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Chlorine isotope composition in chlorofluorocarbons CFC-11, CFC-12 and CFC-113 in firn, stratospheric and tropospheric air

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Abstract

The stratospheric degradation of chlorofluorocarbons (CFCs) releases chlorine, which is a major contributor to the destruction of stratospheric ozone (O₃). A recent study reported strong chlorine isotope fractionation during the breakdown of the most abundant CFC (CFC-12, CCl₂F₂), similar to effects seen in nitrous oxide (N₂O). Using air archives to obtain a long-term record of chlorine isotope ratios in CFCs could help to identify and quantify their sources and sinks. We analyse the three most abundant CFCs and show that CFC-11 (CCl₃F) and CFC-113 (CClF₂CCl₂F) exhibit significant stratospheric chlorine isotope fractionation, in common with CFC-12. The apparent isotope fractionation (ϵ_{app}) for mid- and high-latitude stratospheric samples are (-2.4 ± 0.5) and (-2.3 ± 0.4) ‰ for CFC-11, (-12.2 ± 1.6) and (-6.8 ± 0.8) ‰ for CFC-12 and (-3.5 ± 1.5) and (-3.3 ± 1.2) ‰ for CFC-113, respectively. Assuming a constant source isotope composition, we estimate the expected trends in the tropospheric isotope signature of these gases due to their stratospheric ³⁷Cl enrichment and stratosphere–troposphere exchange. We compare these model results to the long-term $\delta(^{37}\text{Cl})$ trends of all three CFCs, measured on background tropospheric samples from the Cape Grim air archive (Tasmania, 1978–2010) and tropospheric firn air samples from Greenland (NEEM site) and Antarctica (Fletcher Promontory site). Model trends agree with tropospheric measurements within analytical uncertainties. From 1970 to the present-day, we find no evidence for variations in chlorine isotope ratios associated with changes in CFC manufacturing processes. Our study increases the suite of trace gases amenable to direct isotope ratio measurements in small air volumes, using a single-detector gas chromatography-mass spectrometry system.

1 Introduction

CFC-11 (CCl₃F), CFC-12 (CCl₂F₂) and CFC-113 (CClF₂CCl₂F) are the most abundant chlorofluorocarbons (CFCs) in the atmosphere (Table 1). They are very stable and are

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not flammable or toxic, making them suitable as refrigerants, aerosol propellants and foam blowing agents. Measurements performed on air from porous firn (the top part of the polar ice sheet consisting of consolidated snow, progressively transforming into ice), show that they were absent from the atmosphere before the middle of the 20th century (e.g. Butler et al., 1999; Sturrock et al., 2002; Buizert et al., 2012). This suggests that they have no natural sources. CFC-11, CFC-12 and CFC-113 first appeared in the mid-1900s (Sturrock et al., 2002; Martinerie et al., 2009), which coincides with the start of their widespread anthropogenic production and consumption.

Atmospheric removal occurs in the stratosphere, where sink processes are dominated by photo-dissociation (Reaction R1) and reaction with O(¹D) (Reaction R2), e.g.



Cl and ClO released in these reactions catalyse the destruction of O₃ (Molina and Rowland, 1974). Following the identification of a strong decrease in O₃ over Antarctica (Farman et al., 1985), swift international action worked to phase-out the production and consumption of CFCs under the Montreal Protocol and its subsequent amendments.

Measuring isotope ratios in atmospheric trace gases provides additional information regarding their sources and sinks, as well as the ways in which they contribute to biogeochemical cycles (Brenninkmeijer et al., 2003; Goldstein and Shaw, 2003). Long-term studies have investigated a range of species, for example, CH₄ (Etheridge et al., 1998), CO₂ (Francey et al., 1999; Rubino et al., 2013) and N₂O (Röckmann et al., 2003; Kaiser et al., 2006).

Of particular relevance to this work is N₂O, a much studied gas with similar sinks and a comparable lifetime to CFC-12. Various studies have used firn and ice core samples (Sowers et al., 2002; Röckmann et al., 2003; Bernard et al., 2006; Ishijima et al., 2007), as well as direct atmospheric samples (Yoshida and Toyoda, 2000; Kaiser et al., 2006) to investigate large scale N₂O isotope and abundance changes. Source inputs depleted in ¹⁵N and ¹⁸O are thought to have caused the concurrent tropospheric

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N_2O abundance increase and $\delta(^{15}\text{N})$ and $\delta(^{18}\text{O})$ decrease over the last century, while sink processes enrich stratospheric N_2O in heavy isotopes (Röckmann et al., 2003). Altitude and latitude dependence has also been observed, with the isotopic fractionation (ϵ) increasing with altitude and as the sampling point moves towards the equator (Kaiser et al., 2006). Firn air studies are limited in their temporal resolution because diffusion processes smooth interannual and seasonal variations. Direct atmospheric air samples have been used to study these shorter scale changes (Nevison et al., 2005; Röckmann and Levin, 2005; Nevison et al., 2011; Park et al., 2012). The seasonality observed in isotope and mole fraction measurements has not been fully explained, but a potential contributor is the movement of N_2O -depleted air with a correspondingly enriched isotopic composition from the stratosphere to the surface (Park et al., 2012).

In contrast to N_2O , the much lower abundance CFCs are poorly characterised because very few studies have investigated their isotope ratios. Redeker et al. (2007) measured tropospheric $\delta(^{13}\text{C})$ values for 37 hydrocarbons and halocarbons, including CFC-11, CFC-12 and CFC-113, over the course of a year. They found no diurnal or seasonal trends outside their analytical uncertainties. Zuiderweg et al. (2012) measured the stable carbon isotope fractionation of CFC-12 and CFC-11 by simulating stratospheric conditions during UV photolysis experiments. They found a fractionation (ϵ) of -55.3‰ for CFC-12, compared to -23.0‰ for CFC-11 (at 233 K). Zuiderweg et al. (2013) reported a large ^{13}C depletion of CFC-12 measured on NEEM firn air, suggesting that a highly ^{13}C -depleted reservoir has undergone significant enrichment (up to 80‰) from 1950 to the present-day.

The only chlorine isotope ratio study of atmospheric CFCs measured the $\delta(^{37}\text{Cl})$ of CFC-12 in the tropical stratosphere (Laube et al., 2010a). Enrichment with altitude, to a maximum of 27‰, was attributed to sink reactions (R1 and R2) occurring more readily with the lighter isotopologue. Assuming a Rayleigh-type fractionation mechanism, the apparent stratospheric isotope fractionation (ϵ_{app}) was calculated to be $(-12.1 \pm 1.7)\text{‰}$. Laube et al. (2010a) highlighted the large effect of sink reactions on the isotopic composition of CFC-12 in the stratosphere. It is expected that this will lead to

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a relative ^{37}Cl -enrichment of tropospheric CFC-12 with respect to the average source isotope ratio through stratosphere–troposphere air exchange, similar to observations made on other gases, such as CH_4 (McCarthy et al., 2001), H_2 (Batenburg et al., 2012) and N_2O (Röckmann et al., 2003).

The strong stratospheric chlorine isotope fractionation of CFC-12 (Laube et al., 2010a) and the substantial change inferred from $\delta(^{13}\text{C})$ measurements (Zuiderweg et al., 2013), motivated us to conduct a more complete investigation of the chlorine isotope history of this species. This type of study has the potential to better constrain global sources and sinks (Röckmann et al., 2003). Here we use new measurements on stratospheric samples to quantify the apparent stratospheric $^{37}\text{Cl}/^{35}\text{Cl}$ isotope fractionation (ϵ_{app}) of chlorine not only in CFC-12, but also in CFC-11 and CFC-113. Alongside these values, stratosphere–troposphere exchange estimates are used to predict the tropospheric chlorine isotope history of these species. We then present new tropospheric measurements on samples from the Cape Grim air archive, as well as firn air from NEEM and Fletcher Promontory. These are compared to the model projections, leading to an evaluation of long-term chlorine isotope ratio changes in the three most abundant CFCs.

2 Methodology

Table 2 gives details of the samples used in this study. Mid- and high-latitude stratospheric air was collected on an M55 Geophysica research aircraft, as part of the European Union's FP7 project RECONCILE (von Hobe et al., 2013). Background tropospheric samples were sourced from Cape Grim (Tasmania) and firn air was sampled at the North Greenland Eemian Ice Drilling (NEEM) field campaign and Fletcher Promontory.

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of other trace compounds (e.g. sulphur hexafluoride, SF₆). 2% were excluded due to a suspected lack of sample integrity.

A static dilution series was prepared by diluting background air with research-grade nitrogen. These samples were used to quantify the response behaviour of the analytical system. Non-linear isotope ratio responses were found at small mole fractions and were corrected (details in the Supplement). 99% of the measurements were unaffected by this non-linearity and did not have to be corrected. The smallest peaks measured for each species define the effective detection limit for this study. One CFC-113 sample was below this limit and rejected.

The stratospheric samples were used to calculate ε_{app} for each species. This was done by assuming a Rayleigh-type mechanism and applying this linear relationship to the data (Kaiser et al., 2006)

$$\ln(1 + \delta^{37}\text{Cl}) \approx \varepsilon_{app} \ln\left(\frac{y}{y_T}\right) \quad (3)$$

where y and y_T are the stratospheric and tropospheric mole fractions, respectively.

A “bootstrap” analysis with 500 samplings (as described in Laube et al., 2013) was used to determine ε_{app} values and their uncertainties. Anomalous stratospheric samples were omitted using an iterative least squares method, reducing the ε_{app} standard error in all cases. Any sample more than two times the standard error of the y estimate away from the linear regression of the Rayleigh fractionation plot was removed. This process was repeated until all data points fell within two times the standard error of the y estimate. Less than 15% of the measurements were rejected using this method.

2.3 Emissions and transport modelling

Following Kaiser (2002) and Röckmann et al. (2003), a two-box model of the atmosphere is used to project the changing tropospheric isotope signature. The following mass balance equations hold for the model atmosphere, partitioned into a stratosphere (index S) and a troposphere/lowermost stratosphere (index T), where chemical loss

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only occurs in the stratosphere

$$n_T \frac{dy_T}{dt} = P - F_{TS} + F_{ST} = P - F(y_T - y_S) \quad (4)$$

$$n_S \frac{dy_S}{dt} = F_{TS} - F_{ST} - L = F(y_T - y_S) - J n_S y_S. \quad (5)$$

The symbols have the following meaning:

- 5 n : amount of air (in mol)
- y : mole fraction (in mol mol⁻¹)
- F : trace gas flux (in mola⁻¹)
- P : trace gas production (in mola⁻¹)
- L : trace gas loss (in mola⁻¹)
- 10 J : loss rate coefficient (in a⁻¹)

The global average lifetime is calculated as the ratio of total burden and total loss

$$\tau = \frac{n_S y_S + n_T y_T}{J n_S y_S} = \frac{n_S y_S + n_T y_T}{F(y_T - y_S) - n_S \frac{dy_S}{dt}}. \quad (6)$$

The parameters n_T , n_S and F are taken from Holton (1990), Trenberth and Guillemot (1994) and Appenzeller et al. (1996). Using the tropospheric mole fraction history $y_T(t)$ (Velders and Daniel, 2014) and an estimate of the global average lifetime τ , Eqs. (4), (5) and (6) can be solved for y_S , J and P . In practice, we adjust J so that the average τ for the years 1999 to 2008 equals the lifetime estimates as per SPARC (2013) and use a fixed value of J for all model years.

To derive the corresponding isotope budget, we write Eqs. (4) and (5) for another isotopologue (primed quantities)

$$n_T \frac{dy'_T}{dt} = P' - F(y'_T - y'_S) \quad (7)$$

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which can be written in δ notation (relative to an arbitrary reference material) as

$$n_T \frac{d[y_T(1 + \delta_T)]}{dt} = P(1 + \delta_P) - F[y_T(1 + \delta_T) - y_S(1 + \delta_S)]. \quad (8)$$

Together with Eq. (4) this gives the tropospheric isotope budget

$$n_T y_T \frac{d\delta_T}{dt} = P(\delta_P - \delta_T) + F y_S (\delta_S - \delta_T). \quad (9)$$

The equivalent equation for the stratosphere is

$$n_S y_S \frac{d\delta_S}{dt} = F y_T (\delta_T - \delta_S) - n_S y_S J \varepsilon_J (1 + \delta_S) \quad (10)$$

where the photochemical isotope fractionation is

$$\varepsilon_J = \frac{J'}{J} - 1 \quad (11)$$

with J' being the loss rate coefficient for the isotope in the numerator of the isotope ratio (here, ^{37}Cl).

ε_J (or J') are not known, so again following the analogue of N_2O (Röckmann et al., 2003), we use the observed apparent Rayleigh isotope fractionation in the stratosphere (ε_{app}), to solve the isotope-specific part of the budget. Based on this apparent Rayleigh isotope fractionation, the relative isotope ratio difference between the stratosphere and the troposphere δ_{ST} can be calculated according to

$$\delta_{\text{ST}} = \frac{1 + \delta_S}{1 + \delta_T} - 1 = \frac{\delta_S - \delta_T}{1 + \delta_T} = \frac{-\delta_{\text{TS}}}{1 + \delta_{\text{TS}}} = \left(\frac{y_S}{y_T}\right)^{\varepsilon_{\text{app}}} - 1. \quad (12)$$

We adjust ε_J so that the average ε_{app} value for the years 1999 to 2008 equals our stratospheric observations and derive the projected atmospheric history of δ_T and δ_S from Eqs. (10) and (12), assuming a constant δ_P value that results in $\delta_T(2006) = 0$, corresponding to the year our tropospheric air standard was filled.

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2.4 Trace gas transport in firn

The migration of gases from the atmosphere through firn is largely controlled by diffusional and gravitational effects, such that: (1) less diffusive gases move through the firn more slowly than more diffusive gases, (2) heavier gases and isotopologues are enriched at depth relative to lighter gases and isotopologues. Consequently: (1) there is no unique ‘‘age-of-air’’ at a given depth in the firn, rather it is specific to an individual gas or isotopologue, (2) a gas can undergo significant isotope fractionation during its movement through the firn. A model of gas transport in firn must therefore include both diffusive transport and gravitational separation to reconstruct changes in atmospheric abundances and isotope ratios over time. These reconstructions have been accomplished here using the model developed at the Laboratoire de Glaciologie et Géophysique de l’Environnement (LGGE) and Grenoble Image Parole Signal Automatique (GISPA-lab). The physical basis of the model is described in Witrant et al. (2012). The model has been further refined to provide a new and more robust definition of the optimal solution, and a more rigorous treatment of isotopes based on a forward firn model written in terms of delta values (Witrant and Martinerie, 2013). The diffusive behaviour of gases is initialised for each firn column by using a set of gases with well-known atmospheric histories (Witrant et al., 2012).

More details on the firn modelling are given in the Supplement. The dates shown here for the firn data are based on median ages for each gas. The age spectrum associated with these median ages is also discussed in the Supplement. When plotting abundances and isotope ratios against median age, it is necessary to correct for gravitational and diffusive fractionation. In the case of abundances, gravitational settling is calculated as the difference between normal model runs and runs in which gravity is set to zero. For within-firn isotope fractionation, a correction is calculated using the reconstructed scenario of the gas (based on its major isotopologue) and assuming a constant atmospheric isotope delta. Changes in fractionation with depth therefore

et al. (1996), namely the 100 mbar and the 380 K isentrope, (2) the 1σ uncertainty in ε_{app} values, (3) the range of “most likely” lifetime values from SPARC (2013).

3.3 Tropospheric data

Using the GC-MS instrument discussed in this work, Laube et al. (2010a) verified that their isotope ratio measurements were not biased by the response behaviour of the analytical system. They measured CFC-12 to a minimum of 77 pmol mol^{-1} . This study includes two additional species and covers a wider range of sample mole fractions, meaning that a more extensive effort was necessary to remove the possibility of a bias in this work (details in Supplement).

From three independent data sets, a total of 44 (CFC-11), 74 (CFC-12) and 48 (CFC-113) $\delta(^{37}\text{Cl})$ measurements are presented, covering the last 60, 50 and 45 years, respectively (Fig. 3). In general, there is good agreement between the Cape Grim and firn air measurements. This agreement is in line with expectations from well-mixed gases such as these. Before 1970, there are 2 (CFC-11), 1 (CFC-12) and 2 (CFC-113) measurements slightly higher than expected, which could represent source $\delta(^{37}\text{Cl})$ changes. Emissions of these gases earlier in the record represent a larger proportion of their total atmospheric load. Our measurements are more sensitive to source $\delta(^{37}\text{Cl})$ changes before 1970. However, a small correction was applied to 4 out of the 5 samples (details in Supplement) and these measurements were made on the lowest abundance samples leading to larger than mean analytical uncertainties. The limited data available before 1970 make it premature at this stage to report a source $\delta(^{37}\text{Cl})$ change in these gases.

The predicted trends for all three species are small (Fig. 3), with the largest isotope changes expected to have occurred in the last two decades. From 1970 to the present-day, the modelled changes can be accounted for within analytical uncertainties (Fig. 3). This means that for these three CFCs, an isotopically constant source signature and the current understanding of transport and destruction processes are consistent with our observations in this period. Despite the noted parallels in atmospheric chemical

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behaviour between CFC-12 and N_2O , their atmospheric isotope delta histories differ because N_2O has been present in the atmosphere for at least 800 000 years longer than CFC-12 (Spahni et al., 2005; Schilt et al., 2010). When anthropogenic emissions began, atmospheric N_2O isotope deltas were the result of well-established release and removal processes, whereas a balanced system did not precede the industrial release of CFC-12. Also, input from ^{15}N - and ^{18}O -depleted sources have led to a decrease in $\delta(^{15}\text{N}, \text{N}_2\text{O})$ and $\delta(^{18}\text{O}, \text{N}_2\text{O})$ over the last century (Röckmann et al., 2003), while our results suggest that the $\delta(^{37}\text{Cl}, \text{CFC-12})$ source has remained constant, at least since 1970.

Zuiderweg et al. (2013) cite changes in production processes as the most likely cause of the significant ^{13}C enrichment of CFC-12. However, as they concede, there is no direct evidence that the reported methodology changes had any effect on the isotope signature of CFC-12. If variations in production processes did cause such a large change in the carbon isotope signature, it appears that the chlorine isotope signature either changed by a much smaller amount, or was unaffected.

4 Conclusions

Samples from the Cape Grim air archive (Tasmania, 1978–2010) and firn air samples from NEEM (Greenland) and Fletcher Promontory (Antarctica) have been used to infer an atmospheric history of chlorine isotopes in the three most abundant CFCs. These measurements cover the last 60 (CFC-11), 50 (CFC-12) and 45 (CFC-113) years. All three species are isotopically enriched in the stratosphere by destruction processes, leading to negative apparent isotope fractionations (ε_{app}). This measured stratospheric isotope dependence and an assumed constant source $\delta(^{37}\text{Cl})$ were used to reconstruct the long-term changes in their tropospheric isotope signatures. The predicted trends are small due to the long atmospheric lifetimes of the species and can largely be accounted for in our tropospheric measurements. Across all species, 5 pre-1970 $\delta(^{37}\text{Cl})$

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Table 1. Key atmospheric characteristics of CFC-11, CFC-12 and CFC-113. (a) NOAA/AGAGE mean values taken from Montzka et al. (2011) and (b) data from SPARC (2013).

	Compound		
	CFC-11	CFC-12	CFC-113
(a) 2009 mole fraction pmol mol^{-1}	243	535	76
(b) Atmospheric lifetime/years	52	102	93
(a) Ozone Depletion Potential (ODP)	1	0.82	0.85

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Table 2. Details of the samples used in this study.

Sample type	Sampling location	Sampling dates	Approx. air age	Further details
Free tropospheric air	Cape Grim, Tasmania	1978–2010	1978–2010	Sampling procedure previously reported (e.g. Langenfelds et al., 1996)
Tropospheric firn air	Fletcher Promontory, Antarctica	Jan 2012	1970–2012	Drilling and firn air pumping operations conducted by the British Antarctic Survey
Tropospheric firn air	North Greenland	Jul 2009	1950–2009	Drilling and firn air pumping operations during the North Greenland Eemian Ice Drilling (NEEM) field campaign
Stratospheric air	Mid-latitude samples (48.1–53.4°)	2009	–	M55 Geophysica research aircraft, take off from Oberpfaffenhofen, Germany
Stratospheric air	High-latitude samples (66.3–76.2°)	2010	–	M55 Geophysica research aircraft, take off from Kiruna, Sweden

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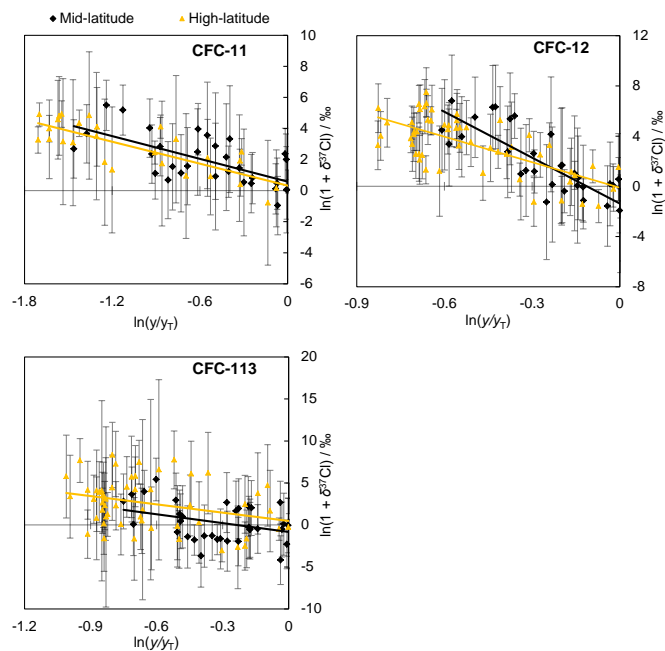


Figure 1. Rayleigh fractionation plots of CFC-11, CFC-12 and CFC-113 chlorine isotope signatures, derived from mid-latitude (black diamonds) and high-latitude (orange triangles) stratospheric samples. Trend lines correspond to apparent fractionations (ϵ_{app}) of (-2.4 ± 0.5) and (-2.3 ± 0.4) ‰ (CFC-11), (-12.2 ± 1.6) and (-6.8 ± 0.8) ‰ (CFC-12) and (-3.5 ± 1.5) and (-3.3 ± 1.2) ‰ (CFC-113) for mid- and high-latitudes, respectively. The standard error of the gradient is quoted for ϵ_{app} values. The average repeatability for individual samples was ± 2.3 ‰ (CFC-11), ± 2.4 ‰ (CFC-12) and ± 3.9 ‰ (CFC-113). 1σ SD error bars are shown.

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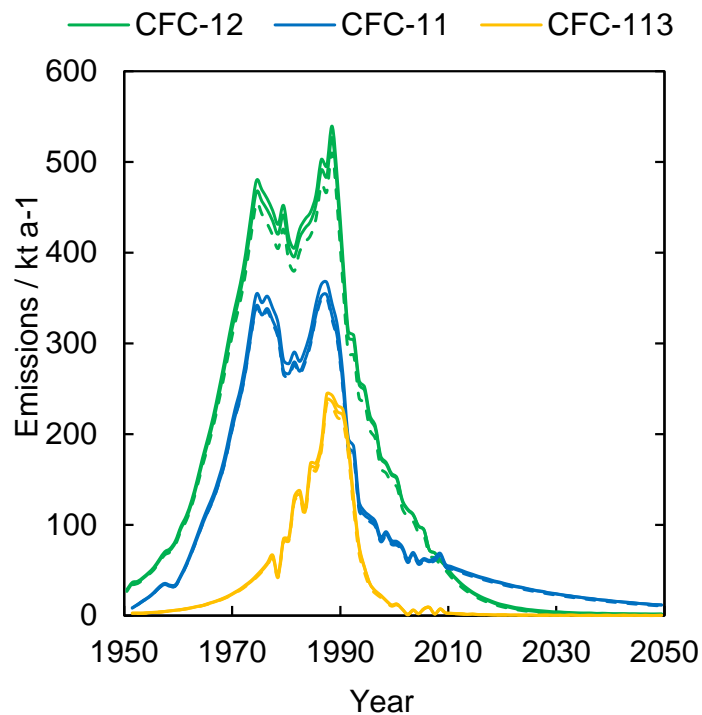


Figure 2. Emissions estimates for CFC-11 (blue), CFC-12 (green) and CFC-113 (orange), based on global tropospheric mole fractions (SPARC, 2013). Each species is represented by three lines. The two emissions estimates from this study (solid lines) are produced using the stratosphere–troposphere exchange flux calculations of Holton (1990) and Appenzeller et al. (1996), with the latter giving slightly higher emissions. These are compared to values from Velders and Daniel (2014) (dashed lines).

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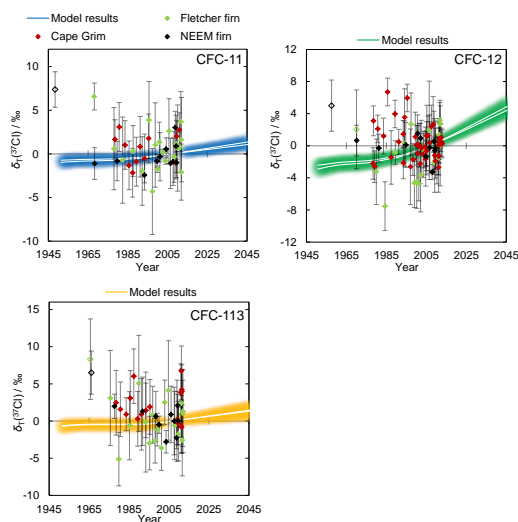


Figure 3. Tropospheric $\delta(^{37}\text{Cl})$ measurements, compared to model predictions. Black diamonds (NEMM firn air), green diamonds (Fletcher Promontory firn air) and red diamonds (Cape Grim air archive) represent the tropospheric chlorine isotope history of CFC-11, CFC-12 and CFC-113. The average repeatability for individual samples was ± 2.7 (CFC-11), ± 2.7 (CFC-12) and ± 3.8 ‰ (CFC-113). 1σ SD error bars are shown. Open symbols indicate that the measurement was subject to a small correction, based on the dilution series analysis (see Supplement). White trend lines represent model estimates of the temporal evolution of $\delta_{\text{T}}(^{37}\text{Cl})$ values. Two lines are displayed per species, one for each stratosphere–troposphere exchange flux estimate (Holton, 1990; Appenzeller et al., 1996). Blue (CFC-11), green (CFC-12) and orange (CFC-113) shading indicates the model uncertainty envelopes. These uncertainties are based on two independent sources of error: (1) the 1σ uncertainty in ε_{app} values, (2) the range of “most likely” lifetime values from SPARC (2013). All delta values are relative to 2006 standard air and the model output has been adjusted so that it passes through 0 ‰ at 2006.