Impact of Antarctic ozone depletion and recovery on Southern Hemisphere precipitation, evaporation and extreme changes

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ABSTRACT

The possible impact of Antarctic ozone depletion and recovery on Southern Hemisphere (SH) mean and extreme precipitation and evaporation is examined using multimodel output from the Climate Model Intercomparison Project 3 (CMIP3). By grouping models into four sets, those with and without ozone depletion in 20th century climate simulations and those with and without ozone recovery in 21st century climate simulations, and comparing their multimodel-mean trends, it is shown that Antarctic ozone forcings significantly modulate extratropical precipitation changes in austral summer. The impact on evaporation trends is however, minimal especially in 20th century climate simulations. In general, ozone depletion has increased precipitation in high-latitudes and decreased it in mid-latitudes, in agreement with the poleward displacement of the westerly jet and associated storm tracks by Antarctic ozone depletion. Although weaker, the opposite is also true for ozone recovery. These precipitation changes are primarily associated with changes in light precipitation (1–10 mm day$^{-1}$). Contributions by very-light precipitation (0.1–1 mm day$^{-1}$) and moderate-to-heavy precipitation (>10 mm day$^{-1}$) are minor. Likewise, no systematic changes are found in extreme precipitation events, although extreme surface wind events are highly sensitive to ozone forcings. This result indicates that, while extratropical mean precipitation trends are significantly modulated by ozone-induced large-scale circulation changes, extreme precipitation changes are likely more sensitive to thermodynamic processes near the surface than to dynamical processes in the free atmosphere.
1. Introduction

Southern Hemisphere (SH) climate changes over the last few decades have been extensively documented in recent studies. They include an expansion of the Hadley cell (Seidel et al. 2008; Johanson and Fu 2009), a shift in atmospheric mass from high- to mid-latitudes (Thompson and Solomon 2002; Marshall 2003), a poleward displacement of the westerly jet and storm tracks (Thompson and Solomon 2002; Marshall 2003; Fyfe 2003), an increase in surface wind speeds over the Southern Ocean (Böning et al. 2008), anomalously dry conditions over southern South America, New Zealand and southern Australia, and anomalously wet conditions over much of Australia and South Africa (Gillett et al. 2006). A freshening and warming of the Southern Ocean (Wong et al. 1999; Gille 2002; Böning et al. 2008) and a significant warming of the Antarctic peninsula (Thompson and Solomon 2002) have also been observed. While some of these changes have been attributed to circulation changes induced by the increase in anthropogenic greenhouse gases (Fyfe et al. 1999; Kushner et al. 2001; Cai et al. 2003), such changes in austral summer have also been influenced by Antarctic ozone depletion (Thompson and Solomon 2002; Shindell and Schmidt 2004; Arblaster and Meehl 2006; Perlwitz et al. 2008; Son et al. 2009, 2010; McLandress et al. 2011; Polvani et al. 2011; Kang et al. 2011). It is known that both increasing greenhouse gases, occurring year-round, and ozone depletion, which occurs most significantly in late spring and summer, have driven SH extratropical circulation changes in a similar way, the cumulative effect resulting in more significant tropospheric climate change in austral summer than in other seasons. In the future, the effects of these two forcings are however predicted to oppose each other (Shindell and Schmidt 2004; Perlwitz et al. 2008; McLandress et al. 2011), as Antarctic ozone concentrations are anticipated to increase due to the implementation of the Montréal Protocol (Austin et al. 2010).

While the surface climate impact of increasing greenhouse gases is relatively well understood, our understanding of stratospheric ozone-related climate change at the surface, especially its mechanisms, is somewhat limited. In particular the impact of stratospheric
ozone changes on the hydrological cycle in the SH is not well understood. A series of recent studies have shown that stratospheric ozone depletion has likely enhanced austral-summer precipitation changes in the subtropics and high-latitudes but reduced them in mid-latitudes, consistent with the poleward displacement of the westerly jet, or equivalently the positive trend in the Southern Annular Mode (SAM) index (Son et al. 2009; McLandress et al. 2011; Polvani et al. 2011; Kang et al. 2011). Although they are crucial for understanding salinity changes in the Southern Ocean, net hydrological changes, including evaporation, are not yet well understood. In addition and arguably more importantly, potential changes in extreme precipitation events are yet to be investigated. It is known that individual precipitation events are likely to get more intense as the climate warms (Emori and Brown 2005; Sun et al. 2007; O’Gorman and Schneider 2009). Previous studies suggest that it is predominantly thermodynamics that control changes in extratropical extreme precipitation (Emori and Brown 2005; O’Gorman and Schneider 2009). Thus, it is questionable whether extreme precipitation events will respond to dynamical changes driven by the Antarctic ozone hole.

The purpose of this study is to bridge the existing gap in understanding the relative contributions of anthropogenic greenhouse gas emissions and stratospheric ozone changes in forcing changes in the hydrological cycle. Multimodel output from the Climate Model Intercomparison Project 3 (CMIP3; Meehl et al. (2007)) are analysed. By grouping models into those with prescribed ozone depletion and recovery, and those without it, we show that Antarctic ozone forcings significantly affect seasonal-mean precipitation trends in the extratropics during austral summer, but play a minimal role in evaporation and extreme precipitation trends.

2. Data and methods

CMIP3 data from the 20th century climate simulations (20C3m) and 21st century climate simulations with the special report on emissions scenarios A1B forcing (A1B) are analysed.
From all available models, the models which archived daily precipitation are first selected. For those models, evaporation is calculated from surface latent heat flux as outlined in Yu et al. (2008). Each model’s precipitation and evaporation climatologies are then compared with Global Precipitation Climatology Project version-2 (GPCP) precipitation (Adler et al. 2003) and Objectively Analyzed Air-Sea Heat Fluxes version-3 (OAFlux) global ocean evaporation data (Yu et al. 2008). Those models with significant biases\(^1\) are discarded and 19 models are selected for the analyses as described in Table 1.

All CMIP3 models have prescribed stratospheric ozone concentrations with a seasonal cycle. However, not all models have incorporated stratospheric ozone depletion in the latter part of the 20\(^{th}\) century and recovery in the 21\(^{st}\) century, as anthropogenic ozone forcings were not mandated in the CMIP3 (Meehl et al. 2007). Ten models prescribed ozone depletion and ozone recovery, whilst nine models simply used climatological ozone fields\(^2\). As such, models are grouped into four sets: those with and without ozone depletion in the 20\(^{th}\) century, and those with and without ozone recovery in the 21\(^{st}\) century. For each group, the multimodel-mean climatologies and trends are calculated for the fields of interest (precipitation and evaporation) over the 20\(^{th}\) and 21\(^{st}\) centuries. As in Son et al. (2009), climatologies and trends are first calculated for each ensemble member and averaged over all available ensemble members of a given model. The ensemble average of each model is interpolated onto a 4° latitude by 4° longitude grid and averaged over all available models within a group. Hatching is used on trend maps to denote where the multimodel mean trend is greater than or equal to one standard deviation of the trends of different models within that group. By comparing the multimodel means of each group, the impact of Antarctic ozone forcings on hydrological climate changes is systematically examined. Although this approach does not necessarily

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\(^1\)FGOALS1.0g (IAP, China) is discarded as 20\(^{th}\) century precipitation in the high-latitude region is found to be unreasonably higher than observations and all other models. GISS-ER (NASA, USA) is discarded as 1971–1999 daily precipitation appears to be erroneous across the extent of the SH.  

\(^2\)Certain CMIP3 models prescribed ozone depletion in the 20\(^{th}\) century but did not prescribe ozone recovery in the 21\(^{st}\) century, however for other reasons, such models were not included in this study.
reveal ozone-related surface climate changes, as each group comprises different models, it is known that trend differences resulting from different ozone forcings are likely larger than those associated with model-dependent internal variabilities (Son et al. 2009).

Since daily data are archived only for selected decades in the A1B runs, long-term trends are estimated in this study using decadal differences. The 20th century change, reflecting the impact of ozone depletion, is defined by the difference between 1990–1999 and 1961–1970 means. Likewise the 21st century change, reflecting the impact of ozone recovery, is defined by the difference between 2056–2065 and 1990–1999 means. Since decadal differences are qualitatively similar to the linear trends computed from monthly-mean data over 1960–1999 and 2000–2079 (not shown), they are simply referred to as “trends” in this study. The possible changes in extreme precipitation events are examined by decomposing seasonal-mean precipitation trends into three regimes (Sun et al. 2007): very-light (0.1–1 mm day$^{-1}$), light (1–10 mm day$^{-1}$) and moderate-to-heavy (>10 mm day$^{-1}$) precipitation changes. Five extreme precipitation indices are also examined. They are the sum of precipitation on all wet days divided by the number of wet days, the sum of rainfall on days exceeding the 95th percentile threshold as determined for the base period of 1961–1990 (hereafter 95th percentile precipitation), the sum of rainfall on days exceeding the 99th percentile threshold as determined for the base period of 1961–1990, seasonal maximum one-day precipitation, and seasonal maximum five-day consecutive precipitation (ETCCDI/CRD 2009).

3. Results

Multimodel-mean trends of austral-summer (DJF) precipitation and evaporation are presented in Fig. 1. Only the extratropics, poleward of 30°S, are shown, as tropical and subtropical trends are noisy and largely insignificant. In the 20th century the poleward displacement of storm tracks by increasing greenhouse gases causes a dipolar trend in precipitation (Yin 2005). This is evident in the models without ozone depletion, however trends are only weak.
A similar pattern but with much stronger magnitude is found in the models with prescribed ozone depletion, indicating the combined effects of increasing greenhouse gases and ozone depletion on extratropical precipitation changes. The opposite is generally true in the 21st century: models with prescribed ozone recovery show relatively weaker precipitation trends than models with fixed ozone forcing. As shown in previous studies (Son et al. 2009; McLandress et al. 2011; Polvani et al. 2011), this sensitivity is observed only in austral summer.

In contrast to the annular-like trends of precipitation, evaporation shows relatively weak trends in the extratropics, which lack organisation. The sensitivity of evaporation trends to ozone forcings is also weak, although there is a hint that models with ozone depletion have a weaker decreasing trend in high-latitude evaporation than those without ozone depletion presumably because of the acceleration of surface westerlies by ozone depletion. This result, combined with the findings related to precipitation trends, indicates that Antarctic ozone forcings modulate long-term trends of surface freshwater flux (or equivalently, precipitation minus evaporation) by primarily affecting precipitation trends. Figure 1 (third row) in particular suggests that Antarctic ozone depletion has likely contributed to the observed freshening of the Southern Ocean (e.g. Böning et al. 2008), but its recovery would have little impact on Southern Ocean freshening, as ozone-related precipitation changes in the 21st century are partly cancelled by evaporation changes.

Figure 2 presents the relative contributions of very-light, light and moderate-to-heavy precipitation changes to the DJF-mean precipitation changes shown in Fig. 1. It is evident that mean precipitation trends are dominated by light precipitation trends in the models (second row), the precipitation regime which is somewhat overestimated in the CMIP3 models (Dai 2006). Note that, although very-light precipitation events also show significant trends, their cumulative impact is very weak (shading interval in the first row is one tenth of that in the second row). More importantly, their sensitivity to ozone forcings is exactly opposite to those of seasonal-mean precipitation trends (compare the first rows of Fig. 1 and Fig. 2). This is somewhat surprising, but similar results, opposite trends between very-light
and light precipitation events, are also found in the precipitation response to anthropogenic warming (e.g. Sun et al. 2007). For instance, as shown in the right-most column of Fig. 2, very-light and light precipitation trends in the 21st century with fixed ozone but with increasing greenhouse gases show opposite trends in the high-latitudes. It appears the similar excitation of the SAM by both increasing greenhouse gases and varying ozone affects the precipitation distribution. The mechanisms for this are however unknown: the response may be accounted for in terms of thermodynamic changes by increasing greenhouse gases (i.e. a warm atmosphere can hold more moisture, likely causing the probability distribution function of precipitation events to shift away from very-light to light precipitation events), but is more perplexing in terms of ozone-forced dynamic changes. Further studies are needed.

The trends in moderate-to-heavy precipitation events (Fig. 2 third row), which are quantitatively similar to 95th percentile precipitation trends (fourth row), are relatively weak. Unlike trends in very-light and light precipitation, they show no dipole pattern: trends are predominantly positive across the extent of the SH, particularly in the 21st century, reflecting more intense and frequent extreme precipitation events with anthropogenic warming. More importantly, no notable sensitivity to ozone forcings is observed. Although not shown, a lack of sensitivity is also found in the other extreme precipitation indices described in the data and methods section. This suggests that Antarctic ozone forcings only play a minimal, if any, role in extreme precipitation changes. Here it should be noted that extreme precipitation events are sensitive to model resolution. Comparisons of MIROC3.2 (hires, T106) with MIROC3.2 (medres, T42), and of CGCM3.1 (T63) with CGCM3.1 (T47), in fact show stronger trends in the higher resolution models (not shown). However, the fact that each model group contains models with a variety of resolutions and that multimodel-mean trends in extreme precipitation are quantitatively similar between models with time-varying ozone and models with fixed ozone, although the former have generally higher resolution than the latter (Table 1), suggests that the conclusion that Antarctic ozone forcings play a minimal role in extreme precipitation changes holds.
The above results suggest that Antarctic ozone forcings affect hydrological climate changes in the SH extratropics by modifying dynamics that in turn modify light precipitation trends. It may be questionable whether results are affected by model groups having different climate sensitivities: in fact models with varying ozone are known to have higher climate sensitivity to increasing greenhouse gases (Miller et al. 2006). However, that in the 21st century models with varying ozone exhibit trends in mean precipitation consistent with ozone recovery, suggests that models with varying ozone are not simply responding more strongly to increasing greenhouse gases, as the response to ozone recovery is in the opposite direction. Nevertheless, results should be treated with caution, as they are based solely on simple multimodel averaging. One limitation of multimodel averaging, amongst others, is due to the fact that individual models have different climatologies (eg. Barnes and Hartmann 2010). For instance, the DJF-mean climatological jet, defined by the maximum westerly wind at 925 hPa, varies from 43°S to 56°S amongst models (not shown). Models with fixed ozone forcing generally have a jet in lower latitudes (46.2°S ± 2.9°) compared to those with ozone depletion (49.8°S ± 2.7°) (as expected, the latter is closer to the real atmosphere (50°S)). Since storm tracks are located on the poleward side of the westerly jet, the simple multimodel averaging used in this study, which ignores individual models’ bias, could over- or underestimate precipitation trends by averaging stormy regions with dry ones.

Figure 3 shows seasonal-mean precipitation climatology and long-term trends for individual models and multimodel means, shifted relative to the climatological jet. Only zonally-averaged fields are shown as precipitation trends are largely homogeneous in the zonal direction (Figs. 1, 2). The CMIP3 models reproduce DJF-mean precipitation reasonably well in the high-latitudes (-30–0° relative to the jet). However, they generally underestimate it in mid-latitudes (0–20° relative to the jet) and overestimate it in the subtropics and tropics (20–50° relative to the jet). Intermodel variation is particularly large in the subtropics and tropics, making any influence of ozone forcings difficult to distinguish. In fact, although a recent study by Kang et al. (2011) has shown that ozone depletion may have enhanced sub-
tropical precipitation in the SH, no sensitivity of subtropical precipitation trends to ozone forcings is found in multimodel-mean trends (second row). A noticeable difference in subtropical precipitation trends is only found in the 21st century (second row, right column). This is however unlikely to be associated with ozone recovery: it is, instead, thought to be a result of the intensification of moderate-to-heavy (fifth row, right column) tropical precipitation, caused by anthropogenic warming, amongst models with different tropical climatologies.

Returning to extratropical precipitation trends, Fig. 3 confirms that ozone depletion tends to increase high-latitude precipitation trends but decrease mid-latitude precipitation trends by modulating light precipitation events. The role of ozone recovery is the reverse: decreasing (increasing) high- (mid-) latitude precipitation trends, relative to greenhouse gas induced trends alone. This sensitivity is statistically significant at the 99 % confidence level as summarised in Table 2. Note that, while the mid-latitude precipitation trend difference in the 20th century is not significant, the decrease in mid-latitude light precipitation by ozone depletion is statistically significant. Note also that although the sensitivity of high-latitude precipitation to ozone forcings is not significant in the 21st century in Table 2, it becomes significant if linear trends of monthly-mean precipitation, which are generally stronger than decadal differences, are used. The identical analyses are further performed for other seasons and no significant difference in multimodel-mean trends between the models with and without time-varying ozone forcings is found.

4. Discussion

The multimodel analyses, based on CMIP3 models, show that Antarctic ozone forcings significantly affect austral-summer precipitation trends in the SH extratropics. Their effects are primarily realised by changes in light precipitation events (1–10 mm day$^{-1}$), with negligible changes in extreme precipitation events attributable to ozone forcings. This lack of sensitivity of extreme precipitation events to ozone forcings is somewhat contradictory to the
atmospheric circulation changes by ozone forcings: ozone depletion is known to strengthen extratropical westerlies, or equivalently to favour the positive polarity of the SAM (Son et al. 2008, 2010; McLandress et al. 2011; Polvani et al. 2011). As shown in Fig. 4, trends of both seasonal-mean and frequency of occurrence of 95th percentile surface westerlies are strengthened by ozone depletion (second row). While mean tropospheric circulation changes are consistent with mean precipitation changes, no link is established between extreme events (compare first and second rows). In other words, strengthening of the 95th percentile wind by ozone depletion does not lead to strengthening of 95th percentile precipitation. This result suggests that thermodynamic effects are likely more important than dynamic effects in the extratropical extreme precipitation changes as discussed by Emori and Brown (2005) and O’Gorman and Schneider (2009). In fact, surface air temperature trends, which control water vapour content in the atmosphere, show a negligible sensitivity to ozone forcings (third row).

As previously stated, the findings of this study are based solely on multimodel averaging and thus should be treated with care. Although multimodel averaging can reduce model biases, it does not allow direct attribution of SH surface climate changes to Antarctic ozone forcings. This approach may also underestimate surface climate responses to ozone forcings by averaging models with realistic climatologies and trends with those without them. More quantitative studies using climate model sensitivity tests (e.g. McLandress et al. 2011; Polvani et al. 2011) are needed for better quantifying stratospheric ozone-related hydrological climate changes in the SH.

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1 Description of CMIP3 models used in this study. Details of each model are described in Randall et al. (2007). Resolutions refer to atmospheric resolution and horizontal resolution is approximate for spectral models, where “T” refers to triangular truncation. The number of ensemble members refers to those used in precipitation analyses. Brackets indicate where different ensemble members are used in evaporation analyses.

2 Summary of differences in percentage change between the models with time-varying ozone forcings and those with fixed ozone forcing. Here percentage change is defined as a decadal change normalised by long-term climatology, calculated for the high-latitude (averaged over 4–24° south of individual model’s climatological jet) and mid-latitude (averaged over 12–0° north of individual model’s climatological jet) regions. Only the values which are statistically significant at the 99 % confidence level are shown. Significance tests are based on a Monte Carlo approach. This approach selects one group of ten models (eight for evaporation) and one group of nine models at random and calculates the percentage change difference between the means of the two groups. This is repeated 50,000 times to get a statistical distribution. The actual difference between the mean of the varying ozone group and the fixed ozone group is then compared with this statistical distribution at the 99 % confidence level. Although not shown, overall results are qualitatively similar to a two-sided Student’s t test at the 99 % confidence level. Only results from DJF are shown, as no significant values are found in other seasons.
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<table>
<thead>
<tr>
<th>Model</th>
<th>Group, country</th>
<th>Horizontal res. (lat. × lon.)</th>
<th>Vertical res. levels, top</th>
<th>20C3m members</th>
<th>A1B members</th>
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* Model documentation claims inclusion of ozone chemistry, however analysis of Antarctic polar-cap temperature by Son et al. (2008) found no ozone impact in either 20C3m or A1B simulations.
Table 2. Summary of differences in percentage change between the models with time-varying ozone forcings and those with fixed ozone forcing. Here percentage change is defined as a decadal change normalised by long-term climatology, calculated for the high-latitude (averaged over 4–24° south of individual model’s climatological jet) and mid-latitude (averaged over 12–0° north of individual model’s climatological jet) regions. Only the values which are statistically significant at the 99 % confidence level are shown. Significance tests are based on a Monte Carlo approach. This approach selects one group of ten models (eight for evaporation) and one group of nine models at random and calculates the percentage change difference between the means of the two groups. This is repeated 50,000 times to get a statistical distribution. The actual difference between the mean of the varying ozone group and the fixed ozone group is then compared with this statistical distribution at the 99 % confidence level. Although not shown, overall results are qualitatively similar to a two-sided Student’s t test at the 99 % confidence level. Only results from DJF are shown, as no significant values are found in other seasons.

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<td>21st C</td>
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<td>Moderate-to-heavy precip.</td>
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List of Figures

1 Multimodel-mean trends of precipitation (first row), evaporation (second row) and precipitation minus evaporation (third row) in DJF. Trends are calculated by differencing 1961–1970 from 1990–1999 averages, and 1990–1999 from 2056–2065 averages, for 20th and 21st century trends, respectively. From left to right, multimodel-mean trends are shown for models with and without ozone depletion in the 20th century, and for models with and without ozone recovery in the 21st century respectively. Cool colours denote an increasing freshwater flux (increasing precipitation or decreasing evaporation) whilst warm colours denote a decreasing freshwater flux (decreasing precipitation or increasing evaporation). Hatched areas denote where the multimodel-mean trend is greater than or equal to one standard deviation. Contours show the climatology, with contour intervals of 100 mm season$^{-1}$ for all panels.

2 Multimodel-mean trends in very-light (0.1–1 mm day$^{-1}$, first row), light (1–10 mm day$^{-1}$, second row), moderate-to-heavy (>10 mm day$^{-1}$, third row), and 95th percentile (fourth row) precipitation in DJF. Other details are the same as Fig. 1. Note that the colour scale is an order of magnitude less for very-light precipitation so that details can be seen. Contour intervals of climatologies are 20 mm season$^{-1}$ for very-light precipitation panels and 50 mm season$^{-1}$ for all other panels.

3 Zonal-mean precipitation climatology (first row) and total (second row), very-light (third row), light (fourth row), and moderate-to-heavy (fifth row) precipitation trends in DJF, plotted as a function of jet-relative latitudes in the SH. Zonal-mean values are shown for individual models (varying ozone models in blue and fixed ozone models in red), multimodel averages (bold blue and red lines) and GPCP precipitation (bold black line). Both 20th century (left column) and 21st century (right column) simulations are shown.
Zonal-mean precipitation (first row), surface zonal wind (second row) and surface air temperature (third row) trends in DJF, plotted as a function of jet-relative latitudes in the SH. Mean trends (left column) and frequency of occurrence of 95\textsuperscript{th} percentile events (right column) are shown. Only 20\textsuperscript{th} century simulations are presented. Note that the frequency trends on the right are different from 95\textsuperscript{th} percentile precipitation trends, as the latter is a cumulative quantity. Due to surface wind data availability, CCSM3.0, PCM1.1 and INM-CM3.0 are not included in these panels. The same colour convention as Fig. 3 is used.
**Fig. 1.** Multimodel-mean trends of precipitation (first row), evaporation (second row) and precipitation minus evaporation (third row) in DJF. Trends are calculated by differencing 1961–1970 from 1990–1999 averages, and 1990–1999 from 2056–2065 averages, for 20th and 21st century trends, respectively. From left to right, multimodel-mean trends are shown for models with and without ozone depletion in the 20th century, and for models with and without ozone recovery in the 21st century respectively. Cool colours denote an increasing freshwater flux (increasing precipitation or decreasing evaporation) whilst warm colours denote a decreasing freshwater flux (decreasing precipitation or increasing evaporation). Hatched areas denote where the multimodel-mean trend is greater than or equal to one standard deviation. Contours show the climatology, with contour intervals of 100 mm season$^{-1}$ for all panels.
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Fig. 3. Zonal-mean precipitation climatology (first row) and total (second row), very-light (third row), light (fourth row), and moderate-to-heavy (fifth row) precipitation trends in DJF, plotted as a function of jet-relative latitudes in the SH. Zonal-mean values are shown for individual models (varying ozone models in blue and fixed ozone models in red), multimodel averages (bold blue and red lines) and GPCP precipitation (bold black line). Both 20th century (left column) and 21st century (right column) simulations are shown.
Fig. 4. Zonal-mean precipitation (first row), surface zonal wind (second row) and surface air temperature (third row) trends in DJF, plotted as a function of jet-relative latitudes in the SH. Mean trends (left column) and frequency of occurrence of 95th percentile events (right column) are shown. Only 20th century simulations are presented. Note that the frequency trends on the right are different from 95th percentile precipitation trends, as the latter is a cumulative quantity. Due to surface wind data availability, CCSM3.0, PCM1.1 and INM-CM3.0 are not included in these panels. The same colour convention as Fig. 3 is used.