

DEVELOPMENT OF A RISK-HEDGING CO₂-EMISSION POLICY, PART I: RISKS OF UNRESTRAINED EMISSIONS

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Abstract. A rational global strategy with respect to greenhouse-gas emissions would seek to minimize total risk, which is the sum of the risk of negative impacts due to climatic change associated with a given level of emissions, and the risks associated with the process of achieving that emission level. Given the existence of reducible uncertainties in estimating these risks, and the possibility that an emission target thought to minimize total risk is later found to be not strict enough, a risk-hedging strategy is a more realistic policy objective. This paper is Part I of a two-part series in which these risks are reviewed and an interim risk-hedging emission level is proposed. Here, the risks associated with unrestrained greenhouse-gas emissions are reviewed. In particular, the carbon-cycle response to continuing CO₂ emissions; the heat trapping of projected greenhouse gas increases in comparison to other anthropogenic and natural heating or cooling perturbations; the climatic response to heating perturbations; and the impacts of projected climatic change on global agriculture, forests, coastal regions, coral reefs, water resources, terrestrial species, stratospheric and tropospheric ozone, and human comfort and welfare are critically examined. It is concluded that unrestrained emissions of greenhouse gases pose real and substantial risks to human societies and to ecosystems, and that these risks are likely to grow substantially if the climate warms beyond that associated with a CO₂ doubling. These risks clearly justify some action to limit emissions. The magnitude of emission restraint that is justified depends not only on the risks reviewed here, but also on the risks associated with measures to limit greenhouse-gas emissions, which are reviewed in Part II.

1. Introduction

The United Nations Framework Convention on Climate Change (UNFCCC), which came into force on 21 March, 1994, declares its ultimate goal to be, "stabilization of greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference with the climate system". The current state of scientific knowledge does not permit quantifying the concentrations at which greenhouse-gas (GHG) emissions should be stabilized, nor does it permit precise quantification of the extent to which emissions would have to be decreased in order to achieve stabilization at any given set of concentrations. The Framework Convention nevertheless establishes three broad objectives which are to be satisfied by whatever stabilization ceilings are adopted: (1) stabilization is to be achieved, "within a time frame sufficient to allow ecosystems to adapt naturally to climate change"; (2) it should "ensure that food production is not threatened"; and (3) it should "enable economic development to proceed in a sustainable manner". These conditions recognize that there are risks to natural ecosystems, food production, and national economies associated with climatic change resulting from unrestrained

emissions of GHG's, and that these risks increase with increasing GHG concentrations. There are, however, additional risks associated with the process of restraining or reducing GHG emissions, and these risks will increase the greater the constraint on GHG emissions.

A rational global strategy with respect to GHG emissions would seek to minimize the sum of these two risks. Some of the risk is due to present ignorance, and as knowledge improves, that portion of the risk will be replaced by knowledge of either certain (100% probability) negative impacts or avoidance of a given impact altogether as a result of a given emission trajectory. Other risk is due to essentially unpredictable internal variability in the climate and biological systems, or in socio-economic systems. Given perfect knowledge of the component probability distributions and covariances associated with a given climatic change one can, in principle, calculate the probability – or risk – of certain 'fatal' combinations of events happening together for a given emission trajectory (for example, a combination high temperature, low precipitation, and pest infestation leading to irreversible dieback of forests in a given region, or a combination of circumstances leading to the extinction of a given species). The same can be said for economic risks associated with CO₂-emission reduction initiatives. One could then determine an emission trajectory which minimizes the total risk, including risk associated with present but reducible ignorance, and the risk associated with irreducible uncertainties. If reducible uncertainties were to be eliminated, one would likely compute a somewhat different risk minimizing trajectory based on the remaining, irreducible uncertainties. That is, an emission trajectory which is risk-minimizing for one state of knowledge would not be risk-minimizing for a later state of knowledge.

That being the case, the preferred strategy is a risk-hedging strategy. This is a strategy which adopts what is likely to be the *minimum* required constraint on emissions associated with any present or future risk minimizing trajectory, while creating the conditions which will facilitate tougher constraints in the future, should this be required to minimize revised assessments of risks.

Complicating the development of either a risk-minimizing or a risk-hedging strategy is that fact that the two sets of risk referred to above – those associated with climatic change and those associated with efforts to limit greenhouse-gas emissions – are fundamentally different in nature. Nevertheless, policy decisions must and are being made in the present (including the decision to do nothing except business-as-usual), so it is important to have a clear understanding of both sets of risks. Because the risks cannot be accurately quantified, and in some cases are unlikely to ever be quantifiable accurately, policy decisions must ultimately rest on subjective judgement concerning the level of GHG emissions which represents the best risk-hedging strategy.

This paper reviews the risks associated with unrestrained emissions of GHG's, while the accompanying Part II (Harvey, 1996a) reviews risks associated with actions to reduce energy-related emissions of carbon dioxide, the primary GHG of concern. Assessment of the first set of risks requires an evaluation of the likely future

carbon-cycle response to continuing CO₂ emissions, of the climatic sensitivity to GHG increases, of the likelihood of positive biogeochemical-climate feedbacks, and of the vulnerability of natural ecosystems, agriculture, and human societies to the resultant climatic change. Assessment of the second set of risks requires evaluating the direct cost and potential of various emission-reduction options, and the hazards associated with alterations to the status quo in order to achieve emission levels that would not otherwise occur. Symmetry in the assessment of risks associated with actions to reduce GHG emissions requires consideration of the potential non-climatic *benefits* of emission reduction, which are therefore also briefly reviewed. Based on these reviews, it will be argued in Part II that an appropriate interim (20–30 year) risk-hedging strategy is one which seeks to stabilize global emissions of carbon dioxide. Climate and carbon-cycle models are used to compare the effect on climate of this strategy compared to typical business-as-usual scenarios as well as more stringent emission strategies.

A number of authors, beginning with Nordhaus (1992) and Peck and Teisberg (1992), have attempted to determine ‘optimal’ global CO₂-emission trajectories for the next century or longer. The term ‘optimal’ signifies a CO₂-emission trajectory such that the discounted value of total consumption summed to the end of the analysis period, and summed over all nations and groups within nations, is maximized. Consumption is assumed to be reduced both by efforts to reduce CO₂ emissions and by climatic change resulting from GHG emissions. The very act of discounting is highly contentious, and strongly influences the results, since even very large costs one or two centuries into the future carry negligible weight. The results of such exercises are also dependent on the assumed relationship between abatement effort and cost, and between climatic change and damage, which in turn is highly dependent on the value attached to non-market assets (such as functioning ecosystems or species) which are damaged or lost as a result of climatic change. Mabey et al. (1996, Chapter 3.1) provide an extensive discussion and quantitative analysis of the optimization approach and conclude that, “using numerical optimization as a prescriptive policy tool is fatally flawed, given current information and methodologies”, and further, “that there is no answer to the question of how much abatement is optimal which does not depend on our attitudes towards what type of environment is right and fair to bequeath to future generations”.

The results of most optimization exercises lead to recommended emission scenarios which do not come close to stabilizing atmospheric GHG concentrations, thereby violating the intent of the UNFCCC, and imply significant environmental losses for future generations, thereby violating the value system of most people (Kempton, 1991; Kempton and Craig, 1993). Political decisions, in reality, have been and likely will continue to be made based on a subjective weighting of a wide range of factors, including the risks of climatic change and of efforts to abate emissions, and the impact of abatement strategies on other policy objectives besides limiting climatic change. This two-part series is intended to facilitate the process of policy development with regard to climatic change by providing a comprehensive

discussion of the risks of both climatic change and of emission abatement efforts, and by suggesting an interim global risk-hedging emission trajectory which, it is hoped, will at the very least stimulate further discussion of what an appropriate near-term global emission target should be.

The risk-hedging framework seeks to minimize net global risk. In reality, potentially large differences in the risks, costs, and benefits of global warming and GHG emission reduction will likely occur both between and within nations. However, such problems can be resolved to some extent through side payments, as discussed by Sand (1991). Furthermore, it is the role of government to make decisions which reflect the interests of society as a whole, including the interests of future generations. For these reasons the problem of disaggregation will not be discussed further here.

2. Risks Associated with Unrestrained Greenhouse-Gas Emissions

To assess the risks associated with a given scenario of CO₂ and other GHG emissions requires (i) estimating the likely buildup of GHG concentrations in the atmosphere, (ii) computing the heating perturbation associated with the projected buildup of GHG's, and comparing it with other natural or anthropogenic perturbations over the time period of interest; (iii) estimating the climatic response to the greenhouse-gas-induced heating perturbation; and (iv) assessing the impact of the climatic change on ecosystems and human societies. Evidence and uncertainties associated with each of these steps are presented below.

2.1. CARBON-CYCLE RESPONSE

During the 1980's an average of 5.0–6.0 Gt (gigatons, or billions of tons) of carbon were released to the atmosphere every year as CO₂ due to the burning of fossil fuels, with an additional 1.6 ± 1.0 Gt C per year due to deforestation and other land use changes (Watson et al., 1992). Atmospheric CO₂ increased at the rate of 3.4 ± 0.2 Gt C per year during this time, which implies that 2–5 Gt C per year, or about one third to two thirds of total anthropogenic emissions, are quickly removed from the atmosphere at present. This carbon is thought to be removed primarily through absorption by the oceans and by an increase in the rate of global photosynthesis due to the stimulatory effect of the 25% increase in atmospheric CO₂ which has occurred since the industrial revolution. Recent analysis of oceanic and atmospheric data indicate that the likely rate of oceanic uptake is 2.0 ± 0.6 Gt C per year (Siegenthaler and Sarmiento, 1993). This implies that the terrestrial biosphere is absorbing an extra 0–4 Gt C per year. Inasmuch as the global rate of net photosynthesis is on the order of 50–60 Gt C per year, the stimulation required to balance the carbon budget is no more than 8%, which will be difficult to detect through direct measurement, particularly since part of the increased photosynthesis is likely to go into increased below-ground biomass production.

Carbon-cycle models can be adjusted to replicate the observed buildup of atmospheric CO₂ for a variety of assumptions concerning land-use emissions, the biosphere response to increasing atmospheric CO₂, and oceanic CO₂ uptake (Harvey, 1989; Wigley, 1993). However, in order to project atmospheric CO₂ concentrations associated with future CO₂ emissions, it is important to know the relative importance of the terrestrial biosphere and oceanic sinks today, and how each sink will respond in the future.

2.1.1. *Terrestrial Biosphere*

The biosphere response to CO₂ increases involves direct stimulation of photosynthesis and inhibition of respiration due to higher atmospheric CO₂ concentration, potential increases in the rate of respiration due to warmer temperatures, and feedbacks involving nutrient cycling through the plant–litter–soil system. An increase in photosynthesis implies a transfer of carbon from the atmosphere to the biosphere, thereby tending to decrease atmospheric CO₂ concentration and increasing carbon storage on land. An increase in respiration has the opposite effect. Woodwell (1990) raises the possibility that the net CO₂ flux from the biosphere to the atmosphere due to the net effect of warming and the atmospheric CO₂ increase could exceed direct human emissions, such that subsequent warming would be largely beyond human control, whereas Idso (1991) argues that CO₂ stimulation of photosynthesis provides a rigid upper limit to the possible atmospheric CO₂ concentration if fossil fuel emissions are held constant at the present level.

Higher temperatures will tend to increase plant maintenance respiration and respiration of litter and soil organic matter. High latitude regions are of concern due to the greater projected warming and large stores of soil carbon at high latitudes (Oechel et al., 1993), although respiration fluxes and the sensitivity of respiration to temperature are largest for tropical soils, which may therefore dominate the global response of soil respiration to warmer temperatures during the next century (Townsend et al., 1992). A number of factors could greatly reduce the loss of plant and soil carbon due to warmer temperatures. First, an increase in soil respiration will tend to increase the availability of nutrients, which alone would tend to increase photosynthesis rates (independently of any increase due to higher CO₂) and reduce the loss of soil organic matter. The net effect of temperature increases on terrestrial carbon storage, after accounting for possible nutrient-cycle feedbacks, could be small or even beneficial (Pastor and Post, 1988; McGuire et al., 1992). Simulations by Rastetter et al. (1992) indicate that, even in steady state, the increase in living biomass carbon more than offsets the decrease in soil carbon due to warming of temperature forests and tundra, although this result may depend on the extent to which the mobilized nutrients are lost from the ecosystem or taken up by soil micro-organisms. Second, a decrease in soil moisture under a warmer climate would tend to limit temperature-induced increases in the respiration of litter and soil organic matter. However, where soils are currently saturated, such as high latitude wet tundra, a lowering of the water table – which could result from thawing

of permafrost – would tend to increase respiration (Billings et al., 1982). Finally, higher CO₂ itself tends to directly inhibit plant respiration in at least some species (Drake, 1992), although opposite effects have also been observed (Ryan, 1991; Wullschlegel et al., 1994). An analysis by Jenkinson et al. (1991), using a model of soil organic carbon turnover, suggests that global warming at a rate of 0.3 °C/decade would cause soils to release about 1 Gt C/year over the next 60 years due to increased respiration, but none of the three possible mitigating effects discussed above were included in this result, nor the effect of a lower water table in regions of wet tundra.

Higher atmospheric CO₂ tends to increase plant photosynthesis directly by inhibiting photorespiration (in C₃ plants) and indirectly by increasing stomatal resistance to water loss, which allows greater productivity for a given water loss. The indirect effect is strongest in cases where soil water is limiting. The direct stimulatory effect on photosynthesis saturates at a CO₂ concentration of 1000 ppmv (about 4 times the pre-industrial concentration) or greater (Oechel and Strain, 1985), although concurrent temperature increases could delay the saturation point. As with temperature effects on respiration, a number of feedbacks involving nutrient cycling could significantly alter the direct or initial effects of higher CO₂. Zak et al. (1993) suggest, on the basis of field data from nutrient-poor soils, that increased plant productivity and associated substrate release from roots due to higher CO₂ increases soil microorganism biomass and rates of nutrient turnover, thereby amplifying the initial stimulation of plant photosynthesis in the short term. In contrast, Diaz et al. (1993) find that in nutrient-rich soils an increase in substrate release results in greater nutrient sequestration by an expanded soil microflora and a nutritional limitation of plant growth. Stimulation of photosynthesis is believed to increase the C : N ratio in plant litter, which in turn tends to reduce the availability of N for plants by reducing decomposition rates, thereby eventually reducing the photosynthetic response of plants to higher CO₂. However, a number of other factors, whose response to increasing CO₂ is unknown, also influence organic matter decomposition rates (Gifford, 1992). Nevertheless, it is possible that the effect of nutrient feedbacks could be of opposite sign between the short term and long term. If plants respond to an increase in the rate of photosynthesis by increasing the proportion of net productivity going to fine roots (which are short-lived) in order to maintain the same C : N ratio in plant tissues, then the increase in carbon storage due to higher rates of photosynthesis can be greatly reduced (Post et al., 1992) or completely eliminated (Norby et al., 1992).

The feedbacks discussed above are summarized in Figure 1, which shows the key pathways by which changes in temperature, atmospheric CO₂, and soil moisture lead to changes in total carbon at a given location. As discussed above and indicated in Figure 1, the terrestrial biosphere response to increasing CO₂ concentration and climatic change involves a number of competing feedbacks whose relative strengths could very well change over time. Thus, even if the biosphere is serving as a net sink of CO₂, as appears to be the case at present, there is no justification for extrapolating this sink indefinitely into the future or in assuming that it will grow with increasing

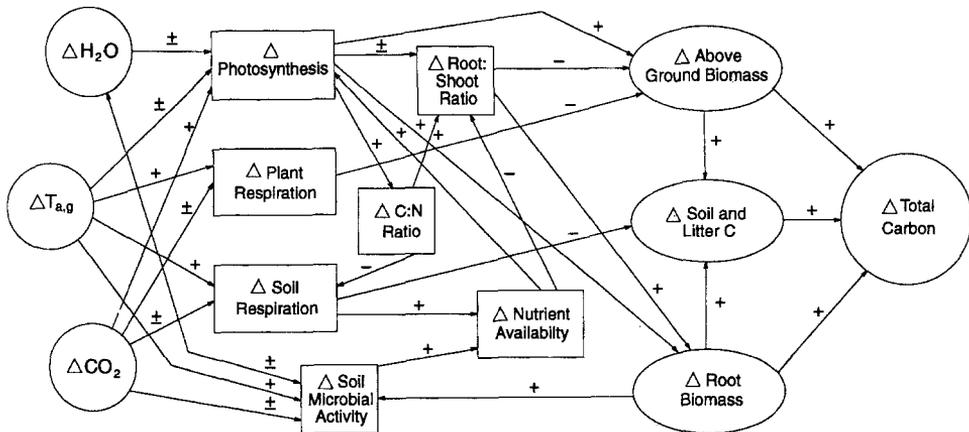


Figure 1. Feedback loops involving the effect of changes in temperature, atmospheric CO₂, and soil moisture on terrestrial carbon storage. Plus signs indicate that an increase in the upstream quantity leads to an increase in the downstream quantity, and vice versa for a negative sign. If the product of all the signs in a given loop is positive, the result is a positive feedback. Not all branches from a given box will necessarily be operative at the same time at a given site. For example, a change in nutrient availability could change the root to shoot ratio or the rate of photosynthesis, but not necessarily both. Not included in this diagram are further feedbacks arising from changes in the species composition at a given site, or lags between the dieback of current species and their replacement by new species.

atmospheric CO₂ concentration (see the exchange between Idso, 1991, and Harvey, 1991). Nutrient limitations in particular could lead to a significant weakening of the terrestrial sink over a period of several decades. Continued destruction of tropical rainforests, to the extent that a large fraction of the present terrestrial sink is in the tropics, will also limit the size of the future terrestrial sink.

A further potentially significant terrestrial biosphere–climate feedback will occur if and when climatic change becomes large enough to induce changes in the species composition at a given site. The lag between dieback of mal-adapted ecosystems and their replacement by new ecosystems as climate changes could lead to significant decadal-to-century time scale decreases in terrestrial carbon storage, even in cases where the net effect of CO₂ and climate changes would be to increase carbon storage once the species distribution is fully adjusted to the new climate (Solomon, 1986; Poster and Post, 1988). Calculations by Dixon et al. (1994) indicate that the average flux over the next century due to climate-induced forest dieback could be as high as 4.2 Gt C yr⁻¹, and would be further increased to as high as 6.0 Gt C yr⁻¹ if climate-induced expansion of agriculture into forested areas is included. However, potential CO₂-fertilization effects are excluded from these calculations. Nevertheless, it is clear that, with unrestrained GHG emissions and global warming, there is a growing risk that what is probably a net carbon sink at present could significantly weaken and possibly become a net carbon source. This may have already happened in some Arctic ecosystems (Oechel et al., 1993).

2.1.2. Oceanic CO₂ Sink

The absorption of atmospheric CO₂ by the oceans involves initial gaseous exchange between the atmosphere and surface layer of the oceans, chemical equilibration with other dissolved C species (namely, HCO₃⁻ and CO₃²⁻) and subsequent downward transfer of excess total dissolved carbon (TDC). Surface waters of the ocean have a low TDC concentration and hence a low CO₂ concentration compared to deep waters, and since the atmosphere tends to equilibrate with the surface water CO₂ partial pressure, this maintains a relatively low atmospheric CO₂ concentration.* The low surface-water TDC concentration is due to two factors: (1) the net uptake of carbon by photosynthetic organisms in the surface layer followed by their death, settling into the deep ocean, and release of carbon at depth, forming the so-called 'biological pump'; and (2) the ability of cold water in equilibrium with the atmosphere to hold more TDC than warm water. Since downwelling water is colder than upwelling water, this temperature difference acting alone causes a net downward transfer of TDC compared to the hypothetical case in which upwelling and downwelling temperatures are the same, and forms the so-called 'solubility pump'. These processes and diffusive mixing are illustrated in Figure 2. Because the surface water as a whole is depleted in TDC relative to deeper water, the downwelling water has a lower TDC than upwelling water, so the net effect of downwelling and upwelling (or ocean 'overturning') is to transfer carbon upward, but the upward transfer tends to be smaller than for the case of uniform temperature. Unlike terrestrial ecosystems, marine ecosystems in general are not limited by the availability of CO₂, so that marine photosynthesis rates in general will not be stimulated by an increase in CO₂ concentration in ocean surface water. Rather, the main factor limiting marine productivity is the availability of nutrients, which in turn is regulated by the rate of upwelling of nutrient-rich deep water.

The flows of carbon between the surface layer and deep ocean could be altered through either (i) the direct effect of a change in ocean overturning; (ii) the indirect effect of a change in ocean overturning by changing the supply of nutrients to the mixed layer and hence the strength of the biological pump; and (iii) a change in the temperature difference between upwelling and downwelling water. Coupled atmosphere–ocean climate models consistently project a weakening of the global scale overturning circulation in the ocean as the climate warms (Harvey, 1994), although local coastal upwelling could increase if continents warm faster than adjacent oceans, as expected (Bakun, 1993). Shaffer (1993) investigated the impact in a one-dimensional model of completely suppressing both the biological and solubility pumps, the former by turning off marine biology and the latter by imposing zero temperature difference between downwelling and upwelling water. Beginning with the pre-industrial atmospheric CO₂ concentration of 280 ppmv, suppression of the biological pump alone increases atmospheric CO₂ concentration to 460 ppmv,

* In particular, the mean CO₂ partial pressure of surface water is around 350 μ atm, whereas the partial pressure of deep water, if brought to the surface without chemical or temperature changes, would be 1500–2000 μ atm (Broecker and Peng, 1980).

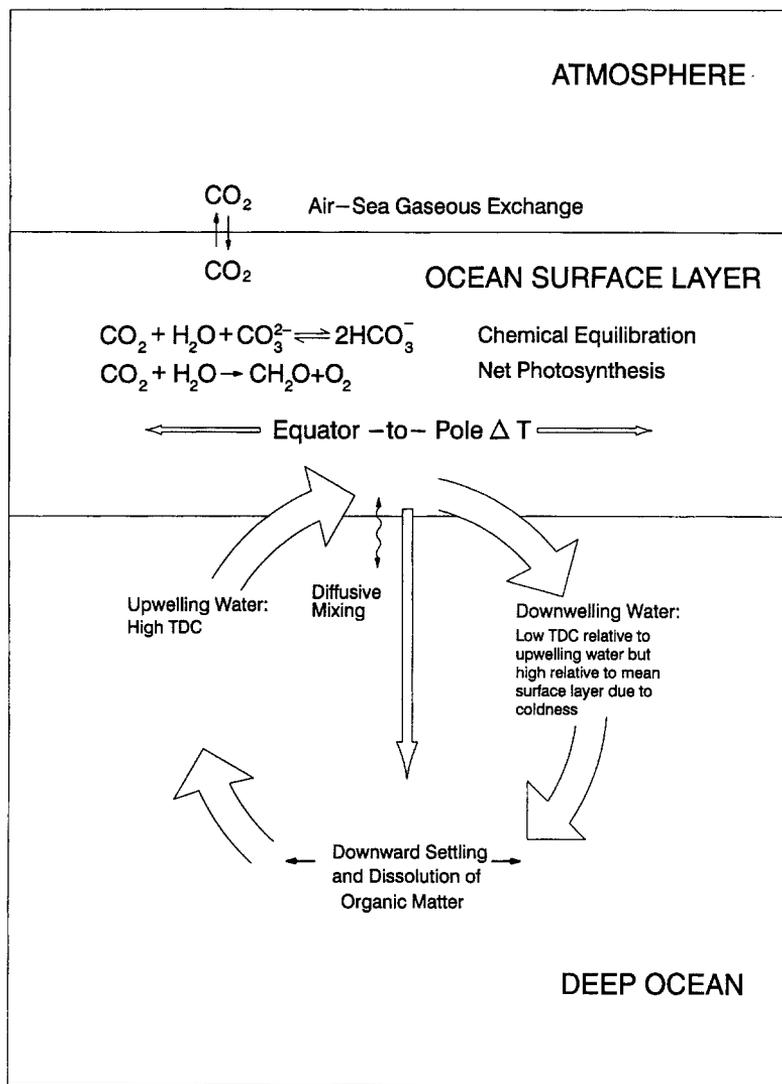


Figure 2. Processes involved in the flows of carbon between the atmosphere, ocean surface layer, and deep ocean. Ocean photosynthesis is, in general, not directly affected by higher atmospheric CO₂, but would be affected by climate-induced changes in the upwelling of nutrients.

while suppression of both pumps increases atmospheric CO₂ to 640 ppmv. Given that projected warming is expected to be greater at high latitudes, a weakening of the solubility pump can be expected. The current temperature difference between upwelling and downwelling water is about 20 °C, and if a 2 °C low latitude warming is accompanied by a 4 °C warming of high latitude downwelling regions, the solubility pump will be weakened by 10%. Based on Shaffer's (1993) result for complete suppression of the solubility pump, this will cause an atmospheric CO₂ increase of about 20 ppmv and constitutes a weak positive feedback.

A significant suppression of the marine biological pump due to climatic change and stratospheric ozone loss seems unlikely, given that reduced vigour by some micro-organisms implies smaller depletion of surface nutrients, which would likely stimulate greater productivity by other organisms more capable of adapting to the environmental changes. However, a significant weakening of oceanic overturning could significantly weaken the biological pump, but this would be at least partially compensated by reduced net upward transfer of carbon by the overturning circulation. This is supported by Bacastow and Maier-Reimer (1990), who find, using a 3-dimensional ocean circulation model, that a 40% reduction in overturning intensity with no change in temperatures reduces atmospheric CO₂ concentration by about 40 ppmv in steady state. In this case the reduction in the biological pump is more than compensated by the reduced net upward transfer of TDC as the overturning weakens. Experiments with a coupled atmosphere–ocean general circulation model by Manabe and Stouffer (1994) suggest that a CO₂ increase of about 250 ppmv could provoke a 40% reduction of overturning intensity, so the resultant 40 ppmv reduction in atmospheric CO₂ concentration represents a weak negative feedback.

Another potential feedback involves the dissolved organic carbon (DOC) pool in the ocean, whose size is estimated at between 480–1100 Gt C (Sundquist, 1985). Considerable uncertainty remains concerning the size of this pool, its turnover time, and the potential impact of changes in DOC turnover or respiration on the oceanic uptake of CO₂ (Mopper et al., 1991). In the extreme (and unlikely) case where half of the pool is oxidized due to warmer oceans, the eventual carbon transfer to the atmosphere would be the same as that which would remain if the same amount of carbon were directly injected into the atmosphere and then allowed to equilibrate with the ocean. This would amount to an atmospheric CO₂ increase of only 13–30 ppmv.

Other feedbacks, such as the gradual decrease in the ocean's ability to absorb CO₂ due to a decrease in CO₃²⁻ concentration as CO₂ enters the ocean, or the effect of ocean-surface warming, are also comparatively small and, in any case, are incorporated in current models. In conclusion, current understanding indicates that unexpected changes in the oceanic sink due to a variety of possible feedback processes are unlikely to be large, in contrast to the terrestrial carbon sink, where large changes could very well occur.

2.1.3. *Biogeochemical-Climate Feedbacks*

A number of potential feedbacks between climate and the carbon cycle have already been discussed. These feedbacks are such that initial anthropogenic GHG emissions may lead to climatic changes which lead to additional fluxes of CO₂ into (or from) the atmosphere, which would thereby amplify (or diminish) the initial GHG and temperature increases. Other potential biogeochemical-climate feedbacks involve CH₄ (methane) which, on a molecule-per-molecule basis, is 26 times stronger as a GHG than CO₂ (Lelieveld and Crutzen, 1992). Two potentially important

feedbacks involving CH₄ are (i) an increase in the CH₄ flux from wetlands; and (ii) destabilization of water–methane compounds, known as clathrates, which are found in terrestrial permafrost at depths of several hundred meters and in continental slope sediments worldwide. Higher temperatures could increase CH₄ emissions from wetlands both by enhancing metabolic rates (Whalen and Reeburgh, 1990) and by enhancing primary productivity (Whiting and Chanton, 1993), but a drop in the water table could lead to decreases in CH₄ production. To the extent that higher methane fluxes from wetlands are caused by higher primary productivity (and associated removal of CO₂ from the atmosphere), feedbacks involving CO₂ and CH₄ would be partially offsetting. Similarly, conditions which would maximize CO₂ emissions from high-latitude peatlands as climate warms (namely, a lower water table) would tend to minimize CH₄ emissions (Gorham, 1991), so that large positive feedbacks involving both CO₂ and CH₄ would not occur simultaneously.

Harvey and Huang (1995) have assessed both methane-wetland and methane-clathrate feedbacks for a range of anthropogenic emission scenarios and climate sensitivities. The impact of a hypothetical positive methane-wetland feedback is small (resulting in about a 5% increase in global mean temperature response), as is the potential methane–clathrate feedback except for cases of very high anthropogenic emissions coupled with high climate sensitivity, in which case the warming is augmented by up to 25% (depending on the details of methane release). However, these circumstances produce extremely large warming (5–6 °C global mean response) even without the additional methane feedbacks, so that the risk of negative impacts is already high and probably not substantially increased.

2.1.4. *Synthesis and Perspectives on the Carbon-Cycle Response*

Current knowledge indicates that the potential change in the ocean CO₂ sink due to circulation changes or warmer temperatures is likely to be small, and that the effect of potential methane–wetland and methane–clathrate feedbacks is also small. However, the present terrestrial carbon sink could significantly weaken with continuing fossil fuel emissions, and there is a substantial risk of a dramatic transient dieback of forests as climatic zones shift, giving a large CO₂ flux to the atmosphere for a century or longer. There will be a tendency for CO₂ and CH₄ feedbacks involving thawing of permafrost to be negatively correlated; that is, if one is comparatively strong, the other will be comparatively weak, and vice versa. Changes in the strength in ocean overturning also have partly compensating effects on atmospheric CO₂, by altering the rate of upwelling of nutrients and of CO₂-rich deep water. Terrestrial nutrient feedbacks appear to limit changes in carbon storage due to both temperature enhancement of respiration and CO₂ enhancement of photosynthesis. However, these conclusions must be tempered by the knowledge that the climate system is being moved by human intervention into a state which is unprecedented in recent geological history, and current models could be omitting key processes which, if included, might give a different CO₂ buildup in response to anthropogenic emissions.

2.2. HEAT TRAPPING DUE TO GREENHOUSE-GAS INCREASES AND OTHER FACTORS

The amount of heat trapped by a given set of GHG increases is the least uncertain of all the quantities required for the computation of climatic change, being largely dependent on laboratory-measured absorption properties of the various gases. The heat trapping is different for clear and cloudy skies, so that the mean trapping depends on the amount of cloud cover, and depends on the initial amounts and distribution of radiatively active gases, such as water vapor, which absorb in the same part of the electromagnetic spectrum as key GHG's such as CO₂ and CH₄. An analysis by Cess et al. (1993) indicates that, after allowing for known shortcomings in the radiative calculations of some models, a CO₂ doubling traps an extra 4.0–4.5 W m⁻² of heat energy. The GHG increases which have occurred since 1765 are trapping about 2.5 W m⁻² heat (Shine et al., 1990).

For the projected GHG increases to be a significant or dominant factor in future climate, it is necessary that the associated heating perturbation be large compared to other natural and anthropogenic heating or cooling tendencies. Likely natural sources of climate variation at the century time scale include variations in energy output from the sun and oscillations in the net heat flow between the deep ocean and surface. Direct observations of the sun indicate that the solar energy output varies by less than ±0.1% during the 11-year sunspot cycle (Hoffert et al., 1988), and observations of other stars suggest that variations of a few tenths of a percent could occur in the sun at an 80-year period (Baliunas and Jastrow, 1990). Reid (1991) claims that a 1% solar energy variation combined with the known greenhouse forcing of the past century gives a good fit to observed temperature variations assuming a CO₂ doubling response of about 2.0 °C, although Kelly and Wigley (1990) claim that a much smaller solar variation gives the best fit to observations. This is supported by a recent thorough statistical analysis of temperature and solar variation by Schonwiese et al. (1994), who conclude that solar variability explains at most 15–20% of global mean temperature changes during the past century. Even the extreme case of a 1% solar constant variation corresponds to a heating change of only 2.4 W m⁻². With regard to global mean deep ocean-surface heat flux oscillations, Harvey (1992) estimated an upper limit of ±1.6 W m⁻² on a century time scale. A 1000-year simulation with a coupled atmosphere–ocean general circulation model indicates that oscillations in the deep ocean-surface heat flow produce surface temperature changes, at the century time scale, which are almost an order of magnitude smaller than the temperature response of the same model to a CO₂ doubling (Stouffer et al., 1994). In contrast to the 1–2 W m⁻² upper limit for the effect of solar and oceanic variability, the projected heat trapping could reach 10 W m⁻² by the end of the next century for some business-as-usual GHG emission scenarios (Harvey, 1989; Shine et al., 1990). Hence, irrespective of the absolute climatic responsiveness to heating perturbations, the projected GHG buildup under business-as-usual scenarios will dominate natural sources of climatic variability during the coming century.

Other anthropogenic sources of climatic change include: (i) depletion of stratospheric ozone, which appears to have caused a cooling tendency of 0.1 W m^{-2} in the 1980's (Ramaswamy et al., 1992); (ii) reflection of solar energy due to the direct and indirect (cloud-related) effects of sulphate particles ('aerosols') produced from sulphur emissions, associated primarily with the burning of coal and oil, which may have caused a globally averaged cooling tendency of about $1\text{--}2 \text{ W m}^{-2}$ (Charlson et al., 1992); and (iii) reflection of solar energy due to the direct and indirect effects of aerosols produced from biomass burning, which could be as large as 2 W m^{-2} in the global average (Penner et al., 1992). These theoretical calculations indicate that the combined cooling tendency due to sulphate and biomass aerosols could be as large or larger than the heating caused by the associated release of GHG's in the global mean, although observations of historical temperature changes, discussed below, suggest that the cooling due to aerosols and aerosol-induced cloud changes is about half of the greenhouse-heating effect. In any case, the cooling effects due to biomass and sulphur aerosol emissions are concentrated geographically, so that significant regional climatic changes can still be expected.

Aerosol particles, whether produced from biomass burning or combustion of coal, have a lifespan in the atmosphere on the order of days, compared to an effective lifespan for anthropogenic CO₂ perturbations of about 100 years. Pressure to preserve remaining tropical forests, or simply destruction of the remaining forests, will likely lead to a reduction in biomass-related emissions within the next few decades, while concerns about acid rain provide continuing pressure to reduce sulphur emissions. As biomass aerosol and sulphur emissions are reduced, the offsetting cooling effect will weaken, while the accumulated CO₂ buildup will persist. Indeed, simply stabilizing both aerosol and CO₂ emissions would stabilize aerosol concentrations (and the associated cooling effect), while CO₂ concentrations would continue to increase indefinitely. Should aerosols emissions increase, that portion of their cooling effect related to induced changes in cloud optical properties (which could be comparable to the direct cooling effect) is expected to saturate (Kaufman et al., 1991). Although an increase in atmospheric relative humidity would increase the cooling effect of aerosols (Piliinis et al., 1995), most climate models project close to constant RH as climate warms (as discussed below). Thus, although biomass and sulphur aerosols might be offsetting a large fraction of the greenhouse heating in the global mean at present, the heating effect of continuing GHG emissions will eventually overwhelm other anthropogenic impacts on global and regional climate. The greater the future concerns over forest destruction and acid rain, the sooner the present uncertain masking of GHG heating effects will diminish.

2.3. CLIMATE-SYSTEM TEMPERATURE RESPONSE

Perhaps the most important link in a global warming risk assessment is the climatic response to a given GHG buildup and associated heat trapping. This can be assessed

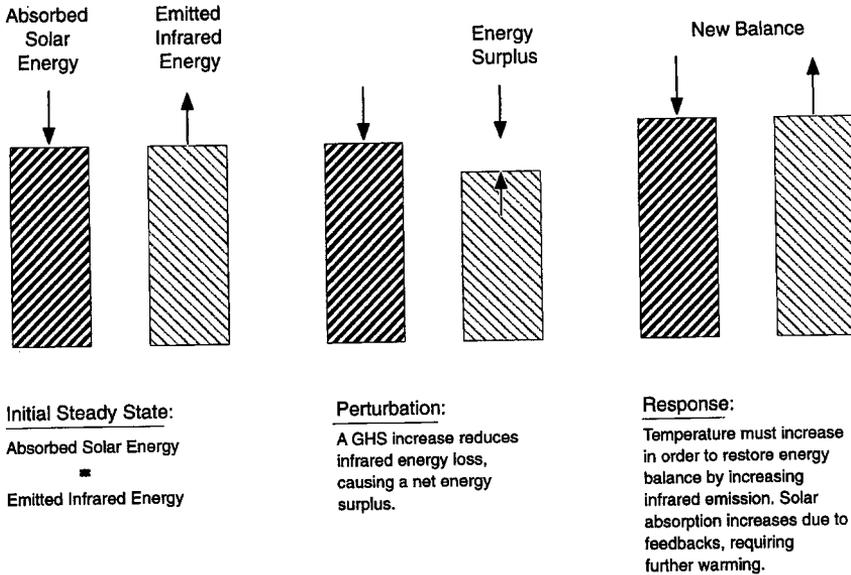


Figure 3. The radiative energy balance of the earth-atmosphere system and the response to a perturbation in the infrared loss to space.

or constrained by three independent methods: (i) based on computer models which simulate individual climatic feedback processes; (ii) by analysis of past climates and estimated associated heating or cooling perturbations; and (iii) by analysis of historical temperature changes. These three approaches, which give comparable results, are briefly reviewed below.

2.3.1. Computer-Based Estimates

The earth's surface and atmospheric temperatures tend to adjust themselves so that the absorbed solar energy is exactly balanced by the heat energy emitted to space. The increased heat trapping due to an increase of GHG concentrations requires that the mean temperature at which the earth radiates heat to space increase so that a balance between absorbed solar energy and emitted heat energy can be restored. This requirement is illustrated in Figure 3. The climatic response to a CO₂ doubling can be decomposed into three steps: (1) the temperature increase required by a perfect radiator, which is 1.06–1.20 °C and not open to dispute; (2) the temperature increase when water vapor and ice and snow feedbacks are taken into account, which is estimated to be about 2.0–2.5 °C according to conventional climate models; (3) and the temperature increase when cloud feedback is taken into account, which yields a range of 1.5–4.6 °C.

The increase of atmospheric water vapor projected by atmospheric models to accompany climatic warming is the single largest feedback in these models. Among the minority of critics sceptical of projections of significant global warming from GHG increases, Lindzen (1990) has published the only potentially serious

objection. His argument is based on the hypothesis that increased cloud convection due to overall warming will lead to a drying of the upper troposphere through the increase in subsidence between clouds which would be induced. Since water vapor in the upper troposphere is particularly effective as a GHG, this could result in a significantly smaller or even negative net water vapor feedback, thereby substantially reducing the projected warming.

One response to Lindzen's hypothesis is to compare trends and variability in upper troposphere moisture and temperature in the tropics. Direct observations of nearly 20 years of tropospheric temperature and humidity through a broad region of the tropics indicate that temperature and humidity have increased together at all heights examined (1000–300 mb), as expected based on conventional climate theory (Hense et al., 1988; Flohn and Kapala, 1989; Gutzler, 1992). Indeed, Gutzler (1992) indicates that relative humidity increases as temperature increased at all heights, although the trend is statistically different from zero at the 5% significance level only below 800 mb. However, interannual variability of temperature and humidity in the tropics exhibits a pattern such that humidity changes by only 3/4, 1/2, and 3/4 that expected based on constant relative humidity at heights of 1000–900 mb, 700 mb, and 400 mb, respectively (Sun and Oort, 1995). It might be that the relative humidity variation is different for interannual and long-term variations. Graham (1995) finds this to be the case for the 1970–1992 time period, and furthermore, that an atmospheric general circulation model forced by observed changes in sea-surface temperature successfully replicates these differences. In any case, observations are suggestive at best.

Another approach is to assess Lindzen's hypothesis from a mechanistic point of view. This requires answers to the following questions: (i) What is the effect of convection on upper troposphere moisture? (ii) What other processes affect upper level moisture and how might they change as climate changes? (iii) How does total column water vapor amount change with climate? and (iv) What is the net effect on radiative heating of hypothetical changes in the vertical water vapor distribution in combination with expected changes in total water vapor amount?

With regard to the first question, direct observations of the relationship between the intensity of convection and atmospheric moisture variations between different seasons and different regions have been interpreted as implying that increasing convection moistens the entire troposphere (Rind et al., 1991; Inamdar and Ramanathan, 1994). However, this correlation could be due in part to differences in atmospheric dynamics (i.e., a shift from lower level convergence to divergence) between the regions and seasons with weak and strong convection. Stronger evidence is provided by Soden and Fu (1995), who show on the basis of six years of data, that enhanced deep convection is associated with increased upper tropospheric moisture even at scales large enough (4000 km × 4000 km) to encompass areas of both large-scale ascent and descent.

The first question can also be addressed theoretically. Simulations with the Goddard Institute for Space Studies (GISS) atmospheric General Circulation Model

(GCM) using two different cumulus convection schemes, both of which account for subsidence-induced drying between cumulus towers, indicate that the entire troposphere moistens as the climate warms (del Genio et al., 1991). Indeed, for a 4-K globally uniform ocean-surface warming, relative humidity is constant near the surface but increases in the global mean by an amount which grows with increasing height to 5% at the 200-mb level. This change is the net effect of all the processes in the GISS model which affect water vapor distribution. Sinha and Allen (1994), on the other hand, examined the effect of convection and convection-induced subsidence in isolation, and conclude that an increase in the intensity of convection can cause the upper troposphere to become dryer in some height intervals, depending on the choice of model parameters which are themselves highly uncertain. For two intermediate cases they obtain peak drying by 10% centered at heights of 500 mb and 250 mb in response to a 1K surface warming, with less drying at other heights in the upper troposphere and moistening close to that expected for constant relative humidity below a height of 700 mb. The model used by Sinha and Allen (1994) adopts one of the key assumptions of Lindzen's hypothesis – that all condensed water vapor in deep clouds falls to the ground as rain rather than partially evaporating and moistening the upper troposphere – an assumption that was first questioned by Betts (1992) and later refuted by model-based analysis of observations (Sun and Lindzen, 1993). Relaxation of this assumption may very well cause the simulated upper troposphere drying to disappear. In contrast, del Genio et al. (1991) allow for detrainment of ice crystals from cumulus cloud tops, which undoubtedly contributes to the upper level moistening in the global mean found in their simulations. Sinha and Allen (1994) indicate that introduction of other processes can also substantially reduce the drying, but that introducing yet further processes can bring it back. Based on Sinha and Allen's (1994) results and the limitations noted above, one might conclude that drying of the upper troposphere of up to 10% could occur in response to a 1K surface warming in regions where upper tropospheric moisture is governed by detrainment from cumulus towers and subsequent subsidence. However, as noted above, direct observations indicate that upper troposphere moistening rather than drying occurs when convection increases.

With regard to the second question, del Genio et al. (1994) find that eddies and mean meridional motions are the dominant factors influencing upper tropospheric moisture outside the tropics. Their analysis of observations and model results indicates that the moistening effect of both of these processes would increase as the climate warms. Thus, even if increasing convection were to dry the upper troposphere in regions surrounding intense convection, this may not imply a drying of the upper troposphere in the global mean.

With regard to the third question, total column water vapor amount clearly increases as sea-surface temperature increases; the relationship in the tropics is weak at the monthly and annual time scales, where concurrent changes in atmospheric dynamics complicate the picture, but is stronger at the decadal time scale (Gaffen et al., 1992). An increase in total column water vapor with increas-

ing surface temperature will lead to an increase in heat trapping as temperatures increase unless a sufficiently strong downward shift in the moisture distribution occurs.

With regard to the fourth question, Harvey (1996b) finds that adoption of the upper limit to the drying effect of convection in low latitudes suggested above, combined with increases in total column water vapor as climate warms similar to that suggested by observations, only reduces the global mean temperature response to a CO₂ doubling to 1.1 °C with fixed cloud top heights and to 1.2 °C with variable cloud top height at low latitudes; the latter case is physically consistent with the hypothesized mechanism for upper troposphere drying.

The evidence reviewed above supports the conventional wisdom that the water vapor feedback is positive. Observational data, taken at face value, indicate that relative humidity is approximately constant or increases slightly throughout the troposphere as climate warms, and that greenhouse trapping increases strongly with increasing surface temperature, in broad agreement with climate models. The theoretical work of Sinha and Allen (1994) shows that it is possible that the effect of enhanced convection, acting alone, could be to dry parts of the upper troposphere. However, even in the extreme case in which a 1K surface warming provokes a 10% drying of the upper troposphere in tropical regions, the global mean temperature response to a CO₂ doubling – in the absence of cloud feedbacks other than increasing cloud height at low latitudes – is still about 1.2 °C.

The likely overall impact of cloud feedbacks is highly uncertain. This is due to the fact that clouds have large but opposing effects on global temperature (they reflect solar energy but also trap heat energy radiated from the surface), and that a number of individual cloud properties affect both the cooling and heating effects of clouds. In an assessment of cloud feedback in 19 different models, Cess et al. (1990) found that cloud feedback varied from slightly negative to strongly positive, a pattern which persists in more recent analyses of individual models (Taylor and Ghan, 1992; Senior and Mitchell, 1993). However, cloud schemes are very crude in all climate models, so these results could be misleading. Nevertheless, the evidence reviewed above indicates that the climatic response to a CO₂ doubling is almost certain to exceed 1.0 °C, and supports the long-standing contention that the climatic warming due to a CO₂ doubling is highly likely to exceed 1.5 °C (Gates et al., 1992).

2.3.2. *Analyses of Past Climates*

By comparing estimates of the global mean temperature differences for geologic periods with distinctly warmer or colder climates than at present with estimates of the radiative heating or cooling perturbation relative to the present, an empirical estimate of the climatic responsiveness to radiative perturbations can be obtained. The most recent such exercise (Hoffert and Covey, 1992) yields an estimated climatic response for CO₂ doubling of 2.3 ± 0.9 °C, which falls within the lower half of the 1.5–4.5 °C range derived from climate models (Gates et al., 1992).

In general, times of inferred high CO₂ concentration throughout geologic history coincide with times of inferred warm climates, while times of low atmospheric CO₂ coincide with episodes of extensive glaciation (Berner, 1990). This implies either that atmospheric CO₂ variations do indeed exert a significant influence on climate, which is consistent with the evidence presented above, or that atmospheric temperature controls atmospheric CO₂ concentration at multi-million year time scales and longer. The latter contradicts the main competing geochemical models of the carbon cycle (i.e., Berner et al., 1983; Francois and Walker, 1992). These models have been successful in explaining key features of the observed geochemical record, and contain negative feedbacks between atmospheric temperature and CO₂ concentration but no mechanism whereby warmer temperatures would induce higher atmospheric CO₂ at multi-million year time scales. Hence, the good correlation between atmospheric CO₂ and climate at multi-million year time scales and longer is further evidence that the climate is sensitive to variations in CO₂ concentration.

2.3.3. *Analysis of Historical Temperature Changes*

The analysis of global mean temperature changes during the past century could, in principle, provide an independent estimate of climate sensitivity. However, because of uncertainty in the magnitude of cooling effects associated with biomass and sulphur aerosols, and in the magnitude of solar radiation variations during the past century, the observed global mean temperature changes are consistent with a climatic responsiveness to CO₂ doubling from as little as 0.8 °C (assuming no offsetting heating effects due to anthropogenic aerosols have occurred) to as large as 12 °C (assuming a recent global mean aerosol cooling effect of 1.4 W m⁻²) (Schlesinger and Ramankutty, 1992). Thus, although historical temperature changes are not inconsistent with model projections of significant warming for CO₂ doubling, the global mean trend does not provide a meaningful constraint on model projections. However, recent analyses of oscillations about the long-term trend and of the geographical pattern of climatic change indicate that both solar variability and atmosphere–ocean variability are modest contributors to the long-term trend (Kelly and Wigley, 1992; Schlesinger and Ramankutty, 1994), with about half of the long-term warming attributable to net anthropogenic effects (primarily GHG's and aerosols).

Analysis of the diurnal pattern of temperature change provides an estimate of the magnitude of the aerosol-cooling effect. A recent analysis covering 50% of Northern Hemisphere land and 10% of Southern Hemisphere land indicates that the average minimum nighttime temperature increased by 0.84 °C during the period 1951–1990, while the average maximum temperature increased by only 0.28 °C (Karl et al., 1993). Hansen et al. (1995) tested the effect on diurnal temperature variation of a wide variety of heating perturbations, and find that the only mechanism capable of simultaneously generating changes in the mean temperature and in the diurnal amplitude comparable to that observed is localized

anthropogenic aerosol and aerosol-induced cloud changes *in combination with* a large-scale warming factor, such as GHG increases. Their analysis implies that the combined aerosol-cloud cooling perturbation is about half as large as the anthropogenic greenhouse-heating perturbation in the global mean. This, combined with the above-noted constraints on solar and atmosphere–ocean variability, implies a climatic responsiveness for CO₂ doubling in excess of 2.0 °C based on Schlesinger and Ramankutty (1992; their Figure 1a).

2.3.4. *Synthesis and Perspective on Temperature Response*

Several lines of evidence, reviewed above, indicate that the global climate will respond significantly to the increase in heat trapping which would accompany the projected buildup of GHG's. Motivated by our desire to develop a risk-hedging strategy, as defined in the introduction, this review has focused on the lower limit to climatic sensitivity, and leads to the conclusion that climatic response to a CO₂ doubling is almost certain to exceed 1 °C, and is highly likely to exceed 1.5 °C. Although we have not dealt explicitly with the upper limit, analyses of cloud feedbacks combined with our assessment of the water vapor feedback supports the longstanding conclusion, reiterated by the Intergovernmental Panel on Climate Change (IPCC) in 1992 that, "the sensitivity of global mean surface temperature to doubling CO₂ is unlikely to lie outside the range 1.5 to 4.5 °C" (IPCC, 1992, p. 5). Global mean temperature change near the lower limit of this range would still have significant effects on many ecosystems and human societies, while temperature change near the upper limit – being comparable to the difference between the present climate and the last ice age – would have serious and far-reaching effects. However, barring constraints on the use of fossil fuels, the heating effect of GHG's is expected to eventually far exceed the equivalent of a CO₂ doubling, so that even for a low climate responsiveness, significant climatic change would be delayed but not averted.

As noted above, recent observed warming of minimum nighttime temperatures over land has been three times larger than the increase in daytime temperature maxima. This, however, does not imply that global warming poses a significantly reduced threat. First, the observed asymmetry in day–night warming seems to be linked to the effects of anthropogenic aerosols and aerosol-induced changes in cloudiness (Hansen et al., 1995) – an effect which can be expected to weaken as local and regional air pollution problems are addressed. Secondly, in transient response simulations carried out by Hansen et al. (1995) with both GHG and aerosol-cloudiness increases, most of the change in diurnal cycle due to aerosol and cloud effects occurs quickly, while the change in mean temperature is delayed by 2–3 decades. Thus, as global mean temperatures come closer into balance with the GHG heating perturbation (which must inevitably happen), the difference between daytime and nighttime warming will diminish. Thirdly, warming concentrated at nighttime would still be important for many anticipated impacts of climatic change.

2.4. CLIMATE SYSTEM MOISTURE RESPONSE

Computer-based estimates of the moisture response to increasing temperature indicate that, although there will be an increase in global mean precipitation as the climate warms, there will also be an increasing tendency for drying of continental interiors in summer because continental evaporation tendency will increase faster than the rate of precipitation increase over the continents. Kellogg and Zhao (1988) and Zhao and Kellogg (1988) compare soil moisture changes for a doubled CO₂ climate as simulated by several models. The models which simulate the greatest drying appear to be over-estimating the response and models which simulate minimal drying appear to be underestimating the response, based on known deficiencies in the models (Mitchell and Warrilow, 1987; Meehl and Washington, 1988). The GFDL (Geophysical Fluid Dynamics Laboratory) model in particular appears to over-estimate summer drying because it simulates soils to be saturated at the end of winter for the present climate. Hence, the projected increase in winter precipitation (which is a common model result) is lost as increased runoff, whereas in other models the winter precipitation increase leads to higher end-of-winter soil moisture, which partly compensates for an increase in evaporation during spring and summer. Many other models simulate soils to be so dry in summer for the present climate that significant further drying is not possible, such that summer drying is underestimated (Rind et al., 1990). On the whole, climate models indicate that there will be tendency for drier summer soils and for an increase in the frequency of drought in continental interiors under a doubled CO₂ climate. This tendency will likely be reduced to some extent by increased stomatal resistance in a higher CO₂ world (Henderson-Sellers et al., 1995).

In contrast to computer-based results, analyses of inferred moisture patterns associated with warm intervals in the geological past suggest that continents will become moister rather than drier as the climate warms. However, past warm climates are not suitable guides to the moisture conditions that would occur in association with future warming for at least two reasons: (i) as discussed by Crowley (1990), the seasonal solar radiation pattern (in the case of the mid-Holocene and last interglacial warm periods) or polar ice cover (in the case of earlier warm periods) were quite different from that which will be associated with warming during the next few centuries, and these differences could readily produce a substantially different precipitation response to warming; and (ii) a significant part of the moisture response to past warm intervals likely involved adjustments in soil properties which require over 1000 years to occur (Lapenis and Shabalova, 1994), and hence would not be involved in moisture changes during the coming century. Hence, data from past climates cannot be taken as evidence that a warmer world would lead to moister continents. Rather, simple principles indicate that, for fixed soil properties, increasing warming should lead to an increasing tendency for drying in continental interiors (Rind et al., 1990).

2.5. IMPACTS OF CLIMATIC CHANGE

Under business-as-usual scenarios, GHG concentrations will continue to increase well beyond the end of the next century, leading to global warming, sea-level rise, and ecosystem responses for hundreds of years. Initial impacts could therefore be quite different from later impacts. Impacts of climate change arise both from the nature of the climatic change, and from the rate at which it occurs. The magnitude of climatic change is likely to be most important for agriculture, water resources, and human comfort, whereas the rate of climatic change (and associated sea-level rise) is likely to be most important for forests, coastal wetlands, coral reefs, estuaries, and human infrastructure. Climatic changes which are more rapid than the rate at which ecosystems can adjust will have negative transient effects which could lead to a significant number of species extinctions, even if the final climatic changes improve biological productivity. Potential impacts in selected sectors are outlined below.

2.5.1. *Agriculture*

With modest warming, some regions are likely to see increases in agricultural yields while other regions are likely to see decreases, but it cannot be predicted with any confidence who will gain and who will lose. As noted above, many climate models indicate a summer drying of continental interiors under a doubled CO₂ climate.

Two assessments of the impact of climatic change on food production have recently been completed which link food crop and macro-economic considerations, and are used here as examples: the global assessment of Rosenzweig et al. (1993) (summarized by Rosenzweig and Parry, 1994), and the assessment for the Missouri–Iowa–Nebraska–Kansas (MINK) region of the US by Easterling et al. (1993). The Rosenzweig study uses scenarios of temperature and precipitation change for CO₂ doubling as simulated by three different climate models, while the Easterling study uses the actual climatic conditions of the 1930's as a possible analog for conditions in 2030. Both assessments take into account the direct physiological effects of higher atmospheric CO₂ concentration on food crops as measured in experimental settings, as well as varying degrees of adaptation to the changed climate by farmers. The direct physiological effects include stimulation of photosynthesis, increased yield, and increased water use efficiency. Table I summarizes the mean temperature, precipitation, and CO₂ changes used in the two studies, as well as the changes in total cereal production for cases of no CO₂ physiological effects and adaptation, physiological effects without adaptation, and physiological effects with adaptation. In the best case climate scenario, global food production decreases by 11% (no CO₂ effects and no adaptation) or increases by 1% (with CO₂ effects and full adaptation) in the Rosenzweig study, while in the worst case the range is from a 20% decrease to a 2.5% decrease. The Easterling study, although applied to a different climatic change scenario and to a specific region only, is consistent with the global results of the Rosenzweig study in suggesting that a combination of CO₂ effects and full

Table I

Comparison of climatic-change scenarios and changes in food production obtained by the study of Rosenzweig and Parry (1994) and Easterling et al. (1993). Impacts are given for cases with and without direct physiological effects of higher CO₂ and plants, and with no or varying degrees of adaptation to the change in climate

Attribute	Rosenzweig and Parry	Easterling et al.
ΔTemperature	4.2 °C, 4.0 °C, 5.2 °C ^a	0.9 °C
ΔPrecipitation	11%, 8%, 15% ^a	-3% to -15%
ΔCO ₂	277.5 ppmv	100 ppmv
<i>Impact on food production (global or regional)</i>		
No CO ₂ effect or adaptation	-11% to -20%	-22%
Full CO ₂ effect, no adaptation	-1% to -8%	-8%
Full CO ₂ effect, weak adaptation	0% to -5%	-
Full CO ₂ effect, strong adaptation	+1% to -2.5%	+3%

^a Global mean values.

adaptation can change a substantial (about 20%) yield loss into a very small loss or a slight increase.

In spite of these favourable results, there is reason for concern. First, Bazzaz and Fajer (1992) question whether the full carbon-fertilization effect would occur in actual growing conditions, since the beneficial effects require adequate water and nutrients, which might not be readily available in developing countries (water could also be a problem in developed countries if irrigation is not possible). Ground-level ozone pollution, which coincides with many major food-producing regions and is a byproduct of fossil fuel use (Chameides et al., 1994), also tends to suppress the beneficial effects of higher CO₂ (Barnes and Pfirrmann, 1992). Wolfe and Erickson (1993) cast further doubt on the applicability of full experimental CO₂-fertilization effects to real-world growing conditions. Second, even in instances where global net effects are small in the Rosenzweig study, developing countries are generally projected to experience significant losses while developed countries are generally projected to experience gains. The greater projected losses in developing countries arise from their reliance on rice, which is particularly sensitive to temperature extremes (see below). These losses would be exacerbated by sea-level rise, which was not included in the Rosenzweig study, and would also have the greatest agricultural impacts in the developing world. Third, the Easterling study might not be a good analog for a doubled CO₂ climate because they may have used too large a ratio of CO₂ increase to climatic change compared to what would occur in reality under a doubled-CO₂ climate. The projected summer warming for the Easterling study area under a doubled CO₂ climate is about 4–7 times that used by Easterling et al. (about 4–6 °C instead of 0.9 °C), based on maps of summer temperature change for three different models given in Mitchell et al. (1990). However, the actual CO₂ increase when this amount of warming occurs could be as little as 200

ppmv and is unlikely to be more than 300 ppmv – that is, only 2–3 times that used by Easterling et al. Even if the Easterling study used an appropriate ratio of CO₂ increase to temperature change, the balance of these two changes is likely to be less favourable when both absolute changes are larger, because of saturation of the beneficial CO₂ effects and the nonlinear increase of potential evapotranspiration with temperature.

The Rosenzweig study imposes model-generated changes in mean temperature and precipitation on present-day time series of interannual temperature and precipitation changes, and therefore does not allow for possible changes in temperature and precipitation variability as the climate warms. Mearns et al. (1992) examined the relative effects of changes in temperature and precipitation means and variabilities for wheat grown at two sites in Kansas. They find that, in general, changing the interannual variability of monthly mean temperatures by a factor of four has a much smaller effect on mean yields than increasing the mean temperature by 4.5 °C. If mean moisture levels are high, increasing precipitation variability decreases the mean yield (since losses in years with reduced precipitation are greater than gains in years with increased precipitation), whereas if mean moisture is low, an increase of precipitation variability increases the mean yield. In all cases, increases in temperature or precipitation variability increase the frequency of crop failures. As discussed below, climate models indicate that precipitation variability will increase in a warmer world, particularly at low latitudes. When the crop models used by Mearns et al. (1992) are driven with GCM-generated changes in precipitation mean and variability, the change in precipitation variability (an increase) has a larger impact on mean yield than the change (a decrease) in mean precipitation. In short, potential increases in precipitation variability could be important, could reduce mean yields in many regions while increasing it in others, and invariably increase the frequency of crop failures.

Even if the impact of a CO₂ doubling on global food production is small or beneficial in aggregate, there are several reasons for expecting the risk of significant negative net effects to grow with increasing climatic warming. First, as temperatures increase, the tendency for evaporation of soil water to increase is likely to increasingly offset any precipitation increases (Rind et al., 1990). Second, agricultural losses along the present equatorward margins of agriculture due to a doubled-CO₂ climate can, to some extent, be compensated by gains along the poleward margins. However, with each climate warming increment, less land with suitable soils along northern margins will be available, in general, to compensate for agricultural losses on southern margins. Third, the effect of higher CO₂ concentration in stimulating photosynthesis is likely to saturate with increasing CO₂ concentration. For present growing conditions, the saturation point is as low as 550 ppmv (in the case of maize, according to Parry, 1990), although warmer temperatures would likely shift the saturation point to a higher CO₂ concentration. Fourth, the frequency with which critical temperature thresholds are crossed will increase as climate continues to warm. Rice in particular is sensitive to high temperatures;

yields decrease by a factor of two in going from a daytime temperature of 28 °C and 330 ppmv CO₂ to 34 °C and 660 ppmv CO₂ (Baker et al., 1992). Relatively small changes in mean temperatures will likely lead to large increases in the frequency with which critical thresholds are crossed, given the existence of temperature variability. Fifth, sea-level rise could significantly reduce the area of highly productive agricultural land in the densely populated, fertile deltas of the world.*

In summary, although agriculture is one of the most adaptable sectors which will be impacted by global warming, it will be increasingly difficult to offset losses through adaptation as the magnitude of warming increases beyond that of a CO₂ doubling. Substantial risks also occur at the regional level with as little as a CO₂-doubling equivalent. The risks are particularly pronounced for developing countries because of their reliance on rice (which already grows close to its critical heat-stress threshold in some regions), because full realization of the beneficial physiological effects of higher CO₂ requires adequate supplies of water and fertilizer, and because of the limited purchasing power of large segments of third-world populations (economic development, to the extent that it occurs by the time a doubled-CO₂ climate is reached, could partly alleviate the latter two factors). Risks to both developed and developing countries grow substantially with increasing warming.

2.5.2. *Forests*

Estimated impacts of doubled-CO₂ climate on regional forest bioproductivity, once the forest distribution is fully adjusted to the new climate which itself is assumed to be no longer changing, range from modest increases (i.e., Williams, 1985; Bonan et al., 1990) to decreases (i.e., Pastor and Post, 1988; Reed and Desanker, 1992; LeBlanc and Foster, 1992), depending on the change in soil moisture. However, climatic zones are anticipated to shift by several hundred kilometers during the coming century under business-as-usual GHG-emission scenarios, which is greatly in excess of the rate at which forests can migrate through seed dispersal (Davis, 1989). Where forested areas are highly fragmented due to habitat destruction, rates of forest migration will be substantially slower than would occur under natural conditions (Schwartz, 1992a). Studies of the transient response of forests to climatic change indicate that present forests will be increasingly stressed as climate changes, particularly along their equatorward margins, resulting in significant losses in forest biomass (Solomon, 1986; Solomon and Bartlein, 1992; Schwartz, 1992b). Fire frequency and severity are likely to increase as a result of increased drought and a greater supply of dead wood. Although removal of maladapted forests through fires will allow faster replacement by species better suited to the changing climate (Overpeck et al., 1990), there would be several centuries of ecological imbalance, reducing the aesthetic, recreational, and commercial value of forests.

As in the case of agricultural crops, there is increasing evidence that the direct physiological effect on forests of a higher atmospheric CO₂ concentration is to

* For example, a 1-m rise of sea level would eliminate 15% of Egypt's present agricultural land through flooding of the Nile delta.

stimulate photosynthesis and reduce water loss through transpiration, all else being equal (Bazzaz, 1990). However, reduction in plant transpiration would tend to increase leaf-surface temperatures, especially for broadleaf plants, while the stimulatory effect on plant photosynthesis of higher CO₂ concentration could increase total leaf area. These two effects together could increase transpiration rates enough to nullify the effect of stomatal constriction (Allen, 1989). The partial control of humidity within a forest canopy by forest transpiration itself will also weaken the reduction in transpiration rates from stomatal constriction (Jarvis and McNaughton, 1986). At multi-decadal time scales, feedbacks with nutrient cycles are likely to reduce the long-term stimulatory effect of higher CO₂ concentrations, as discussed above in the context of the carbon-cycle response to CO₂ emissions. Thus, in spite of beneficial direct physiological effects of higher CO₂ on forest species, significant negative impacts on forests can be expected based on anticipated rates of climatic change.

Significant indirect effects of climatic change on forests will occur if global agricultural productivity decreases as a result of climatic change, since this will place enormous pressure to clear remaining forested areas for agricultural use. This interaction has been neglected in some previous economic assessments of the impact of climatic change (i.e., Kane et al., 1992).

2.5.3. *Coastal Regions*

The sea-level response to a continuously warming climate involves thermal expansion of ocean water and changes in the mass of glacier ice. For a CO₂ doubling, thermal expansion alone could produce an eventual sea-level rise of 1–2 m (Harvey, 1994). It is likely that initial warming could result in an increase in snow accumulation over Antarctica (Budd, 1991), which would tend to offset sea-level rise due to a reduction in glacier mass elsewhere and thermal expansion of ocean water (Miller and Vernal, 1992; Schneider, 1992). Beyond a certain warming, however, collapse of the marine-based West Antarctic ice cap could begin, resulting in a sea-level rise of 4–5 m over a period of several millennia. Model simulations by Budd (1991) indicate that a 4 °C warming around the Antarctic coast, which would easily be associated with a CO₂ doubling, would remove most of the floating ice shelves within 50 years. This would be followed by gradual disintegration of the West Antarctic ice cap, leading to a sea-level rise of 0.6 m 100 years after the melting of the ice shelves, 3 m after 1000 years, and 4.5 m after 5000 years. Disintegration of the West Antarctic ice cap would therefore have no effect on sea level during the coming century, for which the expected global mean sea-level rise is 0.3–1.0 m.

A 1-m sea-level rise would displace on the order of 100 million people in the densely populated, fertile deltas of the world (in China alone, 72 million people would be displaced, while 17% of Bangladesh would be inundated (Cline, 1992)). Additional populations would be impacted in the target areas to which those displaced by sea-level rise migrate, with a risk of social tension and an increased risk of spread of disease. Titus et al. (1991) estimated the quantifiable costs of a

1-m sea-level rise to be \$270 to \$475 billion for the US alone. A 1-m sea-level rise would cause beaches to erode 50–1000 m along the US east coast and 200–400 m in California, yet most US recreational beaches are less than 30 m wide at high tide (Titus et al., 1991). A 0.5–1.0-m sea-level rise within one century would also lead to significant losses of coastal wetlands and mangroves worldwide, which perform a variety of important ecological and economic functions. Coastal wetlands serve as breeding grounds for many commercially important marine fish and crustacean species, with mangroves also serving as storm barriers and sources of fuel wood. Assessments for the US east and Gulf coasts indicate that a 0.5-m sea-level rise by 2100 would result in a loss of 14–26% of present coastal wetland area, while a 1-m sea-level rise by 2100 would result in a 46–52% areal loss (Reid and Trexler, 1992). Human and economic impacts of wetland and mangrove losses are likely to be most severe in developing countries, where these ecosystems often play an important role in the social and economic well-being of coastal populations (Hong, 1993). Estuaries – defined as partly enclosed coastal bodies in which salinity is noticeably diluted by river inflow – are another threatened coastal ecosystem. As sea-level rises and high salinity waters push inward, estuarine species will be pushed into habitats which in some instances will not be suitable due to differences in sediment character. As temperatures increase, heat-sensitive species will need to move to estuaries further poleward, but such movement is expected to take centuries or millennia – much too slow to keep up with projected climatic change (Kennedy, 1990).

In short, even a 1-m sea-level rise would have enormous economic, biological, and human impacts. The possibility exists that a significantly greater sea-level rise could eventually occur in association with a CO₂ doubling. The risk of very large (5–6 m) sea-level rise likely grows rapidly as warming increases beyond the CO₂-doubling level, but the time scale for a sea-level rise this large appears to be millennia. As with agricultural effects, the worst impacts in human terms are likely to occur in developing countries, although the greatest economic impacts are likely to occur in the US (Cline, 1992).

2.5.4. *Coral Reefs*

Coral reefs – a legacy of gradual post-glacial sea-level rise during the past 10,000 years – are threatened by rates of sea-level rise more rapid than their ability to grow upward, by shifts in the location of water masses of suitable temperature faster than the ability of coral to colonize new locations, and through changes in ocean chemistry as anthropogenic CO₂ is absorbed by the oceans. Coral reefs can grow upward at a maximum rate of about 1 cm/year (Gornitz, 1991), which matches the maximum anticipated rate of sea-level rise during the coming century. However, reefs which are stressed due to heat, pollution, or increasing UV-B radiation will not grow as fast (Smith and Buddemeier, 1992). Water temperatures too warm for coral appear to cause expulsion of the algae upon which they depend (a phenomenon known as coral bleaching), leading to death of the coral if sustained

(Roberts, 1990, 1991). Tropical surface-ocean water at present is 520% saturated with respect to calcite, and 340% saturated with respect to aragonite – the two most common structural materials in corals. Increasing atmospheric and surface water CO₂ partial pressure to 600 ppmv from the present value of about 355 ppmv would decrease the saturation states with respect to calcite and aragonite to 360% and 240%, respectively. Complicating the issue is the finding that higher ocean-water CO₂ in some cases inhibits growth by coralline algae, while in other cases it is enhanced. Although very little is known about the impact of any of these changes on coral communities, a plausible expectation is that the ability of coral reefs to keep up with rising sea level will be diminished (Smith and Buddemeier, 1992). If upward coral growth lags behind sea-level rise, bioproductivity will decrease and the value of corals as storm surge barriers will be reduced.

2.5.5. *Water Resources, Rainfall, and Storms*

Stress on water supply might be one of the most important consequences of global warming, affecting the timing and magnitude of runoff, reducing lake levels and groundwater availability, and adversely affecting water quality. Results for two of the most intensively studied basins, the Sacramento Basin (Glieck, 1987) and the Great Lakes Basin (Cohen, 1986; Croley, 1990), both indicate significant reductions in water supply. Semi-arid regions are particularly vulnerable to reductions in water supply.

Modelling studies indicate that the intensity of convective rainfall will increase but that the amount of large-scale (and gentler) rainfall will decrease in a warmer world (Gordon et al., 1992). Hence, even where total precipitation increases, a larger fraction is likely to be lost as runoff, resulting in less soil moisture and greater flood hazard. Modelling studies also suggest that the interannual variability of rainfall due to El Niño-related events will increase in the tropics under a doubled-CO₂ climate (Meehl et al., 1993), with associated negative impacts.

A widely held opinion is that the intensity of tropical storms will increase in a warmer world. Although there are theoretical reasons for believing that warmer sea-surface temperatures will produce more intense storms if all else is unchanged (i.e., Emanuel, 1988), concurrent warming of the entire troposphere could significantly reduce the expected increase in storm intensity according to high resolution model sensitivity tests (Drury and Evans, 1993). Empirically, sea-surface temperature is a poor predictor of storm intensity today (Evans, 1993), and it is far from clear what effect, if any, a warmer climate will have on storm intensity. There is also no compelling reason for expecting the frequency of tropical storms to increase, in spite of an increase in the areal extent of ocean water above 26 °C (Lighthill et al., 1994). The effect of global warming on midlatitude storms also involves competing factors, so the net effect is not clear; modelling studies indicate that storminess could increase in selected regions such as the North Atlantic and Western Europe (Hall et al., 1994).

2.5.6. *Species Abundance and Diversity*

Climatic change will lead to species extinctions when a species' rate of adaptation or ability to relocate are exceeded. World Wildlife Fund (1992) contains an excellent discussion of the ecosystems most at risk from climatic change. These can be categorized as 'front line' ecosystems (coral reefs, mangrove swamps, arctic marine ecosystems), for which there is compelling evidence of an acute near-term threat; 'highly vulnerable' ecosystems (montane, tundra, and boreal and temperate forest ecosystems), for which there is strong evidence of a significant near-term threat; and 'high-risk' ecosystems (savannahs with seasonal rainfall and tropical forests) which are likely in jeopardy but for which the probable impacts of climatic change are unclear. Prairie wetlands can be added to the list of threatened ecosystems, as a warmer climate is expected to lead to decreases in the extent and quality of prairie wetlands (Poiani and Johnson, 1993). Currently protected ecological regions will be pushed out of the existing protected areas as climatic zones shift, but relocation in many cases will be impossible due to human obstacles. Natural ecosystems are often poorly understood, but in many cases extinction of a single-key species can lead to a cascade of unexpected effects which are undesirable from a human point of view (Ehrlich and Daily, 1993). Long lags in the full response are likely in many cases.

2.5.7. *Impacts on Stratospheric and Tropospheric Ozone*

The direct radiative effect of increasing CO₂ is to cool the stratosphere (in contrast to its warming effect on the surface and lower atmosphere). The imposition of normal year-to-year variability in stratospheric conditions on this cooling is projected to result in Arctic ozone losses as large as the present Antarctic ozone hole in about 20% of the winters once CO₂ reaches a concentration of 660 ppmv (Austin et al., 1992; Austin and Butchart, 1994). According to these analyses, only a reduction in both projected chlorine and CO₂ concentrations will avoid conditions which can create an Arctic ozone hole during the next 50 years. Once again, however, there is the possibility that indirect effects might partially mitigate the direct effect; in this case, a warmer climate may increase the frequency of polar stratospheric 'sudden warmings' through changes in atmospheric dynamics (Mahfouf et al., 1994).

At the same time that stratospheric ozone loss is likely to be worsened, warmer surface air temperatures are expected to increase the occurrence of urban photochemical smog (ozone), an increase which would be amplified further by increased UV-B radiation in the troposphere due to reductions in stratospheric ozone. Under present conditions, a 2.5 °C temperature increase would increase peak ozone concentrations in various US cities by 2–10% (Rouviere et al., 1990). However, if emissions of photochemical smog precursors are reduced, which could occur as a by-product of actions to reduce GHG emissions, then ground-level ozone concentrations could decrease.

2.5.8. *Human Comfort and Welfare*

Although it has been argued that human beings are highly adaptable and can respond to warmer temperatures by, for example, increased use of air conditioning (Ausebel, 1991), such options are generally restricted to the affluent members of affluent societies. Most of the world's population, and those in developing countries in particular, have few options in responding to increased temperatures. Peak temperatures in many of the world's major cities (i.e., Cairo, Calcutta, Bombay) are already close to the physiological limits of human tolerance; more intense extremes could have severe impacts (Pachauri, 1992). If climatic warming is slow and population growth restrained, it is possible that rapid economic development, if realized, could limit mortality increases due to heat stress in developing countries. Although temperature extremes in temperate latitudes will be less severe than in tropical latitudes, these extremes often strike suddenly with little or no time for acclimatization, leading to mortality rates which can be as high as at lower latitudes where temperatures are higher but acclimatization has occurred (Kalkstein, 1991).

Another factor directly impacting human health is the increase in the range of vector-borne diseases that would occur with a warmer climate. As one example, under a doubled CO₂ climate the range of malaria could move about 400 km north and south, thereby introducing or re-introducing the disease to the southern USA, Australia, and the Mediterranean region of Europe and exposing an additional 200 million people (Hohmeyer and Gartner, 1992). Malaria today does not occur in many regions where it is biologically capable of occurring, thanks to health system expenditures, and the impact of future warming on the incidence of malaria will depend strongly on socio-economic development. If the rate of climatic change is sufficiently slow, improvements in the health system associated with economic development in the developing countries which would experience increased risk of malaria could offset the effect of climatic change. The rate of global mean temperature increase at which this offset could occur is estimated to be 0.1 °C/decade for the period 1990–2100, which is substantially smaller than business-as-usual projections of 0.2–0.3 °C/decade (Martens et al., 1994). Here is another example where the rate of climatic change appears to be as important as climatic change itself, and where risks increase with increasing rates of climatic change.

2.5.9. *Risk of Surprises*

A final consideration in assessing the risks posed by climatic change is the possibility of surprises which, by definition, cannot be predicted in advance. Surprises could occur either through unexpected climate–carbon-cycle feedbacks, jumps in climate sensitivity, sudden changes in the global ocean circulation or the sea-level response to climatic change, or sudden reductions in the ability of ecosystems to cope with rapid climatic change as key thresholds are crossed. The climatic response to GHG increases could be erratic rather than smooth, and might even entail century time scale, regional cooling, as found in the transient response simulations of Harvey (1994). Although the risk of such surprises cannot be quantified,

it is reasonable to assume that the greater the magnitude and rate of climatic change, the greater such risks. As long as the magnitude of climatic change remains within the envelope experienced in recent geological history (the last one or two million years, for example), it is safe to assume that the risk of certain kinds of surprise is small. In particular, the sea-level and carbon-cycle response to 1–2 °C warmer temperatures in the future is not likely to be very different from that during the past two million years, when global mean temperatures appear to have been no more than 1–2 °C warmer than at present. However, greater warming will result in temperatures warmer than experienced in recent geological history, so that the recent geological past – about which we have the most information and which is most similar to the present – no longer serves as a reliable guide on which to base future expectations.

A particular example of a nonlinearity which is unlikely to lead to unpleasant surprises as long as climatic change is restricted to the range experienced during the past one to two million years involves methane clathrates (Harvey and Huang, 1995). There is the possibility that climatic warming could destabilize vast stores of methane, which are currently locked up in permafrost regions and in marine continental slope sediments worldwide. This would lead to further warming, which could destabilize more methane, and so on. It is possible that much of the current methane–clathrate accumulated gradually during the past 60 million years or so, as global temperatures steadily fell. Any clathrate that would have been susceptible to destabilization by 1–2 °C future warming would have been destabilized by 1–2 °C warming in the recent geological past and might not have fully reformed; hence, a 1–2 °C future warming is very unlikely to initiate a strong positive climate–clathrate destabilization feedback. This argument does not apply to, say, 4–5 °C warming because global mean temperatures have not been that warm for several tens of millions of years, during which time substantial methane–clathrate could have accumulated at depths where it would be susceptible to release by 4–5 °C warming.

3. Summary

Table II summarizes the risks associated with unrestrained emissions of GHG's. It is difficult to quantify these risks, but they have been subjectively rated as Low (<20% probability), Moderate (20–60% probability), High (60–90% probability), and Very High (>90% probability) based on the preceding review. This review indicates that (i) the importance of the terrestrial biosphere as a sink of atmospheric CO₂ is likely to diminish, and could very well become a net source of CO₂ if climatic change is rapid, thereby exacerbating the atmospheric CO₂ buildup and resultant climatic change as a positive feedback; (ii) other climate–carbon–cycle feedbacks, involving methane clathrates and ocean circulation, are likely to be small, even for large climatic changes; (iii) the heating perturbation due to the

buildup of GHG's under business-as-usual scenarios will dominate other heating or cooling perturbations due to natural or other anthropogenic factors during the coming century; (iv) the climate will respond significantly to the projected heating perturbation, with a global mean warming for a CO₂ doubling greater than 1.0 °C a near certainty and warming greater than 1.5 °C highly likely; and (v) there is a significant risk of substantial negative impacts associated with a doubled-CO₂ climate, and this risk will grow as climate warms beyond the doubled-CO₂ climate.

At present, the cooling effect due to sulphur emissions appears to be offsetting about half of the heating effect due to GHG increases, but this partial offsetting will weaken whether or not sulphur emissions are reduced in response to acid rain concerns. Sulphur pollution appears to be responsible for a greater nighttime than daytime warming over midlatitude continents so far, but, even in the absence of cutbacks in sulphur emissions, daily maximum temperature will increase by almost as much as the mean temperature in the long run.

Climatic change associated with as little as a CO₂ doubling poses significant risks to global and regional agriculture, water resources, and to coastal regions (as a result of sea-level rise). With modest warming, some regions are likely to see increases in agricultural yields while other regions are likely to see decreases, but it cannot be predicted with confidence who will gain and who will lose. Two recent major studies suggest that a combination of direct CO₂ effects on photosynthesis and water loss and full adaptation can change a substantial (20%) regional or global loss in agricultural production into a very small loss or a slight increase. However, even when global net effects are small, developing regions are expected to see significant negative impacts. Furthermore, even if the impact of a CO₂ doubling on global food production is small or beneficial in aggregate, there are several reasons for expecting the risk of significant negative effects to grow with increasing climatic warming. In particular, as climate continues to warm, the soil evaporation tendency is likely to increase faster than precipitation over continents, less land will be available along the polar margins of current agricultural areas to offset yield losses along the equatorward margins, the beneficial direct effects of higher CO₂ will tend to saturate, and the frequency with which critical temperature thresholds are exceeded will increase. Stresses on regional water supply might be one of the most important consequences of global warming, particularly since the amount of intense rainfall as a fraction of total rainfall and interannual variability of rainfall are expected to increase in many regions as climate warms.

Anticipated rates of climatic change are expected to cause significant stresses and biomass losses for forests, and to lead to significant wetland losses and damage to coral reef and estuarine ecosystems. Anticipated rates of climatic change in mid-latitudes under business-as-usual scenarios are an order of magnitude greater than the rate at which many forest species can migrate under natural conditions; where suitable forest habitats are highly fragmented due to human activities, rates of forest migration will be even lower. The imbalance between the rates at which climatic zones and forests can shift will lead to forest dieback and impoverishment,

Table II

Summary of the risks associated with unrestrained emissions of greenhouse gases. Risks are subjectively rated as Low (<20%), Moderate (20–60%), High (60–90%), and Very High (>90%)

Risk	Likelihood
Unrestrained CO ₂ emissions lead to several times pre-industrial atmospheric CO ₂ concentration	Very High
Anthropogenic warming effects dominate natural variability and anthropogenic cooling effects over the next century	Very High
Climate is sensitive to GHG increases (>1.5 °C global mean warming for a CO ₂ doubling)	Very High
Large regions in mid-continent become drier in summer on average	High
Significant positive carbon cycle-climate feedback from the terrestrial biosphere occurs	Moderate
Significant positive feedback from the ocean part of the carbon cycle occurs	Low
Other biogeochemical feedbacks constitute a significant net positive feedback	Low
A CO ₂ doubling equivalent provokes:	
– significant declines in agricultural production in some important regions	High
– a non-trivial net decrease in global food production compared to the case without climatic change	Low
– increased precipitation variability and tropical storminess	Moderate
– accelerated species loss	Very High
– decrease in human comfort and welfare (exclusive of agricultural impacts)	Very High in developing countries and for poor members of affluent societies
A CO ₂ doubling equivalent by 2050 provokes:	
– significant negative transient effects on forests	High
– significant negative effects on coastal ecosystems and coral reefs in many regions	High if existing stresses not alleviated; Otherwise, Moderate
– exacerbation of stratospheric O ₃ loss	Moderate (due to CO ₂ increase)

Table II
(Continued)

Risk	Likelihood
Significant GHG increases beyond a CO ₂ doubling equivalent provoke:	
– a significant (>10%) net decrease in global food production compared to the case without climatic change	High
– major (>30%) decreases in food production in some important regions compared to the case without climatic change	High
– significantly greater negative effects on natural ecosystems than for CO ₂ doubling	High
Risk of surprises	Low for <2 °C warming Unknown for larger warming

reducing their recreational, commercial, and biological value. A sea-level rise of 0.5–1.0 m by the end of the next century would lead to significant losses of ecologically important coastal wetlands worldwide. Healthy coral reefs should be able to accommodate a sea-level rise of 1 cm per year, but stresses due to higher temperatures (which can provoke coral bleaching), pollution, or increased UV-radiation, or the impact of increasing CO₂ on ocean chemistry, will reduce the ability of corals to accommodate rising sea level. A significant number of species extinctions can be expected to occur, leading to an impoverishment of the global gene pool.

The worst impacts of global warming are likely to occur in developing regions, which at present are least able to cope with the anticipated climatic changes. Although some developing regions will have experienced significant economic development by the time a doubled-CO₂ climate occurs, which will lessen the human impact of some anticipated changes, populations will also be much larger. Negative impacts are likely to increase sharply as climate continues to warm above the doubled-CO₂ level, and at some point the impacts will almost certainly become so severe that preventive action will be required. Those changes which do occur will be irreversible for all practical purposes, which argues strongly against a wait-and-see approach. Many of the impacts cannot be meaningfully quantified in economic terms, as they involve loss of human life or of biological functioning and diversity.

In short, unrestrained emissions of GHG's pose real and substantial risks to human societies and to ecosystems. These risks clearly justify some action to limit emissions. The magnitude of emission restraint that is justified depends not only on the risks reviewed above, but also on the risks – and non-climatic benefits – associated with measures to limit GHG emissions. These risks and benefits are reviewed in Part II.

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