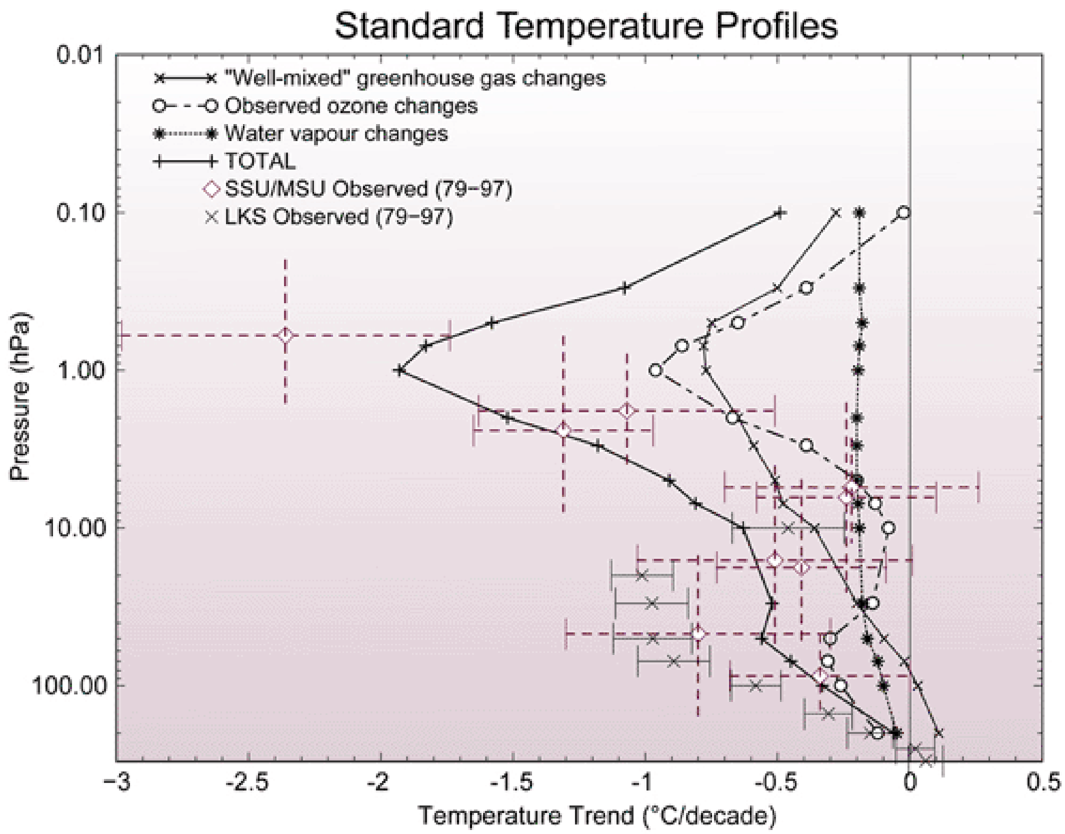

Figure 1.2. Global- and annual-mean temperature change over the 1979-1997 period in the stratosphere. Observations: LKS (radiosonde), SSU and MSU (satellite) data.
Vertical bars on satellite data indicate the approximate span in altitude from where the signals originate, while the horizontal bars are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

Owing to the cooling of the lower stratosphere, there is a decreased infrared emission from the stratosphere into the troposphere (Ramanathan and Dickinson, 1979; WMO, 1999), leading to a radiative heat deficit in the upper troposphere, and a tendency for the upper troposphere to cool. In addition, the depletion of ozone in the lower stratosphere can result in ozone decreases in the upper troposphere due to reduced transport from the stratosphere (Mahlman et al., 1994). This too affects the heat balance in the upper troposphere. Further, lapse rate near the tropopause can be affected by changes in radiatively active trace constituents such as methane (WMO, 1986; WMO/SROC Report, 2004).

The height of the tropopause (the boundary between the troposphere and stratosphere) is determined by a number of physical processes that make up the integrated heat content of the troposphere and the stratosphere. Changes in the heat balance within the troposphere and/or stratosphere can consequently affect the tropopause height. For example, when a volcanic eruption puts a large aerosol loading into the stratosphere where the particles absorb solar and longwave radiation and produce stratospheric heating and tropospheric cooling, the tropopause height shifts downward. Conversely, a warming of the troposphere moves the tropopause height upward (e.g., Santer et al., 2003). Changes in tropopause height and their potential causes will be discussed further in Chapter 5.

The episodic presence of volcanic aerosols affects the equator-to-pole heating gradient, both in the stratosphere and troposphere. Temperature gradients in the stratosphere or troposphere can affect the state of the polar vortex in the northern latitudes, the coupling between the stratosphere and troposphere, and the propagation of temperature perturbations into the troposphere and to the surface. This has been shown to be plausible in the case of perturbations due to volcanic aerosols in observational and modeling studies, leading to a likely causal explanation of the observed warming pattern seen in northern Europe and some other high latitude regions in the first winter following a tropical explosive volcanic eruption (Jones et al., 2003; Robock and Oppenheimer, 2003; Shindell et al., 2001; Stenchikov et al., 2002). Ozone and well-mixed greenhouse gas changes in recent decades can also affect stratosphere-troposphere coupling (Thompson and Solomon, 2002; Gillett and Thompson, 2003), propagating radiatively-induced temperature perturbations from the stratosphere to the surface in the high latitudes during winter.

1.4 Simulated responses in vertical temperature profile to different external forcings

Three-dimensional computer models of the coupled global atmosphere-ocean-land surface climate system have been used to systematically analyze the expected effects of different forcings on the vertical structure of the temperature response and compare these with the observed changes (e.g., Santer et al., 1996; 2003; Hansen et al., 2002). A climate model can be run with time-varying observations of just one forcing over the 20th Century to study the temperature response in the vertical. Then, by running more single forcings, a picture emerges concerning the effects of each one individually. The model can then be run with a combination of forcings to determine the degree to which the simulation resembles the observations made in the 20th Century. Note that a linear additivity of responses, while approximately valid for certain combinations of forcings, need not hold in general (Ramaswamy and Chen, 1997; Hansen et al., 1997; Santer et al., 2003; Shine et al., 2003). To first order, models are able to reproduce the time evolution of globally averaged surface air temperature over the 20th Century, with the warming in the first half of the century generally ascribed to natural forcings (mainly volcanoes and solar) or unforced variations, and the warming in the late 20th Century mostly due to human-induced increases of GHGs (Stott et al., 2000; Mitchell et al., 2001; Meehl et al., 2003; 2004; Broccoli et al., 2003). Such modeling studies used various observed estimates of the forcings, but uncertainties remain regarding details of such factors as solar variability (Frohlich and Lean, 2004), historical volcanic forcing (Bradley, 1988), and aerosols (Charlson et al., 1992; Anderson et al., 2003).

Analyses of model responses to external forcings also require consideration of the internal variability of the climate system for a proper causal interpretation of the observed surface temperature record. For example, the mid-1970s saw the beginning of a significant increase of global temperatures that was associated with an apparent regime shift in the Pacific (Trenberth and Hurrell, 1994). While it has been argued that the warming could have been a delayed response to a regime shift due to the heat capacity of the ocean (Lindzen and Giannitsis, 2002), this increase in temperature starting in the 1970s is also simulated as a response due to changes in external forcing in the models noted above. The relationship between external forcing and internal decadal variability of the climate system (i.e., can the former influence the latter, or are they totally independent) is an intriguing research problem that is being actively studied.

In addition to the analyses of surface temperatures outlined above, climate models can also show the expected effects of different forcings on temperatures in the vertical. For example using simplified ocean representations for equilibrium $2XCO_2$, Hansen et al. (2002) show that changes of various anthropogenic and natural forcings produce different patterns of temperature change horizontally and vertically. Hansen et al. (2002) also show considerable sensitivity of the simulated vertical temperature response to the choice of ocean representation, particularly for the GHG-only and solar-only cases. For both of these cases, the “Ocean A” configuration (SSTs prescribed according to observations) lacks a clear warming maximum in the upper tropical troposphere, thus illustrating that there could be some uncertainty in our model-based estimates of the upper tropospheric temperature response to GHG forcing (see Chapter 5).

An illustration of the effects of different forcings on the trends in atmospheric temperatures at different levels from a climate model with time-varying forcings over the latter part of the 20^{th} Century is shown in Figure 1.3. Here, the temperature changes are calculated over the time period of 1979-1999, and are averages of four-member ensembles. As in Hansen et al., this model, the NCAR/DOE Parallel Climate Model (PCM, e.g., Meehl et al., 2004) shows warming in the troposphere and cooling in the stratosphere for an increase of GHGs, warming through most of the stratosphere and a slight cooling in the troposphere for volcanic aerosols, warming in a substantial portion of the atmosphere for an increase in solar forcing, warming in the troposphere from increased tropospheric ozone and cooling in the stratosphere due to the decrease of stratospheric ozone, and cooling in the troposphere and slight warming in the stratosphere from sulfate aerosols. The multiple-forcings run shows the net effects of the combination of these forcings as a warming in the troposphere and a cooling in the stratosphere. Note that these simulations may not provide a full accounting of all factors that could affect the temperature structure, e.g., black carbon aerosols, land use change (Ramaswamy et al., 2001; Hansen et al., 1997; 2002; Pielke, 2001; NRC, 2005; Ramanathan et al., 2005).

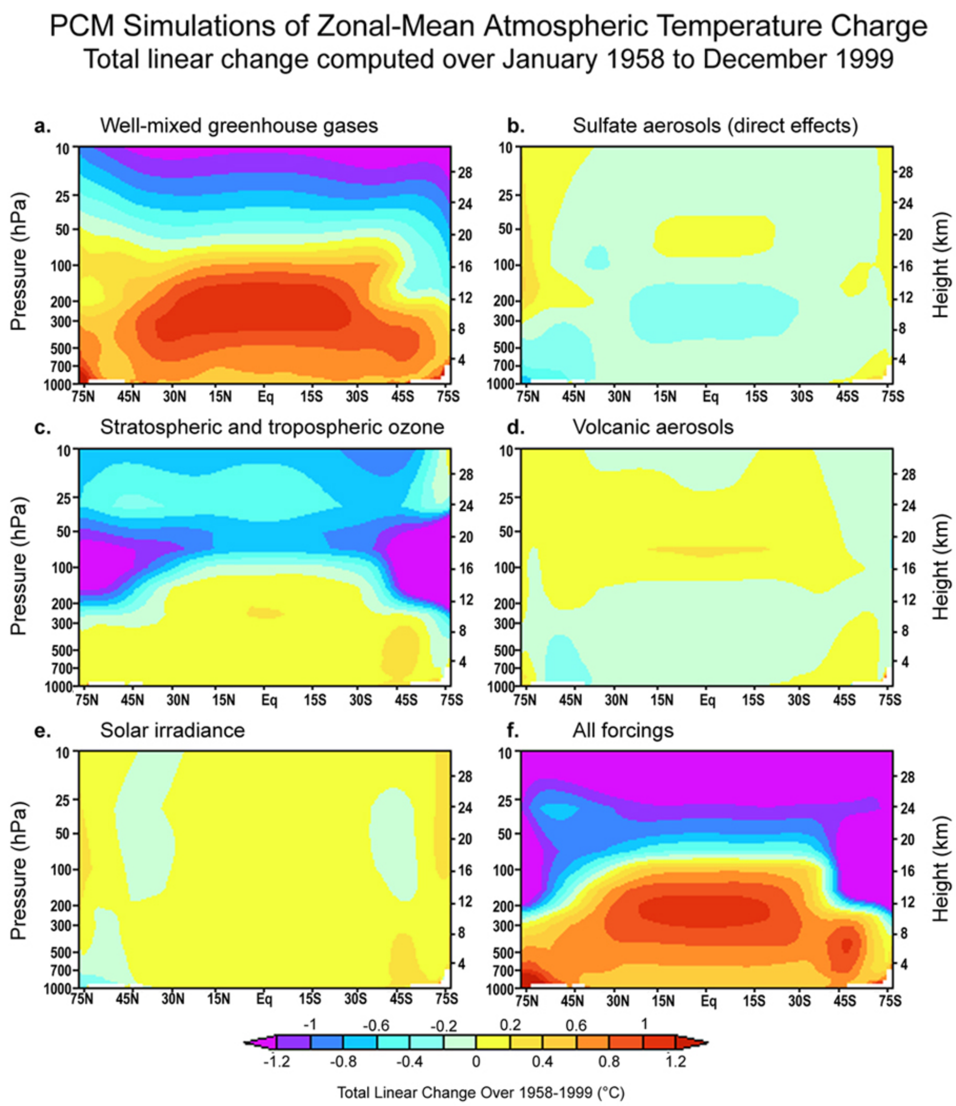

Figure 1.3. PCM simulations of the vertical profile of temperature change due to various forcings, and the effect due to all forcings taken together (after Santer et al., 2000).

The magnitude of the temperature response for any given model is related to its climate sensitivity. This is usually defined either as the equilibrium warming due to a doubling of CO_2 with an atmospheric model coupled to a simple slab ocean, or the transient climate response (warming at time of CO_2 doubling in a 1% per year CO_2 increase experiment in a global coupled model). The climate sensitivity varies among models due to a variety of factors (Cubasch et al., 2001; NRC, 2004).

The important conclusion here is that representations of the major relevant forcings are important to simulate 20th Century temperature trends since different forcings affect temperature differently at various levels in the atmosphere.

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

Tropospheric and surface temperatures, although linked, are separate physical entities (Trenberth et al., 1992; Hansen et al., 1995; Hurrell and Trenberth, 1996; Mears et al., 2003). Insight into this point comes from an examination of the correlation between anomalies in the monthly-mean surface and tropospheric temperatures over 1979-2003 (Figure 1.4). The correlation coefficients between monthly surface and tropospheric temperature anomalies (as represented by temperatures derived from MSU satellite data) reveal very distinctive patterns, with values ranging from less than zero (implying poor vertical coherence of the surface and tropospheric temperature anomalies) to over 0.9. The highest correlation coefficients (>0.75) are found across the middle and high latitudes of Europe, Asia, and North America, indicating a strong association between the surface and tropospheric monthly temperature variations. Correlations are generally much less (~0.5) over the tropical continents and the North Atlantic and North Pacific Oceans. Correlations less than 0.3 occur over the tropical and southern oceans and are lowest (<0.15) in the tropical western Pacific. Relatively high correlation coefficients (>0.6) are found over the tropical eastern Pacific where the ENSO signal is large and the sea-surface temperature fluctuations influence the atmosphere significantly.

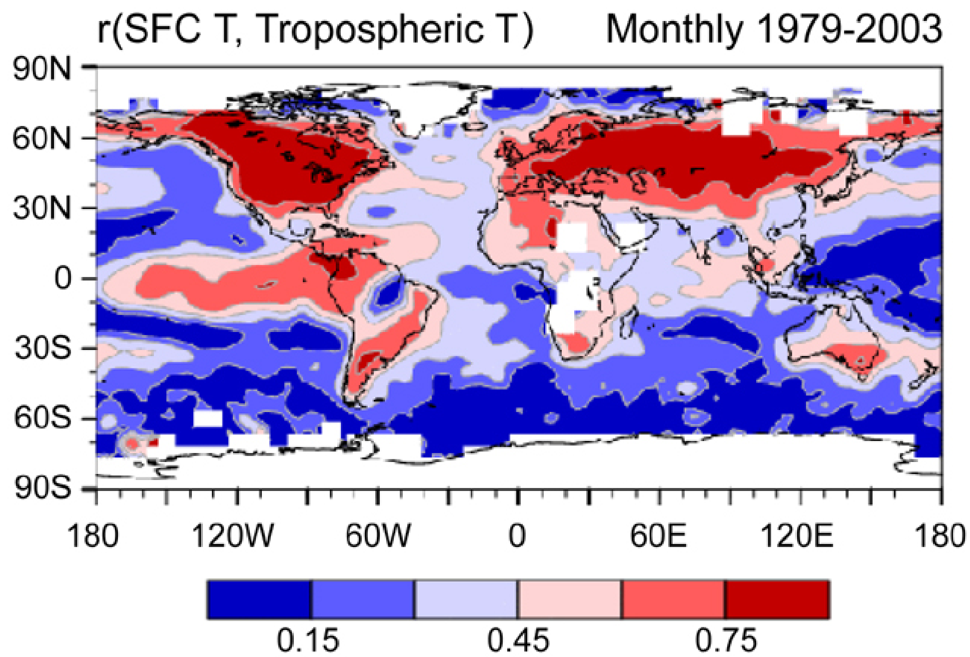

Figure 1.4. Gridpoint correlation coefficients between monthly surface and tropospheric temperature anomalies over 1979-2003. The tropospheric temperatures are derived from MSU satellite data (Christy et al., 2003).

Differences between the surface and tropospheric temperature records are found where there is some degree of decoupling between the layers of the atmosphere. For instance, as discussed earlier, over portions of the subtropics and tropics, variations in surface temperature are disconnected from those aloft by a persistent trade-wind inversion. Shallow temperature inversions are also commonly found over land in winter, especially in high latitudes on sub-seasonal timescales, so that there are occasional large differences between monthly surface and tropospheric temperature anomalies.

More important than correlations for trends, however, is the variability of the two temperature records, assessed by computing the standard deviation of the measurement samples of each record (Figure 1.5). The figure exhibits pronounced regional differences in variability between the surface and tropospheric records. Standard deviations also help in accounting for the differences in correlation coefficients, because they yield information on the size and persistence of the climate signal relative to the noise in the data. For instance, large variations in eastern tropical Pacific sea surface temperature associated with ENSO dominate over measurement uncertainties, as do large month-to-month swings in surface temperatures over extratropical continents.

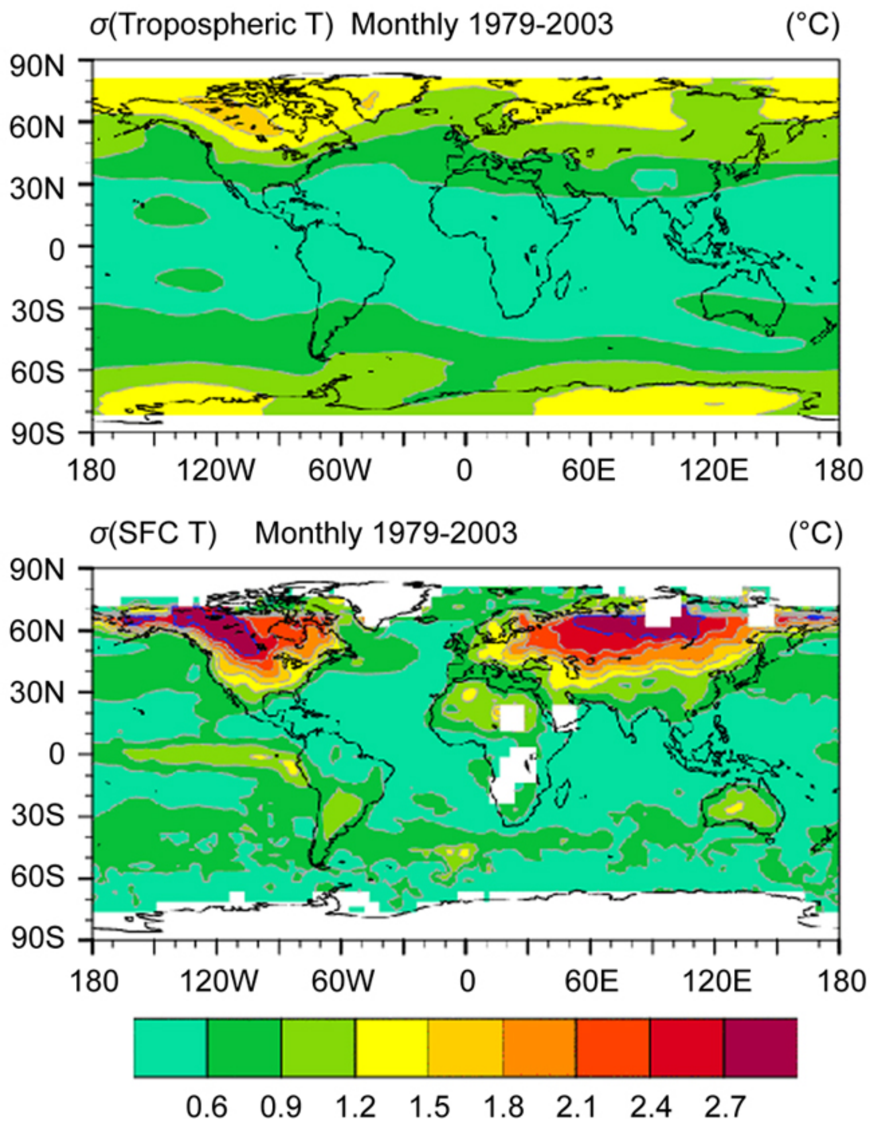

Figure 1.5. Standard deviations of monthly mean temperature anomalies from the surface and tropospheric temperature records over 1979-2003. The tropospheric temperatures are derived from MSU satellite data (Christy et al., 2003).

The largest variability in both surface and tropospheric temperature is over the Northern Hemisphere continents. The standard deviation over the oceans in the surface data set is much smaller than over land except where the ENSO phenomenon is prominent. The standard deviations of tropospheric temperature, in contrast, exhibit less zonal variability. Consequently, the standard deviations of the monthly tropospheric temperatures are larger than those of the surface data by more than a factor of two over the North Pacific and North Atlantic. Over land, tropospheric temperatures exhibit slightly less variability than surface temperatures. These differences in variability are indicative of differences in physical processes over the oceans versus the continents. Of particular importance are the roles of the land surface and ocean as the lower boundary for the atmosphere and their very different abilities to store heat, as well as the role of the atmospheric winds that help to reduce regional differences in tropospheric temperature through the movement of heat from one region to another.

Over land, heat penetration into the surface involves only the upper few meters, and the ability of the land to store heat is low. Therefore, land surface temperatures vary considerably from summer to winter and as cold air masses replace warm air masses and vice versa. The result is that differences in magnitude between surface and lower-atmospheric temperature anomalies are relatively small over the continents: very warm or cold air aloft is usually associated with very warm or cold air at the surface. In contrast, the ability of the ocean to store heat is much greater than that of land, and mixing in the ocean to typical depths of 50 meters or more considerably moderates the sea surface temperature response to cold or warm air. Over the northern oceans, for example, a very cold air mass (reflected by a large negative temperature anomaly in the tropospheric record) will most likely be associated with a relatively small negative temperature anomaly at the sea surface. This is one key to understanding the differences in trends between the two records.

Long-term changes in the atmospheric circulation can be reflected by trends in indices of patterns (or modes) of natural climate variability such as ENSO, the North Atlantic Oscillation (NAO; also known as the Northern Hemisphere annular mode, or NAM), and the Southern Hemisphere (SH) annular mode (SAM). The exact magnitudes of the index trends depend on the period of time examined. Over the past several decades, for instance, changes in atmospheric circulation (reflected by a strong upward trend in indices of the NAO) have contributed to a Cold Ocean Warm Land (COWL) surface temperature pattern over the Northern Hemisphere (Hurrell, 1996; Thompson and Wallace, 2000). In the lower atmosphere, winds blowing from ocean to land to ocean are much stronger than at the surface, and this moderating influence of the winds contributes to less east-west variability in the tropospheric data (Figure 1.5). Thus, the recent warm anomalies over the continents are roughly cancelled by the cold anomalies over the oceans in the tropospheric dataset. This is not the case for the surface temperature record, which is dominated by the warmth over the continents. The result is that the changes in the Northern Hemisphere (NH) atmospheric circulation over the past few decades have produced a significant difference in surface and tropospheric temperature trends (Hurrell and Trenberth, 1996). Similarly, Thompson and Solomon (2002) showed that recent tropospheric temperature trends at high southern latitudes were related to changes in the SAM.

Physical differences between the two measures of temperature are also evident in their dissimilar responses to volcanic eruptions and ENSO (e.g., Santer et al. 2000). These phenomena have a greater effect on tropospheric than surface temperature, especially over the oceans (Jones, 1994). However, changes in ENSO over the past several decades do not explain long-term changes in tropical tropospheric temperatures (Hegerl and Wallace, 2002). Changes in concentrations of stratospheric ozone could also be important, as the troposphere is cooled more by observed ozone depletion than is the surface (Hansen et al., 1995 ; Ramaswamy et al., 1996). Another contributing factor could be that at the surface, the daily minimum temperature has increased at a faster rate than the daily maximum, resulting in a decrease in the diurnal temperature range over many parts of the world (e.g., Easterling et al., 1997; Dai et al., 1999). Because of nighttime temperature inversions, the increase in the daily minimum temperatures likely involves only a shallow layer of the atmosphere that would not be evident in upper-air temperature records.

These physical processes provide indications of why trends in surface temperatures are expected to be different from trends in the troposphere, especially in the presence of strong interannual variability, even if both sets of measurements were perfect. Of course they are not, as described in more detail in Chapter 2, which deals with the strengths and limitations of current observing systems. An important issue implicit in Figure 1.5 is that of spatial sampling, and the accompanying caveats about interpretation of the truly global coverage provided by satellites versus the incomplete space and time coverage offered by radiosondes. These are discussed in depth in Chapters 2 and 3.

References

Anderson, T.L., R.J. Charlson, S.E. Schwartz, R. Knutti, O. Boucher, H. Rodhe, and J. Heitzenberg, 2003: Climate forcing by aerosols—a hazy picture. *Science*, **300**, 1103—1104.

Andrews, D. G., J. R. Holton, and C. B. Leovy, 1987: Middle atmosphere dynamics. Academic Press, Florida, 489pp.

Bradley, R.S., 1988: The explosive volcano eruption signal in Northern Hemisphere continental temperature records. *Climatic Change*, **12**, 221—243.

Broccoli, A., K. Dixon, T. Delworth, T. Knutson and R. Stouffer, 2003: Twentieth-century temperature and precipitation trends in ensemble climate simulations including natural and anthropogenic forcing.

Charlson, R.J., S.E. Schwartz, J.M. Hales, R.D. Cess, J.A. Coakley, J.E. Hansen and D.J. Hoffman, 1992: Climate forcing by anthropogenic aerosols. *Science*, **255**, 423—430.

Christy, J.R., R.W. Spencer, W.B. Norris, W.D. Braswell and D.E. Parker, 2003: Error estimates of Version 5.0 of MSU/AMSU bulk atmospheric temperatures. *J. Atmos. Oceanic Tech*., 20, 613-629.

Chung, C., V. Ramanathan and J. Kiehl, 2002: Effects of the South Asian absorbing haze on the northeast monsoon and surface-air heat exchange. *J. Climate*, **15**, 2462-2476.

Cubasch, U., 2001: Projections of future climate change. In Climate Change 2001: The Scientific Basis. *Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change* [Houghton, J. T. et al (eds.)]. Cambridge University Press, Cambridge, UK, and New York, NY, USA, 881 pp.

Dai, A., K. E. Trenberth, and T. R. Karl, 1999: Effects of clouds, soil moisture, precipitation, and water vapor on diurnal temperature range. *J. Climate*, **12**, 2451– 2473.

Donner, L., C. J. Seman, R. S. Hemler, and S. Fan, 2001: A cumulus parameterization including mas fluxes, convective vertical velocities, and mesoscale effects: thermodynamic and hydrological aspects in a general circulation model. *J. Climate,* **14**, 3444-3463.

Easterling, D. R., and Coauthors, 1997: Maximum and minimum temperature trends for the globe. *Science*, **277**, 364–367.

Erlick, C. E., and V. Ramaswamy, 2003: Sensitivity of the atmospheric lapse rate to solar cloud absorption in a radiative-convective model. *J. Geophys. Res.*, **108(D16)**, 4522, doi:10.1029/2002JD002966.

Frohlich, C. and J. Lean, 2004: Solar radiative output and its variability: Evidence and mechanisms. *Astron. Astrophys. Rev.*, **12**, 273—320.

Gillett, N., and D. W. J. Thompson, 2003: Simulation of recent Southern Hemisphere climate change. *Science*, **302**, 273-275.

Goody, R. M. and Y. L. Yung, 1989: Atmospheric radiation: A theoretical basis. Oxford University Press, New York, 519pp.

Hansen, J. E. et al., 1995: satellite and surface temperature data at odds? *Clim. Change*, **30**, 103-117.

Hansen, J. E., M. Sato and R. Ruedy, 1997: Radiative forcing and climate response, *J. Geophys. Res*., **102**, 6831-6864.

Hansen, J. and co-authors, 2002: Climate forcings in Goddard Institute for Space Studies SI2000 simulations, *J. Geophys. Res*., **107**(D18), 4347, doi:10.1029/2001JD001143.

Hegerl, G. C., and J. M. Wallace, 2002: Influence of patterns of climate variability on the difference between satellite and surface temperature trends. *J. Climate*, **15**, 2412- 2428.

Holton, J.R., 1979: An introduction to dynamic meteorology. Academic Press, New York, p. 286.

Houghton, J. T., 1977: The physics of atmospheres. Cambridge University Press, UK, 203pp.

Hoskins, B. J., 2003: Atmospheric processes and observations. *Philos. Tr. Soc. S.-A* **363**, 1945-1960.

Hurrell, J. W., 1996: Influence of variations in extratropical wintertime teleconnections on Northern Hemisphere temperatures. *Geophysical Research Letters*, **23**, 665-668.

Hurrell, J. W., and K. E. Trenberth, 1996: Satellite versus surface estimates of air temperature since 1979. *Journal of Climate*, **9**, 2222-2232.

Jones, P. D., 1994: Hemispheric surface air temperature variations: a reanalysis and an update to

1993. *J. Climate*, **7**, 1794-1802.

Jones, P.D., A. Moberg, T.J. Osborn, and K.R. Briffa, 2003: Surface climate responses to explosive volcanic eruptions seen in long European temperature records and mid-to-high latitude tree-ring density around the Northern Hemisphere. In *Volcanism and the Earth's Atmosphere.* A. Robock and C. Oppenheimer, eds. Washington D.C.: American Geophysical Union, 239—254.

Kalnay, E. et al., 1996: The NCEP/NCAR 40-year reanalysis project. *Bull. Amer. Meteor. Soc*., **77**, 437-471.

Kiehl, J. T., 1992: *Atmospheric general circulation modeling*. In Climate System Monitoring (K. Trenberth, editor). Cambridge University Press, pp319-369.

Langematz, U., M. Kunze, K. Kruger, K. Labitzke, and G. Roff, 2003: Thermal and dynamical changes of the stratosphere since 1979 and their link to ozone and CO2 changes. *J. Geophys. Res.*, **108**, 4027, doi: 1029/2002JD002069.

Lean, J., G. Rottman, J. Harder and G. Kopp, 2005: SORCE contributions to new understanding of global change and solar variability. *Solar Phys*., in press.

Lindzen, R. S., and K. Emanuel, 2002: The greenhouse effect. Pp. 562-566 in Encyclopedia of Global Change, Environmental Change and Human Society, Vol. 1 (Andrew S. Goudie, ed.) New York. Oxford University Press, 710 pp.

Lindzen, R. S., and C. Giannitsis, 2002: Reconciling observations of global temperature change. *Geophys. Res. Lett.*, 29: 10.1029/2001GL014074.

Lohmann, U., and J. Feichter, 2005: Global indirect aerosol effects: a review. *Atmos. Chem. Physics*, **5**, 715-737.

Lohmann, U., J. Feichter, J. Penner, and R. Leaitch, 2000: Indirect effect of sulfate and carbonaceous aerosols: A mechanistic treatment. *J. Geophys. Res*., **105**, 12193- 12206.

Mahlman, J. D., J. P. Pinto, and L. J. Umscheid, 1994: Transport, radiative, and dynamical effects of the Antarctic ozone hole: A GFDL "SKYHI" experiment. *J. Atmos. Sci*., **51**, 489-508.

Manabe, S. and R. Wetherald 1967: Thermal equilibrium of the atmosphere with a given distribution of relative humidity. *J. Atmos. Sci*., **24**, 241-259.

McClatchey, R. A., et al., 1972: Optical properties of the atmosphere, third ed., AFCRL-72-0497, Air Force Cambridge Research Labs., Hanscom, MA, 110 pp.

Mears, C. A., M. C. Schabel, and F. W. Wentz, 2003: A reanalysis of the MSU channel tropospheric temperature record. *J. Climate*, **16**, 3650-3664.

Meehl, G.A., W.M. Washington, T.M.L. Wigley, J.M. Arblaster, and A. Dai, 2003: Solar and greenhouse gas forcing and climate response in the 20th century. *J. Climate*, **16**, 426--444.

Meehl, G.A., W.M. Washington, C. Ammann, J.M. Arblaster, T.M.L. Wigley, and C. Tebaldi, 2004: Combinations of natural and anthropogenic forcings and 20th century climate. *J. Climate*, in press.

Menon, S., J. Hansen, L. Nazarenko, and Y. Luo, 2002: Climate effects of back carbon aerosols in China and India. *Science*, **297**, 2250-

Mitchell, J. F. B. et al., 2001: Detection of climate change and attribution of causes. *In* Climate Change 2001: The Scientific Basis*. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change* [Houghton, J. T. et al (eds.)]. Cambridge University Press, Cambridge, UK, and New York, NY, USA, 881 pp.

National Research Council, 2003: *Understanding Climate Change Feedbacks*. The National Academies Press, Washington, D. C., 152pp.

National Research Council, 2004: *Estimating Climate Sensitivity*, in preparation.

National Research Council, 2005: *Radiative Forcing of Climate Change: Expanding the Concept and Addressing Uncertainties*, in preparation.

Oort, A. H. and J. Peixoto, 1992: Physics of Climate. American Institute of Physics, New York, 520pp.

Pielke, RA Sr., 2001: Influence of the spatial distribution of vegetation and soils on the prediction of cumulus convective rainfall. *Review of Geophysics*, **39**, 151—177.

Pielke, RA Sr., C. Davey and J. Morgan, 2004: Assessing “global warming” with surface heat content. *EOS*, **85**, no. 21, 210—211.

Ramanathan, V., and J. A. Coakley, 1978: Climate modeling through radiative-convective models. *Rev. Geophys. Space Phys*., **16**, 465-489.

Ramanathan, V. and R. Dickinson, 1979: The role of stratospheric ozone in the zonal and seasonal radiative energy balance of the Earth-tropopshere system. *J. Atmos. Sci*., **36**, 1084-1104.

Ramanathan, V., P. Crutzen, J. Kiehl and D. Rosenfeld, 2001: Aerosols, climate and the hydrological cycle. *Science*, **294**, 2119-2124.

Ramanathan, V. et al., 1989: Cloud radiative forcing and climate: Results from the earth budget radiation experiment. *Science,* **243**, 57-63.

Ramanathan, V., E. Pitcher, R. Malone and M. Blackmon, ,1983: The response of a general circulation model to refinements in radiativ processes. *J. Atmos. Sci*., **40**, 605-630.

Ramanathan, V. et al., 2005: Atmospheric brown clouds: Impacts on South Asian climate and hydrological cycle. *Proc. Natl. Acad. Sci.*, **102**, 5326-5333.

Ramaswamy, V. and C-T. Chen, 1997: Climate forcing-response relationships for greenhouse and shortwave radiative perturbations. *Geophys. Res. Lett*., **24**, 667-670.

Ramaswamy, V. and M. D. Schwarzkopf, 2002: Effects of ozone and well-mixed gases on annual-mean stratospheric temperature trends. *Geophys. Res. Lett*., **29**, 2064, doi: 10.1029/2002GL015141.

Ramaswamy, V. and V. Ramanathan, 1989: Solar absorption by cirrus cloudsand the maintenance of the tropical upper troposphere thermal structure. *J. Atmos. Sci*., **46**, 2293-2310.

Ramaswamy, V. M. D. Schwarzkopf and W. Randel, 1996: Fingerprint of ozone depletion in the spatial and temporal pattern of recent lower-stratospheric cooling. *Nature*, **382**, 616-618.

Ramaswamy, V. et al., 2001: *Radiative forcing of climate change. In* Climate Change 2001: The Scientific Basis*. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change* [Houghton, J. T. et al (eds.)]. Cambridge University Press, Cambridge, UK, and New York, NY, USA, 881 pp.

Ramaswamy, V. et al., 2001: Stratospheric temperature trends: Observations and model simulations. *Reviews of Geophys*., **39**, 71-122.

Robock, A. and C. Oppenheimer, 2003: *Volcanism and the Earth's Atmosphere.* AGU Geophysical Monograph Series, 139. Washington D.C.: American Geophysical Union. 360pp.

Santer, B. D., K. E. Taylor, T. M. L. Wigley, T. C. Johns, P. D. Jones, D. J. Karoly, J. F. B. Mitchell, A. H. Oort, J. E. Penner, V. Ramaswamy, M. D. Schwarzkopf, R. J. Stouffer, and S. Tett, 1996: A search for human influences on the thermal structure of the atmosphere**.** *Nature*, **382(6586)**, 39-46.

Santer, B. D. et al., 2000: Interpreting differential temperature trends at the surface and in the lower troposphere. *Science*, **287**, 1227-1232.

Santer, B. D. et al., 2003: Contributions of anthropogenic and natural forcing to recent tropopause height changes. *Science*, **301**, 479-483.

Sarachik, E., 1985: A simple theory for the vertical structure of the tropical atmosphere. *Pure and Appl. Geophys*., **123**, 261-271.

Schwarzkopf, M. D. and V. Ramaswamy, 2002: Effects of changes in well-mixed gases and ozone on stratospheric seasonal temperatures. *Geophys. Res. Lett*., **29**, 2184, doi: 10.1029/2002GL015759.

Schneider, T., 2004: The tropopause and the thermal stratification in the extratropics of a dry atmosphere**. *J. Atmos. Sci***., **61**, 1317-1340.

Sherwood, S. C., 2002: A microphysical connection among biomass burning, cumulus clouds, and stratospheric moisture. *Science*, **295**, 1271-1275.

Sherwood, S. C. and A. E. Dessler, 2003: Convective mixing near the tropical tropopause: insights from seasonal variations. *J. Atmos. Sci*., **60**, 2674-2685.

Shindell D. T., G. A. Schmidt, R. L. Miller and D. Rind, 2001: Northern Hemisphere winter climate response to greenhouse gas, ozone, solar and volcanic forcing. *J. Geophys. Res.*, **106**, 7193-7210.

Shine, K. P. et al., 2003: An alternative to radiative forcing for estimating the relative importance of climate change mechanisms. *Geophys. Res. Lett*., **30**, 2047, doi: 10.1029/2003GL018141.

Shine, K. P. et al., 2003: A comparison of model-simulated trends in stratospheric temperatures. *Quart. J. Roy. Met. Soc*., **129**, 1565-1588.

Stenchikov, G. et al., 2002: Arctic Oscillation response to the 1991 Mount Pinatubo eruption: Effects of volcanic aerosols and ozone depletion. *J. Geophys. Res*., **107**, 4803, doi: 10.1029/2002JD002090.

Stephens, G. L. and P. J. Webster, 1980: Clouds and climate: sensitivity of simple systems. *J. Atmos. Sci*., **38**, 235-247.

Stocker T. F. et al., 2001: Physical climate processes and fedbacks. *In* Climate Change 2001: The Scientific Basis*. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change* [Houghton, J. T. et al (eds.)]. Cambridge University Press, Cambridge, UK, and New York, NY, USA, 881 pp.

Stott, P. A. et al., 2000: External control of 20th century temperature by natural and anthropogenic forcings. *Science,* **290**,2133-2137.

Stouffer, R. J., G. Hegerl and S. Tett, 2000: A comparison of surface air temperature variability in three 1000-year coupled ocean-atmosphere model integrations. *J. Climate*, **13**, 513-537.

Thompson, D. W. J. and J. M. Wallace, 2000: Annular modes in the extratropical circulation. 1. Month-to-month variability. J. Climate, 13, 1000-1016.

Thompson, D. W., and S. Solomon, 2002: Interpretation of recent Southern Hemisphere climate change, *Science*, 296, 895—899.

Trenberth, K. E., and J. W. Hurrell, 1994: Decadal atmosphere-ocean variations in the Pacific. *Climate Dynamics*, **9**, 303-319.

Trenberth, K. E., J. R. Christy and J. W. Hurrell, 1992: Monitoring global monthly mean surface temperatures. *Journal of Climate*, **5**, 1405-1423.

Wielicki, B. A., et al., 2002: Evidence for large decadal variability in the tropical mean radiative energy budget. *Science*, **295(5556)**, 841-844.

World Meteorological Organization, 1986: *Atmospheric ozone*. Global Ozone Research and Monitoring Project, Report no. 16, 1095pp, Geneva.

World Meteorological Organization, 1999: *Scientific assessment of ozone depletion: 1998*. Global Ozone Research and Monitoring Project, Report no. 44, chapter 5, Geneva.

World Meteorological Organization, 2003: *Scientific assessment of ozone depletion: 2002*. Global Ozone Research and Monitoring Project, Report no. 47, 498pp, Geneva.

World Meteorological Organization, 2005: Special Report on Ozone and Climate. *In press*.