Thermal expansion of sea water associated with global warming

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The relationship between greenhouse-gas forcing, global mean temperature change and sea-level rise due to thermal expansion of the oceans is investigated using upwelling-diffusion and pure diffusion models. The sensitivities of sea-level to short-timescale forcing and deep-water formation rate changes are examined. The greenhouse-gas-induced thermal expansion contribution to sea-level rise between 1880 and 1985 is estimated at 2-5 cm. Projections are made to the year 2025 for different forcing scenarios. For the period 1985-2025 the estimate of greenhouse-gas-induced warming is 0.6-1.0 °C. The concomitant oceanic thermal expansion would raise sea level by 4-8 cm.

FUTURE increases in the atmospheric concentrations of the greenhouse gases (carbon dioxide, methane, nitrous oxide and chlorofluorocarbons) are expected to result in substantial global-scale warming in future decades. In response to this warming, global mean sea level should change owing to thermal expansion of the oceans and the melting (or accumulation) of land ice[1-7]. Prediction of these sea-level changes is of importance, because many coastal regions could be adversely affected by even a small sea-level rise. As a prerequisite to such predictions we need to be able to understand past sea-level changes and to predict the future climatic conditions that will affect sea level. Here, we improve on previous estimates of the past and future contributions to sea-level change arising from thermal expansion of the oceans.

Over the past 100 years, while global mean temperature has increased by ~0.5 °C (ref. 8), sea level has risen by 10-15 cm[2,5,6]. The relative contributions of thermal expansion and ice melting to this sea-level rise are uncertain and estimates vary widely, from a small expansion effect[4-6] through roughly equal roles for expansion and ice melting[2,9] to a dominant expansion effect[10].

In principle, modelling the thermal expansion effect would appear to be exceedingly difficult, as a precise determination would require one to be able to model the three-dimensional details of oceanic temperature changes. This is beyond present capabilities; indeed, our knowledge of how deep-ocean temperatures have varied in recent decades and of the physical processes that control any such variations is still rudimentary. In spite of this, simple diffusion or upwelling-diffusion models of the ocean can be expected to give reasonable results for the amount of thermal expansion that might occur in response to greenhouse-gas forcing, even though such models oversimplify oceanic mixing processes. This is because the main contribution to thermal expansion is concentrated in the near-surface layers; this is where both the warming and the thermal expansion coefficient are largest.

To date, only pure diffusion (PD) models have been used to estimate the thermal expansion effect (see, for example, ref. 2), although Revelle[3] has included an upwelling term in an approximate way. A pure diffusion model leads to an isothermal steady-state ocean temperature profile. Inclusion of an advective, upwelling term (balanced by high-latitude downwelling) in an upwelling-diffusion (UD) model ensures a realistic steady-state temperature profile. For small times the differences between the thermal expansion predictions of PD and UD models will be small, provided one begins with a realistic initial profile and both models are calibrated to match past observations. But for times of the order of centuries PD and UD models may give noticeably different results because of the different ways in which they distribute surface heating effects vertically. Here, we use a UD model, calibrated to match past temperature changes within the limits of uncertainty in the model parameters (compare ref. 2), to estimate the thermal expansion effect from 1880 to the present and to predict the range of possible future expansion-related sea-level changes. The results are compared with those obtained using a similarly calibrated PD model.

The model

The model used is a box-upwelling-diffusion energy-balance climate model, an elaboration of the PD model used by Wigley and Schlesinger[11]. It is similar to the model of Harvey and Schneider[12] and is typical of models currently used to study the transient response of the climate system to continuously changing external forcing[13]. The model differs from that in ref. 11 in that it differentiates land and ocean in both hemispheres, and has upwelling-diffusive oceans. Land-ocean and inter-hemispheric exchange coefficients have been set at values that produce good matches to the seasonal cycles of temperature over land and ocean in response to the annual insolation cycle. The model has an oceanic mixed layer the depth of which is set at a constant 100 m. The results presented here are insensitive to the choice of mixed-layer depth.

The model's output is determined by the imposed forcing, and by internal model parameters which define the rates of land-sea and inter-hemispheric exchange, the strength of ocean mixing processes (mixed-layer depth, diffusivity and upwelling velocity), and the sensitivity of the climate system. The latter is conveniently specified by the equilibrium CO_2-doubling temperature change (ΔT_{2x}), that is, the global mean surface air temperature change which would eventually result if the CO_2 concentration were doubled. The parameters that most affect model output are the diffusivity (κ) and the climate sensitivity. Possible feedbacks involving ocean mixing processes[12,13] are not considered. We concentrate on a range of κ values (0.5-2.0 $cm^2\ s^{-1}$) and ΔT_{2x} values (1.5-4.5 °C) which span the limits of current uncertainty[13-15]. We consider two upwelling cases, a constant upwelling rate of 4 $m\ yr^{-1}$ (the standard estimate based on isotope tracer studies[13,16]) and time-varying upwelling. We also consider the PD case by setting the upwelling rate to zero.

The model computes oceanic thermal expansion using expansion coefficient (β) data from Leyendekkers[17]. The

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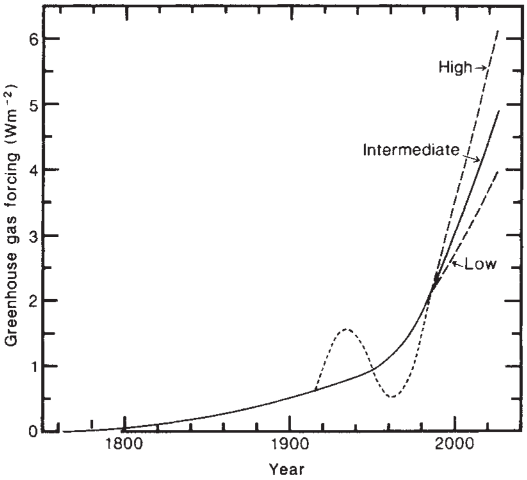

Fig. 1 Assumed forcing based on measured changes in CO_2, CH_4, N_2O and CFC concentrations to 1985 and on projections to 2025. The dashed curve shows the superimposed forcing perturbation applied in the Northern Hemisphere for generating the results shown in Fig. 4. The future projections correspond to low, intermediate and high forcing scenarios.

expansion coefficient β ($=\beta(T, p, S)$ where T is temperature, p is pressure and S is salinity) varies widely with temperature and hence with latitude and depth. Vertical variations in β are included explicitly in the model. To account for latitudinal variations we divided each hemisphere into polar, mid-latitude and tropical zones and used the equilibrium mixed-layer results of Manabe and Stouffer[18] to relate the zonal to the hemispheric mean temperature changes (compare ref. 3). (The latitudinal distribution of these changes is similar to that for observed changes over the past century[19].) The thermal expansion results are relatively insensitive to the details of this partitioning. The effect of S on β is small and S is assumed equal to 35‰.

Past temperature and sea-level changes

As temperature changes and expansion are so intimately linked, we must first evaluate the model's simulations of past changes in global mean surface air temperature in relation to the observed warming, namely 0.5 °C (ref. 8) between 1880 and 1985 with an uncertainty of at least ±0.1 °C[20,21]. The forcing changes at the top of the troposphere due to greenhouse-gas concentration changes since 1765 are shown in Fig. 1. Concentration information used comes from refs 22-26 for CO_2, refs 25 and 27-29 for CH_4, refs 25 and 27 for N_2O and refs 27 and 30-32 for the CFCs. Concentrations have been converted to radiative forcing using the model results of Ramanathan *et al.*[32] and Kiehl and Dickinson[33], with due allowance for overlap effects. Full details are given in ref. 34.

Figure 2 shows the modelled temperature changes from 1880 to 1985, ΔT_0, for various values of κ and ΔT_{2x}. If greenhouse-gas forcing were the sole mechanism responsible for the 1880-1985 temperature rise, then Fig. 2 would give the range of possible ΔT_{2x} values required for compatibility between model and observations. For an observed warming in the range 0.4-0.6 °C, the inferred ΔT_{2x} range is 1.2-2.2 °C, values much lower than recent general circulation model (GCM) estimates which give ΔT_{2x} at ~4 °C[31,35,36]. However, similarly low values for the climate sensitivity have been obtained in other model-based empirical analyses (for example, 1.6 °C in ref. 37).

This apparent discrepancy between observations and recent GCM results can be interpreted in a number of ways: either recent GCM experiments have overestimated the climate sensitivity to a CO_2 change; and/or some additional forcing factor exists which has contributed an overall cooling effect over the past 100 yr, partly offsetting the greenhouse-gas forcing; and/or the upwelling-diffusion parameterization of ocean mixing grossly underestimates the extent of vertical mixing; and/or the 0.4-0.6 °C estimate of global warming is considerably less than the true warming. GCM uncertainties and/or neglected forcings are likely to be the most important factors. For example, changes in cloud optical properties, which may have a strong negative feedback effect[38-41] and are not accounted for in current GCMs, could explain at least part of the discrepancy. The available evidence for additional forcings that are of comparable magnitude to the greenhouse-gas forcing on the century timescale is debatable, but various possibilities have been hypothesized[37,42-44].

A useful way to reduce speculation in interpreting Fig. 2, which we will exploit further below, is to assume only that it gives the greenhouse-gas contribution to 1880-1985 temperature changes. If future model results or observations were to show that the 'correct' values for ΔT_{2x} and κ, say, were 3.0 °C and 1 $cm^2\,s^{-1}$, then from Fig. 2, the greenhouse-gas contribution to the 1880-1985 warming (namely, ΔT_0) would be 0.78 °C. This would then require the existence of a compensating cooling of 0.28 ± 0.1 °C due to other factors.

Figure 3 shows the modelled sea-level change (Δz_0) for the period 1880-1985 due to thermal expansion of the oceans. The range of Δz_0 values compatible with $\Delta T_0 = 0.4$-0.6 °C is 2.3-4.8 cm. This range of values is insensitive to the above-described uncertainties surrounding the greenhouse-gas contribution to the observed warming. We demonstrate this with an example. Suppose that some other external forcing factor (X) operating over the interval 1880-1985 has offset the greenhouse-gas forcing (G) to give a total forcing $T = G - X$, where $X = 0.4G$. With a reduced total forcing compared with G alone, the implied ΔT_{2x}

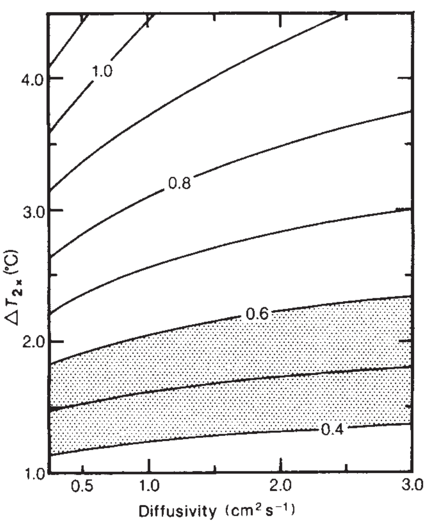

Fig. 2 1880-1985 warming (°C) due to observed increases in greenhouse-gas concentrations for different diffusivities (κ) and equilibrium CO_2-doubling temperature changes (ΔT_{2x}). For varying κ, the upwelling velocity w has been changed to keep the equilibrium vertical temperature profile unchanged (w/κ kept constant). The results differ very little from those obtained using a constant $w = 4$ m yr^{-1}. The observed warming range of 0.4-0.6 °C is shown stippled.

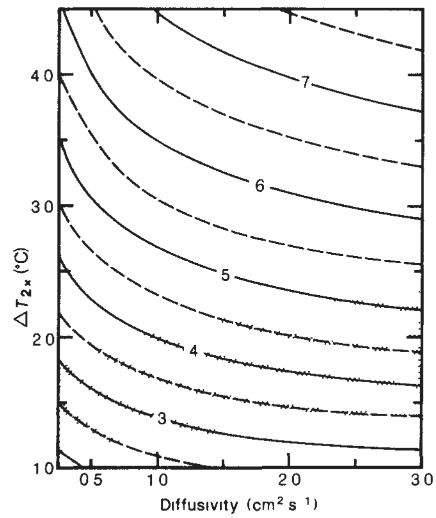

Fig. 3 1880-1985 thermal-expansion-induced sea-level rise (cm) due to observed increases in greenhouse-gas concentrations (Fig. 2) for different diffusivities (κ) and equilibrium CO_2-doubling temperature changes (ΔT_{2x}). The stippled area gives the range of κ and ΔT_{2x} values compatible with the observed 1880-1985 global warming. For $0.5 \leq \kappa \leq 2.0\ cm^2\ s^{-1}$ this implies a contribution of 2-5 cm to sea-level change from thermal expansion.

to give a 0.5 °C warming over the period 1880-1985 must be larger—in this case, 3.3 °C for $\kappa = 1\ cm^2\ s^{-1}$ (as compared with 1.6 °C for greenhouse-gas forcing alone). In spite of the very different forcing and the larger implied ΔT_{2x} value, the corresponding Δz is virtually unchanged. For G-minus-X forcing and with the model tuned to give $\Delta T_0 = 0.5$ °C, Δz is 3.44 cm compared with 3.48 cm for the case of G alone. Thus, for century-timescale forcing, Δz is largely determined by the 1880-1985 temperature change, independent of the magnitude of the external forcing which produced this change.

The results shown in Figs 2 and 3 and the link between Δz_0 and ΔT_0 are well approximated by the empirical expression

$$\Delta z_0 = 6.89 \Delta T_0 \kappa^{0.221} \quad (1)$$

for Δz_0 in cm, ΔT_0 in degrees Celsius and κ in $cm^2\ s^{-1}$. As Δz_0 and ΔT_0 have approximately the same ΔT_{2x} dependence, the ratio $\Delta z_0 / \Delta T_0$ depends only on κ. A similar result is obtained with a PD model,

$$\Delta z_0(\mathrm{PD}) = 7.98 \Delta T_0 \kappa^{0.315} \quad (2)$$

As the diffusivity value for a PD model fitted to tracer data is higher than for a UD model, PD estimates of Δz_0 may be considerably greater than those obtained with a UD model.

Response to short-timescale forcing

Forcing factors operating on annual to decadal timescales undoubtedly exist and are superimposed on the long-timescale greenhouse effect. These will also affect the sea-level rise at any given time. How significant are these effects? We can say *a priori* that they must be relatively small because the damping effect of oceanic thermal inertia on the temperature response to external forcing increases as the timescale of the forcing decreases. A further evaluation of these effects is, however, of considerable interest because different forcings can produce quite different vertical temperature profile changes (and, hence, sea-level changes) even if the surface temperature changes are the same.

To illustrate this, we consider an extreme example. The largest short-timescale perturbation on the overall global warming trend is the Northern Hemisphere cooling that occurred between ~1940 and 1975. (Note that the warming between 1910 and 1940, which exceeded that expected to result from greenhouse-gas forcing, can be considered as part of this perturbation.) If we simulate these changes in different ways, we can maximize the possible shorter-timescale thermal-expansion-related sea-level fluctuations which might be superimposed on the greenhouse-gas-induced rise of 2.3-4.8 cm. As the temperature perturbations are largest in the Northern Hemisphere, we will use hemispherically specific forcing, as suggested in refs 37 and 45. The vertical ocean diffusivity is assumed to be $1\ cm^2\ s^{-1}$, an acceptable assumption in a sensitivity study like this given the results of Figs 2 and 3.

The two possible causes we consider are an external forcing and a change in the rate of upwelling. Both perturbations are taken to be one-cycle sinusoidal changes spanning the period 1915-85. The variable-upwelling case simulates a change in North Atlantic Deep Water (NADW) formation rate, as the North Atlantic is the main source of deep water in the Northern Hemisphere.

The effects of deep-water formation rate changes on global mean temperatures have been considered previously,[13,46] but this is the first time that hemispherically specific changes have been considered. (For evidence pointing towards recent NADW changes, see refs 47-49. Major NADW changes are thought to have occurred on the ice-age timescale[50,51].) Although NADW changes may well be an important factor in explaining recent temperature fluctuations, especially in the Northern Hemisphere, the main reason for considering them here is because their influence on the vertical ocean temperature profile is radically different from that of an external forcing change.

In both cases the model is calibrated (by varying the climate sensitivity and the amplitude of the forcing or upwelling changes) to produce similar hemispheric and global mean temperature changes. The global mean warming over the period 1880-1985 is set to 0.5 °C.

For the external forcing perturbation (shown in Fig. 1), the required amplitude is $0.78\ W\ m^{-2}$ (that is, a decrease of $1.56\ W\ m^{-2}$ over 1932.5 to 1967.5; compare this with the greenhouse-gas-forcing increase of $1.81\ W\ m^{-2}$ over 1880-1985) and the ΔT_{2x} value is 1.77 °C. This gives a maximum modelled Northern Hemisphere cooling of 0.20 °C, during which the Southern Hemisphere warmed by 0.07 °C. Global mean changes are shown in Fig. 4. Both hemispheric and global changes agree well with observations. For the upwelling rate changes, the required amplitude is $0.91\ m\ yr^{-1}$ (see Fig. 4) and the ΔT_{2x} value is 1.76 °C. The corresponding temperature perturbations lead those for the forcing perturbation case by a few years, but they are of the same magnitude—that is, maximum Northern Hemisphere cooling of 0.20 °C with concomitant Southern Hemisphere warming of 0.07 °C. Global mean changes are shown in Fig. 4.

The sea-level effects of these two perturbations are quite different (see Fig. 4), even though both produce similar surface temperature effects. For the two cases, the vertical profiles of the temperature changes agree only at the surface. Clearly, for any given surface temperature history, there is no unique history for sub-surface temperature changes, and hence no unique sea-level time series. In our analysis, however, this is a secondary effect; both perturbations are small relative to the century-timescale influence of the greenhouse effect—deviations of

<0.7 cm compared with a total change of 3.4 cm. The implied uncertainty in the overall rise to 1985 is therefore $\sim \pm 20\%$, owing to short-timescale forcing uncertainties.

The 2.3-4.8 cm range of values for Δz obtained here is compatible with the previous estimates of Gornitz *et al.*[2]. These authors were the first to examine systematically the thermal expansion effect using a model-based approach; a number of other estimates (for instance, refs 7 and 9) are based on different interpretations of Gornitz *et al.* They used a PD model from ref. 42 which considered the world's oceans as a single column. They applied CO_2, volcanic and solar forcing and tuned ΔT_{2x} and κ to match model output with the Hansen *et al.*[42] estimate of global mean temperature changes since 1880. Because they ignored the forcing of other greenhouse gases, they were able to obtain a match for much higher ΔT_{2x} values than obtained here. Also, because their forcing history contains shorter-timescale fluctuations than we have used, their modelled fluctuations in sea level show considerable short-timescale variability. In spite of these differences, they obtain similar thermal expansion estimates to ours, that is, 2.3 cm over the period 1880-1980 and 4.1 cm over the period 1900-80 (for $\Delta T_{2x} = 2.8$ °C and $\kappa = 1.2\ \mathrm{cm^2\ s^{-1}}$). Further support for the values calculated here comes from the empirical estimates of Barnett[5,6], who judges the 1880-1980 expansion effect to be <5 cm.

Future thermal expansion effects

We now return to the constant-upwelling-rate case and project the thermal expansion calculations to the year 2025. The many uncertainties involved in making such a projection may be grouped in two categories, those associated with the projected forcing and those associated with modelling the implied thermal expansion. To cover forcing uncertainties, we have used low, intermediate and high values for the forcing (Fig. 1). The intermediate case uses the best estimates of future greenhouse-gas concentrations from ref. 32, but with due consideration given to other estimates in the literature. Concentrations assumed for the year 2025 are (with 1985 values given in brackets): CO_2, 436 p.p.m.v. (346 p.p.m.v.); CH_4, 2,235 p.p.b.v. (1,643 p.p.b.v.); N_2O, 366 p.p.b.v. (309 p.p.b.v.); F11, 0.93 p.p.b.v. (0.22 p.p.b.v.); F12, 1.59 p.p.b.v. (0.37 p.p.b.v.). (Many other CFCs have been included in pre-1985 and future forcing estimates; F11 and F12 contribute over three-quarters of the total CFC forcing.) Full details of this intermediate-forcing scenario are given in ref. 34. For the low-forcing scenario we have used post-1985 forcing changes that are two-thirds of those used in the intermediate case; for the high-forcing case the post-1985 forcing changes have been multiplied by 1.5. The corresponding projections agree well with other estimates in the literature.

It is somewhat more difficult to account for model uncertainties. We do this by using some remarkable, robust features of the model projections. First, we use the fact that the temperature change ratio, $\Delta T_1/\Delta T_0$ (where subscript 0 refers to 1880-1985 and 1 refers to 1985-2025), is practically independent of the assumed values of ΔT_{2x} and κ[52]. $\Delta T_1/\Delta T_0$ depends only on the 1985-2025 to 1880-1985 forcing ratio, $\Delta F_1/\Delta F_0$. The relationship is nearly linear over a wide range of values of ΔF_1. (For the three models here, $\Delta F_1 = 1.77$, 2.65 and $3.97\ \mathrm{W\,m^{-2}}$, while $\Delta F_0 = 1.81\ \mathrm{W\,m^{-2}}$.) For $\Delta T_{2x} = 1.5$-4.5 °C and $\kappa = 0.5$-$2.0\ \mathrm{cm^2\ s^{-1}}$, we can estimate ΔT_1 to within a few per cent using

$$\Delta T_1 = (0.308 + 0.49\Delta F_1)\Delta T_0 \qquad (3)$$

A similar relationship results if one uses a PD rather than a UD model; indeed, equation (3) gives results which are accurate for a PD model to within a few per cent. As an example, equation (3) implies that, if the greenhouse-gas contribution to ΔT(1880-1985) were 0.6 °C, then (for the intermediate-forcing case) the greenhouse-gas contribution to ΔT(1985-2025) would be $\sim 1.6 \times 0.6 = 0.96$ °C, independent of ΔT_{2x}, κ and the model structure.

Next we use the fact that the ratio $\Delta z_1/\Delta T_1$ is virtually independent of ΔT_{2x} and depends only weakly on κ. $\Delta z_1/\Delta T_1$ does, however, depend on the chosen forcing scenario. This result can be expressed in the form (compare equation (1))

$$\Delta z_1 = f(\Delta F_1)\Delta T_1 \kappa^{0.221} \qquad (4)$$

Equations (3) and (4) can be combined to give

$$\Delta z_1 = (4.13 + 2.65\Delta F_1)\Delta T_0 \kappa^{0.221} \qquad (5)$$

Equation (5) describes the model results to better than $\pm 4\%$ for a wide range of ΔT_{2x} and κ values (ΔT_{2x} is included implicitly through the term ΔT_0). Equation (5) also implies that Δz_1 is non-zero even if $\Delta F_1 = 0$ (for this extreme case, equation (5) has a maximum error of ~20%), a result which is a necessary consequence of the oceanic lag effect and the current disequilibrium between global mean temperature (and sea level) and the greenhouse-gas forcing.

An expression similar to equation (5) may be derived for PD model results:

$$\Delta z_1(\mathrm{PD}) = 1.14(4.13 + 2.65\Delta F_1)\Delta T_0 \kappa^{0.301} \qquad (6)$$

Thus, PD estimates of future expansion may noticeably exceed UD estimates (for $\kappa > 0.2$).

Consider an example based on the UD result, equation (5). Suppose that the greenhouse-gas contribution to the 1880-1985 global warming is 0.4-0.6 °C, and that the future forcing follows the intermediate scenario. The implied 1985-2025 greenhouse-gas-induced warming is 0.64-0.96 °C, and the corresponding range for Δz_1, allowing for uncertainties in κ in the range 0.5-$2.0\ \mathrm{cm^2\ s^{-1}}$, is 3.8-7.8 cm. We consider this range of values to be the best estimate of future oceanic thermal expansion effects over the 1985-2025 interval. Details of changes as a

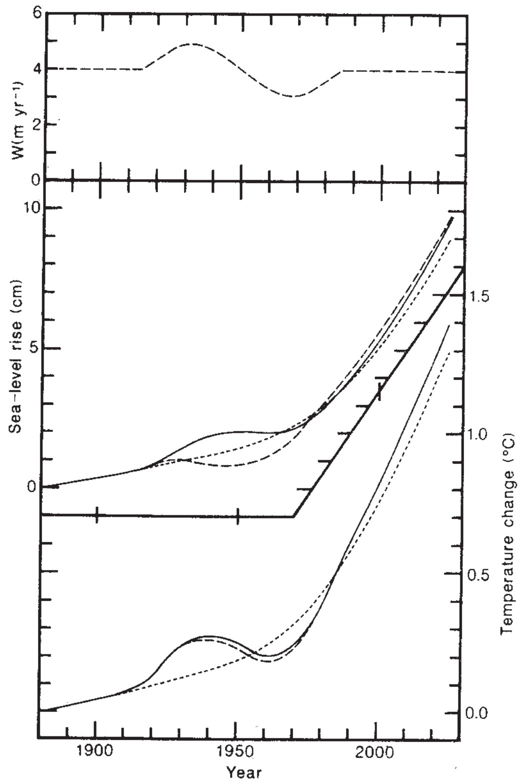

Fig. 4 Global mean temperature and sea-level changes for two different explanations of the mid-twentieth-century cooling of the Northern Hemisphere and the disparate Northern and Southern Hemisphere temperature trends. Dotted curves, greenhouse-gas forcing alone; dashed curves, greenhouse-gas forcing coupled with upwelling velocity changes in the Northern Hemisphere as shown in the top panel; full curves, greenhouse-gas forcing with additional sinusoidal forcing in the Northern Hemisphere only, as shown in Fig. 1. In all three cases the model has been calibrated to produce a global warming of 0.5 °C between 1880 and 1985.

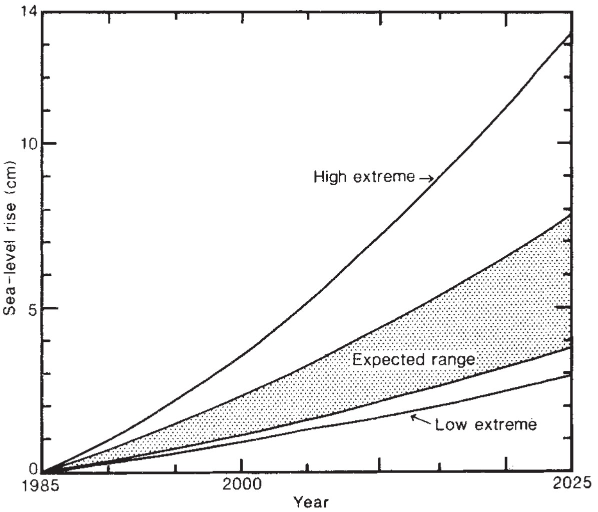

Fig. 5 Thermal-expansion-related sea-level changes over the period 1985-2025. The lower and upper curves correspond to estimated extreme limits (see text). The stippled area gives the range of most likely future changes.

function of time between 1985 and 2025 are shown in Fig. 5.

The lower value is probably close to a lower bound, as it is unlikely that the greenhouse-gas contribution to ΔT(1880-1985) is <0.4 °C (0.4 °C would require ΔT_{2x} to be 1.2-1.3 °C). For the low-forcing case and $\Delta T_0 = 0.4$ °C, Δz_1 is 2.9 cm (equation (5) gives 3.0 cm). To obtain an upper bound, consider first the accepted upper bound for ΔT_{2x} of 4.5 °C. With $0.5 \leq \kappa \leq 2.0\ \mathrm{cm^2\,s^{-1}}$, this would imply a greenhouse-gas contribution to 1880-1985 temperature changes of $\Delta T_0 = 0.93$-1.09 °C (see Fig. 2). Such large values would require a very large additional (negative) forcing to be operating on the century timescale to be compatible with observations. A more realistic upper bound to ΔT_0 is 0.8 °C (implying ΔT_{2x} in the range 2.8-3.5 °C). For the high-forcing case, the corresponding value for ΔT_1 is 1.80 °C and the maximum ($\kappa = 2.0\ \mathrm{cm^2\,s^{-1}}$) value of Δz_1 is 13.4 cm (equation (5) gives 13.7 cm). Figure 5 shows the time evolutions of these upper and lower extremes.

The Δz_1 range of 3.8-7.8 cm, with extreme limits of 2.9-13.4 cm, is noticeably less than other estimates of the future thermal expansion effect. Gornitz *et al.*[2] give a value of 20 cm for the change from 1980 to 2050 which can be translated to a 1985-2025 change of ~10 cm. Hoffman *et al.*[9] give values of 6.5, 13.1 and 18.3 cm as low, intermediate and high values for 1980-2025 (equivalent to 6.0, 12.5 and 17.6 cm for 1985-2025). More recently these authors give lower and upper bounds of 5.7-10.8 cm (1985-2025)[53]. These are the only other publications where time-dependent, transient-response estimates have been made which can be compared with the present work. The differences between these estimates and the present work arise from model differences (all other work has been based on PD models which produce larger expansion effects; see equations (2) and (6)), from differences in the assumed future forcing scenarios, and from differences in the way spatial variations in β have been accounted for and model parameter ranges have been chosen.

Conclusions

These results imply only a small thermal expansion contribution to past sea-level changes (2-5 cm compared with the estimated observed rise of 10-15 cm over the period 1880-1980). The contribution of the melting of small glaciers over 1900-61 has been estimated at $0.46 \pm 0.26\ \mathrm{mm\,yr^{-1}}$ (ref. 54), that is, 2-7 cm over the 1880-1980 period. Long-timescale isostatic rebound effects could account for a further 2 cm (ref. 2). These figures could imply a substantial additional contribution to sea-level rise from melting of the large ice sheets in Antarctica and Greenland, but they are also compatible with a negligible contribution from this source. Our projections of greenhouse-gas-induced sea-level rise due to thermal expansion between 1985 and 2025 are also relatively small, 4-8 cm, accompanied by a global mean warming in the range 0.6-1.0 °C. Estimating future changes in sea level therefore depends crucially on predicting the future melting of land-based glaciers and ice sheets, a daunting task.

This work was supported by grants from the US Department of Energy, Carbon Dioxide Research Division and the European Economic Community Climate Programme. The National Center for Atmospheric Research is sponsored by the NSF. Comments from R. A. Warrick helped to improve this paper substantially. Radiative transfer model results were kindly provided by J. T. Kiehl.

Received 26 May; accepted 17 September 1987.

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