CLIMATE CHANGE 1995

The Science of Climate Change

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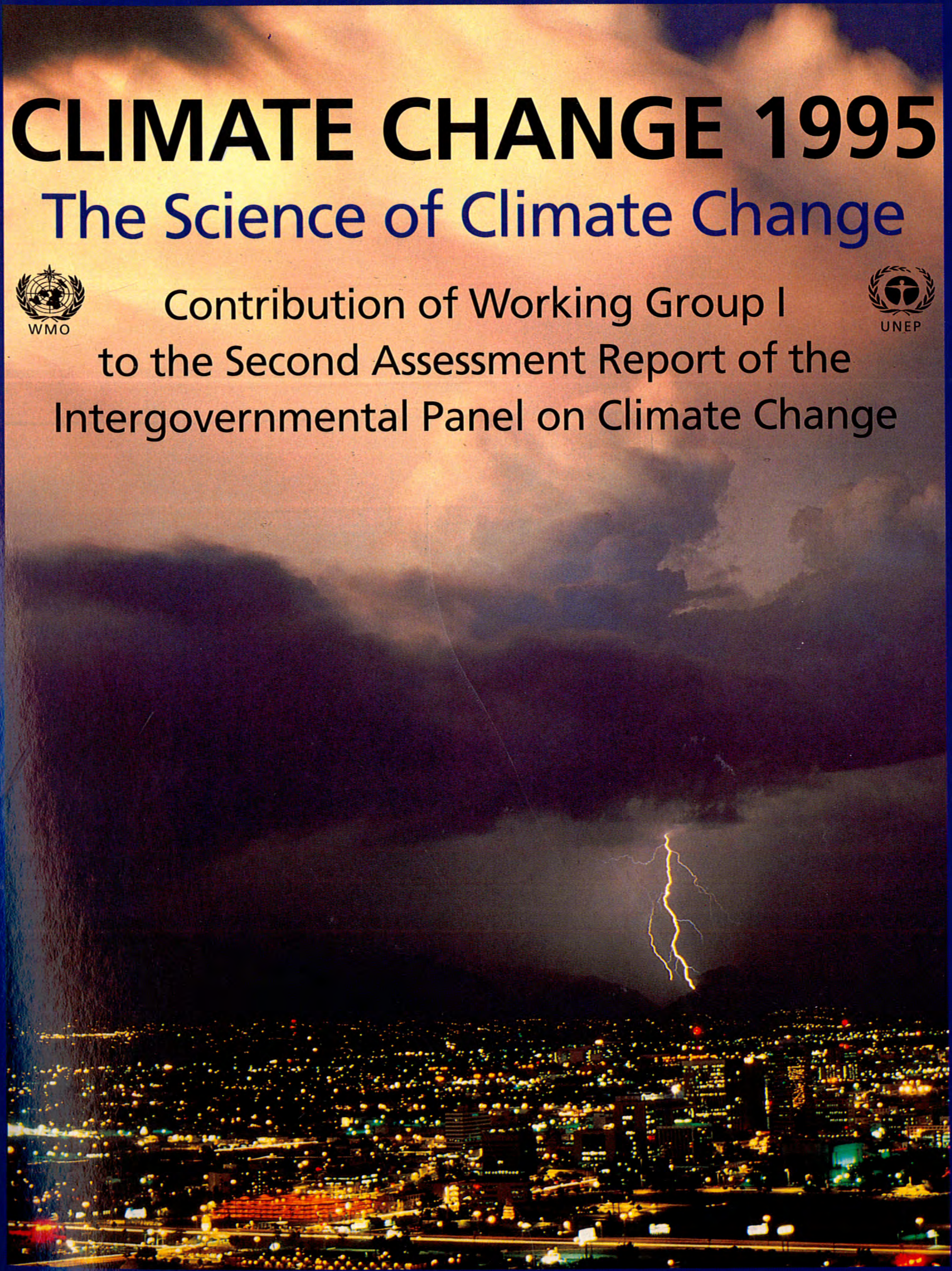
Contribution of Working Group I
to the Second Assessment Report of the
Intergovernmental Panel on Climate Change



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SUMMARY

This chapter assesses the processes in the climate system that are believed to contribute the most to the uncertainties in current projections of greenhouse warming. Many of these processes involve the coupling of the atmosphere, ocean, and land through the hydrological cycle. Continued progress in climate modelling will depend on the development of comprehensive data sets and their application to improving important parametrizations. The large-scale dynamical and thermodynamical processes in atmospheric and oceanic models are well treated and are one of the strengths of the modelling approach. As previously indicated in IPCC (1990, 1992), the radiative effects of clouds and their linkages to the hydrological cycle remain a major uncertainty for climate modelling. The present report, however, goes into much more detail than the past reports in summarising the many facets of this question, recent progress in understanding different feedbacks, and in the development of climate model parametrization treating these processes. This is now a very active research area with much that has been accomplished since IPCC (1990).

Current climate models are highly sensitive to cloud parametrizations, and there are not yet satisfactory means for evaluating the correctness of such treatments. Progress will require improved understanding, observational data sets, sub-grid scale parametrization, and improved modelling of the distribution of atmospheric water in its vapour, liquid, and solid forms, and will not be achieved quickly. Sub-grid scale parametrizations are especially difficult to improve. The determination of cloud-dependent surface radiative and precipitation fluxes is a significant source of uncertainty for both land-surface and ocean climate modelling, making attempts to assess regional climate change problematic.

Clear-sky feedbacks involving changes in water vapour distribution and lapse rates are also uncertain but their global sum varies little between models. The processes determining the distribution of upper tropospheric water vapour are still poorly understood. Water vapour feedback in the lower troposphere is undoubtedly positive, and the preponderance of evidence also points to it being positive in the upper troposphere.

A large-scale dynamical framework for treating the ocean component of climate models is now being used for climate change projections. Sub-grid scale parametrizations in these models are important for surface energy exchange and the thermohaline circulation. A new parametrization for interior mixing appears promising in providing an improved simulation of the global thermocline. Both high latitude and tropical elements of ocean climate models involve important and still inadequately represented processes. In high latitudes, coupling to sea ice models and deep convection are especially important. In the tropics, ocean models need to simulate the large-scale sea surface temperature variability of the El Niño-Southern Oscillation systems.

Climate model treatments of land processes have advanced rapidly since the last assessment. However, there are lags in the validation of these models, in the development of required data sets, in an adequate assessment of how sub-grid scale processes should be represented, and in their implementation in models for climate change projection. None of the land parametrizations include a physically based and globally validated treatment of runoff.

4.1 Introduction to Climate Processes

Climate processes are all the individual physical processes that separately contribute to the overall behaviour of the climate system. They are also the interactions and feedbacks among the individual processes that determine the response of the climate system to external forcing, including the response to global anthropogenic forcing. There are a myriad of such climate processes. This report is focused by the present community experience with numerical modelling. The success of numerical simulations of future climate change hinges on the adequate inclusion of all the climate processes that are responsible for determining the behaviour of the system. A triage approach is used here; processes are treated lightly that are already included adequately in numerical models or those for which there is little or no evidence supporting their importance. The present report emphasises those processes known to contribute substantially to the uncertainties of current numerical simulations of long-term climate response. These all are physical processes. Chemical and biological components are treated in Chapters 2, 9 and 10 respectively, and at present would be de-emphasised by the triage approach for lack of substantial effort to include them in climate models.

A globally averaged temperature increase is the response most easily related to global greenhouse forcing. Current climate models project global temperature increases that vary by over a factor of two to three for a given forcing scenario. These differences are directly attributable to the treatment of cloud processes, their links to the hydrological cycle and their interaction with radiation in the models. Indeed, individual models can give this range of answers, depending on changes only in the cloud parametrization (Senior and Mitchell, 1993). Thus, all the likely contributors to the ultimate representations of cloud processes in climate models need emphasis, as do other aspects of the hydrological cycle and model dynamics that interact with cloud evolution and properties.

Another current focus for climate process studies is the evolving interpretation of the past climate record. Most notable is that of the observed warming of about half a degree over the last century. This is about a factor of two smaller than conventional estimates according to models including only trace gas increases but is consistent with the inclusion of the expected effects of sulphate aerosols.

Observations over continental surfaces show that the warming has been more pronounced at night than during the day (*cf.* Chapter 3). Hansen *et al.* (1995) have hypothesised from a large number of numerical simulations with a simple sector General Circulation Model (GCM) that the observed global trends and continental predominance of night-time warming can be explained if clouds and aerosol have increased over continental surfaces with a global average radiative cooling of about half that of the warming by greenhouse gases increase. For the same direct radiative effects, a cloud increase would have a larger impact on diurnal range than would dry aerosols, since clouds not only cool during daytime but warm the surface at night. GCM calculations by Mitchell *et al.* (1995) indicate that the direct radiative effect of aerosols alone can only partially explain the observed decrease in diurnal range. Further progress will require understanding the time history and distribution of anthropogenic aerosol as well as its effects on cloud properties and its consequent direct and indirect radiative effects.

As mentioned above, ocean surface temperatures, and hence much of climate variability and change are strongly influenced by net energy exchanges between the oceans and the atmosphere. In high latitudes, sea ice is of major significance for modifying these energy exchanges. The possible rapid variations of the thermohaline circulation, first suggested by Bryan (1986), indicate that rearrangements of heat in the ocean can occur relatively quickly so that surface temperatures can respond rapidly to changes in the thermohaline circulation. Further, the thermohaline circulation is particularly sensitive to changes in the high latitude temperature and hydrology; variability on decadal-to-millennial time-scales can arise. On somewhat shorter time-scales, coupled atmosphere-ocean processes in the tropics are a major source of interannual climate variability. Therefore, dynamically active oceans are included in coupled models for a comprehensive evaluation of the response to increases in the radiatively active gases.

Land-surface processes are also now highlighted for several reasons. The land surface is readily modified on the large-scale by human activities. Such modifications may have important regional consequences; historical land modifications may have had larger regional climate impacts up to now than has had greenhouse gas warming. Land-surface processes probably are of lesser importance for the future globally averaged temperature response to greenhouse warming than are cloud processes, but how climate changes over land is of the greatest practical importance to humans and depends substantially on land-surface processes. Furthermore, land-surface processes strongly affect the overlying atmospheric hydrological cycle including clouds, so they arguably will be of major importance as feedbacks for determining changes in regional climate patterns.

This chapter identifies and assesses important processes for incorporation into climate models for projecting climate change and highlights major gaps in our understanding of these processes. These processes occur in the atmosphere, ocean and land surface and involve their coupling through the hydrological cycle.

4.2 Atmospheric Processes

The processes of large-scale dynamics, thermodynamics and mass balance in the atmospheric and oceanic components of the climate models are now included in models with considerable confidence. This is not to imply that the modelling of large-scale circulation and temperature structure can be viewed as completely successful, since it depends not only on the relatively robust components but also on weak elements. Diagnostic comparisons with observations of potential vorticity transport and mixing could help improve confidence in the treatments of large-scale dynamics. Stable dynamical modes in which the coupled atmosphere-ocean system may respond as part of climate change are inadequately understood (e.g., Palmer, 1993). The hydrological cycle and radiation budget are so intimately coupled that they should not be treated separately. Therefore, wherever possible in the following sections, these two aspects are discussed together. Clear-sky water vapour feedback is first examined, then the various feedbacks attributed to clouds that affect the Earth-atmosphere radiative balance, then the additional issues of surface radiative fluxes, and then coupling to precipitation processes. The intention here is not to describe all the physical processes operating in the atmosphere, as that would be far too broad an approach, but rather to focus on those processes that are especially relevant to the various feedbacks, as identified in studies of global warming with climate models.

4.2.1 Water Vapour Amounts

A positive water vapour feedback was hypothesised in the earliest simulations of global warming with simple radiative-convective models (Manabe and Wetherald, 1967). It arises for water vapour near the surface from the strong dependence of the saturation vapour pressure on temperature, as given by the Clausius-Clapeyron equation. Increases in temperature are thus expected to lead to increases in the atmospheric water vapour mixing ratio. Since water vapour is the most important greenhouse gas, such increases in water vapour enhance the greenhouse effect; that is, they reduce the thermal infrared (long-wave) flux leaving the atmosphere-surface system, providing a positive feedback amplifying the initial warming. This feedback operates in all the climate models used in global warming and other studies. However, intuitive arguments for it to apply to water vapour in the upper troposphere are weak; observational analyses and process studies are needed to establish its existence and strength there.

Changes in the vertical decrease of temperature with altitude change surface temperature for a given radiative balance and are known as "lapse rate feedback". Changes in lapse rate act as an additional feedback that can also be substantial and that generally oppose the water vapour feedback. The sum of the water vapour and lapse rate feedbacks comprise the clear-sky feedback. Cess *et al.* (1990) show that the magnitude of the clear-sky feedback is very similar in a wide range of models. However, the partitioning between lapse rate and water vapour feedback may vary substantially between models (Zhang *et al.*, 1994) depending on how convection is parametrized and may depend on the time-scale of the climate change or climate fluctuations considered (Bony *et al.*, 1995).

The consensus view that water vapour provides a strong positive feedback has been challenged by Lindzen (1990), who emphasised the sensitivity of the water vapour feedback to poorly understood processes, such as the profile of cumulus detrainment and the related distribution of water vapour in the upper troposphere. Water vapour is physically most closely controlled by temperature in the lower troposphere, and by transport processes in the upper troposphere. Both regions contribute comparably to the water vapour greenhouse effect.

How upper tropospheric water vapour is distributed and varies with other climate parameter variations is best studied with satellite data (e.g., Soden and Bretherton, 1994). Soden and Fu (1995) relate climatic variations of upper tropospheric water vapour to clear-sky long-wave radiation and to moist convection indicated by the International Satellite Cloud Climatology Project (ISCCP), over a five-year period. These are shown to be highly correlated in the tropical half of the world but uncorrelated outside the tropics. Thus, these data support the conventional view that in the tropics, water vapour is supplied to the upper troposphere primarily by moist convection. They find that this result is maintained, even averaging over the whole tropical belt of 30°S-30°N, and so conclude that the net effect of convection is moistening even allowing for compensating regions of subsidence. Chou (1995) in a case study with two months of data (April 1985 and 1987) over 100°W–100°E infers a somewhat contradictory conclusion that increased convection in the tropics leads to a net reduction in the atmospheric clear-sky greenhouse effect and hence a net drying.

Soden and Fu show that occurrence of tropical convection in the Geophysical Fluid Dynamics Laboratory (GFDL) GCM has a very similar correlation with upper tropospheric water vapour as observed, suggesting that the model captures the essential upward transport processes of water in the tropics. Sun and Held (1995), on the other hand, show that for specific humidities averaged over the tropics, the correlation with surface values declines with height to much smaller values for observed data than in the same model. As they discuss, lack of radiosonde coverage over the eastern and central Pacific may throw into question their observational analyses.

Detailed process studies of cumulus detrainment and of the water budget in mesoscale cumulus convection are also being made through field experiments, to provide a more solid physical basis for testing the GCM treatments of water vapour. Lau *et al.* (1993) used the Goddard Cumulus Ensemble model (GCEM) to investigate the water budget of tropical cumulus convection. Their results on changes in temperature and water vapour induced by surface warming are in agreement with those from GCMs which use only crude cumulus parametrizations.

The details of water vapour feedback in the extra-tropical upper troposphere are also poorly characterised observationally. In the extra-tropics, the relative contributions to upper tropospheric water vapour of lateral transport from the tropics, versus upward transport by large-scale motions or by moist convection are poorly known. Lacis and Sato (1993) showed, in the Goddard Institute for Space Science (GISS) GCM, that the water vapour feedback was almost as strong at middle and high latitudes as it was at low latitudes. Pierrehumbert and Yang (1993) have emphasised the potential small-scale complexity of latitudinal exchanges of water vapour by large-scale eddies at high latitudes, and Kelly *et al.* (1991) have shown some observational evidence for temperature-dependent large-scale high latitude exchanges of water vapour creating a hemispheric asymmetry in upper tropospheric water vapour at these latitudes. Del Genio *et al.* (1994) find, for the GISS model, that large-scale eddies dominate the seasonal variation of upper troposphere water vapour outside the tropical rainbelt.

Feedback from the redistribution of water vapour remains a substantial uncertainty in climate models. That from the lower troposphere seems least controversial. Much of the current debate has been addressing feedback from the tropical upper troposphere, where the feedback appears likely to be positive. However, this is not yet convincingly established; much further evaluation of climate models with regard to observed processes is needed.

Somewhat independent of feedbacks affecting top of the atmosphere long-wave fluxes, water vapour in the lower troposphere affects the long-wave contribution to surface radiation. This feedback has lately been emphasised as contributing somewhat to a reduction in diurnal temperature range with increasing water vapour concentrations from global warming (Mitchell *et al.*, 1995).

4.2.2 Cloud Amounts

The first cloud feedback to be studied in detail in global models involves changes in cloud amounts. These can be extremely complicated because of the many different types of clouds, whose properties and coverage are controlled by many different physical processes, and which affect the radiation budget in many different ways. In the global and annual mean, clouds have a cooling effect on the present climate (the surface and atmosphere), as evaluated from the Earth Radiation Budget Experiment (ERBE) and other satellite measurements. That is, the 31 Wm^{-2} enhancement of the thermal greenhouse effect is exceeded by a 48 Wm^{-2} increase in the reflection of short-wave radiation to space (Ramanathan *et al.*, 1989). But there are large variations in the net cloud forcing with geography and cloud type; indeed some clouds contribute a net warming. For low clouds, the reflected short-wave dominates so that an increase in amount would cool the climate and be a negative feedback on global warming. But thin tropical cirrus clouds are much colder than the underlying surface and act more to enhance the greenhouse effect, so an increase in the amount of this cloud type would be a positive feedback.

Climate model simulations of global warming have found the tropical troposphere to become higher and, in most models, the amounts of high cloud to increase. This increase (which depends on the somewhat uncertain changes of water vapour and relative humidity) enhances the greenhouse effect and produces a positive feedback (e.g., Wetherald and Manabe, 1988; Mitchell and Ingram, 1992). However, its importance relative to other changes in cloud amounts and the detailed changes in the three dimensional cloud distribution, varies significantly between models.

Until recently, climate models have used cloud prediction schemes based largely on presumed relationships between cloud amount and relative humidity and giving cloud amount as the only parameter. Such schemes introduce or remove condensed water instantaneously in amounts that are prescribed or depend only on temperature. Many modelling groups are now moving to prognostic cloud water variables that explicitly determine the amount of liquid water in each grid cell (e.g., Sundqvist, 1978; Le Treut and Li, 1988; Sundqvist *et al.*, 1989; Roeckner *et al.*, 1990; Smith, 1990; Ose, 1993; Tiedtke, 1993; Del Genio *et al.*, 1995; Fowler *et al.*, 1996). The predicted cloud water may vary more smoothly, persisting for hours after the agencies that formed it have ceased (as is especially true for cirrus); it can be used to determine interactively the optical properties of the clouds, as well as the precipitation rate. A prognostic cloud water variable thus improves a model's physical basis, but not without considerable difficulties, including proper representation of mixed-phase clouds (e.g., Senior and Mitchell 1993) and various numerical issues.

Some features of this new approach for predicting clouds are:

- Prediction of the condensed water and its partitioning between liquid and ice (Sundqvist, 1978; Li and Le Treut, 1992). Separate treatment of the liquid and ice cloud particles is important, because they undergo significantly different microphysical and thermodynamical processes and have different optical properties. The transition of water to ice is a critical process and empirically dealt with in some of the latest models.

- Action of cumulus clouds as liquid and/or ice sources for the stratiform clouds in some models (e.g., Ose, 1993; Tiedtke, 1993; Del Genio *et al.*, 1995; Fowler *et al.*, 1996). This important physical link between two types of cloud systems and its explicit incorporation into GCMs may mark a significant step forward in cloud parametrization.

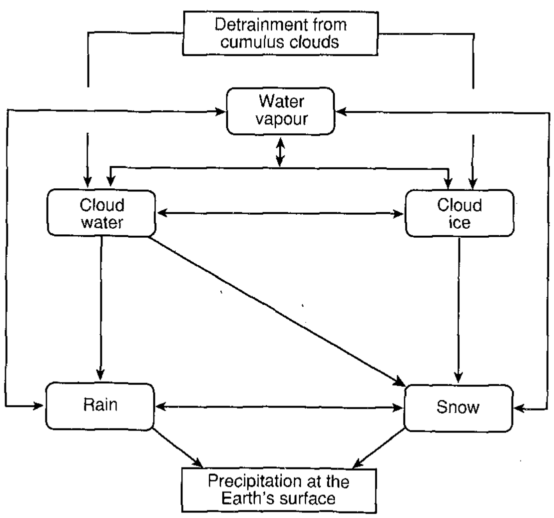

Figure 4.1: Diagram illustrating prognostic variables in a cloud microphysics scheme, and the processes that affect them.

- Physically based parametrization of the various cloud microphysical processes, such as the evaporation of cloud water and ice to water vapour in subsaturated air, conversion of cloud water and ice to rain and snow in supersaturated air, and parametrization of cloud droplets and ice crystals depending on cloud condensation nuclei (CCN) and ice nuclei. Figure 4.1 schematically illustrates some of these processes. For example, cumulus detrainment can produce small ice crystals that combine to make larger falling snowflakes. The snow falling through a warm lower atmosphere will melt to become rain, although ice and liquid may coexist in a range of temperatures whose width is somewhat uncertain. The efficiency of the so-called Bergeron-Findeisen mechanism in this range of coexistence has a decisive impact on how effective the release of precipitation is in the cloud. As a consequence, this mechanism strongly influences the resulting amount of cloud water, and indirectly the optical properties of the cloud.

- Physically based parametrization of the cloud optical properties and fractional cloud amount. It is far from obvious how to determine the cloudy fraction of a GCM grid-box; this cloud fraction depends on the sub-grid scale distribution of the water and ice contents (e.g., Kristjánsson, 1991; Kvamstø, 1991). The dependence of cloud optical properties on sub-grid scale spatial heterogeneity and mixed phases remains an important problem.

The new generation of cloud parametrization has led to some improvements in simulations of the Earth's radiation budget (e.g., Senior and Mitchell, 1993; Del Genio *et al.*, 1995; Fowler and Randall, 1996). Comparisons of the global distribution of simulated cloud water and ice concentrations against observations are problematic. Although satellite observations of the macroscopic distribution of total-column liquid water content are available over the oceans (Njoku and Swanson, 1983; Prabhakara *et al.*, 1983; Greenwald *et al.*, 1993), these show substantial differences, possibly a result of different algorithms (Lin and Rossow, 1994). Also, observations of macroscopic ice water content are lacking. Field data for local cloud ice measurements have been obtained in regional experiments such as the First ISCCP Regional Experiment (FIRE) (Heymsfield and Donner 1990) and the International Cirrus Experiment (ICE) (Raschke *et al.*, 1990), but these are very difficult to convert into macroscopic averages.

High clouds

High clouds are very effective in trapping outgoing long-wave radiation, and so tend to warm the Earth. The net radiative effect on climate of anvils and cirrus clouds associated with deep convection in the tropics is near zero (e.g., Ramanathan *et al.*, 1989) because short-wave cooling and long-wave warming nearly cancel each other. The solar cooling acts mainly at the Earth's surface; however, how much is absorbed rather than reflected by clouds is currently controversial, as discussed in Section 4.2.6. Long-wave warming by anvils and cirrus clouds associated with deep convection in the tropics acts mainly on the atmosphere and can influence the general circulation of the atmosphere (Slingo and Slingo, 1988; Harshvardhan *et al.*, 1989).

Recent work has led to simple ice crystal scattering parametrizations for climate models (Ebert and Curry, 1992; Fu and Liou, 1993). Improvements may be needed to account for ice crystal size and shape effects, especially as climate models develop the capability to compute crystal sizes and realistic distributions of the ice water path. Comparison of cirrus properties from ISCCP with those generated by GCM parametrization using observed large-scale dynamic and thermodynamic fields from operational analyses shows encouraging agreement in the spatial patterns of the cirrus optical depths (Soden and Donner, 1994). However, determining cirrus modification of net radiative fluxes to the accuracies desirable for climate models may require accuracies in measurement of cirrus cloud temperature, ice water content and/or scattering properties that, in some cases, are beyond current observational and computational abilities (Vogelmann and Ackerman, 1996), suggesting that improved accuracies in these will be needed for climate studies.

Middle clouds

Frontal cloud systems are major sources of precipitation and cloud cover in the mid-latitudes. Climate models are incapable of explicitly resolving frontal circulations, although they do resolve "large-scale" cloud cover and rain associated with these systems. However there is a dearth of observations over the ocean and generally in the Southern Hemisphere to validate the microphysical parametrizations used to calculate the radiative and precipitation processes with these mid-level clouds (Ryan, 1996).

Low clouds

The marine stratocumulus clouds that commonly occur on the eastern sides of the subtropical oceans (e.g., Hanson, 1991) are important for their solar reflection (e.g., Slingo, 1990), and are, at present, under-predicted by many atmospheric GCMs. Similar clouds occur in the Arctic in summer (e.g., Herman and Goody, 1976), as well as over the mid-latitude oceans (Klein and Hartmann, 1993). They must be simulated successfully in order to obtain realistic sea surface temperature (SST) distributions in coupled atmosphere-ocean models (e.g., Robertson *et al.*, 1995).

Low clouds, such as marine stratocumulus, are favoured by strong capping temperature inversions (e.g., Lilly, 1968; Randall, 1980; Klein and Hartmann, 1993) as, for example, when a subsidence inversion associated with a subtropical high pressure cell confines moisture evaporated from the ocean within a thin, cool marine layer. At the same time, the radiative cooling associated with the clouds helps to maintain such inversions by lowering the temperature of the cloudy air. Turbulent entrainment, driven in part by radiative destabilisation, also maintains the inversion. Under suitable conditions, an external perturbation that reduces the SST favours a change in cloudiness that further reduces the SST (e.g., Hanson, 1991).

The positive feedback maintaining low clouds may be suppressed by various dynamical processes. Mesoscale circulations are forced in the marine boundary layer by the strong cloud-top radiative cooling that, in turn, breaks stratiform clouds into mesoscale cloud patches (Shao and Randall, 1996). Such broken clouds allow more solar radiation to warm the sea surface. Furthermore, temperature, moisture, and the subsidence rate of the subsiding air above the inversion may be changed to weaken the inversion (Siems *et al.*, 1990), thus increasing the likelihood of the stratiform clouds breaking up (Deardorff, 1980; Randall, 1980). Still controversial are details of the criteria and relative importance of other mechanisms, such as drizzle, absorbed sunlight, and entrainment decoupling that might detach planetary boundary layer (PBL) clouds from the surface (Kuo and Schubert, 1988; MacVean and Mason, 1990; Siems *et al.*, 1990).

Since subsiding air in the subtropics is connected to the sinking branch of a Hadley cell, subtropical boundary layer cloud properties may be partly determined by remote processes. A model with inadequate dynamical coupling to other regions may exaggerate the strength of the low cloud feedback and have a tendency to produce excessive low-level cloudiness. Coupled atmosphere-ocean models may be particularly susceptible because they have the ability to produce negative SST anomalies in response to increases in low-level cloudiness. On the other hand, over-emphasis of coupling to other regions may lead the feedback to another positive loop. Miller and Del Genio (1994) found in a version of the GISS GCM that the reduced low cloud amount led to stronger surface solar radiation, therefore a warming of the subtropical sea surface, and a further reduction in low cloud amount. The initial reduction of low cloud amount was due to the temperature changes at remote grid points, apparently weakening the inversion. They found this mechanism to give oscillations with periods on the order of a few decades. They are appropriately cautious about concluding that the feedbacks in the GISS model also operate in the real climate system, but suggest that positive cloud feedback could enhance variability.

Arctic clouds

Clouds are the dominant modulators of the Arctic radiation climate, affecting sea ice characteristics such as temperature, albedo and ice volume, (Curry and Ebert 1990, 1992), and so indirectly possibly altering the rate of sea ice transport to the North Atlantic. Sea ice and related high latitude physical processes including Arctic cloud formation are incorporated into current models with many unverified assumptions, so that the reliability of the simulated Arctic climate change scenarios is not high and the model-to-model differences are not surprising.

Ingram *et al.* (1989) used the UK Meteorological Office climate model to investigate the sea ice feedback on greenhouse warming. They performed sensitivity tests with prescribed, fixed sea ice distributions and compared their results with those of a simulation in which the sea ice distribution was permitted to change in response to climate change. They found that cloud-ice feedbacks were very important; the clouds obscure surface albedo exchanges, thus minimising their effects.

4.2.3 Cloud Water Content

As already noted, the climate models used for the first studies of global warming ignored the possibility of changes in cloud-water content, but are now beginning to include cloud formulations which explicitly predict the liquid- and ice-water content. The distinction between liquid-water and ice is important, not only for thermodynamic reasons but also because of differing radiative properties. The need to include formulations of cloud water and ice comes in part from the strong dependence of the short-wave, and for high clouds, long-wave radiative properties on water content and the possibility that the latter will change during global warming. Rennó *et al.* (1994) suggest that climate is sensitive to cloud microphysical processes and in particular, precipitation efficiencies.

Liquid water feedback might be substantially negative, as was emphasised by the simple radiative-convective model study carried out by Somerville and Remer (1984). Subsequently, a different sign for this feedback was inferred in a GCM by Roeckner *et al.* (1987) through a large increase in the long-wave greenhouse effect from thin cirrus clouds. The complexity of such cloud feedbacks is further illustrated by the studies of Li and Le Treut (1992) and Taylor and Ghan (1992).

In climate change studies with the UK Meteorological Office (UKMO) GCM, Senior and Mitchell (1993) found that in a warmer atmosphere, water clouds (lasting longer because of slower fall rates) replace ice clouds in the mixed phase region and hence cloud amount, especially at low and mid-levels, is increased. This "change of phase" feedback led to larger cloud amounts and to an increased short-wave cloud cooling with a warming climate. Consequently climate sensitivity was reduced with a 2.8°C warming in response to a doubling of CO_2, in comparison with an older relative humidity dependent cloud scheme where the global mean warming was 5.4°C. The inclusion of interactive cloud radiative properties further reduced the global mean warming to 1.9°C. In regions where cloud amount increased, the optical depth also increased. The net cloud feedback in this experiment was negative.

Changes in the long-wave cloud feedback depend on the representation of clouds in the model (Senior and Mitchell, 1995). A sensitivity experiment, in which the assumed statistical distribution of cloud water in a grid box is changed, produced a slightly improved validation of radiative fluxes against ERBE data (Barkstrom, 1984), but increased the climate sensitivity of the model from 3.4°C to 5.5°C (Senior and Mitchell, 1995). The effect of the change was to reduce high cloud amount for a given cloud water content. The reduced high cloud amount led to a much smaller "change of phase" feedback and so a higher climate sensitivity.

Several of the GCMs incorporate new cloud microphysics parametrizations for stratiform clouds and consequently produce weak negative cloud feedbacks in the climate sensitivity experiments reported by Cess *et al.* (1996). For example, the new version of the GISS GCM (Del Genio *et al.*, 1995) contains a prognostic cloud water parametrization and incorporates interactive cloud optical properties. When subjected to globally uniform increases and decreases of SST, the new model's climate sensitivity is only about half that of the earlier GISS model, and its cloud feedback is slightly negative as opposed to the substantial positive cloud feedback in the earlier model. This change in the cloud feedback is largely a result of a dramatic increase in both the cloud cover and cloud water content of tropical cirrus anvil clouds in the warmer climate. This result is sensitive to the type of climate experiment conducted. Tests with an SST perturbation that reduces the tropical Pacific SST gradient and weakens the Walker circulation give a higher climate sensitivity (Del Genio *et al.*, 1995).

The crude nature of current parametrization of detrainment of cumulus ice and the sensitivity to this term as discussed above, suggest that this may be an important area for future work in cloud parametrization. Low-level clouds may not produce the large negative feedback that is characteristic of models with only temperature-dependent cloud water because increasing cloud water content in a warmer climate could be offset by a decreasing geometric thickness of these clouds. This is consistent with the behaviour of satellite-derived optical thickness in the ISCCP data set (Tselioudis *et al.*, 1992). Thus the sign of the cloud liquid-water feedback in the real climate system is still unknown. Further study will be needed to reach the goal to determine the overall sign and magnitude of the real world's cloud feedback, as cloud feedback varies considerably with cloud type and geography and presumably with time. Nevertheless, the results from the new models provide valuable new insight into the physical issues that must be confronted before this goal can be achieved.

4.2.4 Cloud Particle Size

Studies of the effects of possible changes in cloud particle size on global warming integrations are at an early stage, but mechanisms by which changes might occur have been suggested. Cloud drops are formed by condensation on submicron diameter hygroscopic aerosol particles, which are present throughout the troposphere. The concentrations of CCN are highly variable and are influenced by air pollution as well as by natural processes as addressed in Chapters 3 and 4 of IPCC (1994) and Sections 2.3 and 2.4 of this report. The number of nuclei available has a strong effect on the number of cloud particles formed, and this in turn affects both the cloud optical properties and the likelihood of precipitation. Changes in the CCN concentration might alter the number of cloud drops that can grow, and hence alter the mean radius for a given liquid-water path. Some parametrization are attempting to represent the effects of nuclei availability on cloud particle number density and cloud particle size.

Twomey *et al.* (1984) noted that pollution produced by burning fossil fuels consists not only of carbon dioxide (CO_2) but also sulphur dioxide (SO_2), the gaseous precursor of sulphate aerosols, and that cloud drop sizes may be sensitive to changes in sulphate aerosol concentrations. Increased pollution could therefore increase the number of these CCN, increase the number of cloud drops, and hence reduce the mean particle size (provided that the water content does not change). This would increase the cloud optical thickness and hence the cloud albedos, leading to a cooling influence on climate. Efforts to include effects of sulphate aerosols from pollution in climate model cloud parametrization have been reported by Ghan *et al.* (1993, 1995), Jones *et al.* (1994), Boucher *et al.* (1995) and Boucher and Lohmann (1995). Kiehl (1994) argues that the difference between continental and oceanic cloud droplet sizes needs to be accounted for in determining climate model cloud albedos.

Charlson *et al.* (1987) suggested that the primary source of sulphate aerosol and hence CCNs over the ocean involves biological organisms, so that much of the sulphate aerosol over the remote oceans comes not from pollution but from dimethylsulphide (DMS), excreted by marine phytoplankton (discussed further in Section 10.3.4). Charlson *et al.* discussed the possibility of a regulation of the climate system by such marine organisms. Attempts to establish the details of processes by which this regulation might occur have not been successful, and there is no observational evidence that DMS sources would change with climate change.

Relatively shallow clouds may develop drizzle. This drizzle production depends on cloud depth, cloud liquid-water content and CCN distribution, as well as on cloud dynamics and lifetime. Drizzle depletes the cloud liquid-water, and reduces the cloud reflectivity. Below the cloud the drizzle may evaporate and cool the air, thus possibly leading to a decoupling of the cloud from the air below. If future anthropogenic emissions of SO_2 increase, thus leading to more CCN, then cloud droplets may become more numerous and smaller. This will likely impede drizzle production and possibly lead to longer cloud lifetimes.

4.2.5 Model Feedback Intercomparisons

The intercomparison reported by Cess *et al.* (1990) of climate models' response to changing the SST by $\pm 2°C$ provided a snapshot of the feedbacks operating in GCMs at that time. The models agreed in the magnitudes of the clear-sky feedbacks, as noted earlier, but cloud feedbacks varied considerably and were responsible for a threefold variation in the overall climate sensitivity between the participating models. This exercise has recently been repeated by Cess *et al.* (1996), who find that the disparity between the models has been reduced significantly (Figure 4.2). The most notable change is the removal of the largest (positive) values of cloud feedback. A detailed analysis of the reasons for this convergence has not yet been made. A small, overall cloud feedback can result from a cancellation of much larger feedbacks of opposite signs in the long-wave and short-wave regions of the spectrum. These components of the overall feedback vary considerably between the models. In addition, several modelling centres have produced a wide range of cloud feedbacks from slightly different cloud parametrizations in the same model. Hence, the convergence found by Cess *et al.* (1996) may not represent a reduction in the uncertainty of the magnitude of the cloud feedback. The idealised SST perturbation applied in these comparisons is very different from the more complex patterns obtained in coupled model simulations of global warming, and consequently the cloud feedback produced by a model forced by the ±2°C SST perturbation might be completely different from that found in global warming simulations (Senior and Mitchell, 1993). These intercomparisons may provide only limited guidance as to the cloud feedback to be expected during greenhouse warming.

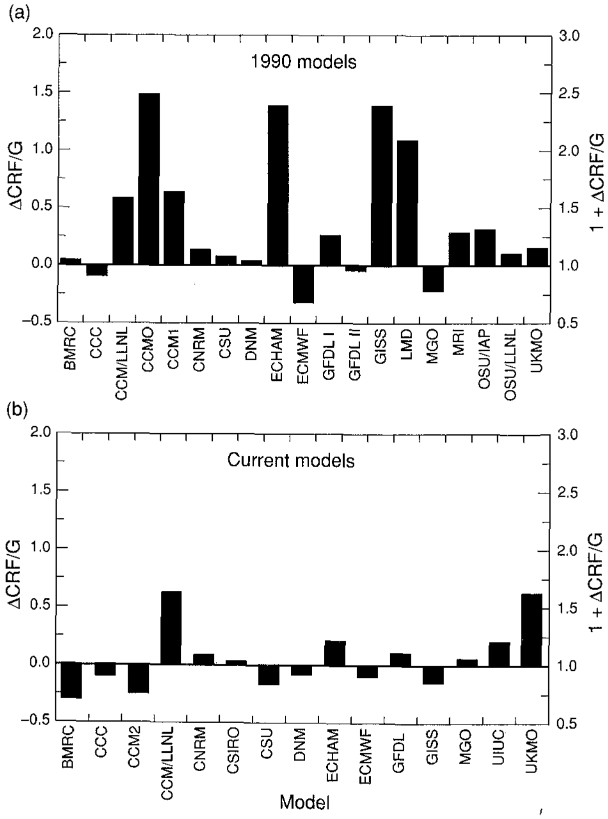

Figure 4.2: (a) The cloud feedback parameter, ΔCRF/G, as produced by the 19 atmospheric GCMs used in the Cess *et al.* (1990) study, where ΔCRF is the Wm^{-2} due to cloud changes and G is the overall Wm^{-2} change, both as a result of the prescribed SST change. (b) The same as (a) but for the Cess *et al.* (1996) study.

A further intercomparison that gives insights into the impact of differences between cloud simulations in models is provided by Gleckler *et al.* (1995). They analysed the ocean energy transports implied by the ensemble of GCMs participating in the Atmospheric Model Intercomparison Project (AMIP) (Gates, 1992). The models were run for ten simulated years, using prescribed seasonally and interannually varying SST and sea ice distributions, as observed for the years 1979 to 1988. The models calculated the net radiation at the top of the atmosphere and the net energy flux across the Earth's surface.

The pattern of ten-year-averaged net radiation at the top of the atmosphere implies a pattern of total energy transport inside the system, since there is a net energy input in some parts of the world, and a net energy output in other parts, giving a global total very close to zero. This energy transport is accomplished by the circulation of the atmosphere and of the oceans. A pattern of ocean meridional energy transports is implied by the ten-year averages of the net ocean surface energy flux for each atmospheric GCM.

The implied ocean energy transports T_O are represented by the thin lines in the upper panel of Figure 4.3. The grey stippling shows the range of observationally derived upper and lower bounds of T_O (*cf.* Trenberth and Solomon, 1994, for the most recent observational study). Most of the simulations imply ocean energy transports that differ markedly from those inferred from observations, particularly in the Southern Hemisphere, where the implied T_O for many of the models is towards the equator.

Gleckler *et al.* (1995) determined the total meridional energy transport by the atmosphere and ocean combined, denoted by T_{A+O}, from the simulated ten-year averages of the top-of-the-atmosphere net radiation for each model. They then determined the simulated atmospheric energy transport, T_A by subtracting the ocean transport from the sum of the ocean and atmosphere transport. Finally, they computed a "hybrid" value of T_O, denoted by $T_{O\ hybrid}$, by subtracting each model's simulated T_A from the ERBE-observed T_{A+O} . This "hybrid" combines the simulated T_A with the observed T_{A+O} . The results for $T_{O\ hybrid}$ are shown in the lower panel of Figure 4.3. On the whole, the various curves for $T_{O\ hybrid}$ bear a much closer resemblance to the observations than do the model curves, indicating that the simulated atmospheric meridional energy transports are relatively realistic in most cases. Evidently, improved cloudiness parametrization and improved simulations of the effects of clouds on the radiation budget are needed to improve oceanic forcing in coupled atmosphere-ocean models.

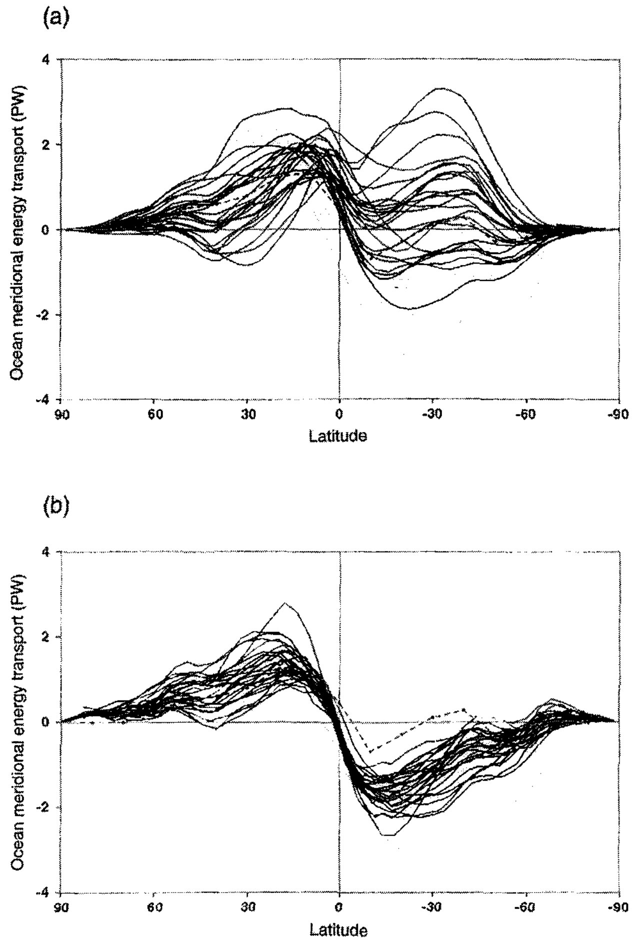

Figure 4.3: Ocean meridional energy transport from models and observations (a) model results from a range of atmosphere GCMs, derived from the ten-year averages of their implied net surface energy flux (thin lines); bounds on observed ocean transport (shaded area); model results from Semtner and Chervin (1992) in a numerical simulation of the general circulation of the oceans, forced with the observed atmospheric climate (dashed line). (b) As in (a), except that the thin lines show the "hybrid"ocean transport (see text).

4.2.6 Coupling of Clouds with the Surface

The effects of clouds on net solar radiation at the top of the atmosphere are largely mirrored in the effects of clouds on surface solar radiative fluxes. The uncertainties in the input of solar radiation to ocean and land models are a major source of uncertainties in determining the regional and global response to increasing greenhouse gases. Long-wave radiation would appear to be a smaller source of error, except in high latitudes, because of the atmosphere's large opacity to long-wave, even without clouds. In any case, the surface solar fluxes are more easily estimated from remote sensing since long-wave fluxes depend on cloud bases which are not easily seen from space.

Clouds are sensitive to the changes of both atmospheric circulation and the surface boundary conditions. Small changes in the surface boundary layer can cause substantial differences in convective clouds over the tropical oceans (e.g., Fu *et al.*, 1994). However, a change of large-scale circulation can modify atmospheric conditional instability and thus clouds, even without a change at the surface (Lau *et al.*, 1994; Fu *et al.*, 1996). Because the prediction of future cloud changes depends on the correctness of these responses in GCM cloud schemes, more stringent tests, using satellite and *in situ* observations, are needed to ensure that the observed sensitivities of clouds to the changes of atmospheric circulation and surface conditions are adequately simulated in GCMs.

Consideration of ocean surface-atmosphere interactions have led to a controversial hypothesis. Figure 4.4 schematically illustrates what is known as the Thermostat Hypothesis (Ramanathan and Collins, 1991). If a positive SST perturbation leads to an increase in surface evaporation and moisture convergence (Lindzen and Nigam, 1987) then the increased moisture supply induces more convection, which leads to the formation of more high, bright clouds, which reflect more solar energy back to space. The resulting reduction in the solar radiation absorbed at the sea surface thus acts to dampen the postulated positive SST perturbation. The initial perturbation might be the climatological differences between east and west Pacific, or warming of the east Pacific associated with El Niño, or overall SST increases associated with greenhouse warming. Similar mechanisms would not necessarily apply to all these situations.

To support their idea, Ramanathan and Collins presented observational evidence that SST fluctuations associated with El Niño are accompanied by changes in the solar cloud radiative forcing (CRF) that would tend to dampen the SST fluctuations regionally. Where the ocean warms, the solar radiation reaching the sea surface diminishes, and where the ocean cools, the increased solar radiation tends to warm it. Ramanathan and Collins argued that convection and high bright clouds increase when the SST increases to about 30°C. They suggested that the increased solar cloud forcing associated with deep convection might act to prevent much higher SSTs.

Although Ramanathan and Collins explicitly discussed only regional climatological effects, their paper has been widely interpreted as suggesting that the global surface temperature of the Earth may also be limited in this way.

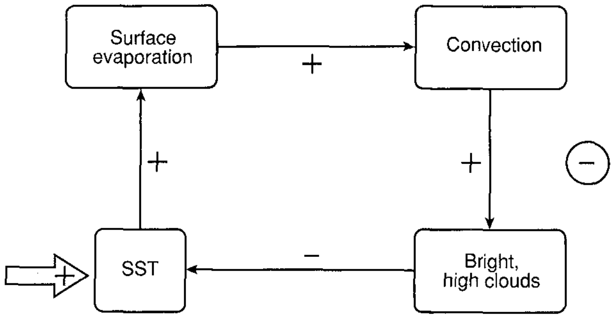

Figure 4.4: The "Thermostat Hypothesis" of Ramanathan and Collins (1991). An external perturbation leads to an increase in the SST, either locally or globally. The signs in the diagram are "+" for amplification of the next term in the loop and "−" for reduction. Increased SST promotes stronger evaporation and moisture convergence, which then lead to more vigorous convection. The convection generates high, bright clouds, which reduce the insolation of the ocean, thus counteracting the external perturbation. This is, therefore, a negative feedback and indicated by the circled "−".

This Thermostat Hypothesis has been very controversial and remains an active research topic. It has been criticised by Wallace (1992), Hartmann and Michaelson (1993) and Lau *et al.* (1994) for failure to recognise the importance of regional effects associated with large-scale dynamics, and also for under-emphasising the tendency of surface evaporation to cool the oceans. Several critical papers have emphasised a more conventional view of the tropical energy balance (e.g., Pierrehumbert, 1995). Fu *et al.* (1992) have argued on the basis of satellite data that the strong regional cloud radiative forcing anomalies associated with El Niño average to near zero over the tropics as a whole. The reality of locally negative short-wave cloud radiative forcing anomalies in response to local positive SST anomalies in the central tropical Pacific is apparent, but the importance of such short-wave cloud radiative forcing anomalies relative to other processes, and also their importance for the globally averaged surface temperature, are still in dispute.

Miller and Del Genio (1994) found, with their version of the GISS GCM, that a negative tropical evaporation anomaly resulted in a warming of the SST, leading to enhanced convection and rainfall. This convection decreased the solar radiation incident on the sea surface, and, not unlike the Thermostat Effect envisioned by Ramanathan and Collins (1991), dampened the initial warming of the sea surface. Spectral analysis of the model results showed that this mechanism led to oscillations with periods on the order of years up to a decade.

Three recent papers argue that clouds absorb much more solar radiation than current physical understanding and radiation codes would allow. Suggestions that some clouds absorb more short-wave radiation than can be accounted for on the basis of the known radiative properties of water and ice have appeared at various times in the literature over several decades (see the review by Stephens and Tsay, 1990). This possibility has been invoked by Ramanathan *et al.* (1995), in the context of the West Pacific warm pool. According to their results, the effect of clouds on short-wave radiation at the surface needs to be 50% more than that at the top of the atmosphere, whereas current radiation schemes predict an enhancement of less than 20%. The implication is that the clouds are absorbing several times more short-wave radiation than previously believed. Ramanathan *et al.* do accept that, by making "extreme (but plausible)" changes to their numbers, it is possible to close the heat budget without recourse to anomalous absorption. However, the companion paper by Cess *et al.* (1995) claims that the anomalous absorption is a global phenomenon, showing evidence from a wide variety of locations to support the 50% enhancement required by Ramanathan *et al.* This enhancement was also obtained by Pilewskie and Valero (1995) in an analysis of observations from research aircraft in the tropics.

A global enhancement of the magnitude proposed by Ramanathan is in conflict with many other documented studies with research aircraft where the measured absorption was not substantially different from the theoretically predicted value (Stephens and Tsay, 1990). Additionally, a critical examination of the anomalous absorption papers has revealed major flaws in the analysis methods which appear to invalidate the conclusions (Stephens, 1995). Hayasaka *et al.* (1995) have concluded "that the anomalous absorption pointed out by aircraft observations in previous studies does not exist". An extensive survey of surface and satellite data (Li *et al.*, 1995a; Whitlock *et al.*, 1995) suggests that short-wave budgets are unlikely to be in error by more than 15 Wm^{-2}. Li *et al.* (1995b) carried out extensive analyses of data "following the same methodologies" as Cess *et al.* and Ramanathan *et al.* They do not find anomalous absorption except possibly in the tropics where the data are most uncertain. The disagreement with Cess *et al.* is attributed to the use of different data sets, not due to different methodologies. Thus, at present, the evidence is weak for the claim that clouds absorb substantially more short-wave radiation than is predicted by models.

4.2.7 Precipitation and Cumulus Convection

Cumulus convection provides very rapid mass, energy, and momentum exchanges between the lower and upper troposphere. Much of the precipitation that falls to Earth is produced during this convective overturning, and a substantial fraction of the cloudiness in the tropics is produced by cumulus convection, either directly in the cumulus clouds themselves, or indirectly in the cirrus and other debris that cumuli generate. The proper treatment of these processes in climate models is still far from established. Errors in treating momentum exchange may seriously affect surface winds in the tropics and hence coupling to ocean models.

Cumulus convection is a manifestation of a buoyancy-driven instability that occurs when the vertical decrease of temperature is sufficiently rapid (i.e., when the "lapse rate" of temperature is sufficiently strong) and, at the same time, sufficient moisture is available. Because of the latter condition, cumulus instability is often called "conditional instability". The degree to which buoyancy forces can drive cumulus convection thus depends on both the lapse rate and the humidity. The time-scale for convective release is on the order of an hour – very short, compared to the multi-day time-scale of large-scale weather systems. This disparity of time-scales implies that ensembles of cumulus clouds must stay nearly in balance with large-scale weather systems. If a large-scale motion system or surface heating tries to promote cumulus instability, convection releases the instability restoring the system to a near-neutral state almost as rapidly as the instability is generated.

"Large-scale precipitation" refers to a somewhat old-fashioned but still widely used parametrization forming stratiform clouds such as cirrus or stratus and the accompanying precipitation that occurs when the mean state relative humidity reaches or tries to exceed a threshold value, such as 100%. Although generally larger than cumulus clouds, these systems are still typically sub-grid scale in climate models. Some such schemes accordingly include a sub-grid scale distribution of humidities, so that precipitation occurs with mean state relative humidities <100%. Relative humidities exceeding the threshold can be produced, for example, by large-scale rising motion which leads to adiabatic cooling and a decrease of the saturation mixing ratio. The excess humidity is typically assumed to condense and fall out as precipitation. In prognostic schemes, it may also be stored as liquid water. In many models, the falling precipitation is permitted to evaporate or partially evaporate on the way down.

Precipitation and convection are coupled to atmospheric radiative cooling in several ways. Slingo and Slingo (1988) discuss a positive feedback between the horizontal gradients of atmospheric radiative warming/cooling associated with localised high clouds produced by deep convection and the large-scale rising motion associated with the convection. In convectively active regions, long-wave radiation is trapped by anvils and cirrus produced by convective detrainment, and so the long-wave radiative cooling of the atmospheric column is reduced, and may even be transformed into a heating. The convectively active column is consequently radiatively warmed relative to the surrounding, convectively inactive regions, reinforcing the latent heating. The combination of these two heatings, together with the radiative cooling in the surrounding radiatively inactive regions, amplifies, on the average, the rising motion in the convectively active column.

Evidently, the strengths of the cloud feedbacks on precipitation must be further quantified. They do not occur in isolation, but coexist not only with each other, but also with many other powerful processes that can affect weather and climate. Idealised numerical experiments with GCMs can be designed to focus on such feedbacks in relative isolation, and so are particularly well suited to investigating their relative strengths.

The distribution of modelled precipitation and its changes with climate change needs more extensive validation as it is a major coupling link to both the land hydrological cycle and oceanic buoyancy forcing of the thermohaline circulation. High latitude precipitation is especially important for the latter.

4.2.8 Assessment of the Status of Moist Processes in Climate Models

In the previous IPCC report the radiative feedbacks of clouds were identified as a major source of uncertainty for modelling future climate change. Considerable research efforts have addressed this issue since the last assessment and have further reinforced this conclusion. They have also added considerably to our understanding of the complexity of this issue. Some conclusions from these studies are:

- Different cloud parametrizations in current GCMs give a wide range of radiative feedbacks, affecting global and regional energy balances and the occurrence and intensities of atmospheric precipitation. These are derived from plausible physical assumptions and parametrizations, but the issue is extremely complex and many assumptions or approximations are made. Models including only cloud amount feedbacks have indicated that these could amplify global warming. Feedbacks involving cloud liquid-water and phase could also have major impacts on the global energy balance. At present, it is not possible to judge even the sign of the sum of all cloud process feedbacks as they affect greenhouse warming, but it is assessed that they are unlikely either to be very negative or to lead to much more than a doubling of the response that would occur in their absence. Improved treatments are vigorously being pursued.

- The cloud feedback processes are intimately linked to the atmospheric hydrological cycle, and can only be simulated satisfactorily if there is a comprehensive treatment of water in all its phases – vapour, liquid and ice. Many of these cloud and linked hydrological processes occur on scales not resolved by current GCMs. The sub-grid scale parametrizations treating these processes should be physically based and carefully evaluated with observational data.

- Inadequate simulation of cloud amounts and optical properties in GCMs contributes major errors to the simulation of surface net radiation, and thereby introduces errors in simulation of regional ocean and land temperatures. Uncertainties in the simulation of changes in these properties with climate change have a major impact on confidence in projections of future regional climate change.

- There is an important, but poorly understood, linkage of cloud optical properties to the CCN distribution. Inclusion of this linkage for models of climate change will require an improved description of the time and spatially varying distribution of global sulphate and other aerosols, as well as detailed microphysical treatments of cloud droplet size distributions.

- There is a consensus among different GCMs as to the sign and magnitude of clear-sky feedbacks but not for water vapour feedbacks alone. With these clear-sky feedbacks but with fixed cloud properties these GCMs would all report climate sensitivities in the range 2-3°C. There is no compelling evidence that the water vapour feedback is anything but the positive feedback indicated by the models. However, the partitioning between water vapour and lapse rate feedback is not well established, and the processes maintaining water vapour in the upper troposphere are poorly understood.

4.3 Oceanic Processes

The ocean covers about 70% of the surface area of the Earth, has most of the thermal inertia of the atmosphere-ocean-land-ice system, is a major contributor to total planetary heat transport, and is the major source of atmospheric water vapour. Its interaction with the atmosphere through its surface quantities occurs through: SST, sea ice extent and thickness, surface albedo over ice-covered and ice-free regions, sea surface salinity and the partial pressure of CO_2 at the surface (pCO_2). It is a major component of the climate system in determining the mean (annually averaged) climate, the annual variations of climate, and climate variations on time-scales as long as millennia. While it is only through ocean surface variations that the atmosphere and land can be affected, these surface variations, in turn, depend on the thermal and saline coupling between the deeper ocean and the surface. Thus, the thermal and saline structure and variations in the deeper ocean must be simulated in order to determine surface variations on long time-scales. In general (except in those few regions of deep convection and other water mass transformation regions), the longer the time-scale of interest, the greater the depth of ocean that communicates with the surface. In turn, the ocean is driven by fluxes from the atmosphere of heat, momentum, and fresh water at the surface of the ocean, so that the only consistent way of simulating the evolution of the climate is through coupled atmosphere-ocean models.

The first atmosphere-ocean coupled models studying greenhouse warming concentrated on the sensitivity and response of climate to sudden and transient changes of radiatively active gases by using well-mixed (slab) oceans of fixed depth with no (or specified) transport of thermal energy. While such simplifications are useful for understanding atmospheric responses and for qualitative estimates of certain ocean responses, work over the last few years has indicated that the ocean circulation itself is sensitive to changes in forcing at the surface. Thus as the radiatively active gases increase, the ocean circulation may change and these changes may affect the mean climate and its variability. These changes may be significant, so that only coupled models that include the relevant parts of the ocean circulation are capable of simulating the entire range of possible climatic responses.

The ocean, like the atmosphere, has complex internal processes that must be parametrized. It has its own unique properties of boundaries and a density (buoyancy) structure that is affected by salt as well as temperature. Although venting of heat from fissures in the deep ocean may contribute to its circulation (Riser, 1995), to a good first approximation the ocean is driven entirely at the surface by the input of heat, fresh water and momentum fluxes from the atmosphere.

4.3.1 Surface Fluxes

Because the inertia (mechanical, thermal, and chemical) of the ocean is large compared to that of the atmosphere, changes in ocean surface properties for the most part occur relatively slowly. These slow changes depend on the surface fluxes of heat, momentum, fresh water, and CO_2 from the atmosphere. Models of the mean circulation of the ocean are sensitive to changes in heat flux (Maier-Reimer *et al.*, 1993) and fresh water flux (Mikolajewicz and Maier-Reimer, 1990; Weaver *et al.*, 1993).

The momentum input at the surface of the ocean, as wind stress, depends mostly on the winds near the surface and on the wave response. The heat input to the ocean surface consists of latent and sensible heat exchange between the ocean and the atmosphere, (depending on near surface winds, air temperature, and humidity), state of the sea, and radiative inputs (which depend on the overlying atmospheric column). The fresh water input to the surface of the ocean is composed of the difference between precipitation and evaporation, of runoff from land, and of the difference between the melting and freezing of ice (Schmitt, 1994). The carbon dioxide flux into the ocean depends on pCO_2, the concentration of CO_2 in the atmosphere, and on the near-surface winds. The pCO_2 is in turn controlled by oceanic transport processes, geochemical processes, and upper ocean biotic processes. The possibility of large errors in incident solar flux and latent heat flux are especially of concern for climate modelling; such errors may account for much of the flux adjustment needed for many models.

4.3.2 Processes of the Surface Mixed Layer

The near surface ocean is usually well-mixed by sub-grid scale processes involving stirring by the wind and by its convection. Therefore, quantities at the surface are determined by their mixed-layer values. The mixed-layer temperature is determined by heat fluxes at the surface, by mixing and advection of temperature horizontally, by the depth of the mixed layer, and by the entrainment heat flux at the bottom of the mixed layer (the interface between the near-surface turbulence and the relatively non-turbulent ocean interior). Skin effects at the atmospheric interface may give departures from mixed-layer values, especially under low winds. The interior ocean affects the surface

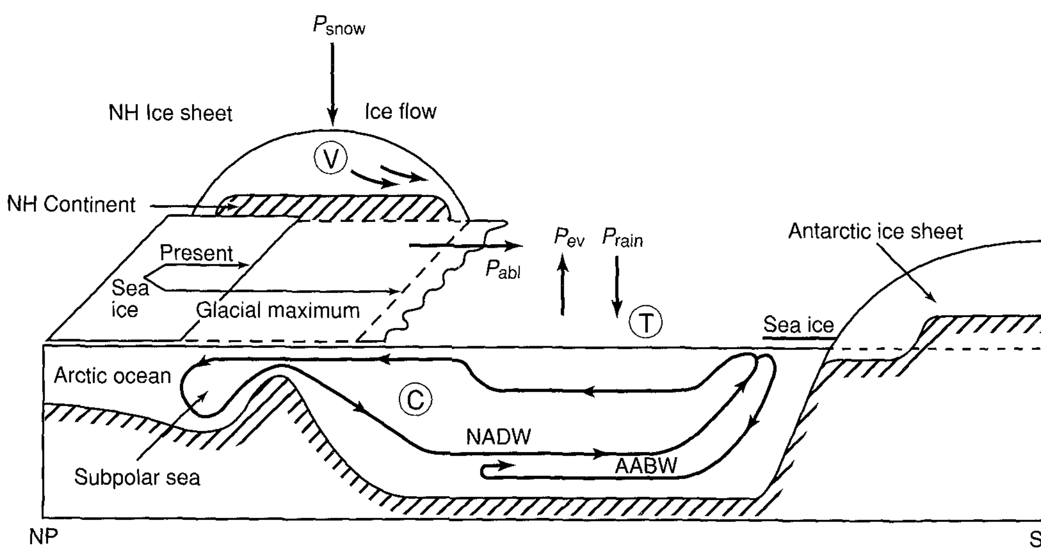

Figure 4.5: Diagram of an Atlantic meridional cross-section from North Pole (NP) to South Pole (SP), showing mechanisms likely to affect the thermohaline circulation on various time-scales. The change in hydrologic cycle, expressed in terms of water fluxes, $P_{rain} - P_{evaporation}$, for the ocean and water fluxes, $P_{snow} - P_{ablation}$, for the snow and ice, is due to changes in ocean temperature. Deep-water formation in the North Atlantic Subpolar Sea (North Atlantic Deep Water: NADW) is affected by changes in ice volume and extent (V), and regulates the intensity of the thermohaline circulation (C); changes in Antarctic Bottom Water (AABW) formation are neglected in this approximation. The thermohaline circulation affects the system's temperature (T) and is also affected by it (Ghil and McWilliams, 1994).

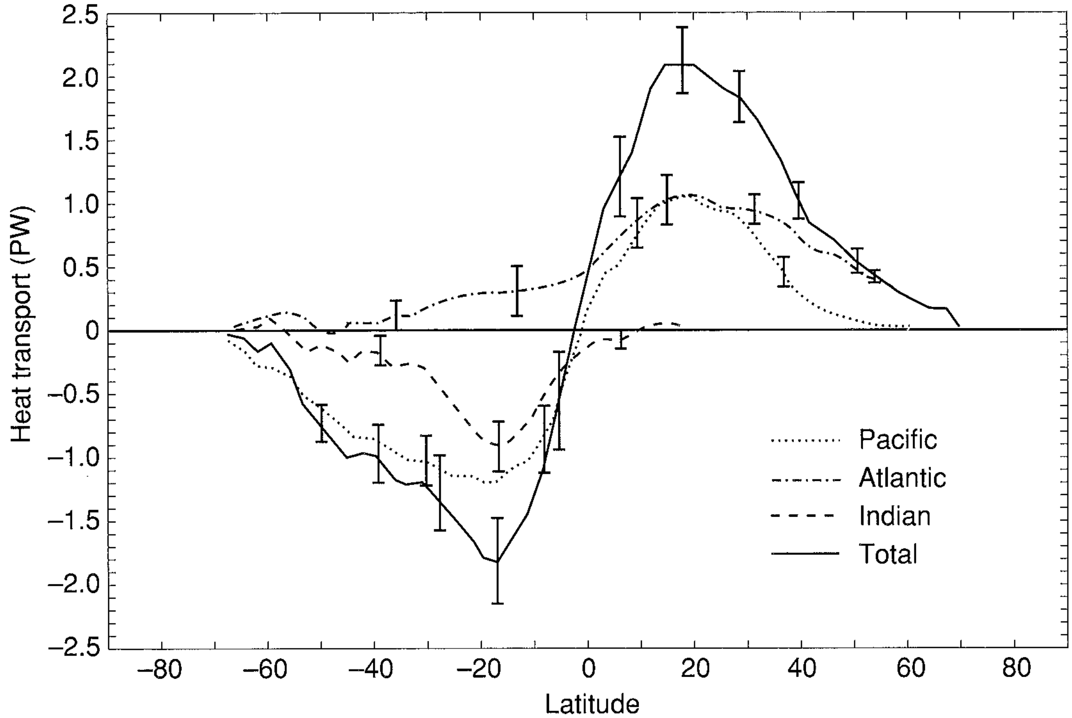

Figure 4.6: The poleward ocean heat transports in each ocean basin and summed over all oceans as calculated indirectly from energy balance requirements using ERBE for top of the atmosphere radiative fluxes and ECMWF data for atmospheric energy fluxes (from Trenberth and Solomon, 1994).

ocean only through this entrainment at the bottom of the mixed layer. Entrainment depends on the strength of the mixed-layer turbulence, on the motion of the bottom of the mixed layer itself, and on upwelling in the interior through the bottom of the mixed layer. Sterl and Kattenberg (1994) examine the effects on an ocean model of a mixed-layer parametrization and suggest that wind-stirring has important consequences not captured in the present ocean GCMs.

4.3.3 Wind Driven and Thermohaline Ocean Circulation

The wind driven circulation is that directly driven by the wind stress: because it is slow and large-scale in the interior of the ocean, it can conveniently be expressed by the conservation of vorticity (the so-called Sverdrup balance) which says that the vertically integrated meridional flow, where not affected by lateral boundaries and bottom topography, is given directly by the curl of the wind stress. This is an absolute constraint that would exist whatever the internal stratification of the ocean.

The thermohaline circulation is driven by changes in sea water density arising from changes in temperature versus salinity. Its functioning in the Atlantic Ocean is illustrated schematically in Figure 4.5. Formation of sea ice increases the salinity of adjacent unfrozen water. At low temperatures and relatively high salinity, cold dense waters sink convectively and spread throughout the oceanic depths, thereby maintaining the stable vertical stratification. Warmer surface waters flow toward these sinking regions and are cooled along their journey by heat fluxes from the ocean to the atmosphere. The sinking regions are highly restricted in area: "deep waters" are formed only in the North Atlantic Greenland, Norwegian, Iceland, and Labrador seas. The world's "bottom waters" are formed only in restricted regions of the Southern Oceans near the coast of Antarctica in the Weddell and Ross Seas. Waters from these sinking regions spread at depth to fill the world's entire basin and thereby help maintain the vertical stratification even in oceans where no deep sinking exists. These processes and the thermohaline circulation may change on various time-scales.

In the North Atlantic the combination of warm surface water flowing northward and cold water flowing southward at depth gives a net northward heat transport which has been estimated by direct measurements to be about 1 PW at 24°N (see Bryden, 1993 and references therein) and verified by indirect methods to be about the same (Figure 4.6, from Trenberth and Solomon, 1994). Deep cold waters flow southward across the equator from the North Atlantic and thereby imply a northward heat flux even in the South Atlantic. The Pacific is less saline than the Atlantic, is bounded further south by Alaska, and has no deep water formation. Its transport is more nearly symmetrical about the equator and similarly has a magnitude at 24°N of about 1 PW (Bryden *et al.*, 1991 and Figure 4.6). The properties of the Atlantic and Pacific rapidly interact through the Antarctic Circumpolar Current.

A coupled simulation (Manabe and Stouffer, 1988) has shown that during periods of no thermohaline circulation, the high northern Atlantic would be ice-covered to south of Iceland and be much colder than now. Palaeo-records indicate that oscillations in the thermohaline circulation leading to warming and cooling at high latitudes prevailed during the last glacial interval (Bond *et al.*, 1993). If the thermohaline circulation were to weaken with the expected larger inputs of fresh water to high latitudes during global warming, the net effect would be to either weaken the warming in high latitudes and amplify it in lower latitudes, and/or to make the thermohaline circulation and, hence latitudinal temperature gradients, more variable as discussed in Section 4.3.9.

Ocean only models are useful for isolating the ocean processes that may be present in more comprehensive coupled models of the climate although such may be considerably altered when coupled to the atmosphere. The idea that the stability and variability of the thermohaline circulation depends on the relative strength of high latitude thermal to fresh water forcing was introduced in coarse resolution ocean GCMs by Weaver *et al.* (1991, 1993). A steady thermohaline circulation can only exist if the fresh water input by high latitude precipitation, runoff, and ice melt is balanced by the fresh water export by that same thermohaline circulation. As the high latitude fresh water flux is increased, the ability of the thermohaline circulation to remove the fresh water is limited and the thermohaline circulation may have multiple equilibrium solutions (Stommel, 1961; Bryan 1986; Marotzke, 1988) and large variability. Such variability on decadal (Weaver and Sarachik, 1991) to millennial (Winton and Sarachik, 1993, Winton, 1993) time-scales has been demonstrated in ocean models with relatively large fresh water stochastic forcing. While the imposed boundary conditions on temperature have come into question, the mechanism is physically plausible and may survive the transition to a responsive atmosphere. If so, the implications are considerable: in a warmer world with warmer high latitudes and a stronger hydrologic cycle the thermohaline circulation could become less stable and more variable.

Recent simulations of the coupled transient response to increases of radiatively active gases that reach twice the pre-industrial concentration of CO_2 (Manabe and Stouffer, 1994) have indicated that the thermohaline circulation first weakens and then returns. The experiments that reach four times the pre-industrial concentration of CO_2 (in 140 years) have the thermohaline circulation weaken and stay weak for up to 500 years. While this model has a severe flux correction which stabilises the thermohaline circulation, it raises the important question of the response of the thermohaline circulation to changes in the greenhouse forcing and the subsequent effects of this response on climate.

Very high resolution models are being used to explicitly examine small-scale orographic, topographic and eddy processes in the ocean (e.g., Semtner and Chervin 1988, 1992). These models are useful for studying relatively short-lived phenomena and have given impressive simulations of the wind driven annual variation of heat transports in the Atlantic (Bönning and Herrmann, 1994) but, because of the huge computational overhead, can only be run for relatively short periods of model time. Thus, they cannot yet be coupled to model atmospheres for use in climate simulations nor to examine the thermohaline circulation. Such simulations must still use coarser-resolution models with parametrized eddies.

4.3.4 Ocean Convection

Given the surface fluxes and a formulation of mixed-layer processes near the surface, ocean general circulation models have to solve the advective equations for temperature and salt in the presence of convective overturning and stable ocean mixing. Convection arises when the density stratification becomes unstable and when relatively salty water is cooled to low temperature and becomes so dense that a water column becomes unstable. Convection homogenises the column and allows overturning to occur (see Killworth (1983) for a complete review).

Convection occurs over scales of a few km and is therefore hard to observe (but see MEDOC, 1970). Detailed simulations of individual convective elements in neutral stratification show the process to be complex and rife with small-scale features (Jones and Marshall, 1993; Legg and Marshall, 1993). Nevertheless, it is important to represent this process in ocean models with resolution coarser than the scales on which convection is known to occur. Simple convective adjustment parametrization (e.g., Cox, 1984) has long been used with improvement developed over time (e.g., Yin and Sarachik, 1994).

Parametrization of deep convection based on the detailed simulations of convective elements cited above are under development but it is not yet known how this will affect the simulation of the thermohaline circulation.

4.3.5 Interior Ocean Mixing

Parametrization of the mixing of stably stratified water in the interior of the ocean is crucial for the simulation of ocean circulation. Traditionally, eddy mixing coefficients have been used with values of order $1cm^2/s$ for vertical diffusion and viscosity and order$10^7cm^2/s$ for horizontal diffusion. These diffusion values are sometimes derived from tracer experiments, but more often they are selected to ensure numerical stability of the simulation. Parcels tend to stay and move on surfaces of constant density, which are predominantly horizontal in the interior ocean, and only small values of cross-density diffusion are expected. A tracer experiment (Ledwell *et al.*, 1993) has recently indicated that the correct vertical diffusion coefficient for the ocean interior is closer to 0.1 cm^2/s, an order of magnitude smaller than often used. Vertical mixing in the regions of lateral boundary currents or perhaps sea mounts is likely to be larger. Large *et al.* (1995) review a new scheme for mixed-layer dynamics and for ocean vertical mixing.

Further insights into turbulent diffusion in the ocean depend on a detailed knowledge of the precise mechanisms of that mixing. Mesoscale eddies, for example, are plentiful in the ocean and their finer scales would imply enhanced mixing through the production of frontal type gradients in temperature and tracers. A recent parametrization (Gent and McWilliams, 1990; Danabasoglu *et al.*, 1994) of these eddies follows an approach pioneered in stratospheric tracer modelling; that is, the mixing is primarily accomplished by advection by the "residual circulation" induced by the eddies rather than by the direct effects of the eddies themselves. Experiments with this parametrization show striking improvements in simulation of the depth and sharpness of the global thermocline and of the meridional heat transport.

4.3.6 Sea Ice

In high latitudes, sea ice is a major modulator of energy exchange between ocean and atmosphere. It is an insulating and highly reflecting surface. In the Arctic, much of the ice lasts permanently through the summer so that it has a substantially larger impact on surface albedos per unit area than do continental snow surfaces.

In winter, sea ice controls the transfer of heat from the relatively warm ocean to the cold atmosphere. The sea ice-cover is normally not complete. Even in the Central Arctic, 1-2% of open water permanently exists in the winter and a larger fraction in summer. Observation and model simulations indicate that surface sensible heat fluxes from the open water are one to two orders of magnitude larger than from the surface of the pack ice in the Arctic Basin (e.g., Meleshko *et al.*, 1991). Albedo of the sea ice-covered region also depends on sea ice concentration. Given the significant low-frequency variability of sea ice-cover and the persistence of its anomalies, its variable control of heat fluxes may significantly affect the atmospheric circulation and climate of the mid-latitude regions.

In the Northern Hemisphere, sea ice reaches its minimum extent in September when it covers $8.5 \times 10^6\,km^2$. It attains a maximum extent of $15 \times 10^6\,km^2$ in March. Its interannual variability varies from $1.1 \times 10^6\,km^2$ in winter to $1.8 \times 10^6\,km^2$ in summer, mainly in the marginal zones, and depending on atmospheric circulation and oceanic currents (Parkinson and Cavalieri, 1989).

Many GCMs and other climate models still treat the sea ice-cover as a single slab and do not take into account the always present but randomly distributed open waters ("leads"). Climate simulations with GCMs that incorporate sea ice inhomogeneities in a single grid box (sea ice concentration), show additional and substantial heating of the atmosphere in winter (Kattsov *et al.*, 1993; Groetzner *et al.*, 1994). This heating amounts to 10 Wm^{-2} over the polar cap of 60°N–90°N, increasing surface air temperature by 3.2°C over the same region. The warming is confined to the lower troposphere by the high stability of the polar atmosphere. The leads have a comparable heating effect to that produced by the observed sea ice anomalies (Kattsov *et al.*, 1993).

The largest surface air temperature increase with greenhouse warming is expected in the high latitudes of the Northern Hemisphere, because of the large atmospheric stability and the positive feedbacks between sea ice albedo and the surface temperature of the mixed ocean layer. These feedbacks are inadequately characterised, in part because of uncertain cloud feedbacks at the ice margins. Climate models that do not account for open waters over the ice-covered ocean probably overestimate the effect of the sea ice albedo in summer and underestimate the ocean cooling in winter. Changing snow cover may also feed back on ice thickness (Ledley, 1993).

The distribution of leads and thickness of sea ice is complex, of small scale and depends substantially on the dynamics of the ice as forced by wind and water drag. The required drag from the atmospheric model is, furthermore, not always of the correct strength and direction. Flato and Hibler (1992) have proposed a practical method for sea ice dynamics that is now being used in several GCMs.

4.3.7 The El Niño-Southern Oscillation as a Climate Process

The El Niño-Southern Oscillation (ENSO) is the coupled atmosphere-ocean phenomenon (Figure 4.7) wherein the normally cool, dry, eastern and central tropical Pacific becomes warmer, wetter, and with a lower sea level pressure every few years. The entire global tropics warm by about 1°C, and, by virtue of the large fraction of global area covered by the tropics, affect the globally averaged surface temperature on seasonal-to-interannual time-scales (e.g., Yulaeva and Wallace, 1994). The ENSO cycle affects the distribution and concentration of global CO_2 (e.g., Feely *et al.*, 1987 and Chapter 10). Major progress in understanding, simulating, and predicting ENSO has been made in the last 10 years (e.g., Battisti and Sarachik, 1995).

A number of recent papers have shown that:

- The recent land temperature record over the last few decades can be modelled by an atmospheric GCM forced by the observed record of global SST (Kumar *et al.*, 1994; Graham, 1995), and,
- These results are dominated by tropical rather than mid-latitude SST (Graham *et al.*, 1994; Lau and Nath, 1994; Smith, 1994).

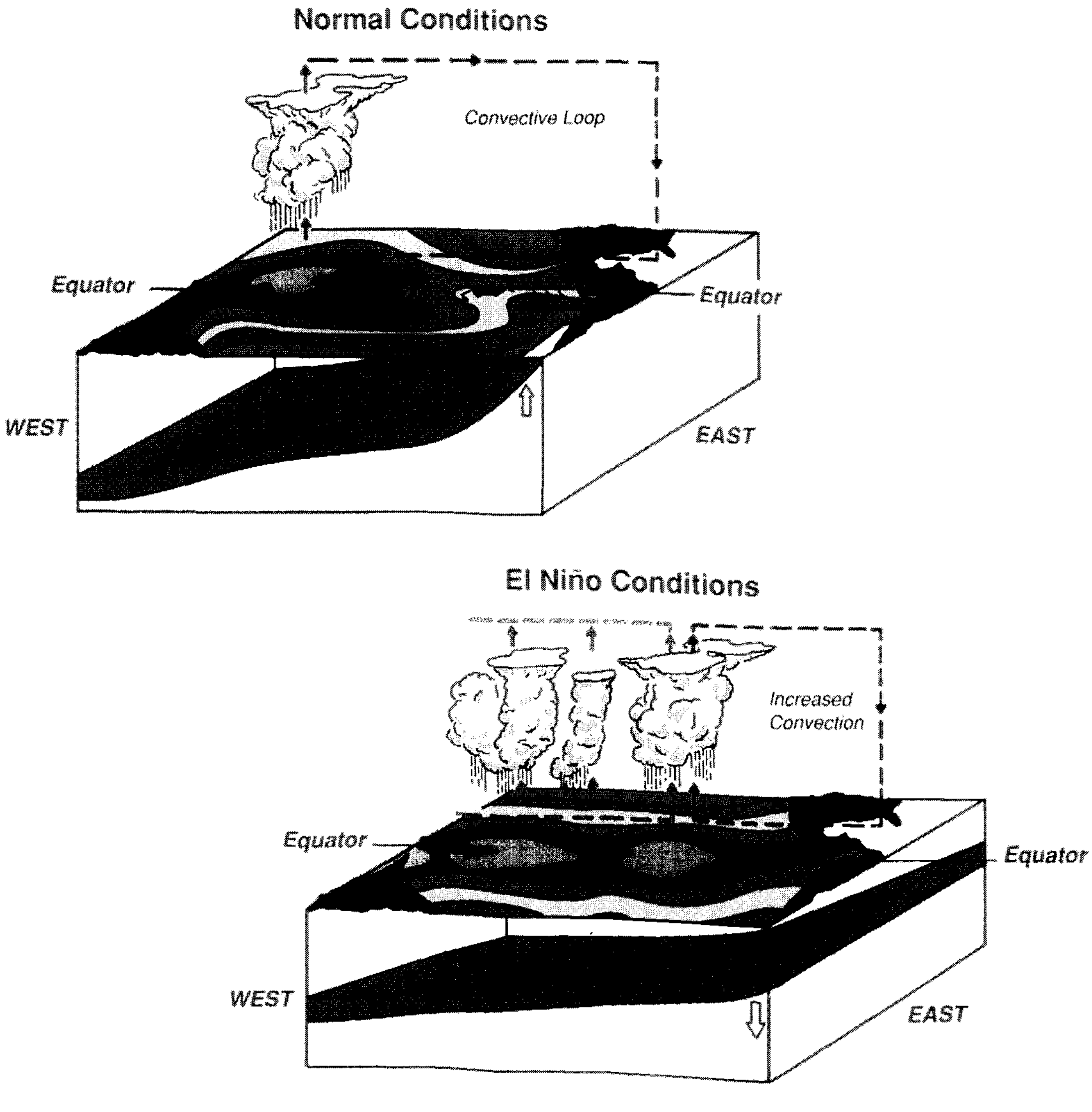

Figure 4.7: Schematic illustration of the differences in tropical climate between normal and El Niño conditions. For the latter, the thermocline becomes less tilted, SSTs increase in the eastern Pacific and regions in the central and eastern Pacific see increased convection.

These studies indicate that the unusual warmth of the tropical SST since the mid 1970s, during the warm phase of ENSO, has imprinted an unusual warmth on the entire global circulation. ENSO is evidently a major contributor to the natural variability of SST in the tropics, and arguably, also globally.

Warm phases of ENSO coincide with a warming of the tropical troposphere of close to 1°C. The net effect of ENSO is to warm the surface waters of the eastern Pacific. Ocean models that try to simulate the tropical Pacific SST in response to a repeating annual cycle of surface fluxes tend to simulate too cold an eastern Pacific. The simulation where the ocean is forced by long records of observed fluxes is more difficult and has not yet been done, though it is likely that such more realistic forcing would ameliorate the cold eastern Pacific problem.

Interestingly, all coupled model simulations (Mechoso *et al.*, 1995) of eastern Pacific SST show a result which contrasts with the ocean-only simulations: an eastern Pacific that is too warm, a feature attributed to the inadequacy of stratus cloud simulations in the atmospheric component of the coupled model (Koberle and Philander, 1994). To the extent that tropical Pacific temperatures matter for global climate and to the extent that ENSO variability dominates SST and land temperatures in the tropics (Wallace, 1995), it is clear that ENSO must be considered a vital part of the global climate system and should be accurately simulated. In order to correctly simulate ENSO, the meridional resolution at the equator must be a fraction of a degree in order to simulate wave processes and the meridional extent of the upwelling, both crucial. To date, no coupled model used for projecting the response to greenhouse warming has such resolution.

There is one final and intriguing possibility that the above cited papers imply: the possibility that global warming not only affects ENSO by affecting the background state (Graham *et al*, 1995; and a contrary view by Knutson and Manabe, 1994, in a coarse resolution GCM) but that indeed much of the effects of greenhouse warming might be modulated through changes in the magnitude and regularity of the warm and cold phases of ENSO.

4.3.8 *Assessment of the Status of Ocean Processes in Climate Models*

- Comprehensive ocean GCMs, coupled to the atmosphere through fluxes of energy, momentum, and fresh water, are required for assessment of the rate, magnitude and regional distribution of climate change.

- The large-scale dynamics of current ocean models seem reasonably realistic, but are not completely validated, in part because of a dearth of appropriate observations. An outstanding question to be resolved is the response of the thermohaline circulation in coupled models to increased high latitude inputs of fresh water. This question needs to be answered to assess how latitudinal gradients of SST may respond to global warming. Current suggestions point to either a large increase in or a highly variable latitudinal temperature gradient as possible responses.

- Fluxes at the ocean-atmosphere interface in coupled models have not yet been fully examined. In some cases there may be serious errors – for example, surface radiative fluxes depending on inadequately parametrized cloud processes, and high latitude inputs of fresh water depending on poorly characterised changes in the atmosphere-hydrological cycle.

- Oceanic models still use relatively crude parametrizations of sub-grid scale processes for near surface and interior mixing and for deep convection. New parametrizations for interior mixing associated with mesoscale eddies are likely to improve the simulated depth and sharpness of the global thermocline.

- Details of sea ice treatments in GCMs are still questionable, although some improvements are being examined. The role of sea ice in climate change is especially uncertain because of poorly known interface feedbacks; that is, overlying clouds modifying radiation, surface wind stress, ocean currents and changes in oceanic heat transport underneath the sea ice.

- ENSO processes have major effects on the tropical climate system, with a strong impact on hydrological processes and surface temperatures on interannual time-scales. Some coupled atmosphere-ocean models appear to give reasonable simulations of this system and show promise for providing useful predictions. However, the current generation of models used for projection of greenhouse gas response do not satisfactorily simulate ENSO processes, in part because the spatial resolution required to do so is not computationally feasible for century-long climate simulations.

4.4 Land-surface processes

Fluxes of heat and moisture between land and the atmosphere are central to the role of land processes in the climate system. These fluxes determine the overlying distributions of atmospheric temperature, water vapour, precipitation, and cloud properties. Atmospheric inputs of precipitation and net radiative heating are crucial for determining land-surface climate (climate over land is of greatest practical importance) and in turn are modified by land process feedbacks. Solar fluxes at the surface are currently highlighted as being significantly in error compared to observations, in some and perhaps most climate models due to the inadequate treatments of clouds (e.g., Garratt, 1994; Ward, 1995). These comparisons are being made possible by the recent availability of satellite-derived surface solar fluxes (e.g., Pinker *et al.*, 1995; Whitlock *et al.*, 1995).

Compared to the ocean, the land's relatively low heat capacity and limited capacity for water storage lead to strong diurnal variations in surface conditions and direct local responses to radiative and precipitation inputs. These limited storage capacities combined with the heterogeneous nature of the underlying soils, vegetation, and slope (e.g., Figure 4.8) imply potentially large heterogeneities in sensible and latent fluxes which may drive mesoscale atmospheric effects.

4.4.1 Soil-Vegetation-Atmosphere Transfer Schemes

Sensitivity studies with GCMs have shown that the treatment of land in climate models has major effects on the model climate and especially near the land surface (e.g., Koster and Suarez, 1994). The schemes to represent land in climate models are called soil-vegetation-atmosphere transfer schemes (SVATs). Important elements of these schemes are their storage reservoirs and their mechanisms for the exchange with the atmosphere of water and thermal energy. Water storage occurs in soil reservoirs and in some models also as fast time-scale canopy or surface terms. The water intercepted by the canopy may be stored or evaporated. This store generally holds about 1 mm of water. The soil store, usually in the range of 50 to 500 mm of water, depends on soil porosity, wilting point for vegetation, soil drainage rates and especially the depth from which water can be extracted from the soil. This depth is associated with the rooting depth of vegetation and the root distribution.

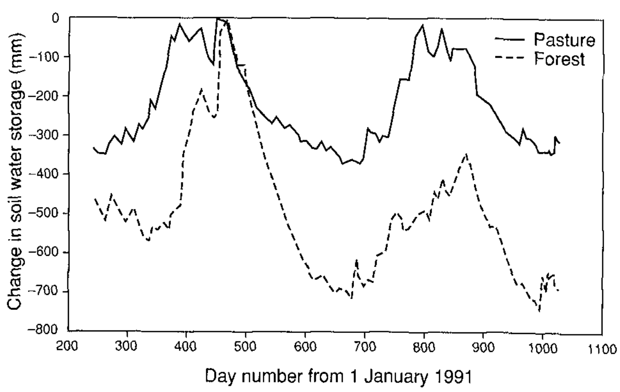

Figure 4.8: The seasonal variation in soil moisture storage for a tropical forest in Rondonia, Brazil (dashed line), compared to an adjacent pasture (solid line) (Institute of Hydrology, 1994).

Canopy transpiration is a physiological process depending on water transfer from the soil through roots, stems and leaves. The canopy resistance measures the effectiveness of this moisture transfer. It is primarily the integrated stomatal resistances. The transpiration, as mediated by the stomates, is limited by the supply of water from the roots and atmospheric conditions of demand. Neglect of this canopy resistance, now included in the SVAT models, has perhaps been the largest source of error in the older "bucket" models. The importance of canopy resistance is illustrated by two independent studies with GCM simulations of the effect of doubling stomatal resistance within the model SVATs (Henderson-Sellers *et al.*, 1995; Pollard and Thompson, 1995). The computed effects are largest for forests, and hence the largest areas affected are the boreal and tropical forests. In the boreal forests, summertime evapotranspiration (ET) is reduced by at least 20%, and surface air temperature increased by up to several degrees.

Surface roughness is the basis for determining the aerodynamic drag coefficient, C_D , for a surface. In the early GCMs, C_D for land was specified as 0.003, a typical value for short vegetation and for conditions of neutral stability. To achieve adequate accuracy, it is necessary to represent drag coefficients in terms of surface similarity theory, where transfer coefficients for momentum, heat and moisture are determined from a roughness length, Z_0 and the thermal stability of the near surface air. It may be necessary to distinguish between coefficients for momentum, heat and moisture. In particular, all sub-grid scale roughness elements and topography may contribute to momentum transfer but it is likely that only those on the scale of individual vegetation elements contribute to heat and moisture transfer. Schmid and Bünzli (1995) suggest a new approach for scaling roughness elements to model resolution and emphasise the importance of surface texture. The largest departures in newer models from the earlier ones are over forests, where C_D can readily exceed 0.01.

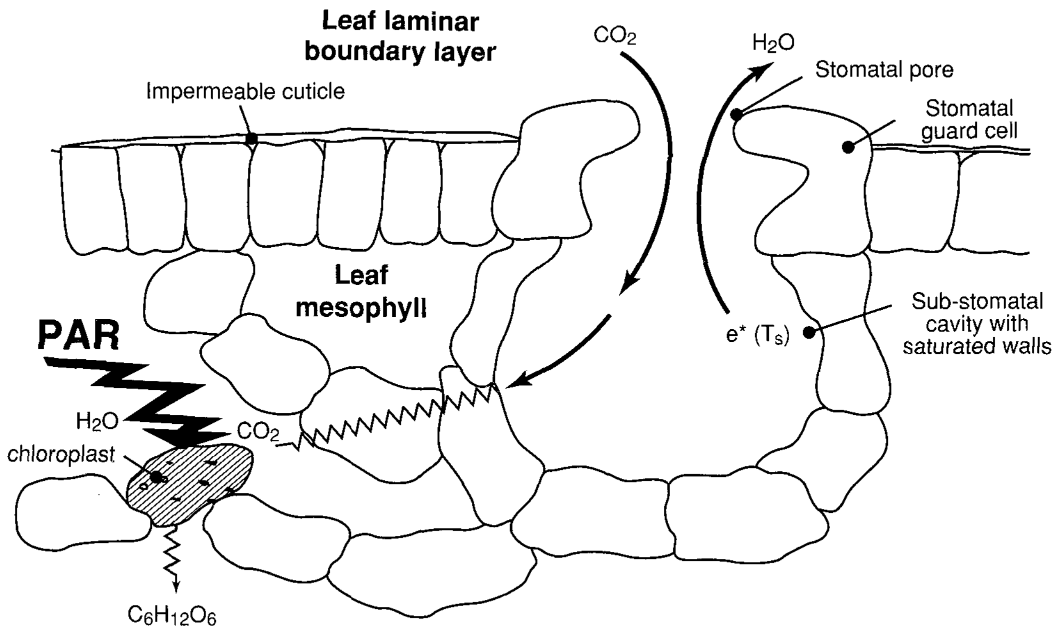

Figure 4.9: Schematic illustration of a leaf cross-section showing links between stomatal gas exchange (CO_2, H_2O) and photosynthesis. The stomatal conductance is related to the area-averaged value of the stomatal pore width which is of the order of 10 μm. The stomatal pores are under active physiological control and appear to act so as to maximise the influx of CO_2 for photosynthesis for a minimum loss of leaf water. Thus photosynthesis and transpiration are dependent on PAR flux, atmospheric CO_2 concentration, humidity, temperature, and soil moisture (Sellers et al., 1992). e*(Ts) is the saturation vapour pressure at the leaf surface.

Parametrization of vegetation properties related to canopy architecture determines significant features of the treatment of vegetation for evapotranspiration. The total surface of photosynthesising leaves and stem surfaces influences canopy resistance and transfer of heat and moisture from the canopy to the atmosphere. The flux of photosynthetically active radiation (PAR) normal to leaf surfaces, as required for stomatal parametrization, depends on canopy and leaf architecture. Furthermore, the net radiative loading over the surface of a given canopy element depends on these properties. Because of the large effect that canopy resistance has on SVAT models, they are sensitive to such details.

Recently, some canopy and soil schemes have included the uptake, storage and release of carbon through carbon dioxide exchanges with the atmosphere as illustrated in Figure 4.9. These sub-schemes will be increasingly important as physical models are coupled to biogeochemical models (*cf.* Chapters 2 and 9).

Vegetation cover and properties are now, for the most part, included in climate models as prescribed from inadequate observations. However, interactions between vegetation and climate may have significant effects. On seasonal time-scales, such interaction includes the effects of drought on vegetation cover, and on longer time-scales, possible changes in structure, e.g., transition between forest and grassland.

Canopy albedo determines the fraction of incident solar radiation that is absorbed. Current model parametrizations of surface albedo are largely inferred from some limited surface measurements for various kinds of vegetation canopies. Satellites are, in principle, the only means of globally establishing surface albedos. Because albedo can change substantially with vegetation cover it has been a major parameter in studies of the response of regional climate to land-use change (Section 4.4.3).

Runoff depends on soil moisture, properties of the incident precipitation, and characteristics of soils and topography. A theoretical foundation exists for the local vertical infiltration of water in soil, given soil hydraulic properties. In reality, runoff rarely results from precipitation exceeding maximum infiltration. Rather, much more often lateral down-slope flows carry soil water to low regions where the water table reaches the surface and into streams. However, there are intrinsic difficulties in parametrizing slope effects to determine runoff in a climate model. Furthermore, soil properties are highly heterogeneous both horizontally and vertically so that specifying them as constants over a model grid square or in a soil column is questionable. Changes of soil hydraulic properties with depth may strongly affect runoff. The recent intercomparisons through the Project for the Intercomparison of Land-surface Parametrization Schemes (PILPS) (Shao *et al.*, 1995) indicate a wide range of runoff rates between different land models, leading to substantial differences in annual average evapotranspiration. The current lack of a physically based and adequately validated treatment for runoff may be the biggest single obstacle to achieving an SVAT adequate for climate modelling.

In cold regions and seasons, processes involving snow cover and soil freezing become important for surface energy and water balances. Depending on its depth, snow masks some part of the underlying surface (e.g., Robinson and Kukla, 1985 and Baker *et al.*, 1991). Atmospheric models provide snow to the surface in liquid water equivalent, depending on criteria for transition between rain and snow. The surface model must determine the snow's density, temperature, albedo, and spatial heterogeneities. Long-term data records of snow cover and other surface conditions allow validation of the parametrization of snow processes in climate models (e.g., Foster *et al.*, 1996; Yang *et al.*, 1996). New treatments of snow processes in climate models have been proposed by Loth *et al.* (1993) and Lynch-Stieglitz (1995).

Although our understanding of how to model land processes has advanced considerably, there has been a substantial lag in implementing this understanding in models of greenhouse gas response. Some models with detailed land process treatments are now being used for such studies, but it is not yet possible to assess how future climate projections are influenced by these treatments. Current intercomparisons of off-line models by PILPS indicate a considerable divergence of results between different land-surface models for the same prescribed forcing. Improved criteria for accuracy and validation will be needed before the current conceptual improvements of land-surface models can be translated into increased confidence in climate change projection.

4.4.2 Questions of Spatial Heterogeneity

One of the common criticisms of present treatments of land processes in climate models is their failure to include many of the essential aspects of sub-grid scale heterogeneity. Heterogeneity is manifested in the precipitation and radiative inputs and in modelling the land processes themselves. The issue of precipitation heterogeneity (Milly and Eagleson, 1988) has been addressed through a simple statistical model in several GCMs, as reviewed by Thomas and Henderson-Sellers (1991). In this approach, precipitation is assumed to occur over some fraction of a model grid square, within which the precipitation is assumed to have an exponential distribution of intensities. Simple runoff models having a non-linear dependence on precipitation are integrated across the distribution of precipitation intensities to provide a grid-square runoff. This approach has been generalised to interception (Shuttleworth, 1988). Pitman *et al.* (1990) have demonstrated, for prescribed atmospheric forcing, that the partitioning between ET and runoff in a land model can be very sensitive to the fractional area of precipitation. On the other hand, Dolman and Gregory (1992), who allow for atmospheric feedbacks with a 1-D model, find very little sensitivity of average ET to assumptions about the rainfall distribution, but that the partitioning between interception versus evaporation and hence short time-scale rates of ET, can vary widely with assumptions about the precipitation distribution.

All of the above approaches assume variability of precipitation but retain a homogeneous water storage. However spatial variability of storage can also have a substantial effect (Wood *et al.*, 1992). Entekhabi and Eagleson (1989) have developed separate statistical models for precipitation and soil moisture that have been tested in the GISS GCM (Johnson *et al.*, 1993). They find a large variation in ET and runoff depending on the assumed model. Eltahir and Bras (1993) have generalised Shuttleworth's approach to a statistical description of interception. In particular, they assume a statistical distribution of leaf water stores and find, for prescribed atmospheric forcing, that interception changes substantially but that little change from homogeneous conditions would be realised if the leaf water stores were assumed uniform.

Another issue is the inclusion of heterogeneities in land-surface cover (Avissar and Pielke, 1989; Seth *et al.*, 1994) and hence inferred parameters such as roughness. At least three scales need to be considered. On a very fine scale, canopy air interacts between different surfaces, and surface roughness lengths cannot be associated with individual surfaces. Koster and Suarez (1992) refer to this scale of heterogeneity as "mixture" and for a given roughness give a simple model for deriving a total canopy temperature and water vapour from individual elements. This type of heterogeneity needs to be provided as part of the overall land cover description.

On a somewhat coarser scale, surfaces independently interact with an overlying homogeneous atmosphere. Koster and Suarez (1992) refer to this as "mosaic", and Shuttleworth (1988) as "disordered" heterogeneity. This is the scale, for example, of typical agricultural fields or small stands of forest. Finally, at scales of at least a few tens of kilometres and certainly at the scales of GCM resolution, each surface has a different PBL overlaying it. Shuttleworth (1988) refers to these as "ordered" heterogeneity. Mesoscale circulations on this scale may substantially add to boundary layer fluxes (e.g., Pielke *et al.*, 1991). They may also modify processes of clouds and convection in ways not accounted for by grid box-mean information.

Another issue of heterogeneity is in the distribution of incident radiation. Sub-grid scale clouds, perhaps associated with precipitation, may be important. For example, if precipitation occurs over some fraction of a grid square and spatial variation of surface wetness is included, but radiation is assumed homogeneous, the estimated evaporation from the wet surfaces will be excessive. Besides clouds, surface slope can be a major cause of heterogeneity in the amounts of absorbed surface solar radiation (Avissar and Pielke, 1989), as well as determining further heterogeneities in clouds and precipitation. Barros and Lettenmaier (1994) review the role of orography in triggering clouds and precipitation.

Some aspects of heterogeneity can be treated with relatively straightforward approaches including how to determine an average over a wide range of surface types with different characteristics. Raupach and Finnigan (1995) have considered energetic constraints on areally averaged energy balances in heterogeneous regions. In some cases, particularly if surface characteristics do not differ strongly, surface parameters can be aggregated (Claussen, 1990; Blyth *et al.*, 1993). For instance, albedo can simply be linearly averaged, (when the underlying surface is otherwise fairly homogeneous) whereas for roughness length a more complex aggregation is necessary (Taylor, 1987; Mason, 1988; Claussen, 1990).

The calculation of regional surface fluxes is made difficult by the non-linear dependence between fluxes and driving mean gradients. For example, where parts of the area are snow-covered with surface temperatures held at the freezing point, the mean vertical temperature structure over the area may imply a downward heat flux but, because the transfer coefficients are larger in the snowfree, statically unstable part of the area, the actual mean heat flux can be upward. Moreover, if surface types strongly differ, "parameter aggregation" becomes unfeasible. For example, definition of an aggregated soil temperature diffusivity does not make sense if parts of the area consist of bare soil, where heat conduction is diffusive, and of open water, where heat can be advected horizontally as well as vertically. In these more complex landscapes, "flux aggregation" is preferred (Avissar and Pielke, 1989; Claussen, 1991). Flux aggregation implies the computation of surface fluxes for each type in a grid box separately. Consequently, a regional surface flux is obtained by a linear average. Flux and parameter aggregation may be combined or use can be made of the intermediate approach of averaging exchange coefficients (Mahrt, 1987). The various aggregation methods require specification of a "blending height" where the aggregated information is matched to the overlying GCM grid squares (Claussen, 1990, 1991; Wood and Mason, 1991; Dolman, 1992; Blyth *et al.*, 1993). For ordered heterogeneity, this concept is less useful since the blending height would be above the surface layer. Practical application of these averaging procedures to global models will require high resolution global data sets on vegetation properties from future satellite sensors.

The aggregation or averaging methods may effectively treat the prescribed distribution of surface heterogeneities, but are not readily generalised to the dynamic interactions with sub-grid scale atmospheric inputs of water and radiation. Alternatively, a model grid box can be subdivided into sub-elements with both distinct surface characteristics and distinct atmospheric inputs. The "tile" or "mosaic" scheme by Koster and Suarez (1992) emphasises representation of the land heterogeneity. Different approaches may be needed to efficiently treat the heterogeneous distribution of atmospheric inputs; these atmospheric inputs may provide the largest overall departure in results from that inferred for homogeneous conditions. The possible importance of mesoscale circulation effects as a function of the spatial scale of individual homogeneous element on surface and boundary layer fluxes still needs to be assessed, although some preliminary work has been done (e.g., Zeng and Pielke, 1995) to parametrize mesoscale and turbulent fluxes in the boundary layer over inhomogeneous surfaces.

4.4.3 Sensitivity to Land-use Changes

Processes at the land surface influence the atmosphere through fluxes of heat and moisture into the PBL. These in turn affect atmospheric stability and the occurrence of precipitation and cloud radiative effects. Betts *et al.* (1993, 1994) have shown a close coupling between errors in a model's surface parametrization and its PBL, clouds, and moist convection. Sensitivity studies with GCMs have looked at the question of the possible effect of land cover modifications. These studies have indicated major surface influences on the atmospheric hydrological processes. This should perhaps not be too surprising since variations in ocean surface latent heat fluxes driven by small changes in oceanic surface temperatures have long been known to have large regional climate effects. However, the feedbacks of land-surface processes to the atmosphere are more complex and still far from understood.

Recent studies of the sensitivity of the Amazon Basin climate to a change from forest to grassland have been published, e.g., by Nobre *et al.* (1991), Henderson-Sellers *et al.* (1993) and Lean and Rowntree (1993). The response of climate to deforestation in the Amazon has been found to be sensitive to the specification of surface properties such as albedo (Dirmeyer and Shukla, 1994) and surface roughness (Sud *et al.*, 1995). A contrasting question is the effect of changing a semi-arid grassland such as the Sahel in Africa to a desert (e.g., Xue and Shukla, 1993). Reductions in absorbed solar energy due to higher surface albedos reduce ET but it is not clear how the change in precipitation should be related to change in ET. Many of the simulations have shown precipitation to reduce substantially more than ET, which implies a reduction in the convergence of moisture from the ocean.

The sum of various regional climate changes from land-use may contribute an overall effect on global climate. Estimates of global radiative forcing depending on albedo change are usually small compared to energy flux change from greenhouse gases. However, effects of changing land surface on the ratio of sensible to latent fluxes and on precipitation are not accounted for in such estimates and may be significant not only regionally but globally.

4.4.4 Assessment of the Current Status of Land Processes in Climate Change Simulations

- Land processes, driven by incoming solar radiation and precipitation, play an important role in the determination of near-surface climate, surface temperature, soil moisture, etc., and hence regional climates. Biases and uncertainties in the surface energy balance, and radiation and water budgets, are a significant source of error in simulations of regional climate.

- Detailed treatments of land processes are now available, replacing the previously used bucket models, and are being incorporated into numerous climate models. These treatments are substantial conceptual improvements, but it is premature to judge how they will modify or improve our confidence in climate change simulations. A number of issues must be resolved before their inclusion in climate models may be viewed as satisfactory. These include: the reasons for the current divergences in answers between conceptually similar models in the PILPS model intercomparisons; the relative importance of various sub-grid scale processes, their interaction, and how to represent those that are important in the model; what level of observational detail is needed to prescribe the land properties needed by the models; and how these observations will be made available.

- Modelling of runoff has large uncertainty in global models, there are no convincing treatments of the scaling of the responsible processes over the many orders of magnitude involved, and in high latitudes of the effects of frozen soils. Global data on soils, topography, and water holding capacities at the relevant scales will be urgently needed to make progress on this issue.

- The question of the averaging of heterogeneous land surfaces has been clarified and reasonable approaches proposed.

- Future models should begin to address the issue of how land-surface characteristics might change with climate and CO_2 change and how important that is as a feedback. Sensitivity studies of the impact of doubling stomatal resistance indicates some climatic effects over forest regions comparable to those anticipated from global warming over the next century. How the integrated stomatal resistance for global vegetation might change with changing climate and CO_2 concentrations is largely unknown.

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