Chapter 4

Climate Modeling in the Global Warming Debate

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I. INTRODUCTION

Akio Arakawa played a key role in the development of the Goddard Institute for Space Studies (GISS) global climate models (GCMs). Along with Jule Charney, Arakawa also motivated us to use those models to analyze climate sensitivity and processes involved in global warming. The current suite of GISS models, ranging from the global ocean to the Earth's mesosphere and Mars, continues to have dynamical cores that are fundamentally based on Arakawa's numerical methods.

We summarize the origins of climate modeling at GISS in the 1970s and later extension into a family of global models. Our first model application was to the fundamental question of how sensitive the Earth's climate is to external forcings, such as changes of atmospheric composition and solar irradiance. We also discuss climate predictions based on models driven by realistic transient climate forcings. The topical question of "missing atmospheric absorption" is considered in the penultimate section. Finally, we

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present a summary perspective of global warming issues. For the sake of informality, this chapter is written mainly in the first person by the first author, Jim Hansen.

II. GISS GLOBAL CLIMATE MODELS

A. Weather Model Prelude

When I came to GISS as a postdoctoral candidate in the late 1960s my primary interest was in planetary atmospheres, especially the clouds of Venus, and I focused on radiative transfer theory as a tool to study the Venus clouds. But at about that time the director of GISS, Robert Jastrow, concluded that the days of generous NASA support for planetary studies were numbered, and he thus began to direct institutional resources toward Earth applications.

The principal upshot was a concerted effort for GISS to get involved in testing the value of space observations for improving weather forecasts. Jule Charney of MIT, serving as a scientific consultant to GISS, provided the intellectual underpinnings, arguing that daily global measurements of atmospheric temperature profiles, if inserted continuously in a global weather prediction model, could sufficiently constrain the temperature, pressure, and wind fields in the model and hence lead to more accurate weather forecasts.

The first requirement for testing this hypothesis was a good weather prediction model, i.e., a computer program solving the fundamental equations for atmospheric structure and motion: the conservation equations for energy, mass, momentum and water substance, and the ideal gas law. That is where Akio Arakawa came in. Charney recommended that GISS import the UCLA two-layer atmospheric model of Yale Mintz and Arakawa and increase the model's vertical resolution, thus making full use of the temperature profiles measured by satellites and presumably increasing the model's forecast capability. Because Arakawa was the architect of the model, it was only through his enthusiastic cooperation that the model could be adapted for the GISS project. Milt Halem was the project director, Richard Somerville led the meteorological analysis of model capabilities, and Peter Stone was the principal consultant on atmospheric dynamics.

I had only a minor responsibility in the GISS modeling project, specifically to calculate the solar radiative heating, a term in the energy equation that is of little importance for weather forecasts. But this project, together with a Venus spacecraft project, provided resources that permitted hiring someone to work with me, and I used that opportunity to bring Andy Lacis, who was just completing his Ph.D. thesis in astrophysics at the University of Iowa, to GISS. Although our main interest was in planetary studies, our involvement with the weather model made it practical for us to initiate a climate modeling effort several years later.

Andy soon became the GISS expert in modeling of atmospheric radiation. We developed a method for calculating solar heating of the atmosphere (Lacis and Hansen, 1974) that used a crude eight-point k distribution to represent water vapor absorption over the entire spectrum. We also parameterized ozone absorption and cloud and molecular scattering, using analytic formulas fit to off-line radiative transfer calculations. This parameterization was cited by Paltridge and Platt (1976) as "a classic example of the derivation of a parameterization scheme whose validity has been tested by comparison with the results of complex but precise numerical solutions" (p. 91) and it was adopted in a number of GCMs and regional models. Although this parameterization of solar heating was sufficiently accurate for weather models, and was used in the GISS weather model (Somerville *et al.*, 1974), it did not include aerosols and was not designed for or ever used in any of our climate models. Decades later it became inadvertently involved in the current issue about "missing atmospheric absorption," but we argue in Section V that this missing absorption is primarily a misunderstanding.

Perhaps our main (inadvertent) contribution during the weather modeling era was to improve the lighting in the GISS building. Andy and I always worked until about 9 P.M., by which time everyone else had gone home. Just before leaving we would have a contest of hardball Frisbee standing at opposite ends of the hallway. The object was to throw the Frisbee so hard that the opponent would fail to catch it. We soon became sufficiently skilled that the only good way to induce a miss was via the sudden change of direction that accompanied a skip off a light fixture. Unfortunately, these plastic fixtures were not always as strong as the Frisbee and cracks occasionally appeared in a light cover. Fortunately, the fixtures were identical throughout the building and it was easy to interchange them. Within several years there was more light getting through the fixtures throughout the building, which was good because they were grimy and fuliginous. And, fortunately, by the 1990s when the building was renovated and the lights replaced, we had retired from hardball Frisbee.

B. Initial GISS Climate Model

Our interest in global climate was an outgrowth of radiation calculations. Following the approach of Suki Manabe (Manabe and Moller, 1961;

Manabe and Strickler, 1964), we used a one-dimensional (1-D) radiative-convective model to estimate the effect of various human-made greenhouse gases (GHGs) on global mean temperature (Wang *et al.*, 1976). This 1-D modeling allowed us to be involved in climate studies while we were seeking support for 3-D climate modeling. In addition to greenhouse calculations, we used the 1-D model to test the climate effect of volcanic aerosols, simulating a cooling after the largest volcanic eruption of the previous 50 years, Mt. Agung in 1963, in reasonable agreement with observations (Hansen *et al.*, 1978).

The problem with 1-D models was that climate feedbacks were specified, rather than computed from first principles, so climate sensitivity was essentially prescribed. Realistic study of climate problems required a 3-D global climate model (GCM), so that physical processes involved in climate feedbacks could be modeled more explicitly. The need was for a model that could be run on climatic time scales, and it seemed to me that it could define the main features of the atmospheric general circulation without having a resolution as fine as that in a weather model. Peter Stone, referring to a paper by Merilees (1975), argued that the important large-scale eddies could be represented with resolution as coarse as about 1000 km.

That is where Arakawa's model came in, in a crucial way. Other studies suggested that fine resolution (a few hundred kilometers or less) was required in global models, but those studies used unrealistic horizontal viscosity that tended to damp out not only the numerical instabilities at which it was aimed, but also real atmospheric motions when the resolution was coarse (Merilees, 1975). However Arakawa had designed the finite-differencing schemes in his model to conserve fundamental integral properties, thus permitting stable integration of the equations with little artificial diffusion or smoothing. And because the computing time varies roughly in proportion to the cube of the horizontal resolution, the long simulations needed for climate studies are much more feasible with coarse resolution.

I presented a proposal to NASA in 1975 to develop a climate model from the GISS weather model. Although this first proposal was not supported, Kiyoshi Kawabata, a Venusian scholar in our planetary group, volunteered to test Arakawa's model at coarse resolution, as a part-time activity. We were delighted to find that the simulated general circulation looked reasonably realistic at $8° \times 10°$ resolution, and it was qualitatively similar at $4° \times 5°$, $8° \times 10°$, and even $12° \times 15°$ resolutions. This meant that Arakawa's model could provide the dynamical core that we needed for an efficient climate model, although we would need to provide "physics" required for climatic time scales.

Our practical need was for someone with complete command of the model, including the finite-differencing methods and model programming. As fate would have it, in 1977 Milt Halem moved his weather modeling group to the parent Goddard Center in Greenbelt, Maryland. That provided the opportunity for us to acquire from Halem's group a brilliant young mathematician, Gary Russell, who had been the principal programmer for the GISS weather model. Gary not only had the confidence and ability to completely overhaul parts of the model when necessary, but also an insight about the physics that is crucial for model development.

The other key player soon added to our group was David Rind, coming from Bill Donn's group at Columbia's Lamont Observatory. His background in atmospheric dynamics, including the upper atmosphere, was an essential complement to the others, particularly since many climate change mechanisms involve the stratosphere. David developed a broad interest in climate modeling, including paleoclimate studies, thus also providing a working connection with paleoclimate researchers and to their invaluable perspective on climate change. For more than a decade David has been the most effective person at GISS in spurring model development and applications, and he has been our most active researcher in the crucial area of evaluating model performance relative to observations.

This internal GISS climate group (Fig. 1) has been guided by regular consultations with Peter Stone from the time of our first musings about developing a model. Although Peter is best known as an atmospheric dynamicist, he advises on the entirety of the model and is a collaborator on many of the model applications. The other main contributors to our early modeling, all coauthors on the paper describing our first model (Hansen *et al.*, 1983), were Reto Ruedy, Larry Travis, and Sergej Lebedeff.

Tony Del Genio arrived at GISS at about the time we finished that paper, and since then he has been responsible for clouds and moist convection, leading to some of the most significant model improvements. Other important model improvements came from Greg Hartke for the planetary boundary layer, Michael Prather for quadratic upstream differencing for atmospheric tracers, Cynthia Rosenzweig and Frank Abramopoulos for ground hydrology, and Elaine Matthews for global vegetation properties.

The gestation period for our first 3-D climate model paper, published in 1983, was more than 5 years. In addition to model development being laborious (we included 61 sensitivity experiments in our first paper) and our innate tendency to be deliberate, other factors contributed to this long gestation. First, we were pursuing multiple objectives. Although my aim was to study global change, e.g., the greenhouse effect, the GISS director asked us to focus on the "farmer's forecast." Thus, in addition to model

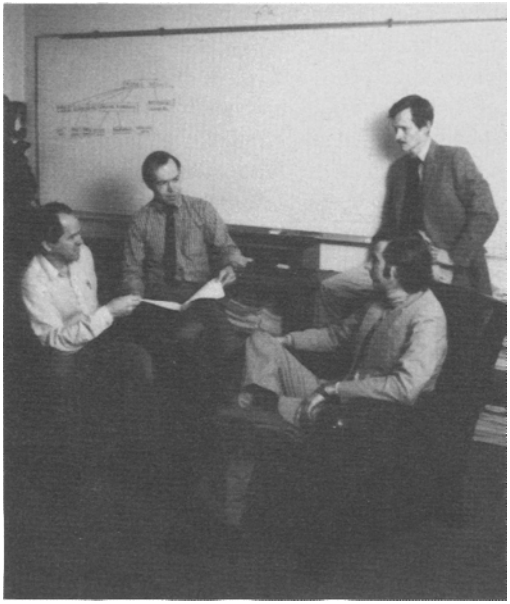

Figure 1 *Left to right*: A. Lacis, J. Hansen, D. Rind, and G. Russell in the early 1980s.

development, we carried out experiments to test the influence of sea surface temperature and initial land surface and atmospheric conditions on 30-day forecasts. Second, we worked on simpler models that provided guidance for more detailed study, as exemplified by our 1981 paper "Climate impact of increasing atmospheric CO_2" based on a 1-D model (Hansen *et al.*, 1981). Third, it took us a long time to convince referees that a coarse resolution model was a legitimate climate model.

This last factor warrants a comment here, and it is touched on implicitly under our "philosophy" below and in the concluding section. It is inappropriate to equate model validity with resolution, in our opinion. Resolution should relate to science objectives and the phenomena to be represented. Our aim is to employ a resolution sufficient to define the general circulation, including transports by large-scale atmospheric eddies, to allow simulation of seasonal climate on global and regional scales. Although a weather prediction model must attempt to resolve and follow midlatitude synoptic storms precisely, that is not necessarily required of a climate model. Model intercomparisons indicate that our coarse model does a good job of simulating seasonal variation of precipitation over the United States (Boyle, 1998), for example. Improvements obtained with finer resolution must be weighed carefully against improvements obtained with better physics and against the advantages of an efficient model.

C. Model Variations and Philosophy

The model that we documented in 1983, dubbed model II, was basically a tropospheric model. It was used for a number of climate studies in the 1980s, usually with a simple "Q-flux" treatment of the ocean, as described in Section III. The descendants of the original GISS climate model now form a family of models that can be used for more comprehensive investigations of climate change.

The most direct descendant of the original GISS model based on Arakawa's B Grid is the series of models SI95, SI97, SI99, which have been used and tested by students and faculty in the GISS Summer Institute on Climate and Planets (Hansen *et al.*, 1997c). These models, so far, have been run at $4° \times 5°$ resolution. Changes of model physics subsequent to model II include the moist convection parameterization (Del Genio and Yao, 1993), prognostic clouds (Del Genio *et al.*, 1996), the planetary boundary layer representation (Hartke and Rind, 1997), ground hydrology and evapotranspiration (Rosenzweig and Abramopoulos, 1997), numerical differencing schemes, including use of a quadratic upstream scheme (Prather, 1986) for heat and moisture, and various minor factors (Hansen *et al.*, 1997c). The SI95 model had the same 9 layers as model II, while the SI97 and SI99 models have 12 layers with 3 or 4 of these in the stratosphere. Current development gives priority to improved vertical resolution and better representation of physical processes.

The first major extension of the GISS model was to the stratosphere and mesosphere, with the development of the GISS global climate/middle atmosphere model (Rind *et al.*, 1988). That model is used with different choices for vertical resolution and model top, as high as about 80 km, and with increasingly sophisticated treatments of gravity wave drag. Recent applications of that model to solar cycle and ozone climate forcings (Shindell *et al.*, 1999a,b), including successful simulation of observed solar cycle changes, provide an incentive for improving the vertical resolution in other model versions. Inclusion of this model in the GISS stable allows testing of the model resolution and vertical extent required to simulate different climate phenomena.

Another variation of the GISS model is Gary Russell's coupled atmosphere–ocean model (Russell *et al.*, 1995). Both atmosphere and ocean use Arakawa's C Grid with the linear upstream method of Russell and Lerner (1981) for heat and water vapor. In addition Gary modified and simplified physics parameterizations, including replacement of the surface/boundary layer formulation with an extrapolation from the lowest model layer and replacement of the Del Genio *et al.* prognostic clouds with a simpler scheme having cloud optical thickness proportional to the square root of water vapor amount. The resulting model is faster and has an improved climatology for several climate diagnostics including sea level pressure distribution. A criticism that has been made is that the model yields an increasing cloud optical thickness with increasing temperature, contrary to observations at most places in the world (Tselioudis and Rossow, 1994; Del Genio and Wolf, 2000). But the model's efficiency has allowed it to be used for many climate studies and comparison of its results with other models has been valuable for model development and analysis of climate experiments. Also Russell's ocean model has been coupled with the B Grid atmosphere model, providing a useful comparison with the community ocean models used in most climate studies.

Still another variation is the Wonderland model (Hansen *et al.*, 1997b). This uses the physics of the 1983 model with $8° \times 10°$ resolution and an idealized cyclic geography, which makes the model fast enough for numerous century and millennium time scale simulations. The Wonderland model has been used for systematic analysis of the climate response to a wide range of radiative forcings (Hansen *et al.*, 1997c), and it has potential for paleoclimate studies. The Wonderland model has been temporarily abandoned because of its outdated physics, but, once we have model physics that we are satisfied with, we intend to revive it with the updated physical parameterizations.

Finally, I offer a few comments on our modeling philosophy. Our emphasis is on improved representation of the "physical" (including biological) processes. In our opinion, inadequate treatment of the physics is the primary restraint on understanding of long-term climate change. But better physics includes a need for higher vertical resolution in the atmosphere, where our present focus is on the planetary boundary layer and the upper atmosphere. Also Gary Russell emphasizes the need to handle nonlinear advection (the momentum equation) more accurately, which may require fundamental changes in the differencing schemes. Horizontal resolution in the atmosphere warrants continued examination, i.e., experimentation with finer grids. But, as we discussed in our 1983 paper, increased horizontal resolution is very expensive in resource requirements and relatively ineffective; when it is overemphasized, it limits the ability to attack fundamental issues. In comparison, there is a better justified need for improved resolution in ocean models. Along with the need for better physics in the atmosphere, this provides a primary drive for improved computer power.

A corollary of emphasis on model physics is the need to involve the research community in our model development and applications. GISS researchers can cover only a few topics in depth. But, if we can demonstrate that our model simulates characteristics of decadal climate change realistically and that it can help investigate the causes of long-term climate change, that should promote collaborations and interactions with leading researchers, and that in turn may provide a positive feedback advancing modeling capabilities.

Modeling philosophy must also relate to computing technology. It is commonly assumed that the fastest supercomputer is most productive for climate modeling. But the speed of a single run is only one consideration. Other factors include cost, the fraction of time available on the computer, the need for special programming, and especially how the computing approach meshes with the research objectives. We were among the first to emphasize the potential of workstations; for example, the ensembles of runs with the SI95 model (Hansen *et al.*, 1997c) were carried out on individual workstations. Now we have a 64-processor cluster that is well suited for ensembles of runs, but also, using a fraction of the processors in parallel, it permits use of a 32-layer $2° \times 2.5°$ model.

Ongoing technological advances in computing, data storage, and communications capabilities open new possibilities to advance modeling capabilities and understanding of long-term climate change. These advances will make it possible not only to include more realistic physics and higher model resolutions, but to systematically carry out ensembles of simulations and make the results readily available to the research community. This is an approach that we will pursue vigorously.

III. CLIMATE SENSITIVITY

A. Charney Report

In 1979 the president's science advisor requested the National Academy of Science to study the carbon dioxide and climate issue. This resulted in the famous Charney (1979) report from a group of climate researchers, including Akio Arakawa, who met at Woods Hole in the summer of 1979.

Jule Charney, the panel chairman, decided to focus on a well-defined question: If the amount of atmospheric CO_2 were doubled, how much would the global average temperature increase by the time the system came to a new equilibrium? This question allowed use of the doubled CO_2 GCM studies of Suki Manabe that were already published (Manabe and

Wetherald, 1975) and in preparation (Manabe and Stouffer, 1980). The Charney panel also employed other tools, especially 1-D climate models, to analyze the topic.

Charney and Arakawa were interested personally in 3-D global models, which provided us opportunities for interactions with them. After Charney learned that we had initiated a doubled CO_2 experiment, we had several discussions with him and he asked Arakawa to visit GISS and work with us for a week. It was a good opportunity for us to talk with Akio not only about the doubled CO_2 results, but also about climate model development in general.

Our model result differed from the most recent model of Manabe, ours yielding a global warming of almost 4°C, while Manabe and Stouffer obtained 2°C. The conclusion that we reached with Arakawa, under the assumption that both models calculated the radiation accurately, was that differences between the models probably were caused by different strengths of climate feedback processes, especially sea ice and clouds. Specifically, there was relatively little Southern Hemisphere sea ice in the control run of Manabe and Stouffer, which would limit that positive feedback. Also their model used fixed clouds, while our model calculated reduced cloud cover with global warming, thus yielding more positive feedback.

Based on these model studies and their other deliberations, the Charney report estimated that equilibrium global climate sensitivity to doubled CO_2 was 3 ± 1.5°C. The range 1.5 to 4.5°C was broad and the stated uncertainty range was not meant to exclude the possibility of a sensitivity outside that range. Perhaps the best summary of the Charney report was their statement: "To summarize, we have tried but have been unable to find any overlooked or underestimated physical effects that could reduce the currently estimated global warming due to doubling of atmospheric CO_2 to negligible proportions" (p. 3).

The interactions with Charney and Arakawa stimulated us to analyze the contributions from each of the radiative feedbacks in our climate sensitivity experiments by inserting the changes (of sea ice, clouds, and water vapor) found in the GCM into a 1-D radiative model. This feedback analysis, developed by Andy Lacis, was used to help interpret our first published doubled CO_2 experiment (Hansen *et al.*, 1984). The separation of the climate response into that which would occur without feedbacks, ΔT_O, plus feedback contributions is the fundamental distinction between radiative forcing and climate response. ΔT_O measures the forcing in °C; the proportionality factor needed to convert this to a forcing in W/m^2 is 3.33. Thus the forcing for doubled CO_2 is $\Delta T_O \sim 1.25$°C or $\Delta F \sim 4.2\ W/m^2$.

B. Ice Age

Climate models by themselves can never yield an accurate and convincing knowledge of climate sensitivity. It is possible to change model parameters, e.g., in the cloud representation, that greatly alter the model sensitivity. And one can always think of climate feedbacks that may exist in the real world, but are entirely unrepresented in the model.

A more accurate measure of climate sensitivity can be obtained from analysis of empirical data with the help of climate models. Probably the best measure of climate sensitivity that we have now is that inferred from the last ice age, about 20,000 years ago. We now have a rather good knowledge of both the climate change between the last ice age and the current interglacial period as well as the change in the climate forcing that maintained the changed climate.

The important point is that, averaged over, say, 1000 years, the Earth had to be in near radiation balance with space during the middle of the last glacial period as well as during the current interglacial period. An imbalance of even 1 W/m^2 would have caused a rate of ocean temperature change or a change in the mass of glacial ice much greater than actually occurred.

The composition of the Ice Age atmosphere has been measured well from samples of air trapped in the polar ice sheets at the time of their formation (e.g., Lorius *et al.*, 1990). Planetary surface conditions, including the distribution of ice sheets, shorelines, vegetation, and surface albedo, have also been reconstructed (CLIMAP, 1981). The resulting radiative forcings that maintained the Ice Age cold were increased reflection of sunlight by the Earth's surface due mainly to larger ice sheets and altered vegetation distributions, decreased amounts of GHGs, and increased atmospheric aerosol loading (Hansen *et al.*, 1984, 1993; Hoffert and Covey, 1992). These surface and atmospheric changes caused a total forcing of -6.6 ± 1.5 W/m^2 (Fig. 2).

This forcing maintained a global mean temperature change of about 5°C. CLIMAP (1981) reconstructions of ocean temperature, which had the last Ice Age being warmer than at present in much of the tropics, implied a global cooling of about 3.7°C during the last Ice Age. But recent data indicate that the tropics did cool by at least a few degrees (e.g., Guilderson *et al.*, 1994; Schrag *et al.*, 1996), so that a better estimate of the global mean Ice Age cooling is 5 ± 1°C.

Thus the climate sensitivity implied by the last Ice Age is about $5°C/(6.6\ W/m^2) = 0.75°C$ per W/m^2, equivalent to 3 ± 1°C for doubled CO_2, in remarkable agreement with the analysis of Charney and Arakawa.

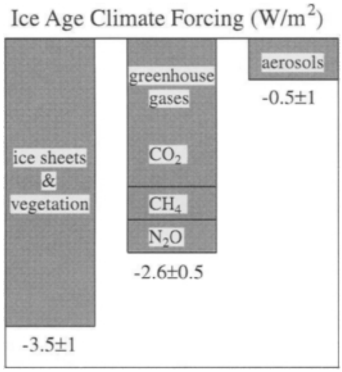

Figure 2 Climate forcings during the Ice Age 20,000 years ago relative to the current interglacial period. This forcing of -6.6 ± 1.5 W/m^2 and the 5°C cooling of the Ice Age imply a climate sensitivity of 0.75°C per 1 W/m^2.

The great thing about this empirical derivation is that it includes all climate feedbacks; any feedback that exists in the real world, whether we have thought about it yet or not, is incorporated, and that includes any changes of ocean heat transports.

A concern that can be raised about this empirical sensitivity is that climate sensitivity depends on the mean climate state. Variations of past climate and climate models both suggest that climate sensitivity is greater for a colder climate than for a warmer climate, and thus climate sensitivity inferred from comparison with the last Ice Age may not be accurate for the present climate. But, for several reasons, this concern is less substantial than it may appear. First, much of the higher sensitivity toward a colder climate is a consequence of increasing land ice cover with colder climate, and this factor is taken out in our present evaluation that treats land ice changes as a forcing, i.e., the inferred sensitivity refers only to the "fast" feedbacks, such as water vapor, clouds, and sea ice (Hansen *et al.*, 1984). Second, although the sea ice feedback is expected to increase toward colder climates, the nonlinearity should be moderate for small changes of the mean climate. Third, the sensitivity 0.75°C per W/m^2 if calculated to two decimals yields 3.2°C for our current estimate of doubled CO_2 forcing (Hansen *et al.*, 1998b) with this result representing the mean sensitivity between the last Ice Age and today. We conclude that 3 ± 1°C for doubled CO_2 is the appropriate estimate of climate sensitivity for today's global temperature.

IV. TRANSIENT CLIMATE: CLIMATE PREDICTIONS

A. Climate Response Time: Simple Ocean Models

The Charney report discussed only briefly the issue of how long it takes the climate system to more or less fully respond to a climate forcing. Charney realized that it was necessary to account for the ocean heat capacity beneath the mixed layer, and I recall him suggesting that the response time to increased CO_2 could be a few decades, on the basis of overturning times for near surface ocean layers in the tropics and subtropics. What was not realized at that time was that the climate response time is a function not only of the ocean's overturning rate, but of climate sensitivity itself. In fact, it is a very strong function of climate sensitivity. This issue does not alter Charney's analysis, because he focused on the equilibrium response to doubled CO_2. But climate sensitivity and response time become intimately connected if one attempts to infer climate sensitivity from observed transient climate change, and the climate response time raises a severe problem for policy makers.

I became especially interested in climate response time with the publication of the Carbon Dioxide Assessment Committee report (CDAC, 1983). This report seemed to be aimed at damping concern about anthropogenic climate change; at any rate, that was a likely effect of their conclusion that climate sensitivity was probably near the lower end of the range that Charney had estimated (1.5°C for doubled CO_2). But their conclusion was based on the magnitude of observed global warming in the past century and the assumption that most of the warming due to human-made GHGs should already be present. Specifically, their analysis assumed that the climate response time could be approximated as being 15 years and that the response time was independent of climate sensitivity.

The fact that climate response time is a strong function of climate sensitivity is apparent from the following considerations. First, note that climate feedbacks, such as melting sea ice or increasing atmospheric water vapor, come into play only in conjunction with temperature change, not in conjunction with the climate forcing. Thus, even if the ocean's heat capacity could be represented as that of a simple slab mixed layer ocean, the response time would increase in proportion to the feedbacks (and thus in proportion to climate sensitivity). And, second, while the feedbacks are coming into play, the heat perturbation in the ocean mixed layer can mix into the deeper ocean, further delaying the surface response to the forcing.

Investigation of this issue requires a realistic estimate of the rate of heat exchange between the ocean surface (well-mixed) layer and the deeper ocean. Our approach to this problem in the early 1980s was to attach a simple representation of the ocean to our atmospheric GCM. We used this ocean representation for our transient climate predictions, described in the next section, as well as for investigation of climate response time. The objectives of the ocean representation were (1) to obtain a realistic climate response time at the Earth's surface and (2) to achieve a realistic distribution of surface climate in the model's control run despite the absence of a dynamic simulation of the ocean.

One part of the ocean representation was vertical exchange of heat anomalies beneath the ocean mixed layer. For our 1-D radiation model we had used a vertical diffusion coefficient based on observed global penetration of transient tracers. For the 3-D model Inez Fung determined local diffusion coefficients by using transient ocean tracer observations to establish a relationship between the vertical mixing rate and the local stability at the base of the winter mixed layer. This relationship and the Levitus ocean climatology were then used to obtain effective mixing coefficients beneath the mixed layer for the entire ocean, as described in our Ewing symposium paper (Hansen *et al.*, 1984).

The second part of the ocean representation was a specification of horizontal heat transports in the ocean suggested by Peter Stone and developed by Gary Russell, as described briefly in our Ewing paper and in more detail by Russell *et al.* (1985). Specifically, we employed the ocean heat transports implied by the energy balance at the ocean surface in our GCM when the model was driven by observed sea surface temperatures. This approach of specifying the horizontal ocean heat transport has come to be known as the Q-flux ocean model and is used with the mixed layer model alone as well as with the mixed layer attached to a diffusive ocean.

The upshot of our climate simulations was that climate response time is a strong function of climate sensitivity. The response time is only about 15 years if climate sensitivity is near the lower limit estimated by Charney (1.5°C for doubled CO_2), but more than 100 years if climate sensitivity is 4.5°C for doubled CO_2. The climate sensitivity inferred from paleoclimate data, about 3°C for doubled CO_2, suggests that the climate response time is at least 50 years.

Such a long response time raises a severe conundrum for policy makers. If, as seems likely, GHGs are the dominant climate forcing on decadal time scales, there may be substantial warming "in the pipeline" that will occur in future decades even if GHGs stop increasing. Such yet to be realized warming calls into question a policy of "wait and see" for dealing with the uncertainties in climate prediction. The difficulty of halting climate change once it is well under way argues for commonsense measures that slow down the climate experiment while a better understanding is developed.

B. Global Climate Predictions

We used the model described above, with Q-flux horizontal ocean transports and spatially variable diffusive mixing of temperature anomalies beneath the mixed layer, for the first transient climate predictions with a 3-D global climate model (Hansen *et al.*, 1988). Climate change in this model was driven by observed and projected GHG changes and secondarily by aerosols from volcanic eruptions.

Figure 3 compares observed global surface temperature with the simulations, which were carried out in 1987. The large interannual variability makes it difficult to draw inferences based on only 11 years of data subsequent to the calculations. But so far the world has been warming at a rate that falls within the range of scenarios considered.

Scenarios A, B, and C differed in their growth rates of GHGs and in the presence or absence of large volcanic eruptions. Scenario A assumed that GHGs would increase exponentially at rates characteristic of the preceding 25 years and that there would be no large volcanic eruptions. Scenario

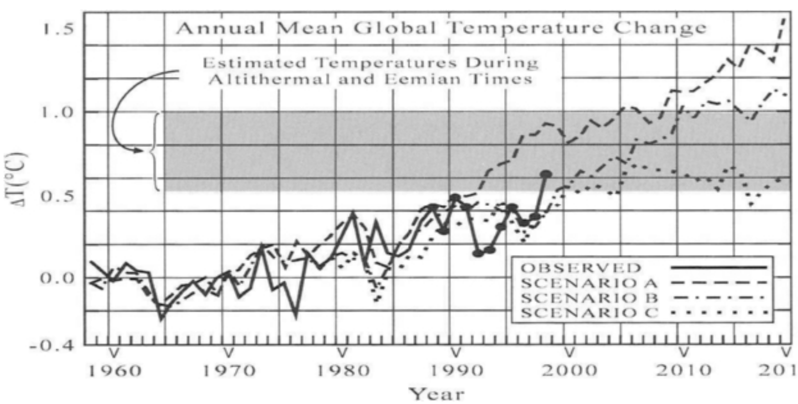

Figure 3 Global surface air temperature computed with GISS model in 1987 (Hansen *et al.*, 1988) and observed global temperature based on meteorological station measurements (Hansen *et al.*, 1999), including update subsequent to model predictions (· · · · ·).

A was designed to reach the equivalent of doubled CO_2 by about 2030, consistent with the estimate of Ramanathan *et al.* (1985). Scenario B had an approximately linear growth of GHGs, reaching the equivalent of doubled CO_2 at about 2060. Scenario B included occasional cooling from volcanic eruptions in 1995 and 2015. Scenario C had a still slower growth rate of GHGs with a stabilization of GHG abundances after 2000 and the same volcanos as in scenario B.

What is the climate forcing in the real world? Both GHGs and volcanic aerosols have been well measured in recent decades. The observed GHG changes and volcanic aerosols both correspond closely to scenarios B and C (Hansen *et al.*, 1998a,b), which are practically the same until year 2000. The main difference is that the large volcano in the 1990s occurred in 1991 in the real world, while in the model it occurred in 1995. Scenario C, with terminating GHG growth in 2000, is not expected to be realistic in the future. Thus scenario B is the most realistic.

The global temperature in scenario B increases by 1°C in 50 years (Fig. 3), with a rather steady warming rate of about 0.2°C/decade. This is in good agreement with observations of the past few decades, as described in detail by Hansen *et al.* (1999). But the absence of information on all climate forcings makes it difficult to draw substantive conclusions even from the 40-year record.

One important conclusion that can be drawn is that the rate of growth of GHGs in the real world is significantly less than in scenario A, the "business as usual" scenario with continued exponential growth of GHGs that is similar to the principal IPCC (1996) scenarios. The climate forcing due to observed growth rates of GHGs during the past several years is only about half of that in the scenarios commonly used by IPCC, such as IS92a or 1% CO_2 increase per year (Hansen *et al.*, 1998b). The slowdown in growth rates provides hope that the more drastic climate changes can be avoided.

Clarification of GHG scenarios is important for the global warming debate (Section VI) and for interpretation of present and future observed climate change. Although IPCC defines a broad range of scenarios, the full range is not emphasized. It is a common practice of modelers to employ a single scenario with a strong GHG growth rate. A strong forcing has the merit of yielding a large "signal-to-noise" ratio in the climate response. But use of a single scenario can be taken as a prediction in itself, even if that is not intended. Multiple scenarios are especially useful for problems that may involve nonlinear processes in a significant way. Thus we argue (Hansen *et al.*, 1998b) for use of a range of scenarios bracketing plausible rates of change, which was the intention of our scenarios A, B, and C.

C. Forcings and Chaos

We present an example of calculations with the current GISS GCM to bring the modeling discussion up to date. Specifically, we use the model version based on Arakawa's B Grid atmosphere that is employed by the Forcings and Chaos research team in the GISS Institute on Climate and Planets. Examples of recent results from the other principal variations of the GISS GCM are given by Shindell *et al.* (1999b) for simulated climate effects of solar cycle and ozone variability using the GISS climate/middle atmosphere model and by Russell *et al.* (2000) for simulated climate trends due to increasing CO_2 using the C Grid coupled atmosphere–ocean version of the GISS model.

The objective of the Forcings and Chaos group is to shed light on the roles of climate forcings and unforced climate variability ("chaos") in climate variability and change during recent decades. The approach is to make ensembles of simulations, adding various radiative forcings to the model one by one, and running the model with several different treatments of the ocean (Hansen *et al.*, 1997c). Initial simulations were made for the period 1979–1996 with the SI95 model, which was frozen during the Summer Institute of 1995. Trial simulations for the period 1951–1997 were made with the SI97 and SI99 models, and a larger array of simulations for 1951–1999 is planned for the SI00 model.

1. SI95 Simulations

The SI95 model, documented by Hansen *et al.* (1997c), had nine layers in the atmosphere with one or two layers in the stratosphere. This model was run with four representations of the ocean: (A) observed SST, (B) Q-flux ocean, (C) GISS ocean model (Russell *et al.*, 1995), and (D) an early GFDL ocean model (Bryan and Cox, 1972; Cox, 1984). The SI95 model was flawed by excessive absorption of solar radiation by sea ice, as illustrated by Fig. 1 of Hansen *et al.* (1997c). It was realized later that the excessive absorption was the result of a programming error that caused sea ice puddling to be active independent of surface temperature.

The SI95 simulations illustrated that most of the interannual variability of *regional* climate on an 18-year time scale at middle and high latitudes is chaotic, i.e., unforced. But a natural radiative forcing (volcanic aerosols) and an anthropogenic forcing (ozone depletion) were found to leave clear signatures in the simulated *global* climate that were identified in observations. The SI95 simulations were also used to infer a planetary radiation imbalance of about 0.5 W/m^2, leading to prediction of a new record global temperature that has subsequently occurred.

2. SI97 Simulations

Significant modifications in the SI97 model include the use of 12 atmospheric layers, changes to the planetary boundary layer (Hartke and Rind, 1997) and the clouds and moist convection (Del Genio *et al.*, 1996), correction of the programming error in the sea ice puddling, and addition of a parameterization for ice cover of lakes. The three additional layers increase the resolution in the tropopause and lower stratosphere region with the model top remaining at 10 mb. These modifications will be described in a future paper documenting the SI99 model and, in some cases, in future papers defining specific aspects of the model physics.

Improvements in the SI97 climatology over the SI95 model include (1) more realistic stratospheric temperatures, especially the longitudinal variations, although the stratosphere remains too warm at the winter pole and too cool at the summer pole; (2) more realistic poleward heat transports; (3) more accurate computations of stratospheric radiative forcings, especially due to stratospheric aerosol and ozone changes, resulting in accurate representation of stratospheric temperature change after large volcanos; (4) more accurate albedos for sea ice, improving the sea ice cover in coupled atmosphere ocean runs; and (5) more accurate winter temperatures in Canada.

Known outstanding problems with the SI97 model include (1) deficiencies in boundary layer stratus cloud cover off the west coast of the continents, resulting in a solar radiation flux at the ocean surface that is excessive by as much as 50 W/m^2 in the summer; (2) buildup of snow cover along the northeast coast of Siberia that fails to melt in the summer, a problem that was exacerbated by improved physical representations of the PBL and clouds; and (3) a still very crude representation of the stratosphere, including the rigid top at 10 mb and a sponge-layer drag in the top layer, resulting in errors in the stratospheric temperature distribution and circulation.

We carried out several simulations for the period 1951–1997 with the SI97 model that helped assess the model capabilities and deficiencies. Figure 4 (see color insert) shows the degree to which the SI97 model simulates observed surface temperature change during that 47-year period. Observed change of the surface temperature index, which consists of surface air temperature over land and SST over the ocean, is shown in Fig. 4b. The left column, Figs 4a, 4c, and 4e, shows climate model simulations of surface air temperature change driven only by observed changes of SST and sea ice, with the three cases providing an indication of the impact of uncertainties in these boundary conditions. Figures 4d and 4f add the two most accurately known radiative forcings, greenhouse gases (Hansen *et al.*, 1998b) and stratospheric aerosols (Sato *et al.*, 1993).

Two features in the observed climate change are of special interest: (1) high latitude warming over Siberia and the Alaska region, which is strongest in the winter, and (2) cooling over the contiguous United States, which is strongest in the summer. We discuss each of these briefly.

a. High-Latitude Warming

The model simulates the Alaska warming, but it does not simulate the Siberia warming well. The results may improve with the SI99 model, which eliminates the problem of growing glaciers in northeast Siberia, but that seems unlikely to be important in the winter. Additional climate forcings, including ozone, solar irradiance, and aerosol direct and indirect effects may be important. But it is likely that simulation of the winter warming in Siberia will require a better representation of the stratosphere. Shindell *et al.* (1999a) find that greenhouse gas forcing yields greater Siberian warming in the GISS climate/middle atmosphere model, associated with an intensification of the stratospheric polar vortex. This topic requires further study as the climate/middle atmosphere model has a sensitivity of 5.5°C for doubled CO_2, which may be larger than reality, and the climate forcing used by Shindell *et al.* (1999a) is similar to IPCC IS92a, which exceeds the observed greenhouse gas forcing. The Siberian warming is a part of the Arctic oscillation (Thompson and Wallace, 1998) that seems to be a natural mode not only of the real world but of climate models. Thus the stronger response in the experiment by Shindell *et al.* (1999a) might be in part a consequence of the bell being rung harder in that model. But the important point is the evidence that adequate representation of stratospheric dynamics is needed for simulating tropospheric climate.

This is an important practical matter for climate model development because the higher model top (80 km) and sophisticated gravity wave drag treatment in the climate/middle atmosphere model increase the computation time by a factor of 7. The plans for the SI model series, which is aimed at studies of surface climate, were to make moderate improvements in the representation of the stratosphere, perhaps increasing the model top to 50 km and including a simple representation of gravity wave effects. But if the suggestion of Shindell *et al.* (1999a), that even the mesosphere must be included to simulate the effects of solar variability on surface climate, is borne out, we will need to reconsider this strategy for model development.

b. United States Cooling

It is interesting that the GISS model driven by observed SST anomalies consistently simulates a cooling trend in the United States during the past 50 years. This cooling trend is not an accident, because it is captured by all of the five ensembles of SI97 model runs. All five ensembles yield greater cooling in the summer than in the winter, in agreement with observations. This suggests that the observed regional climate trend is a tropospheric phenomenon driven immediately by SST anomalies, and that the model can represent, at least in part, the immediate mechanisms for change. Although it will be a challenge to determine whether the SST anomalies are themselves forced or chaotic, it may be easier to make progress in partial understanding of this climate change by making simulations in which the SST anomalies are restricted to specific parts of the ocean. However, because of inherent limitations in the ability of specified SST experiments to deliver correct atmosphere to ocean flux changes, it will be necessary to also carry out experiments with other ocean representations that more realistically portray ocean-atmosphere interactions.

We point out elsewhere (Hansen *et al.*, 1999) the practical importance of understanding this climate change in the United States. During the past century, temperatures have increased slightly in the United States, but not as much as in most of the world, and the warmest temperatures in the United States occurred in the 1930s (Fig. 8 of Hansen *et al.*, 1999). Although long-term climate change in recent years seems to be reaching a level that is noticeable to the layperson in some parts of the world (Hansen *et al.*, 1998a), this is less so in the contiguous United States. However, if the SST patterns that are giving rise to the recent cooling tendency in the United States are a temporary phenomenon, there could be a relatively rapid change to noticeably warmer temperatures in the near future.

3. SI99 Simulations

The SI99 model was recently frozen to allow an array of simulations for 1951–1999 to be carried out. Principal changes in the SI99 model are (1) modification of the snow albedo parameterization to eliminate the growth of glaciers in northeast Siberia, (2) replacement of the tropospheric aerosol distribution of SI95 and SI97 with a new distribution based mainly on assumed aerosol sources and tracer transport modeling by Ina Tegen and Dorothy Koch, and (3) optional replacement of the fourth-order differencing scheme for the momentum equation with second-order differencing. The new aerosol distribution reduces solar heating of the surface by several watts per square meter, as shown in Section V. The second-order differencing eliminates excessive noise and model instability caused by the fourth-order scheme while reducing the computing time by about 25%. However, midlatitude storms move more slowly and do not cross the continents as realistically, so the fourth-order differencing is retained in the model coding and employed in many experiments.

The SI99 model will be documented in conjunction with a paper describing the array of simulations for 1951–1999. These experiments will differ from the array described by Hansen *et al.* (1997c) in several ways: (1) The period of simulation will be about five decades rather than two decades; (2) forcings each will be run individually rather than cumulatively, but some experiments will also include all or most of the forcings; (3) tropospheric aerosols will be included as a forcing; (4) dynamic ocean models are expected to include the GISS model, an up-to-date version of the GFDL MOM model, and the global isopycnal (MICOM) ocean model of Shan Sun and Rainer Bleck; and (5) access to model results will be provided via the GISS World Wide Web home page (www.giss.nasa.gov).

V. MISSING ATMOSPHERIC ABSORPTION

A prominent issue concerning climate models in the 1990s has been "missing atmospheric absorption." Surface, satellite, and *in situ* observations have been used to surmise that most climate models underestimate solar radiation absorbed in the atmosphere by 20–40 W/m^2 and overesti mate solar radiation absorbed at the planetary surface by a similar amount. Such errors could affect the simulated atmospheric circulation and the drive for oceanic temperatures and motions.

Comprehensive review of this topic is beyond the scope of our paper. We refer instead to a few recent papers, which lead to many others. John Garratt and colleagues (1998) and Bob Cess and colleagues (1999) have been especially productive in providing observational data and interpretations in a series of papers going back at least to 1993. These scientists and others (cf. semipopular review by Kerr, 1995) deserve credit for stimulating discussions about atmospheric physics and verification of models, in the best spirit of scientific investigation.

The focus has been on identifying missing or underrepresented absorbers in the models. Arking (1996) argues that water vapor absorption is underestimated. Garrett *et al.* (1998) suggest that inaccurate water vapor calculations and aerosols contribute to the problem. Cess *et al.* (1999), however, present data that they interpret as indicating that the missing absorber is present only in cloudy skies, not clear skies. There has been much speculation about possible exotic mechanisms for absorption, such as water vapor dimers, that are not included in present models.

Not long ago Bob Cess presented a seminar at GISS summarizing evidence that he interpreted as requiring the presence of a missing absorber. He commented that Paul Crutzen not only agreed with this conclusion but stated that it was time to stop arguing about it. Although Bob took some solace in the support of a Nobel prize winner, somehow the thought that jumped to my mind on hearing this was one of Oscar Wilde's epigrams: "When people agree with me, I always feel that I must be wrong."

Observationally it is difficult, if not impossible, to obtain a clean separation of clear and cloudy skies, especially with satellite observations. For this reason, and because it is the total absorption that drives the atmosphere and ocean, it seems best to examine first the all-sky case. Martin Wild has presented extensive comparisons of modeled and "observed" solar radiation absorption (see Wild *et al.*, 1998, and references therein) that we will use for quantitative discussion.

We focus on three numbers: (1) the amount of solar radiation hitting the Earth's surface, $S\downarrow$, (2) the amount of solar radiation absorbed by the Earth's surface, $\alpha \times S\downarrow$, where α is the surface co-albedo, i.e., 1 minus the albedo), and (3) the amount of solar radiation absorbed by the atmosphere (A_{atm}). The debate in the literature has focused on atmospheric absorption, but we argue that A_{atm} is a tertiary quantity and is not observed. Thus it is better to consider the three quantities in the order listed here.

The solar radiation hitting the Earth's surface, $S\downarrow$, is a primary quantity, i.e., it can be measured and, indeed, has been measured at hundreds of stations around the world. The solar radiation absorbed by the Earth's surface, $\alpha \times S\downarrow$, is a secondary quantity. It cannot practically be measured with the needed accuracy, because it varies on small spatial scales. One must assume a global distribution of surface albedos, so $\alpha \times S\downarrow$ includes the uncertainties in both $S\downarrow$ and α. Similarly, the absorption in the atmosphere, A_{atm}, is a tertiary quantity and cannot be measured directly on a global scale, and its calculation requires additional input. That input can be an assumed (or measured) planetary albedo, which is often taken as 30%, or detailed information on clouds and other atmospheric properties required for radiative transfer calculations across the solar spectrum.

The GEBA (Global Energy Balance Archive) data for $S\downarrow$ are shown in Fig. 5a, top left (see color insert), based on measurements at about 700 stations (Ohmura *et al.*, 1998). Where there is more than one measurement within a $4° \times 5°$ gridbox, we average the results. The mean over all gridboxes having data, weighted by gridbox area, is 184 W/m^2, in agreement with Fig. 18 of Wild *et al.* (1998). The true global mean is uncertain due to the limited sampling, but this difficulty can be minimized by averaging the model results over the GEBA gridboxes (G Grid). In Table I we include the modeled $S\downarrow$ integrated over the G Grid and the true global average; these two ways of averaging over the world yield results that tend to differ by several W/m^2, not always in the same sense.

Table I compares the estimates of Wild *et al.* (1998) for global radiation quantities with values obtained in recent GISS global climate models. Model results are 5-year means, years 2–6 of 6-year runs. The SI95 model is described by Hansen *et al.* (1997c). One difference between SI99 and earlier models is more absorbing aerosols in the SI99 model, as quantified below. Another change that may affect these results is improvement in the cloud physics beginning with the SI97 model (Del Genio *et al.*, 1996). The radiation scheme is the same in all models: It uses the k distribution method for gaseous absorption and the adding method for multiple scattering with spectrally dependent aerosol and cloud scattering parameters to ensure self-consistency between solar and thermal regions. Clear comparisons can be made among the runs with the SI99 model, which differ only in atmospheric composition. Differences among the runs are meaningful only if they exceed a few W/m^2, because the cloud cover fluctuates from run to run, especially for the G Grid. The clearest demonstration of the aerosol effect is the run with all aerosols removed. This shows that the assumed 1990 aerosol distribution reduces $S\downarrow$ by 11 W/m^2 for the true global average and by 18 W/m^2 averaged over the GEBA gridboxes.

$S\downarrow$, as simulated in the GISS climate model, agrees well with the GEBA data, as summarized in Table I and Fig. 5. SI95 has 5–10 W/m^2

Table I
Global Radiation Quantities[a]

	$S\downarrow$ (W/m^2)		$\alpha \times S\downarrow$ (W/m^2)	A_{atm} (W/m^2)	Albedo (%)
	G Grid	Global			
Wild estimates	184	—	154	85	30
SI95 model, 1980 atmos.	194	190	167	66	30.8
SI99 model, 1950 atmos.	188	188	163	66	32.9
SI99 model, 1990 atmos.	179	182	159	70.4	33.0
SI99 model, no aerosols	197	193	168	63.5	32.3

[a] Estimated by Wild *et al.* (1998) and as calculated in recent versions of the GISS global climate model. Results are global, but for $S\downarrow$ results are also given for the GEBA network of stations.

more solar radiation hitting the surface than observed. But SI99, with its more absorbing aerosols, agrees even more closely with observations. Sulfate, black carbon, and organic aerosols are time dependent in the SI99 model, so results are given for both 1950 and 1990. The observations were taken over a few decades, so an average of 1950 and 1990 seems appropriate for comparison. With this choice the SI99 model agrees with GEBA data within 1 W/m^2 on average (Fig. 5, lower left), but if aerosols are removed there would be a significant discrepancy of 13 W/m^2 with GEBA (Fig. 5, lower right).

$\alpha \times S\downarrow$, the solar radiation absorbed by the Earth's surface, is at least 5 W/m^2 more in our current model than estimated by Wild *et al.* (1998), implying that our surface is slightly darker. Surface albedo in recent GISS models is specified in detail, with ocean albedo including effects of whitecaps as a function of wind speed (Gordon and Wang, 1994) and subsurface particulate scattering (Gordon *et al.*, 1988), while the land albedo varies seasonally with vegetation and snow cover and depends on soil properties (Matthews, 1983; Hansen *et al.*, 1983, 1997c). We believe that our largest error is an *underestimate* of surface absorption in the Himalayas in the summer. But the discrepancy with the estimate of Wild *et al.* (1998) for surface absorption is small in any case.

A_{atm}, the solar radiation absorbed in the atmosphere, is almost 15 W/m^2 less in our model than in the estimate of Wild *et al.* (1998). Much of this difference is associated with the planetary albedo in our model being higher (32–33%) than the observed albedo of 30%, which is based mainly on Earth Radiation Budget Experiment (ERBE) data (Barkstrom *et al.*, 1989).

In summary, there is no discrepancy between the model and observations of solar radiation reaching the Earth's surface. Our calculated atmospheric absorption of 70–71 W/m^2 is 14–15 W/m^2 less than that estimated by Wild. We argue below that absorbers omitted or underestimated in our model can increase atmospheric absorption to only about 75 W/m^2. Before considering the likely sources of the remaining 10 W/m^2 discrepancy with Wild's estimate for A_{atm}, we discuss how the near agreement of the GCM with GEBA observations can be reconciled with the conclusion that most models underestimate absorption by 20–40 W/m^2.

We believe, in agreement with Garrett (see above), that absorption by aerosols and water vapor has been underestimated in some models. That is why we said that the Lacis and Hansen (1974) parameterization for solar absorption may have inadvertently contributed to the "missing" atmospheric absorption issue. That parameterization, adopted by a number of GCM groups, does not include aerosols, and for that reason we never used it in our climate models. We use the more general correlated k distribution method (Lacis and Oinas, 1991) with explicit integration over the spectrum to achieve accurate scattering and absorption by clouds and aerosols. The water vapor parameterization of Lacis and Hansen, though quite accurate given its simplicity, underestimates absorption of solar radiation by 5–10% for typical water vapor amounts, as judged by the more general k distribution method or line-by-line calculations (Ramaswamy and Freidenreich, 1992). Especially when combined with the low water vapor amounts in many atmospheric models, this also contributes to underestimates of absorption of solar radiation.

The effect of aerosols is illustrated in Fig. 5 (and Table I), where we compare results from our SI99 model with and without aerosols. The aerosols in our SI99 model are a combination of sulfates, organics, black carbon, soil dust, and sea salt as summarized and compared with other aerosol climatologies in Table II. The sulfates, organics, and black carbon each contain a time-dependent anthropogenic component as well as a natural component. Time dependence is not included in either the soil dust or biomass burning (which contributes both organics and black carbon) because of insufficient available information. The aerosol distributions, based in part on aerosol transport models (Tegen *et al.*, 1997; Koch *et al.*, 1999), will be described in more detail elsewhere. The principal change of aerosols that has occurred in successive GISS climate models has been the addition of more absorbing aerosols, as illustrated in Fig. 6, which shows that the global mean aerosol single-scatter albedo decreased from 0.954 in the SI95 model to 0.927 in the SI99 model.

Absorption by SI99 aerosols is due principally to black carbon and soil dust, and only slightly to organics. The black carbon distribution, based on a transport model (Tegen *et al.*, 1997), is especially uncertain; if it is reduced by a factor of 2 the net single-scatter albedo increases from 0.927 to 0.943. The small absorption by organics, presumably occurring mainly at ultraviolet wavelengths, is based on measurements of Tica Novakov (private communication, 1999). Sea salt amount is very uncertain; we multiply the optical depth of Tegen *et al.* (1997) by 4 to account for submicron particles (Quinn and Coffman, 1999). But sea salt is nonabsorbing, so it has little effect on atmospheric absorption.

How realistic is the aerosol absorption in the SI99 model? Although we have concern that the black carbon amount could be exaggerated, other factors work the other way. Actual aerosols often are mixtures of compositions, which tends to decrease the net single-scatter albedo. Also satellite data (Nakajima, *et al.*, 1999) reveal greater aerosol amount in the tropical Western Pacific and Indian Ocean regions than in our model, perhaps in part a consequence of the fact that we did not have data to include

Table II
Aerosol Optical Depth and Single-Scatter Albedo

	Optical depth				Single-scatter albedo	
	Andreae	Seinfeld	SI95 model	SI99 model (1950/1990)	SI95 model	SI99 model
Sulfates						
Trop. natural	0.021	0.014	0.045	0.0067	1.00	1.00
Trop. anthro.	0.032	0.019	0.030	0.0090/0.0222	0.99	1.00
Black carbon						
Industrial	0.006	0.003	0.011	0.0021/0.0067	0.48	0.31
Biomass burning	[a]	[a]	—	0.0014	—	0.48
Organic carbon						
Natural	0.019	0.014	—	0.0032	—	0.98
Industrial	0.003	0.002	—	0.0086/0.0267	—	0.96
Biomass burning	0.027[a]	0.017[a]	—	0.0124	—	0.93
Soil dust	0.023	0.023	0.042	0.0324	0.96	0.89
Sea salt	0.003	0.003	0.012	0.0267	1.00	1.00
Other						
Volcanic	0.004	0.001	0.012	0.005 + variable (total = 0.0065/0.011)	1.00	1.00
NO_x	0.003	0.002	—	—	—	—
Industrial dust	—	0.004	—	—	—	—
Total	0.144	0.102	0.152	0.109/0.149	0.954	0.935 (1950) 0.927 (1990)

From Andreae, 1995, and Seinfeld, 1996.
[a] Black carbon included with organic aerosol optical depth.

time-dependent biomass burning and did not include a Western Pacific biomass source. Because of the complexity of aerosols, the best verification of aerosol absorption is probably field data for the net aerosol single-scatter albedo. Data from field campaigns off the eastern United States and near Europe and India suggest that absorption as great as that in Fig. 6 is not unrealistic, but more extensive and precise data are needed.

What about other possible absorption, besides aerosols? Several minor effects are not included in our present radiation calculations, for example, oxygen dimer (Newnham and Ballard, 1998) and nitrogen continuum (Boissoles *et al.*, 1994) absorption, but these are likely to produce at most a few W/m^2. A popular idea, championed by Bob Cess, is that clouds somehow absorb more sunlight than calculated. However, as a GCM experiment, we doubled the calculated absorption by liquid and ice cloud particles and found the effect to be negligible because of absorption by water vapor in the same spectral regions. Finite (horizontal) cloud extent

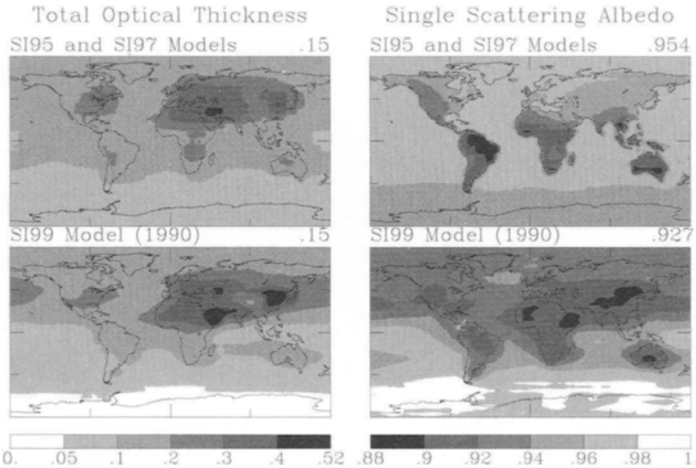

Figure 6 Optical depth and single-scatter albedo of aerosols in GISS GCM.

needs to be accounted for, but it does not introduce substantial absorption. Water vapor absorption is underestimated in our and many other models because the troposphere tends to be about 1–2°C cooler than observed, and thus also drier than observed, but at most this could produce a few W/m^2 of additional absorption. For these reasons we believe that atmospheric absorption is at most about 75 W/m^2.

Finally, assuming atmospheric absorption is not more than 75 W/m^2, how is the remaining 10 W/m^2 difference with Wild's estimate of 85 W/m^2 accounted for? In our present model 5 W/m^2 of this difference is in our larger surface absorption and the other 5 W/m^2 is in our planetary albedo being larger than 30% (our calculated albedo is about 31.5% if atmospheric absorption is 75 W/m^2). The ERBE planetary albedo of 30% is uncertain by at least 1% because it depends on uncertain models for the angular distribution of reflected sunlight and on detectors that do not have a uniform response over the solar spectrum. We suspect that an Earth albedo of 31–32% is possible. But the division of this 10 W/m^2 between surface absorption and planetary albedo can be shifted, and such detailed discussion pushes the data beyond current accuracy levels.

The bottom line is that we find no evidence for a 20–40 W/m^2 radiation mystery and no need for an exotic absorber. The solar radiation reaching the planetary surface is in good agreement between our climate model and observations. This does not mean that a better understanding of absorption of solar radiation, especially by atmospheric aerosols, is unimportant. On the contrary, we must have improved knowledge of aerosols and their changes to predict long-term climate change (Hansen *et al.*, 1998b), and the results and discussion in this section only reinforce the need for better aerosol observations.

VI. GLOBAL WARMING DEBATE

It has been 20 years since the global warming discussions of Charney and Arakawa in 1979. Is our understanding of this topic improving? The picture drawn by the media is one of opposing camps in perpetual fundamental disagreement. Opposing interpretations of the science seem likely to persist, because of the perceived economic stakes associated with energy policies.

The public debate is not as scientific as we would prefer. It can be difficult to find references for public statements or positions of participants. Publication of your own research does not ensure that it will be represented accurately. An egregious example, from my perspective, was congressional testimony of Patrick Michaels in 1998 in which he extracted from our Fig. 3 (see earlier section) the simulated global temperature for scenario A, compared this with observed global temperature, and concluded that my congressional testimony in 1988 had exaggerated global warming. If he had used the entire figure, and noted that real-world climate forcings have been following scenario B, he would have been forced to a very different conclusion.

Recently I had the opportunity to debate "global warming" with Richard Lindzen (Schlumberger, 1998), who has provided much of the intellectual underpinnings for global warming "skeptics." It seemed to me that it may aid future progress to delineate our fundamental differences of opinion, thus providing a way to pin each other down and a basis to keep tabs on progress in understanding. So I went through Dick's publications and made a list of our key differences, for use in my closing statement at the debate. As it turned out, closing statements were eliminated from the debate format at the last minute. But I used this list (Table III) in a debate with Patrick Michaels (AARST, 1998), and, with the same objective of pinning down key issues, I briefly discuss each of the six items here.

A. Reality of Warming

Lindzen (1989) and others have questioned the reality of global warming. Many "greenhouse skeptics" continue to argue that it is only an urban

Table III

Fundamental Differences with R. Lindzen, as Prepared for Schlumberger (1998) Discussion and Used in AARST (1998) Debate

1. Observed global warming: real or measurement problem?
   Hansen: Warming 0.5–0.75°C in past century; $\geq$ 0.3°C in past 25 years.
   Lindzen: Since about 1850 "more likely... 0.1°C.
2. Climate sensitivity (equilibrium response to doubled CO_2).
   Hansen: 3 $\pm$ 1°C *Lindzen*: $\leq$ 1°C
3. Water vapor feedback
   Hansen: Positive (upper tropospheric H_2O increases with warming)
   Lindzen: Negative (upper tropospheric H_2O decreases with warming)
4. CO_2 contributions to the ~33°C natural greenhouse effect
   Lacis and Hansen: Removing CO_2 and trace gases with water vapor fixed would cool the Earth 5–10°C; with water vapor allowed to respond, it would remove most of the greenhouse effect.
   Lindzen: If all other GHGs (such as CO_2 and CH_4) disappeared, over 98% of the natural greenhouse effect would remain.
5. When will global warming and climate change be obvious?
   Hansen: With the climatological probability of a hot summer represented by two faces (say, painted red) of a six-faced die, judging from our model by the 1990s, three or four of the six die faces will be red. It seems to us that this is a sufficient "loading" of the dice that it will be noticeable to the man in the street
   Lindzen: I personally feel that the likelihood over the next century of greenhouse warming reaching magnitudes comparable to natural variability remains small.
6. Planetary disequilibrium
   Hansen: Earth is out of radiative equilibrium by at least 0.5 W/m^2.

effect. We summarize elsewhere (Hansen *et al.*, 1999) evidence that global surface temperature has risen sharply in recent decades and that there has been 0.5–0.75°C global warming since 1880. The warming is largest in remote ocean and high-latitude regions, where local human effects are minimal, and the geographical patterns of warming clearly represent climatic phenomena, not patterns of human development. The instrumental temperature measurements are supported by borehole temperature profiles from hundreds of locations around the world (Harris and Chapman, 1997; Pollack *et al.*, 1998) and by analysis of the near-global meltback of mountain glaciers during the past century (Oerlemans, 1994).

The issue of the reality of global warming survives only because tropospheric temperatures showed essentially no warming over the first 19 years of satellite measurements, 1979–1997. For such a brief period it is not expected that surface and tropospheric temperature changes must coincide, especially in view of measured and suspected changes of atmospheric ozone, aerosols, and clouds. Indeed, tropical surface temperatures hardly increased during 1979–1997, so we would not anticipate much increase of global tropospheric temperature (Hansen *et al.*, 1999). Because of the small temperature change during 1979–1997, small measurement errors can add to real differences in surface and tropospheric trends and cause a qualitative impact on their comparison. But tropospheric warming becomes obvious when one includes (radiosonde) data from several years preceding 1979 and as data following 1997 are added to the record. Temperature measurements deserve continued attention, but the reality of long-term warming is already apparent to most analysts and it is our expectation that this topic will recede as an issue as additional data are collected.

B. Climate Sensitivity

Lindzen argues that climate sensitivity is less than or approximately 1°C for doubled CO_2 and may be as small as 0.3–0.5°C (Lindzen, 1997). We have presented an analysis of paleoclimate data (Hansen *et al.*, 1984, 1993, this paper) that we maintain not only confirms the climate sensitivity estimated by Charney and Arakawa, but sharpens it to 3 ± 1°C. It is our expectation that confidence in this high climate sensitivity will increase as paleoclimate data continue to improve and as their significance for analyzing climate sensitivity is more widely accepted. Climate models can contribute further to this discussion by showing that the details of paleoclimate changes can be simulated realistically.

The approach of attempting to infer climate sensitivity from the current rate of global warming, as discussed in CDAC (1983) and IPCC (1996), will remain fruitless as long as major climate forcings remain unmeasured (Hansen *et al.*, 1998b). A more meaningful constraint on climate sensitivity could be obtained from observations of ocean heat content, as discussed in Subsection F below, but full interpretation of changes in ocean heat content also requires that climate forcings be measured.

C. Water Vapor Feedback

This feedback is related to climate sensitivity, but it is so fundamental that it deserves specific attention. Lindzen has argued that with global warming tropospheric water vapor will decrease at altitudes above 2–3 km (Lindzen, 1990). This contrasts sharply with our expectation based on global climate modeling that water vapor will increase through most of the troposphere with global warming (Hansen *et al.*, 1984).

Water vapor feedback has resisted definitive empirical assessment, because water vapor is not accurately measured and tropospheric temperature change in the past 20 years has been small. Ozone depletion, which cools the upper troposphere, complicates empirical assessment, because it tends to counteract upper tropospheric warming due to increasing carbon dioxide (Hansen *et al.*, 1997c). But ozone depletion is expected to flatten out, while the well-mixed greenhouse gases continue to increase. Thus it should be possible to verify this feedback empirically, if upper tropospheric water vapor is accurately monitored.

D. CO_2 Contribution to Natural Greenhouse

Lindzen (1992) has argued that "Even if all other greenhouse gases (such as carbon dioxide and methane) were to disappear, we would still be left with over 98% of the current greenhouse effect" (p. 88) and makes a similar statement elsewhere (Lindzen, 1993). We believe that this contention, also made in essence by other greenhouse skeptics, illustrates a lack of understanding of the basic greenhouse mechanism that in turn contributes to their expectation that climate should be stable. Although water vapor is the strongest greenhouse gas, the other greenhouse gases contribute a large portion of the present 33°C greenhouse effect on Earth.

Radiation calculations are straightforward, but they need to be made in the context of a climate model to be relevant. And because climate models are complex, results can be debated and obfuscated, which discourages any effort to invest time in addressing this somewhat academic issue per se. But the history of the Earth includes dramatic changes of both climate and atmospheric composition. Ongoing improvements in the knowledge of these changes will provide an opportunity to study the Earth's climate over a large range, and this will incidentally illuminate the contribution of CO_2 to the Earth's natural greenhouse effect.

E. When Will Climate Change Be Obvious?

Lindzen (1989) has said that he believes it unlikely that warming will reach magnitudes comparable to natural variability in the next century. On the contrary, we argue that global mean warming is already comparable to natural variability of global temperature and the warming should soon reach a level comparable to the natural variability of local seasonal mean temperature (Hansen *et al.*, 1988, 1998a). This topic is important because agreement on substantial efforts to curb global warming may require that climate change first be apparent to people.

We have examined practical measures of climate such as seasonal heating degree days, defining an index of change in units of the local standard deviation (Hansen *et al.*, 1998a). We find that in large parts of the world this index is at or near a level such that climate change should be noticeable to the perceptive layperson. If global warming continues as in our scenario B simulations, climate change should be more generally obvious in the next decade.

F. Planetary Disequilibrium

The most fundamental measure of the state of the global greenhouse effect is the planetary "disequilibrium" (imbalance between incoming and outgoing radiation). Averaged over a few years, this imbalance is a simple measure of all climate forcings, measured and unmeasured. Specifically it is the integral over time of past forcings weighted by their exponential decay, with the decay constant being the ocean response time. But this imbalance is not a simple measure of the forcings, because the ocean response time, as discussed in Section IV.A, is not just a function of ocean mixing rates, but rather is a strong function of climate sensitivity. A planetary radiation imbalance must exist today, if climate sensitivity is as high (and thus the ocean response time as long) as we estimate and if increasing greenhouse gases are the dominant climate forcing.

Lindzen has not addressed specifically planetary radiation imbalance, as far as I know, but his positions regarding climate sensitivity and ocean response time would yield a negligible imbalance. We have inferred a planetary disequilibrium of at least approximately 0.5 W/m^2 based on climate simulations for 1979–1996 (Hansen *et al.*, 1997c). An imbalance of this magnitude has practical implications, implying that at least 0.4°C future global warming is still "in the pipeline."

It will be difficult to measure the radiation imbalance directly; we noted in Section V that the Earth's albedo presently is uncertain by at least 1% (3.4 W/m^2). But the imbalance can be deduced indirectly, because the only place the excess energy can go is into the ocean and into melting of ice. A global mean rate of even 0.1 W/m^2 used for melting ice would raise sea level by about 1 cm/year, well above observed rates. Thus most of the energy imbalance must raise the ocean temperature, which can be measured accurately.

White *et al.* (1998) find a substantial positive rate of heat storage between the sea surface and the top of the main pycnocline at latitudes

60°N–20°S for years 1955–1996. Our coupled atmosphere–ocean simulations (Plate 4 of Hansen *et al.*, 1997c) suggest that heat storage at higher latitudes may be large and that storage beneath the top of the main pycnocline is significant. Although temperature changes beneath the ocean mixed layer are small, the mass of water is so great that heat storage at depth can be important. Temperature measurements are needed globally for the full ocean depth.

The aim should be to measure the heat content with an accuracy sufficient to determine the rate of energy storage over a period as short as a year. Climate fluctuations such as El Niños cause a variability in the heat storage rate, but would not prevent use of it to infer information on climate forcings and the long-term energy imbalance. The rate of heat storage for the entire ocean would provide a crucial measure of the state of the planet, a measure that, in our opinion, is more fundamental than the mean global temperature.

VII. A CAUTIONARY CONCLUSION

Nostalgia can cloud perceptions, yet it is clear that the scientific approach of Arakawa and Charney, toward building of models and their application to climate problems, is a paragon for researchers. The essence of that approach is a focus on the relevant climate physics and design of models to represent that physics. A close corollary is use of the models to define needed observations, with continual iterations between data and models.

Technological advances in computing capabilities are opening the potential to advance our modeling capabilities and understanding of climate change. But achievement of that potential requires continued emphasis on the climate physics, on brainpower over megaflops. This may seem obvious, and any commentary perceived as criticism will be met with the response that the focus is on climate physics.

Yet it is difficult to witness current discussions of national climate research plans without concern. The most common measure of modeling prowess seems to be model resolution, or what is worse, the number of simulations that are added to the set of IPCC simulations for the 21st century. It is useful to have a number of such simulations, and we have argued for using and emphasizing a broad range of scenarios, yet with current uncertainties in the models and in the climate forcings driving the models, the law of diminishing returns with additional projections is reached quickly.

We are all pursuing the goal of understanding the climate system so that people and policy makers have information to help make the best decisions. The issue is how to get there. Moving in the direction of a centralized top-down approach is deleterious, in my opinion, because it opens too much of a danger of specification of what to compute and how to do it. That may be good for converging on a single answer, which might even be a goal of some people, but it is hardly in the interests of the best science and thus the long-term interests of the public.

These concerns should not mask an underlying optimism about the prospects for improved understanding of long-term climate change. The spectacular technical improvements in computing, data handling, and communication capability are ideal for increasing scientific cooperation and communication. At the same time there are improving capabilities for global observations that promise to make the modeling and scientific collaborations more productive.

Two topics of this chapter illustrate the potential for improved understanding of climate change: the cooling in the United States in the past 50 years and heat storage in the ocean. We found that models, notably of Arakawa's pedigree and with a relatively coarse resolution of 400–500 km, can simulate U.S. cooling. This provides the potential to investigate the mechanisms behind this regional climate trend, and in turn the possibility of anticipating future change. It should be straightforward to isolate the ocean regions driving the continental temperature change, but it may be more challenging to understand the causes of the ocean changes. A complete analysis will depend on having appropriate observations of climate forcings.

The rate of heat storage in the ocean is important for studies of regional climate change, and it is crucial for analysis of global climate change. An accurate current heat storage rate would provide an invaluable constraint on the net global climate forcing and climate sensitivity. Continued monitoring of heat storage, along with satellite monitoring of the major climate forcings, and preferably ice sheet and ocean topography, would serve as an integral measure of the state of the climate system and provide important data for analyzing mechanisms of long-term global climate change. Technology exists for the temperature measurements, but it must be deployed globally and measure the entire depth of the ocean.

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REFERENCES

AARST (American Association for the Rhetoric of Science and Technology), "Science Policy Forum," New York, Nov. 20, 1998, (G. R. Mitchell and T. M. O'Donnell, eds.), Univ. Pittsburgh.

Andreae, M. O. (1995). Climatic effects of changing atmospheric aerosol levels. *In* "World Survey of Climatology, Vol. 16: Future Climates of the World" (A. Henderson-Sellers, ed.), pp. 341–392. Elsevier, Amsterdam.

Arking, A. (1996). Absorption of solar energy in the atmosphere: Discrepancy between model and observations, *Science* **273,** 779–782.

Barkstrom, B., E. Harrison, G. Smith, R. Green, J. Kibler, and R. Cess (1989). Earth radiation budget experiment (ERBE) archival and April 1985 results. *Bull. Am. Meteor. Soc.* **70,** 1254–1262.

Boissoles, J., R. H. Tipping, and C. Boulet (1994). Theoretical study of the collision-induced fundamental absorption spectra of N_2–N_2 pairs for temperatures between 77 and 297K. *J. Quant. Spectrosc. Radiat. Transfer* **51,** 615–627.

Boyle, J. S. (1998). Evaluation of the annual cycle of precipitation over the United States in GCMs: AMIP simulations, *J. Climate* **11,** 1041–1055.

Bryan, K., and M.D. Cox (1972). An approximate equation of state for numerical model of ocean circulation. *J. Phys. Oceanogr.* **15,** 1255–1273.

CDAC (1983). "Changing Climate, Report of the Carbon Dioxide Assessment Committee." National Academy Press, Washington, DC.

Cess, R. D., M. Zhang, F. P. J. Valero, S. K. Pope, A. Bucholtz, B. Bush, C. S. Zender, and J. Vitko (1999). Absorption of solar radiation by the cloudy atmosphere: Further interpretations of collocated aircraft observations. *J. Geophys. Res.* **104,** 2059–2066.

Charney, J. (1979). "Carbon Dioxide and Climate: A Scientific Assessment." National Academy Press, Washington, DC.

CLIMAP Project Members (1981). Seasonal reconstruction of the Earth's surface at the last glacial maximum, *Geolog. Soc. Am.* Mmap and chart series, **MC-36.**

Cox, M. D. (1984). A primitive equation three-dimensional model of the ocean, GFDL Ocean Group Tech. Rep. 1. Geophys. Fuid Dyn. Lab., Princeton, NJ.

Del Genio, A. D., and M. S. Yao (1993). Efficient cumulus parameterization of long-term climate studies: The GISS scheme. *Am. Meteor. Soc. Monogr.* **46,** 181–184.

Del Genio, A. D., and A. Wolf (2000). Climatic implications of the observed temperature dependence of the liquid water path of low clouds in the Southern Great Plains. *J. Climate*, in press.

Del Genio, A. D., M. S. Yao, W. Kovari, and K. K. W. Lo (1996). A prognostic cloud water parameterization for global climate models, *J. Climate* **9,** 270–304.

Garratt, J. R., A. J. Prata, L. D. Rotstayn, B. J. McAvaney, and S. Cusack (1998). The surface radiation budget over oceans and continents, *J. Climate* **11,** 1951–1968.

Gordon, H. R., and M. Wang (1994). Influence of oceanic whitecaps on atmospheric correction of SeaWIFS, *Appl. Opt.* **33,** 7754–7763.

Gordon, H. R., O. B. Brown, R. H. Evans, J. W. Brown, R. C. Smith, K. S. Baker, and D. K. Clark (1988). A semi-analytic radiance model of ocean color. *J. Geophys. Res.* **93,** 10,909–10,924.

Guilderson, T. P., R. G. Fairbanks, and J. L. Rubenstone (1994). Tropical temperature variations since 20,000 years ago: Modulating interhemispheric climate change. *Science* **263,** 663–665.

Hansen, J. E., W. C. Wang, and A. A. Lacis (1978). Mount Agung eruption provides test of a global climate perturbation. *Science* **199,** 1065–1068.

Hansen, J., D. Johnson, A. Lacis, S. Lebedeff, P. Lee, D. Rind, and G. Russell (1981). Climatic impact of increasing atmospheric carbon dioxide. *Science* **213,** 957–966.

Hansen, J., G. Russell, D. Rind, P. Stone, A. Lacis, S. Lebedeff, R. Ruedy, and L. Travis (1983). Efficient three-dimensional global models for climate studies: Models I and II, *Mon. Wea. Rev.* **111,** 609–662.

Hansen, J., A. Lacis, D. Rind, G. Russell, P. Stone, I. Fung, R. Ruedy, and J. Lerner (1984). Climate sensitivity: Analysis of feedback mechanisms. *Geophys. Mono.* **29,** 130–163.

Hansen, J., I. Fung, A. Lacis, D. Rind, S. Lebedeff, R. Ruedy, G. Russell, and P. Stone (1988). Global climate changes as forecast by the Goddard Institute for Space Studies three-dimensional model. *J. Geophys. Res.* **93,** 9341–9364.

Hansen, J., A. Lacis, R. Ruedy, M. Sato, and H. Wilson (1993). How sensitive is the world's climate? *Natl. Geogr. Res. Explor.* **9,** 142–158.

Hansen, J. R. Ruedy, A. Lacis, G. Russell, M. Sato, J. Lerner, D. Rind, and P. Stone (1997a). Wonderland climate model. *J. Geophys. Res.* **102,** 6823–6830.

Hansen, J., M. Sato, and R. Ruedy (1997b). Radiative forcing and climate response. *J. Geophys. Res.* **102,** 6831–6864.

Hansen, J., and 42 Others (1997c). Forcings and chaos in interannual to decadal climate change. *J. Geophys. Res.* **102,** 25,679–25,720.

Hansen, J., M. Sato, J. Glascoe, and R. Ruedy (1998a). A common-sense climate index: Is climate changing noticeably? *Proc. Natl. Acad. Sci.* **95,** 4113–4120.

Hansen, J., M. Sato, A. Lacis, R. Ruedy, I. Tegen, and E. Matthews (1998b). Climate forcings in the industrial era. *Proc. Natl. Acad. Sci.* **95,** 12753–12758.

Hansen, J., R. Ruedy, J. Glascoe, and M. Sato (1999). GISS analysis of surface temperature change. *J. Geophys. Res.* **104,** 30997–31022.

Harris, R. N., and D. S. Chapman (1997). Borehole temperatures and a baseline for 20th-century global warming estimates. *Science* **275,** 1618–1621.

Hartke, G. J., and D. Rind (1997). Improved surface and boundary layer models for the GISS general circulation model. *J. Geophys. Res.* **102,** 16,407–16,442.

Hoffert, M. I., and C. Covey (1992). Deriving global climate sensitivity from paleoclimate reconstructions. *Nature* **360,** 573–576.

Intergovernmental Panel on Climate Change (1996). "Climate Change 1995" (J. T. Houghton, L. G. Meira, Filho, B. A. Callandar, N. Harris, A. Kattenberg, and K. Maskell, eds.). Cambridge Univ. Press, Cambridge, UK.

Kerr, R. A. (1995). Darker clouds promise brighter future for climate models. *Science* **267,** 454.

Koch, D., D. Jacob, I. Tegen, D. Rind, and M. Chin (1999). Tropospheric sulfur simulation and sulfate direct radiative forcing in the GISS GCM. *J. Geophys. Res.* **104,** 23799–23822.

Lacis, A. A., and J. E. Hansen (1974). A parameterization for the absorption of solar radiation in the Earth's. *J. Atmos. Sci.* **31,** 118–133.

Lacis, A. A., and V. Oinas (1991). A description of the correlated *k* distribution method for modeling nongray gaseous absorption, thermal emission, and multiple scattering in vertically inhomogeneous atmospheres. *J. Geophys. Res.* **96,** 9027–9063.

Lindzen, R. (1989). EAPS' Lindzen is critical of global warming prediction. *MIT Tech Talk,* **34,** No. 7, 1–6.

Lindzen, R. S. (1990). Some coolness concerning global warming. *Bull. Am. Meteorol. Soc.* **71,** 288–299.

Lindzen, R. S. (1992). Global warming: The origin and nature of the alleged scientific consensus. *Cato Rev. Bus. Govt.* **2,** 87–98.

Lindzen, R. (1993). Absence of scientific basis. *Nat. Geog. Res. Explor.* **9,** 191–200.

Lindzen, R. S. (1997). Can increasing carbon dioxide cause climate change? *Proc. Natl. Acad. Sci.* **94,** 8335–8342.

Lorius, C., J. Jouzel, D. Raynaud, J. Hansen, and H. Le Treut (1990). The ice-core record: Climate sensitivity and future greenhouse warming. *Nature* **347,** 139–145.

Manabe, S., and F. Moller (1961). On the radiative equilibrium and heat balance of the atmosphere. *Mon. Wea. Rev.* **89,** 503–532.

Manabe, S., and R. J. Stouffer (1980). Sensitivity of a global climate model to an increase of CO_2 concentration in the atmosphere. *J. Geophys. Res.* **85,** 5529–5554.

Manabe, S., and R. F. Strickler (1964). Thermal equilibrium of the atmosphere with a convective adjustment. *J. Atmos. Sci.* **21,** 361–385.

Manabe, S., and R. T. Wetherald (1975). The effects of doubling the CO_2 concentration on the climate of a general circulation model. *J. Atmos. Sci.* **32,** 3–15.

Matthews, E. (1983). Global vegetation and land use: New high resolution data bases for climate studies. *J. Clim. Appl. Meteor.* **22,** 474–487.

Merilees, P. E. (1975). The effect of grid resolution on the instability of a simple baroclinic model. *Mon. Wea. Rev.* **103,** 101–104.

Nakajima, T., A. Higurashi, N. Takeuchi, and J. R. Herman (1999). Satellite and ground-based study of optical properties of 1997 Indonesian forest fire aerosols. *Geophys. Res. Lett.* **26,** 2421–2424.

Newnham, D. A., and J. Ballard (1998). Visible absorption cross sections and integrated absorption intensities of molecular oxygen O_2 and O_4). *J. Geophys. Res.* **103,** 28,801–28,816.

Oerlemans, J. (1994). Quantifying global warming from the retreat of glaciers. *Science* **264,** 243–245.

Ohmura, A., and 13 coauthors (1998). Baseline Surface Radiation Network (BSRN/WCRP), a new precision radiometry for climate research. *Bull. Am. Meteor. Soc.* **79,** 2115–2136.

Paltridge, G. W., and C. M. R. Platt (1976). "Radiative Processes in Meteorology and Climatology." Elsevier, New York.

Pollack, H. N., H. Shaopeng, and P. Y. Shen (1998). Climate change record in subsurface temperatures: A global perspective. *Science* **282,** 279–281.

Prather, M. J. (1986). Numerical advection by conservation of second-order moments. *J. Geophys. Res.* **91,** 6671–6680.

Quinn, P. K., and D. J. Coffman (1999). Comment on "Contribution of different aerosol species to the global aerosol extinction optical thickness: Estimates from model results" by Tegen *et al. J. Geophys. Res.* **104,** 4241–4248.

Ramanathan, V., R. J. Cicerone, H. J. Singh, and J. T. Kiehl (1985). Trace gas trends and their potential role in climate change. *J. Geophys. Res.* **90,** 5547–5566.

Ramaswamy, V., and S. M. Freidenreich (1992). A study of broadband parameterization of the solar radiative interactions with water vapor and water drops. *J. Geophys. Res.* **97,** 11,487–11,512.

Reynolds, R. W., and T. M. Smith (1994). Improved global sea surface temperature analyses. *J. Clim.* **7,** 929–948.

Rind, D., R. Suozzo, N. K. Balachandran, A. Lacis, and G. L. Russell (1988). The GISS global climate/middle atmosphere model, I. Model structure and climatology. *J. Atmos. Sci.* **45,** 329–370.

Rosenzweig, C., and F. Abramopoulos (1997). Land surface model development for the GISS GCM. *J. Climate* **10,** 2040–2054.

Russell, G. L., and J. A. Lerner (1981). A new finite-differencing scheme for the tracer transport equation. *J. Appl. Meteorol.* **20,** 1483–1498.

Russell, G. L., J. R. Miller, and L. C. Tsang (1985). Seasonal oceanic heat transports computed from an atmospheric model. *Dynam. Atmos. Oceans* **9,** 253–271.

Russell, G. L., J. R. Miller, and D. Rind (1995). A coupled atmosphere-ocean model for transient climate change studies. *Atmos. Oceans* **33,** 683–730.

Russell, G. L., J. R. Miller, D. Rind, R. A. Ruedy, G. A. Schmidt, and S. Sheth (2000). Comparison of model and observed regional temperature changes during the past 40 years. *J. Geophys. Res.*, in press.

Sato, M., J. E. Hansen, M. P. McCormick, and J. B. Pollack (1993). Stratospheric aerosol optical depth, 1850–1990. *J. Geophys. Res.* **98,** 22,987–22,994.

Schlumberger Research, Climate Change and the Oil Industry, A Debate, Ridgefield, CT, Oct. 15, 1998, available at www.slb.com/research/sdr50.

Schrag, D. P., G. H. Hampt, and D. W. Murray (1996). Pore fluid constraints on the temperature and oxygen isotopic composition of the glacial ocean. *Science* **272,** 1930–1932.

Seinfeld, J. H. (1996). "Aerosol Radiative Forcing of Climate Change." National Research Council, National Academy Press, Washington, DC.

Shindell, D. T., R. L. Miller, G. A. Schmidt, and L. Pandolfo (1999a). Greenhouse gas forcing of Northern Hemisphere winter climate trends. *Nature* **399,** 452–455.

Shindell, D. T., D. Rind, N. Balachandran, J. Lean, and P. Lonergan (1999b). Solar cycle variability, ozone and climate. *Science* **284,** 305–308.

Smith, T. M., R. W. Reynolds, R. E. Livesay, and D. C. Stokes (1996). Reconstruction of historical sea surface temperature using empirical orthogonal functions. *J. Clim.* **9,** 1403–1420.

Somerville, R. C. J., P. H. Stone, M. Halem, J. Hansen, J. S. Hogan, L. M. Druyan, G. Russell, A. A. Quirk, and J. Tenenbaum (1974). The GISS model of the global atmosphere. *J. Atmos. Sci.* **31,** 84–117.

Tegen, I., P. Hollrig, M. Chin, I. Fung, D. Jacob, and J. Penner (1997). Contribution of different aerosol species to the global aerosol extinction optical thickness: Estimates from model results. *J. Geophys. Res.* **102,** 23,895–23,915.

Thompson, D. W. J., and J. M. Wallace (1998). The Arctic oscillation signature in the wintertime geopotential height and temperature fields. *Geophys. Res. Lett.* **25,** 1297–1300.

Tselioudis, G., and W. B. Rossow (1994). Global, multiyear variations of optical thickness with temperature in low and cirrus clouds. *Geophys. Res. Lett.* **21,** 2211–2214.

Wang, W. C., Y. L. Yung, A. A. Lacis, T. Mo, and J. E. Hansen (1976). Greenhouse effects due to man-made perturbations of trace gases. *Science* **194,** 685–690.

White, W. B., D. R. Cayan, and J. Lean (1998). Global upper ocean heat storage response to radiative forcing from changing solar irradiance and increasing greenhouse gas/aerosol concentrations. *J. Geophys. Res.* **103,** 21,355–21,366.

Wild, M., A. Ohmura, H. Gilgen, E. Roeckner, M. Giorgetta, and J. J. Morcrette (1998). The disposition of radiative energy in the global climate system: GCM-calculated versus observational estimates. *Clim. Dynam.* **14,** 853–869.

Chapter 5

A Retrospective Analysis of the Pioneering Data Assimilation Experiments with the Mintz–Arakawa General Circulation Model

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I. Introduction
II. Description of Experiments
III. Results of GEOS Simulation Experiments
IV. Conclusions
References

I. INTRODUCTION

We have performed a retrospective analysis of a simulation study, published about 30 years ago, which had a profound impact on satellite meteorology. The paper had the strange title "Use of incomplete historical data to infer the present state of the atmosphere." It was authored by J. Charney, M. Halem, and R. Jastrow, and appeared in the *Journal of the Atmospheric Sciences*, in September 1969 (Charney *et al.* 1969). We decided that the numerical experiments which formed the basis of that paper should be repeated using a contemporary model, particularly in view of their relevance to upcoming satellite missions.

General Circulation Model Development

Secondly, by the end of 2000, NASA plans to launch the EOS PM platform, which will carry a new generation of temperature sounders, the Atmospheric Infra-Red Sounder (AIRS) and the Advanced Microwave Sounding Unit (AMSU). These sounders will have substantially increased spectral and spatial resolutions and are expected to produce an increase in accuracy over that of today, perhaps attaining 1 K accuracies throughout the column in clear and cloudy regions. AIRS will also provide greatly improved vertical humidity profiles, which really are not feasible with today's instruments. These expectations are reminiscent of the situation in July 1969, just after the launch of NIMBUS 3, which carried the first of a new class of remote sensors, namely, the Space Infra-Red Sounder (SIRS-A), which could acquire global vertical temperature profiles, with a potential accuracy of 1 K in clear tropical regions. Shortly thereafter, Dr. Morris Tepper, NASA program manager, visited Goddard Institute for Space Studies (GISS) to meet with Charney, Jastrow, and Halem to ask what impact such data could have in numerical weather prediction. It was then that Charney proposed that we conduct an experiment to assimilate complete temperature fields synoptically into a GCM, in order to infer the geostrophic winds. He called Mintz and Arakawa to ask them to lend GISS their model to perform such experiments, and they agreed to do so.

Those experiments produced some very interesting results that initially raised some skepticism in the community. Most modelers had expected that the insertion of "foreign" temperature fields without balancing would generate spurious disturbances in the model. Another conclusion which generated considerable discussion was that a knowledge of the temperature fields alone could lead to adjustments of the wind and pressure fields even in the tropics, where the geostrophic approximation is not accurate. The retrospective analysis reported here investigates the model dependencies of those results. At that time, the Mintz–Arakawa model had a very coarse spatial resolution by present standards, $7° \times 9°$ by two levels, and very crude physical parameterizations compared with today's models. Clearly, the simulation experiment of Charney *et al.* (1969) ignored the operational world weather observing system with hundreds of upper air radiosondes and thousands of ground surfaces observing systems and focused mainly on a conjecture that Charney (1969) had earlier presented at the 1968 International Numerical Weather Prediction Conference in Tokyo, Japan. The Charney conjecture was based on a simplified linear hydrodynamical model. In Chapter 6 of this volume, Schubert shows that the relevant system of first-order equations in several variables can be reduced to a single equation of higher order in a single unknown with a forcing term expressed in terms of higher order temporal and spatial derivatives. Initial conditions of state variables are replaced with higher order temporal derivatives of the single unknown variable. Such a linear higher order differential equation can be solved by the method of Green's functions, but Charney conjectured that the GCM would produce such a solution "automatically" if provided with the temperature history over a sufficiently long integration period. Although this conjecture was not at all obvious at the time, it is generally accepted today.

Ghil *et al.* (1977, 1979) analytically proved the Charney conjecture for certain simple atmospheric models. These results were extended by Ghil (1980). In practice, numerous problems with real data and with complexities of current atmospheric models render Ghil *et al.*'s theory not strictly applicable. However, the power of the process whereby continuous assimilation of temperature profiles can be used to infer complete global states or even just extratropical atmospheric states is still of considerable interest today. Thus, we set out to repeat the experiments of Charney *et al.* (1969) using a contemporary GCM.

II. DESCRIPTION OF EXPERIMENTS

In this retrospective study, we conduct a simulation experiment that is as nearly as possible identical to the original experiment of Charney *et al.* (1969), except that we employ the Goddard Earth Observing System (GEOS) GCM (Takacs *et al.*, 1994) in place of the Mintz–Arakawa GCM (Langlois and Kwok, 1969).

The satellite system configuration that the original Charney *et al.* (1969) experiments were designed to simulate consisted of one polar orbiting NIMBUS 3 satellite carrying infrared and microwave scanning sounders capable of providing temperature profiles throughout the atmosphere under clear and cloudy conditions, including the radiative surface temperatures. Based on today's NOAA operational satellite configuration, we assume for these experiments that two satellites can provide synoptic global coverage every 6 hr.

The original experiments consisted of generating a "history" record to represent the synoptic state of the atmosphere by conducting a long integration with a GCM. The Charney *et al.* (1969) experiment employed the Mintz–Arakawa GCM two-level model at 400 and 800 mb and $7° \times 9°$ grid spacing in latitude and longitude, respectively. A second integration was performed with the Mintz–Arakawa model starting with initial conditions from the "history" file at day 85 with a random perturbation error of 1 K added to the temperature field at all grid points. This integration was carried out to day 95 to produce an initial state that was considerably different from the history tape. Experiments all starting from this initial state of day 95 were then conducted assimilating the "history" temperature field with different random perturbation errors. The experiments tested a parametric range of assumed temperature accuracy and frequency of insertions.

Our current experiments used the GEOS GCM with 20 levels and $4° \times 5°$ grid spacing in latitude and longitude and much more detailed physical parameterizations (clouds, radiations, turbulence, surface processes, etc.); see Takacs *et al.* (1994). The "history" record was started from an atmospheric state provided by L. Takacs and integrated for 90 days. At day 30, a second integration was started for 60 days with a 1 K random perturbation introduced into the temperature field. The atmospheric state at day 60 was then used as the initial condition for two parametric temperature assimilation experiments. Synoptic temperature fields from the "history" record with random root mean square (rms) errors of 0, 1, and 2.5 K were assimilated into the GEOS GCM at different time intervals (every hour, 3, 6, and 12 hr) for 30 days. A fourth experiment assimilation was carried out with both the temperature field and the surface pressure field.

The following section compares the results of Charney *et al.* (1969) with those obtained by a contemporary model.

III. RESULTS OF GEOS SIMULATION EXPERIMENTS

As mentioned in the previous section, a "history" file was generated by carrying out the numerical integration of the GEOS GCM for 90 days. This file is treated throughout the remainder of the study as an exact measurement notwithstanding all of the limitations of the model. At day 30, a random perturbation or "error" of 1 K is introduced in the temperature fields at all grid points and all levels, and the flow is then recalculated from this initial state for 60 days. The resulting atmospheric state of the "perturbation" run will be compared with the "history" run to confirm that their respective fields are randomly correlated. Results are presented in terms of rms differences of the sea level pressure and 400-mb zonal winds. The results, summarized in Figs. 1 and 2, demonstrates that the sea level pressure and 400-mb wind differences between the perturbed integration and the unperturbed history files grow rapidly with time and then reach asymptotic error differences of 10 mb and 12 m s^{-1}, respectively. After 30 days, an examination of contour plotted differences shows that the sea

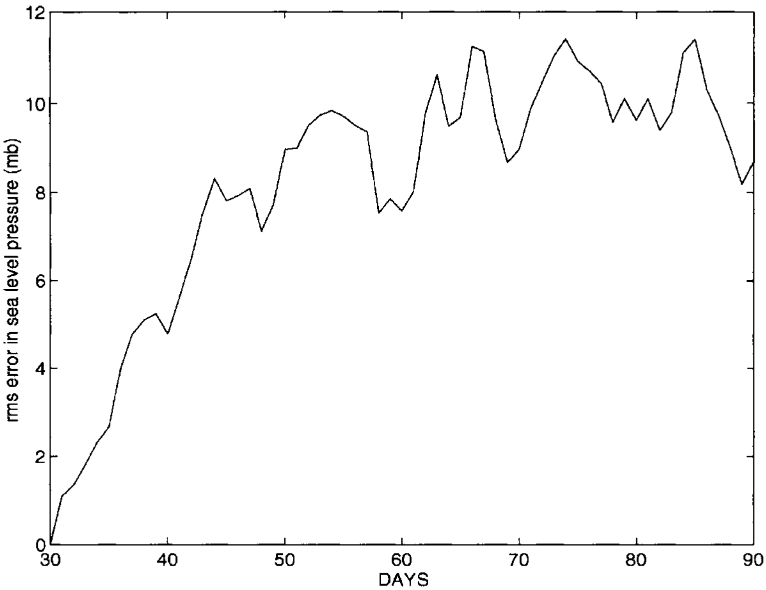

Figure 1 The rms differences in sea level pressure between the history and perturbed runs, from day 30 to day 90.

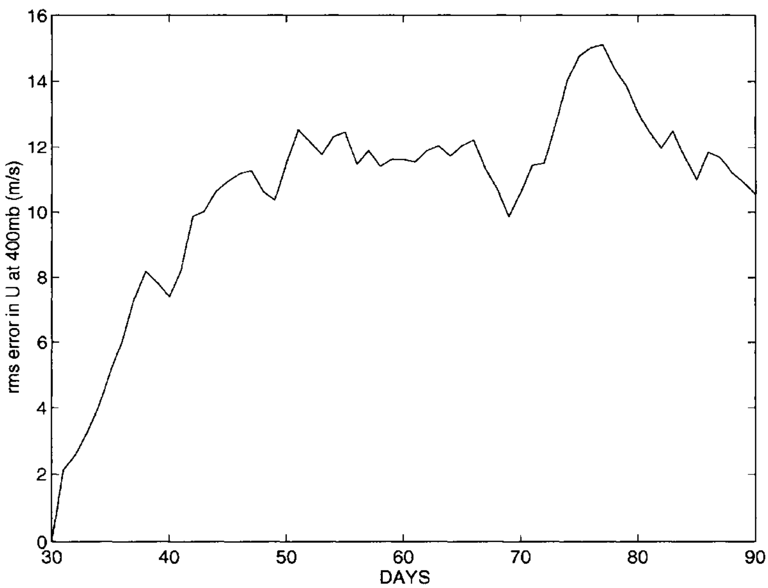

Figure 2 The rms error in the 400-mb zonal wind between history and perturbed files, from day 30 to day 90.

level pressure and the winds are meteorologically distinct and uncorrelated, with no remaining sign of their common parentage.

The next set of runs is designed to investigate the sensitivity of our results to the frequency of data insertion. Charney *et al.* (1969) found that a 12-hr insertion frequency was optimal, but we wanted to find out what would be optimal for the GEOS GCM. The sensitivity experiments were performed by starting from the perturbed file at day 60 and integrating the GCM with exact temperatures inserted from the history file at specified time intervals.

Figures 3 and 4 show the results of inferring the sea level pressure and 400-mb zonal wind fields by inserting data from the history temperature file at intervals of 1, 3, 6, and 12 hr, respectively. It is seen that continuous temperature insertions immediately arrest the growth in the sea level pressure differences (Fig. 1), and reduce the differences to approximately 3 mb for insertions every 3 and 6 hr after 30 days. Insertions of temperature fields every hour and every 12 hr produce asymptotic differences of 3.8 and 4.6 mb, respectively. The 400-mb zonal wind behaves similarly, reducing the differences to about 3.5 m s^{-1} for insertions at intervals of 3 and 6 hr, and to approximately 5.5 m s^{-1} for 1- and 12-hr insertion intervals. This is

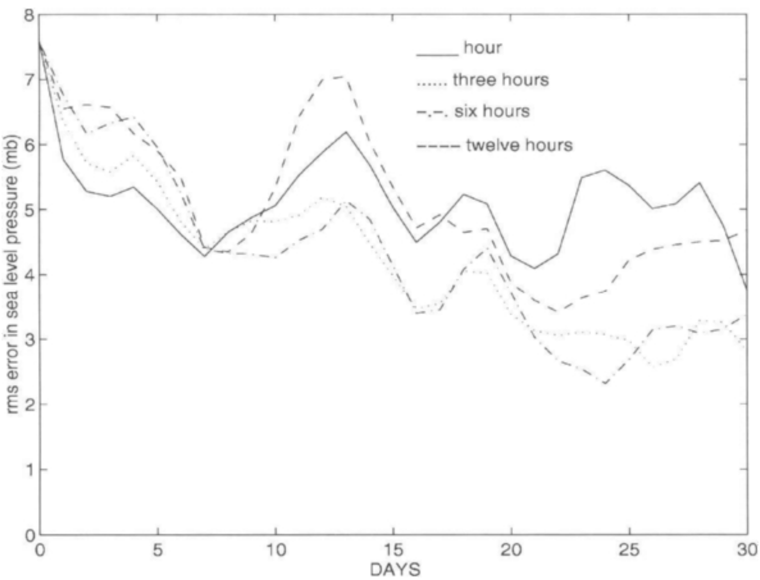

Figure 3 The rms error in sea level pressure in cases for which exact temperatures are inserted every 1, 3, 6, and 12 hr at all grid points.

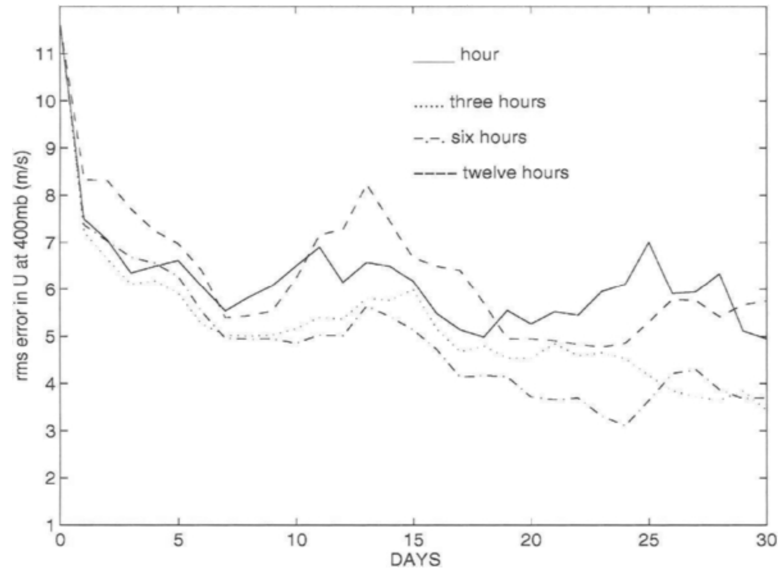

Figure 4 The rms error in 400-mb zonal wind (m s^{-1}), in cases for which exact temperatures are inserted every 1, 3, 6, and 12 hr at all grid points.

in contrast to the results without temperature corrections, given in Figs. 1 and 2, which show divergences from the history sea level pressure and 400-mb zonal wind, with amplitudes of 8 mb and 12 m s^{-1}, respectively, after 30 days.

The greatest reduction of rms error, i.e., the smallest rms error, was achieved when the "correct" temperatures were inserted every 3 or 6 hr. A more frequent insertion (every hour for instance) gives rise to oscillations in the wind field. The 6-hr interval was chosen for the experiment described below. This choice is consistent with an observing system consisting of two satellite overpasses a day. Operational weather forecasting systems today employ two satellites in this manner.

A second set of runs was performed in which temperatures were inserted at each grid point with random errors of 1 or 2.5 K, representing a range of observational errors, for comparison with exact temperature insertion. Figure 5 indicates that the insertion of temperatures with 1 K errors at 6-hr intervals reduces the global sea level pressure difference to approximately 3 mb, the same level as the insertion of exact temperatures. With temperature errors of 2.5 K, corresponding to the current estimated accuracies of today's operational sounders, the asymptotic differences are on the order of 4 to 5 mb. Figure 6 shows similar behavior with the global

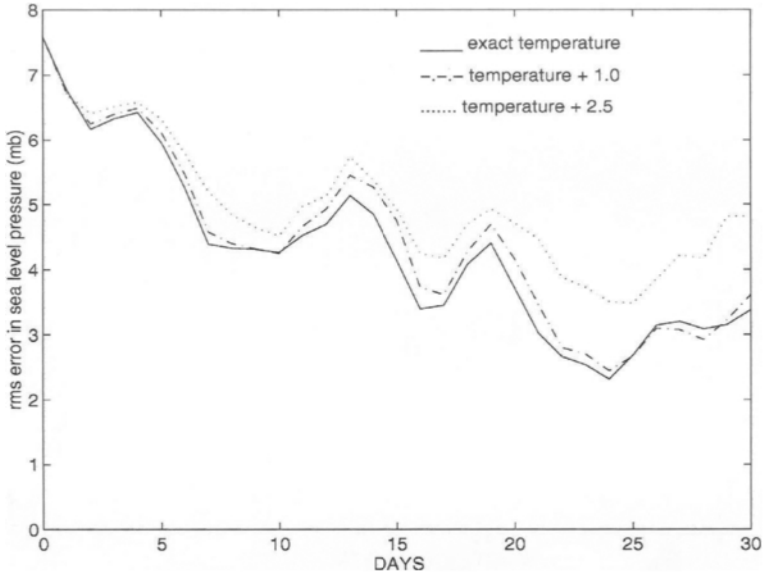

Figure 5 The rms error in sea level pressure, in cases for which temperatures with random error perturbations of 0, 1, and 2.5 K are inserted every 6 hr at all grid points.

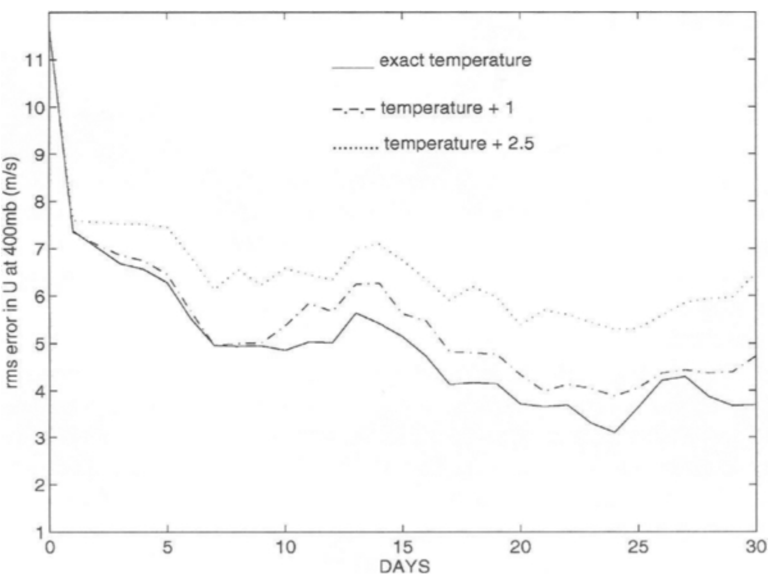

Figure 6 The rms error in 400-mb zonal wind (m s^{-1}), in cases for which temperatures with random error perturbations of 0, 1, and 2.5 K are inserted every 6 hr at all grid points.

wind adjustments, which reduce the wind errors to 4 and 6.5 m s^{-1}, respectively. Although this is a significant reduction of errors from the initial state, it falls somewhat short of the desired 3 m s^{-1} global wind errors.

We next wish to compare the results of the experiments described above with those derived earlier obtained by Charney *et al.* (1969). Figures 7 and 8, taken from Charney *et al.* (1969), show that the 400-mb extratropical and tropical zonal winds are reduced to below 1 m s^{-1} with 1 K temperature errors. These very favorable results, referred to earlier in the introduction, generated both skepticism and excitement over the prospective use of temperature sounders to infer the global wind fields.

Figure 9 shows that, for the GEOS GCM with 1 K sounding errors, the 400-mb wind differences at 48°N are reduced to about 4 m s^{-1}, while with 2.5 K temperature errors they are reduced to 6 m s^{-1}. These results are similar to those of Charney *et al.* (1969), but differ in the magnitude of the asymptotic errors. At the equator, shown in Fig. 10, the 1 K sounder errors lead to oscillatory wind adjustments ranging from 4 to 6 m s^{-1}, down from an uncorrected error of 7 m s^{-1}. Temperatures with 2.5 K errors also produce oscillations with magnitudes between 6 and 8 m s^{-1}, with a mean of 7 m s^{-1}, effectively showing no reduction relative to the uncorrected

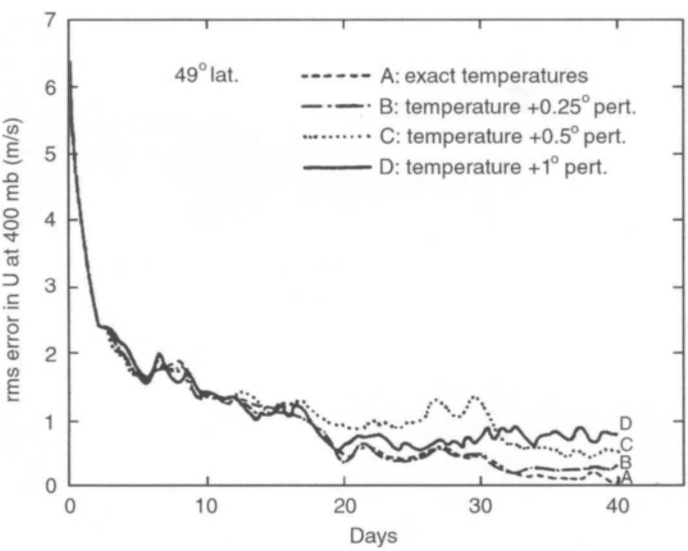

Figure 7 Charney *et al.* (1969) results with the two-level Mintz–Arakawa GCM: the rms error in 400-mb zonal wind (m s^{-1}) at 49°N, in cases for which temperatures with random error perturbations of 0, 0.25, 0.5, and 1 K are inserted every 12 hr at all grid points. (From Charney *et al.* (1969).)

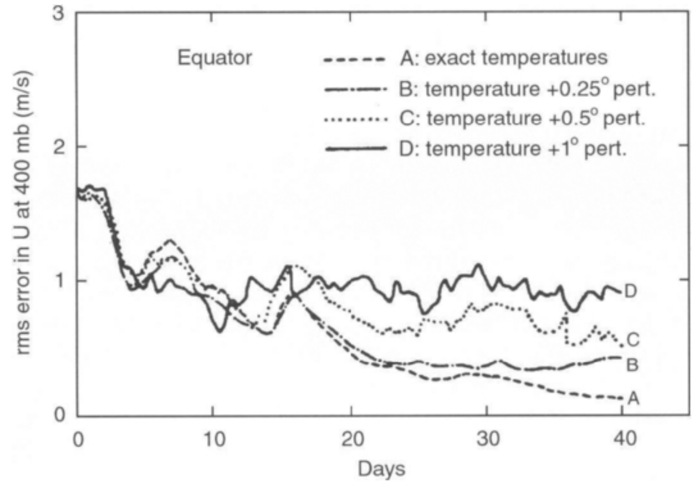

Figure 8 Charney *et al.* (1969) results with the two-level Mintz–Arakawa GCM: the rms error in the 400-mb zonal wind (m s^{-1}) at the equator, in cases for which temperatures with random error perturbations of 0, 0.25, 0.5, and 1 K are inserted every 12 hr at all grid points. (From Charney *et al.* (1969).)

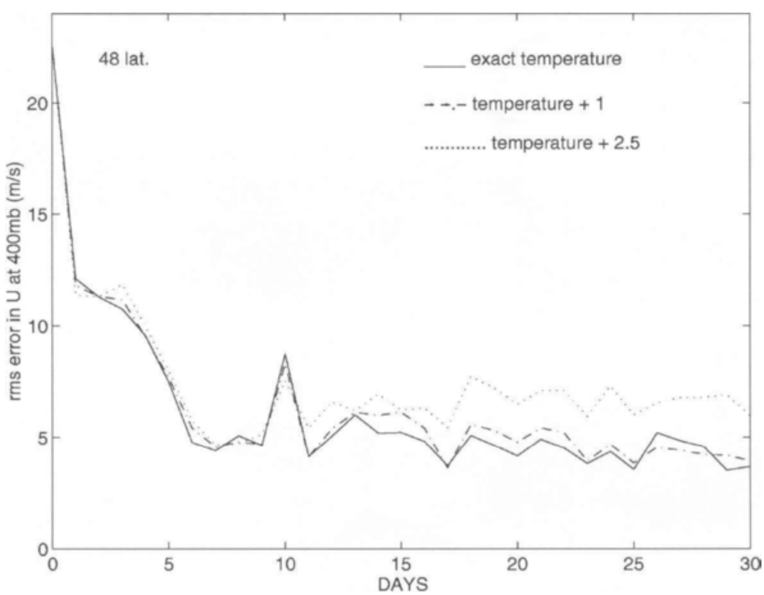

Figure 9 The rms error in 400-mb zonal wind (m s^{-1}) at 48°N, in cases for which temperatures with random error perturbations of 0, 1, and 2.5 K are inserted every 6 hr at all grid points.

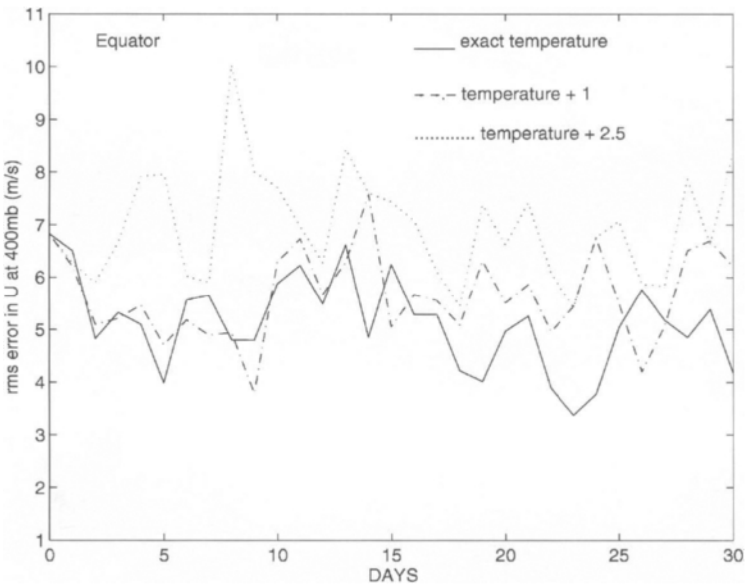

Figure 10 The rms error in 400-mb zonal wind (m s^{-1}) at the equator, in cases for which temperatures with random error perturbations of 0, 1, and 2.5 K are inserted every 6 hr at all grid points.

wind errors. This disagrees with the results of Charney *et al.*, which indicated that highly accurate tropical winds can be inferred from sounding data.

The last experiment was designed to explore whether combining surface pressure data together with temperature data helps in dynamical balancing, especially in the tropics. Figures 11, 12, and 13 compare the global zonal wind errors and meridional wind errors at 48°N and at the equator, for exact temperature insertions, with and without sea level pressure insertions. We see from Fig. 11 that the error reductions in the global winds are significantly greater when surface pressure fields are combined with temperature fields. A more noticeable reduction is achieved at 48°N (Fig. 12), in very close agreement with the results of Charney *et al.* (1969). However, even with exact observations of sea level pressure, there is very little improvement in the inferred equatorial zonal winds (Fig. 13).

IV. CONCLUSIONS

We have performed observing-system simulation studies whose basic objective is the determination of the relationship between the temperature

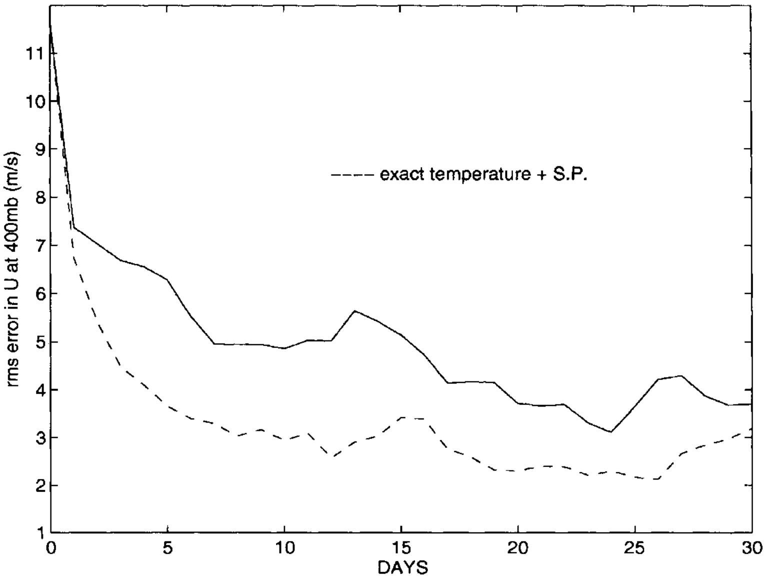

Figure 11 The rms error in 400-mb zonal wind (m s^{-1}), in cases for which exact temperatures are inserted with and without surface pressure every 6 hr at all grid points.

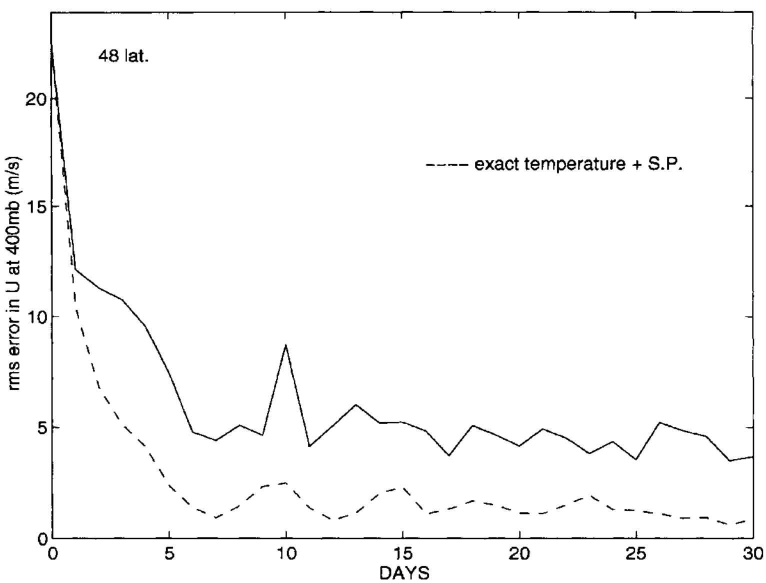

Figure 12 The rms error in 400-mb zonal wind (m s^{-1}) at 48°N, in cases for which exact temperatures are inserted with and without surface pressure every 6 hr at all grid points.

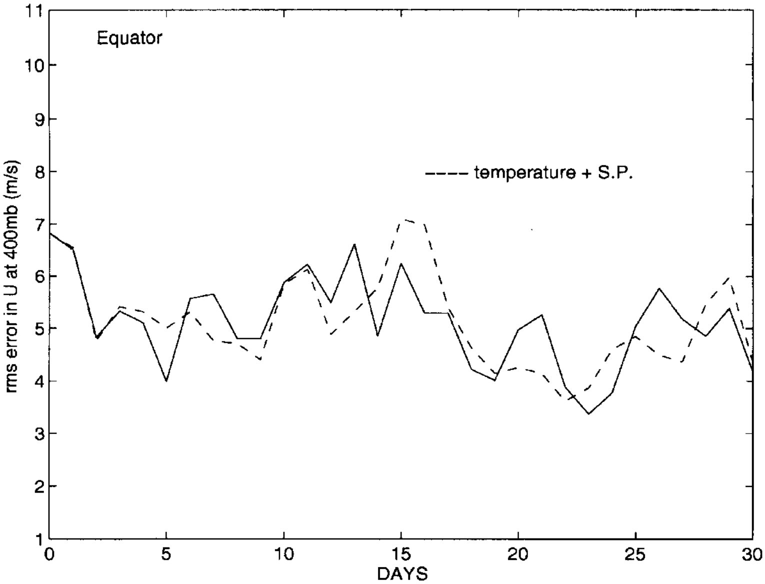

Figure 13 The rms error in 400-mb zonal wind (m s^{-1}) at the equator, in cases for which exact temperatures are inserted with and without surface pressure every 6 hr at all grid points.

errors and the inferred global winds and pressures, for realistic configurations of a proposed earth observing system with advanced vertical temperature sounders. Numerical results obtained with the GEOS GCM indicate that if a continuing day-by-day sequence or history of temperature profiles is inserted into the numerical integrations at appropriate time intervals, wind components and sea level pressures can be determined to a useful degree of accuracy. More precisely, we can draw the following conclusions:

- Based on limited idealized simulations with 1998 GEOS GCM, the gross accuracies of the inferred wind and sea level pressure fields are consistent with the findings of Charney *et al.* (1969), but with somewhat larger asymptotic errors.
- GCMs of higher spatial and vertical resolution assimilate temperature data to substantially improve the inferred winds and sea level pressure where no data are available.
- A system of two polar orbiting satellites with temperatures sounders of 1 K accuracy in clear and cloudy regions, combined with surface pressure observations, should be capable of inferring the global wind fields to the required accuracies of 3 m s^{-1}.

- The conclusion of Charney *et al.* (1969) that it is possible to infer tropical winds from temperature profiles may have been a model-dependent result.
- Assimilating surface pressure greatly improves the rate of adjustment and the asymptotic accuracies of the extratropical winds, but does not significantly improve the inferred tropical winds.

As mentioned earlier, the new integrations reported here were performed with a resolution of $4° \times 5°$ by 20 levels. We plan to carry out further simulations employing finer resolution versions of the same model, as well as additional experiments with other models, to assess the effects of model dependence.

ACKNOWLEDGMENTS

The study of Charney *et al.* (1969) was made possible by Profs. Arakawa and Mintz, who shared the Mintz–Arakawa GCM with our organization at NASA as early as 1964, and again in 1969, agreeing that we could conduct and publish independent research results based on the use of the model.

We also want to take this opportunity to acknowledge that we at NASA are deeply indebted to Professor Arakawa for encouraging so many of his students and colleagues at UCLA to visit the NASA Goddard Space Flight Center. Some have stayed on to become permanent members of our staff. Many have continued to work closely with Arakawa, in introducing his concepts into the NASA model-development effort. Arakawa has often shared with us at NASA his latest ideas and models, well before he publishes them. For example, in 1972 he provided to us an early version of his three-level model, which subsequently evolved into the GISS nine-level model.

We are grateful to R. Rood for making the GEOS GCM available for use in this study. We also wish to thank L. Takacs and S. Nebuda for implementing the GEOS GCM code on the NASA Center for Computational Science (NCCS) computing facilities. The computations presented were all performed on the SGI/CRAY J90 system at the NCCS at Goddard Space Flight Center. We are indebted to the NCCS for making their computing environment and resources available to the authors. We also thank J. Raymond, who provided support in the preparation of this document.

REFERENCES

Charney, J. G. (1969). "Proceedings 1968 WMO/IUGG Symp. on Numerical Weather Prediction," Tokyo, March 1969. Meteorological Society of Japan.

Charney, J., M. Halem, and R. Jastrow (1969). Use of incomplete historical data to infer the present state of the atmosphere. *J. Atmos. Sci.* **26,** 5, 1160–1163.

Ghil, M. (1980). The compatible balancing approach to initialization, and four-dimensional data assimilation. *Tellus* **32,** 198–206.

Ghil, M., B. Shkoller, and V. Yangarber (1977). A balanced diagnostic system compatible with a barotropic prognostic model. *Mon. Wea. Rev.* **105**, 1223–1238.

Ghil, M., M. Halem, and R. Atlas (1979). Time-continuous assimilation of remote-sounding data and its effect on weather forecasting. *Mon. Wea. Rev.* **107**, 140–171.

Langlois, W. E., and H. C. Kwok (1969). Numerical simulation of weather and climate, Technical Report 3. Dept. of Meterorology UCLA.

Takacs, L., A. Molod, and T. Wang (1994). Documentation of the Goddard Earth Observing System GEOS general circulation model, Version 1, Technical Memorandum 104606. NASA.