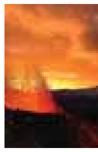
Model Climate Sensitivity



The response of climate to a perturbation such as a change in carbon dioxide concentration, or in the flux of energy from the sun, can be divided into two factors: "radiative forcing" due to the perturbation in question and "climate sensitivity," characterizing the response of the climate per unit change in radiative forcing. Climate response is then the product of radiative forcing and climate sensitivity. This distinction is useful because of two approximations: radiative forcing often can be thought of as independent of the resulting climate response, and climate sensitivity can often be thought of as independent of the agent responsible for perturbation to the energy balance. When two or more perturbations are present simultaneously, their cumulative effect can be approximated by adding their respective radiative forcings (Hansen et al. 2006).

Climate sensitivity as traditionally defined refers to the global mean temperature, but a model's global mean temperature response is very relevant to its regional temperature responses as well. This "pattern scaling" effect is discussed at the end of this chapter.

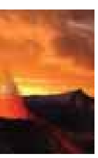
Radiative forcing typically is calculated by changing the atmospheric composition or external forcing and computing the net trapping of heat that occurs before the climate system has had time to adjust.[1] These direct heat-trapping properties are well characterized for the most significant greenhouse gases. As a result, uncertainty in climate responses to greenhouse gases typically is dominated by uncertainties in climate sensitivity rather than in radiative forcing (Ramaswamy et al. 2001). For example, suddenly doubling the atmospheric amount of carbon dioxide would add energy to the surface and the troposphere at the rate of about 4 W/m^2 for the first few months after the doubling (Forster et al. 2007). Eventually, lower tropospheric temperatures would increase (and climate would change in other ways) in response to this forcing, Earth would radiate more energy to space, and the imbalance would diminish as the system returned to equilibrium.

4.1 CHARACTERIZING CLIMATE RESPONSE

4.1.1 Equilibrium Sensitivity and Transient Climate Response

The idea of characterizing climate response using a single number represented by climate sensitivity appeared early in the development of

[1] Because the stratosphere cools rapidly in response to increasing carbon dioxide and this cooling affects the net warming of the lower atmosphere and surface, it has become standard to include the effects of this stratospheric cooling in estimating radiative forcing due to carbon dioxide.



climate models (e.g., Schneider and Mass 1975). Today, two different numbers are in common use. Both are based on changes in global and annual mean-surface or near-surface temperature. *Equilibrium sensitivity* is defined as the long-term near-surface temperature increase after atmospheric carbon dioxide has been doubled from preindustrial levels but thereafter held constant until the Earth reaches a new steady state, as described in the preceding paragraph. *Transient climate response* or TCR is defined by assuming that carbon dioxide increases by 1% per year and then recording the temperature increase at the time carbon dioxide doubles (about 70 years after the increase begins). TCR depends on how quickly the climate adjusts to forcing, as well as on equilibrium sensitivity. The climate's adjustment time itself depends on equilibrium sensitivity and on the rate and depth to which heat is mixed into the ocean, because the depth of heat penetration tends to be greater in models with greater sensitivity (Hansen et al. 1985; Wigley and Schlesinger 1985). Accounting for ocean heat uptake complicates many attempts at estimating sensitivity from observations, as outlined below.

Equilibrium sensitivity depends on the strengths of feedback processes involving water vapor, clouds, and snow or ice extents (see, e.g., Hansen et al. 1984; Roe and Baker 2007). Small changes in the strengths of feedback processes can create large changes in sensitivity, making it difficult to tightly constrain climate sensitivity by restricting the strength of each relevant feedback process. As a result, research aimed at constraining climate sensitivity—and evaluating the sensitivities generated by models—is not limited to studies of these individual feedback processes. Studies of observed climate responses on short time scales (e.g., the response to volcanic eruptions or the 11-year solar cycle) and on long time scales (e.g., the climate of last glacial maximum 20,000 years ago) also play central roles in the continuing effort to constrain sensitivity. The quantitative value of each of these observational constraints is limited by the quality and length of relevant observational records, as well as the necessity in several cases to simultaneously restrict ocean heat uptake and equilibrium sensitivity. Equilibrium warming is directly relevant when considering paleoclimates, where observations represent periods that are very long compared to the climate's adjustment time. The transient climate response is more directly relevant to the attribution of recent warming and projections for the next century. For example, Stott et al. (2006) show that global mean warming due to well-mixed greenhouse gases over the 20th Century, in the set of models they consider, is closely proportional to the model's TCR. In the following, we discuss individual feedback processes as well as these additional observational constraints on sensitivity.

Equilibrium warming in an AOGCM is difficult to obtain because the deep ocean takes a great deal of time to respond to changes in climate forcing. To avoid unacceptably lengthy computer simulations, equilibrium warming usually is estimated from a modified climate model in which the ocean component is replaced by a simplified, fast-responding "slab ocean model." This procedure makes the assumption that horizontal redistribution of heat in the ocean does not change as the climate responds to the perturbation. Current climate models generate a range of equilibrium and transient climate sensitivities. For the models in the CMIP3 archive utilized in the Fourth Assessment of the IPCC, the range of equilibrium sensitivity is 2.1 to 4.4°C with a median of 3.2°C. This ensemble of models was not constructed to systematically span the plausible range of uncertainty in climate sensitivity; rather, each development team simply provided its best attempt at climate simulation. Complementary to this approach is one in which a single climate model is modified in a host of ways to explore more systematically the sensitivity variations associated with the range of uncertainty in various key parameters. Results with a Hadley Centre model give a 5 to 95 percentile range of ~2 to 6°C for equilibrium sensitivity (Piani et al. 2005; Knutti et al. 2006).

Charney (1979) provided a range of equilibrium sensitivities to CO_2 doubling of 1.5 to 4.5°C, based on the two model simulations available at the time. Evidently, the range of model-implied climate sensitivity has not contracted significantly over three decades. The current range, however, is based on a much larger number of models subjected to a far more comprehensive comparison to observations and containing more detailed treatments of clouds and other processes that are fundamental to climate sensitivity. We understand in much more detail why models differ in their equilibrium climate sensitivities: the source of much of this spread lies in differences in how clouds are modeled in AOGCMs. Questions remain as to whether or not the substantial spread among models is a good indication of the uncertainty in climate sensitivity, given all the constraints on this quantity of which we are aware. There also is a desire to know the prospects for constraining equilibrium climate sensitivity more sharply in the near future.

The variation among models is less for TCR than for equilibrium warming, a consequence of the interrelationship between the climate's adjustment time and its sensitivity to forcing noted above (Covey et al. 2003). The full range for TCR in the CMIP3 archive is 1.3 to 2.6°C, with a median of 1.6°C and 25 to 75% quartiles of 1.5 to 2.0°C (Randall et al. 2007). Systematic exploration of model input parameters in one Hadley Centre model gives a range of 1.5 to 2.6°C (Collins, M., et al. 2006).

Table 4.1 Equilibrium and Transient Sensitivities in Some U.S. Models Contributing to CMIP3

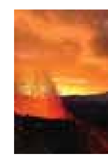
MODEL	TCR (ºC)	Equilibrium Warming (ºC)*
CSM1.4	1.4	2.0
CCSM2	1.1	2.3
CCSM3	1.5	2.5
GFDL CM2.0	1.6	2.9
GFDL CM2.1	1.5	3.4
GISS Model E		2.7 to 2.9

**Equilibrium warming was assessed by joining a simplified slab ocean model to the atmosphere, land, and sea-ice AOGCM components.*

[Sources of Information in table. First three lines – J.T. Kiehl et al. 2006: The climate sensitivity of the Community Climate System Model: CCSM3. *J. Climate*, **19**, 2584–2596. Next two lines – R.J. Stouffer et al. 2006: GFDL's CM2 global coupled climate models. Part IV: Idealized climate response. *J. Climate*, **19**, 723–740. Last line – J. Hansen et al. 2007: Climate simulations for 1880–2003 with GISS ModelE. *Climate Dynamics*, **29**(7–8), 661–696.]

The equilibrium and transient sensitivities in some models developed by U.S. centers contributing to CMIP3 are listed in Table 4.1. In the last column, the larger of the two GISS ModelE values is obtained using a full ocean model in which the circulation is allowed to adjust. All other values of equilibrium warming in the table are obtained with the ocean component replaced by a slab ocean model. The close agreement in transient climate sensitivity among models in this subset should not be overinterpreted, given the larger range among the full set of CMIP3 models.

Climate sensitivity is not a model input. It emerges from explicitly resolved physics, subgrid-scale parameterizations, and numerical approximations used by the models—many of which differ from model to model—particularly those related to clouds and ocean mixing. The climate sensitivity of a model can be changed by modifying parameters that are poorly constrained by observations or theory. Influential early papers by Senior and Mitchell (1993, 1996) demonstrated how a seemingly minor modification to the cloud-prediction scheme alters climate sensitivity. In the standard version of the model, the effective size of cloud drops was fixed. In two other versions, this cloud-drop size was tied to the total amount of liquid-water cloud through two different empirical relationships. The equilibrium sensitivity ranged from 1.9 to 5.5°C in these three models. In general, the nonlinear dependence of equilibrium sensitivity on the strength of feedback processes allows relatively small changes in feedbacks to generate large changes in sensitivity (see, e.g., Hansen et al. 1984; Roe and Baker 2007).



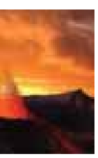
Studies of the CCSM family of models provide another example of this problem. Kiehl et al. (2006) found that a variety of factors is responsible for differences in climate sensitivity among the models of this family. However, the lower TCR of CCSM2 (relative to CSM1.4 and CCSM3), evident in Table 4.1, results primarily from a single change in the model's algorithm for simulating convective clouds. Table 4.2 shows how equilibrium sensitivity varied during development of the most recent GFDL models. The dramatic drop in sensitivity between model versions p10 and p12.5.1 was unex-

Table 4.2 Equilibrium Global Mean Near-Surface Warming Due to Doubled Atmospheric Carbon Dioxide from Intermediate (“p”) Model Versions Leading to GFDL’s CM2.0 and CM2.1

MODEL VERSION	Equilibrium Warming (ºC)*
p7	3.87
p9	4.28
p10	4.58
p12.5.1	2.56
p12.7	2.65
p12.10b	2.87
p12b	2.83
CM 2.0	2.90
CM 2.1	3.43

**Equilibrium warming was assessed by joining a simplified slab ocean model to the atmosphere, land, and sea-ice AOGCM components.*

[Source of information for table: Personal communication with Thomas Knutson, NOAA GFDL laboratory.]



pected. It followed a reformulation of the model’s treatment of processes in the lower atmospheric boundary layer, which, in turn, affected how low-level clouds in the model respond to climate change.

4.1.2 Observational Constraints on Sensitivity

Climate models in isolation have not yet converged on a robust value of climate sensitivity. Furthermore, the actual climate sensitivity in nature might not be found in the models’ range of sensitivities, since all the models may share common deficiencies. However, observations can be combined with models to constrain climate sensitivity. The observational constraints include the response to volcanic eruptions; aspects of the internal variability of climate that provide information on the strength of climatic “restoring forces”; the response to the 11-year cycle in solar irradiance; paleoclimatic information, particularly from the peak of the last Ice Age some 20,000 years ago; aspects of the seasonal cycle; and, needless to say, the magnitude of observed warming over the past century.

4.1.2.1 Volcanic Eruptions

Volcanoes provide a rapid change in radiative forcing due to the scattering and absorption of solar radiation by stratospheric volcanic aerosol. Of special importance, recovery time after the eruption contains information about climate sensitivity that is independent of uncertainties in the magnitude of the radiative forcing perturbation (e.g., Lindzen and Giannitsis 1998). Larger climate sensitivity implies weaker restoring forces on Earth’s temperature, and, therefore, a slower relaxation back toward the unperturbed climate. However, this time scale also is affected by the pathways through which heat anomalies propagate into the ocean depths, with deeper penetration increasing the relaxation time. Several modeling studies have confirmed that this relaxation time after an eruption increases as climate sensitivity increases in GCMs when holding the ocean model fixed (Soden et al. 2002; Yokohata et al. 2005), encouraging the use of volcanic responses to constrain sensitivity. On the other hand, Boer, Stowasser, and Hamilton (2007) study two models with differing climate sensitivity and different ocean models; they highlight the difficulty in determining which model has the higher sensitivity from the surface-temperature responses to volcanic forcing in isolation, without quantitative information on ocean heat uptake.

Some studies have argued that observations of responses to volcanoes imply that models are overestimating climate sensitivity (e.g., Douglass and Knox 2005; Lindzen and Giannitsis 1998). These studies argue that observed relaxation times are shorter than those expected if climate sensitivity is as large as in typical AOGCMs. Studies that directly examine the volcanic responses in AOGCMs, however, find no such gross disagreement with observations (Wigley et al. 2005; Boer, Stowasser, and Hamilton 2007; Frame et al. 2005) consistent with earlier studies (e.g., Hansen et al. 1996; Santer et al. 2001). They nevertheless consistently suggest (Frame et al. 2005; Yokohata et al. 2005) that climate sensitivities as large as

6ºC are inconsistent with observed relaxation times. It is important to note that these "observational" studies of climate sensitivity that do not utilize GCMs still make use of models, but they use simple energy balance "box" models rather than GCMs. The value of these studies depends on the relevance of the simple models as well as on the techniques for estimating parameters in models that control climate sensitivity. From these analyses, one can infer that further research isolating changes in ocean heat content after eruptions, such as that of Church, White, and Arblaster (2005), will be needed to strengthen constraints on climate sensitivity provided by responses to volcanic eruptions.

4.1.2.2 Natural Climate Variability

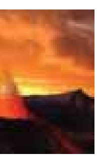
Natural variability of climate also provides a way of estimating the strength of the restoring forces that determine climate sensitivity. Just as investigators learn something about sensitivity by watching the climate recover from a volcanic eruption, they can hope to obtain similar information by watching the climate relax from an unforced period of unusual global warmth or cold. This approach to constraining the response to a perturbation by examining the character of a system's natural variability, discussed by Leith (1975) in the context of climate sensitivity, is referred to as "fluctuation-dissipation" analysis in other branches of physics. In the case of equilibrium statistical mechanics, this relationship between characteristics of natural variability and response to an external force has been placed on a firm theoretical footing, but application to the climate is more heuristic, generally depending on approximation of the climate system by a linear stochastically forced model. The power of the approach is illustrated by Gritsun and Branstator (2007) in a study of the extratropical atmosphere's response to a perturbation in tropical heating. A recent attempt to apply this approach to climate sensitivity can be found in Schwartz (2007). This technique deserves more attention, with careful analysis of uncertainties. Its value likely will be determined by its ability to infer an AOGCM's sensitivity from an analysis of its internal variability.

4.1.2.3 Solar Variations

The 11-year solar cycle has potential for providing very useful information on climate sensitivity. Total solar irradiance is known to vary by roughly 0.1% over this cycle (Frölich 2002). The expected response in global mean temperature is only ~0.1ºC, so the technique is limited in value by the quality and length of the observational record, both of which restrict our ability to isolate this small signal. Recent results show promise in more cleanly identifying the climatic response to this cyclic perturbation (Camp and Tung 2007). Since ultraviolet wavelengths play a disproportionately larger role in these cyclic variations, detailed representations of the stratosphere and mesosphere, where ultraviolet radiation is absorbed, along with ozone chemistry are required for quantitative analysis of climatic response to the solar cycle (e.g., Shindell et al. 2006). Solar variations also have been invoked repeatedly to explain early 20th Century warming and to connect the Little Ice Age to the Maunder Minimum in sunspot number. While these connections may very well have a valid basis, using them to constrain climate sensitivity remains difficult as long as variations in insolation on time scales longer than the 11-yr cycle are not better quantified. To illustrate the difficulty, we note the substantial reduction in estimated insolation variations in the 20th Century between the Third and Fourth IPCC Assessments (Forster et al. 2007). Further analyses of responses to the sunspot cycle in models and observations seem likelier to lead to stronger constraints on climate sensitivity in the near term.

4.1.2.4 Glacial-Interglacial Variations

The glacial-interglacial fluctuations of the Pleistocene (the Ice Ages) are thought to be forced by changes in the Earth's orbit on time scales of 20,000 years and longer—the astronomical theory of the Ice Ages. Since this theory assumes that the mean temperature of the Earth can be altered by changing the distribution of the incoming solar flux without changing its global mean, it suggests important limitations to simple models based solely on global mean radiative forcing. For the limited purpose of constraining climate sensitivity, we need not understand how glacial-interglacial variations of ice sheets and of carbon dioxide are forced by changes in the Earth's orbit. Since we have knowledge from ice cores of greenhouse gas concentrations at the peak of the last major gla-



cial advance 20,000 years ago as well as considerable information on the extent of continental ice sheets, one may ask if climate models can simulate the ocean-surface temperatures inferred from a variety of proxies, given these greenhouse gas concentrations and ice sheets (Manabe and Broccoli 1985). A logical assumption is that models that are more sensitive to doubling of carbon dioxide would also simulate larger cooling during the low carbon dioxide levels 20,000 years ago. Crucifix (2006) describes some of the difficulties with this simple picture. Annan and Hargreaves (2006) argue that the tropics and Antarctica are regions where this connection may be the strongest. Model results generated in the Paleoclimate Modelling Intercomparison Project (Braconnet et al. 2007a, b; Crucifix et al. 2006)) provide a valuable resource for analyzing these relationships. Despite these complications, several studies agree that past climates are difficult to reconcile with the low end of the equilibrium-sensitivity range generated by models (e.g., Hansen et al. 1993; Covey, Sloan, and Hoffert 1996). Models of the last glacial maximum also provide some of the strongest evidence that climate sensitivity is very unlikely to be larger than 6°C (Annan et al. 2005; Annan and Hargreaves 2006). As paleoclimatic reconstructions for this period improve, these simulations will become of greater quantitative value. Uncertainty in Ice Age aerosol concentrations may be the most difficult obstacle to overcome.

4.1.2.5 Seasonal Variation

The seasonal cycle is a familiar forced climate response to changes in the Earth-sun geometry and, therefore, should yield information on climate sensitivity. Although the seasonal cycles of global (Lindzen 1994) and hemispheric (Covey et al. 2000) mean temperature are not themselves strongly related to equilibrium climate sensitivity, regional variations and other aspects of the seasonal cycle may constrain sensitivity. Knutti et al. (2006) provide an example of a methodology using ensembles of climate model simulations to search for variables, or combinations of variables, that correlate with climate sensitivity (see also Shukla et al. 2006). If such a variable that predicts climate sensitivity in models is found, investigators can then examine its value in observations and hope thereby to constrain climate sensitivity. Knutti et al. (2006) use a neural network to look for aspects of the seasonal cycle with this predictive capability, with some success. Their study favors sensitivity in the middle of the typical model range (near 3˚C).

The work of Qu and Hall (2006) provides an especially straightforward example of this approach. They do not address climate sensitivity directly but only the strength of one feedback mechanism that contributes to sensitivity: snow-albedo feedback (the decrease in reflection of solar radiation by snow as the snowcover retreats in a warming climate). Qu and Hall demonstrate that the strength of this feedback in models is strongly correlated to the seasonal cycle of the snow cover simulated by the models. Comparison of observed and simulated seasonal cycles of snow cover then suggest which model simulations of snow albedo feedback are the most reliable. These studies suggest that detailed comparisons of modeled and observed seasonal cycles should provide valuable information on climate sensitivity in the future.

The observed 20th Century warming is a fundamental constraint on climate models, but it is less useful than one might think in constraining sensitivity because of the large uncertainty in forcing due to anthropogenic aerosols. Twentieth Century simulations are important in demonstrating the consistency of certain combinations of sensitivity, aerosol forcing, and ocean-heat uptake, but they do not provide a sharp constraint on sensitivity in isolation (Kiehl 2007). Further discussion of 20th Century simulations can be found in Chapter 5.

Rather than focusing on one particular observational constraint or on models in isolation, attempts to combine some or all of these observational constraints with model simulations are recognized as the most productive approaches to constraining climate sensitivity (Bierbaum et al. 2003; Randall et al. 2007; Stott and Forest 2007). As an example, while model ensembles in which parameters are varied sys-

[2] Estimating the probability of very high climate sensitivities above the high end of the CMIP3 model range, even if these probabilities are low, can be relevant for analyses of unlikely but potentially catastrophic climate change. It is not within the scope of this report to attempt to quantify these probabilities.

tematically can include models with sensitivities larger than 6ºC (Stainforth et al. 2005; Roe and Baker 2007), these very high values can be excluded with high confidence through comparisons with observations of volcanic relaxation times and simulations of the last glacial maximum. As summarized by Randall et al. (2007) in the Fourth IPCC Assessment, these multiconstraint studies are broadly consistent with the spread of sensitivity in the CMIP3 models.[2]

4.2 FEEDBACKS

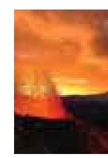
Better understanding of Earth's climate sensitivity, with potential reduction in its uncertainty, will require better understanding of a variety of climate feedback processes (Bony et al. 2006). We discuss some of these processes in more detail below.

4.2.1 Cloud Feedbacks

Clouds reflect solar radiation to space, cooling the Earth-atmosphere system. Clouds also trap infrared radiation, keeping the Earth warm. The integrated net effect of clouds on climate depends on their height, location, microphysical structure, and evolution through the seasonal and diurnal cycles. Cloud feedback refers to changes in cloud amounts and properties that can either amplify or moderate a climate change. Differences in cloud feedbacks in climate models have been identified repeatedly as the leading source of spread in model-derived estimates of climate sensitivity (beginning with Cess et al. 1990). The fidelity of cloud feedbacks in climate models therefore is important to the reliability of their prediction of future climate change.

Soden and Held (2006) evaluated cloud feedbacks in 12 CMIP3 AOGCMs and found weakly to strongly positive cloud feedback in the various models. The highest values of cloud feedback raise the equilibrium climate sensitivity (for CO_2 doubling) from values of about 2 K to roughly 4 K. In comparison with the earlier studies of Cess (1990) and Colman (2003), the spread of cloud feedbacks among GCMs has become somewhat smaller over the years but is still very substantial. Indeed, intermodel differences in cloud feedback are the primary reason that models disagree in their estimates of equilibrium climate sensitivity; which (if any) models give accurate cloud simulations remains unclear (Randall et al. 2007) as debate over specific processes continues (Spencer et al. 2007)

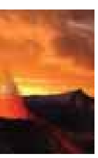
Examples of competing hypotheses concerning high clouds (for which the infrared trapping effects are large) are the IRIS hypothesis of Lindzen, Chou, and Hou (2001) and the FAT (Fixed Anvil Temperature Hypothesis) of Hartmann and Larsson (2002). The IRIS hypothesis asserts that warmer temperatures cause the area coverage of clouds in the tropical upper troposphere to decrease, a negative feedback since these clouds are infrared absorbers. The FAT hypothesis asserts that the altitude of these tropical high clouds tends to increase with warming, minimizing the temperature change at the cloud tops—a positive feedback since the lack of warming at cloud top prevents the increase in outgoing radiation needed to balance the heat trapping of greenhouse gases. Observational studies aimed at evaluating these mechanisms are difficult because clouds in the tropics are strongly forced by circulations that are, in turn, driven by temperature gradients and not by the local temperature in isolation. These circulation effects must be eliminated to isolate effects relevant to global warming. Very high resolution simulations in localized regions have some potential to address these questions. The FAT hypothesis, in particular, has received some support from high-resolution modeling (Kuang and Hartmann 2007).



Although these studies focus on high clouds, the intermodel differences in model responses of low-level clouds are responsible for most of the spread of cloud feedback values in climate models (Bony et al. 2006). While tempting, assuming that this implies that low-cloud feedbacks are more uncertain than high-cloud feedbacks probably is premature. The strengths and weaknesses of cloud-cover simulations for present-day climate are described in Chapter 5.

As discussed in Chapter 6, a new class of much higher resolution global atmospheric simulations promises fundamental improvements in cloud simulation. Using the surrogate climate change framework of Cess (1990) in which ocean temperatures are warmed uniformly, Miura et al. (2005) carried out experiments using a global model with 7-km resolution, obtaining results suggestive of negative cloud feedback outside the tropics, and Wyant et al. (2006) describe results from a multigrid technique in which high-resolution cloud models are embedded in each grid box of a traditional GCM. Much work will be required with these new types of models before they can be given substantial weight in discussions of the most probable value for cloud feedback, but they suggest that real-world feedback is less positive than the typical CMIP3 AGCMs and that midlatitude cloud feedbacks may be more important than hitherto assumed. Results from this new generation of models will be of considerable interest in the coming years.

Several questions remain to be answered about cloud feedbacks in GCMs. Physical mechanisms underlying cloud feedbacks in different models must be better characterized. How best to judge the importance of model biases in simulations of current climate and in simulations of cloud changes in different modes of observed variability is not clear. In particular, how to translate these biases into levels of confidence in simulations of cloud feedback processes in climate change scenarios is unclear. New satellite products such as those from active radar and lidar systems should play a central role in cloud research in coming years by providing more comprehensive space-time cloud datasets.



4.2.2 Water-Vapor Feedbacks

Analysis of radiative feedbacks in the CMIP3 models (Soden and Held 2006) reaffirms that water-vapor feedback—the increase in heat trapping due to the increase in water vapor as the lower atmosphere warms—is fundamental to the models' climate sensitivity. The strength of their water-vapor feedback typically is close in magnitude to but slightly weaker than that obtained by assuming that relative humidity remains unchanged as the atmosphere warms.

A trend toward increasing column water vapor in the atmosphere consistent with model predictions has been documented from microwave satellite measurements (Trenberth, Fasullo, and Smith 2005), and excellent agreement for this quantity has been found between satellite observations and climate models constrained by the observed ocean-surface temperatures (Soden 2000). These studies increase confidence in the models' vapor distributions more generally, but column water vapor is dominated by changes in the lower troposphere, whereas water-vapor feedback is strongest in the upper troposphere where most outgoing terrestrial radiation to space originates. The results of Soden and Held (2000) imply that at least half the global water-vapor feedback arises from the tropical upper troposphere in models in which relative humidity changes are small. Studies of vapor trends in this region are therefore of central importance. Soden et al. (2005) present analysis of radiance measurements, implying that relative humidity has remained unchanged in the upper tropical troposphere over the past few years, which, combined with temperature measurements, provides evidence that water vapor in this region is increasing.

Observations of interannual variability in water vapor can help to judge the quality of model simulations. Soden et al. (2002) concluded that a GCM appropriately simulates water-vapor variations in the tropical upper troposphere during cooling associated with the Pinatubo volcanic eruption. Minschwaner, Essler, and Sawaengphokhai (2006) compared the interannual variability of humidity measured in the highest altitudes of the tropical troposphere with CMIP3 20th Century simulations. Both models and observations show a small negative correlation between relative humidity and tropical temperatures, due in large part to lower relative humidity in warm El Niño years and higher relative humidity in cold La Niña years. However, there is a suggestion that the magnitude of this covariation is underestimated in most models. There also is a tendency for models with larger interannual variations in relative humidity to produce larger reductions in this region in response to global warming, suggesting that this deficiency in interannual variability might be relevant for climate sensitivity. (This is another example, analogous to the Qu and Hall (2006) analysis of snow feedback, in which the strength of a feedback in models is correlated with a more readily observed aspect of climatic variability.) In short, the study of Minschwaner, Essler, and Sawaengphokhai (2006) suggests that water-vapor feedback in the very highest levels of the tropical troposphere may be overestimated in models, but it does not imply that a significant correction is needed to the overall magnitude of the feedback.

Positive water-vapor feedback, resulting from increases in vapor that keep the relative humidity from changing substantially as the climate warms, has been present in all GCMs since the first simulations of greenhouse gas–induced warming (Manabe and Wetherald 1975). It represents perhaps the single most robust aspect of global warming simulations. Despite the fact that the distribution of water vapor in the atmosphere is complex, we are aware of no observational or modeling evidence that casts doubt of any significance on this basic result, and we consider the increase in equilibrium sensitivity to roughly 2ºC from this feedback to be a solid starting point from which the more uncertain cloud feedbacks then operate.

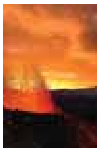
4.3 TWENTIETH CENTURY RADIATIVE FORCING

Radiative forcing is defined as a change that affects the Earth's radiation balance at the top of the tropopause between absorbed energy received in the form of solar energy and emitted infrared energy to space, typically expressed in terms of changes to the equilibrium preindustrial climate. Uncertainties in 20^{th} Century radiative forcing limit the precision with which climate sensitivity can be inferred from observed temperature changes. In this section, we briefly discuss the extent to which models provide consistent and reliable estimates of radiative forcing over the 20^{th} Century. Further information is provided by Forster et al. (2007).

Radiative forcing in models can be quantified in different ways, as outlined by Hansen et al. (2005). For example, the radiative forcing for the idealized case of CO_2 doubling can be computed by (1) holding all atmospheric and surface temperatures fixed, (2) allowing the stratospheric temperatures to adjust to the new CO_2 levels, (3) fixing surface temperatures over both land and ocean and allowing the atmosphere to equilibrate, or (4) fixing ocean temperatures only and allowing the atmosphere and land to equilibrate. Comparing model forcings in the literature is complex because of differing calculations in different papers. An important objective for the climate modeling community is to improve the consistency of its reporting of radiative forcing in models.

4.3.1 Greenhouse Gases

Greenhouse gases like carbon dioxide and methane have atmospheric lifetimes that are long, compared to the time required for these gases to be thoroughly mixed throughout the atmosphere. Trends in concentration, consistent throughout the world, have been measured routinely since the International Geophysical Year in 1958. Measurements of gas bubbles trapped in ice cores give the concentration prior to that date (with less time resolution). Nevertheless, the associated radiative forcing varies somewhat among climate models because GCM radiative calculations must be computationally efficient, necessitating approximations that make them less accurate than the best laboratory spectroscopic data and radiation algorithms. Using changes in well-mixed greenhouse gases measured between 1860 and 2000, Collins et al. (2006b) compared the radiative forcing of climate models (including CCSM, GFDL, and GISS) with line-by-line (LBL) calculations in which fewer approximations are made. The median LBL forcing at the top of the model by greenhouse gases is 2.1 W/m^2, and the corresponding median among the climate models is higher by only 0.1 W/m^2. However, the standard deviation among model estimates is 0.30 W/m^2 (compared to 0.13 for the LBL calculations). Based on these most-recent comparisons with LBL computations, we can reasonably assume that radiative forcing due to carbon dioxide doubling in individual climate models may be in error by roughly 10%.



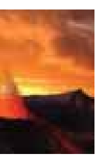
4.3.2 Other Forcings

While increases in the concentration of greenhouse gases provide the largest radiative forcing during the 20^{th} Century, other smaller forcings must be considered to quantitatively model the observed change in surface air temperature. The burning of fossil fuels that release greenhouse gases into the atmosphere also produces an increase in atmospheric aerosols (small liquid droplets or solid particles that are temporarily suspended in the atmosphere). Aerosols cool the planet by reflecting sunlight back to space. In addition, among other forcings are changes in land use that alter the reflectivity of the Earth's surface, as well as variations in sunlight impinging on the Earth.

4.3.2.1 Aerosols

Aerosols have short lifetimes (on the order of a week) that prevent them from dispersing uniformly throughout the atmosphere, in contrast to well-mixed greenhouse gases. Consequently, aerosol concentrations have large spatial variations that depend on the size and location of sources as well as changing weather that disperses and transports the aerosol particles. Satellites can provide the global spatial coverage needed to observe these variations, but satellite instruments cannot distinguish between natural and anthropogenic contributions to total aerosol forcing. The anthropogenic component can be estimated using physical models of aerosol creation and dispersal constrained by available observations.

Satellites increasingly are used to provide observational estimates of the "direct effect" of aerosols on the scattering and absorption of radiation. These estimates range from –0.35 +/– 0.25 W/m^2 (Chung et al. 2005) to –0.5 +/– 0.33 W/m^2 (Yu et al. 2006) to –0.8 +/– 0.1 W/m^2 (Bellouin et al. 2005). The fact that two of these three estimates do not overlap suggests incomplete uncertainty analysis in these studies. In particular, each calculation must decide how to extract the anthropogenic fraction of aerosol. Global direct forcing by aerosols is estimated by the IPCC AR4 as –0.2 +/– 0.2 W/m^2, according to models, and –0.5 +/– 0.4 W/m^2, based upon satellite estimates and models. This central estimate is smaller in magnitude than the 2001 IPCC estimate of –0.9 +/– 0.5 W/m^2.



In addition to their direct radiative forcing, aerosols also act as cloud condensation nuclei. Through this and other mechanisms, they alter the radiative forcing of clouds (Twomey 1977; Albrecht 1989; Ackerman et al. 2004). Complex interactions among aerosols and cloud physics make this "aerosol indirect effect" very difficult to measure, and model estimates of it vary widely. This effect was generally omitted from the IPCC AR4 models, although, among the U.S. CMIP3 models, it was included in GISS ModelE where increased cloud cover due to aerosols results in a 20th Century forcing of –0.8 W/m^2 (Hansen et al. 2007).

4.3.2.2 Variability of Solar Irradiance and Volcanic Aerosols

Other climate forcings include variability of solar irradiance and volcanic aerosols. Satellites provide the only direct measurements of these quantities at the top of the atmosphere. Satellite measurements of solar irradiance are available from the late 1970s and now span about 3 of the sun's 11-year magnetic or sunspot cycles. Extracting a long-term trend from this relatively brief record (Wilson et al. 2003) is difficult. Prior to the satellite era, solar variations are inferred using records of sunspot area and number and cosmic ray–generated isotopes in ice cores (Foukal et al. 2006), which are converted into irradiance variations using empirical relations The U.S. CMIP3 models all use the solar reconstruction by Lean, Beer, and Bradley (1995) with subsequent updates.

Volcanic aerosols prior to the satellite era are inferred from surface estimates of aerosol optical depth. The radiative calculation requires aerosol amount and particle size, which is inferred using empirical relationships with optical depth derived from recent eruptions. The GFDL and GISS models use updated versions of the Sato et al. (1993) eruption history, while the CCSM uses Ammann et al. (2003). As with solar variability, different reconstructions of volcanic forcing differ substantially (see, e.g., Lindzen and Giannitsis 1998). Land-use changes also are uncertain, and they can be of considerable significance locally. Global models, however, typically show very modest global responses, as discussed in Hegerl et al. (2007).

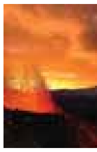
Studies attributing 20th Century global warming to various natural and human-induced forcing changes clearly are hindered by these uncertainties in radiative forcing, especially in the solar and aerosol components. The trend in total solar irradiance during the last few decades (averaging over the sun's 11-year cycle) apparently is negative and thus cannot explain recent global warming (Lockwood and Fröhlich 2007). The connection between solar energy output changes and the warming earlier in the 20th Century is more uncertain. With the solar reconstructions assumed in the CMIP3 models, much of the early 20th Century warming is driven by solar variations, but uncertainties in these reconstructions do not allow confident attribution statements concerning this early-century warming. The large uncertainties in aerosol forcing are a more important reason that the observed late 20th Century warming cannot be used to provide a sharp constraint on climate sensitivity. We do not have good estimates of the fraction of greenhouse gas forcing that has been offset by aerosols.

4.4 OCEAN HEAT UPTAKE AND CLIMATE SENSITIVITY

As noted above, the rate of heat uptake by the ocean is a primary factor determining transient climate response (TCR): the larger the heat uptake by the oceans, the smaller the initial response of Earth's surface temperature to radiative forcing (e.g., Sun and Hansen 2003). Studies show (e.g., Völker, Wallace, and Wolf-Gladrow 2002) that CO_2 uptake by the ocean also is linked to certain factors that control heat uptake, albeit not in a simple fashion. In an AOGCM, the ocean component's ability to take up heat depends on vertical mixing of heat and salt and how the model transports heat between low latitudes (where heat is taken up by the ocean) and high latitudes (where heat is given up by the ocean). The models make use of several subgrid-scale parameterizations (see Chapter 2), which have their own uncertainties. Thus, as part of understanding a model's climate-sensitivity value, we must assess its ability to represent the ocean's mixing processes and the transport of its heat, as well as feedbacks among the ocean, ice, and atmosphere.

The reasons for differing model estimates of ocean uptake are incompletely understood. Assessments typically compare runs of the same model or output from different AOGCMs. Raper, Gregory, and Stouffer (2002) examined climate sensitivity and ocean heat uptake in a suite of then-current AOGCMs. They calculated the ratio of the change in heat flux (from the surface to the deep ocean) to the change in temperature (Gregory and Mitchell 1997) and found in general that models with lower ocean-uptake efficiency had lower climate sensitivity, as expected (Hansen et al. 1985; Wigley and Schlesinger 1985). Uptake efficiency can be thought of as the amount of heat the ocean absorbs through mixing relative to the change in surface temperature (e.g., to reproduce the observed 20th Century warming despite a high climate sensitivity, a model needs large heat export to the deep ocean). Comparing the current generation of AOGCMs with the previous generation, however, Kiehl et al. (2006) found that the atmospheric component of the models is the primary reason for different transient climate sensitivities, and the ocean component's ability to uptake heat is of secondary importance. Ocean heat-uptake efficiency values calculated in this study differ substantially from those in Raper et al. (2002).

Despite these complexities, modern ocean GCMs are able to transport both heat (AchutaRao et al. 2006) and passive tracers such as chlorofluorocarbons and radiocarbon (Gent et al. 2006; Dutay et al. 2002) consistent with the limited observations available for these quantities. Better observations in the future—particularly of the enhanced ocean warming expected from the anthropogenic greenhouse effect—should provide stronger constraints on modeled ocean transports.



4.5 IMPACT OF CLIMATE SENSITIVITY ON USING MODEL PROJECTIONS OF FUTURE CLIMATES

This chapter has emphasized the global and annual mean of surface temperature change even though practical applications of climate change science involve particular seasons and locations. The underlying assumption is that local climate impacts scale with changes in global mean surface temperature (Santer et al. 1990). In that case, time histories of global mean temperature—obtained from a simple model of global mean temperature, run under a variety of forcing scenarios—could be combined with a single AOGCM-produced map of climate change normalized to the global mean surface temperature change. In that way, the regional changes expected for many different climate-forcing scenarios could be obtained from just one AOGCM simulation using one idealized forcing scenario such as atmospheric CO_2 doubling (Oglesby and Saltzman 1992) or 1% per year increasing CO_2 (Mitchell et al. 1999). This "pattern scaling" assumption also permits the gauging of effects on regional climate change that arise from different estimates of global climate sensitivity. For example, if an AOGCM with TCR = 1.5 K predicts temperature and precipitation changes ΔT and ΔP as a function of season and location in a 21st Century climate simulation, and if investigators believe that TCR = 1.0 K is a better estimate of the real world's climate sensitivity, then, under the pattern-scaling assumption, they would reduce the local ΔT and ΔP values by 50%.

Although it introduces its own uncertainties, the pattern-scaling assumption increasingly is used in climate impact assessments (e.g., Mitchell 2003; Ruosteenoja, Tuomenvirta, and Jylha 2007). For example, the annual mean temperature change averaged over the central United States during the 21st Century for any of the projections in the IPCC *Special Report on Emissions Scenarios* shows that about 75% of the variance among the CMIP3 models is explained by their differing global mean warming (B. Wyman, personal communication). (The central United States is defined in this context following Table 11.1 in Christensen et al. 2007.) Precipitation patterns, in contrast, do not scale as well as temperature patterns due to sharp variations between locally decreasing and locally increasing precipitation in conjunction with global warming.