

# THE ROLE OF HALOCARBONS IN THE CLIMATE CHANGE OF THE TROPOSPHERE AND STRATOSPHERE

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**Abstract.** Releases of halocarbons into the atmosphere over the last 50 years are among the factors that have contributed to changes in the Earth's climate since pre-industrial times. Their individual and collective potential to contribute directly to surface climate change is usually gauged through calculation of their radiative efficiency, radiative forcing, and/or Global Warming Potential (GWP). For those halocarbons that contain chlorine and bromine, indirect effects on temperature via ozone layer depletion represent another way in which these gases affect climate. Further, halocarbons can also affect the temperature in the stratosphere. In this paper, we use a narrow-band radiative transfer model together with a range of climate models to examine the role of these gases on atmospheric temperatures in the stratosphere and troposphere. We evaluate in detail the halocarbon contributions to temperature changes at the tropical tropopause, and find that they have contributed a significant warming of  $\sim 0.4$  K over the last 50 years, dominating the effect of the other well-mixed greenhouse gases at these levels. The fact that observed tropical temperatures have not warmed strongly suggests that other mechanisms may be countering this effect. In a climate model this warming of the tropopause layer is found to lead to a 6% smaller climate sensitivity for halocarbons on a globally averaged basis, compared to that for carbon dioxide changes. Using recent observations together with scenarios we also assess their past and predicted future direct and indirect roles on the evolution of surface temperature. We find that the indirect effect of stratospheric ozone depletion could have offset up to approximately half of the predicted past increases in surface temperature that would otherwise have occurred as a result of the direct effect of halocarbons. However, as ozone will likely recover in the next few decades, a slightly faster rate of warming should be expected from the net effect of halocarbons, and we find that together halocarbons could bring forward next century's expected warming by  $\sim 20$  years if future emissions projections are realized. In both the troposphere and stratosphere CFC-12 contributes most to the past temperature changes and the emissions projection considered suggest that HFC-134a could contribute most of the warming over the coming century.

## 1. Introduction

Humans directly emit small amounts of the gases collectively known as halocarbons into the atmosphere. These gases do not occur naturally and although they have very small mixing ratios in the atmosphere they are often strong greenhouse gases due to the fact that most of their absorption occurs in the 8–13  $\mu\text{m}$  atmospheric window

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region of the outgoing thermal infrared spectrum and that they often have very long atmospheric lifetimes. Collectively, halocarbons are estimated to have led to a positive radiative forcing of  $0.34 \text{ W m}^{-2}$ , which is  $\sim 14\%$  of the total forcing from well-mixed greenhouse gases (WMGHGs) (IPCC, 2001). This would be expected to warm the surface-troposphere system. However, the chlorinated and brominated halocarbons have been shown to be responsible for stratospheric ozone loss, whose negative radiative forcing could have offset  $\sim 50\%$  of their direct positive forcing (WMO, 2003).

In climate models, halocarbons have seldom been assessed individually: rather, they are usually included collectively in the WMGHGs. For simplicity a generic halocarbon absorption cross-section is often used in their radiation schemes. Further, these radiation schemes typically employ only a few wide bands and are not well suited for modelling the sharp spectral features of the halocarbons that overlap with other absorbers (e.g., Jain et al., 2000). Typically offline detailed radiative transfer calculations (e.g., Jain et al., 2000; Sihra et al., 2001) are used to calculate a radiative forcing efficiency, which is combined with information on a gases atmospheric lifetime to estimate its Global Warming Potential (GWP), which is defined as the time-integrated radiative forcing from a 1 kg pulse emission of a gas, compared to that of a reference gas (carbon dioxide). This gives some measure of the gases surface climate impact and is used extensively in policy decisions, like the Kyoto protocol. However, such an approach tells one very little about the expected surface temperature change and alternative suggestions based on using simple analytical climate models to explicitly calculate temperature change have been proposed (e.g., Shine et al., 2004). Also radiative forcing makes some basic assumptions about the equitable distribution of energy through the surface-troposphere system. However, it is helpful to recognize that halocarbons have quite different spectral absorption characteristics to the other WMGHGs and interact very differently with the Earth's radiation field. The very strong  $15 \mu\text{m}$  band dominates the role of carbon dioxide, whereas the halocarbons tend to absorb weakly in the  $8\text{--}13 \mu\text{m}$  atmospheric window, a region of the spectrum where other gases have only a small effect on outgoing longwave radiation. This, combined with a typical vertical temperature profile, means that halocarbons usually warm the atmosphere locally, as opposed to carbon dioxide, which generally cools (the atmosphere only warms as a response to the induced surface warming). Further, this effect is largest at the tropical tropopause, where temperatures are most different from those of the underlying surface (e.g., Dickenson, 1978; Wang et al., 1992; Forster et al., 1997).

In this paper we first examine the role of halocarbons on temperatures near the tropopause (Section 2) and discuss the consequences for climate sensitivity (Section 3). We then use a very simple climate model to examine the likely role of halocarbons on present and future global average temperatures (Section 4).

## 2. Temperature Change near the Tropopause

In this section, we examine the role of halocarbons in temperature change above the surface. We concentrate our analysis on the tropopause region, as changes here can strongly affect the strength of climate feedback (especially the water vapour feedback). Temperature changes here are also important for stratospheric chemistry. For example, the temperature at the tropical tropopause largely controls the amount of water vapour entering the stratosphere.

### 2.1. APPROACH

As noted in Section 1 the wide-band radiation scheme of climate models can have difficulty modelling the role of halocarbons. Fixed dynamical heating (FDH) models are a common approach to modelling stratospheric temperature change with sophisticated radiation schemes (e.g., Shine et al., 2003). They assume that there is no dynamical response to a radiative forcing above a “tropopause”, and this “tropopause” is chosen to separate out the relatively fast response of the atmosphere that is unconnected to the much slower response of the ocean, which has a much larger heat capacity. It is common to use FDH models to examine stratospheric temperature changes. However, they may also prove a worthwhile tool in the tropical upper troposphere. In the tropics there is a well-defined tropical transition layer (TTL) (e.g., Gettelman and Forster 2002; Folkins et al., 2000). Papers such as these have shown that the influence of convection on temperature rapidly tails off upwards of its base at the mean convective outflow region ( $\sim 12$  km), and is insignificant by its top, at the cold-point tropopause ( $\sim 16$  km). Therefore, the balance of radiative heating and the meridional circulation largely control the temperatures near the cold-point. Forster et al. (1997) used a radiative convective model to show that in a globally averaged atmosphere that is in radiative–convective equilibrium, a TTL-type region is heated by halocarbons independently from the rest of the troposphere. In this paper, we choose to adopt an FDH approach with a lower than usual “tropopause”, which we set at the base of the TTL. We do this to examine temperature changes in the TTL and stratosphere in isolation of the surface response.

### 2.2. MODEL

For the FDH calculations we use the Reading Narrowband Model (NBM) (Shine 1991), with updates and halocarbon absorption cross-section data from Sihra et al. (2001). This was combined with the solar scheme of Fu and Liou (1992), for experiments involving other gases and forcing mechanisms. The most important 17 present and future halocarbons, in terms of their emissions and GWPs were examined (shown in Table I). The first set of experiments analysed the tropical

TABLE I  
The tropopause temperature change efficiencies of the gases assessed in this paper

Gas	2000 concentration (pptv)	Radiative efficiency ( $\text{W m}^{-2} \text{ ppbv}^{-1}$ )	TTCEs ( $\text{K ppbv}^{-1}$ )	
			Clear	Cloudy
CFC-11	261 <sup>a</sup>	0.25	0.36	0.24
CFC-12	535 <sup>a</sup>	0.32	0.50	0.32
CFC-113	82 <sup>a</sup>	0.30	0.44	0.29
CFC-114	17 <sup>a</sup>	0.31	0.43	0.28
CFC-115	8 <sup>a</sup>	0.18	0.33	0.22
HCFC-22	142 <sup>a</sup>	0.20	0.34	0.22
HCFC-141b	13 <sup>a</sup>	0.14	0.23	0.15
HCFC-142b	13 <sup>a</sup>	0.12	0.30	0.19
CF <sub>4</sub>	82	0.08	0.28	0.19
C <sub>2</sub> F <sub>6</sub>	3	0.26	0.25	0.18
SF <sub>6</sub>	5	0.52	0.71	0.47
HFC-23	15	0.16	0.31	0.20
HFC-32	0	0.09	0.18	0.12
HFC-125	0	0.23	0.39	0.26
HFC-134a	12	0.15	0.28	0.18
HFC-143a	0	0.13	0.27	0.17
HFC-227ea	0	0.30	0.45	0.29
HFC-245ca	0	0.23	0.24	0.16
HFC-42-10mee	0	0.40	0.39	0.26

*Notes:* Values of concentration are taken from either WMO (2003) (a superscript) or IPCC(2001). Radiative efficiencies are from WMO (2003). For each gas we list its present day concentration (ppbv) and its radiative forcing efficiency ( $\text{W m}^{-2} \text{ ppbv}^{-1}$ ), taken from either WMO (2003) or IPCC (2001). The tropopause temperature change efficiency (TTCE) ( $\text{K ppbv}^{-1}$ ) is the temperature change at the tropical cold-point tropopause (16.4 km) from adding 1 ppbv of the gas to either clear-sky or all-sky climatologies.

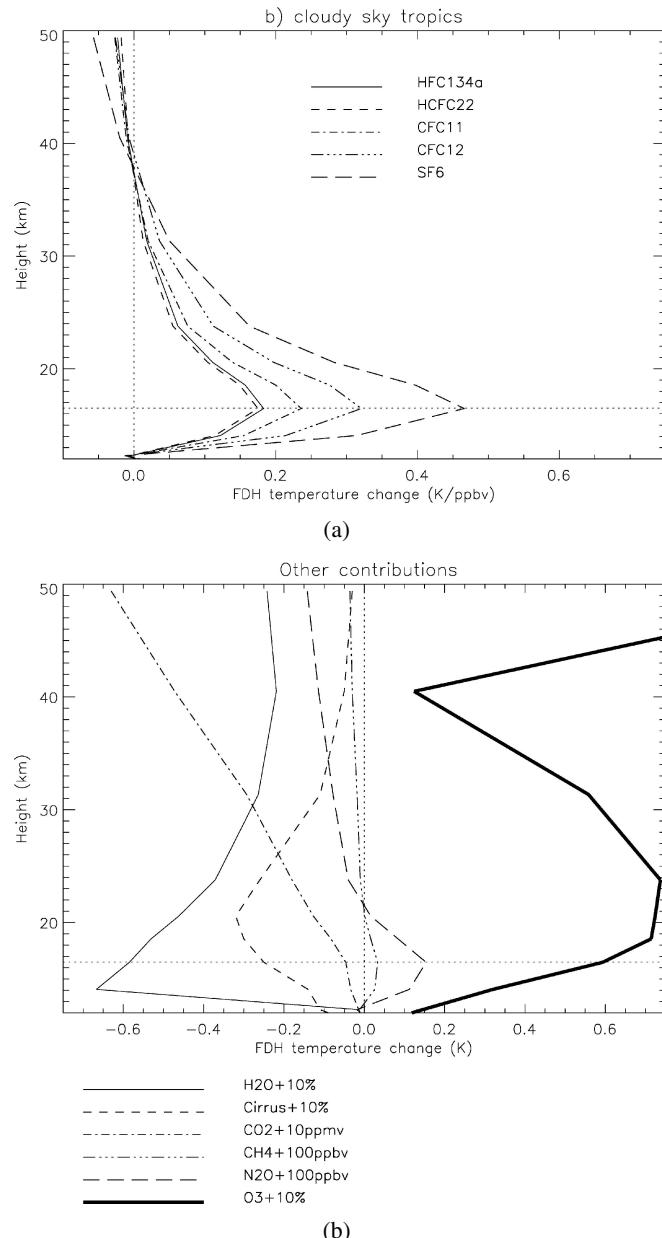
temperature changes near the cold-point tropopause. For these we employed a 15° N–15° S zonally averaged background atmosphere for each season. We used a present day climatology of H<sub>2</sub>O, CO<sub>2</sub>, CH<sub>4</sub>, N<sub>2</sub>O and O<sub>3</sub> and separately added a uniform 1 ppbv of each halocarbon to the background atmospheres and calculated the FDH temperature change above a 12 km level (termed the “FDH tropopause”). The FDH temperature change at the cold point is termed the tropopause temperature change efficiency (TTCE) (see Table I). Clear-sky and all-sky experiments were performed. When clouds were included (in the all-sky experiments), they were taken from International Satellite Cloud Climatology Project (ISCCP) (Rossow and Schiffer, 1991) data. The second set of experiments looked at global effects and employed a latitudinally resolved 10° resolution zonally averaged seasonal background climatology. For these experiments all latitudes had a 5 km “FDH tropopause”. It is certainly not justifiable to assume that all temperature changes above 5 km could be independent of the surface. However, this low-altitude tropopause is only chosen to compare qualitative patterns of response and we found that varying the height of

this tropopause by a few kilometers made very little difference to our results. We also make the assumption that halocarbon mixing ratios are constant with height. In practice, halocarbon mixing ratios decrease in the stratosphere, so our approach overestimates their stratospheric role. However, as we concentrate our investigation and discussion to the TTL layer, this assumption has a minimal effect on our findings. The temperature changes we calculate would be in addition to any change introduced by an changes in surface temperature.

### 2.3. RESULTS AND DISCUSSION

Table I gives the TTCEs for the 17 gases under clear- and all-sky conditions and also shows their radiative efficiencies, as quoted in WMO (2003). This radiative efficiency is a radiative forcing per unit concentration, and is a measure of a halocarbons ability to alter surface temperature. Clouds absorb upwelling radiation from the surface and emit this at a cooler temperature, hence they reduce a halocarbons ability to warm the TTL. Using our cloud climatology, the all-sky TTCE is approximately 2/3 of the clear sky TTCE. The vertical profile of temperature change for 5 of the most important gases is shown in Figure 1a. As in Table I, a range of temperature changes was found, with SF<sub>6</sub> giving over twice the warming at the cold point of many of the other halocarbons. For comparison, Figure 1b illustrates the temperature change from other possible perturbations to the radiative balance of the region (see figure legend for details). These perturbations are for illustrative purposes and are not designed to be realistic.

Using present day mixing ratios, Figure 2 shows the latitudinally resolved total expected temperature change from all halocarbons and compares it to that associated with stratospheric ozone loss and carbon dioxide. For simplicity a linear sum of the individual calculations was used to estimate this halocarbon total. Tests at two single latitudes showed that this assumption was robust to within a few percent. The stratospheric temperature changes associated with ozone loss have been extensively studied (see the discussion in Shine et al., 2003; WMO, 2003). Here we simply note that our FDH model essentially reproduces the results of past models that employed the same ozone change data set (see figure legend for details). For the calculation of a globally averaged temperature change, the warming in the lower stratosphere can be thought of as partially cancelling out the cooling effect of ozone depletion. However, the two patterns are quite distinct and, as a result, the equator-to-pole temperature gradient in the lower stratosphere, where temperature normally increases towards higher latitudes, is reduced. This is similar to the situation when the surface radiative forcings are compared for halocarbon increases and stratospheric ozone loss. Although the forcings partially cancel out in the global average, the cancellation does not happen between the two patterns of forcing: their sum gives a positive forcing at low latitudes and a negative forcing at high latitudes (Shine and Forster, 2000).



*Figure 1a.* (a) Tropical annual average FDH temperature changes (K) from various climate perturbations: (a) 1 ppbv increase in five halocarbons; (b) seven other perturbations. For (b) water vapour mixing ratios were increased by 10% above 12 km; cirrus cloud fractional coverage was increased by 10% (from 20 to 22% total cloud cover); CO<sub>2</sub> was increased by 10 ppmv, from a background value of 350 ppmv; CH<sub>4</sub> was increased by 100 ppbv, from a 1400 ppbv background; N<sub>2</sub>O was increased by 100 ppbv from a 250 ppbv background; O<sub>3</sub> mixing ratios were increased by 10% throughout the atmosphere. The horizontal thin dotted line marks the height of the cold-point tropopause and the vertical line marks the zero temperature change

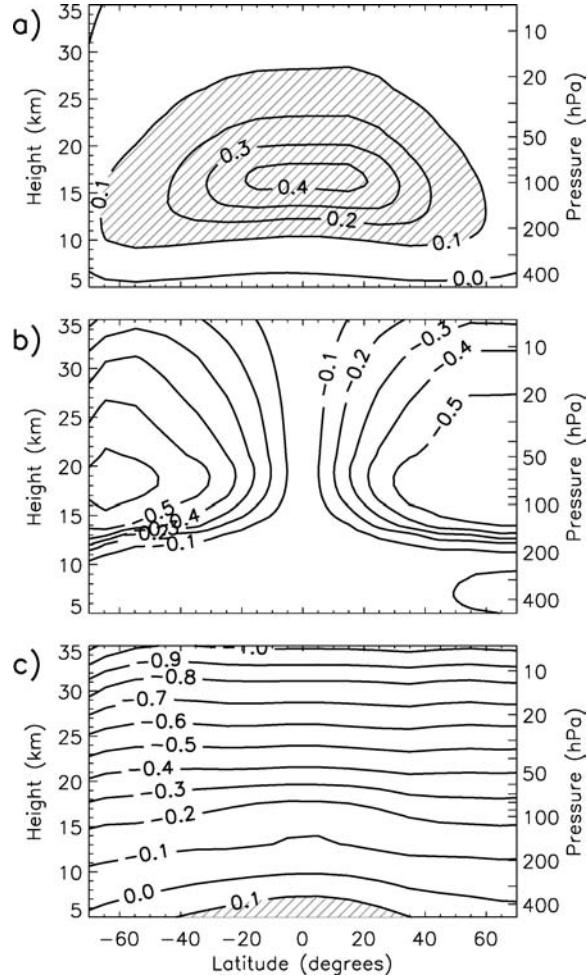


Figure 2. The annually averaged equilibrium FDH temperature change (K) for (a) the total observed halocarbon change (since  $\sim 1950$ ), (b) the 1979–1997 stratospheric ozone change from Randel and Wu (1999) and (c) the 1980–2000 change in carbon dioxide. For the purposes of the FDH calculation the “FDH tropopause” is set at a uniform 5 km altitude.

Figure 3 shows the contribution to the time-evolution of cold-point temperature change for different groups of gases, using the TTCEs from Table I and halocarbon scenario A1B from IPCC-SRES (2000). As the emission scenario for each of the gases is very different, the figure gives a different sense of the importance of each gas, compared to the TTCEs themselves. Halocarbons currently give a total warming of 0.34 K largely from CFC-12 and CFC-11. CFC-12 has an atmospheric lifetime over 100 years, and hence will continue as one of the most important contributors for the remainder of this century. However, the scenario specified here suggests that HFC-134a could dominate by 2100 (nearly all the HFC contribution

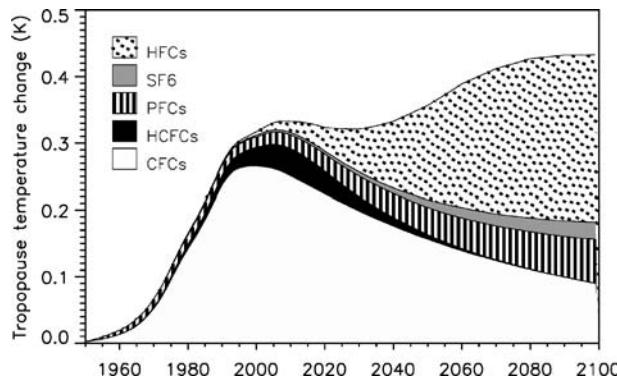


Figure 3. The contribution of halocarbon groups to the tropical cold-point temperature change since 1950 (K). Temperature changes are calculated by multiplying mixing ratios (taken from IPCC-SRES 2000, scenario A1B) with the TTCE for each gas, given in Table I.

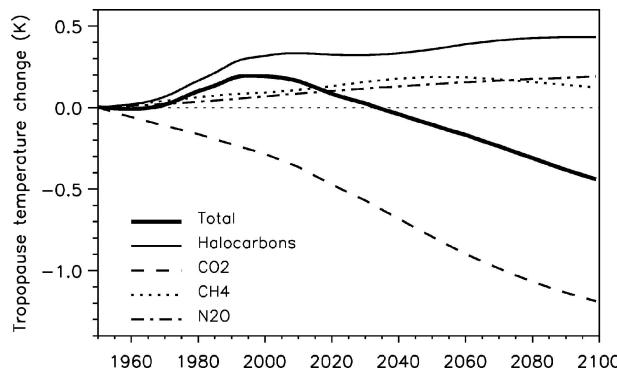


Figure 4. The contribution of different greenhouse gases to the tropical cold-point temperature change since 1950 (K). The net total effect is also shown. Mixing ratios are taken from IPCC (2000), scenario A1B and these are converted to temperature change using the methodology described in the text.

is from HFC-134a). Clearly, the importance of HFC-134a as a strong absorber of infrared radiation and hence warming agent depends not only on its well-established molecular properties but also on future emissions projections, which are uncertain. For the next few decades, a slight reduction in the CFCs may cause a slight cooling; however, overall a further small warming of  $\sim 0.1$  K is expected by the end of the century. Figure 4 compares the halocarbon contribution to that of the three main well-mixed greenhouse gases. For these gases the linear assumption used for summing the halocarbon contribution does not hold. For example a 10 ppmv increase in CO<sub>2</sub> leads to a tropical tropopause temperature change of  $-0.046$  K for a 300 ppmv background value, but only a  $-0.013$  K for a 800 ppmv background value. To address these nonlinearities we calculated the TTCE timeseries by linear interpolating between the results of 20 FDH calculations for each gas spread over their respective mixing ratio ranges. The value for CO<sub>2</sub> includes a component from its

heating at solar wavelengths, as including this was found to reduce the magnitude of its cooling by 35%. It can be seen that halocarbons presently dominate the effects of the other gases. However, towards the latter half of the century we expect CO<sub>2</sub> to dominate and lead to an overall cooling of the tropical cold point. Presently, the total expected warming of the region from WMGHGs since 1950 is  $\sim 0.2$  K. Although potentially important, it is unlikely that this mechanism dominates the response of the tropical tropopause as there is good observational evidence for a general cooling of the tropical tropopause over the last few decades (e.g., Zhou et al., 2001). The size of this cooling is uncertain and its cause the subject of ongoing investigations. However, the halocarbon-related warming suggests that the possible process(es) causing the observed cooling must be operating with greater efficiency than might otherwise be thought. Figure 1b illustrates the potential for other processes to cause temperature changes in this region 1b. For example, it shows that a 10% increase in cirrus cloud fraction or only a 3% increase in water vapour mixing ratio, above 12 km, could compensate for the halocarbon warming. Other potential mechanisms exist for changing tropical cold-point temperatures. For example, the strength of the upwelling and adiabatic cooling associated with the Brewer–Dobson circulation could also have increased.

### 3. Climate Sensitivity

Radiative forcing is defined in IPCC (2001) as the change in net flux at the tropopause after allowing stratospheric temperatures to adjust to radiative equilibrium. It utilizes the notion that the globally averaged radiative forcing ( $\Delta F$ ) is related to the globally averaged equilibrium surface temperature change ( $\Delta T_s$ ) through a constant known as the climate sensitivity ( $\lambda$ ).

$$\Delta T_s = \lambda \Delta F \quad (1)$$

The true climate sensitivity depends on the various feedbacks in the atmosphere and is poorly known (IPCC, 2001 give a factor of 3 uncertainty estimate in  $\lambda$ , from about 0.4 to 1.2 K (Wm<sup>-2</sup>)<sup>-1</sup>). However, climate model studies have shown that  $\lambda$  in an individual climate model is more-or-less independent of forcing mechanism (e.g., Hansen et al., 1997; Joshi et al., 2003). This enables the climate-change potential of different mechanisms to be compared through their radiative forcings, at least to a first-order. Halocarbons and ozone changes may not fit this simple constant-sensitivity picture. Climate models typically show different responses to equivalent forcings from WMGHG changes and CO<sub>2</sub> (Wang et al., 1991, 1992; Govindasamy et al., 2001). In particular, Hansen et al. (1997) found halocarbons to have a  $\sim 20\%$  larger climate sensitivity than CO<sub>2</sub>, mainly due to a stronger positive cloud feedback in their halocarbon experiments. For ozone changes several climate models have now found that ozone perturbations often have a  $\sim 50\%$  different climate sensitivity when compared to that from a carbon dioxide change in same

model and, further, this climate sensitivity depends strongly on the nature of the ozone change employed (Joshi et al., 2003).

As discussed in Section 2, the radiative-convective models results of Forster et al. (1997) have suggested that a lower tropopause at the base of the TTL may be more appropriate for radiative forcing calculations. Forster et al. (1997) find in their radiative-convective model that the climate sensitivity at the lapse-rate tropopause is 6% smaller than that of  $\text{CO}_2$ . This was due to halocarbons preferentially warming the TTL, rather than the surface. Here we test this mechanism in the Reading Intermediate General Circulation Model (RIGCM) (Forster et al., 2001; Joshi et al., 2003). The experimental methodology is described in Joshi et al. (2003) and the RIGCM in Forster et al. (2000). For our experiments we use radiative transfer scheme which was an updated version of that used in Zhong et al. (1996). First, CFC-12 is added to a halocarbon free present day climatology to give a  $1 \text{ W m}^{-2}$  radiative forcing at the lapse-rate tropopause. Equilibrium temperature changes were then calculated for this perturbation, employing perturbed and control integrations of the RIGCM with a 100 m slab ocean. The last 20 years of 50-year integrations were used. A carbon dioxide experiment was also performed, increasing its present day concentration to give a radiative forcing of  $1 \text{ W m}^{-2}$ . Figure 5 shows the zonally averaged temperature change from the two experiments. The different response in the TTL is clearly in evidence (Figure 5c). As a direct consequence of the warming in the TTL the halocarbon change gives  $\sim 6\%$  less warming at the surface, compared to carbon dioxide (0.441 K for CFC-12 and 0.469 K for  $\text{CO}_2$ ). This 6% smaller climate sensitivity for halocarbons compared to  $\text{CO}_2$  agrees well with the earlier radiative convective model result. It also qualitatively agrees with Govindasamy et al. (2001), who found a 20% smaller climate sensitivity for WMGHG changes, compared to that for equivalent  $\text{CO}_2$ . However, it does not agree with the finding of Hansen et al. (1997), who found a  $\sim 50\%$  larger climate sensitivity for halocarbons. Resolving these differences could be an important avenue of future research.

The climate sensitivity for stratospheric ozone changes has been investigated in studies with the RIGCM (Forster and Shine, 1999; Joshi et al., 2003) and other climate models (Hansen et al., 1997; Christiansen 1999; Joshi et al., 2003). It is one of the radiative forcings whose climate sensitivity differs most from that of carbon dioxide. The results of past ozone experiments are comprehensively reviewed and discussed in Joshi et al. (2003). The most important findings relevant to our paper are that, firstly, Forster and Shine (1999) found a 40% larger climate sensitivity for an observed 1979–1997 latitude-height resolved ozone change applied in the stratosphere, compared to that for  $\text{CO}_2$ . Secondly, Joshi et al. (2003) find that three quite different climate models including the RIGCM have a climate sensitivity for idealised lower stratospheric ozone changes that is consistently 25–80% higher than their sensitivity to  $\text{CO}_2$  changes. For these models changes in stratospheric water vapour provided a positive feedback that was the primary cause of their larger sensitivities.

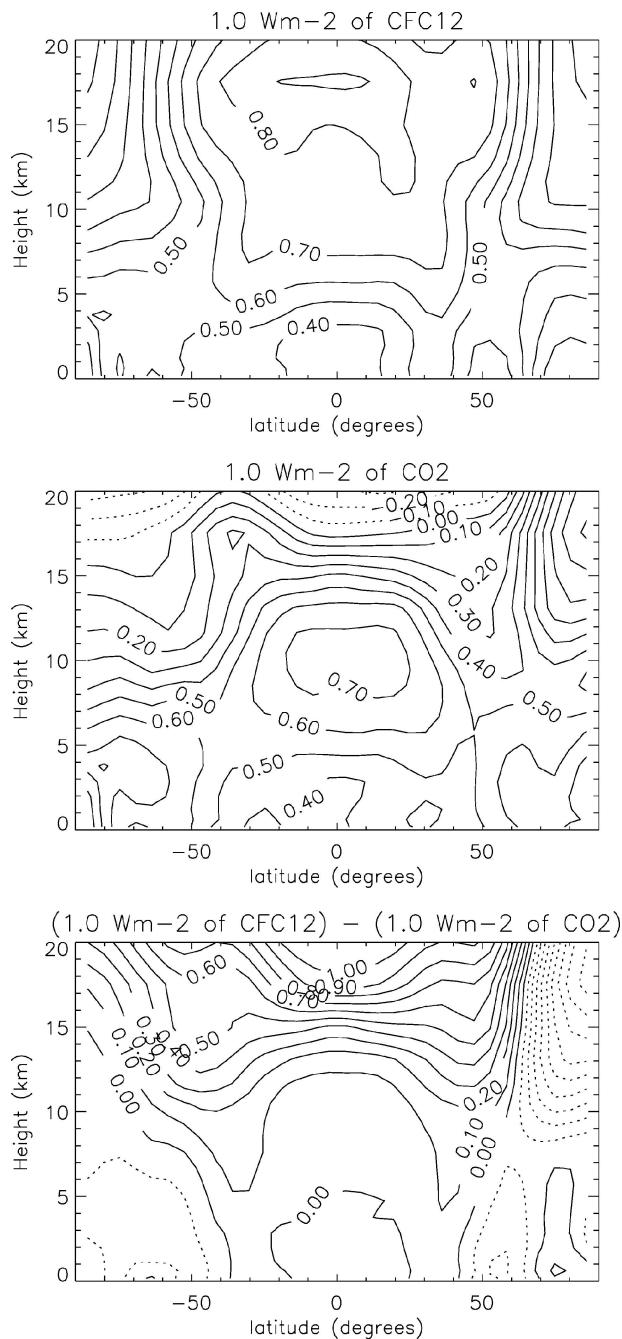


Figure 5. The zonally averaged equilibrium temperature changes from 30-year integrations with the ICGM, using a standardized global mean radiative forcing of  $1 \text{ Wm}^{-2}$ . (a) For a uniform increase of CFC-12; (b) for a uniform increase in CO<sub>2</sub> and (c) shows the difference of the CFC-12 response compared with the CO<sub>2</sub> response. The contour interval is 0.1 K and dotted lines are negative values.

In the next section, where the direct and indirect surface temperature changes are evaluated for the halocarbons, the same climate sensitivity is used for all forcing mechanisms. Differences in the climate sensitivities discussed in this section could easily be taken into account by, for example, reducing the halocarbon sensitivity by 6% and increasing the stratospheric ozone sensitivity by 40%. However, we have chosen not to do this as we feel that there is still considerable uncertainty in both the stratospheric ozone and halocarbon sensitivities.

#### 4. Surface Temperature Changes

As stated in Section 1, the role of halocarbons in surface climate change can be assessed through their GWP. This is a simple measure of the radiative impact on climate over a given time frame per kilogram emitted, relative to a reference gas. It is possible to use a similar methodology to calculate an indirect GWP associated with the ozone loss of the halocarbon; examples of this are in previous ozone assessments (e.g., Table I–VIII, WMO, 2003). Several assumptions enter into this indirect ozone-loss calculation; the most important of these is the background climate into which the halocarbon is introduced, as, for example, in the present climate there is little room for further Antarctic ozone destruction (WMO, 2003). The possible maximum and minimum values for this net GWP also differ by roughly a factor of 5, based on the range of calculated stratospheric ozone radiative forcings taken from IPCC (2001). Given this large degree of uncertainty and the complexity of ozone–climate interactions, we choose to adopt an even simpler metric and compare ozone and halocarbon radiative forcings directly and then using these calculate the likely surface temperature change, using a simple model. Assumptions are still needed; however, in lieu of multiple chemistry–climate model integrations for each halocarbon that are in consensus, a rough zero-order estimate is arguably the most practical way of assessing their net effect on surface climate.

Another quantity of interest, related to but different from the GWP, is radiative forcing. Radiative forcing estimates are indicative of *equilibrium* surface temperature changes due, not to the per kilogram emission, but to the increase of gas observed in the atmosphere today compared to pre-industrial times. In reality, the Earth's surface temperature change is transient and a good deal, but not all, of the temperature change(s) associated with the WMGHG radiative forcing would have already have been realised in the observed surface warming of the last 150 years (IPCC, 2001). The *rate of change* of the radiative forcing is potentially more important for evaluating *short-term* climate change (Solomon and Daniel, 1996). Their paper estimates a positive forcing change of  $0.3 \text{ Wm}^{-2}$  between 1985 and 2005, which is more than five times larger than the rate of change of radiative forcing due to other WMGHGs. They conclude that the combined effects of ozone recovery and halocarbon increases are expected to have a large role in surface temperature change over the next few decades. However, due to the relatively short timescale

of these radiative forcing changes not all of them will probably be realised in the ocean temperature response. Therefore, in an attempt to gauge a more complete picture of their role we use a simple model to convert radiative forcing timeseries into surface temperature timeseries.

#### 4.1. RADATIVE FORCING TIMESERIES

The direct and indirect halocarbon radiative forcings are compared in Figure 6 (see figure legend for details). The halocarbon timeseries used are the same as those in Section 2 and are converted to radiative forcings using the radiative efficiencies from WMO (2003). The indirect forcing from ozone loss is based on  $-0.15 \text{ Wm}^{-2}$  over 1979–1997, from IPCC (2001). When comparing the direct and indirect effects of the halocarbons a factor of two uncertainty in the ozone-loss indirect forcing needs to be kept in mind (illustrated in Figure 6e). Nevertheless, the figure gives a first-order estimate of the likely global mean tradeoffs involved with assessing halocarbons role in surface climate. Whilst CFCs are declining they are still forecast to be the most important halocarbon for the next  $\sim 40$  years. HFCs are forecast to grow to dominate the radiative forcing in the latter half of this century (Figure 6c). Halocarbon direct radiative forcing has risen sharply since 1970. However, a significant portion of this has been offset by the ozone loss due to the halocarbons. Whilst little direct forcing increase is expected over the next 100 years, ozone recovery is expected to effectively double the net halocarbon forcing over the next  $\sim 50$  years. (This  $\sim 0.1 \text{ Wm}^{-2}$  is 5% of the increase likely from  $\text{CO}_2$ .) The CFCs, HCFCs, HFCs and PFCs all can be seen to have a net positive radiative forcing, despite the ozone loss associated with the CFCs and HCFCs. For the other ozone-depleting gases their net forcing is negative over the next 50 years (Figure 6d). In summary, in terms of the global mean radiative forcing, since 1980 ozone depletion has more-or-less effectively masked the effect of halocarbon increases. However, in the next few decades, as ozone recovers we are likely to realise relatively large increases in the halocarbon forcing and we would therefore expect them to play an important role in the next few decades of temperature change. This is evaluated next.

#### 4.2. SIMPLE CLIMATE MODEL

To crudely investigate the impact of the radiative forcing timeseries, a simple energy-balance model coupled to a deep ocean via diffusion is used. We employ a climate sensitivity parameter of  $0.67 \text{ K} (\text{Wm}^{-2})^{-1}$  (approximately equivalent to a  $2.5 \text{ K}$  equilibrium surface warming for doubling  $\text{CO}_2$ ). The resulting transient surface temperature changes are shown in Figure 7. Figure 7a shows that with this choice for climate sensitivity the net effect of halocarbons is to increase surface

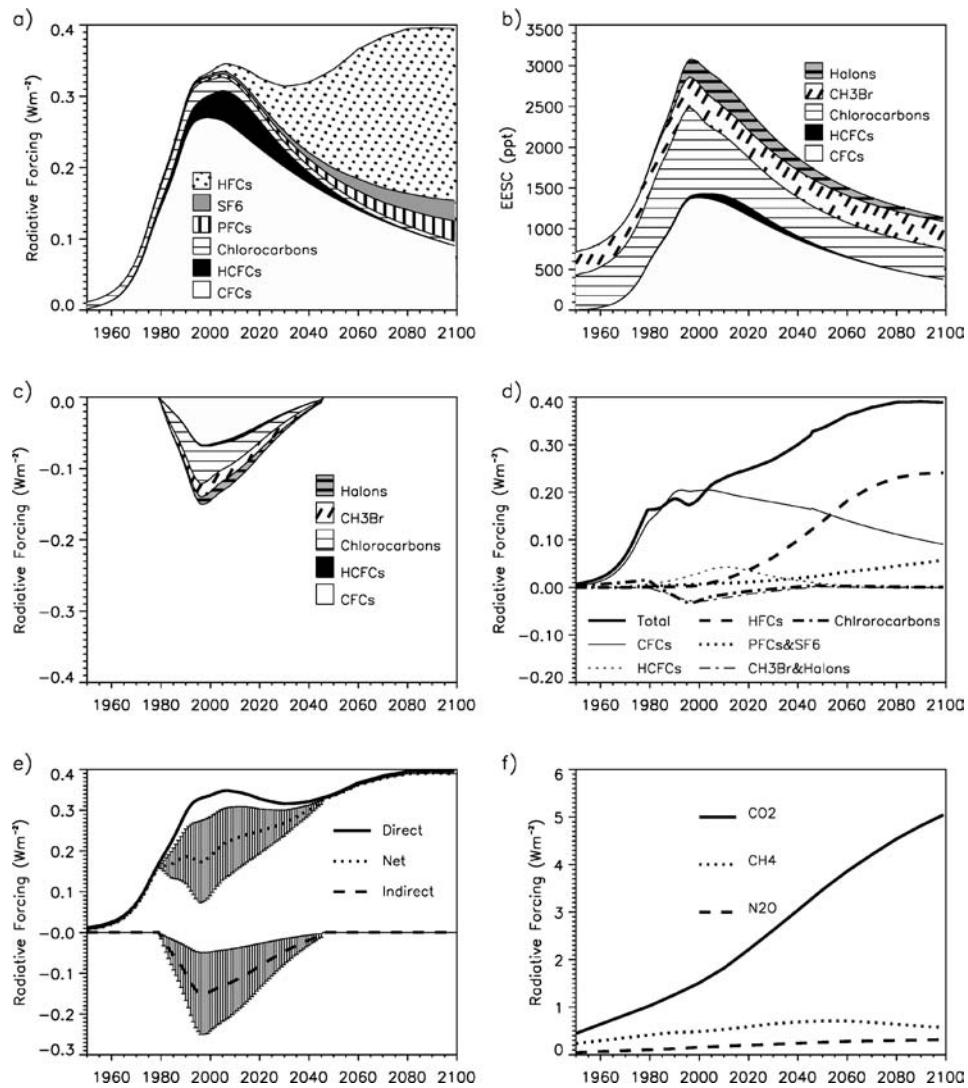


Figure 6. The breakdown of the (a) total direct radiative forcing, (b) the equivalent effective stratospheric chlorine (EESC), (c) indirect radiative forcing and (d) net radiative forcing timeseries for individual halocarbon groups. In (e) the total direct, indirect and net radiative forcing is shown, with error bars arising from uncertainties in the stratospheric ozone radiative forcing from IPCC (2001). For (a)–(c) the individual forcings have been stacked to show the total forcing. Scenarios are taken from IPCC-SRES (2000), A1B. The EESC has been calculated using data in WMO (2003), assuming that bromine is 45 times more effective at destroying ozone than chlorine. The change in EESC between 1979 and 1997 is assumed to give an indirect radiative forcing of  $-0.15 \text{ W m}^{-2}$  (IPCC, 2001). This radiative forcing scales with EESC amounts above the 1979 threshold. For comparison, (f) shows the SRES A1B radiative forcings for the other WMGHGs.

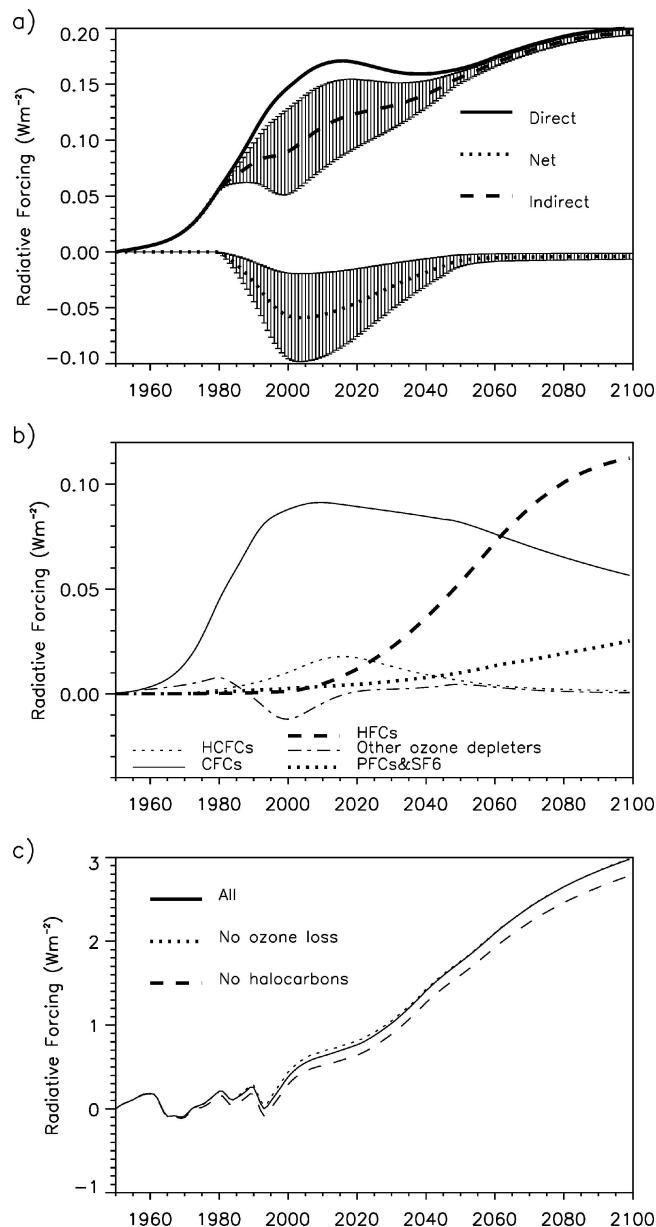


Figure 7. Shows the temperature change calculated with our simple climate model, using the radiative forcing timeseries in Figure 6. (a) The direct, indirect and net temperature change and its uncertainty; (b) the breakdown of the temperature change into species components; and (c) the total temperature change, using many natural and anthropogenic past forcings and a future WMGHG radiative forcing – see text for details. Lines are also shown for the same timeseries without stratospheric ozone loss or without both the direct and indirect effects of halocarbons.

temperature by  $\sim 0.2$  K by 2100 and stratospheric ozone depletion has offset up to approximately half of the predicted past increases in surface temperature due to the direct effect of halocarbons. As ozone will likely recover in the next few decades, we predict a slightly faster rate of warming. CFCs have contributed most to these temperature changes (Figure 7b) and the HFCs (HFC-134a, in particular) are predicted to dominate temperature changes over the coming century.

To put the halocarbon and ozone-loss surface temperature change into context with the overall expected temperature change, we performed additional calculations using realistic past and future total radiative forcing timeseries (Figure 7c). The timeseries of the extra natural and anthropogenic forcings given by Myhre et al. (2001) are used for the past, and IPCC SRES scenario A1B (IPCC, 2001) is used for the future. The past time-dependent forcing estimate includes forcings due to other well-mixed greenhouse gases, tropospheric ozone, direct and indirect aerosol, solar and volcanism. It is similar to that in IPCC (2001) (Figure 6.8) but, in addition, includes a mid-range estimate for the indirect aerosol forcing. No solar or volcanic forcing is assumed in the future scenarios. It can be seen that the net effect of halocarbons is expected to bring next century's warming forward 20 years based upon the adopted emission scenario, which is one near the middle of the range suggested in IPCC (2001). Stratospheric ozone depletion has led to a slightly reduced temperature rise over the past  $\sim 15$  years. However, its temperature change signal is swamped by the large temperature rise from the combination of the post-Pinatubo (1991) increase in radiative forcing and an increasing solar forcing.

## 5. Conclusions

We have shown that halocarbons significantly warm the TTL and, in this region, since 1950 they would be expected to cause a 0.3–0.4 K temperature increase, dominating the effect of the other WMGHGs. This number can be placed in the context of the  $\sim 0.5$  K surface warming over the last 100 years and the  $\sim 2$  K cooling of the lower stratosphere over the last 20 years. The fact that such temperature increases are not seen suggest that some other mechanism(s) are highly likely to be compensating for this by cooling the TTL. As stratospheric ozone depletion has cooled the high latitude lower stratosphere, the net effect of halocarbons on temperatures in the lower stratosphere is to reduce the equator–pole temperature gradient. These increases in the temperature at the tropical cold point could be important for a number of potential climate forcing mechanisms, such as stratospheric water vapour changes and meridional circulation changes. We suggest that as well as surface-based metrics (radiative forcing, GWP, etc.), it is also useful to evaluate climate change mechanisms away from the surface, using metrics such as the TTCE to obtain a fuller picture of a given mechanism's climate role.

We find halocarbons (specifically CFC-12) have a  $\sim 6\%$  lower global climate sensitivity in the RIGCM than carbon dioxide, when radiative forcing is calculated

at the lapse rate tropopause. This result agreed with two earlier studies, but not with the results of Hansen et al. (1997). Reasons for these differences remain unclear.

For surface temperature changes over the past 50 and next 100 years halocarbons are also important, but are by no means dominant. Based upon the scenario chosen here for emissions projections, it is possible that they could bring forward the expected warming over the coming century by  $\sim 20$  years or 0.2 K, with ozone depletion providing only a small offset to this, which is near a maximum for the climate of today.

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