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A comprehensive estimate for loss of atmospheric carbon tetrachloride (CCl$_4$) to the ocean

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Abstract. Extensive undersaturations of carbon tetrachloride (CCl$_4$) in Pacific, Atlantic, and Southern Ocean surface waters indicate that atmospheric CCl$_4$ is consumed in large amounts by the ocean. Observations made on 16 research cruises between 1987 and 2010, ranging in latitude from 60° N to 77° S, show that negative saturations extend over most of the surface ocean. Corrected for physical effects associated with radiative heat flux, mixing, and air injection, these anomalies were commonly on the order of −5 to −10 %, with no clear relationship to temperature, productivity, or other gross surface water characteristics other than being more negative in association with upwelling. The atmospheric flux required to sustain these undersaturations is 12.4 (9.4–15.4) Gg yr^{-1}, a loss rate implying a partial atmospheric lifetime with respect to the oceanic loss of 183 (147–241) yr and that ∼ 18 (14–22) % of atmospheric CCl$_4$ is lost to the ocean. Although CCl$_4$ hydrolyzes in seawater, published hydrolysis rates for this gas are too slow to support such large undersaturations, given our current understanding of air–sea gas exchange rates. The even larger undersaturations in intermediate depth waters associated with reduced oxygen levels, observed in this study and by other investigators, strongly suggest that CCl$_4$ is ubiquitously consumed at mid-depth, presumably by microbiota. Although this subsurface sink creates a gradient that drives a downward flux of CCl$_4$, the gradient alone is not sufficient to explain the observed surface undersaturations. Since known chemical losses are likewise insufficient to sustain the observed undersaturations, this suggests a possible biological sink for CCl$_4$ in surface or near-surface waters of the ocean. The total atmospheric lifetime for CCl$_4$, based on these results and the most recent studies of soil uptake and loss in the stratosphere is now 32 (26–43) yr.

1 Introduction

CCl$_4$ is a strong ozone-depleting gas for which production for dispersive use (e.g., fire suppression, dry cleaning, fumigation) has been banned through the Montreal Protocol (1987) and its amendments and adjustments. Although the concentration of atmospheric CCl$_4$ has been declining in the atmosphere since the early 1990s, its rate of decline is slower than predicted from estimates of emissions suggested by production data reported to the Ozone Secretariat (< 11 Gg yr^{-1} since 2007; Carpenter and Reimann, 2014) and its atmospheric lifetime (e.g, Liang et al., 2014; Carpenter et al., 2014). The dominant loss for CCl$_4$ is through photolysis in the upper atmosphere, which, based on the most recent evaluations, would yield an atmospheric lifetime of 44 (36–58) yr

(Ko et al., 2013; Laube et al., 2013; Rontu Carlon et al., 2010). The oceanic sink, previously calculated as 94 (82–191) yr (Yvon-Lewis and Butler, 2002), also removes significant amounts of CCl$_4$ from the atmosphere. The CCl$_4$ lifetime owing to uptake by soils, previously determined at 90–195 yr, but recently assessed at 375 (288–536) yr, is considered a lesser component (Happell et al., 2014; Rhew and Happell, 2016). These additional sinks had previously brought the overall calculated lifetime of CCl$_4$ in the atmosphere down to 23–35 yr. The oceanic sink for CCl$_4$ determined by Yvon-Lewis and Butler (2002) and used in subsequent Scientific Assessments of Ozone Depletion (e.g., Carpenter et al., 2014), however, was based almost entirely on surface data from four research cruises in the Pacific Ocean from 1987 to 1992 (Butler et al., 1997). Considerable data exist for a deficit of CCl$_4$ in deeper ocean waters, particularly those associated with low oxygen (e.g. Krysell et al., 1994; Tanhua and Olsson, 2005). This study focuses on surface data from the original four cruises and a dozen additional expeditions to enhance the earlier analysis, to examine the oceanic sink for potential sampling and analytical biases, to evaluate the potential cause of the sink, and to provide more confidence in the estimated mean rate of atmospheric CCl$_4$ removal by the ocean. This analysis also takes advantage of significant improvements in determining air–sea exchange rates, which have a substantial impact on the final estimate. Finally, we draw on hydrographic data from selected cruises to evaluate the role of subsurface processes. With extensive surface data from numerous cruises coursing three of the world's major oceans, we provide here a more representative picture of oceanic removal of this gas from the atmosphere.

2   Sampling and analysis

CCl$_4$ mole fractions in air were measured hourly in equilibrated surface water and the atmosphere on most of 16 research cruises, crossing many of the major ocean basins over a period of 23 years (1987–2010; Fig. 1, Table 1). (On four cruises, SAGA II in the West Pacific, CLIVAR-01 in the Southern Ocean, and A16N and A16S in the Atlantic, we sampled daily from surface Niskin bottles; WOCE P18 data, likewise from Niskin bottles, were used only for depth profile analyses and were not accompanied by continuous air measurements.) On the remaining cruises, air samples were collected from the ship's bow and surface samples were obtained with an underway, Weiss-type equilibrator (e.g., Johnson, 1999; Butler et al., 1988). Gases in all samples were separated by gas chromatography on OV-101 or similar columns, some packed, most capillary, and detected with electron capture detectors (GC-ECD) or mass spectrometers (GCMS; Hewlett Packard or Agilent 5971 or 5973). Both types of detectors and different columns were used on some cruises to evaluate potential analytical biases. To evaluate
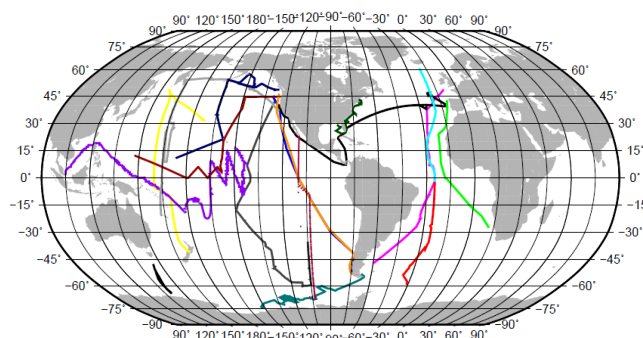

Figure 1. Research cruises contributing to this study: SAGA-2, Leg I (1987, W Pacific; yellow), RITS89 (1989, E Pacific; orange), SAGA-3 (1990, Equatorial Pacific; lime), OAXTC (1992 N. Pacific; light gray, part of WOCE P13), BLAST1 (1994, E Pacific; olive), BLAST2 (1994, Atlantic; cyan), BLAST3 (1996, Southern Ocean; dark gray), GasEx98 (1998, N. Atlantic, Gulf of Mexico, NE Pacific; pink), RB9906 (1999, NE Pacific; cyan), CLIVAR01 SR3 (2001, Southern Ocean; brown), A16N (2003, N Atlantic; dark green), A16S (2005, S Atlantic; red), PHASE (2004, Central and North Pacific; purple), P18 (2008, Tropical and SE Pacific; dark green), GOMECC, Coastal NW Atlantic; olive), HalocAST-P (2010, E Pacific; blue, underlying BLAST 1, HalocAST-A (2009, E Atlantic; green).

potential bias introduced by the equilibrator, surface samples from hydrocasts or bucket samples were obtained and analyzed on several occasions. Gases were extracted from these samples with a purge-and-trap technique (Bullister and Weiss, 1988; Yvon-Lewis et al., 2003), and subsequently analyzed by GCMS or GC-ECD. Results from these grab samples agreed well with those from underway equilibrated surface water (Fig. 2).

Full-depth water column profiles of dissolved CCl$_4$ concentrations, along with CFC-11 and CFC-12, were also collected on several expeditions during this period. These profiles typically consisted of measurements made on discrete water samples collected at 24–36 depths, using purge-and-trap techniques and analyzed with GC-ECD. Depth profiles of CCl$_4$ were obtained in some instances to identify potential zones of CCl$_4$ loss relative to that of CFC-11, a gas of similar physical properties, but unreactive in seawater except under virtually anoxic conditions (Bullister and Lee, 1995; Shapiro et al., 1997). However, to further understand the potential cause of the surface deficits of CCl$_4$, we also evaluated extensive data from World Ocean Circulation Experiment (WOCE) Repeat Section P-18 (2008), which runs from 21° N to 70° S in the East Pacific Ocean, and used CFC-12 as the conservative tracer. This section runs through an extensive, well-developed oxygen minimum, thus allowing a close evaluation of the correlation between dissolved CCl$_4$ and O$_2$.

Table 1. Cruise details including season, ship, analytical technique used to measure CCl$_4$, water sampling technique, and color code for Figs. 1 and 5. Gas chromatography with electron capture detection (GC-ECD) and gas chromatography with mass spectrometry (GCMS) are the analytical techniques.

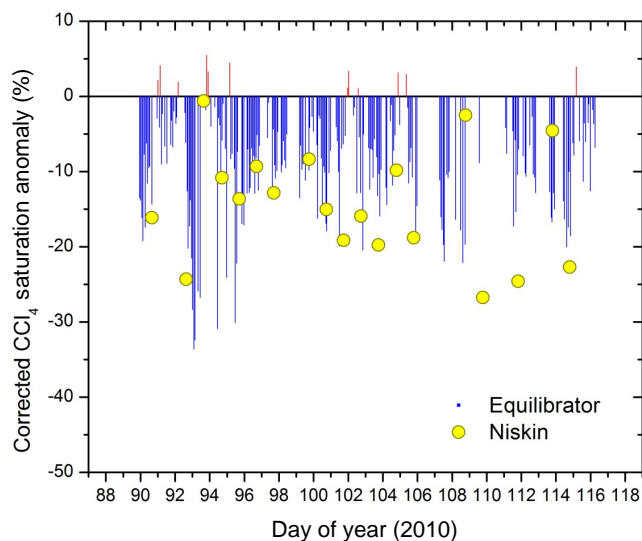| Cruise Name	Dates	Ship	Analytical Technique	Sample Technique	Color Code
SAGA-2	1 May–9 Jun 1987	*Akademik Korolev*	GC-ECD	Equilibrator	Dark Gray
RITS89	6 Feb–19 Apr 1989	*Discoverer*	GC-ECD	Equilibrator	Light gray
SAGA-3	10 Feb–12 Apr 1990	*Akademik Korolev*	GC-ECD	Equilibrator	Blue
OAXTC	4 Aug–21 Oct 1992	*John V. Vickers*	GC-ECD	Equilibrator	Red
BLAST 1	28 Jan–17 Feb 1994	*Discoverer*	GCMS	Equilibrator	Navy
BLAST 2	18 Oct–21 Nov 1994	*Polarstern*	GC-ECD, GCMS	Equilibrator	Cyan
BLAST 3	26 Feb–7 Apr 1996	*Nathaniel B. Palmer*	GCMS	Equilibrator	Magenta
GasEx98	7 May–27 Jul 1998	*Ronald H. Brown*	GCMS	Equilibrator	Black
RB9906	14 Sep–23 Oct 1999	*Ronald H. Brown*	GCMS	Equilibrator	Purple
CLIVAR01	29 Oct–13 Dec 2001	*Aurora Australis*	GCMS	Surface Niskin	Maroon
A16N	4 Jun–11 Aug 2003	*Ronald H. Brown*	GCMS	Surface Niskin	Yellow
PHASE	22 May–2 Jul 2004	*Wecoma*	GCMS	Equilibrator	Brown
A16S	11 Jan–26 Feb 2005	*Ronald H. Brown*	GCMS	Surface Niskin	Tan
GOMECC	10 Jul–4 Aug 2008	*Ronald H. Brown*	GCMS	Equilibrator	Dark Green
HalocAST-P	30 Mar–27 Apr 2010	*Thomas Thompson*	GCMS	Equilibrator and Surface Niskin	Orange
HalocAST-A	25 Oct–26 Nov 2010	*Polarstern*	GCMS	Equilibrator	Green

Figure 2. Corrected CCl$_4$ saturation anomalies determined from measurements of mole fractions in ambient air and air equilibrated with surface water in a Weiss-type equilibrator (blue bars) or extracted from near-surface grab samples (Niskin bottles; yellow points), showing no substantial bias in the equilibration technique for CCl$_4$ (Halocast-P, 2010; see Fig. 1, Table 1). Saturation anomalies in grab samples are calculated using solubility data from Bullister and Wisegarver (1998). This good agreement suggests that the equilibrator does not bias results for CCl$_4$ and further suggests that the solubilities used to calculate saturation anomalies from discrete surface samples are correct.

3 Computations

"Instantaneous" CCl$_4$ fluxes were estimated from the observed difference between partial pressures of CCl$_4$ in the atmosphere and those in the surface ocean, the air–sea exchange velocity, and the solubility and diffusivity of the gas. Influences of in situ physical effects, such as warming, cooling, and mixing (Kester, 1975), on CCl$_4$ saturation anomalies were minimized by subtracting saturation anomalies of CFC-11 (e.g., Butler et al., 1991). The saturation anomaly (Δ_g) for a dissolved gas is expressed as the percent departure of the observed dissolved amount from equilibrium. This is computed from the difference in partial pressures:

$$\Delta_g = 100 \left(\frac{p_{gw} - p_{ga}}{p_{ga}} \right) [\%], \tag{1}$$

where p_{gw} and p_{ga} are the partial pressures of the gas in water and air. If the saturation anomaly of a gas is positive, it indicates that the water is supersaturated and the net flux is from the ocean to the atmosphere. If it is negative, then the net flux of the gas is from the atmosphere to the ocean. However, the saturation anomaly alone is not sufficient for detecting or estimating in situ consumption of the gas in the water. This is because physical processes, such as radiative warming and cooling of surface waters, dissolution of bubbles, or mixing of water masses, can give rise to positive or, in the case of radiative cooling, negative saturation anomalies (Kester, 1975). The magnitude of this effect depends upon the diffusivity of the gas, its solubility, and the temperature dependence of its solubility. In practice, the difference in saturation anomaly for similar compounds is small; saturation anomalies for CFC-11 and CFC-12 typically differ by 1–2 %. Their molecular diffusivities are not much different, nor is

Table 2. Properties of CFC-11, CFC-12 and CCl$_4$ 18.9 °C.

	CFC-11	CFC-12	CCl4
Physical Properties			
Diffusivity (D; 10^5 cm^2 s^{-1})	0.9076	0.9844	0.8465
$(\Delta D/D)/\Delta T$ (% deg^{-1})	2.7	2.7	2.7
Solubility (S; m^3 atm mol^{-1})	0.1116	0.3983	0.0344
$(\Delta S/S)/\Delta T$ (% deg^{-1}; 0–30° C)	3.9	3.6	4.1

the temperature dependence of their solubilities, but their absolute solubility differs by a factor of 3.6 (Table 2).

Consequently, to determine if there is any consumption of a gas such as CCl$_4$ in the water, we computed its *corrected* saturation anomaly to capture that portion of the saturation anomaly that is largely free of physical influences. For CCl$_4$ this is done by subtracting the CFC-11 saturation anomaly from the observed CCl$_4$ saturation anomaly. CFC-11 is chosen over CFC-12, because its physical properties more closely resemble those of CCl$_4$ than do those of CFC-12, which is much less soluble and has a similar percent change in solubility with temperature. Calculated this way, a corrected saturation anomaly that is negative indicates that the gas is probably being consumed in the water, regardless of its non-corrected anomaly. In some of our studies we noted that saturations of CFC-11 and CFC-12, which also have similar physical properties (Table 2), could differ by as much as ∼ 2 % (Butler et al., 1988; Lobert et al., 1995). As a result, we considered in situ consumption of CCl$_4$ significant if the corrected saturation anomaly was more negative than −2 %.

The corrected saturation anomaly of CCl$_4$ should be roughly proportional to its in situ loss or production. If we assume steady-state conditions, the loss or production rate can be calculated from the flux across the surface of the water that is required to maintain the corrected saturation anomaly:

$$F_{\text{CCl}_4,\text{parcel}} = \frac{K_w \, p_{\text{CCl}_{4,a}} A}{H_{\text{CCl}_4}} \left(\frac{\Delta_{\text{CCl}_4} - \Delta_f}{100} \right). \tag{2}$$

Here $F_{\text{CCl}_4,\text{parcel}}$ is the emission of CCl$_4$ across a given parcel (e.g., a 1° × 1° section) of the ocean surface (mol m^{-2} d^{-1}; negative values imply uptake), K_w is the air–sea transfer velocity (m d^{-1}), H_{CCl_4} is Henry's Law constant for CCl$_4$ (m^3 atm mol^{-1}), $p_{\text{CCl}_{4,a}}$ is its partial pressure in the atmosphere, A is the area of the parcel (m^2), Δ_{CCl_4} is its measured saturation anomaly (%), and Δ_f is the saturation anomaly of CFC-11. There is some uncertainty in this kind of flux estimate, mainly associated with K_w, which varies considerably with wind speed and sea surface roughness, but also because the CFC-11 correction is only an approximation.

The rate constant for oceanic removal of atmospheric CCl$_4$ through the parcel, $k_{\text{CCl}_4,\text{parcel}}$, is computed as the ratio of the flux across that surface to the total amount of CCl$_4$ in the

atmosphere,

$$k_{\text{CCl}_4,\text{parcel}} = \frac{K_W \text{Ar}_{\text{CCl}_4}}{n_{\text{tr}} H_{\text{CCl}_4}} \left(\frac{\Delta_{\text{corrected}}}{100} \right), \tag{3}$$

where n_{tr} is the number of moles of air in the troposphere (1.46×10^{20}), r is the fraction of atmospheric CCl$_4$ that resides in the troposphere (0.886), and $\Delta_{\text{corrected}}$ is the difference between Δ_{CCl_4} and Δ_f. To account for spatial variations in surface ocean properties and surface wind speeds, we applied Eq. (3) to a monthly mean, 1° × 1° gridded data set of sea surface temperature (SST), salinity, and wind speed (NOAA, 1988). Providing CCl$_4$ saturation anomalies on this same scale requires an extension of the available measurement data, since we have not made measurements in every 1° × 1° grid cell. To do this, we considered correlations between the measured CCl$_4$ saturations and other properties in the 1° × 1° gridded data set (e.g., temperature, wind speed, season), but found none to be significant in a global sense. Measured saturations do vary somewhat consistently over latitudes, so we used the mean saturation anomalies within latitudinal bands in the gridded data set to compute the global flux. This approach is not perfect – it does not capture unique coastal influences all that well, nor does it capture all areas of upwelling as a category, but it does accommodate the equatorial upwelling influences and those of some other fronts, which appear significant in the data. It also captures the temporal and spatial variability of wind speed and sea surface temperature. We ran additional sets of computations using the global mean and median corrected saturation anomalies and the median saturation anomaly in 10° latitudinal bands to test the uncertainty associated with which averaging approach we selected. Finally, we further evaluated the sensitivity of our results to these choices by alternatively considering an extreme case for the high-latitude Southern Ocean, where we have little data, by selecting a saturation anomaly of −99 % for that region. This last approach examines the highly unlikely possibility that circulation in the Southern Ocean could lead to extreme undersaturations and how it might influence the global flux estimates derived.

4   Results and discussion

4.1   Explaining the observed undersaturations

A comparison of CCl$_4$ mole fractions measured in air from the ship's bow and from the equilibrator headspace shows that CCl$_4$ is largely undersaturated by about 5–10 % in the surface ocean virtually everywhere, nearly all the time (Figs. 3, 4a). Larger undersaturations are measured in equatorial regions and, occasionally, other areas of upwelling, which is consistent with the delivery of subsurface waters undersaturated in CCl$_4$ with respect to CFC-11. Exceptions to this general picture are periodic, small supersaturations, often, but not always, measured in coastal waters or in rough
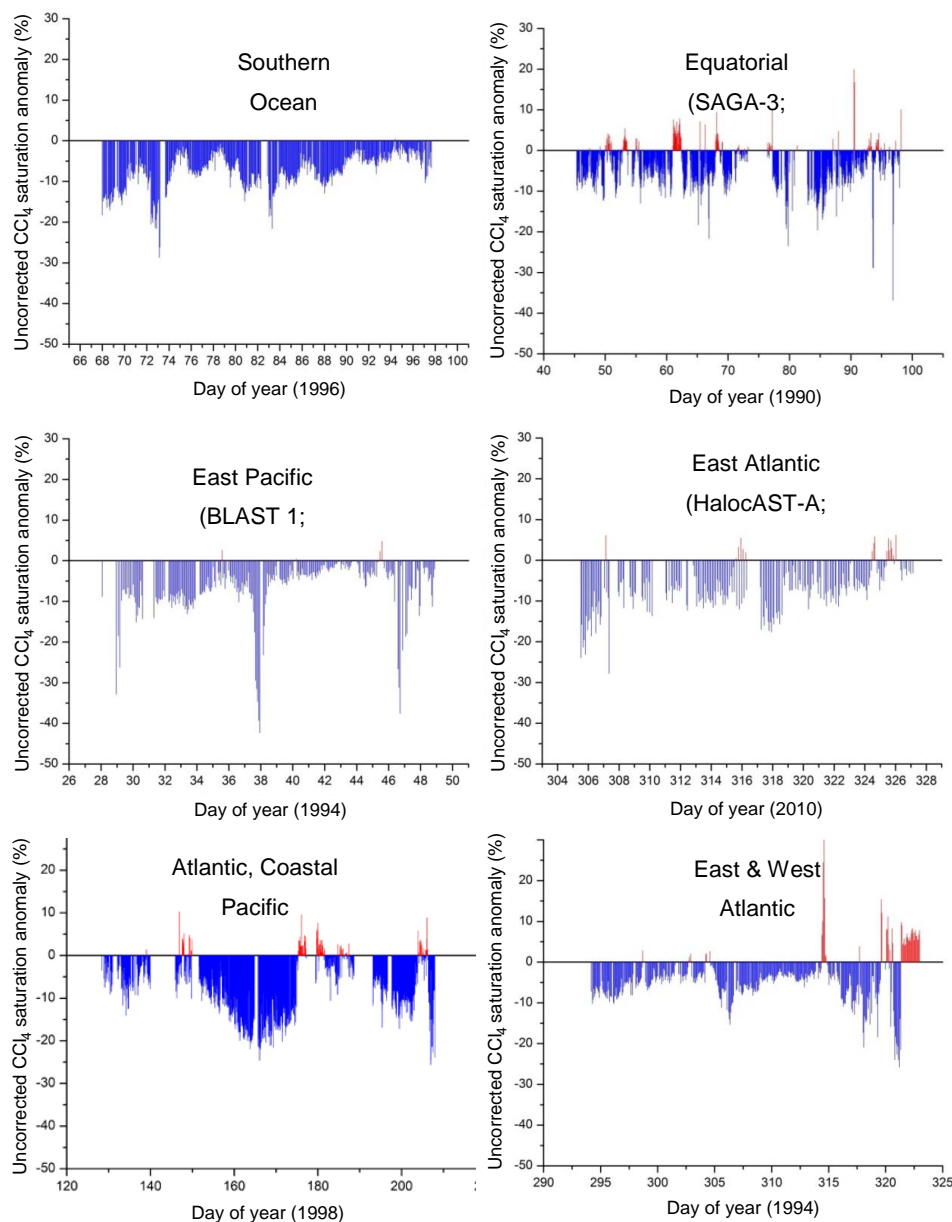
Figure 3. Regionally representative examples of observed (i.e., uncorrected) CCl4 undersaturations from six cruises. Corrections for physical effects can alter these anomalies in either direction, but largely make them more negative by 1–3 %. Negative saturations are indicated with blue lines, positive with red lines.

seas. These could be evidence of periodic CCl4 contamination from the ship or a localized anthropogenic source of CCl4, such as riverine runoff. They could also result from an inadequate correction of physical influences on the saturation anomaly. No evidence exists for production of CCl4 in seawater. Three of the earlier cruises, SAGA II, RITS-89 and OAXTC-92, were conducted at a time when contamination or analytical artifacts associated with these compounds was not uncommon and difficult to avoid, particularly in rough seas, owing to their widespread use and ubiquitous nature as solvents and in other common materials. Consequently, we

have removed these positive values in our flux computations because including them would bias the results. Our goal is to determine an air–sea flux that best represents CCl4 that is irreversibly removed by reactions in the ocean, which, in this instance is estimated by the corrected saturation anomaly.

These widespread undersaturations exceed those that might be expected from physical effects, such as radiative cooling of surface waters. The correction for physical effects that we use here, i.e., subtracting the CFC-11 saturation anomaly, which more often than not is positive, makes the CCl4 saturation anomaly more negative. Although one might
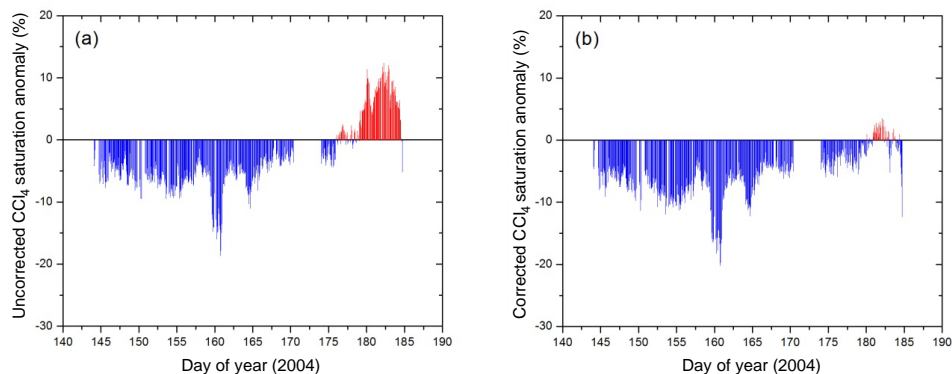
Figure 4. Observed **(a)** and corrected **(b)** CCl$_4$ saturation anomalies, central and NE Pacific, show the influence of physical processes on the observed values. Corrected saturation anomalies were derived by subtracting the observed anomalies of the largely unreactive gas of similar properties, CFC-11. Negative saturation anomalies are indicated with blue lines, positive with red lines.

expect the effects of warming and cooling to balance out on a global basis, effects such as dissolution of bubbles and mixing of waters tend to elevate surface saturation anomalies of all gases (e.g., Kester, 1975; Bowyer and Woolf, 2004). Nevertheless, the undersaturations calculated without these corrections still generally fall within 5–10 %. Large departures from this range were usually associated with high mixing rates (e.g., upwelling) or rapid heating and were corrected somewhat, though not completely, with the algorithms applied here (e.g., Fig. 4a, b). Corrected saturation anomalies more negative than the −5 to −10 % range might still be due in part to differences in the physical properties of CCl$_4$ and CFC-11 that are not fully corrected in our approach (e.g., Table 2). They may also be a result of differences in the atmospheric histories of CCl$_4$ and CFC-11, where CCl$_4$ has a longer and slightly differently shaped historic accumulation in the atmosphere than CFC-11, which is reflected somewhat in its concentration-depth profiles in the ocean as additional scatter in correlation plots of these gases. Finally, trace gases in surface waters may not have always equilibrated with an atmosphere identical to that observed above it. All in all, the corrections used here for physical effects largely reduce error and remove biases in the computation of atmospheric lifetime with respect to oceanic loss and thus lead to a better estimate (e.g., Butler et al., 1991). It is important to note that the corrections for physical effects are most often smaller than the gross flux of CCl$_4$ into the water and therefore do not change the overall picture of widespread undersaturation observed for CCl$_4$.

Plotting the corrected saturation anomalies from our cruises shows a reasonable degree of scatter, but remarkably consistent means and medians (Fig. 5; Table 3). Larger undersaturations near the equator are likely associated with increased upwelling, which brings up water more deficient in CCl$_4$ than CFC-11. Causes of the larger undersaturations in other areas are less clear but include, for example, a Gulf Steam ring (GasEx98), coastal waters, and larger scale
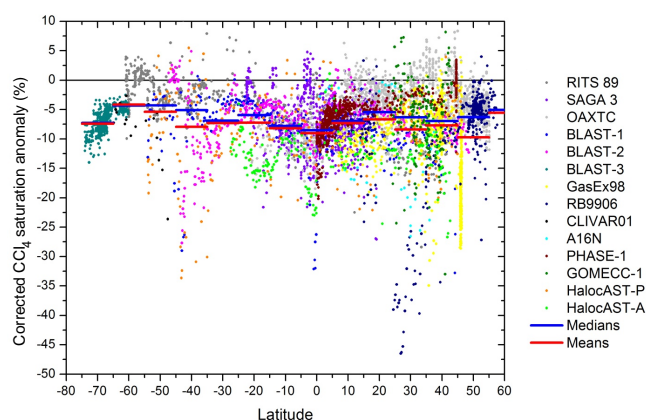

Figure 5. Zonal distribution of the corrected saturation anomaly from all cruises. Because observed positive values are likely the result of contamination, external influences, or sampling or analytical artifacts, means and medians for each 10° latitudinal band (shown as the red and blue bars) are computed from negative saturations only (see text).

ocean fronts, all of which are associated with some degree of upwelling. Yet, some areas of large, negative saturation anomalies, such as the central gyre north of Hawaii during BACPAC-99, are not so readily understood and cannot be attributed to sampling or analytical artifacts.

This relative consistency of undersaturation in surface waters during all seasons, regardless of sea surface temperature and biological regime, might suggest that CCl$_4$ is removed not in surface waters, but at depth, with CCl$_4$ being advected and mixed to deeper waters, where it is consumed. First-order computations of the time required to mix waters between the surface and intermediate depths, however, suggest that, on average, the loss at depth cannot fully support the observed surface water deficits. Air–sea exchange renews gases in surface waters on the order of 20–30 days, whereas, with an eddy diffusivity of $\sim 1\,\mathrm{cm^2\,s^{-1}}$ through the thermo-

Table 3. Corrected saturation anomalies for 10° latitude-bins. Positive values not included (see text and Fig. 5).

Bin	Assigned latitude	mean	median	N	Standard deviation	Standard error
< -65	-70	-7.40	-7.29	372	1.78	0.09
-55 to -65	-60	-4.16	-4.35	176	1.84	0.14
-45 to -55	-50	-5.40	-4.31	148	4.27	0.35
-35 to -45	-40	-7.94	-5.13	192	7.80	0.56
-25 to -35	-30	-7.31	-6.88	129	4.63	0.41
-15 to -25	-20	-7.24	-5.95	255	4.83	0.30
-5 to -15	-10	-8.20	-7.80	331	3.43	0.19
-5 to $+5$	0	-9.02	-8.53	723	4.71	0.18
5 to 15	10	-7.31	-6.97	722	3.64	0.14
15 to 25	20	-6.68	-5.46	539	4.92	0.21
25 to 35	30	-8.39	-6.33	530	8.00	0.35
35 to 45	40	-7.69	-6.97	656	5.46	0.21
45 to 55	50	-9.73	-6.29	880	7.32	0.25
> 55	60	-5.56	-5.06	41	2.67	0.42

cline (Quay and Stuiver, 1980; Li et al., 1984), transport to depths of hundreds of meters from the ocean surface requires times of years to decades. Exceptions are apparent in areas of upwelling, where water from depth can be advected as well as mixed toward the surface in a matter of days (e.g., Tanhua and Liu, 2015). Depletion of CCl$_4$ at depth would be in agreement with other reports suggesting a loss of CCl$_4$ in low oxygen waters (e.g., B.-S. Lee et al., 1999), although the rate of a few percent depletion per year (e.g., Huhn et al., 2001; Min et al., 2010) at depth is still not sufficient to sustain the observed undersaturations at the surface. A number of depth profiles of CCl$_4$ along with CFCs 11 and 12 on our cruises suggest a sink for CCl$_4$ in intermediate waters, typically near the oxygen minimum (Fig. 6). The relationships of the relative saturations of CCl$_4$ vs. CFC-12 and apparent oxygen utilization (AOU) throughout the water column, where AOU is the difference between the calculated atmospheric equilibrium concentration of dissolved oxygen and the measured oxygen concentration, suggest a clear, though water-mass dependent, relationship between the ratio of CCl$_4$ saturation to that of CFC-12 and AOU, particularly in waters of low oxygen concentration (high AOU; Fig. 7). This invokes the possibility of CCl$_4$ degradation by microorganisms in oxygen-deficient waters. The actual mechanisms are not discernible from these data, nor is the degree of loss in waters of different oxygen content, though the loss would be consistent with anaerobic or heterotrophic metabolism.

So, if exchange with deeper waters where CCl$_4$ is anaerobically degraded, presumably by microorganisms (e.g., Krone et al., 1991; C. H. Lee et al., 1999; Tanhua et al., 1996), cannot account for widespread undersaturations in surface waters, then there must also be some mechanism for in situ removal in oxygenated surface water. With a calculated degradation time of ~ 2600 yr, hydrolysis does not provide a removal rate that could balance air–sea exchange (e.g.,
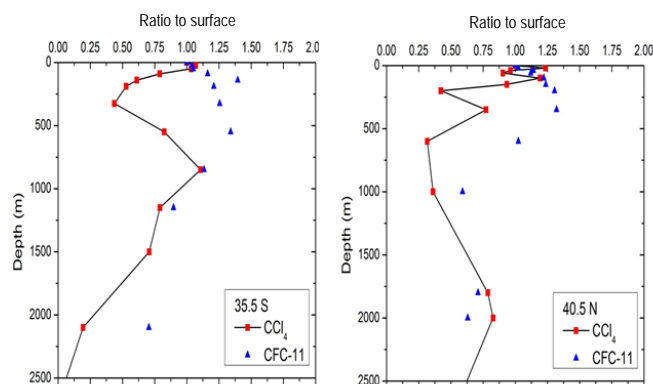

Figure 6. Depth profiles of the concentrations of CCl$_4$ and CFC-11, normalized to the surface value, strongly suggest consumption of CCl$_4$ in intermediate waters. If there were no in situ consumption and the distributions entirely a function of mixing and transport, the curves would be similar if not identical. Plots shown are from the Atlantic Ocean.

Yvon-Lewis and Butler, 2002; Jeffers et al., 1994), so that leaves us with invoking biological or other unknown mechanisms removing CCl$_4$ in well oxygenated surface waters. This is an area that requires further investigation, as there is no direct evidence to date for such pathways in the ocean, although there is evidence for microbial removal of CCl$_4$ in well oxygenated soils (e.g., Mendoza et al., 2011).

4.2 Estimating the air–sea flux

4.2.1 Uncertainties

Because the observed undersaturations in surface waters appear somewhat independent of SST or surface biological activity, the major remaining variable in the distribution of the
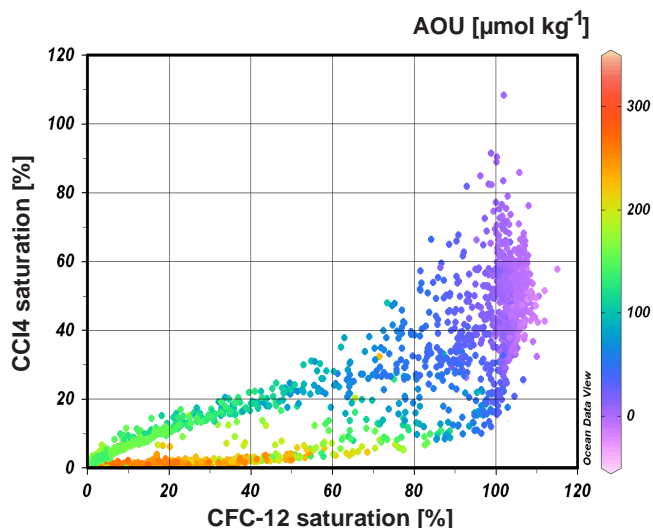
Figure 7. Measured saturations of CCl$_4$ and CFC-12 in sub-surface waters during the P18 cruise in the eastern Pacific as a function of Apparent Oxygen Utilization (AOU). Each point represents a single sample extracted from a Niskin bottle and analyzed by electron-capture gas chromatography. Points are colored according to the AOU in each measured sample.

Figure 8. Various wind speed relationships for the air–sea exchange velocity ($K_{w,660}$) determined in previous studies. Curves from each of these studies were normalized to CO$_2$, but as shown here, have been adjusted to represent CCl$_4$ according to its Schmidt number, which, as a function of viscosity of the medium and diffusivity of the gas, is largely a function of sea surface temperature.

oceanic sink is the air–sea exchange rate, driven largely by wind speed and not, for example, SST, as has been observed for CH$_3$CCl$_3$ and CH$_3$Br (e.g., Butler et al., 1991; King et al., 2002). Our understanding of the dependence of air–sea exchange on wind speed has evolved substantially over time and studies over the past 15 years have been converging (Fig. 8 and Wanninkhof, 2014). Perhaps the most influential of these is by Naegler et al. (2006), which was a re-evaluation of bomb ^{14}CO$_2$ uptake that has led to substantially lower estimates of air–sea exchange (Sweeney et al., 2007; Wanninkhof et al., 2009; Wanninkhof, 2014). These newer studies lower the air–sea flux substantially relative to Wanninkhof (1992), which was based on earlier estimates of the ocean inventory of bomb and natural ^{14}CO$_2$ (e.g., Broecker et al., 1985) and which we used in previous assessments of the loss of atmospheric CCl$_4$ to the ocean (Yvon-Lewis and Butler, 2002). This change significantly decreases the estimated flux of CCl$_4$ and increases the calculated lifetime of the gas in the atmosphere with respect to loss to the ocean. To estimate overall flux uncertainties, we added in quadrature the value of 20 % derived in Wanninkhof (2014) for uncertainty in K_w and the standard deviation of the CCl$_4$ flux ($\sim$ 14 %; Table 4) for the four scenarios under one air–sea exchange relationship (Wanninkhof, 2014). Other uncertainties are either captured within these are comparatively insignificant in the calculations.
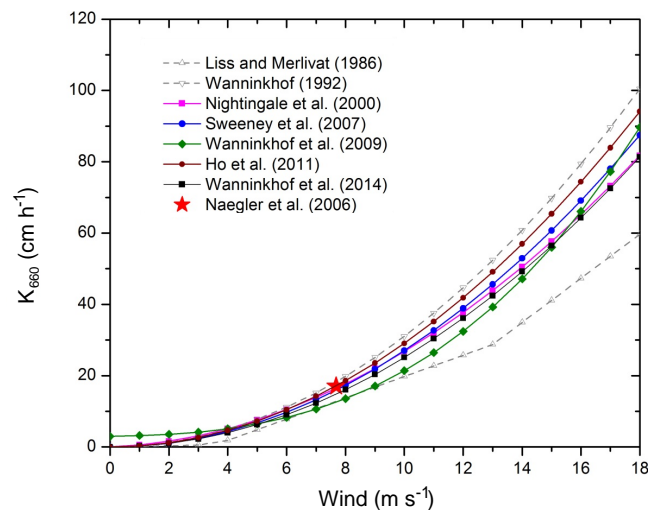
4.2.2   Air–sea flux and lifetime

Propagating the rate constant for oceanic removal of atmospheric CCl$_4$ across the globe with consideration of parameters in Eq. (3) and the updated dependence of K_w on wind speed, our model derives a rate constant for uptake of atmospheric CCl$_4$ by the ocean that is highest where winds are strongest and turnover of surface waters fastest, but that is also influenced by the saturation anomaly value (Fig. 9). As noted in the Methods section, we used four approaches for estimating the overall removal of atmospheric CCl$_4$ by the global ocean, but our preferred approach uses the latitudinally binned, mean saturation anomalies along with the air–sea exchange relationship of Wanninkhof (2014). We note that we obtain the same results by simply using the median of all negative corrected saturation anomalies or by averaging fluxes from all approaches. (Note: our original version of this paper (doi:10.5194/acp-2016-311) preferred the Sweeney et al. (2007) relationship for computing air–sea fluxes and subsequent lifetimes. We had selected that parameterization because it was formulated similarly to Wanninkhof (1992), which had been used in the earlier calculations of the ocean sink of CCl$_4$ (Yvon-Lewis and Butler, 2002), accounted for the change in ocean bomb-^{14}C inventory, and was centered among the distribution of wind speed relationships considered (e.g., Fig. 8). We have since updated that and prefer the Wanninkhof (2014) approach, which has all of these features, includes a rigorous evaluation of the biases and uncertainties in estimating air–sea exchange, and takes additional studies into account.)

Table 4. Oceanic uptake rates and partial atmospheric lifetimes with respect to ocean uptake from a 90 ppt atmosphere using the Wanninkhof (2014) air–sea exchange parameterization applied to the global median corrected saturation anomaly, global mean corrected saturation anomaly, median corrected saturation anomalies for 10° latitudinal bins, and mean corrected saturation anomalies for 10° latitudinal bins. The latitudinal bins are defined in Table 3. The variability in these results demonstrates the uncertainty associated with computational approach. The partial lifetimes are independent of atmospheric amount.

Computational Approach	Global uptake (Gg yr^{-1})	NH uptake (Gg yr^{-1})	SH uptake (Gg yr^{-1})	τ_{ocean} (yr)	$\tau_{NH,ocean}$ (yr)	$\tau_{SH,ocean}$ (yr)
Global median corrected Δ (%)	12.3	3.9	8.4	184	581	270
Global mean corrected Δ (%)	14.4	4.6	9.8	157	492	231
Latitudinally binned, median corrected Δ (%)	10.3	3.6	6.7	220	629	338
Latitudinally binned mean corrected Δ (%)	12.4	4.4	8.0	183	514	283

Summing uptake across the world's oceans suggests that the oceanic sink represents a partial atmospheric lifetime of 183 (147–241) yr, a number considerably longer than the 94 yr of Yvon-Lewis and Butler (2002) used in the 2002–2014 WMO/UNEP Scientific Assessments of Ozone Depletion. Though we have selected the latitudinally binned, global mean, corrected saturation anomaly as our best approach for interpreting the saturations, the difference among selected approaches is relatively small (Table 4). The scenario where we tested the extreme possibility of the Southern Ocean being 99 % undersaturated south of 65° S changed the flux by about 3 Gg yr^{-1} (a ~30 % increase in uptake), but, given the observed undersaturations both north and south of 65° S where we obtained measurements, that scenario is unrealistic; it was run only to understand the limits of Southern Ocean influence. Barring that, and given the hemispheric asymmetries in wind speed and ocean area, the model calculates uptake by the ocean in the Southern Hemisphere that is twice that in the Northern Hemisphere (Table 4).

This updated estimate is based on four times as many observations as used in Yvon-Lewis and Butler (2002), which account for all seasons and cover almost all major ocean basins. The average saturation anomaly used in this study is 10–20 % less than the average used in Yvon-Lewis and Butler (2002). Binning the surface data in our preferred approach (rather than applying a global mean anomaly as done before) to reflect better the actual distribution over the oceans accounts for another 10–20 % decrease (Table 3). The model used by Yvon-Lewis and Butler (2002) was based on the 2° × 2° COADS data set for sea surface temperatures and wind speeds and our new estimate is based on a different, newer data set with 1° × 1° bins. The mean or median wind speed for the 1° × 1° data set is ~5 % lower and winds were distributed differently than in the COADS data set. The most influential change, however, is the use of an updated air–sea exchange coefficient, based on a revised inventory of bomb-^{14}CO$_2$ (Naegler et al., 2006; Wanninkhof, 2014). Yvon-Lewis and Butler (2002) used the Wanninkhof (1992) relationship, which was normalized to an earlier assessment of bomb-^{14}CO$_2$. We evaluated the impact of this change on
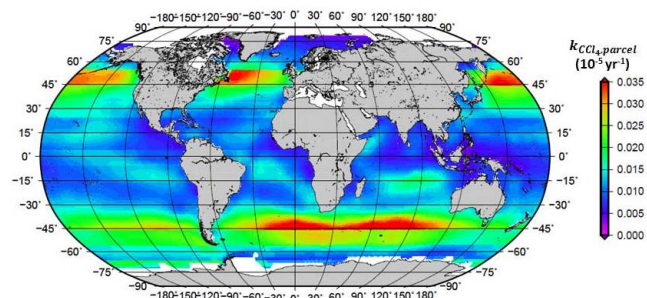

Figure 9. Global distribution of the oceanic uptake rate constant ($k_{CCl_4,parcel}$) based on the mean zonal saturation anomalies (Fig. 5, Table 3) and calculated monthly air–sea exchange rates and other parameters given in Eq. (3) for 1° × 1° cells of the Global Oceanographic Data Set Atlas, Climatologies (NOAA, 1988). Because the saturation anomaly is largely similar among zonal regions of the ocean, variability in removal of CCl$_4$ from the atmosphere is controlled primarily by wind speed, which is highest in higher latitudes of both hemispheres.

CCl$_4$ flux over the ocean and determined that it alone accounts for a 24 % lower flux with Wanninkhof (2014) than with Wanninkhof (1992). Additional reductions came from use of a simpler computational approach that differs from that of Yvon-Lewis and Butler (2002), which was designed for gases where in situ loss rates are known and which required estimates of mixed layer depth and loss during downward mixing through the ocean thermocline. The newly revised estimate for CCl$_4$ uptake provided here is based simply on the air–sea difference in partial pressure and the kinetics of air–sea exchange. It is more robust for this gas, for which there is little understanding of the loss mechanisms, and suggests that the ocean sink is responsible for about 18 % (vs. 32 % previously) of the CCl$_4$ removed from the atmosphere.

5 Implications for atmospheric CCl$_4$

If we consider the revised oceanic sink derived here, the even weaker soil sink of 375 (288–536) yr, and the partial life-

time for CCl_4 removal in the stratosphere of 44 (36–58) yr, the mid-range estimate for the lifetime of CCl_4 in the atmosphere would be 32 (26–43) yr, somewhat longer than the 26 (23–33) yr used in the past four quadrennial assessments on ozone depletion and virtually identical to the 33 (28–41) yr in the recent SPARC report on CCl_4 (Liang et al., 2016). As noted, the ocean is responsible for removing $\sim 18\%$ of the CCl_4 in the atmosphere and the difference in uptake by the ocean in the two hemispheres is $4 \pm 1\,\mathrm{Gg\,yr^{-1}}$ for a 90 ppt atmosphere, so can account for only about 14 % of the current interhemispheric difference of 1.2 ppt in the atmosphere (Table 4). The interhemispheric difference in CCl_4 mixing ratio is still larger than can be accounted for based on known emissions. This suggests additional emissions in the Northern Hemisphere, as the larger oceanic sink reported here for the Southern Hemisphere cannot account for it. Nevertheless, the overall budget of atmospheric CCl_4 is now much closer to being balanced, owing largely to the findings presented in this paper and to the re-evaluation of the soil sink (Happell et al., 2014; Rhew and Happell, 2016).

Considering this new, longer lifetime for atmospheric CCl_4 and the atmospheric trends and distributions given in Carpenter et al. (2014), the remaining discrepancy between potential emissions, as suggested by data on CCl_4 production and destruction reported to the UNEP Ozone Secretariat (e.g., Montzka et al., 2011), and emissions computed from atmospheric lifetime, is now of the order of $10\text{–}20\,\mathrm{Gg\,yr^{-1}}$. While it is possible that historical natural fluxes could in part account for this additional source, no evidence exists for its presence in any significant quantity, especially given the small mole fractions measured in the oldest firn air. Butler et al. (1999) noted that the lowest firn air values were near detection limits in their samples and suggested that CCl_4, though at times not differing from zero, could have been as high as 5–10 ppt in the atmosphere before 1900; recent, unpublished firn air data we have obtained, however, suggest that it is more likely between 3 and 4 ppt in the late 19th century. Consequently, most of the emission discrepancy must arise from heretofore unquantified, anthropogenic sources, predominantly in the Northern Hemisphere, to be consistent with the observed rate of decline for the global mole fraction, given our understanding of the global lifetime and the mean hemispheric difference measured for atmospheric CCl_4 (e.g., Fraser et al., 2014; Liang et al., 2014; Carpenter et al., 2014). Only with an excess of northern hemispheric sources would the deficit identified in this study and its distribution be fully consistent with the observed rate of decline of CCl_4 in the atmosphere (1.2–1.4 % yr^{-1}), the observed interhemispheric gradient of ~ 1.2 ppt in recent years, and an interhemispheric exchange time of the order of 1 year (Carpenter et al., 2014).

6   Data availability

Data used in this report, along with metadata for the cruises, are available at http://www.esrl.noaa.gov/gmd/hats/ocean/. We ask that anyone planning to use the shared data in a publication or product contact James Butler, Shari Yvon-Lewis, or other authors of this manuscript and acknowledge NOAA and Texas A&M as data sources.

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