MIT Joint Program on the Science and Policy of Global Change

Evaluating the Use of Ocean Models of Different Complexity in Climate Change Studies

Andrei P. Sokolov, Stephanie Dutkiewicz, Peter H. Stone and Jeffery R. Scott

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To inform processes of policy development and implementation, climate change research needs to focus on improving the prediction of those variables that are most relevant to economic, social, and environmental effects. In turn, the greenhouse gas and atmospheric aerosol assumptions underlying climate analysis need to be related to the economic, technological, and political forces that drive emissions, and to the results of international agreements and mitigation. Further, assessments of possible societal and ecosystem impacts, and analysis of mitigation strategies, need to be based on realistic evaluation of the uncertainties of climate science.

This report is one of a series intended to communicate research results and improve public understanding of climate issues, thereby contributing to informed debate about the climate issue, the uncertainties, and the economic and social implications of policy alternatives. Titles in the Report Series to date are listed on the inside back cover.

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Abstract

The study of the uncertainties in future climate projections requires large ensembles of simulations with different values of model characteristics that define its response to external forcing. These characteristics include climate sensitivity, strength of aerosol forcing and the rate of ocean heat uptake. The latter can be easily varied over a wide range in an anomaly diffusing ocean model (ADOM). The rate of heat uptake in a three-dimensional ocean general circulation model (OGCM) is, however, defined by a large number of factors and is far more difficult to vary. The range of the rate of the oceanic heat uptake produced by existing Atmosphere-Ocean General Circulation Models (AOGCMs) is narrower than the range suggested by available observations. As a result, simpler models, like an ADOM, are useful in probabilistic climate forecast type studies as they can take into account the full uncertainty in ocean heat uptake.

To evaluate the performance of the ADOM on different time scales we compare results of simulations with two versions of the MIT Integrated Global System Model (IGSM): one with an ADOM and the second with a full three dimensional OCGM. Our results show that in spite of its inability to depict feedbacks associated with the changes in the ocean circulation and a very simple parameterization of the ocean carbon cycle, the version of the IGSM with ADOM is able to reproduce important aspects of the climate response simulated by the version with the OCGM through the 20^{th} and 21^{st} century and can be used to obtain probability distributions of changes in many of the important climate variables, such as surface air temperature and sea level, through the end of 21^{st} century. On the other hand, the ADOM is not able to reproduce results for longer term climate change and specifically those concerned with details of the feedbacks on the heat and carbon storage. Such studies will require the use of the OGCM and uncertainties in those results will be limited.

Contents

1. INTRODUCTION

Projections of climate change over the next century are complicated by significant uncertainties in the climate system properties that determine the response to transient forcing, such as climate sensitivity and the rate at which the deep ocean absorbs heat and CO_2. There are additional uncertainties in the forcing itself, especially in the indirect forcing by aerosols (IPCC, 2001). Unfortunately, the available observations for the 20^{th} century can only place limited constraints on these key quantities (Andronova & Schlesinger, 2002; Gregory *et al.,* 2002; Forest *et al*., 2002, 2006; Frame *et al*., 2005).

Due to these uncertainties, there is no single best climate model or best set of key climate parameters for projecting climate change. A sensible approach is therefore to produce probability distributions for the changes in the most important climate variables. Such probabilistic approaches are also more useful for policy makers than a single model result. However, even with much greater computational power than is available today, it will be difficult to perform such probabilistic studies using state-of-the-art Atmosphere-Ocean General Circulation Models (AOGCMs). In addition to the large computational demand, the use of AOGCMs in probabilistic studies has been restricted by the difficulty in changing the rate of heat uptake in ocean general circulating models (GCMs) (discussed more below); accounting for the uncertainty in the amount of heat taken up by the ocean is crucial for our understanding of possible climate change. Probalistic studies including heat uptake uncertainties have therefore usually been carried out with models of intermediate complexity (*e.g.,* Wigley & Raper, 2001; Knutti *et al.*, 2003, 2005; Webster *et al*., 2003). The recent study with the HadCM3 AOGCM (Collins *et al.,* 2006) only considered the uncertainty in climate sensitivity, and not in heat uptake.

The heat uptake by the deep ocean also in ocean GCMs depends on many factors, including representation of small-scale processes (*e.g.*, Stone, 2004). Dalan *et al.* (2005b) showed that versions of a 3-dimensional (3D) ocean model with different rates of heat uptake can be produced by changing the vertical/diapycnal diffusion coefficients. However, changing the diffusion coefficient alters the ocean circulation as a whole, in particular the strength of North Atlantic overturning (Dalan *et al.,* 2005a). It appears almost impossible (certainly without changes to parameterizations in the 3D models) to vary the heat uptake over the full range suggested by climate change observations during the 20th century (Forest *et al.,* 2006) and to maintain a reasonable circulation. It is worth noting that while different AOGCMs differ significantly in the rate of heat uptake (IPCC, 2001; Raper *et al.,* 2002; Sokolov *et al.,* 2003) they also do not cover the range suggested by observations. Studies based on multi-model ensembles (*e.g.,* Covey *et al.,* 2003) therefore do not take into account the full uncertainty in the rate of heat uptake.

The MIT Integrated Global System Model (IGSM), described by Prinn *et al.* (1999) and updated in Sokolov *et al.* (2005), was designed to be used in a probabilistic framework. The IGSM provides the flexibility and computational speed required for uncertainty analysis while still including the representations for all the major components of the climate system.

The IGSM consists of a 2-dimensional (2D) (zonally averaged) statistical-dynamical atmospheric model with interactive chemistry coupled to a model of terrestrial ecosystem and an ocean model. In the first version of the IGSM (IGSM1, Prinn *et al.,* 1999), the oceanic component of the climate system was represented by a zonally averaged mixed layer anomaly-diffusing ocean model (ADOM) (Hansen *et al.,* 1984; Sokolov & Stone, 1998). The second version of the IGSM (IGSM2, Sokolov *et al.,* 2005) was developed in two different configurations: with either a two-dimensional (latitude-longitude) ADOM (IGSM2.2) or with a three-dimensional ocean GCM (IGSM2.3). The ADOM has several advantages: it is computationally efficient and it is flexible. The rate of heat mixing into the deep ocean can be varied over a wide range just by changing the coefficient of effective diffusion of the heat anomalies.

As shown by Sokolov *et al.* (2003), the MIT 2D climate model with the mixed layer/ADOM can (with an appropriate choice of the vertical diffusion coefficient and climate sensitivity) simulate the behavior of different coupled AOGCMs, in terms of surface warming and sea level rise, on time scales of about 100-150 years. The simple anomaly-diffusing ocean model works well because the mixing of the heat into the deep ocean is a linear response to the forcing on century time scales in typical global warming simulations (*e.g.,* Keen & Murphy, 1997; Huang *et al.,* 2003). Thus the mixed layer/ADOM seems to be an appropriate tool for obtaining probability distributions for uncertain climate parameters (Forest *et al.,* 2002, 2006; Frame *et al.,* 2005), as well as studying uncertainty in possible climate change for time scales from a few decades to a century (Webster *et al.,* 2003). The ADOM coupled to the GISS AGCM has been used for simulating both past and future climate for a number of years (*e.g.,* Hansen *et al.,* 1988, 2002).

However, in some cases much longer simulations are required to fully evaluate the impact of proposed economic policies, for instance stabilization of atmospheric GHG concentrations. Feedbacks associated with changes in the ocean circulation, not simulated by the ADOM, may become crucially important on the longer time scales. The goal of this study is to investigate on what time scales a simplified ocean model can capture the climate response of a 3D model.

The model components, and especially the difference in the two versions of the ocean, are described in Section 2. Section 3 provides a comparison of results between the IGSM2.2 and IGSM2.3 for different future emission scenarios, for different climate sensitivities, and for different time scales. Conclusions are provided in Section 4.

2. MODEL COMPONENTS

The IGSM is a fully coupled model of the Earth climate system that allows simulation of critical feedbacks between components. The second version of the IGSM (IGSM2, Sokolov *et al.,* 2005) includes the following components:

- An atmospheric dynamics, physics and chemistry model, which includes a sub-model of urban chemistry,
- Either mixed layer/ADOM, or 3D general circulation ocean model, both with carbon cycle and sea ice sub-models,
- A set of coupled land models, the Terrestrial Ecosystem Model (TEM), a Natural Emissions Model (NEM), and the Community Land Model (CLM), that encompass the global, terrestrial water and energy budgets and terrestrial ecosystem processes.

The time steps used in the various sub-models range from 20 minutes for atmospheric dynamics to 1 month for TEM, reflecting differences in the characteristic timescales of the various processes simulated by the IGSM. The atmospheric and ocean sub-model are briefly described below. Descriptions of the other components of the IGSM2 can be found in Schlosser *et al.*

(2006), Liu (1996), Wang *et al.* (1998), Wang (2004), and Xiao *et al.* (1997, 1998). A comparison between the old version, IGSM1, and the newer version, IGSM2 can be found in Sokolov *et al.* (2005).

2.1 Atmospheric Dynamics and Physics

The MIT two-dimensional (2D) atmospheric dynamics and physics model (Sokolov & Stone, 1998) is a zonally averaged statistical-dynamical 2D model that explicitly solves the primitive equations for the zonal mean state of the atmosphere and includes parameterizations of heat, moisture, and momentum transports by large scale eddies based on baroclinic wave theory (Stone & Yao, 1987, 1990). The model's numerics and parameterizations of physical processes, including clouds, convection, precipitation, radiation, boundary layer processes, and surface fluxes, are built upon those of the Goddard Institute for Space Studies (GISS) GCM (Hansen *et al.,* 1983). The radiation code includes all significant greenhouse gases (H_2O, CO_2, CH_4, N_2O, CFCs and O_3) and eleven types of aerosols. The model's horizontal and vertical resolutions are variable, but in the standard version of IGSM2 it has 4° resolution in latitude and eleven levels in the vertical.

The MIT 2D atmospheric dynamics and physics model allows up to four different types of surface in each grid cell (ice free ocean, sea-ice, land, and land-ice). The surface characteristics (*e.g.*, temperature, soil moisture, albedo) as well as turbulent and radiative fluxes are calculated separately for each kind of surface. The atmosphere above is assumed to be well-mixed horizontally in each latitudinal band. The area weighted fluxes from the different surface types are used to calculate the change of temperature, humidity, and wind speed in the atmosphere. The atmospheric model's climate sensitivity can be changed by varying the cloud feedback (Sokolov & Stone, 1998; Sokolov, 2006).

2.2 Ocean Component

In the older IGSM1 (Prinn *et al.,* 1999), a zonally (longitudinally) averaged mixed layer ocean model with 7.8° latitudinal resolution was used. In the new IGSM2 the ocean component has been replaced by either a two-dimensional (latitude-longitude) mixed layer anomaly-diffusing ocean model (hereafter denoted as IGSM2.2) or a fully three-dimensional ocean GCM (denoted as IGSM2.3).

2.2.1 The Three-Dimensional (3D) Ocean General Circulation Model

The 3D ocean component is a major advance in the capabilities of the IGSM. The IGSM1 atmospheric model (with lower resolution than in the IGSM2) had previously been coupled to the Modular Ocean Model 2 (MOM2) ocean GCM for studies of ocean response to climate change (Kamenkovich *et al.,* 2002, 2003; Dalan *et al.,* 2005a, b; Huang *et al.,* 2003a, b). This version has also been used in a number of model intercomparison studies (Gregory *et al.,* 2005; Petoukhov *et al.,* 2005; Stouffer *et al.,* 2005). However, as detailed by Dutkiewicz *et al.* (2005), the 3D ocean-sea ice-carbon cycle component of the IGSM2.3 is now based on the 3D MIT ocean general circulation model (Marshall *et al.,* 1997a, b). As configured for the IGSM2.3, the

MIT ocean model has realistic bathymetry, and 4° x 4°resolution in the horizontal with fifteen layers in the vertical (ranging from 50m at the surface to 500m thick at depth). Mesoscale eddies, which are not captured in this coarse resolution, are represented by the Gent & McWilliams (1990) parameterization. Embedded in the ocean model is a thermodynamic sea-ice model based on the 3-layer model (two ice layers and a snow layer) of Winton (2000) and the LANL CICE model (Bitz & Lipscome, 1999).

The ocean model has a biogeochemical component with explicit representation of the cycling of carbon, phosphate, dissolved organic phosphorus, and alkalinity. The physical ocean model velocities and diffusion are used to transport these tracers; in addition chemical and biological processes are parameterized. Air-sea exchange of carbon dioxide follows Wannikof (1992), and carbonate chemistry is calculated following Najjar & Orr (1998), Millero (1995), and the DOE Handbook (1994). There is also a parameterization of the export of organic carbon from the surface waters: biological productivity is modelled as a function of available nutrient (phosphate) and photosythetically available radiation (see Dutkiewicz *et al.,* 2005). A fraction of the biological production in the sunlit surface layers enters a dissolved organic pool that has an *e*-folding timescale of remineralization of 6 months (following Yamanaka & Tajika, 1997). The remaining fraction of the productivity is instantaneously exported as particulate matter to depth (Yamanaka & Tajika, 1996), where it is remineralized according to the empirical power law relationship of Martin *et al.* (1987). There is also a representation of the calcium carbonate cycle following the parameterization of Yamanaka & Tajika (1996).

The coupling between the atmospheric and 3D oceanic sub-models takes place once a day. The atmospheric model calculates 24-hour averaged surface heat, freshwater and momentum fluxes, and passes these to the ocean model. After receiving these fluxes, the ocean and sea ice sub-models are integrated for 24 hours (two ocean tracer time steps). At the end of this period, sea surface temperatures, surface sea ice temperatures, and sea ice coverage are passed back to the atmospheric sub-model.

The atmospheric sub-model provides heat and fresh-water fluxes separately for open ocean and sea ice, as well as derivatives with respect to surface temperature. Total heat and freshwater fluxes for the oceanic sub-model can therefore vary by longitude as a function of ocean sea surface temperature, *i.e.* warmer ocean locations undergo greater evaporation and receive less downward heat flux. The heat flux (F_H) at the longitude-latitude point (i, j) is calculated as:

$$F_H(i,j) = F_{HZ}(j) + \frac{\partial F_{HZ}}{\partial T}(j)(Ts(i,j) - Tsz(j)), \qquad (1)$$

where $F_{HZ}(j)$ and ${\partial F_{HZ}}/{\partial T}(j)$ are the zonally averaged heat flux and its derivative with respect to surface temperature, and $Ts(i,j)$ and $Tsz(j)$ are the surface temperature and its zonal mean.

Fluxes of sensible and latent heat are calculated in the atmospheric model by bulk formulas with turbulent exchange coefficients dependent on the Richardson number. The atmosphere's turbulence parameterization is also used in the calculation of the flux derivatives with respect to surface temperature. To account for partial adjustment of near surface air temperature to changes in fluxes, the derivatives are calculated under the assumption that the exchange coefficients are fixed. A more detailed discussion of technical issues involved in the calculations of these fluxes and their derivatives, is given in Kamenkovich *et al.* (2002).

Wind stresses from the atmospheric sub-model are weaker than observations, especially in the Southern Ocean. The oceanic sub-model therefore uses the technique of anomaly coupling: the mean wind stresses, including zonal variations, are taken from the climatology of Trenberth *et al.* (1989), while the anomalies are taken from the atmospheric sub-model. The oceanic sub-model requires adjustments to the atmospheric heat and freshwater fluxes in order to replicate the ocean sea surface temperature and salinity for the later part of the 20th century. The adjustments are calculated during the initial stage of the ocean sub-model spin-up in which sea surface temperature (SST) and sea surface salinity (SSS) are relaxed toward observed values. These adjustments are then held fixed for a pre-industrial (1860) spin-up of several thousand years and then are also held fixed for the 1860-onward simulation. In this 3D configuration, the ocean-carbon-atmospheric component must be spun-up for several thousand years to reach a pre-industrial (1860) steady state for each set of model parameters (*e.g.*, vertical diffusivity). More details of the ocean-carbon-sea ice sub-model and its coupling to the atmosphere are provided by Dutkiewicz *et al.* (2005).

2.2.2 The Two-Dimensional Mixed Layer Anomaly Diffusing Ocean Model

The ocean component of the IGSM2.2 consists of an Q-flux model of an upper ocean layer with horizontal resolution of 4° in latitude and 5° in longitude, and a 3000m deep anomaly diffusing ocean model beneath. The upper ocean layer is divided into two sub-layers, namely a mixed layer and the layer between the mixed layer depth and its annual maximum (seasonal thermocline). The mixed layer depth is prescribed based on observations as a function of season and location (Hansen *et al.,* 1983). In addition to the temperature of the mixed layer, the model also calculates the averaged temperature of the seasonal thermocline and the temperature at the annual maximum mixed layer depth (Russell *et al.,* 1985). Changes in the temperature of the two layers due to the increase/decrease of the mixed layer depth are calculated under the assumption that temperature is constant throughout the mixed layer and changes linearly with depth in the seasonal thermocline. Temperature at the bottom of the seasonal thermocline is updated when mixed layer depth reaches its annual maximum. In contrast with conventional upwelling–diffusion models, diffusion in the model used here is not applied to temperature itself but to the temperature difference from its values in a present-day climate simulation (Hansen *et al.,* 1984; Sokolov and Stone, 1998). Since this diffusion represents a cumulative effect of heat mixing by all physical processes, the values of the diffusion coefficients are significantly larger than those used in sub-grid scale diffusion parameterizations in OGCMs. The spatial distribution of the diffusion coefficients used in the diffusive model is based on observations of tritium mixing into the deep ocean (Hansen *et al.,* 1984).

The coupling between the atmospheric and oceanic components takes place every hour. Surface fluxes required to force ocean and sea ice models are calculated in the way described in previous section (Equation 1).

The mixed layer model also includes a specified vertically-integrated horizontal heat transport by the deep oceans, a so-called "Q-flux", allowing zonal as well as meridional transport. This flux is calculated from a simulation in which sea surface temperature (SST) and sea ice distribution are relaxed toward their present-day climatology with relaxation coefficient of 300 $W/m^2/K$, which corresponds to an *e*-folding time scale of about 15 days for a 100 m deep mixed layer. Monthly averaged values of Q-flux were used as a flux adjustment in 20th century and future climate simulations. Relaxing SST and sea ice on such short time scale, while being virtually identical to specifying them, avoids problems with calculating the Q-flux near the sea ice edge. The use of a two-dimensional (longitude-latitude) mixed layer ocean model instead of the zonally averaged one used in IGSM1 results in an improved simulation of both the present day sea ice distribution and sea ice changes in response to increasing radiative forcing (Sokolov *et al.,* 2005). Taking into account the seasonal cycle and latitudinal distribution of the mixed layer depth allows better simulations of the asymmetry in the transient changes in sea surface temperature. We understand that this distribution can change in the future, but lacking a basis for making predictions, we use present-day climate distribution in all simulations.

A thermodynamic ice model is used for representing sea ice. This model has two ice layers and computes ice concentration (the percentage of area covered by ice) and ice thickness.

The IGSM2.2 includes a significantly modified version of the ocean carbon model (Holian *et al.,* 2001) used in the IGSM1. Formulation of carbonate chemistry (Follows *et al.*, 2006) and parameterization of air-sea fluxes in this model are similar to the ones used in the IGSM2.3. Vertical and horizontal transports of the total dissolved inorganic carbon, though, are still parameterized by diffusive processes. The values of the horizontal diffusion coefficients are taken from Stocker *et al.* (1994), and the coefficient of vertical diffusion of carbon (K_{vc}) depends on the coefficient of vertical diffusion of heat anomalies (K_v). In IGSM1, K_{vc} was assumed to be proportional to K_v (Prinn *et al.,* 1999; Sokolov *et al.,* 1998). This assumption, however, does not take into account the vertical transport of carbon due to the biological pump. In the IGSM2.2 K_{vc} is defined as follows:

$$K_{vc} = K_{vco} + rK_v \tag{2}$$

where the values of K_{vco} and r were estimated by comparing results of the simulations with the IGSM2.2 and IGSM2.3 (see section 3.1 for details).

Since K_{vco} is a constant, the vertical diffusion coefficients for carbon have the same latitudinal distribution as coefficients for heat. For simulations with different rates of oceanic uptake, the diffusion coefficients are scaled by the same factor in all locations. Therefore rates of both heat and carbon uptake by the ocean are defined by the global mean value of the diffusion coefficient for heat. In the rest of the paper the symbol K_v is used to designate the global mean value.

Preliminary comparison (see Appendix) has shown that the assumption that changes in ocean carbon can be simulated by the diffusive model with fixed diffusion coefficient, as used in Holian *et al.* (2001), works only for about 150 years (most likely due to relatively small changes in the atmospheric CO_2 concentration). On longer timescales the simplified carbon model severely overestimates that ocean carbon uptake. However, if K_{vc} is assumed to be time dependent, the IGSM2.2 reproduces changes in ocean carbon simulated by the IGSM2.3 on multi century scales (see Appendix for details). Thus, for the runs discussed here, the coefficient for vertical diffusion of carbon was calculated as:

$$K_{vc}(t) = (K_{vco} + rK_v) \cdot f(t) \qquad (3)$$

where f(t) is a time dependent function discussed in the Appendix as illustrated in Figure A3.

3. SIMULATIONS OF THE PAST AND FUTURE CLIMATE

As discussed previously, there are significant uncertainties in the characteristics of climate models defining their response to changes in radiative forcing. For obtaining probability distributions for the future climate a large number of climate change simulations must be carried out. For example, the distributions presented by Webster *et al.* (2003) are based on the results of 250 simulations with different values of climate sensitivity, strength of aerosol forcing and the rate of oceanic heat and carbon uptake.

Carrying out such an ensemble requires knowledge of the probability distributions for the above mentioned climate characteristics. Such distributions can be produced by comparing observed temperature changes with results of 20th century simulations in which these model parameters were varied. **Figure 1** shows the marginal, that is integrated over all possible values of climate sensitivity and aerosol forcing strength, probability density function for the rate of heat uptake by the ocean (measured by the square root of the effective diffusion coefficient) suggested by observations (Forest *et al.*, 2006).

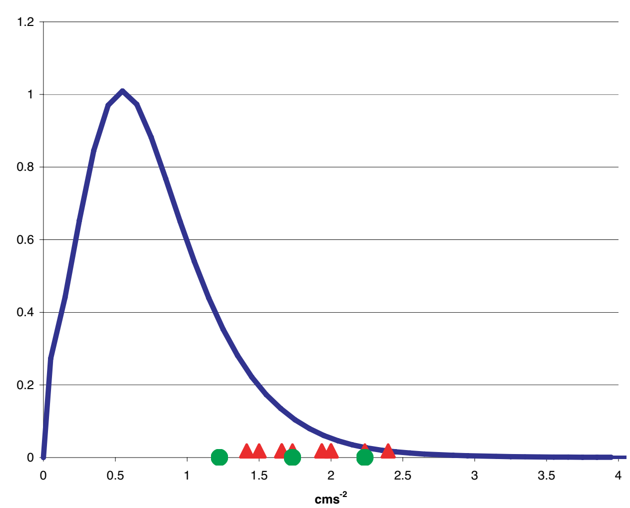


Figure 1. Probability density function for the square root of the effective anomaly diffusion coefficient from Forest *et al.* (2006) and values corresponding to the three versions of the IGSM2.3 with different values of vertical diffusion coefficient (large dots) and some AOGCMs (triangles).

The IGSM2.3 version was spun up to steady state equilibrium for 1860 atmospheric composition with three different values for ocean vertical diffusion coefficients, K_z, namely 0.2, 0.4 and 0.6 cm^2/sec. Actual measurements of *in-situ* vertical diffusivity in the ocean are rather sparse, although they suggest rather weak values, typically less than 0.2 cm^2/sec, with greatly elevated diffusivities in the vicinity of rough topography (Ledwell *et al.,* 1993, 2000; Polzin *et al.,* 1997). Schneider and Bhatt (2000) used a volume integrated "dissipation method" to estimate the vertical diffusivities, also finding O (0.1 cm^2/sec) for much of the ocean except the warmest ocean water masses. On the other hand, mixing from other sources such as boundary mixing (Moum *et al.,* 2002) and mixing from tropical storms (Stiver & Huber, 2007) remain poorly constrained. Global mean values of the effective diffusion coefficients (K_v) for the IGSM2.2 required to match the behavior of these three versions of the IGSM2.3 were determined from simulations with 1% per year increase in the atmospheric CO_2 concentration using the approach described in Sokolov *et al.* (2003) and are shown in the Figure 1 by green circles. The results from the 3D ocean (consistent with other AOGCMs) are in the upper part of the range for the rate of the heat uptake that is consistent with observations.

Changes in the diffusion coefficient in the 3D ocean model affect the ocean circulation as a whole: the strength of the meridional overturning circulation (MOC) in the North Atlantic in the simulations with these three versions is 9, 14 and 17 S_v, respectively. Further decrease in the vertical diffusion coefficient in our specific model setup leads to an even weaker MOC. A smaller diffusion coefficient would be possible with additional changes in the structure of the ocean model; however, such manipulations still do not guarantee a decrease in the rate of heat uptake. Heat uptake rates for some of the AOGCMs used in the AR4 IPCC simulations are shown in Figure 1 by triangles: All of them lie above the median of the distribution suggested by observations. These models have significant differences in the parameterizations and values of parameters, and some have smaller vertical diffusions.

Data required to estimate K_v values for other model (namely, sea level change) are not available, but judging from values of the heat uptake efficiencies for those models, the corresponding values of K_v are likely to be larger than 1cm^2/sec.

The fact that K_v values for all AOGCMs are larger than the median of the marginal PDF suggests that more than a half of the values of effective vertical diffusivity that cannot be rejected by available observations are smaller that values suggested by existing AOGCMs. This, in its turn, suggests that the results of uncertainty studies based on the range of the rates of oceanic heat uptake suggested by AOGCMs will be biased. On the other hand, for any values of K_v falling in the range of acceptable values it is possible to find values of climate sensitivity and aerosol forcing strength that will produce results consistent with observations (see section 3.2).

The climate response of the MIT climate model with ADOM was previously compared with the responses of different coupled AOGCMs only in simulations with prescribed changes in atmospheric CO_2 concentration. The existence of two versions of the IGSM2 that differ only by the ocean sub-component allows us to conduct a more detailed comparison and to define time scales on which a mixed layer anomaly-diffusing ocean model can reproduce behavior of the more sophisticated 3D ocean model.

3.1 Simulations Design

The climate change simulations shown here start in 1861 from the end of the corresponding spin up simulation and are conducted in two stages: a simulation with historical forcings and a future climate projection. During the first stage, from 1861 to 1990, the model is forced by the observed changes in GHG concentrations (Hansen *et al.,* 2002), tropospheric and stratospheric ozone (Wang & Jacob, 1998), the solar constant (Lean, 2000), sulfate aerosols (Smith *et al.,* 2004), and volcanic aerosols (Sato *et al.,* 1993). For this historical forcing stage, carbon uptake by the ocean and terrestrial ecosystems are calculated but not fed back to the atmospheric model. Based on data for anthropogenic carbon emissions and atmospheric CO_2 concentrations, the net land plus ocean carbon uptake should equal about 4.1 GtC/year for the 1980s. In these experiments, the difference between the model actual total land-ocean uptake and this observed value is determined and this additional sink/source is then kept constant during the subsequent forward stage of the simulations. The use of the fixed additional term for carbon uptake is inappropriate in long-term simulations and something we plan to remedy in future model simulations. However, since magnitudes of additional terms in corresponding simulations with two different versions of the model are very close, it does not affect comparison presented in this paper.

In the second-stage of the simulations, which begins in 1991, the full version of IGSM2 is forced by the GHG emissions. Historical GHG emissions are used through 1996 and emissions projected by the MIT Emissions Predictions and Policy Analysis model (Paltsev *et al.,* 2005) from 1997. In this future climate stage of the simulations, all components of the IGSM2 were fully interactive; concentrations of all gases and aerosols were calculated by the atmospheric chemistry sub-model based on anthropogenic and natural emissions and the terrestrial and oceanic carbon uptake provided by the corresponding sub-components.

In this study, two different emission scenarios are used: a "reference" no policy case (Paltsev *et al.,* 2005) and a "stabilization" scenario. In the first scenario (REF) GHGs emissions grow at a rather high rate up to year 2100. To compare model responses under strong forcing, simulations with this scenario were continued until year 2200 with emissions being fixed at their 2100 values. In the second case (STAB), emissions were constructed to ensure stabilization of different GHGs and thereby radiative forcing over a few hundreds years. Simulations with the stabilization scenario were carried out through year 2400. CO_2 emissions for these two scenarios are shown in **Figure 2**.

For a thorough comparison of the climate responses simulated by the IGSM2.2 and IGSM2.3, three simulations with each version of the IGSM were conducted for each of the two emission scenarios. These three simulations differ in parameter choices, as shown in **Table 1**. Climate sensitivities (S) and the strengths of the aerosol forcing were chosen so as to ensure consistency between simulated and observed climate for 20th century (**Figure 3** and Table 1).

Values of the parameters in the equation for K_{vc} (Equation 2) were estimated so as to ensure consistency of the oceanic carbon uptakes in the historical stage of simulations with the versions of the IGSM2.2 and the IGSM2.3 with similar rates of heat uptake (**Table 2**). The values of K_{vco} and r that satisfy this requirement are 2.85 $cm^2\ s^{-1}$ and 0.6 respectively.

In spite of the fact that all versions of the IGSM2.3 were spun up for 3000 years with the pre-industrial conditions, there are small drifts in the deep ocean temperature, salinity and, therefore, sea level. In all results discussed below these drifts are removed.

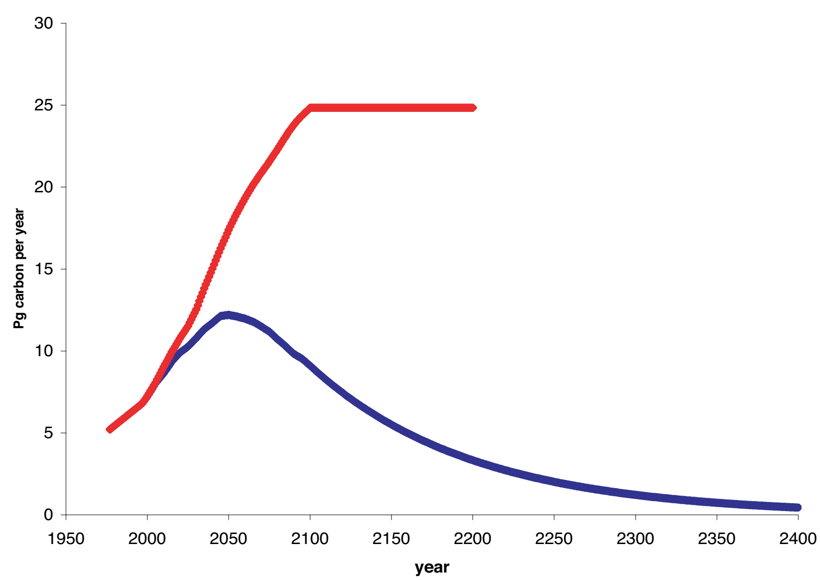
Table 1. Parameters settings in the simulations with the IGSM2.2 and IGSM2.3.

Simulation	Model	Emission scenario	Climate sensitivity (°C)	Aerosol forcing for 1980s ($w\ m^{-2}$)	Vertical diffusion coefficient ($cm^2\ s^{-1}$)	Effective diffusion coefficient ($cm^2 s^{-1}$)
REF31	IGSM2.3	REF	1.5	-0.10	0.2	
REF32	IGSM2.3	REF	2.0	-0.35	0.4	
REF33	IGSM2.3	REF	3.0	-0.70	0.6	
REF21	IGSM2.2	REF	1.5	-0.10		1.5
REF22	IGSM2.2	REF	2.0	-0.35		3.0
REF23	IGSM2.2	REF	3.0	-0.70		5.0
STAB31	IGSM2.3	STAB	1.5	-0.10	0.2	
STAB32	IGSM2.3	STAB	2.0	-0.35	0.4	
STAB33	IGSM2.3	STAB	3.0	-0.70	0.6	
STAB21	IGSM2.2	STAB	1.5	-0.10		1.5
STAB22	IGSM2.2	STAB	2.0	-0.35		3.0
STAB23	IGSM2.2	STAB	3.0	-0.70		5.0

Table 2. Oceanic carbon uptake (GtC/year) averaged over years 1981-1990 in the simulations with the IGSM2.2 and IGSM2.3.

K_z/K_v	IGSM2.3	IGSM2.2
0.2/1.5	1.61	1.60
0.4/3.0	1.69	1.69
0.6/5.0	1.87	1.83
IPCC 2001	1.9 ± 0.6	

Figure 2. CO_2 emissions (in Pg of carbon) in the reference (red line) and stabilization (blue line) simulations.



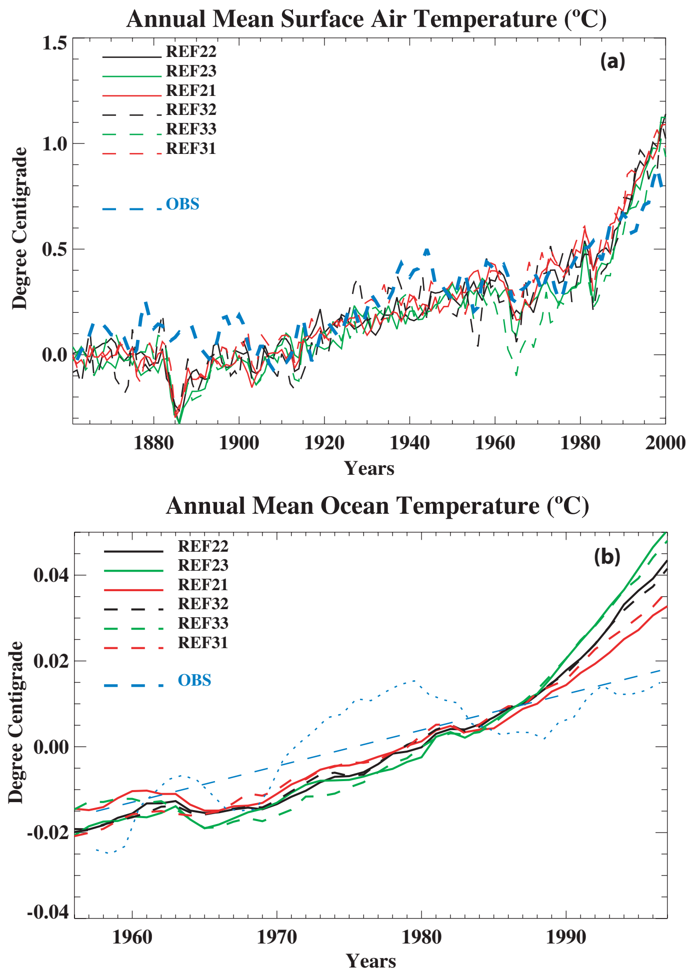


Figure 3. Changes in global mean annual mean surface air temperature **(a)** and ocean temperature for top 3000 meters **(b)** in simulations with IGSM2.2 and IGSM2.3. Observations are from Jones, 2003 and Levitus *et al.* (2005), respectively. Dotted line on the Figure 3b shows five-year means and dashed line estimated linear trend for observations.

3.2 Results

The two versions of the IGSM produce similar warming trends in the historical forcing stage (Figure 3), illustrating that the IGSM2.2 matches the response of the IGSM2.3 in simulations with multiple anthropogenic and natural forcings.

The IGSM2.2 reproduces reasonably well the changes in the annual global mean surface air temperature (SAT) projected by the IGSM2.3 for all combinations of parameters and for both emission scenarios (**Figure 4**). SAT increases predicted by the two versions of the model agree in the corresponding simulations within 0.5°C. Moreover, zonally averaged distributions of changes in temperature in the last decade of 21st century as simulated by IGSM2.2 and IGSM2.3 (**Figure 5**a and 5c) are overall very close except in the polar regions, where slightly different changes in sea ice cover cause corresponding differences in temperature changes. These zonal distributions are shown for the simulations with reference emission scenarios. Differences are even smaller in the stabilization simulations where the forcing is weaker. The differences in sea ice cover (Figure 5b) are, to a large part, related to differences in how the flux adjustment is calculated in the IGSM2.2 and IGSM2.3. In the spin-up simulation with the ISGM2.2 both sea surface temperature and sea ice are relaxed toward the observations, while in the IGSM2.3 spin-up relaxation is applied only to temperature and only from 60°S to 60°N. Sea ice sub-models used in the IGSM2.2 and IGSM2.3 are also different. As a result, sea ice cover in the equilibrium pre-industrial climate simulations with the IGSM2.2 and IGSM2.3, as well as sea ice changes in the simulations discussed here, are somewhat different. In all simulations with the reference emission scenario the IGSM2.2 produces noticeably larger decrease in sea ice cover (Figures 5b and 6a). In the stabilization case large differences occur only between the simulations with high climate sensitivity (**Figure 6**b).

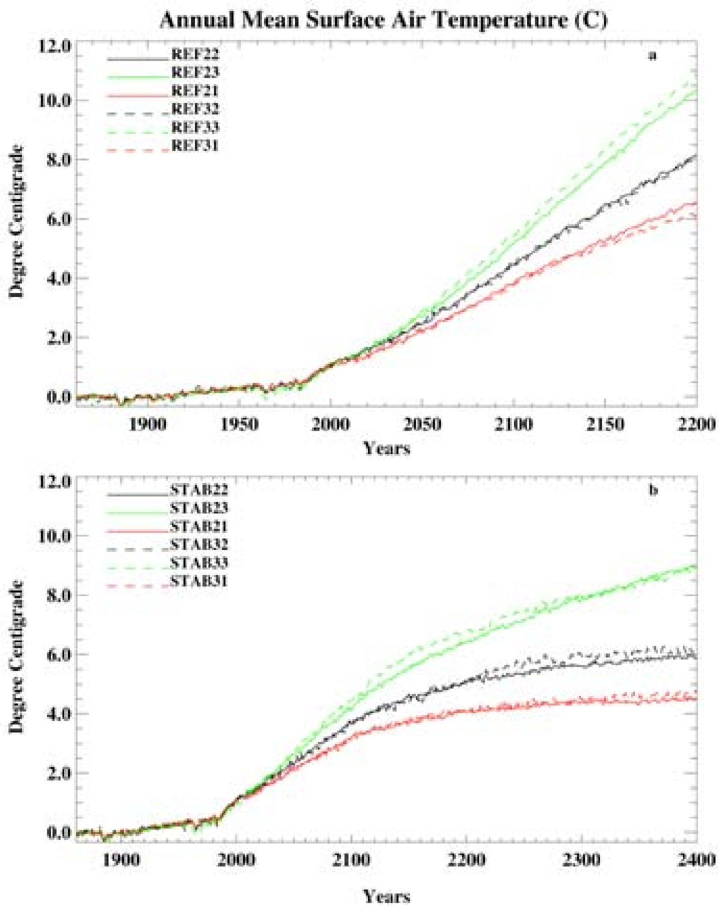


Figure 4. Changes in global mean annual mean surface air temperature in simulations with reference **(a)** and stabilization **(b)** emission scenarios.

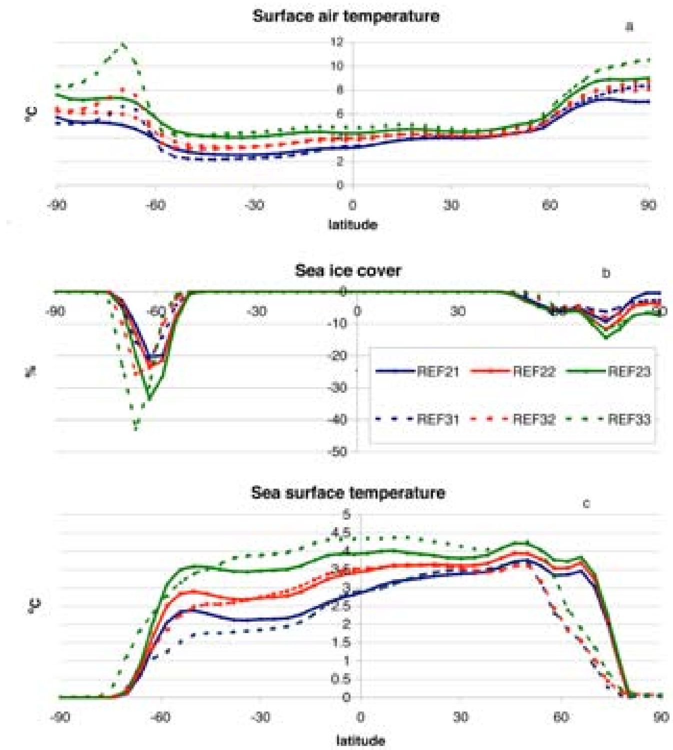


Figure 5. Changes in zonally averaged surface air temperature **(a)**, sea ice cover **(b)**, and sea surface temperature **(c)** in the simulations with reference emission scenario. Difference between decadal means 2091-2100 and pre-industrial equilibrium climate.

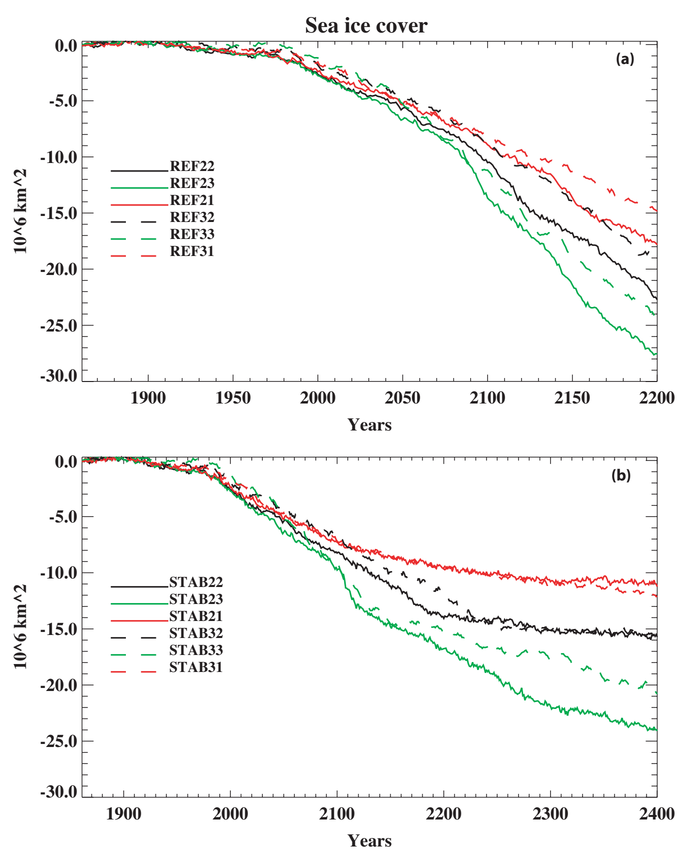


Figure 6. The same as Figure 4, but for sea ice cover.

There is a good agreement in sea level rise due to thermal expansion of the ocean as projected by the two versions of the IGSM for the historical forcing stage, and for about 100 years of the future forcing stage. By year 2100 the two versions differ by less than 2 cm in the corresponding simulations (**Figure 7**). However, the IGSM2.2 begins to overestimate the increase in sea level after about year 2150. At the end of the simulations with the stabilization emissions scenario, sea level rise simulated by the IGSM2.2 is about 20-25% larger that that simulated by the IGSM2.3.

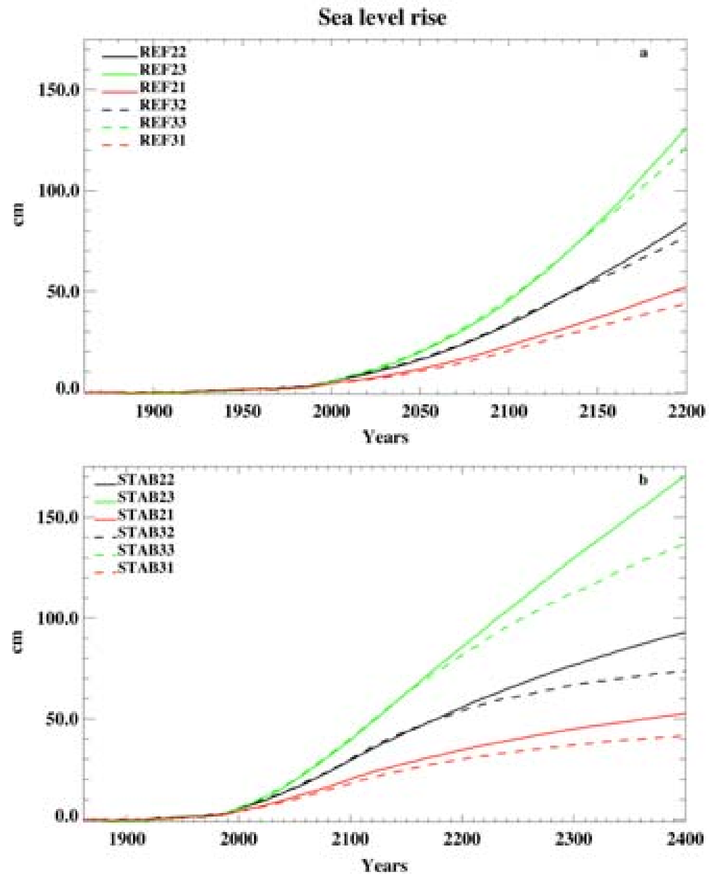


Figure 7. The same as Figure 4, but for sea level rise due to thermal expansion.

Figure 8 shows the zonally averaged temperature changes with depth in the STAB22 (left column) and STAB32 (right column) simulations. In spite of its simplicity, the ADOM reproduces ocean temperature changes from the pre industrial equilibrium state simulated by the 3D ocean model through the middle of the 21st century. However by year 2100 the structure of the deep ocean warming for the two versions of the model begins to look quite different. The IGSM2.2 overestimates the depth of the warming at high latitudes more severely later in the integration. During the second half of the 21st century, an excessive warming at high latitudes is compensated by an underestimate of the warming of tropical waters to 3000m; this compensation leads to good agreement in the global sea-level rise (Figure 7). However, later there is no longer sufficient compensation and an increasing overestimate of the sea-level rise occurs. It should be noted that zonal distribution of changes in sea level simulated by two versions of the IGSM agree with each other for much shorter time than global means. Latitudinal structure in the deep ocean temperature changes simulated by the IGSM2.2 is defined by the distribution of the diffusion coefficients which, as noted in section 2.2.2, were derived from observations on tritium mixing (Hansen *et al.*, 1984). As such, values of the diffusion coefficients in high latitude are an order of magnitude larger than in the equatorial region.

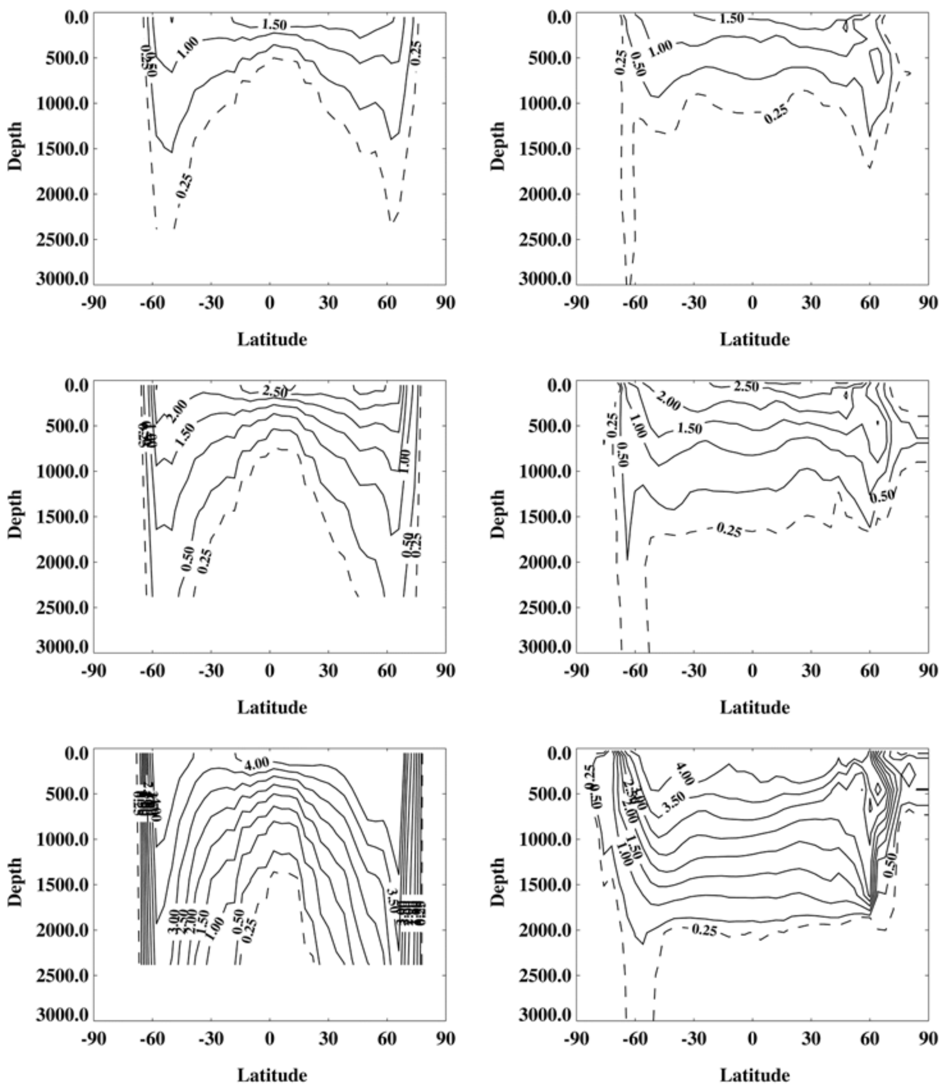


Figure 8. Changes in zonally averaged ocean temperature from the pre industrial equilibrium averaged over years 2041-2050 (top), 2091-2100 (middle) and 2291-2300 (bottom), in simulations STAB22 (left) and STAB32 (right).

It is interesting to note that mixing of heat to the deep ocean can be approximated by the diffusion of mixed layer temperature anomalies in spite of significant changes in the strength of the Atlantic meridional overturning circulation occurring during 21st century (**Figure 9**). On the long time scales, though, this approximation does break down.

As shown in the Appendix, IGSM2.2 with a time dependent coefficient for the vertical diffusion of carbon well reproduces the changes in the carbon uptake by the ocean and oceanic carbon storage produced by the IGSM2.3 in the simulations with prescribed changes in the atmospheric CO_2 concentration. The agreement between the results of the two versions of the IGSM2 remains good in the simulations with an interactive carbon cycle. Atmospheric CO_2 concentrations simulated by the IGSM2.2 and IGSM2.3 are almost identical through year 2200 in all simulations (**Figure 10**). Even in 2400 the difference between the two models does not exceed 50 ppm.

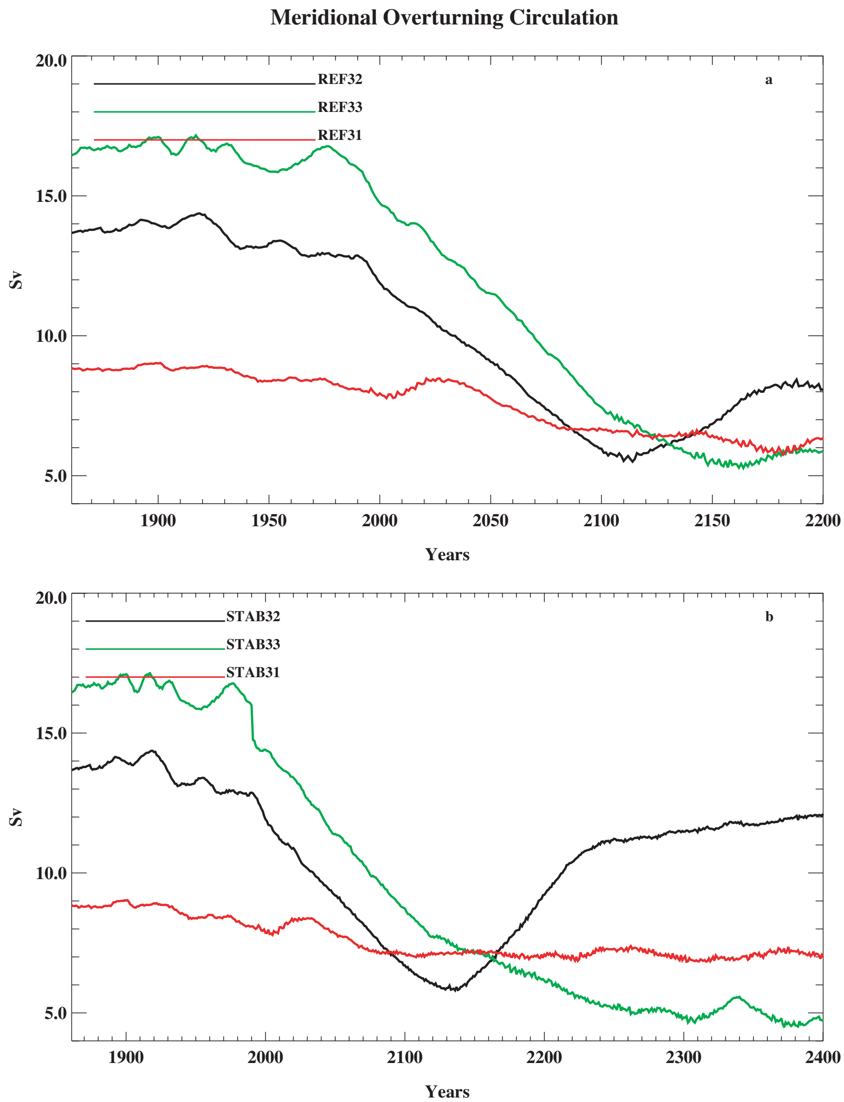


Figure 9. The same as Figure 4, but for the maximum value of the Atlantic meridional overturning circulation.

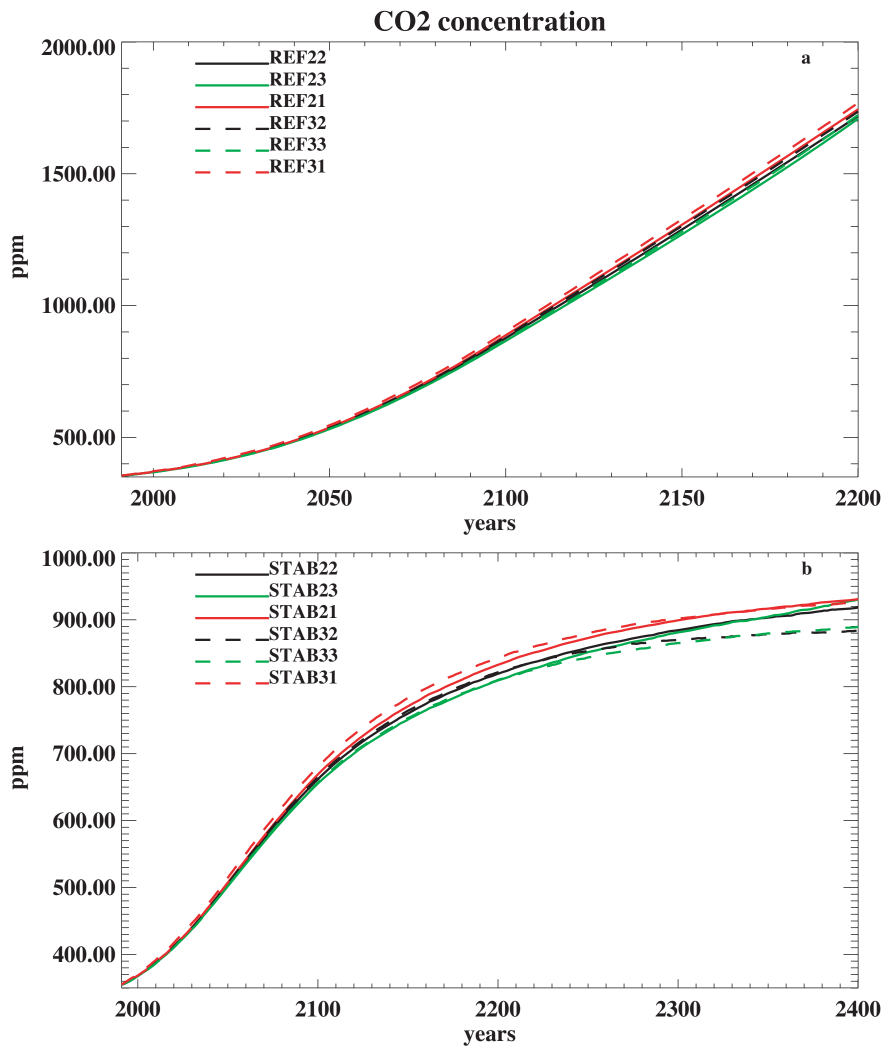


Figure 10. The same as Figure 4, but for atmospheric CO_2 concentration.

4. CONCLUSIONS

The MIT IGSM was designed to provide probabilistic forecasts of future climate under different GHG emission scenarios. Such forecasts require large ensembles of climate change simulations in which climate system parameters, including the rate of heat and carbon uptake by the ocean, must be varied. Changing the rate of ocean heat uptake over a wide enough range is easy to achieve in an ADOM but is rather difficult in the full 3D ocean GCM. Our 3D model, similar to other models, takes up heat at rates lying in the upper part of the range suggested by observations (Forest *et al.,* 2006). Changes in the 3D model parameters that would lead to decreases in the rates to the lower part of the range consistent with observations produce unrealistic ocean circulations. There appears to be no easy remedy to this problem with the 3D models given their current ocean parameterizations. Thus we find it necessary to use the simpler 2D ADOM in studies aimed at making probabilistic climate change forecasts that take into account the full uncertainty.

The goal of this study was to define time scales for which the mixed layer anomaly diffusive ocean model is able to capture the climate response of the 3D ocean GCM.

Comparison of climate change simulations with the two versions of the IGSM2 shows that the IGSM2.2 reproduces changes in both surface and ocean temperatures simulated by the IGSM2.3 from pre-industrial time through the end of the 21st century. However, on longer time scales the assumption that changes in the ocean temperature can be described by the diffusion of the mixed layer temperature anomalies breaks down, leading to overestimation of the deep ocean warming and sea level rise due to thermal expansion by the IGSM2.2. The need for time dependence in the carbon diffusion suggests that the linear assumptions made for the heat diffusion breaks down even earlier for carbon (see Appendix). However, with appropriate choice for the time dependent vertical diffusion of carbon, the simplified ocean carbon model used in the IGSM2.2 can match the changes in the total amount of carbon sequestered by the ocean seen in the three dimensional model.

These results suggests that the IGSM2.2, in spite of simplicity of its ocean component, can nevertheless be used to study changes in atmospheric GHGs concentrations, global air temperature and sea level through the end of the 21st century. These proxies have been used frequently in global change studies, and are arguably the main climate change indicators that will affect human societies. We note here that 2100 is not a "bifurcation point", but merely a useful date, since it is often used in climate change studies.

Thus, to the extent that we accept the IGSM 2.3 model predictions, our comparison has shown that IGSM2.2 is an adequate tool for producing probability distributions of possible changes in SAT and sea level over the 21st century. The use of 3D ocean model is essential for understanding feedbacks or studying changes in climate variables that are not properly simulated by the simplified ocean model as well as in long-term climate simulations.

It should be acknowledged that the need to use simpler ocean models (such as our ADOM) in uncertainty studies has been dictated by the results of Forest *et al.* (2006): different constraints might be put on the rate of oceanic heat uptake as more data will become available.

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APPENDIX: Simplified Ocean Carbon Model

As noted in Section 3, simplified ocean carbon model with fixed vertical diffusion coefficient can reproduce behavior of the 3D ocean carbon model only on very short time scale. When changes in the distribution of the ocean carbon became significant the assumption of the linearity of the response breaks down.

To more thoroughly evaluate the performance of the simplified carbon ocean model, in particular effect of the different assumptions on time dependency of the vertical diffusion coefficient, we carried out simulations with prescribed changes in atmospheric CO_2 concentration from year 1861 to year 2300 following the Bern SP550 and SP1000 scenarios (atmospheric CO_2 stabilization at 550 and 1000 ppm, Plattner *et al.*, 2007). In one set of simulations the changes in atmospheric CO_2 concentration, while affecting ocean carbon model, did not affect the climate simulated by the atmospheric model (FC: fixed climate). In a second set of simulations the ocean carbon model responded to changes in both atmospheric CO_2 and climate (CC: changing climate).

Results for these simulations for IGSM2.3 (performed for the IPCC AR4, Plattner *et al.*, 2007) are shown with black lines in **Figure A1** (dashed for CC and solid for FC). The results of

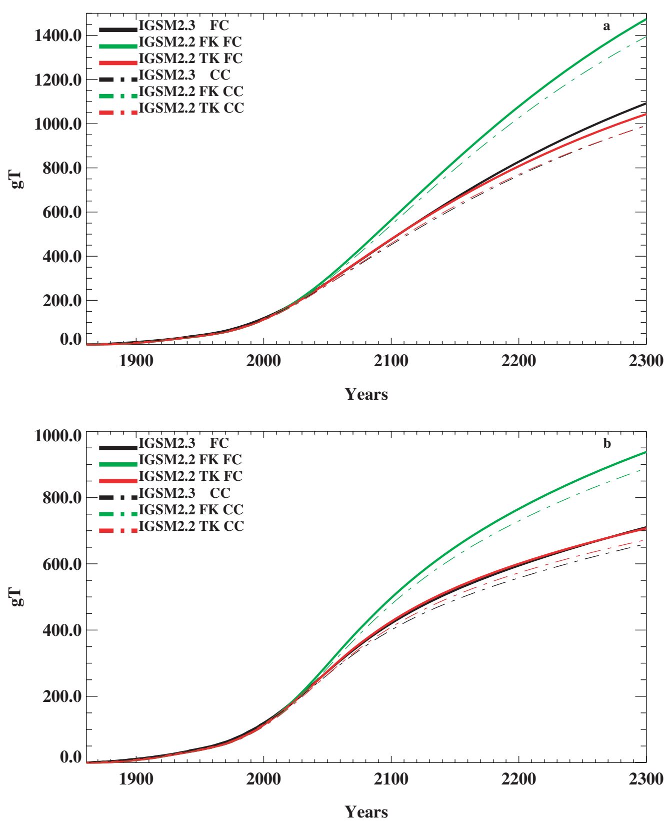

Figure A1. Changes in the ocean carbon in the SP550 **(b)** and SP1000 **(a)** simulations with the IGSM2.3 (black lines), IGSM2.2 with fixed (green line) and time dependent (red lines) K_{vc}. Changes in simulations with fixed climate (FC) are shown by solid lines and in simulations with changing climate (CC) by dashed lines.

the simulations for IGSM2.2 with a K_{vc} that is fixed in time (FK) are shown in green. The values of S, K_v and K_z used in the simulations discussed here are identical to ones in REF22 and REF32 simulations (Table 1). The IGSM2.2 with FK significantly overestimates increase in the ocean carbon storage. However, the difference between the CC and FC runs are similar for the IGSM2.2 and IGSM2.3 implying that changes to circulation and biological pump (simulated by the IGSM2.3 but not IGSM2.2) in the future climate scenarios is not the cause of the mismatch between the simple diffusive and the three dimensional ocean models.

The ability of water to take up carbon from the atmosphere is reduced markedly when it has a higher dissolved inorganic carbon concentration and when the water is warmer; these feedback mechanisms have been cited by several authors (*e.g.,* Matear *et al*., 1999 and Chuck *et al*., 2005). Different ocean carbon models can be compared in terms of carbon uptake sensitivities to an increase in CO_2 and to surface warming (*e.g.,* Plattner *et al.*, 2007), and the results of the FC and CC runs can be used to estimate these sensitivities. Assuming that the change in ocean carbon (ΔC) can be approximated by a linear function of the changes in CO_2 (ΔCO_2) and surface temperature ($\Delta Tsrf$),

$$\Delta C=\beta o\ \Delta CO_2+\gamma o\ \Delta Tsrf,$$

the sensitivity to changes in CO_2 (βo) can be calculated from the change in ocean carbon in the simulations with FC (ΔCfc), as

$$\beta o=\Delta Cfc/\ \Delta CO_2\ ,$$

and the sensitivity to changes in surface temperature (γo) can be calculated from the difference in the change in ocean carbon in simulations with changing (ΔCcc), and fixed climate (ΔCfc), as

$$\gamma o=(\Delta Ccc-\ \Delta Cfc)/\ \Delta Tsrf\ .$$

As discussed by Plattner *et al.* (2007), values of sensitivity parameters depend on both CO_2 change scenario and a time frame. Values shown in the **Table A1** are calculated from changes for 1861-2100 period. As can be seen, the IGSM2.2 with the fixed K_{vc} particularly overestimates sensitivity to CO_2 . The values of βo for this version of the IGSM2.2, in fact, lie outside of the range produced by existing ocean carbon models (see Plattner *et al.*, 2007).

Both versions of the IGSM parameterize the air-sea flux of carbon-dioxide and mixed layer chemistry in a similar manner. However, while the 3D ocean model transports carbon away from the surface by ocean dynamical processes and by an explicit (if simple) parameterization of the sinking of organic material, the IGSM2.2 relies entirely on effective diffusion. (Note however the component of vertical diffusion of carbon independent of the diffusion of heat anomalies (Equation 2) can be considered as a very simplified representation of the biological pump). As a result, the IGSM2.2 poorly represents the movement of carbon, moving it too quickly away from the surface, leading to carbon concentrations in the surface water that are too low (especially when there is significant increase in atmospheric CO_2). This leads to higher carbon fluxes in the IGSM2.2; the maximum flux into the ocean is about 30% larger than in simulations with the IGSM2.3 for both CO_2 scenarios (**Figure A2**).

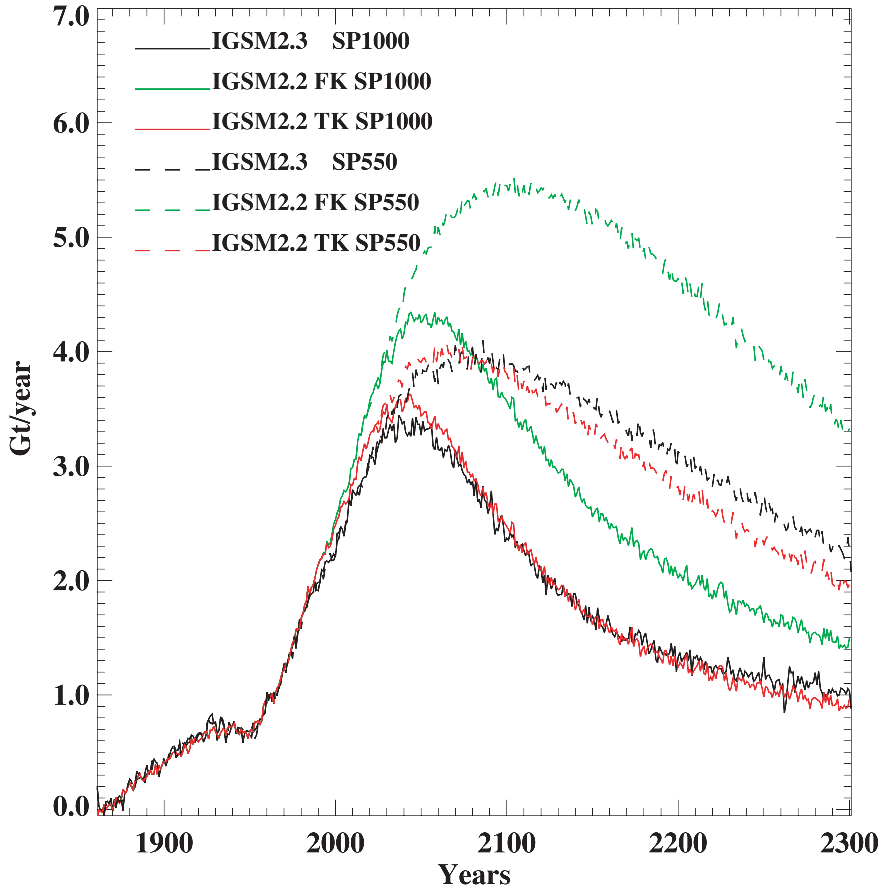

Figure A2. Carbon uptake by the ocean in the SP550 and SP1000 simulations with fixed climate (FC), with the IGSM2.3 (black lines), IGSM2.2 with fixed (FK, green line) and time dependent (TK, red lines) Kvc.

The depth to which a tracer is mixed in the deep ocean as simulated by the diffusive model is proportional to the square root of the vertical diffusion coefficient. Analysis of the carbon mixing into the deep ocean as simulated by the IGSM2.3 in the fixed climate (FC) simulation suggests that an effective diffusion coefficient for carbon should decrease with time as shown on **Figure A3**. This suggests that carbon is not mixed downward in a continuous manner as the diffusion model assumes.

Table A1. Sensitivity of ocean carbon cycle to changes in CO_2 and climate.

	SP550		SP1000	
	βo	γo	βo	γo
IGSM2.3	1.49	-12.2	1.31	-12.3
IGSM2.2 fixed K_{vc}	1.74	-13.5	1.54	-12.7
IGSM2.2 time dependent K_{vc}	1.54	-12.0	1.35	-12.1

We carried out simulations with the IGSM2.2 using K_{vc} calculated from Equation 3 (section 3.2.2), where f(t) is a function shown in figure A3. Results obtained in these simulations (TK), are shown as red lines in Figure A1. They are in considerably better agreement with the results for the IGSM2.3. The sensitivity of the model to changes in carbon is also in a much better agreement with the result produced by the IGSM2.3 (Table A1), and within ranges suggested by other models (Plattner *et al.*, 2007). The fluxes then also compare well between the IGSM2.3 and IGSM2.2 (Figure A2). The IGSM2.3 carbon uptake compares well to other ocean models (Plattner *et al.,* 2007); and thus to the extent that we can believe the 3D model results, the IGSM2.2 with the time dependent K_{vc} is able to reproduce results in the future through the 21st century.

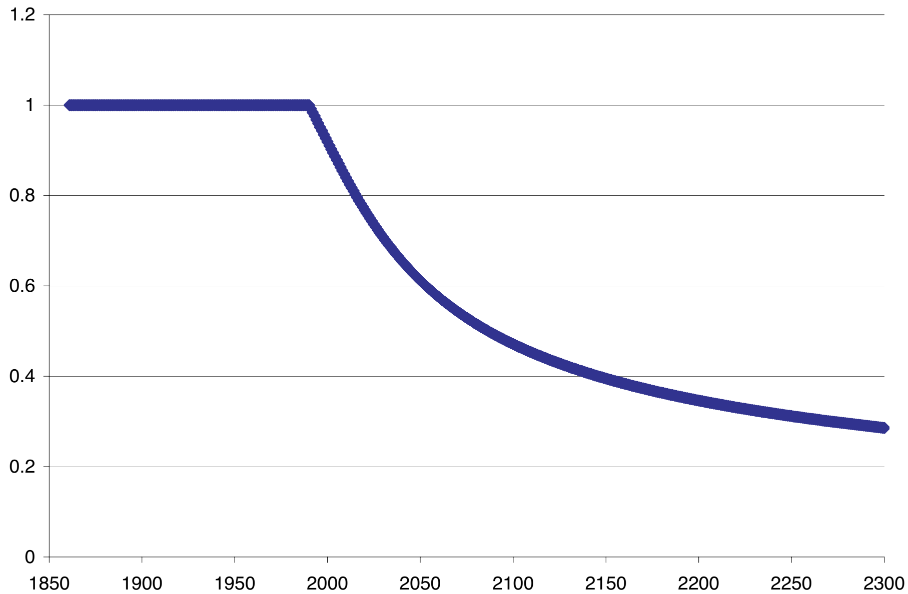

Figure A3. Time dependence of the coefficient for effective vertical diffusion of carbon implied by the results of the simulations with the IGSM2.3

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