Contributors to Future Stratospheric Climate Change: An Idealized Chemistry–Climate Model Sensitivity Study

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Abstract

Within the framework of an idealized model sensitivity study, three of the main contributors to future stratospheric climate change are evaluated: increases in greenhouse gas concentrations, ozone recovery, and changing sea surface temperatures (SSTs). These three contributors are explored in combination and separately, to test the interactions between ozone and climate; the linearity of their contributions to stratospheric climate change is also assessed.

In a simplified chemistry–climate model, stratospheric global mean temperature is most sensitive to CO_2 doubling, followed by ozone depletion, then by increased SSTs. At polar latitudes, the Northern Hemisphere (NH) stratosphere is more sensitive to changes in CO_2, SSTs and O_3 than is the Southern Hemisphere (SH); the opposing responses to ozone depletion under low or high background CO_2 concentrations, as seen with present–day SSTs, are much weaker and are not statistically significant under enhanced SSTs. Consistent with previous studies, the strength of the Brewer–Dobson circulation is found to increase in an idealized future climate; SSTs contribute most to this increase in the upper troposphere/lower stratosphere (UT/LS) region, while CO_2 and ozone changes contribute most in the stratosphere and mesosphere.

1 Motivations

Braesicke et al., 2006; hereafter BHP2006) examined the stratospheric sensitivity to ozone depletion and to the doubling of CO_2. Their study used a sea surface temperature (SST) climatology with a repeating annual cycle, representative of the late 20th century. That is, SSTs did not increase in response to increased greenhouse gas concentrations; their experiments primarily examined the stratospheric radiative impact of increased CO_2.

Using the same simplified chemistry–climate model (CCM) as in BHP2006, the relative response to changes in CO_2 and O_3 concentrations and sea surface temperatures (SSTs) is explored; this approach considers the combined stratospheric response to warming from both the troposphere and the upper ocean by prescribing 'future' SSTs. To separate the three proposed contributions to stratospheric climate change, global mean temperature, eddy heat flux, winds and ozone are diagnosed in each of a set of idealized time–slice experiments. The strength of the Brewer-Dobson circulation, and the connection between tropical upwelling and polar ozone in an idealized present–day climate scenario is compared with an idealized future climate.

BHP2006 found that global mean temperatures cooled in response to both an O_3 change (2000–like – 1980–like) and a CO_2 change (704ppmv – 352ppmv), throughout the middle atmosphere. However, neither change much affected tropospheric temperatures because the same SST climatology was prescribed in all experiments. Recent trends in observed SSTs, as well as coupled ocean–atmosphere simulations of the 21st century, suggest that anthropogenic climate change will continue to affect the temperature of the sea surface (i.e. Johns et al., 2003). When 'future' SSTs are used in conjunction with a doubled–CO_2 atmosphere, the troposphere should respond by warming more significantly than for a doubling of CO_2 alone. Also, there may be feedbacks between these increased tropospheric temperatures and global mean temperature in the stratosphere that can be considered when CO_2 concentrations and SSTs are increased together.

BHP2006 showed that two dynamical relationships held for a set of four idealized climate change simulations in a simplified CCM. First, the negative correlation between zonal mean zonal wind (a proxy for polar vortex strength) and total O_3 at Northern Hemisphere (NH) high latitudes in January, as was originally discussed by Braesicke and Pyle (2004). Second, high–latitude temperature was seen to mimic changes in mid-latitude heat flux (the 'tropospheric forcing' by planetary waves) in the December–January–February (DJF) season (e.g., as shown by Newman et al., 2001). This paper will address how warmer SSTs affect the character of these two relationships.

In the time–slice experiments with present–day SSTs evaluated by BHP2006, it was noted that the behavior of the NH polar vortex (and thus of polar ozone) depended on the background CO_2 concentration: The single–CO_2 experiments responded to ozone depletion in the opposite sense to the doubled CO_2 experiments; this effect provided an example of the competition between radiative and dynamical processes in the polar stratosphere. This paper will determine whether increased SSTs enhance cooling in the middle atmosphere, thus favoring stronger polar vortices in all experiments. Do 'future' SSTs affect the coupling between ozone depletion and tropospheric forcing? Sections 3, 4 and 5 will assess the response of the UM chemistry–climate model to changes in O_3, CO_2 and SSTs with a set of transport, dynamical, radiative and chemical diagnostics. Section 6 will summarize the main conclusions.

2 Methods

2.1 Model Description

The following discussion refers to a set of eight time–slice integrations conducted with version 4.5.1 of the MetOffice Unified Model (UM). In this configuration, the UM has 3.75° x 2.5° horizontal resolution and 64 vertical levels, with ~1.3km resolution in the stratosphere. The climate model is coupled non-interactively with the Cariolle and Déqué (1986) parameterized stratospheric ozone chemistry scheme. The chemical module contains a cold tracer (X) used to mimic the impact of polar stratospheric clouds (PSCs) on polar ozone: when temperatures drop below a given threshold (~195K for nitric acid trihydrate at 50hPa) the cold tracer is produced exponentially with a time constant of four hours; the cold tracer decays with a ten–day time constant. This model setup has been used previously and documented by Braesicke and Pyle (2003, 2004) and Pyle et al. (2005).

2.2 Experimental Design

Eight 20–year time–slice experiments will be discussed in this paper (see Table 1). Each experiment tests the combination of one of two CO_2 concentrations (1xCO_2 or 2xCO_2), stratospheric ozone climatologies (1980–like or 2000–like), and SST and sea ice climatologies (present–day or future). Differences between pairs of experiments can be examined so as to isolate the effects of changes in each of the three parameters (ozone, CO_2 and SSTs) on the climate of the middle atmosphere. Since one of the time–slice experiments represents an idealized present–day climate (1B; low CO_2, depleted ozone layer and present–day SSTs) and another represents the likely stratospheric climate in the mid– to late 21st century (2C; high CO_2, recovered ozone layer and future SSTs), the 2C–1B difference can be interpreted as the 'climate change signal' (see WMO, 2007).

A 1980–like ozone climatology is prescribed in experiment 1A, whereas a 2000–like ozone climatology, with substantial polar ozone deficits as compared with the 1980–like climatology, is prescribed in experiment 1B. Experiments 1A and 1B use a background CO_2 concentration of 352ppmv. Experiments 2A and 2B are designed to investigate the same change under doubled CO_2 (704ppmv) conditions. As discussed by BHP2006, annually repeating boundary conditions are imposed in all four experiments: AMIP II[1] sea surface temperature (SST) and sea ice climatologies representative of the late 20th century[2]. Volcanic aerosols and the solar cycle are not considered. Experiments 1C to 2D are identical to experiments 1A to 2B, except that they use the 'future' SST climatology described in the next section. The difference in the setup between experiments 1A and 1C, for example, is solely a switch from the present–day to the future SST climatology, as is the difference between experiments 1B and 1D, 2A and 2C, and 2B and 2D.

2.3 Construction of a 'Future' SST Climatology

Including SSTs as a factor in this study, and thus examining the impact of the ocean surface on the stratospheric chemistry–climate system, requires a 'future' SST climatology. While some CCMs now include an interactive ocean model, UM 4.5.1 is an atmosphere–only model and future SSTs derived from another ocean–atmosphere model simulation must be prescribed. In the present model study, a future SST dataset is constructed by adding a twelve–month set of SST anomalies to the existing present–day SST climatology. A MetOffice SST dataset, spanning from 1970 to 2020, merges HadISST[3] data (1970–1995) with SST and sea ice output from HadGEM1 simulations (beginning in 1995). The 'climate change' SST anomalies are defined as the difference between mean SSTs in the 1970s (1971–1980) and mean SSTs the 2010s (2011–2020) from this MetOffice dataset.

[1] http://www-pcmdi.llnl.gov/projects/amip/AMIP2EXPDSN/BCS_OBS/amip2_bcs.htm.

[2] The 'present–day' SST and sea ice climatologies were defined as the AMIP II 1979-1996 mean.

[3] http://hadobs.metoffice.com/hadisst.

The 2010s – 1970s differences are large enough to simulate differences between present–day SSTs and those projected for the mid–21st century, and thus provide a stratospheric response.

The idealized 'climate change' SST anomalies (i.e. future – present–day) are generally positive. In the tropics and at mid–latitudes, anomalies are of the order of 1–2K (consistent with coupled ocean–atmosphere predictions of SST changes by the mid–21st century; see IPCC, 2007). The largest positive differences occur at high latitudes. Near the Gulfstream and Kuroshio currents, the SST changes exceed 10K; this is larger than predicted by most ocean–atmosphere models (IPCC, 2007). In the tropical Pacific Ocean, the climate change anomalies are positive but small (up to 1.5K) in the Intertropical Convergence Zone (ITCZ), and negligible or slightly negative to the north and south of this region. The strongest positive–negative–positive pattern occurs in the DJF season; this is the pattern of SST anomalies that defines El Niño events. There is a positive trend in the 1970–2020 timeseries of HadGEM1 SSTs in the Niño 3.4[4] region. This finding is in agreement with Timmermann et al. (1999), who predict a climate change–induced shift toward an increasingly positive Niño 3.4 index, and thus toward more frequent El Niño events, in future.

3 Radiative and Dynamical Response to Changes in Ozone, CO_2 and SSTs

3.1 Global Mean Temperature

Profiles of global and annual mean temperature differences allow for an easy assessment of overall radiative changes, suppressing dynamical changes (important at seasonal timescales) and their effects on the thermal structure of the atmosphere. Consistent with Shine et al. (2003, 2008), in the idealized time–

[4] Monthly mean SST anomalies from the 1950–1999 period in the 120°W–170°W, 5°S–5°N region (see Trenberth, 1997).

slice experiments, the middle atmosphere cools in response to both ozone depletion (with a small peak in the lower stratosphere and a larger peak centered at 1hPa) and increased CO_2 concentrations (with the strongest cooling at the stratopause). Profiles of the four sets of responses to ozone depletion (Fig. 1, left–hand panel) are indistinguishable from 1000 to 0.1hPa; the responses are the same for both SST climatologies (i.e. 1B–1A ≈ 1D–1C) and for both CO_2 concentrations (1B–1A ≈ 2B–2A). That is, the global mean temperature response to ozone depletion is independent of both background CO_2 and thermal forcing from the sea surface. Similarly, the response to doubled CO_2 (Fig. 1, centre panel) is the same for both ozone climatologies (i.e. 2A–1A ≈ 2B–1B) and for either present–day or future SSTs (2A–1A ≈ 2C–1C).

The troposphere and lower stratosphere are warmer in the future SST experiments than in the experiments using present–day SSTs (Fig. 1, right–hand panel). The magnitude of the SST–related change in global mean stratospheric temperature is far smaller than found for CO_2 doubling or for ozone depletion: Under future SSTs, the troposphere warms as much as 1.3K, while the lower stratosphere warms by ~0.25K. Similarly to the response to ozone depletion and doubled CO_2, the response to the change in SSTs is similar under the differing ozone and CO_2 conditions. Small differences between the four pairs of experiments arise in the upper troposphere/lower stratosphere region, likely due to mismatches in the thermal forcing between the prescribed SSTs and atmospheric greenhouse gas concentrations.

As in the annual mean, future SSTs enhance warming in the upper troposphere/lower stratosphere region and raise the level where no temperature difference occurs throughout the seasonal cycle (not shown; see Hurwitz, 2008). This finding agrees with previous studies (i.e. Chakrabarty et al., 2001; Fomichev et al., 2007; Lorenz and DeWeaver, 2007) that have found a link between climate change and tropopause height.

3.2 Wintertime Lower Stratospheric Temperature, Eddy Heat Flux and Geopotential Height

Tropospheric forcing by planetary waves has a large influence on stratospheric temperatures, particularly in winter. Previous studies (e.g., Newman et al., 2001; Austin et al., 2003; Cagnazzo et al., 2006) have used the zonal mean eddy heat flux at 100hPa, averaged over a mid–latitude band (40°N/S and 80°N/S) and over a two–month time period, to diagnose the tropospheric forcing. Fig. 2 shows the modeled heat flux with respect to the 50hPa polar temperature. For the NH, December–January (positive) heat flux is plotted against January–February temperature, while for the SH, August–September (negative) heat flux is plotted against September–October temperature. Fig. 2 shows that, for the set of eight experiments, the relationship between tropospheric forcing and polar temperature is linear and positive[5]: Increased tropospheric forcing in the early winter leads to increased polar temperatures at higher altitudes somewhat later in the winter. As expected, Southern polar temperatures are lower (by roughly 20K) than are northern polar temperatures.

The slopes of the regression lines are steeper in the NH than in the Southern Hemisphere (SH). The mean slope of the eight experiments is 1.29 m^{-1} s in the NH and –0.80 m^{-1} s in the SH. These values are in agreement with the analysis by Austin et al. (2003) that found the heat flux–temperature slopes based on the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis (Kalnay et al., 1996) to be 1.49 ± 0.27 m^{-1} s (NH) and –0.89 ± 0.16 m^{-1} s (SH). This result suggests that the NH polar region is

[5] Heat flux is poleward in both hemispheres: positive (negative) values indicate northward (southward) heat flux.

more sensitive to changes in tropospheric forcing than is the southern polar region.

For the NH winter season stronger heat fluxes and lower temperatures are found in experiment 2C (idealized future climate; burgundy) relative to experiment 1B (idealized present–day climate; turquoise). This result agrees with the work of Manzini et al. (2003), who found a downward shift of the heat flux versus temperature regression (perpendicular to the original slopes)) between 1960–like and 2000–like time–slice simulations. In the UM experiments, however, the shift toward stronger heat fluxes and lower temperatures is not significantly larger than the signal due to interannual variability.

In contrast to the NH, the SH interannual variability is lower, and there is little change in the heat flux and temperature values between the eight experiments. This suggests that the dynamics of the SH stratosphere are not as sensitive to changes in greenhouse gases, SSTs or ozone climatologies, though SH winters are slightly cooler in the doubled–CO_2 experiments. Rather, the SH polar stratosphere is closer to being in radiative equilibrium as compared with the NH.

As noted by BHP2006, when present–day SSTs are prescribed, DJF heat flux differences due to ozone depletion are positive in a single CO_2 atmosphere (1B–1A) but negative in a doubled CO_2 atmosphere (2B–2A). These differences[6] are shown in the two leftmost bars in Fig. 3. While the responses to ozone depletion have the opposite sign when future SSTs are prescribed, the 1C–1D heat flux difference is not statistically distinct from the 2D–2C difference. Consistent with the heat flux differences described above, temperature and geopotential

[6] In figure 3, heat flux differences due to ozone depletion are shown as 1A–1B, 2A–2B, etc. as this clarifies the three contributions to the climate change signal, i.e. 2C–1B = (2C–2A) + (2A–1A) + (1A–1B).

height differences due to ozone depletion are small in the experiments using future SSTs (see Hurwitz, 2008).

The error bars in Fig. 3 reveal the high degree of interannual variability in tropospheric forcing during the NH winter season, and thus the difficulty in evaluating heat flux differences and their subsequent effects on stratospheric dynamics. The only heat flux differences from that are statistically different from zero are the three changes of parameter relative to experiment 1A (ozone depletion, doubling of CO_2 and increasing SSTs). While heat flux increases due to climate change (2C–1B), dynamical warming of the lower stratosphere is overwhelmed by the radiative cooling associated with doubled CO_2 and increasing SSTs. In the NH winter, 2C–1B temperature and geopotential height differences at 50hPa are generally negative (up to 3K and 30m, respectively; not shown).

4 Chemical Response to Changes in Ozone, CO_2 and SSTs

4.1 Relationship Between Polar Vortex Strength and Ozone at NH High Latitudes

The strong relationship between NH polar vortex strength and high–latitude total ozone seen in other CCM experiments (e.g., Braesicke and Pyle, 2004; BHP2006) extends to the four experiments using future SSTs. The slope of the regression line fitting each set of 20 points in experiments 1C through 2D is very similar to that seen in experiments 1A, as is the range of zonal wind and total ozone values (Fig. 4a).

Fig. 4b shows the mean regression line of all eight 20–year time–slice experiments as well as the mean zonal wind and total ozone values for each experiment. As noted by BHP2006, experiments where present–day SSTs are prescribed (1A–2B) exhibit a 'flip flop' response to ozone depletion. The means of the future SST experiments (1C through 2D) are situated between these two states (1A/2B and 1B/2A); differences in the mean zonal winds and ozone in the future SST experiments are not statistically significant.

4.2 Polar Ozone Loss as a Function of PSC Volume

Rex et al. (2004, 2006) found a linear relationship between column ozone loss and PSC volume during the NH winter, in observations of the past two decades. BHP2006 examined this same relationship in four UM experiments using present–day SSTs (1A–2B). The strongest correlation between ozone loss and PSC volume was found in experiment 1A (1980–like ozone and 1xCO_2; r = 0.98), while somewhat weaker linear relationships were found in two other experiments (1B and 2A). In experiment 2B (2000–like ozone and 2xCO_2), ozone losses were clustered around 90DU despite interannual variation in PSC volumes and polar temperatures.

The relationship between ozone loss and PSC volume is examined in the future SST experiments (1C–2D). Just as for experiments 1A–2B, PSC volume is defined as the sum of all grid cells where cold tracer values exceed a fixed threshold (0.95). For each time–slice experiment and for each winter season, the DJF average PSC volume is then sorted into size classes (bin width $\Delta = 1 \cdot 10^7$ km^3; bin overlap $\delta = 0.5 \cdot 10^7$ km^3). Also, the modeled January mean polar temperature (at 30hPa, north of 85°N), and the wintertime column ozone loss from November to March (within the 400 to 550K potential temperature layer) are calculated for each winter and sorted according to the associated PSC volume size class. The linear relationship between ozone loss and DJF PSC volume in experiments 1C and 2D (1A and 2B with future SSTs) have slopes nearly identical to that found in 1A and a good fit (r ~ 0.85). Another of the future SSTs experiments (1D) has a similar slope but a lower correlation coefficient. The idealized future climate scenario (2C) exhibits the same 'saturation' behavior as does experiment 2B: wintertime column ozone losses do not correlate with PSC volumes.

In the NH winter, PSC volumes and polar temperatures are related by a power law. The highest mean wintertime PSC volumes generally correspond with the lowest mean January polar temperatures, as PSC formation is highly temperature-dependent. Three of the future SST experiments have continuous temperature distributions, similarly to experiment 1A (refer to BHP2006, Fig. 7). The idealized future climate scenario (2C) has a bimodal distribution, as seen in two of the experiments with present–day SSTs (1B and 2A).

5 Sensitivity of the Brewer–Dobson Circulation to Ozone Depletion and Climate Change

Tropospheric forcing changes in response to ozone depletion/recovery and climate change, particularly during the NH winter season, are likely to be linked to changes in the strength of the Brewer–Dobson circulation (BDC; originally described by Brewer (1949) and Dobson (1956)). Tropospheric forcing correlates not only with stratospheric mid-winter temperatures in the lower stratosphere (e.g., as shown in Fig. 2) and with polar ozone, but is also connected to the residual circulation in the middle atmosphere. Newman et al. (2001) calculated that a 10% reduction in the 100hPa eddy heat flux would weaken the BDC by 10%. Conversely, recent increases in tropospheric forcing (as discussed by Dhomse et al., 2006) may have caused a strengthening of the stratospheric circulation. Modeling studies by Butchart and Scaife (2001), Austin and Li (2006) and Li et al. (2008) have shown that increased greenhouse gas concentrations lead to a strengthened BDC in the middle atmosphere. Increased heating near the equator and thus increased upward mass flux in the tropics is a key part of the mechanism that links greenhouse gas concentrations with a stronger Brewer–Dobson circulation (Eichelberger and Hartmann, 2005). Butchart and

Scaife (2001) and Li et al. (2008) found increased tropical upwelling in conjunction with increased downwelling at high latitudes, in climate change simulations, but could not provide an unambiguous mechanism for the strengthening. Climate change–related strengthening of the BDC has not been observed as yet (Engel et al., 2009).

The present study will separate the effects of each parameter change (ozone depletion/recovery, increased greenhouse gas concentrations and increased SSTs) on the strength and character of the BDC. This study complements a recent study by Oman et al. (2009), which examined changes in the age of stratospheric air in transient simulations of the recent past and future.

5.1 Qualitative Streamfunction Analysis

For each of the eight time–slice experiments, the residual streamfunction is calculated following Andrews et al. (1987). A latitude–height cross–section of the residual streamfunction in the idealized present–day climate simulation (1B), for the DJF season, is shown in Fig. 5a. Many features of the observed meridional circulation are reproduced: first, the separation of transport toward the South Pole (negative contours) from transport toward the North Pole (positive contours) is located just south of the equator (due to the southward shift of the ITCZ during the NH winter season). The strongest transport occurs in the troposphere; the Hadley and Ferrell cells can be seen in the tropics and at mid-latitudes, respectively, and in both hemispheres. The larger but weaker BDC (i.e. following the ±0.1 kg s^{-3} contours) is characterized by upwelling in the tropics, poleward transport through the stratosphere, and downwelling at high latitudes (particularly in the NH, during the DJF season). Mesospheric transport also features equatorial upwelling and downwelling at high latitudes of the winter hemisphere, though the winter pole is favored.

Pressure–weighting the streamfunction highlights the behavior of the middle atmosphere; an example is shown in Fig. 5b. In experiment 1B, during the DJF season, the largest magnitudes occur above 5hPa, with another region of strong upwelling in the equatorial upper troposphere.

For an in depth comparison of streamfunction values, seven regions are defined. As shown in Fig. 5b, these regions provide good coverage of the features of the BDC. Each of the seven boxes covers a 20° latitude band and spans four model pressure levels. Region 1 is located in the equatorial lower stratosphere (10°S to 10°N). Regions 2, 3 and 4 are located in the upper stratosphere; region 3 is centered at the equator, while regions 2 and 4 are located in the SH and NH high latitudes (60° to 80° latitude) respectively. Regions 5, 6 and 7 are located in the mesosphere, region 6 at the equator, and regions 5 and 7 in the SH and NH mid- to high latitudes (40° to 60° latitude).

Seasonal differences in pressure–weighted streamfunction values generally have the same sign in the equatorial upper troposphere (region 1) as in the mesosphere (regions 5, 6 and 7; not shown). Doubling the background CO_2 concentration (for example, 2A–1A or 2C–1B) leads to small changes in streamfunction values in the equatorial UT/LS and larger changes in the upper stratosphere, in the winter hemisphere. These streamfunction changes are positive in the DJF and MAM seasons, but negative in the JJA and SON seasons; streamfunction changes in the NH winter are generally larger than in the SH winter. The difference between the two ozone climatologies (i.e. the difference between experiments 1A and 1B) has little effect on streamfunction values in the troposphere and stratosphere; changes in the mesosphere tend to be smaller than for the difference seen when the background CO_2 concentration is doubled. For the change from the present–day to the future SST climatology, streamfunction values increase (decrease) during the NH (SH) winter. The magnitude of the changes is larger in the upper troposphere than in the middle atmosphere.

While this initial, qualitative analysis hints at the influence of CO_2, O_3 and forcing from the ocean surface on tropical upwelling and the overturning circulation in the middle atmosphere, a more quantitative approach (which follows) evaluates the relative importance of changes in these three parameters, as well as their variation with altitude and season, on the strength of the BDC under climate change.

5.2 Quantifying Seasonal Differences in the Mean Streamfunction

The impact of changes in ozone, CO_2 and SSTs on the seasonal mean strength of the meridional overturning circulation is assessed quantitatively by grouping together pairs of experiments differing by the same boundary conditions. In Fig. 6, the bars shown in blue represent pairs of experiments which differ only by their O_3 climatology (1980–like – 2000–like); the pink bars represent pairs of experiments which differ only by their background CO_2 concentration (704ppmv – 352ppmv); the green bars represent pairs of experiments which differ only by their SST climatology (future – present–day); the yellow bars represent the climate change signal (2C–1B). Positive values (increased streamfunction) indicate increased transport toward the North Pole, whereas negative values (decreased streamfunction) indicate increased transport toward the South Pole. Values not significantly different from zero indicate that changing a particular parameter has not affected the meridional circulation. Error bars shown in region 5 (in Fig. 6) denote ±1 standard deviation; often, the uncertainties are comparable to the magnitudes of the differences themselves.

Fig. 6 shows pressure–weighed streamfunction differences between pairs of experiments, in the seven atmospheric regions defined in section 5.1, for the DJF season. Streamfunction values are larger in the idealized future climate scenario (experiment 2C) than in the idealized present–day climate (1B). Furthermore, this figure shows that the relative contribution of the three types of parameter changes is altitude dependent. The SST change dominates the climate change signal in the upper troposphere (region 1) while the CO_2 change dominates in the NH high–latitude upper stratosphere (region 4), and the CO_2 and O_3 changes dominate in the mesosphere (regions 5–7).

Streamfunction differences have a seasonal cycle. Streamfunction differences are generally positive in the DJF season (i.e., increased transport toward the North Pole; see Fig. 6) and negative in the June–July–August (JJA) season (increased transport toward the South Pole; not shown). That is, the strength of the meridional circulation increases in both the NH and SH winter seasons. Generally, differences are positive in the March–April–May (MAM) season and negative in the September–October–November (SON) season, though the magnitudes of these differences are smaller than in the two winter seasons. In regions 2 and 4 (located in the high–latitude upper stratosphere), pressure–weighted streamfunction magnitudes in each of the simulations are small (see Fig. 5b) and the seasonal cycles are much weaker than in other parts of the atmosphere.

5.3 Relationship Between Tropical Upwelling and Polar Ozone

Time–slice simulations with increased CO_2 predict that the polar vortex will strengthen and, assuming the continued presence of anthropogenic chlorine, greater wintertime ozone loss should occur by the mid– to late 21st century (note the differences between 2C and 1B in Fig. 4). The same simulations predict that the strength of the BDC will increase as greenhouse gas concentrations continue to rise and the ozone layer recovers (see Fig. 6). Combining these two predictions, increased tropical upwelling in early or mid–winter should correlate with a decrease in total column ozone at NH high latitudes in late winter.

For individual time–slice experiments, the correlation between tropical streamfunction and high–latitude total ozone is low, due to the high degree of interannual variability within the 20–year analysis period; García–Herrera et al. (2006) note that relating changes in tropical upwelling and circulation changes at high latitudes is made difficult because of various sources of climate variability, such as the quasi–biennial oscillation (QBO). A more robust relationship between tropical upwelling and polar ozone emerges when the means of each experiment are examined (Fig. 7). As expected, least squares fitting of the eight means yields a negative slope; that is, relative to present–day (experiment 1B), there will be stronger tropical upwelling but lower total ozone near the north pole in March in a future climate (2C)[7]. Though differences between experiments 1B and 2C are statistically significant, the linear regression of January pressure–weighted streamfunction in region 3 as a function of March total ozone at 80°N (Fig. 7) is not.

Correlations of the annual cycles of the pressure–weighted streamfunction between two of the seven atmospheric regions are much higher (generally exceeding 0.90; see Hurwitz, 2008) than for the tropical streamfunction–polar ozone link. That is, increased tropical upwelling corresponds with increased meridional transport in the mesosphere. (Correlations with region 2, which is outside the region of meridional overturning circulation for much of the year, are not statistically significant.) The magnitude of these correlations is generally consistent from experiment to experiment. Thus, although the strength of the BDC is likely to be affected by changes in greenhouse gas concentrations and

[7] Note that this result is a dynamical signature and does not take into account potential changes in ozone chemistry.

other climate forcings, the structure of the circulation pattern itself will remain unchanged.

6 Discussion

This study assessed the roles of three contributors to future stratospheric climate change: increasing CO_2, ozone recovery and generally warmer sea surface temperatures. Stratospheric temperatures, dynamics, ozone and the strength of the Brewer-Dobson circulation were examined in various idealized climate scenarios, using a chemistry–climate model with parameterized ozone chemistry. The 'climate change' signal (2C–1B), the difference from an idealized present–day climate and one predicted for the mid– to late 21st century, corresponded with an increase in polar vortex strength, increased poleward heat fluxes, decreases in stratospheric temperature and a strengthening of the BDC.

In experiments where future SSTs were prescribed (1C-2D; see Table 1), the global mean temperature responses to decreased ozone and increased CO_2 concentrations matched those seen in experiments using present–day SSTs (see Fig. 1). In the stratosphere, the SST–related global mean temperature response was weaker than was the response to doubling CO_2 or to ozone depletion. Nevertheless, the switch from present–day to future SSTs enhanced tropospheric warming and slightly increased global mean temperatures in the lower stratosphere, elevating the tropopause.

A time–lagged linear relation between heat flux and temperature held for all eight time–slice experiments in both winter seasons (see Fig. 2). The NH and SH heat flux–temperature relationships had different slopes, though both showed a positive association between poleward heat fluxes in the upper troposphere and increased polar temperatures in the lower stratosphere. SH dynamics were less sensitive to changes in CO_2, O_3 and SSTs than were NH dynamics.

As noted by BHP2006, the response to ozone depletion was CO_2–dependent: the NH stratospheric vortex weakened under present–day CO_2 conditions (1B–1A) but strengthened in a doubled–CO_2 atmosphere (2B–2A). This 'flip flop' response was not seen in the four experiments where future SSTs were prescribed. Under future SSTs, prescribed ozone depletion had no significant effect on temperatures, heat fluxes, ozone concentrations or zonal winds at NH high latitudes in winter. This may have resulted from the reduction in baroclinicity in the atmosphere, when SSTs and greenhouse gas concentrations were increased simultaneously.

A strong anti–correlation between 10hPa zonal wind at 60°N and total ozone at 80°N is common to experiments using present–day SSTs (BHP2006) and to the future SST experiments. The lines of best fit were nearly identical. That the relationship between polar vortex strength and polar ozone remained robust, despite large changes in the temperature structure and dynamics of the middle atmosphere, points to the fundamental interdependence of chemistry and climate in the NH polar stratosphere.

Changes in tropical upwelling and meridional overturning in the middle atmosphere were quantified by examining regional streamfunction variations. In the model experiments, increased tropical upwelling, reduced mid–winter polar ozone and increased polar vortex strength occurred in a climate forced by warmer SSTs and higher greenhouse gas concentrations. The idealized climate change signal (2C–1B) showed a strengthened streamfunction, particularly for the DJF and JJA seasons. This result, therefore, is consistent with previous modeling studies that suggest that the BDC will strengthen in a future climate. The relative contribution of SSTs, O_3 and CO_2 changes to the enhanced circulation in the middle atmosphere was altitude dependent: SST changes played an important role in the tropical upper troposphere (consistent with Garny et al. (2009)), while changes in CO_2 and O_3 dominated the circulation response in the middle atmosphere (see Fig. 6).

This work predicted no substantial change in the relationship between tropospheric forcing, polar temperature and BDC in a future climate. However, it was not possible to relate climate change–induced increases in tropical upwelling in mid–winter to greater springtime ozone losses at NH high latitudes at statistically significant levels.

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Figure Captions

Table 1: Experimental design.

Figure 1: Summary of global and annual mean temperature response for the set of eight time–slice experiments. The left–hand panel shows the ozone–related response (2000–like – 1980–like); the turquoise line shows 1B–1A differences, the yellow line shows 2B–2A differences, the pink line shows 1D–1C differences and the light grey line shows 2D–2C differences. The central panel shows the CO_2–related response ($2xCO_2$ – $1xCO_2$); the orange line shows 2A–1A differences, the yellow line shows 2B–1B differences, the dark grey line shows 2C–1C differences and the light grey line shows 2D–1D differences. The right–hand panel shows the SST–related response (future – present–day); the burgundy line shows 1C–1A differences, the pink line shows 1D–1B differences, the dark grey line shows 2C–2A differences and the light grey line shows 2D–2B differences.

Figure 2: Zonal mean meridional mid-latitude heat flux at 100hPa versus polar temperature at 50hPa. For the NH winter, heat fluxes are northward (positive); the scatter plot shows December–January heat fluxes versus January–February temperatures for each year of each experiment. For the SH winter, heat fluxes are southward (negative); the scatter plot shows July–August heat fluxes versus August–September temperatures. Experiment 1A is shown in blue, 1B in turquoise, 2A in orange, 2B in yellow, 1C in burgundy, 1D in pink, 2C in dark grey and 2D in light grey; refer to table 1.

Figure 3: DJF 100hPa heat flux differences between pairs of experiments. The blue bars show heat flux differences due to ozone recovery (1980–like – 2000–like); the pink bars show heat flux differences due to doubling CO_2; the green

bars show heat flux differences due to SSTs (future – present–day); the yellow bar shows the climate change signal (2C–1B).

Figure 4: a) Scatter plot of January zonal mean zonal wind versus total ozone for experiments 1A, 1C, 1D, 2C and 2D. b) Scatter plot of the 20–year mean January zonal wind versus total ozone for all eight time–slice experiments. The dotted line shows the mean regression line relating zonal wind to total ozone.

Figure 5: (a) Latitude–height cross–section showing the mean streamfunction for experiment 1B, for the DJF season [1×10^{-9} kg/s^3]. (b) Latitude–height cross–section of the mean pressure–weighted streamfunction (the streamfunction divided by the pressure in hPa) for experiment 1B for the DJF season [1×10^{-9} m/s]. The seven numbered boxes identify the atmospheric regions defined in section 5.1 of the text.

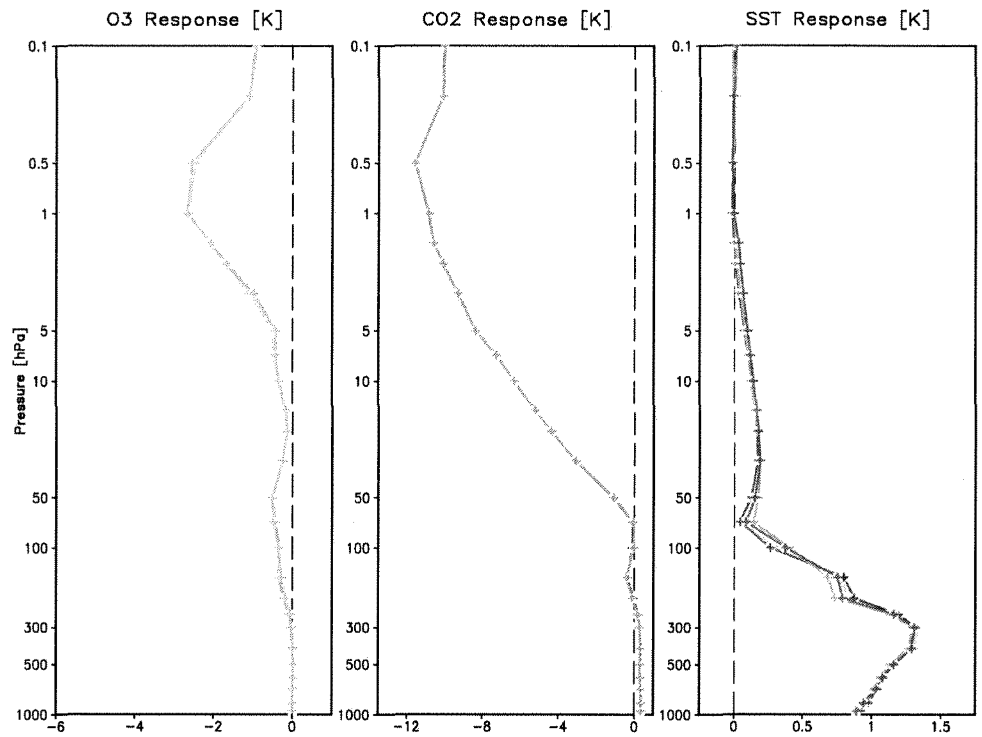
Figure 6: Pressure–weighted streamfunction differences between pairs of experiments, for seven atmospheric regions, for the DJF season. The spatial organization of the regions is as shown in figure 6b. The four bars for each set of differences denote the DJF, MAM, JJA and SON seasons, respectively. The blue bars represent the response to ozone recovery; the pink bars represent the response to doubled CO_2; the green bars represent the response to increased SSTs; the yellow bars represent the response in the climate change signal (2C–1B). Error bars in region 5 denote ± 1 standard deviation.

Figure 7: January streamfunction in region 3 versus March total column ozone at 80°N. The colored circles show the 20–year mean for each time–slice experiment; the dashed line shows the line of best fit, fitting the eight mean values.

Figures

Experiment	O_3 Climatology	Background $[CO_2]$ (ppmv)	SST Climatology
1A	1980	352 ($1xCO_2$)	Present–day
1B	2000	352	Present–day
2A	1980	704 ($2xCO_2$)	Present–day
2B	2000	704	Present–day
1C	1980	352	Future
1D	2000	352	Future
2C	1980	704	Future
2D	2000	704	Future

Table 1: Experimental design.



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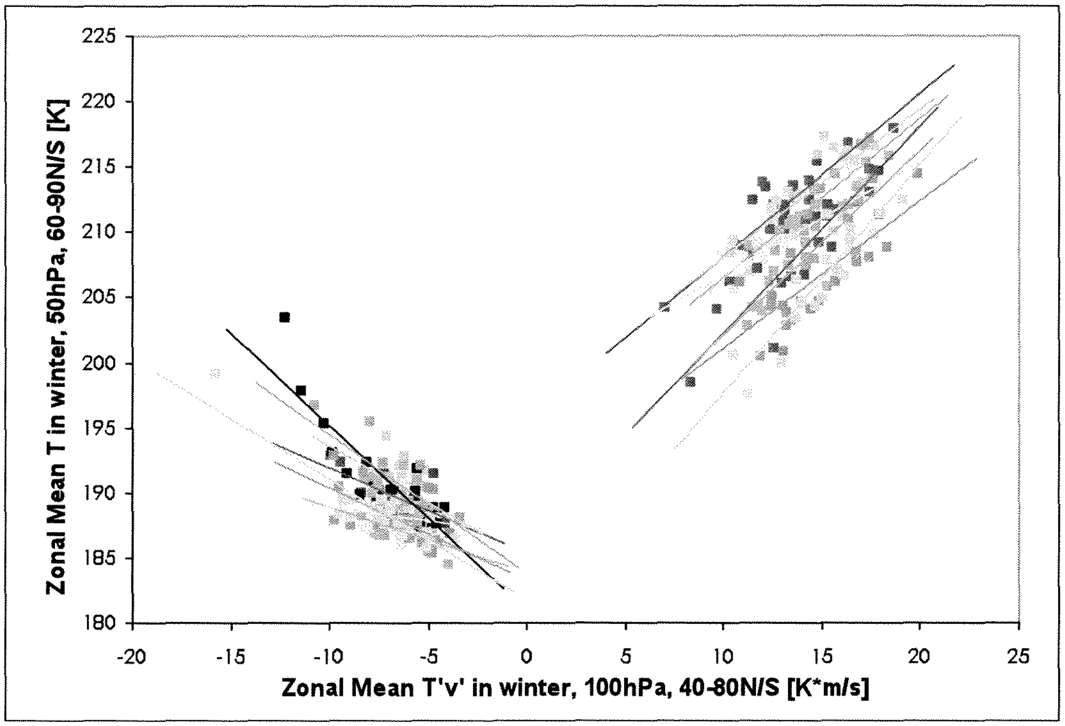

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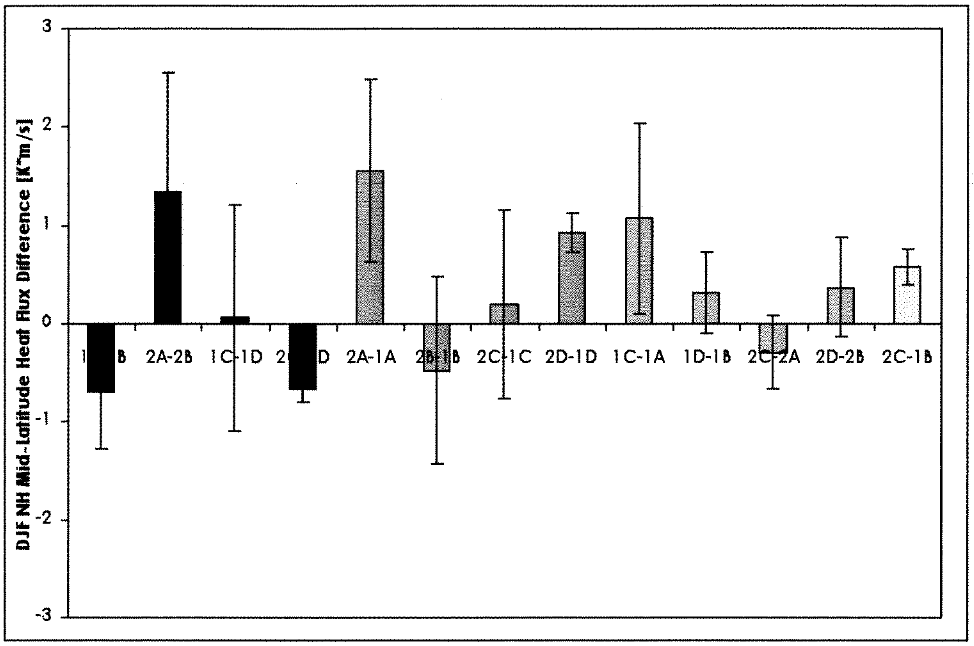

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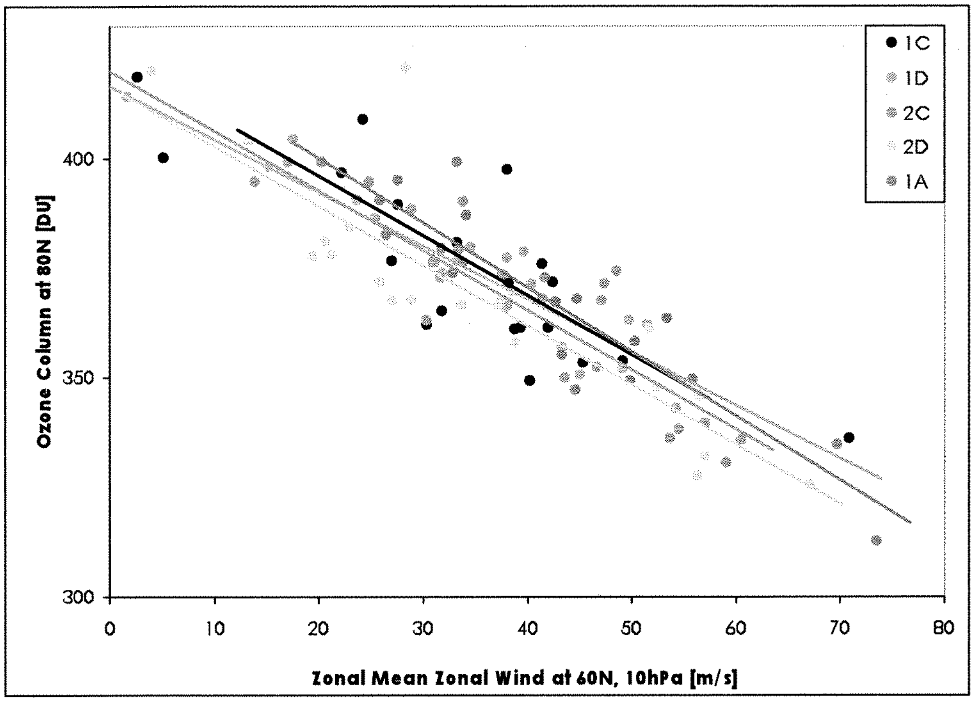

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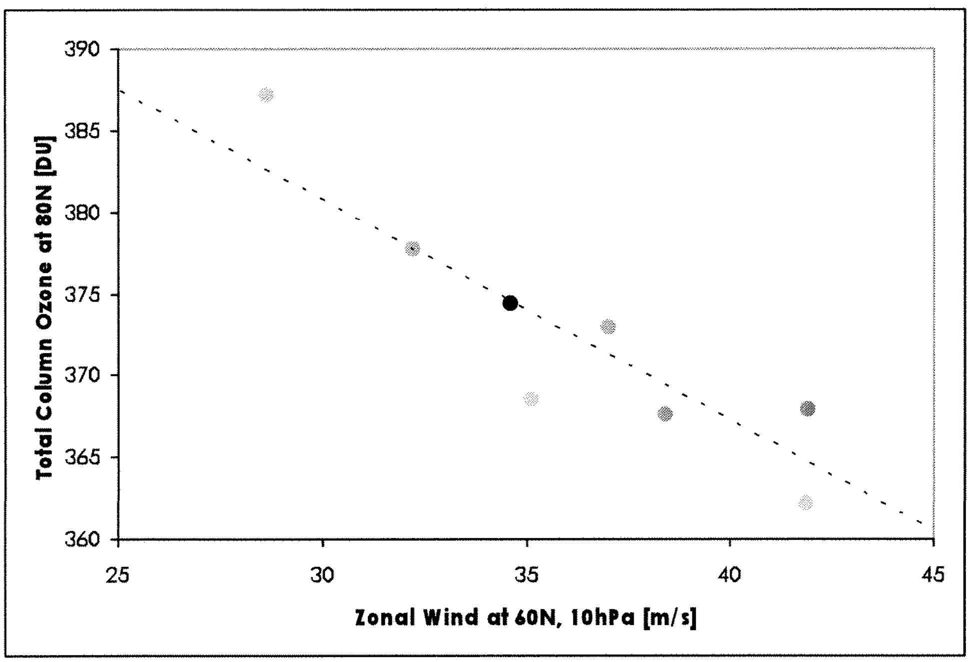

b)

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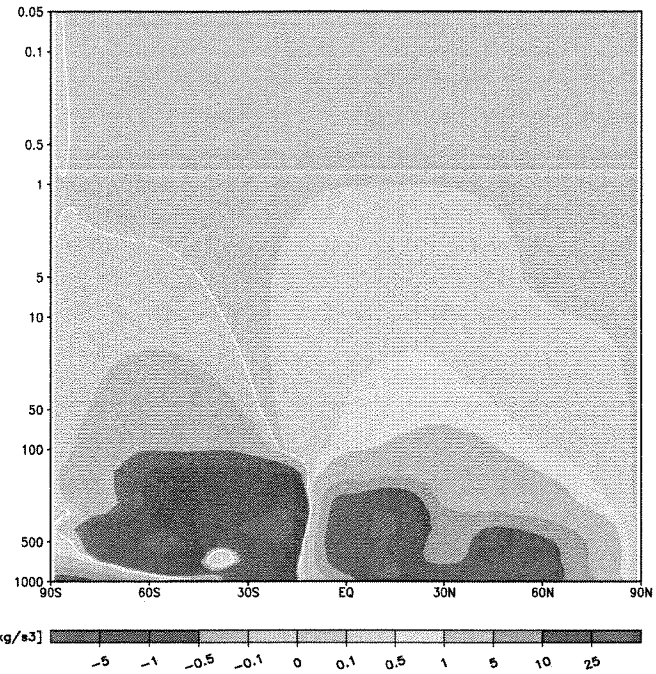

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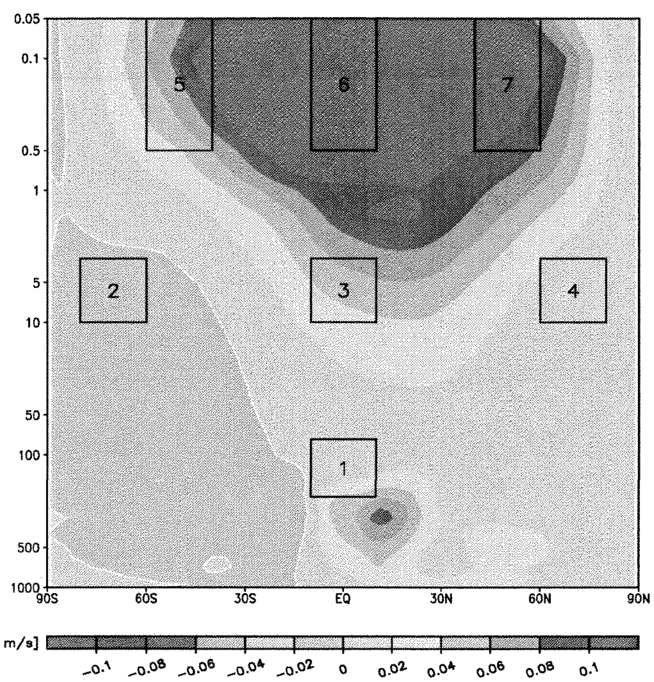

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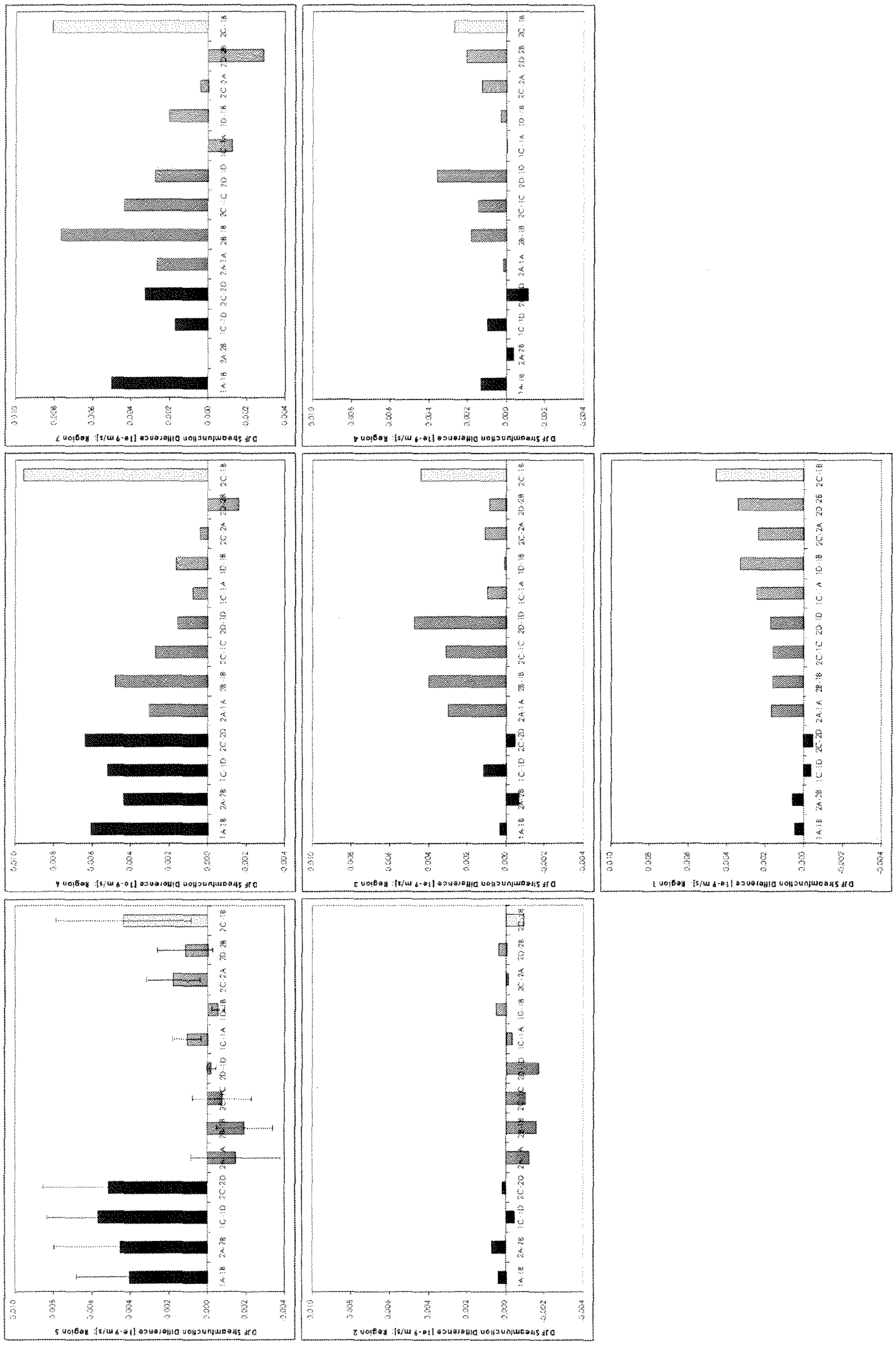

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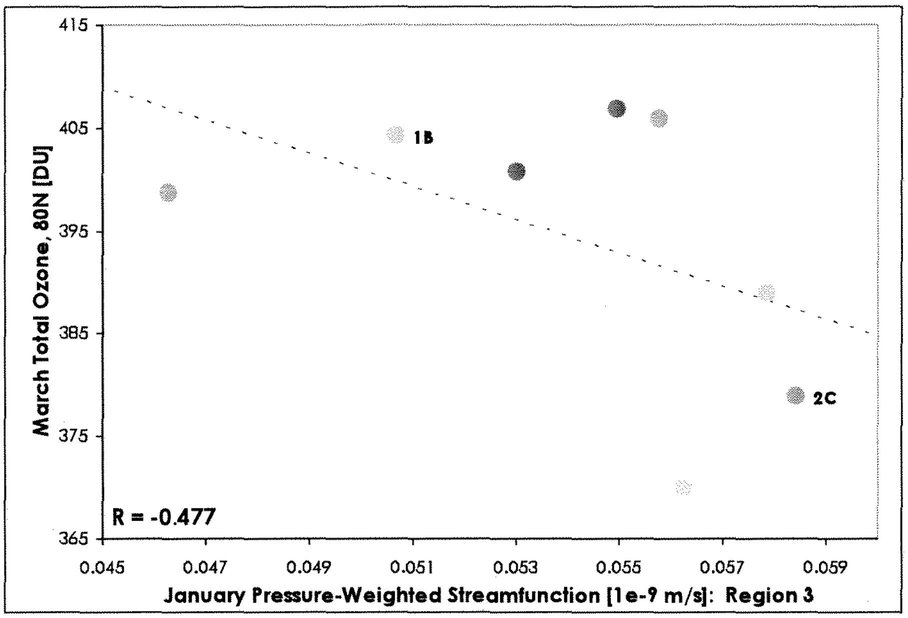

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