Methyl iodide production in the ocean: Implications for climate change

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Methyl iodide concentrations of up to 45 pmol L\(^{-1}\), which flux into the marine boundary layer, have been found in low latitude waters of the Atlantic and Indian oceans. These high concentrations correlate well with the abundance of Prochlorococcus, and we have confirmed the release of methyl iodide by this species in laboratory culture experiments. Extrapolating, we estimate the global ocean flux of iodine to the marine boundary layer from this single source to be \(5.3 \times 10^{11} \text{ g L}^{-1}\), which is a large fraction of the previously estimated total global flux and the implications are far reaching. Climate prediction models suggest increases in sea surface temperature and changes in biogeographical provenances in response to global warming. Such changes are likely to increase the abundance of Prochlorococcus, and we estimate a concomitant \(\sim 15\%\) increase in the release of iodine species to the atmosphere. Potentially, this could help mitigate global warming.


1. Introduction

It is generally accepted that the marine environment is the dominant source of a range of volatile iodinated compounds to the atmosphere [Carpenter et al., 2003, 2000; Class and Ballschmiter, 1988; Klick and Abrahamson, 1992; Moore and Tokarczyk, 1992; Schall and Heumann, 1993]. These compounds rapidly dissociate into very reactive iodine radicals, which then participate in a number of different catalytic cycles in the atmosphere [Vogt et al., 1999], resulting in the destruction of stratospheric and tropospheric ozone [Alicke et al., 1999; Chameides and Davis, 1980; Davis et al., 1996; McFiggans et al., 2000; Solomon et al., 1994]. Iodocarbons differ from their other halogen counterparts, as bonds involving iodine are generally weak and photochemically active. This leads to the rapid release of iodine from source gases and reservoirs and also means that the source gases have lifetimes of only a few days or weeks. Initially, it was thought that iodine only impacted on the lower troposphere and that the short tropospheric lifetimes would preclude significant transport to the stratosphere. However, research has shown that during deep convective events, convective clouds can transport material within a few hours from low altitudes to the lower stratosphere, particularly in the tropics [Danielsen, 1993; Kritz et al., 1993]. In addition, recent studies from the PARFORCE project, which investigated the phenomenon of coastal nucleation bursts, have suggested that iodine compounds play a part in the formation of new particles and cloud condensation nuclei (CCNs) [Hoffmann et al., 2001; Makela et al., 2002; McFiggans et al., 2004; O’Dowd et al., 2002b]. An increase in the production of iodocompounds and the subsequent production of CCNs would potentially result in a net cooling of the earth system and hence in a negative climate feedback mechanism, mitigating global warming. At present, however, the exact nature of the precursor compounds is unknown and there is debate on how extensive the formation processes are in global terms [McFiggans, 2005; O’Dowd et al., 2005]. Rapid episodic increases in particle number concentrations have been observed at coastal sites throughout Europe (Sweden, Scotland, Eire, France) during the CEC funded BMCape and PARFORCE projects [O’Dowd et al., 2002a] and prior to these during the OCEANOX experiment in Brittany, France [Mihalopoulos et al., 1992] as well as over sea ice in Antarctic waters [Davison et al., 1996]. These episodes, typically of a few hours in duration and associated with tidal movements, are characterized by particle counts increasing from a few hundred cm\(^{-3}\) to several thousand cm\(^{-3}\) (and occasionally in excess of \(10^5\) cm\(^{-3}\)) before returning to their previous levels. Measurements show that the particles present during these episodes fall within the nucleation mode size range. It is therefore assumed that they are recently formed particles resulting from gas to particle conversion processes. However, for such processes to be effective in global terms, there needs to be an extensive marine source of iodinated gases [McFiggans, 2005; O’Dowd et al., 2005].
sites of data collection. The FOUREX cruise was a section from Spain to Greenland, the CHAOS cruise was a meridional section along 20°W, and SCIPIO was a cruise around the Mascarene Plateau in the Indian Ocean.

The marine sources of volatile iodocompounds are not well constrained. Release by marine macroalgae and ice algae from various climatic regions [Carpenter et al., 1999, 2000; Giese et al., 1999; Laturnus et al., 1997] is well documented, suggesting a predominantly coastal or polar origin. However, IO concentrations of up to 3 ppt have been observed at Teneriffe in air masses with little or no coastal influence [Allan et al., 2000] suggesting a more widespread marine source. Release by phytoplankton has been observed in laboratory cultures [Manley and De la Cuesta, 1997; Moore et al., 1996] and related to phytoplankton activity in open ocean waters [Moore and Groszko, 1999; Moore and Tokarczyk, 1993, 1992; Smythe-Wright and Boswell, 2001], but estimates of the global release of iodine from algae ($10^9$–$10^{10}$ g I yr$^{-1}$) are insufficient to account for the global flux of $10^{11}$–$10^{12}$ g I yr$^{-1}$ [O’Dowd et al., 2002b]. Moreover, it has recently been suggested that there is a significant missing source (or sources) of iodine over the global oceans that is likely to be crucial to the formation of new particles and cloud condensation nuclei [McFiggans, 2005; O’Dowd et al., 2005]. Methylation of iodine by bacteria [Amachi et al., 2001; Manley and Dastoor, 1988] and photochemical action [Bell et al., 2002; Happell and Wallace, 1996; Moore and Zafiriou, 1994; Richter and Wallace, 2004; Yokouchi et al., 2001] have been proposed as alternative source mechanisms, but these too cannot account for the estimated global ocean flux.

We have examined data from three cruises, two in the Atlantic Ocean and one in the Indian Ocean (Figure 1), focusing on methyl iodide production and the importance of a biologically related source. The FOUREX cruise (RRS Discovery cruise 230) was a quasi-zonal section from Spain to Greenland in August/September 1997. The CHAOS cruise (RRS Discovery cruise 233) was a meridional section approximately along 20°W, from 20°N to Iceland in May 1998 and the SCIPIO cruise (RRS Charles Darwin cruise 141) was a grid of stations between the Seychelles and Mauritius in June 2002.

2. Methods

2.1. Trace Gas Analysis

Trace gas concentrations in seawater were measured at sea using purge-and-trap gas chromatography with either electron capture (Shimadzu GC14a,) or mass spectrometric (Hewlett Packard 6890 GC/5973 MSD) detectors. The basic methodology has been described previously [Boswell and Smythe-Wright, 1996; Smythe-Wright and Boswell, 2001]. Seawater samples were collected at each station directly from the CTD hydrographic bottles into gas-tight syringes and stored under water, at surface water temperatures in diffuse light/dark, prior to analysis.

Air samples were either piped directly from the bows of the ship to the analytical equipment using a continuous air line or collected in individual gas-tight syringes. Direct air samples were analyzed using an HP 6890 GC with an adsorption-desorption system front end that allowed ambient temperature trapping of the halocarbons of interest from a 200-mL sample on an adsorbent filled microtrap [Dimmer, 1999]. Syringe air samples were measured using the same instruments as for purge-and-trap analysis.

All samples were calibrated by injection of known quantities of calibrated gas standards. Three sources of cross-calibrated gas were used, one from a cylinder prepared and calibrated by the National Oceanic and Atmospheric Administration Climate Monitoring and Diagnostics Laboratory, the second from a cylinder prepared by Bristol University for the AGAGE (Advanced Global Atmospheric Gases Experiment) atmospheric monitoring program and the third from a Kin-Tek standards generator.

2.2. Plant Pigment Analysis

Samples for plant pigment analysis were collected at 6–8 depth levels in the top 200 m at each station. Up to 2 L of seawater from the hydrographic bottles were filtered through 25-mm GF/F filters to harvest the pigments, and the filters were stored in liquid nitrogen for laboratory analysis approximately 2 months after the cruise. The pigments were then extracted with acetone and analyzed by high-pressure liquid chromatography [Gibb et al., 2001].

2.3. Prochlorococcus Enumeration

Samples for picophytoplankont identification and enumeration were also collected at 6–8 depth levels in the top 200 m at each station. Exactly 1.8 mL of unfiltered seawater was placed in a cryovial and 0.05 mL of 37% formaldehyde added. The formaldehyde had previously been filtered using in-line filters to remove particles and stored in the refrigerator at 4°C. Following a stabilization period of 24 hours in the 4°C refrigerator, the samples were stored at −20°C for later analysis by flow cytometry [Carr et al., 1996].

2.4. Laboratory Culture Experiments

Two strains of Prochlorococcus, MED4 (Mediterranean, high light adapted) and SS120 (Sargasso Sea, low
light adapted) were grown in duplicate in specially designed and fabricated 1-L, 3-arm gas tight culture flasks [Peckett, 2001] together with identical duplicate flasks containing only culture medium. The medium used was that recommended by Station Biologique de Roscoff [Rippka et al., 2000], who supplied the primary cultures, and all flasks were subjected to 20 μE of light and maintained at temperatures of 20°C ± 0.5°C. Headspace gas was measured for trace gases in all six flasks immediately following inoculation of the cultures and then every 2 days using the GC/MS purge-and-trap technique outlined above. In addition to the 40 mL headspace gas samples, 10 mL of medium were also extracted every 2 days for cell count enumeration [Carr et al., 1996]. Immediately following each extraction, clean carbon dioxide enriched air was introduced back into the culture flask using the same arm that was used for sampling. This led to a 50-mL, in total, dilution of the headspace with “new gas,” and this is taken into account when calculating the final concentrations.

3. Results and Discussion

3.1. Observed Concentrations

[11] During the meridional transect from 20°N to Iceland (CHAOS cruise) concentrations of methyl iodide up to 45 pmol L\(^{-1}\) (±1.2 pmol L\(^{-1}\)) were observed in the top 150 m of the water column (6–8 sampling depths, 40-km station spacing) south of 40°N. These high seawater concentrations were mirrored by (hourly) atmospheric methyl iodide concentrations of up to 100 pptv at 20 m above the sea surface (Figures 2a and 2b). A terrestrial source for the high atmospheric concentrations between 20°N and 30°N has been discounted [Dimmer, 1999]. At around 40°N concentrations fell sharply to <10 pmol L\(^{-1}\), and moving northward continued to decrease to ~2.0 pmol L\(^{-1}\) off the coast of Iceland. The abrupt change at 40°N probably reflects the influence of the Azores Current, which acts as a boundary between the subtropical and subpolar gyres. An analysis of phytoplankton distributions at the time of the cruise revealed an abundance of picoplankton, in particular the species Prochlorococcus together with its marker pigment divinyl chlorophyll-a (see Figures 2c and 2d).

[12] Similar results (from 6–8 sampling depths, 40-km spacing) were observed at 41.5°N from 9°W to 20°W along the zonal section of the FOUREX cruise (Figure 3a). Here concentrations were a maximum of 10 pmol L\(^{-1}\), which is in agreement with that seen at 40°N during the CHAOS cruise, but are much lower than at 20°N–30°N. Again there was a coincidence with the pigment divinyl chlorophyll-a (Figure 3b). At 20°W the section turned northward toward Greenland and concentrations declined, reaching levels of less than 2 pmol L\(^{-1}\) west of 30°W as the ship crossed the Mid-Atlantic Ridge and passed into colder waters.

[13] The distribution of methyl iodide and corresponding pigment data for the Indian Ocean study have been previously reported [Smythe-Wright et al., 2005]. Here again, concentrations of up to 40 pmol L\(^{-1}\) were observed with atmospheric concentrations fairly constant at 12 pptv.

3.2. Comparison With Other Data

[14] The concentrations we observed in the North Atlantic, north of 40°N are consistent with those reported by other researchers. For example, Moore and Groszko [1999] observed concentrations of typically less than 2.5 pmol L\(^{-1}\) in the cool waters of the Labrador Sea and levels closer to 6 pmol L\(^{-1}\) in the waters of the Gulf Stream. In addition, Baker et al. [2000] and Moore and Groszko [1999] observed concentrations of up to 6 pmol L\(^{-1}\) off the west coast of Ireland, Richter [2003] reported values of 4–6 pmol L\(^{-1}\) around 20°W, 45°N and Tanzer and Heumann [1992] up to 8.5 pmol L\(^{-1}\) between 40°N–45°N. All these reported values are consistent with the 2 pmol L\(^{-1}\) we observed in polar waters in the western Atlantic and off Iceland and the 5–10 pmol L\(^{-1}\) we observed in the subpolar eastern Atlantic.

[15] More anomalous are the high (up to 45 pmol L\(^{-1}\)) concentrations we observed between 20°N and 30°N. The closest data are those of Chuck et al. [2005], who reported concentrations of 10 pmol L\(^{-1}\) very close to the northwest African coast, but their measurements are well within the upwelling region where biological community structure is somewhat different. Other data for the Atlantic Ocean are those of Happell and Wallace [1996] who observed up to 12 pmol L\(^{-1}\) at 19°S. Concentrations as elevated as ours have been seen in the Eastern Pacific between 40°N and 32°S. Here Singh et al. [1983] report concentrations of 48 pmol L\(^{-1}\). Even higher concentrations of up to 53 pmol L\(^{-1}\) have also been reported by Reifenhauser and Heumann [1992] off the South Shetland Islands in the Southern Ocean.

[16] There are no seawater measurements for the tropical Indian Ocean to corroborate our reported values, which range up to 40 pmol L\(^{-1}\) in the waters surrounding the Mascarene Plateau [Smythe-Wright et al., 2005]. Again the closest data are that of Chuck et al. [2005] who report concentrations of 18 pmol L\(^{-1}\) around 40°S, 35°E.

3.3. Biological Versus Photochemical Source

[17] The flux of methyl iodide to the atmosphere is given by the equation

\[ \text{Flux} = k\Delta C, \]

where \( k \) is the transfer velocity, which is related to prevailing wind speed [McGillis et al., 2001; Wanninkhof and McGillis, 1992], and \( \Delta C \) is the concentration anomaly = \( (C_w - C_a) \), where \( C_w \) is the concentration in seawater and \( C_a \) the concentration in the atmosphere. In simple terms, the flux to the atmosphere equals net production per unit area \( P_a \), where net production takes into account losses other than those to the atmosphere such as reaction with chloride ion [Elliott and Rowland, 1993] or diffusion through the thermocline. Therefore, rearranging the flux equation, we have

\[ \Delta C = \frac{P_a}{k}. \]
Figure 2
According to McGillis et al. [2001] the transfer velocity for a gas with Schmidt number (Sc) = 660 (i.e. CO$_2$ in seawater at 20°C) can be expressed as

$$k_{660} = 3.3 + 0.026 u^3,$$

where $u$ is the wind speed at 10 m. For other gases

$$k = k_{660} (\text{Sc/660})^{-1/2},$$

where Sc is the Schmidt number specific to the gas. For methyl iodide, this can be calculated from the expression [Richter, 2003]

$$\text{Sc} = 2223 - 103.74 T + 1.54 T^2,$$

where $T$ is temperature in °C.

[Richter] 2003 and Richter and Wallace [2004] have made a thorough investigation of photochemical methyl iodide production in the Atlantic Ocean and, allowing for other losses, have estimated a daily net methyl iodide photochemical production rate of 25.5 nmol m$^{-2}$ d$^{-1}$ for the sunlit tropical Atlantic at 10°N. This represents the maximum photochemical production that can be expected for the North Atlantic. The maximum and minimum temperatures observed between 20°N and 40°N during the CHAOS cruise were 21.5°C and 16.0°C and the wind speed at 10 m ranged between 7 and 11 m s$^{-1}$. Using this temperature data and a mean wind speed of 9 m s$^{-1}$, we have calculated Schmidt numbers for methyl iodide over our observed temperature range and gone on to calculate values for $k$ using the technique described by Richter

Figure 2. Distribution along the CHAOS 20°W meridional section of (a) methyl iodide at 20 m above the sea surface (in parts pert trillion by volume, pptv), (b) methyl iodide in seawater (in pmol L$^{-1}$), (c) Prochlorococcus carbon (in µg L$^{-1}$), (d) divinyl chlorophyll-a (in µg L$^{-1}$), and (e) density, sigma-0 (in kg m$^{-3}$).

Figure 3. Distribution of (a) methyl iodide (in pmol L$^{-1}$), (b) divinyl chlorophyll-a (in µg L$^{-1}$), and (c) temperature (in °C) along the FOUREX section.
therefore looked toward a biological source for the methyl iodide and examined the halide data in relation to prevalent biological species both at the micro and pico plankton size class.

We found little or no relationship between the methyl iodide concentrations and microalgal species, but there was coincidence with picoplankton. The relationship between mixed layer seawater methyl iodide concentrations and Prochlorococcus carbon for the North Atlantic and Indian oceans are shown in Figure 4. Figure 2e shows that the mixed layer depth extended to between 50 and 100 m during the CHAOS cruise and a similar situation occurred during the SCIPIO cruise [New et al., 2005]. In general, there is a good correlation ($r^2$ values of 0.88 and 0.82) between the two parameters indicating that the Prochlorococcus can account for >80% of the variability in the methyl iodide concentrations. The observed scatter in the data is not unexpected since the trace gases will tend to diffuse away from the source location masking a direct one-to-one relationship.

3.4. Prochlorococcus Source

We were able to confirm the release of methyl iodide by Prochlorococcus in laboratory culture studies. Figure 5 shows concentrations of up to 21 pmol L$^{-1}$ in two cultures of Prochlorococcus marinus strain MED4 (Mediterranean, high light adapted). Release of methyl iodide was also seen from the cultures of strain SS120 (Sargasso Sea, low light adapted), but at much lower levels (~3 pmol L$^{-1}$). What causes this release is at present unknown. Our results suggest that it might be related to a decline in cell numbers and this effect may contribute to the data scatter seen in Figure 4. The fact that it is the high light adapted strain that appears to be more productive substantiates the idea that release is associated with the excretion of poisonous oxygen products in high light conditions [Mtolera et al., 1996].

Our results for the MED4 strain of Prochlorococcus are higher than those reported by Moore et al. [1996] for the production of methyl iodide in cultures of Nitzschia sp (up to 15 pmol L$^{-1}$) and Porosira glacialis (less than 6 pmol L$^{-1}$). They are also higher than the <8 pmol concentrations reported by Manley and De la Cuesta [1997] in cultures of Nitzschia punctata, Navicula sp, Porosira glacialis and Phaeocystis sp with similar cell numbers to ours. It is difficult to make direct comparison between the rate of release by Prochlorococcus and macroalgae, because of the variety of different units used in the reported release rates (e.g., dry, DW and fresh, FW weight). Nevertheless we have examined the compilation given by Baker et al. [2001], who suggest that the maximum emission rate of methyl iodide from polar and temperate macroalgae is 2.6 ng gFW$^{-1}$ h$^{-1}$ based on the data of Manley and Dastoor [1988] for Laminaria sp. This equates to 0.439 × $10^{-3}$ pmol µg FW$^{-1}$ d$^{-1}$ or 3.3 × $10^{-3}$ pmol µg DW$^{-1}$ d$^{-1}$ (assuming DW/FW ratio of 13% for Laminaria sp), which is 3 orders of magnitude lower than our observed results of 2.25 pmol µgFW$^{-1}$ d$^{-1}$ for Prochlorococcus. However, Ekdahl [1997] report 850 ng gFW$^{-1}$ h$^{-1}$ (equal to 0.143 pmol µgFW$^{-1}$ d$^{-1}$) for the red algae Falkenbergia hillebrandii, suggesting that tropical macroalgae are capable of producing methyl iodide in two cultures value also quoted by Manley and Dastoor [1988] for Prochlorococcus.

Consequently, from the results presented here and those previously reported for the Indian Ocean [Smythe-Wright et al., 2005] photochemistry cannot account for our observed concentration anomalies of up to 39 pmol L$^{-1}$ and cannot be the primary source of methyl iodide in the subtropical/tropical Atlantic and Indian oceans, but may well be the midlatitude source, for example, between 50$^\circ$N and 60$^\circ$N in the Atlantic (see Figure 2b) where concentration anomalies were less than 6.0 pmol L$^{-1}$. In order to sustain our maximum concentration anomaly we would need a production rate of close to 175 nmol m$^{-2}$ d$^{-1}$. We suggest that it might be related to a decline in cell numbers and this effect may contribute to the data scatter seen in Figure 4. The fact that it is the high light adapted strain that appears to be more productive substantiates the idea that release is associated with the excretion of poisonous oxygen products in high light conditions [Mtolera et al., 1996].

Figure 4. Correlation between seawater methyl iodide (in pmol L$^{-1}$), and Prochlorococcus carbon (in µg L$^{-1}$) concentrations during (a) the CHAOS meridional transect in the North Atlantic, $r^2 = 0.88$, and (b) the SCIPIO cruise in the Indian Ocean, $r^2 = 0.82$. [2003]. Substituting the data in the rearranged flux equation gives concentration anomalies of 4.9–5.8 pmol L$^{-1}$. Even using the higher 40.33 nmol m$^{-2}$ d$^{-1}$ value also quoted by Richter [2003], which does not allow for losses, this can only account for 7.8 –9.2 pmol L$^{-1}$.

Therefore, from the results presented here and those previously reported for the Indian Ocean [Smythe-Wright et al., 2005] photochemistry cannot account for our observed concentration anomalies of up to 39 pmol L$^{-1}$ and cannot be the primary source of methyl iodide in the subtropical/tropical Atlantic and Indian oceans, but may well be the midlatitude source, for example, between 50$^\circ$N and 60$^\circ$N in the Atlantic (see Figure 2b) where concentration anomalies were less than 6.0 pmol L$^{-1}$. In order to sustain our maximum concentration anomaly we would need a production rate of close to 175 nmol m$^{-2}$ d$^{-1}$. We
iodide at higher rates, but even then not as high as Prochlorococcus. It therefore appears from these crude calculations that Prochlorococcus are the most efficient producers of methyl iodide in the marine environment.

It has been suggested that other types of bacteria are responsible for both the production and degradation of methyl halides [Amachi et al., 2001; Goodwin et al., 1998; Manley and Dastoor, 1988], but we do not believe that such bacteria have influenced our results in any major way. Recently, a gene has been identified that controls methyl halide production in Arabidopsis (Brassicaceae) [Rhew et al., 2003]. This gene encodes an enzyme (SAM dependent methyl transferase) that is also found in other higher plants as well as in fungi, bacteria and phytoplankton [Rhew et al., 2003]. Similar sequences have been seen in Prochlorococcus (The Institute for Genomic Research (TIGR) database, 2005, http://www.tigr.org), and we are presently investigating common genomic links.

3.5. Oceanic Distribution of Prochlorococcus

Prochlorococcus is one of two genera that make up the photosynthetic prokaryote fraction of picoplankton in the open ocean. Together with Synechococcus they are the most abundant primary producers in the tropical and sub-tropical oceans [Partensky et al., 1999], outnumbering nano and micro phytoplankton by at least an order of magnitude and contributing about 50% of the biomass. Both genera cohabit, but with Prochlorococcus dominating the open ocean oligotrophic regions and Synechococcus the more nutrient rich coastal regions. In general, Prochlorococcus prefers temperatures above 15°C and is confined to latitudes lower than 40°; outside these limits growth is limited and cell numbers decline rapidly [Partensky et al., 1999]. This is confirmed in our North Atlantic data, which show a rapid decline in Prochlorococcus carbon north of 40°N during the CHAOS cruise, where the surface temperature dropped below 15°C, and the existence of Prochlorococcus to 50°N during the FOUREX cruise, when the prevailing temperatures were up to 21°C in the subpolar gyre at the end of the summer (Figure 3c). However, these conditions are not exclusive since Prochlorococcus is also found in mesotrophic conditions, for example in high nutrient–low chlorophyll areas and in coastal regions [Crosbie and Furnas, 2001; Partensky et al., 1999].

3.6. Flux Calculations

Using the flux equations given above and the Wanninkhof relationship for transfer velocity [Wanninkhof and McGillis, 1992], we have calculated the total mean flux of methyl iodide from the North Atlantic between 20°N and 40°N and from the high methyl iodide region of the Indian Ocean between 8°S and 20°S to be 129 nmol m⁻² d⁻¹ and 141 nmol m⁻² d⁻¹, respectively (mean 135 nmol m⁻² d⁻¹). These values are remarkably similar considering they come from two different ocean basins at seasonally different times (although the latter is not a major issue in low-latitude waters) with different prevailing physical conditions at the time of measurement. Taking into account that 25.5 nmol m⁻² d⁻¹ of our observed flux has likely come from photochemistry [Richter and Wallace, 2004], we estimate the mean flux of methyl iodide from biological sources to be 109.5 nmol m⁻² d⁻¹. If we assume this is from Prochlorococcus, which are ubiquitous in the ocean at latitudes of less than 40°, then the flux is released over 106.7 × 10¹² m² of ocean surface area, giving a global flux of methyl iodide of 4.3 × 10⁹ mol yr⁻¹. This
equates to a global flux of iodine from Prochlorococcus of \(5.3 \times 10^{11}\) g I yr\(^{-1}\) (±10%).

3.7. Implications for Climate Change

[26] Models suggest that even allowing for a weakening of the thermohaline circulation, in general, sea surface temperatures will rise by up to 3.5°C by 2100 [Sarmiento et al., 2004] and waters will become more stratified and nutrient depleted [Boyd and Doney, 2002; Sarmiento et al., 2004]. These models suggest that the subtropical, permanently stratified biome will increase by 4.0% in the Northern Hemisphere and 9.4% in the Southern Hemisphere [Sarmiento et al., 2004]. Such shifts are advantageous for colonization by Prochlorococcus and it is likely that this species will become ubiquitous within the <50° latitude band. This would increase the ocean surface area over which the flux of methyl iodide from Prochlorococcus is released to 127.4 \(\times\) \(10^{12}\) m\(^2\), resulting in a ~15% increase in the release of iodine to the atmosphere that could act as a negative feedback to global warming.

4. Conclusion

[27] We have shown that Prochlorococcus is likely to be responsible for a global flux of 5.3 \(\times\) \(10^{11}\) g I yr\(^{-1}\), which is a large fraction of the total estimated global flux of iodine (10\(^{11}\)–10\(^{12}\) g I yr\(^{-1}\)) [O’Dowd et al., 2002b] to the atmosphere. Other workers have highlighted the importance of atmospheric iodine in relation to tropospheric and stratospheric ozone depletion and the formation of cloud condensation nuclei [Chameides and Davis, 1980; Davis et al., 1996; McFiggans et al., 2004; O’Dowd et al., 2002b; Solomon et al., 1994], processes which cause net global cooling rather than warming. As ocean waters become warmer and more stratified, nutrient concentrations will fall and there will likely be a regime shift away from microalgae toward Prochlorococcus. This implies that as global warming takes hold there will be a concomitant increase in the abundance of Prochlorococcus, and we suggest that colonization within the <50° latitude band will result in a ~15% increase in the release of iodine to the atmosphere. Since atmospheric iodide is now thought to instigate cloud condensation nuclei, our results have important implications for global climate change.

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