Radiative forcing and climate sensitivity: The ozone experience

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(Received 16 April 1998; revised 15 March 1999)

SUMMARY

The climatic impact of ozone perturbations is studied with a general-circulation model by considering both the radiative forcing at the tropopause and the change in the near-surface temperature. In particular, we focus on the vertical sensitivity by imposing the perturbations in different vertical layers. The simulations are done in perpetual-January mode and with climatological sea ice. The main part of the analysis is focused on global-mean values and in particular on the surface energy budget. Moderate sensitivities related to weak positive feedbacks are found when the perturbation is applied to the troposphere or the lower stratosphere, while larger sensitivities are found for perturbations in the higher stratosphere. This difference in the feedbacks is argued to be related to the vertical partitioning of the radiative forcing determined by the relative strengths of the long-wave and short-wave forcings. While the larger part of the short-wave forcing, which dominates for ozone perturbations in the higher stratosphere, penetrates to the surface, the long-wave forcing is predominantly deposited in the free troposphere. When the model is forced at the surface, latent-heat flux effectively transports energy to the free troposphere restoring the lapse rate to its unperturbed value. Conversely, when the forcing is located in the free troposphere transport to the surface is limited. For the experiments under consideration the radiative forcing and the climate sensitivity are found to be robust and only slightly sensitive to the definition of the tropopause. As a measure of climate change the forcing adjusted for thermal changes in the stratosphere is superior to both the instantaneous forcing and to the forcing obtained after dynamical changes in the stratosphere.

KEYWORDS: Climate sensitivity Global climate model Ozone Radiative forcing

1. INTRODUCTION

The radiative forcing is a natural starting point when attempting to assess the climate impact of changes in either the atmospheric composition or the external forcing. The reasons are both physical and economic. A perturbation in, for example, the concentration of one of the atmosphere’s radiatively active constituents (the climate gases) will lead to a causal chain of changes: the radiative transfer and correspondingly the radiative-heating rates will change immediately, the temperature distribution will start to respond according to the new trends, and finally the dynamics will respond to the new temperature distribution.

The radiative-transfer process is virtually instantaneous and for most climatic purposes only coupled vertically. Thermal adjustments have time-scales ranging from weeks to months in the stratosphere and up to decades in the troposphere–surface system, while dynamical changes require fully three-dimensional considerations. Furthermore, in the study of the tropospheric climate one has to deal with feedbacks from poorly understood physical processes (clouds, water vapour, sea ice). In the light of these obstacles the first objective when studying the impact of climate-gas perturbations on the troposphere should be to estimate the radiative forcing on the tropopause. It is generally believed that this parameter can be used to compare the impact of different climate-gas scenarios, although recently some doubt has been cast on this issue (Bintanja et al. 1997; Hansen et al. 1997).

Several definitions of radiative forcing exist, which all measure the radiative imbalance at the tropopause after an imposed perturbation but before the atmosphere has fully relaxed to its new equilibrium (World Meteorological Organization 1992, hereafter WMO92; Shine et al. 1995). The instantaneous radiative forcing (INST) is calculated by keeping all the atmospheric variables fixed at their unperturbed values. The adjusted

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radiative forcing (FDH) is calculated after allowing the stratospheric temperatures to relax into their new thermal equilibrium under the fixed dynamical-heating approximation (Fels and Kaplan 1975), with unperturbed values for the tropospheric variables and the stratospheric dynamics. A third version of the forcing (DYN), defined by Christiansen and Guldborg (1998), extends the FDH forcing by also allowing the stratospheric dynamics to relax into the new equilibrium state.

The relation between radiative forcing $\Delta F$ and the climate change, measured as the change in the 2 m temperature $\Delta T_{2m}$, is usually written as

$$ \Delta T_{2m} = \lambda \Delta F. $$  \hspace{1cm} (1)

This equation should be understood as a relation between global means. The climate feedbacks manifest themselves through $\lambda$, which is called the climate-sensitivity factor. In the simplest case, where the response is that of a black body, the climate-sensitivity factor $\lambda_0$ can be easily calculated from the balance $L = (1 - a)S/4$ where the outgoing long-wave radiation $L$ is related to the temperature by the Stefan–Boltzmann law $L = \sigma T^4$, $a$ is the planetary albedo and $S$ is the solar insolation. The result is

$$ \lambda_0^{-1} = \frac{dL}{dT} = (4\sigma)^{1/4} \{(1 - a)S\}^{3/4}. $$

With $S = 1370$ W m$^{-2}$ and $a = 1/3$ we find $\lambda_0 = 0.28$ K m$^{-2}$ W$^{-1}$. Positive or negative feedbacks will result in larger or smaller values of $\lambda$, respectively. For Eq. (1) to be useful $\lambda$ should be constant for at least a certain class of radiative forcings. In the best case $\lambda$ should fulfil all of the following three requirements: it should be independent of (i) the strength of the forcing, (ii) its latitudinal distribution, and (iii) the frequency spectrum of the forcing, in particular the division of the flux into its long-wave (lw) and short-wave (sw) parts.

A recent study by Bintanja et al. (1997) with an energy-balance atmosphere model coupled to an advection–diffusion ocean model disagreed with the second requirement. The authors concluded that $\lambda$ depends on the latitudinal distribution of the radiative forcing mainly due to the effect of the ice–albedo feedback.

Hansen et al. (1997) calculated the dependence of the sensitivity factor on the height of the ozone perturbation in the Wonderland general-circulation model (GCM) by successively adding 100 DU to each of the model’s 12 layers*. The sensitivity factor shows large variations with height, and of particular interest is the negative sensitivity factor in the lowest atmospheric layers and the high sensitivity factor in the upper stratosphere around 5–10 hPa. The latter result should be considered with some caution as only four to five model layers are situated in the stratosphere and the atmospheric dynamics are only computed for the lowest nine layers. By repeating the experiments with fixed cloud cover the authors showed that much of the vertical variation can be attributed to cloud feedbacks.

One should also note the large discrepancy in the climate sensitivity between different GCMs. In an intercomparison of 19 GCMs Cess et al. (1990) found that the climate sensitivity varied with a factor of three among the models. The experiments were performed by perturbing the sea surface temperatures while keeping the sea ice constant. In this way the different models have nearly the same control climate, and the time for equilibration of the ocean is avoided. It was concluded that most of the discrepancy in the climate sensitivity was caused by differences in the models’ cloud feedbacks. However, from a thorough study of the models’ surface energy budgets Randall et al.

* 1 DU (Dobson Unit) = $10^{-5}$ m equivalent depth at STP of ozone in a vertical column.
(1992) revealed that the differences between the models to a high degree can be related to general differences in the simulated hydrological cycles.

The present paper reports a GCM study of radiative forcing and climate change due to ozone perturbations imposed in different vertical layers. Ozone is radiatively active in both the sw and lw spectrum, and the lw forcing on the tropopause depends strongly on the distance between the perturbed level and the tropopause, whereas this is not the case for the sw forcing. As a result, the division of the net forcing into its lw and sw contributions depends strongly on the height of the ozone perturbations and the experiments form a test of the third of the three requirements stated above. For each vertical layer two experiments have been performed with a huge (twentyfold) difference in the size of the ozone perturbation providing a test of the first requirement. The emphasis of the analysis is put on the energy budgets of the atmosphere and the surface.

The rest of the paper is organized as follows. The model set-up and the unperturbed model atmosphere are briefly described in section 2 and section 3, respectively. In section 4 the perturbation experiments are defined and the main results are presented. Three aspects of the radiative forcing are treated in some details in section 5: the vertical partitioning of the radiative forcing, the influence of thermal and dynamic adjustment in the stratosphere, and the influence of the tropopause height. Section 6 is devoted to the surface energy balance and to an analysis of the feedbacks. The paper is closed with the conclusions in section 7.

2. THE MODEL

The GCM used in this study is Arpège (Déqué et al. 1994), a spectral model embracing both the troposphere and the stratosphere to the top level at about 80 km (0.007 hPa). The model features a complete set of physical parametrization routines including the Morcrette radiation scheme (Morcrette 1991). The horizontal resolution is T21 and the model has 41 vertical levels. All the experiments are done for perpetual-January conditions. The stratospheric response to uniform ozone reductions and a reduction concentrated in the lower stratosphere has previously been reported (Christiansen et al. 1997), as well as the influence of dynamical adjustment in the stratosphere on the radiative forcing (Christiansen and Guldberg 1998). A study of stratospheric vacillations in the perpetual-January model stratosphere was presented in Christiansen (1999).

The standard version of Arpège has climatological sea surface temperatures. To be able to study changes in the near-surface climate we coded a simple slab-ocean algorithm based on the 'q-flux' method introduced by Hansen et al. (1984). The ocean dynamics are supposed to be the same in all experiments. The contribution from the ocean to the heating of the surface layer, \( F_{\text{ocean}} \), is calculated from the balance \( F_{\text{ocean}} + F_{\text{ref}} = 0 \), where \( F_{\text{ref}} \) is the energy flux from the atmosphere into the surface obtained from a reference run with climatological sea surface temperatures \( T_{\text{clim}} \). The flux from the atmosphere into the surface is the sum of the net sw radiation, \( S \), the net lw radiation, \( L \), the latent-heat flux, \( LH \), and the sensible-heat flux, \( H \). In the following experiments, i.e. the control experiment and the perturbation experiments, the temperature of the slab ocean, \( T_{\text{slab}} \), is calculated from

\[
\rho c d \frac{d(T_{\text{slab}} - T_{\text{clim}})}{dt} = F_{\text{slab}} - F_{\text{ref}},
\]

where \( \rho \) and \( c \) are the density and specific heat of water, respectively, and \( d \) is the depth of the slab ocean. In the perpetual-January runs \( T_{\text{clim}} \) is constant and \( F_{\text{ref}} \) is represented
by its long-term mean. The relaxation time of the slab ocean is $\left( \rho c d \right) / \left( 4 \sigma T_{slab}^3 \right)$. The depth $d$ has been chosen to be 15 m without any geographical variations, which gives a relaxation time of 135 days (with $T_{slab}$ set to 288 K, the global mean). Realistic values of the mixed-layer depth are much higher, between 50 and 100 m (Manabe and Stouffer 1980; Lamb 1984; Dool et al. 1984). The low value for the depth has been chosen to speed up the relaxation into the new equilibrium. The drawback is the possibility of an increased variability in the near-surface climate. However, no significant discrepancies are observed between the reference experiment and the control experiment either in the temporal average or in the variability of the 2 m temperature and the surface fluxes. It should be emphasized that the sea ice is not allowed to change.

The reference experiment lasted for 2400 days starting from an analysed January state. The last 1200 days were used to extract the $F_{ref}$. The control run and the perturbation experiments with the slab ocean active, started from day 2400 of the reference run and lasted for 2400 days. Daily means of the last 1200 days of the control experiment and the perturbation experiments were used for diagnostics.

For each perturbation an additional run was performed to calculate the fixed dynamical-heating forcing and the instantaneous forcing as described in Christiansen et al. (1997). These runs lasted for 600 days and the last 300 days are used for diagnostics. We note that both the FDH and DYN temperature adjustments are performed throughout the whole atmospheric column and are not restricted to the stratosphere. This approach is often taken in GCM studies (Fels et al. 1980; Christiansen et al. 1997). The possibility exists that the tropospheric adjustment may influence the temperature response in the lower stratosphere and thereby the radiative forcing. However, we will argue in section 5(b) that this effect is small.

Where nothing else is stated, we have used a standard definition of the tropopause height that varies linearly with latitude from 100 hPa at the equator to 270 hPa at 76.5° (Ramaswamy et al. 1992).

3. THE UNPERTURBED ATMOSPHERE

Although, the present experiments have a coarser resolution (T21 versus T42) the general model climatology agrees with that presented in Christiansen et al. (1997). In this section we focus on the energy balance of the troposphere–surface system.

The global-mean 2 m temperature is 285.77 K with an estimated standard error of 0.03 K. The standard error of the temporal mean $\sigma (\bar{T}_{2m})$ is calculated from $\sigma (\bar{T}_{2m}) / (N / \tau_{cor})^{1/2}$, where $N = 1200$ days and $\tau_{cor}$ is the correlation time. From inspection of the autocorrelation function for $T_{2m}$ we find $\tau_{cor} \sim 12$ days, as the time where the correlation has decreased to $e^{-1}$. This method is used whenever an estimate is given for the standard error of a temporal mean. The global-mean 2 m temperature is somewhat less (0.7 K) than the value obtained for January when the annual cycle is included. As expected the variability is larger in the winter hemisphere than in the summer hemisphere, and larger over land than over sea.

The global energy budget at the top of the atmosphere (TOA), the tropopause, and the surface is shown in Table 1. Downward energy flux is considered positive. The incoming solar flux at the TOA is 1421.5/4 W m$^{-2}$ in agreement with the earth–sun distance in January. The atmosphere is in energy balance; it receives 13.85 W m$^{-2}$ at the TOA and looses 14.13 W m$^{-2}$ at the surface. The imbalance at the surface is a common feature in models based on climatological sea surface temperatures. The imbalance is reduced to about 1 W m$^{-2}$ in global annual mean if the model is run with the annual
cycle. The slab-ocean model described in the previous section ensures energy balance in the perturbation experiments relative to the control experiment.

The stratosphere is radiatively cooled by 1.58 W m\(^{-2}\). Considering the standard error on the radiative fluxes we see that the cooling is statistically significant. As there are no other energy sources than radiation in the model stratosphere this cooling must be balanced by energy transport from the troposphere. The fact that the cooling almost disappears if the troposphere is defined as a surface of constant pressure (13.85–13.48 W m\(^{-2}\), for a tropopause at 100 hPa, and 13.85–14.25 W m\(^{-2}\) for a tropopause at 150 hPa) suggests that the cooling is real and that the energy transport is mainly in the form of dynamical heating.

The albedo of the troposphere–surface system is 0.30 and the albedo for the surface alone is 0.14. With the model values for the insolation and the albedo we get the blackbody climate-sensitivity factor \(\lambda_0 = 0.27\) K m\(^2\)W\(^{-1}\), which only differs slightly from the value calculated in the introduction.

The energy fluxes at the surface compare well with the range of annual-mean values from other simulations and observations (Boer 1993), except for the sensible-heat flux. The sensible-heat flux is a little weaker (−12.0 W m\(^{-2}\)) in the present experiment compared to the range of values (from −14 to −22 W m\(^{-2}\)) reported by Boer (1993). Accordingly the Bowen ratio (\(H/LH = 0.15\)) is at the low end of the reported values. Of particular interest for the present study is the surface greenhouse effect \(L_{\text{surf}}^\uparrow = 328.45\) W m\(^{-2}\) and the amount of the incoming sw radiation which is absorbed or reflected in the atmosphere \(S_{\text{TOA}}^\uparrow - S_{\text{surf}}^\uparrow = 166.16\) W m\(^{-2}\). We note that the variability is by far the largest for the latent-heat flux. The variability in the latent-heat flux is especially large at middle and high latitudes. The variability in the control experiment does not differ significantly from the variability in the reference experiment with prescribed sea surface temperatures.

At the tropopause the variability is limited to the upward fluxes and related to the variability in the cloud cover. The largest variability in the upward lw flux is found over the equator where high convective clouds are present. Although the variability in the upward sw flux has a local maximum in this region, its absolute maximum is found at high southern latitudes (~60°), where low stratiform clouds are abundant.

Details of the zonal-mean climatology can be found in Christiansen et al. (1997). Here we limit ourselves to present a few fields of relevance for the rest of the paper; the ozone mixing ratio, the specific humidity, and the cloud cover are shown in Fig. 1 as functions of latitude and pressure.

4. The Perturbation Experiments

The ozone perturbation experiments include two uniform reduction (UN) experiments of 50% and 75%, respectively, as well as a series of experiments with ozone perturbations in different vertical intervals. The vertical intervals are chosen as the troposphere (TR) where pressure \(p > 150\) hPa, the lower stratosphere (LS) where \(150 > p > 35\) hPa, and the higher stratosphere (HS) where \(p < 35\) hPa. For each vertical interval two ozone perturbation experiments have been performed with a 50% reduction and a tenfold increase, respectively. For reference we have also performed a doubled CO\(_2\) experiment. The perturbation experiments are denoted 0.5UN, 0.25UN, 0.5TR, 10TR, 0.5LS, 10LS, 0.5HS, 10HS, and 2CO2.

Figure 2 shows the absolute ozone content, \(O_3\), vertically integrated over the four intervals. The total column holds on average 310.1 DU with a minimum over the
TABLE 1. ENERGY BALANCE (W m⁻²) AT THE SURFACE, THE TROPOPAUSE, AND THE TOP OF THE ATMOSPHERE (TOA) IN THE UNPERTURBED MODEL ATMOSPHERE. DOWNWARD FLUX IS CONSIDERED POSITIVE. THE NUMBERS IN THE PARENTHESES ARE DAY TO DAY STANDARD DEVIATIONS.

<table>
<thead>
<tr>
<th>Level</th>
<th>( S_{\downarrow} )</th>
<th>( S_{\uparrow} )</th>
<th>( L_{\downarrow} )</th>
<th>( L_{\uparrow} )</th>
<th>( LH )</th>
<th>( H )</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface</td>
<td>189.19 (1.79)</td>
<td>-26.15 (0.30)</td>
<td>328.45 (1.61)</td>
<td>-386.99 (0.97)</td>
<td>-78.38 (3.34)</td>
<td>-12.00 (1.04)</td>
<td>14.13 (4.19)</td>
</tr>
<tr>
<td>Tropopause</td>
<td>337.47 (0.04)</td>
<td>-102.30 (1.33)</td>
<td>18.94 (0.15)</td>
<td>-239.68 (1.26)</td>
<td>-256.19 (1.21)</td>
<td>-15.43 (0.87)</td>
<td>13.85 (0.94)</td>
</tr>
<tr>
<td>TOA</td>
<td>355.38 (0.00)</td>
<td>-105.38 (1.30)</td>
<td>0.04 (0.00)</td>
<td>-236.19 (1.21)</td>
<td>-15.43 (0.87)</td>
<td>13.85 (0.94)</td>
<td></td>
</tr>
</tbody>
</table>

See text for definition of symbols.


<table>
<thead>
<tr>
<th>Experiment</th>
<th>( \Delta T_{2m} ) (K)</th>
<th>( \Delta O_3 ) (DU)</th>
<th>( \Delta T_{2m}/\text{Int}(1 + \frac{\Delta O_3}{O_3}) ) (K)</th>
<th>( \Delta F_{\text{DIR}} ) (W m⁻²)</th>
<th>( \Delta F_{\text{FDH}} ) (W m⁻²)</th>
<th>( \Delta L_{\text{FDH}} ) (K m²W⁻¹)</th>
<th>( \Delta F_{\text{DYN}} ) (W m⁻²)</th>
<th>( \lambda_{\text{FDH}} ) (K m²W⁻¹)</th>
<th>( \lambda_{\text{DYN}} ) (K m²W⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5U0</td>
<td>-0.46</td>
<td>-156.0</td>
<td>0.66</td>
<td>-0.02</td>
<td>0.91</td>
<td>-2.18</td>
<td>-1.27</td>
<td>-1.25</td>
<td>0.36</td>
</tr>
<tr>
<td>0.5SU0</td>
<td>-0.82</td>
<td>-234.0</td>
<td>0.58</td>
<td>0.05</td>
<td>1.57</td>
<td>-3.61</td>
<td>-2.04</td>
<td>-2.02</td>
<td>0.40</td>
</tr>
<tr>
<td>0.5HS</td>
<td>0.21</td>
<td>-73.2</td>
<td>-0.78</td>
<td>0.62</td>
<td>0.74</td>
<td>-0.57</td>
<td>0.37</td>
<td>0.75</td>
<td>0.57</td>
</tr>
<tr>
<td>0.5LS</td>
<td>-0.27</td>
<td>-60.1</td>
<td>1.25</td>
<td>0.17</td>
<td>0.37</td>
<td>-1.07</td>
<td>-0.70</td>
<td>-0.76</td>
<td>0.36</td>
</tr>
<tr>
<td>0.5TR</td>
<td>-0.54</td>
<td>-22.7</td>
<td>4.47</td>
<td>-0.83</td>
<td>-0.24</td>
<td>-0.60</td>
<td>-0.84</td>
<td>-0.79</td>
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</tr>
<tr>
<td>2CO2</td>
<td>1.57</td>
<td>-22.7</td>
<td>3.39</td>
<td>-0.22</td>
<td>4.17</td>
<td>3.95</td>
<td>4.70</td>
<td>0.40</td>
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</tr>
<tr>
<td>10HS</td>
<td>-0.95</td>
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<td>-0.54</td>
<td>-6.78</td>
<td>-7.66</td>
<td>6.00</td>
<td>-1.66</td>
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<td>0.57</td>
</tr>
<tr>
<td>10LS</td>
<td>2.17</td>
<td>1202.0</td>
<td>1.37</td>
<td>-2.60</td>
<td>-4.49</td>
<td>10.82</td>
<td>6.34</td>
<td>8.60</td>
<td>0.34</td>
</tr>
<tr>
<td>10TR</td>
<td>3.43</td>
<td>444.0</td>
<td>3.80</td>
<td>8.21</td>
<td>3.34</td>
<td>5.34</td>
<td>8.68</td>
<td>10.64</td>
<td>0.40</td>
</tr>
</tbody>
</table>

See text for explanation of experiments and symbols.
Figure 1. Zonal-means of (a) the ozone mixing ratio (parts per million by volume), (b) the specific humidity (g kg$^{-1}$) and (c) the cloud cover (%) as functions of latitude and pressure. The thick dashed curve and the thick solid curve are the standard and TOC tropopause, respectively (see section 5(c)).

equator, the absolute maximum over the winter pole and a secondary maximum over the summer pole. The same latitudinal variation, only more pronounced, is found in the lower stratosphere, which includes approximately 40% of the global ozone amount. In the higher stratosphere, which holds 46% of the total ozone, the ozone amount peaks in the equatorial region and decreases smoothly towards the poles. In the troposphere, which hosts the remaining 15%, the latitudinal variation is weak.
Table 2 summarizes some of the main results of the experiments. In this table are shown the global-mean values for the change in the 2 m temperature $\Delta T_{2m}$, the radiative forcings $\Delta F^{\text{INST}}$, $\Delta F^{\text{FDH}}$, and $\Delta F^{\text{DYN}}$ as well as the FDH climate-sensitivity factor $\lambda_{\text{FDH}} = \Delta T_{2m} / \Delta F^{\text{FDH}}$ and the DYN climate-sensitivity factor $\lambda_{\text{DYN}} = \Delta T_{2m} / \Delta F^{\text{DYN}}$.

A histogram of the climate sensitivity based on the FDH forcing is shown in Fig. 3. Before we continue with a detailed discussion in the next sections we emphasize a few results. Firstly, except for the two experiments with ozone perturbations in the higher stratosphere the sensitivity is remarkably constant. The sensitivity ranges from 0.34 K m$^2$W$^{-1}$ in 10LS, over 0.36 K m$^2$W$^{-1}$ in 0.5LS and 0.5UN, to 0.40 K m$^2$W$^{-1}$ in 2CO2, 0.25UN, and the TR experiments. Secondly, the two experiments 0.5HS and 10HS, with ozone perturbations in the higher stratosphere, stand out with a considerable higher sensitivity, 0.57 K m$^2$W$^{-1}$, than the other experiments. In the next sections we will argue that this result is robust and that the reason is to be found in the vertical partitioning of the radiative forcing.

We note the remarkably linear response in the 2 m temperature change. For perturbations in the same vertical layer even a twentyfold difference in the ozone perturbations gives the same value of $\lambda_{\text{FDH}}$.

The uncertainty in the sensitivity factor is determined by the uncertainty in the 2 m temperature response. The variability in the INST and FDH radiative forcing is virtually zero as the forcing is calculated with exactly the same dynamics as in the
unperturbed atmosphere (Christiansen et al. 1997; Christiansen and Guldberg 1998). The standard error in $\Delta T_{2m}$ is calculated as described in section 3. With typical values of $\sigma(\Delta T_{2m}) = 0.32$ K and $\tau_{cor} = 10$ days we have $\sigma(\Delta T_{2m}) = 0.03$ K. The weakest FDH forcing and thus the largest uncertainty in $\lambda_{FDH}$ is found in 0.5HS with a standard error of 0.08 K m$^2$W$^{-1}$. We therefore conclude that the results described above are statistically significant.

As mentioned above, the FDH climate-sensitivity factor $\lambda_{FDH}$ is remarkably constant except for the perturbations in the higher stratosphere. However, this observation does not hold when one considers the sensitivity factor $\lambda_{DYN}$ based on the forcing calculated after dynamical adjustment in the stratosphere. We also note that the sensitivity factor $\lambda_{INST}$ based on the instantaneous forcing varies strongly among the experiments, taking both positive and negative values. Thus, using $\lambda_{INST}$ to predict the climate change could easily give the wrong sign.

The one-dimensional study by Lacis et al. (1990), with a radiative-convective model which did not include climate feedbacks, showed the largest impact on the surface temperature for ozone perturbations close to the tropopause. Hansen et al. (1997) noted that the relative importance of tropospheric ozone change is enhanced when climate feedbacks are included. Owing to the large vertical extent of the layers used in the present study, a direct comparison can not be performed. However, we note that $\Delta T_{2m}/\Delta O_3$ is negative for the HS experiments and positive for the rest, and that the largest values are found for the TR experiments. As reported in Christiansen and Guldberg (1998), the vertical dependence of $\Delta T_{2m}/\Delta O_3$ on the height of the ozone change found by Lacis et al. (1990) has been reproduced with a radiative-convective model based on the Morcrette radiation scheme. We note that the maximum value of $\Delta T_{2m}/\Delta O_3$ near the tropopause found by Hansen et al. (1997) in the absence of climate feedbacks is almost a factor of two lower than the values reported by Lacis et al. (1990). This discrepancy must be related to differences in either the radiation schemes or the unperturbed atmospheres.

5. Radiative Forcing

In this section we consider three aspects of the radiative forcing. The first aspect is the vertical partitioning of the radiative forcing in a part which is felt on the surface and a part which is felt in the free troposphere. We believe that this partitioning is the main cause of the high sensitivity in the HS experiments. The second aspect is the effect of thermal and dynamical adjustments in the stratosphere. Finally, we discuss the robustness of the radiative forcing and the climate sensitivity to the definition of the tropopause.

The radiative forcing at the tropopause can be split into four parts: the downward sw flux $\Delta S_\downarrow$, the upward sw flux $\Delta S_\uparrow$, the downward lw flux $\Delta L_\downarrow$, and the upward lw flux $\Delta L_\uparrow$. The different versions of the radiative forcing are calculated by

$$\Delta F_{INST,FDH,DYN} = \Delta S_\uparrow + \Delta S_\downarrow + \Delta L_\uparrow + \Delta L_\downarrow,$$

as the sw radiation is independent of temperature within the approximations of the Morcrette radiation scheme. We recall from section 2 that $\Delta L_{FDH,DYN}$ is calculated after temperature adjustments in the whole atmospheric column.

(a) The vertical partitioning

Reducing the ozone content only in the troposphere, leads to a sw cooling of the free troposphere and a heating of the surface. In the experiments where the ozone
perturbations are confined to the stratosphere, approximately 50% of the downward sw forcing at the tropopause reaches the surface where 15–20% is reflected. About 50% of the downward sw forcing at the tropopause is reflected from the troposphere–surface as a whole. That is, about 75% of the sw net forcing is deposited on the surface and the rest in the free troposphere. In contrast, the lw effect of a stratospheric perturbation is mainly felt in the free troposphere.

Figure 4 summarizes how the FDH forcing is split between the surface and the free troposphere. As the downward lw FDH forcing on the surface is not directly accessible in our experiments (the tropospheric temperatures change in the FDH experiments) we have obtained this part of the forcing by scaling the adjusted lw forcing at the tropopause with a factor $L_{\text{surf}}^{\text{INST}}/L_{\text{tropo}}^{\text{INST}}$ calculated from the HS experiments. In the HS experiments the ozone perturbation is located entirely in the stratosphere and this factor gives the attenuation in the troposphere of the downward lw flux. Both HS experiments give the same value, $L_{\text{surf}}^{\text{INST}}/L_{\text{tropo}}^{\text{INST}} = 0.3$. Although only an estimate, the figure shows that the HS experiments are unique as the larger parts of the forcings are felt on the surface. The positive values larger than one in the HS experiments, represent a sw forcing which penetrates largely undisturbed to the surface as described above, accompanied by a weaker lw forcing of the opposite sign due to the temperature change in the lower stratosphere which is mainly felt in the free troposphere. In the LS experiments the lw effect dominates at the tropopause, as the thermal adjustment in the stratosphere is situated closer to the tropopause. The forcing on the surface (mainly sw) is now smaller and of opposite sign to the forcing at the tropopause.

In the next section we will argue that it is this vertical partitioning of the radiative forcing which leads to the high sensitivity in the HS experiments. The possible importance of the vertical partitioning of the forcing and the strength of the convective coupling between the surface and the troposphere was originally put forward by Dickinson et al. (1978), and later pointed out in WMO92. Hansen et al. (1997) suggest that this effect, together with the effect of calculated sea ice, partly explains the similarity of the climate sensitivity found in solar forcing and CO$_2$ experiments. Owing to effects of the calculated sea ice, the climate is more sensitive to a forcing at high latitudes than to a forcing at low latitudes. This effect favours CO$_2$ perturbations, while the effect of the vertical partitioning between surface and the free atmosphere favours solar forcings. We emphasize that sea ice is fixed in our experiments.
The importance of thermal adjustment in the stratosphere for the magnitude and sign of the radiative forcing was first pointed out by Ramanathan (1976), and the first quantitative estimates were calculated by Ramanathan and Dickinson (1979). Now this approximation is in widespread use (WMO92; Ramaswamy et al. 1992; Shine et al. 1995). Recently, dynamical adjustments have been shown to be important for explaining the observed cooling in the lower stratosphere (Wang et al. 1993; Ramaswamy et al. 1996). While Christiansen and Guldberg (1998) discussed the effect of dynamical adjustments on the radiative forcing, we will here focus on the effect on the climate sensitivity.

The difference between the instantaneous forcing and the FDH forcing, as seen from Table 2, is in general much larger than the difference between the FDH forcing and the dynamical adjusted forcing. One exception is the 0.5HS experiment, where the FDH adjustment is small because the perturbation is located at a distance from the tropopause. The difference between the global means of the FDH and DYN approximations is mainly a result of the latitudinal variation of the tropopause height. If the tropopause followed a surface of constant pressure, the average dynamical heating on this surface would be zero due to continuity and so would the average of the change in the lw emission.

As noted in Christiansen and Guldberg (1998), only temperature perturbations immediately above the tropopause will have an effect on the radiative forcing. To ease the comparison with other studies, the FDH zonal-mean temperature adjustment is shown in Fig. 5 as a function of height and latitude for the 2CO2, 10HS, 10LS, and 10TR experiments. The adjustment in the UN experiments has been discussed in detail in Christiansen et al. (1997) and here we only give a few general remarks. For the TR, LS and HS experiments the temperature adjustment is largest in the regions of the ozone perturbations, but considerable adjustment is also found outside these regions and large vertical gradients around the tropopause can be found in the LS, HS and UN experiments. Except for the LS experiments (and to a much weaker extent 0.5HS) the temperature adjustment immediately above the tropopause has the same
sign for all latitudes, although the latitudinal gradients can be quite large. The large latitudinal gradients on the winter hemisphere in the LS and HS experiments is a sign of the dominant SW forcing. The cross-over between cooling and heating in the 2CO2 experiment increases from the tropopause level at the winter pole to 40 hPa at the summer pole. In the global mean the cross-over is found at 50 hPa, not far from the annual-mean value found by Hansen et al. (1997).

It is interesting to note that the FDH forcing is larger than the INST forcing for the 2CO2 experiment, while other investigations (Hansen et al. 1997; Myhre and Stordal 1997) have found that the INST forcing exceeds theFDH forcing. To test if this discrepancy originates from the fact that our FDH approach involves temperature adjustments throughout the whole atmospheric column, additional doubled CO$_2$ experiments were performed with the adjustments limited to the stratosphere. We find only a small influence on the FDH forcing from tropospheric adjustment, and in particular we find that the FDH forcing stays well above the INST forcing.

The DYN sensitivity factor $\lambda_{\text{DYN}}$ (Fig. 6) varies much more than $\lambda_{\text{FDH}}$. It ranges from 0.25 K m$^2$W$^{-1}$ in 10LS to 0.43 K m$^2$W$^{-1}$ in 0.5TR with a mean of 0.35 K m$^2$W$^{-1}$. Thus, it is in general smaller than $\lambda_{\text{FDH}}$ except for the 10LS experiment. Experiments with ozone perturbations in the same vertical interval do not have the same DYN sensitivity factor and, in particular, the HS experiments do not differ systematically from the rest. This is in contrast to what was found for the FDH sensitivity factor.

As we will see in the next section, the HS experiments also stand out with respect to the change in the latent-heat flux at the surface and the change in the lapse rate. In combination with the vertical partitioning of the forcing discussed in section 5(a) we are able to link the high values of $\lambda_{\text{FDH}}$ to the physical processes. This strongly suggests that the distribution of $\lambda_{\text{FDH}}$ among the experiments is not spurious, and it confirms the usefulness of the FDH radiative forcing as a measure of climate change.

The same usefulness does not apply to the DYN radiative forcing. However, this version of the forcing is important as it is a direct measurable quantity. The reason why it is $\lambda_{\text{FDH}}$ and not $\lambda_{\text{DYN}}$ that best measures the climate change is not obvious. One guess could be that the dynamical adjustment is strongly coupled to changes in the troposphere. In fact, it is believed that the circulation of the stratosphere is forced by the pumping effect of breaking Rossby waves propagating from the troposphere (see, e.g. Andrews et al. (1987)). Such a connection would make dynamical adjustment in the stratosphere more a part of the response than a part of the forcing.
TABLE 3. THE FDH RADIATIVE FORCING FOR THREE TROPOPAUSES WITH DIFFERENT LATITUDINAL PROFILES. THE LAST COLUMN GIVES THE SENSITIVITY OF THE RADIATIVE FORCING TO THE AVERAGE HEIGHT OF THE STANDARD TROPOPAUSE.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>$\Delta F_{1}^{FDH}(\Delta p = 0)$ (W m$^{-2}$)</th>
<th>$\Delta F_{2}^{FDH}(\Delta p = 0)$ (W m$^{-2}$)</th>
<th>$\Delta F_{3}^{FDH}(\Delta p = 0)$ (W m$^{-2}$)</th>
<th>$(\partial \Delta F_{1}^{FDH} / \partial \Delta p)_{\Delta p=0}$ (W m$^{-2}$ hPa$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5UN</td>
<td>-1.28</td>
<td>-1.24</td>
<td>-1.28</td>
<td>0.51 $\times$ 10$^{-2}$</td>
</tr>
<tr>
<td>0.25UN</td>
<td>-2.05</td>
<td>-1.98</td>
<td>-2.04</td>
<td>0.88 $\times$ 10$^{-2}$</td>
</tr>
<tr>
<td>0.5HS</td>
<td>0.37</td>
<td>0.41</td>
<td>0.38</td>
<td>0.11 $\times$ 10$^{-2}$</td>
</tr>
<tr>
<td>0.5LS</td>
<td>-0.71</td>
<td>-0.71</td>
<td>-0.72</td>
<td>0.49 $\times$ 10$^{-2}$</td>
</tr>
<tr>
<td>0.5TR</td>
<td>-0.83</td>
<td>-0.83</td>
<td>-0.82</td>
<td>-0.14 $\times$ 10$^{-2}$</td>
</tr>
<tr>
<td>2CO2</td>
<td>3.96</td>
<td>3.94</td>
<td>3.95</td>
<td>-0.10 $\times$ 10$^{-2}$</td>
</tr>
<tr>
<td>10HS</td>
<td>-1.66</td>
<td>-1.88</td>
<td>-1.71</td>
<td>-1.20 $\times$ 10$^{-2}$</td>
</tr>
<tr>
<td>10LS</td>
<td>6.36</td>
<td>6.27</td>
<td>6.46</td>
<td>-4.43 $\times$ 10$^{-2}$</td>
</tr>
<tr>
<td>10TR</td>
<td>8.68</td>
<td>8.68</td>
<td>8.63</td>
<td>0.99 $\times$ 10$^{-2}$</td>
</tr>
</tbody>
</table>

See text for explanation of experiments and symbols.

(c) The definition of the tropopause

In this section we consider the influence of the definition of the tropopause on the radiative forcing and the climate-sensitivity factor. No unique definition exists and at least three versions are abundant in the literature: the dynamical tropopause coinciding with a steep gradient in the potential vorticity; the thermal tropopause defined as the lowest height where the lapse rate exceeds $-2$ K km$^{-1}$; and the tropopause defined as the top of the convective layer (TOC). For studies of the radiative forcing the TOC might seem the natural choice, as it separates the at least partially mixed troposphere–surface system from the radiatively governed stratosphere.

Here we restrict the study to three families of tropopauses all without longitudinal dependence. The general form is chosen as $p_{\text{trop}} = \Pi^i(\phi) - \Pi^\perp + \Pi^\parallel + \Delta p$, where $\phi$ is the latitude, $\Pi^i - \Pi^\perp$ is the latitudinal profile of the $i$th family, and $\Delta p$ is an additive constant. The first family is the standard troposphere used in the previous sections of this paper and defined as

$$\Pi^1 = \begin{cases} 
100 + |\phi|(270 - 100)/76.5 \text{ hPa}, & \text{if } |\phi| < 76.5^\circ \\
270 \text{ hPa}, & \text{if } |\phi| > 76.5^\circ.
\end{cases}$$

The second family is chosen as surfaces of constant pressure, i.e. $\Pi^2 - \Pi^\perp = 0$. The third family is the TOC determined from the long-term average of the control experiment as the lowest level without clouds. The TOC varies from 185 hPa at the north pole, over 75 hPa at the equator to 236 hPa at the south pole.

Table 3 shows the FDH radiative forcing $\Delta F^{FDH}$ at $\Delta p = 0$ for all three families and the slope $(\partial \Delta F^{FDH} / \partial \Delta p)_{\Delta p=0}$ for the first family. For the investigated interval, $-50 < \Delta p < 100$ hPa, the forcing changes almost linearly with approximately the same slope for all three families. The largest differences in $\Delta F^{FDH}$ between the three families are found for the HS experiments, where the differences are around 10%, while much lower differences are found for the rest of the experiments. Also the vertical dependence measured as the ratio $(\partial \Delta F^{FDH} / \partial \Delta p) / F^{FDH}$ between the slope of the forcing and the forcing itself is largest for the 10HS experiment, but rather small for 0.5HS. The reason for the relatively strong dependence on the definition of the tropopause in the HS experiments lies in a rather delicate balance between lw forcing and sw forcing as will be seen later.
Figure 7. The sensitivity factor $\lambda_{\text{FDH}}$ as function of the height of the tropopause for the 2CO2, 10HS, 10LS, and 10TR experiments. See text for explanation.

Let us now consider the robustness of the climate-sensitivity factor to changes in the definition of the tropopause. Comparing first the three families for $\Delta p = 0$ we find, in agreement with the discussion of the forcing in the previous paragraph, that only the sensitivity for the HS experiments varies significantly. However, for all three families the largest sensitivity factors are found for the HS experiments. In a study with a one-dimensional radiative–convective model, Forster et al. (1997) found that only the TOC definition gave climate-sensitivity factors that were independent of the type of the atmospheric perturbation. In a one-dimensional radiative–convective model the TOC is time independent and easily recognized. Below the TOC the atmosphere and the surface are rigidly connected through the fixed lapse-rate condition simulating convective processes in this type of model. The situation is dramatically different in three-dimensional GCMs and in the real atmosphere where convective mixing varies rapidly in space and time.

Considering the dependence of the sensitivity factor $\lambda$ on $\Delta p$, the nine perturbation experiments fall into three different groups represented in Fig. 7 by the 10HS, 10LS, and 10TR experiments. The 2CO2 experiment, which has the same dependence as the 10TR experiment, is also shown in Fig. 7. The first family, i.e. the standard tropopause, is used here, but the two other families give similar results. For the UN and the LS experiments the sensitivity increases weakly with decreasing tropopause height (i.e. increasing $\Delta p$), while the sensitivity is almost independent of the tropopause height for the TR experiments and the 2CO2 experiment. The two HS experiments show a decreasing sensitivity with decreasing tropopause height. From the definition of $\lambda$ we have

$$\frac{\partial \lambda}{\partial \Delta p} \bigg|_{\Delta p=0} = - \left( \lambda \frac{\partial F_{\text{FDH}}}{\partial \Delta p} \frac{1}{F_{\text{FDH}}} \right) \bigg|_{\Delta p=0}.$$

Thus, the large negative slope in Fig. 7 for 10HS is in accordance with the values in Table 3. The negative slope in the HS experiments is mainly a consequence of the variation of $\Delta L_{\text{FDH}}$ with $\Delta p$, as can be seen by expanding $\partial F_{\text{FDH}} / \partial \Delta p$ in
components. For 10HS we have:

\[ \frac{\partial \Delta (S_{\uparrow}^{\text{INST}} + S_{\downarrow}^{\text{INST}})}{\partial \Delta p} = 0.21 \times 10^{-2} \text{ W m}^{-2} \text{hPa}^{-1}, \]

\[ \frac{\partial \Delta L_{\uparrow}^{\text{INST}}}{\partial \Delta p} = 0.00 \times 10^{-2} \text{ W m}^{-2} \text{hPa}^{-1}, \]

\[ \frac{\partial \Delta L_{\downarrow}^{\text{INST}}}{\partial \Delta p} = -0.05 \times 10^{-2} \text{ W m}^{-2} \text{hPa}^{-1}, \]

\[ \frac{\partial \Delta L_{\uparrow}^{\text{FDH}}}{\partial \Delta p} = -1.4 \times 10^{-2} \text{ W m}^{-2} \text{hPa}^{-1}. \]

The large value of a \( \frac{\partial \Delta L_{\uparrow}^{\text{FDH}}}{\partial \Delta p} \) is in turn a result of the strength of the FDH temperature adjustment and its strong vertical gradient near the tropopause (Fig. 5). This adjustment would be weaker if the ozone depletion was located higher in the stratosphere than in the HS experiments presented here, and as a consequence the sign of \( (\partial \lambda / \partial \Delta p)|_{\Delta p=0} \) could be positive, as it is for \( \lambda_{\text{INST}} \). Hansen et al. (1997) found a slight heating near the tropopause after removing ozone above 10 hPa.

An important result is the well-defined gap between the HS experiments and the rest which exists for \( \Delta p < 30 \) hPa. Thus, the larger sensitivity in the HS experiments is robust both to variations in the latitudinal profile of the tropopause and to the height of the tropopause. Only for low tropopauses will the sensitivity of the HS experiments be comparable to the sensitivity found in the rest of the experiments. However, for these low tropopauses a large spread is found in the sensitivities. We note that the tropopause height \( \Delta p = 0 \) seems to be a reasonable choice, as all the experiments except the HS experiments have sensitivities in a narrow band around 0.4 K m²W⁻¹.

6. CLIMATE SENSITIVITY AND SURFACE ENERGY BALANCE

The temporal averages of the change in the surface fluxes taken over the last 1200 days are shown in Table 4. The standard variations (not shown) agree very well with those obtained simply by multiplying the values from the control experiment in Table 1 by \( \sqrt{2} \). As mentioned previously, the strength of the variability of the latent-heat flux is a factor of two larger than that of the sw flux and even larger yet when compared to the lw flux and the sensible-heat flux. As discussed in section 3, the standard error of the time average of the total flux may be roughly estimated from

\[ \sigma(\overline{F}) = \sigma(F)/(N/\tau_{\text{cor}})^{0.5}. \]

With \( N = 1200 \) days and \( \tau_{\text{cor}} = 12 \) days we get \( \sigma(\overline{F}) = 0.47 \) W m⁻². Thus, the changes in the total flux are consistent with energy balance at the surface as should be expected from the formulation of the slab-ocean model and from the absence of a climatological lowest layer over land. However, the sea ice may act as a source or sink of energy as the temperature of the topmost layer of the sea ice is not allowed to increase above the melting point. In the global mean this effect is small. Because of its large variability we have checked the results concerning the latent-heat flux by repeating the calculations with a 'corrected' latent-heat flux obtained from energy balance at the surface. No significant changes were found.

In this section we discuss the changes in the surface energy budget and estimate the importance of the feedbacks. The surface energy budget gives direct information on the hydrological cycle through the latent-heat flux and is a natural starting point for analysing the climate feedbacks (Randall et al. 1992). We begin with the doubled CO₂ experiment for which a comparison with previous work is possible.
TABLE 4. CHANGE IN THE ENERGY BALANCE (W m⁻²) AT THE SURFACE IN THE PERTURBATION EXPERIMENTS

<table>
<thead>
<tr>
<th>Experiment</th>
<th>$\Delta S_\downarrow$</th>
<th>$\Delta S_\uparrow$</th>
<th>$\Delta L_\downarrow$</th>
<th>$\Delta L_\uparrow$</th>
<th>$\Delta S_{net}$</th>
<th>$\Delta L_{net}$</th>
<th>$\Delta H$</th>
<th>$\Delta LH$</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.5UN</td>
<td>1.55</td>
<td>-0.46</td>
<td>-3.88</td>
<td>2.42</td>
<td>1.10</td>
<td>-1.44</td>
<td>0.11</td>
<td>-0.21</td>
<td>-0.47</td>
</tr>
<tr>
<td>0.25UN</td>
<td>2.38</td>
<td>-0.58</td>
<td>-7.00</td>
<td>4.33</td>
<td>1.80</td>
<td>-2.67</td>
<td>-0.17</td>
<td>0.59</td>
<td>-0.45</td>
</tr>
<tr>
<td>0.5HS</td>
<td>0.50</td>
<td>-0.05</td>
<td>1.04</td>
<td>-0.82</td>
<td>0.45</td>
<td>0.22</td>
<td>0.23</td>
<td>-0.83</td>
<td>0.07</td>
</tr>
<tr>
<td>0.5LS</td>
<td>0.55</td>
<td>-0.29</td>
<td>-2.17</td>
<td>1.46</td>
<td>0.26</td>
<td>-0.72</td>
<td>0.45</td>
<td>-0.22</td>
<td>-0.22</td>
</tr>
<tr>
<td>0.5TR</td>
<td>-0.19</td>
<td>-0.01</td>
<td>-2.63</td>
<td>1.73</td>
<td>-0.21</td>
<td>-0.91</td>
<td>-0.04</td>
<td>0.89</td>
<td>-0.25</td>
</tr>
<tr>
<td>2CO2</td>
<td>-1.22</td>
<td>0.48</td>
<td>12.14</td>
<td>-8.09</td>
<td>-0.74</td>
<td>4.05</td>
<td>1.33</td>
<td>-3.75</td>
<td>0.89</td>
</tr>
<tr>
<td>10HS</td>
<td>-6.03</td>
<td>0.73</td>
<td>-3.80</td>
<td>4.71</td>
<td>-5.31</td>
<td>0.89</td>
<td>-0.08</td>
<td>4.08</td>
<td>-0.39</td>
</tr>
<tr>
<td>10LS</td>
<td>-6.85</td>
<td>1.55</td>
<td>19.23</td>
<td>-11.81</td>
<td>-5.29</td>
<td>7.38</td>
<td>1.87</td>
<td>-2.93</td>
<td>1.06</td>
</tr>
<tr>
<td>10TR</td>
<td>-1.61</td>
<td>0.84</td>
<td>27.56</td>
<td>-18.17</td>
<td>-0.77</td>
<td>9.39</td>
<td>2.70</td>
<td>-9.51</td>
<td>1.81</td>
</tr>
</tbody>
</table>

See text for explanation of experiments and symbols. $\Delta S_{net} = \Delta S_\downarrow + \Delta S_\uparrow$; $\Delta L_{net} = \Delta L_\downarrow + \Delta L_\uparrow$.

(a) The doubled CO₂ experiment

For the 2CO2 experiment the FDH radiative forcing at the tropopause agrees with the 4 W m⁻² found in the literature. The change in the 2 m temperature is 1.57 K, which falls in the low end of the interval 1.5–4.5 K obtained in other GCM studies (Intergovernmental Panel on Climate Change 1990, hereafter IPCC90). The resulting climate-sensitivity factor is 0.40 K m²W⁻¹ corresponding to a weak positive feedback. It is important to remember that the present model does not include the ice–albedo feedback which is among the strongest positive feedbacks in a doubled CO₂ climate (Mitchell 1989; IPCC90). The value given above for the climate-sensitivity factor was calculated using the FDH version of the radiative forcing. The instantaneous forcing (INST) is about 15% lower and the dynamical forcing (DYN) about 20% higher. Using the largest forcing still gives a positive feedback $\Delta T_{2m}/\Delta F^{DYN} = 0.33$ K m²W⁻¹.

The change in the equilibrium surface energy budget shows the same pattern as that in the four models compared by Boer (1993). The changes in the lw flux and the sensible-heat flux act to warm the surface while the change in the latent-heat flux acts to cool the surface. Only a minor part of the increase in the downward lw flux is due to the direct effect of increasing the CO₂, as can be seen by comparing with the instantaneous forcing on the surface (not shown). The change in the sw flux is negative. In the comparison cited above, a similar negative change in the sw flux is only observed in the Canadian Climate Centre GCM (CCC). The direct effect of increasing CO₂ and the indirect effect of the expected increase of the humidity will decrease the penetration of sw flux to the surface. For a positive change in the sw flux a decrease of sw extinction by clouds must be expected. The sign and magnitude of the sw effect of clouds depend both on the amount of cloud cover and on the optical properties of the clouds. All the compared models show a decrease of the cloud cover, but the CCC and Arpège are the only models which include parametrized cloud optical properties, and these models both show an increase of cloud optical depth. Thus, the increase of the optical depth compensates for the effect of the decrease in cloud cover. In Arpège the content of cloud water and cloud ice depends on the temperature through $\partial q_s/\partial T$, where $q_s$ is the saturation specific humidity, and the increase of cloud optical depth in a warmer climate is a straightforward consequence of the temperature dependence of this derivative. In the present model the equilibrium change of sw flux is approximately equal to the instantaneous sw forcing (not shown), so the increased penetration due to the decreased cloud cover is balanced by the increased cloud optical depth.

The inclusion of parametrizations of cloud optical properties can have a dramatic influence on a model’s sensitivity as demonstrated by Mitchell et al. (1989). Also
in Boer (1993) the weakest surface warmings are found in the models with such parametrizations. We suggest that the low sensitivity found in the present study is due to a combination of this effect and the absence of the ice–albedo feedback.

\( (b) \) The feedbacks

The change in the surface energy balance is shown in Fig. 8. This histogram shows four bars for each experiment. From the left these bars represent the change in latent-heat flux (solid filled), sensible-heat flux (horizontally hatched), net sw flux (diagonally hatched), and downward lw flux (cross hatched). All fluxes have been normalized with the change in the upward lw flux, which is proportional to the change in the surface temperature but of opposite sign. As a consequence negative (positive) values in Fig. 8 represent a change in the energy flux that if unbalanced would drive the surface temperature in the (opposite) direction to the observed change. We might say that negative values represent a forcing of the surface temperature and positive values represent a relaxation. Analysis of the surface energy balance will provide information on the sign and strength of the related feedbacks.

Let us first comment on the signs of the four different contributions. In all experiments except 0.5UN and 0.5LS the latent-heat flux gives a relaxation of the surface temperature, i.e. in experiments with surface heating the evaporation increases result in a transport of energy away from the surface and into the free troposphere. In all experiments the change in the sensible-heat flux has the opposite sign to the change in the latent-heat flux, i.e. the sensible-heat flux is heating the surface whenever evaporation is cooling it and vice versa. Given the small value of the latent-heat flux in 0.5UN and 0.5LS, it is fair to conclude that the change in the latent-heat flux mainly operates as a negative feedback on the surface temperature and that the change in the sensible-heat flux operates as a positive feedback. The sign of the normalized change in the net sw flux varies among the experiments. The normalized downward lw flux (the surface greenhouse effect) is negative for all experiments constituting a forcing of the surface. The explanation is straightforward—a heating of the surface will be accompanied by a heating of the lowermost part of the troposphere. Both the downward lw flux and the net sw flux is a combination of the instantaneous forcing and the effects of feedbacks, and no conclusions about the sign of the feedbacks can be drawn solely from the budget in Fig. 8.
Although positive for eight out of the nine experiments, the size of the normalized change in the latent-heat flux varies strongly. Leaving out the negative value in 0.5UN it ranges from 0.13 in 0.25UN to 0.9 and 1.0 in the two HS experiments. Thus, for these experiments the negative feedback related to the latent-heat flux is almost as important as the black-body feedback. We note that this feedback can not explain the large sensitivity of the two HS experiments as it has the wrong sign. Figure 9 shows the relative change in latent-heat flux ($\Delta LH/LH_{\text{ctrl}}$), where $LH_{\text{ctrl}}$ is the latent-heat flux in the control experiment, as a function of the change in the 2 m temperature. This figure can be compared to a similar figure in Boer (1993) which compares eight different doubled CO$_2$ simulations. The latent-heat flux in the model is determined in part by the difference between the specific humidity of the surface and the lowest level, and in part by a ventilation factor given by the strength of the surface wind. Considering only the first factor, an upper limit can be estimated from the Clausius–Clapeyron criterion. This limit is shown as the thick line in Fig. 9 with a slope of 6.5 K$^{-1}$. There is a large scatter in the data points in this figure, reflecting the difference in the surface response between the different experiments. In Boer (1993) all models fell below the line with slope equal to three. Our doubled CO$_2$ experiment falls exactly on this line indicating a relatively strong change in the surface activity, although care should be taken when comparing perpetual-January experiments with fixed sea ice to experiments with a full annual cycle and interactive sea ice. The two HS experiments exhibit a very strong change in the surface activity approaching the Clausius–Clapeyron criterion, while the LS and the UN experiments show the weakest activity with values falling between the lines with slopes of one and two, respectively. In all experiments except 0.5LS the change in the latent-heat flux is larger than the change in the sensible-heat flux.

As mentioned above, the change in the net sw flux shown in Fig. 8 is a combination of both the instantaneous forcing and possible feedbacks. Figure 10 shows the contribution when the instantaneous forcing has been removed. All experiments, except the tropospheric ozone perturbations, experience a decreased/increased sw flux at the surface in a warmer/colder climate. Thus, changes in the sw flux are related to negative
feedbacks for all experiments except the TR experiments. In particular, the large negative value for 10HS in Fig. 10 is related to a negative feedback and not the positive feedback that could explain the high climate sensitivity. The obvious candidates for causing changes in the sw transfer are clouds and humidity. Figure 11 shows the changes of humidity, cloud cover, and cloud water. The values are shown per unit of the FDH forcing at the tropopause. In all experiments the humidity and the cloud water increase in a warmer climate. The latter is a simple consequence of the parametrization of the cloud optical properties as mentioned in the previous section. Only in the HS experiments will the cloud cover increase/decrease in a warmer/colder climate. In general, a vertical homogeneous increase in the cloud cover is believed to have a cooling effect on the surface, and the cloud change in the HS experiments can not explain the enhanced sensitivity. The effect of cloud changes on the surface temperature is complex, and the global averages of cloud properties are probably a bad measure as both the altitude and the latitude dependence is considerable. However, there is nothing in the vertical or meridional distribution of the cloud changes that distinguish the HS experiments from the other experiments.

The normalized change in the greenhouse effect (the downward lw flux) is close to $-1.5$ for all experiments except the HS experiments, where the value is $-1.2$ for 0.5HS and $-0.8$ for 10HS. We note that the values more negative than $-1$ mean that the surface is forced by the net lw flux. As for the sw flux, these values include contributions both from the instantaneous forcing and from the adjustments in the troposphere. However, here the instantaneous forcing is negligible compared to the effect of temperature adjustment in the lower part of the troposphere. In the control experiment the surface is cooled by the net lw flux, and the ratio of downward lw flux to upward lw flux is 0.85. Under the conditions of constant lapse rate and constant effective depth of the troposphere, we have $\Delta T_{\text{surf}} = \Delta T_{\text{tropo}}^{\text{eff}}$, where $T_{\text{tropo}}^{\text{eff}}$ is the effective black-body temperature of the troposphere, and therefore

$$\frac{\Delta F_{\downarrow}^{\text{surf}}}{\Delta F_{\uparrow}^{\text{surf}}} = \left( \frac{F_{\downarrow}^{\text{surf}}}{F_{\uparrow}^{\text{surf}}} \right)^{3/4} = 0.9.$$  

Thus, except for 10HS and to a weaker extent 0.5HS, the observed changes in the surface lw flux are in disagreement with these conditions. Returning again to Fig. 11, we do not find that changes in the humidity or clouds are qualitatively or quantitatively different for the HS experiments as compared to the rest. The difference must therefore be found in the change of the lapse rate—a hypothesis which is confirmed from the vertical variation of the global-temperature change shown in Fig. 12. The 10TR, 10LS
and 2CO2 experiments all show a decrease of the lapse rate, while the opposite is the case for 0.25UN, 0.5UN and 0.5TR. The HS experiments and 0.5LS show almost no change in the lapse rate.

The above discussion of the feedbacks and the partitioning of the radiative forcing discussed in section 5(a) can now be synthesized as follows. In the HS experiments the main part of the radiative forcing is sw, of which the largest part penetrates to the surface. The forcing of the surface induces changes in the latent-heat flux, which counteract the radiative forcing and effectively redistribute the energy between the surface and the free
troposphere in such a way as to keep the lapse rate undisturbed. The forcing in the other experiments is dominated by the lw part which is mainly felt in the free troposphere. This forcing does not induce the change in the latent-heat flux, and the related vertical mixing and the temperature response is therefore effectively confined to the free troposphere.

7. CONCLUSIONS

We have presented a study of a series of perturbation experiments in a GCM. The series includes a doubled CO₂ experiment, two uniform ozone-reduction experiments,
and experiments with a 50% reduction and a tenfold increase in each of three vertical intervals. The vertical intervals were chosen to represent the troposphere, the lower stratosphere and the higher stratosphere, respectively. This study differs from most GCM studies by considering more than a few experiments with the same model. All experiments are performed under perpetual-January conditions and with fixed sea ice. The emphasis of the study has been put on the radiative forcing, the climate sensitivity, and the surface energy balance, and we have almost exclusively dealt with global-mean values.

The attempt to explain the climate sensitivity from global values of forcings and physical variables is a crude approach. For example, the latent-heat flux, which played a major part in our analysis, depends in part on the surface winds which in turn depend on the horizontal gradients. Therefore, it is not surprising that many details must be left unexplained. Some of the conclusions may have alternatives, e.g. it might be argued from Fig. 7 that a low tropopause should be preferred, although this choice would give a 25% variability in the sensitivity factor between the experiments and this variability would now be uncorrelated with the change in the energy budget at the surface. In GCM studies the evidence is necessarily circumstantial, but we have tried to present a consistent picture leading from the radiative forcing to the climate change. Additional experiments with artificially specified ‘ghost’ forcings or changes in the solar constant would be helpful for investigating the suggested connection between the sensitivity factor and the vertical partitioning of the radiative forcing.

We have found that the climate-sensitivity factor $\lambda_{FDH}$ is remarkably constant, with values in a narrow interval around 0.38 K m$^2$W$^{-1}$, except for the ozone perturbations in the higher stratosphere where it has values close to 0.6 K m$^2$W$^{-1}$. As both values are larger than the black-body climate-sensitivity factor $\lambda_0 = 0.28$ K m$^2$W$^{-1}$ positive feedbacks dominate in all experiments. In fact $\lambda_{FDH} = 0.38$ K m$^2$W$^{-1}$ is close to the value found for most of the models in the intercomparison by Cess et al. (1990). This value seems to represent a lower bound on the sensitivity and was attributed to a positive water-vapour feedback in the presence of a neutral cloud feedback. In our doubled CO$_2$ experiment the sw effect of the reduced cloud cover is balanced by the changes in the cloud optical properties.

The high sensitivity in the HS experiments is accompanied by large changes in the latent-heat flux at the surface and almost vanishing changes in the temperature lapse rate. The radiative forcing at the tropopause in the HS experiments is dominated by sw flux penetrating mostly to the surface. This should be compared to the radiative forcing in the other experiments which is dominated by lw flux felt mainly in the free troposphere. We suggest that the difference in the sensitivity factor is a consequence of this vertical partitioning of the radiative forcing and the nature of the coupling of the surface and free troposphere by convective processes. When the forcing is mainly deposited on the surface large changes in the latent-heat flux are initiated, effectively distributing the temperature response over the full troposphere–surface system. On the other hand, when the forcing is mainly felt in the free troposphere, the change in the latent-heat flux is smaller and the response confined to the free troposphere. We note a possible negative feedback: the deeper in the troposphere the heating is deposited the larger the greenhouse trapping of the cooling lw radiation.

We have shown that the radiative forcing at the tropopause and thus the climate sensitivity is robust to changes in the definition of the tropopause profile as well as to the height of the tropopause. Whether these results hold for other perturbations and carry over to other GCMs remains to be seen. The robustness of the climate sensitivity disagrees with a recent study (Forster et al. 1997) with a one-dimensional
model. We believe the explanation should be found in the more fuzzy transition (varying in both time and space) between the troposphere and the stratosphere in the GCM.

Finally, an alternative definition of radiative forcing has been studied. This definition extends the usual adjusted forcing by allowing dynamical as well as thermal adjustments in the stratosphere. Defined in this way the DYN radiative forcing has the advantage of being a directly observable quantity. However, we find that, as a measure of climate change, this definition is inferior to the usual adjusted forcing. The reason is unknown but might be related to the fact that the general circulation in the stratosphere to a large degree is forced by waves originating from the troposphere. Thus, changes in the stratospheric dynamics should not be taken as part of the forcing but rather as part of the response.

We close the paper by a comparison with the recent work by Hansen et al. (1997) which presents forcings and sensitivities for a large number of perturbations including ozone perturbations in individual levels, doubled CO₂, changed solar constant, as well as ghost forcings. They used the Wonderland GCM with 12 vertical layers coupled to a slab ocean based on the 'q-flux' method with calculated sea ice. They found a larger variation in the climate sensitivity than in the present study. For example, for ozone perturbations in the individual stratospheric layers the sensitivity factor varies between 0.29 and 1.72 K m²W⁻¹, where both these extreme values are found for perturbations in the higher stratosphere. With fixed clouds the range of the sensitivities diminishes, although the sensitivity still varies with a factor of 3. With fixed clouds the largest sensitivities are found in the lower stratosphere. Quantifying the effect of cloud-cover changes, they attempted to explain the difference between the fixed cloud and calculated cloud experiments as a direct, linear effect of the cloud-cover change. However, although this attempt was successful for ozone perturbations in the troposphere, and in particular for explaining a 'paradoxical' cooling when ozone was added to the lowest layer, it was not successful for ozone perturbations in the stratosphere. Experiments with ozone perturbations in extended vertical intervals covering several model layers, like those presented in the present work, show a relatively high sensitivity (0.83 K m²W⁻¹) when ozone is removed above 10 hPa. With fixed clouds this value increases weakly (0.93 K m²W⁻¹). For comparison the sensitivities are 1.44 and 0.67 K m²W⁻¹ with calculated clouds, and 0.66 and 0.75 K m²W⁻¹ with fixed clouds, for ozone perturbations below 400 hPa and near the tropopause, respectively. They conclude that the principal mechanisms of the climate response involve alterations of lapse rate and changes of cloud cover. A direct comparison with the present work is hard to make for several reasons. Firstly, the definition of the adjusted radiative forcing used in Hansen et al. (1997) differs from the FDH forcing used here, by allowing some dynamical adjustment in the lower stratosphere. However, their adjusted forcing also differs from our DYN forcing by not allowing adjustment in the troposphere. Secondly, in general the climate sensitivity found by Hansen et al. (1997) (0.92 and 0.69 K m²W⁻¹ for 2CO₂ with and without interactive clouds, respectively) is considerably larger than the sensitivities reported here, even in their experiments with fixed clouds. With interactive clouds the difference is more than a factor of two. This is probably partly a result of the calculated sea ice itself and partly a result of the interaction between the clouds and the calculated sea ice. Also the parametrizations of cloud optical properties present in Arpège may reduce the sensitivity as discussed in section 6(a). We believe that the main explanation of the differences between Hansen et al. (1997) and the present work is related to the differences in the cloud feedback.
ACKNOWLEDGEMENTS

This work was supported by the Commission of the European Communities (contract ENV4-CT96-0323) and the Nordic Council of Ministers (contract FS/ULF/93003).

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