

Abstract

The potential effect of ocean acidification (OA) on seawater halocarbons in the Arctic was investigated during a mesocosm experiment in Spitsbergen in June–July 2010. Over a period of 5 weeks, natural phytoplankton communities in nine $\sim 50\text{ m}^3$ mesocosms were studied under a range of $p\text{CO}_2$ treatments from $\sim 185\ \mu\text{atm}$ to $\sim 1420\ \mu\text{atm}$. In general, the response of halocarbons to $p\text{CO}_2$ was subtle, or undetectable. A large number of significant correlations with a range of biological parameters (chlorophyll *a*, microbial plankton community, phytoplankton pigments) were identified, indicating a biological control on the concentrations of halocarbons within the mesocosms. The temporal dynamics of iodomethane (CH_3I) alluded to active turnover of this halocarbon in the mesocosms and strong significant correlations with biological parameters suggested a biological source. However, despite a $p\text{CO}_2$ effect on various components of the plankton community, and a strong association between CH_3I and biological parameters, no effect of $p\text{CO}_2$ was seen in CH_3I . Diiodomethane (CH_2I_2) displayed a number of strong relationships with biological parameters. Furthermore, the concentrations, the rate of net production and the sea-to-air flux of CH_2I_2 showed a significant positive response to $p\text{CO}_2$. There was no clear effect of $p\text{CO}_2$ on bromo-carbon concentrations or dynamics. However, periods of significant net loss of bromoform (CHBr_3) were found to be concentration-dependent, and closely correlated with total bacteria, suggesting a degree of biological consumption of this halocarbon in Arctic waters. Although the effects of OA on halocarbon concentrations were marginal, this study provides invaluable information on the production and cycling of halocarbons in a region of the world's oceans likely to experience rapid environmental change in the coming decades.

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1 Introduction

5 Volatile marine halocarbons are produced via a range of biological and photochemical processes in the surface ocean, resulting in a strong flux to the marine atmosphere. Production processes include direct biological synthesis by phytoplankton, bacteria and macroalgae (Tokarczyk and Moore, 1994; Tait and Moore, 1995; Moore et al., 1996; Manley and Cuesta, 1997; Scarratt and Moore, 1998; Amachi et al., 2001; Hughes et al., 2006), and indirect production through reactions between dissolved organic matter and light (Moore and Zafiriou, 1994; Happell and Wallace, 1996; Richter and Wallace, 2004) and/or ozone (Martino et al., 2009). Seawater concentrations of halo-

10 carbons are also controlled by a number of loss processes, including hydrolysis and nucleophilic attack (Zafiriou, 1975; Elliott and Rowland, 1993), photolysis (Jones and Carpenter, 2005; Martino et al., 2005) and bacterial assimilation (King and Saltzman, 1997; Goodwin et al., 1998, 2001). The resultant surface ocean halocarbon pool undergoes sea-air gas exchange, a flux which constitutes the most important source of natural halogens to the atmosphere. Halocarbons are rapidly oxidised in the atmosphere to produce reactive radicals (I, IO, Br, BrO) which exert an important control on ozone (Chameides and Davis, 1980; Solomon et al., 1994; Davis et al., 1996; Read et al., 2008), and act as condensation nuclei for the growth of larger particles and clouds with the potential to influence global climate (O'Dowd et al., 2002).

20 In the Arctic, atmospheric halogen species are implicated in ozone (O₃) depletion events (ODEs) – dramatic losses of tropospheric ozone that occur following the polar sunrise. ODEs are initiated and catalysed by photochemistry that converts inert halide salt ions (e.g. Br⁻) into reactive halogen species (e.g., Br and BrO). The most extensive ODEs occur over the frozen Arctic Ocean, as sea-ice surfaces, brine and frost flowers represent the main source of inorganic bromine during these events (see Simpson et al., 2007 for review). Halocarbons are considered to play an important role in both the initiation and/or the termination of Arctic reactive halogen chemistry (Simpson et al., 2007). Furthermore, results of modelling studies indicate that iodocarbons

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such as diiodomethane (CH_2I_2) have a significantly greater O_3 depletion effect (per molecule) than the addition of further sea-salt derived Br_2/BrCl (Calvert and Lindberg, 2004). Recent work in sub-Arctic Canada has further raised the importance of halocarbons in Arctic atmospheric chemistry. A combination of Differential Optical Absorption Spectroscopy (DOAS) observations of IO and measurements of atmospheric mixing ratios of halocarbons has revealed episodes of elevated IO, accompanied by a variety of iodocarbons (Mahajan et al., 2010). Air-mass back trajectories show that the iodine compounds originated from open water polynyas in the sea-ice covered Hudson Bay. Using the one-dimensional Tropospheric Halogen Chemistry Model (THAMO) (Saiz-Lopez et al., 2008), Mahajan et al. showed that iodocarbon sources from ice-free Arctic waters could account for the observed concentrations of IO. Such levels of IO deplete O_3 at rates comparable to BrO . Furthermore, relatively small amounts of IO can cause a large increase in the O_3 destruction potential of BrO .

The Arctic region is currently experiencing rapid environmental change. Summer sea-ice extent has steadily decreased over the past 30 yr, and the rate of this decline now exceeds any predictions made using IPCC AR4 simulations (Stroeve et al., 2011; Wang and Overland, 2009). The duration of melt season has increased by about 20 days over the last 30 yr for the Arctic as a whole (Markus et al., 2009), and in areas where sea ice concentration has decreased in early summer, the timing of the annual phytoplankton blooms has become significantly earlier (Kahru et al., 2011). The Arctic region is particularly susceptible to ocean acidification (OA). Over 400 billion tons of CO_2 has been released to the atmosphere by human activities over the last 200 yr, one third of which has been soaked up by the oceans (Calderia and Wickett, 2003; Sabine et al., 2004). This unprecedented influx of CO_2 is resulting in an increase in H^+ ion concentrations manifested as a drop in surface ocean pH, accompanied by a decrease in the saturation state of calcium carbonate (Ω_{CaCO_3}). This reduces the availability of carbonate ions (CO_3^{2-}) which are a vital component of the skeletons of calcifying marine organisms (Orr et al., 2005). Due to increased CO_2 solubility at lower water temperatures and a rapidly changing climate leading to land- and sea-ice loss, model

predictions suggest that areas of the surface Arctic Ocean will, within the next decade, be the first to experience the effects of OA (Steinacher et al., 2009).

Recently, there has been interest in how the changing Arctic climate may influence the sea-to-air flux of halocarbons. Declining sea ice extent and thickness, accompanied by an increase in open water and marine primary productivity, could result in an upturn in the net production and resulting flux of halocarbons to the atmosphere (Mahajan et al., 2010; Shaw et al., 2011; Zhang et al., 2010). This may elevate the importance of halocarbons in Arctic atmospheric chemistry and O₃ regulation. However, nothing is known of the response of marine halocarbons to OA in the Arctic, a phenomenon that will go hand-in-hand with climatic changes. Species shifts in phytoplankton (see Riebesell and Tortell, 2011 for review) and increased rates of bacterial activity (Piontek et al., 2010) are anticipated responses to future OA. Furthermore, different species of phytoplankton are known to produce differing quantities of halocarbons. Therefore, in order to gain an understanding of the future flux of marine halocarbons to the atmosphere, a whole ecosystem approach must be adopted and mesocosm experiments provide this vital platform.

Previous mesocosm experiments performed in Norwegian temperate coastal waters have given contrasting effects of OA on halocarbons. Wingenter et al. (2007) observed large increases in chloriodomethane (CH₂ClI) under 2 × ambient CO₂ (46 ± 4 %) and 3 × ambient CO₂ (131 ± 11 %) relative to ambient control mesocosms. By contrast, Hopkins et al. (2010) reported large and significant decreases in a variety of iodocarbons under high CO₂ (~ 750 μatm) relative to present day control mesocosms (~ 380 μatm), whilst the bromocarbons showed little response to the future conditions. Now, we need to understand how the net production of halocarbons from other ocean regions may respond to OA. As marine halocarbons may be key players in Arctic atmospheric chemistry now and perhaps to a greater extent in the future, the aim of this study was to investigate how their net production and subsequent sea-to-air flux may be affected by future OA.

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2 Materials and methods

2.1 General experimental set-up

A mesocosm experiment was performed between 31 May and 7 July 2010 in Kongsfjorden (78° 56.2' N; 11° 53.6' E), on the west coast of Spitsbergen, the largest island of the Svalbard Archipelago. A total of nine mesocosms were deployed in the fjord, moored in sets of three with ~ 40 m between each mesocosm and ~ 50 m between the triplets, and each capable of enclosing ~ 50 m³ of seawater. The mesocosms were filled with fjord water which was screened through 3 mm mesh to eliminate larger organisms including pteropods. On 2 June (t_{-5}), the mesocosms were closed, allowing no further exchange with the surrounding fjord water. The mesocosms were fully open to the atmosphere, but were covered with a transparent protective lid to minimise external nutrient inputs from seabirds and rain. Seven of the mesocosms received varying amounts of CO₂-saturated seawater over a period of 5 days (t_{-1} to t_4), resulting in a range of $p\text{CO}_2$ levels, from ~ 185 to 1420 μatm (Phase 0). The two control mesocosms received no CO₂ addition, and represented the in situ carbonate chemistry of the fjord (175–180 μatm). Once the $p\text{CO}_2$ /pH levels had been adjusted, daily experimental sampling of the mesocosms for halocarbons began, commencing on 11 June (t_4) and continuing until 4 July (t_{27}). Table 1 gives a summary of the mean $p\text{CO}_2$ (μatm) and pH (on the total scale) for the period t_4 – t_{27} , as well as mean values for the three experimental phases that are referred to in this paper. Nutrients were added to the mesocosms on 20 June (t_{13}) (mean concentrations: nitrate 5.56 μM , phosphate 0.39 μM , and silicate 1.47 μM). Full details of the experimental setup, evolution of the carbonate systems of the mesocosms and nutrient additions are given by Bellerby et al. (2012) and Schulz et al. (2012) (this issue).

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2.2 Sampling for halocarbon compounds

Samples for halocarbon analysis were taken using a depth integrating water sampler (IWS) (Hydrobios, Kiel, Germany) deployed from a small boat, suitable for the collection of trace gas-sensitive samples. The sampler was manually lowered through the water column to depth, and programmed to collect a 12 m-integrated sample. Once returned to the boat, a length of Tygon tubing was attached to the outlet at the bottom of the sampler and sub-samples for halocarbon analysis were collected in 250 ml amber glass-stoppered bottles. The bottle was rinsed three times before the Tygon tubing was placed to the bottom of the bottle, allowing it to gently fill and overflow three times. On the fourth filling, the bottle was filled to the top and the glass-stopper was replaced, ensuring the absence of bubbles or headspace. Samples were transported in a cool box back to the laboratory onshore, and all were analysed within 6 h of collection.

2.3 Quantification of halocarbons compounds

Seawater sub-samples were gently withdrawn from the amber glass-stoppered bottles using a 100 ml glass syringe and 1/8" nylon syringe extension. The sample was filtered through a 0.7 µm filter (GF/F, Whatman) into a second syringe, ensuring that the introduction of bubbles into the samples was avoided at all times. Following the addition of two deuterated surrogate analytes to monitor instrument sensitivity drift (Martino et al., 2005; Hughes et al., 2006), a 40 ml sample was injected into a glass purge vessel, and the halocarbons were extracted by purging the seawater with ultra-high purity (BIP) nitrogen for 10 min at a flow rate of 90 ml min⁻¹. Aerosols were removed from the purge gas stream using glass wool contained within a section of glass tubing, and a counterflow nafion drier using oxygen-free nitrogen at a flow rate of 180 ml min⁻¹ was used to dry the gas. Halocarbons were trapped on triple-bed stainless steel solid sorbent tubes (Markes International Ltd.) containing Tenax, Carbograph and Carboxen, held at 1–2 °C in a custom-made peltier-cooled metal block. Sample tubes were analysed immediately after trapping using a semi-automated system consisting of an Agilent

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Gas Chromatograph-Mass Spectrometer (GC-MS), coupled to a Markes Unity thermal desorption (TD) platform. The GC was fitted with a 60 m DB-VRX capillary column (0.32 μm film thickness, J & W Ltd.), and the MS was operated in electron ionization (EI)/single ion mode (SIM) throughout the analyses. Within Unity, the sample tubes were heated to 200 $^{\circ}\text{C}$ for 5 min, and the desorbed sample was refocused on a cold trap held at -10°C . Following this, the cold trap underwent rapid heating up to 290 $^{\circ}\text{C}$ at a rate of 100 $^{\circ}\text{C s}^{-1}$ and the sample was introduced to the GC column using a He carrier flow rate of 2 ml min^{-1} . The GC oven was held at 40 $^{\circ}\text{C}$ for 5 min, then heated to 200 $^{\circ}\text{C}$ at a rate of 20 $^{\circ}\text{C min}^{-1}$ and held for 2 min. Finally the oven was heated to 240 $^{\circ}\text{C}$ at a rate of 20 $^{\circ}\text{C min}^{-1}$ and held for 4 min. The total run time was 21 min, and the MS collected data between 6 and 14 min of the run. Calibration and quantification of the compounds was performed using laboratory-prepared liquid standards, by dilution of the pure compounds into ultra-high purity methanol. The primary standards were prepared gravimetrically, the secondary and working standards by serial dilution. The analytical error as based on triplicate samples were: < 5% for Iodomethane (CH_3I), 2-iodopropane (2- $\text{C}_3\text{H}_7\text{I}$), 1-iodopropane (1- $\text{C}_3\text{H}_7\text{I}$), chloriodomethane (CH_2ClI), bromiodomethane (CH_2BrI), < 10% for CH_2I_2 , bromoform (CHBr_3), dibromomethane (CH_2Br_2), dibromochloromethane (CHBr_2Cl), and 10–15% for Iodoethane ($\text{C}_2\text{H}_5\text{I}$), bromochloromethane (CH_2BrCl).

2.4 Sea-to-air flux of halocarbons

The sea-to-air flux of halocarbons, determined by the concentration difference between the air and seawater after correcting for solubility, was estimated for all mesocosms. Gas exchange in the mesocosms was determined by the addition of 3 times-atmospheric concentrations of N_2O and the measurement of the subsequent loss rates, allowing the transfer velocity (k) of N_2O to be derived and enabling the estimation of the flux of other gases. For a detailed description of methods and results, see Czerny et al.

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(2012, this issue). Transfer velocities of halocarbons (k_{halo}) were derived as follows:

$$k_{\text{halo}} = k_{\text{N}_2\text{O}} / (\text{Sc}_{\text{halo}} / \text{Sc}_{\text{N}_2\text{O}})^{0.5} \quad (1)$$

The Schmidt number of halocarbons (Sc_{halo}) was estimated based on experimentally-determined values of molecular diffusivity for CH_3Br (De Bruyn and Saltzman, 1997), using an approach described by Moore and Groszko (1999). Estimated fluxes of halocarbons could then be calculated, using experimentally determined values of the dimensionless Henry's Law Coefficient (Moore et al., 1995), and the only reported atmospheric concentrations of halocarbons from Ny-Ålesund reported by Schall and Heumann (1993). Fluxes were low relative to open ocean measurements due to the sheltered nature of the mesocosm environment and a minimal wind speed component (see Czerny et al., 2012, this issue).

2.5 Ancillary measurements

2.5.1 Chl *a* and additional phytoplankton pigments

Samples for both chl *a* and additional phytoplankton pigments were processed as soon as possible after sampling, and in the meantime, were stored at the in situ temperature of the fjord. For chl *a* 500 ml of seawater was filtered onto GF/F filters (Whatman), and immediately frozen and stored at -20°C . Chl *a* was measured after a minimum of 24 h in the freezer, and extraction was performed with 10 ml acetone (90 %). The filter was homogenised for 4 min with 5 ml acetone, after which an additional 5 ml was added and the sample centrifuged. The supernatant was then analysed fluorometrically after the method of Welschmeyer (1994). For determination of the individual phytoplankton pigments, 2×750 ml were filtered, which was reduced to 1×750 ml at the onset of the bloom. The filters were immediately frozen and stored at -80°C until analysis at GEOMAR. Pigments were extracted with 3 ml acetone and analysed using high pressure liquid chromatography (HPLC), with the addition of Canthaxanthin as an internal standard.

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2.5.2 Microbial abundance

Samples for the quantification of the microbial plankton community were taken directly from the integrated water sampler from the same cast used to collect the halocarbon samples, thus providing plankton counts that were directly comparable to halocarbon concentrations.

2.5.3 Phytoplankton abundance and composition

Phytoplankton composition and abundance were determined by analysis of fresh samples on a Becton Dickinson FACSsort flow cytometer (FCM) equipped with a 15 mW laser exciting at 488 nm and with a standard filter set up. Samples were analysed at high flow rate ($\sim 150 \mu\text{l min}^{-1}$), and specific phytoplankton groups were discriminated in bivariate scatter plots by differences in side scatter and red-orange fluorescence (Tarran et al., 2001).

2.5.4 Total bacteria abundance

Samples for bacterial enumeration were fixed for 30 min at 7°C with glutaraldehyde (25 %, EM-grade) at a final concentration of 0.5 % before snap freezing in liquid nitrogen and storage at -80°C until analysis. Bacteria were counted using a FCM according to Marie et al. (1999). Briefly, thawed samples were diluted with Tris-EDTA buffer (10 mM Tris-HCl and 1 mM EDTA, pH 8) and stained with the green fluorescent nucleic acid-specific dye SYBR-Green I (Molecular Probes, Invitrogen Inc.) at a final concentration of 1×10^{-4} of the commercial stock, in the dark at room temperature for 15 min. Bacteria were discriminated in bivariate scatter plots of green fluorescence versus side scatter.

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2.6 Statistical analyses

In order to identify differences in halocarbon concentrations between mesocosms, one-way analyses of variance (ANOVA) were applied to the data. Initially, tests of normality were applied ($p < 0.05$ = not normal), and if data failed to fit the assumptions of the test, linearity transformations of the data were performed (logarithmic or square root), and the ANOVA proceeded from this point. The results of ANOVA are given as follows: F = ratio of mean squares, df = degrees of freedom, σ = significance of F -test, p = level of confidence. For those data which still failed to display normality following transformation, a rank-based Kruskal–Wallis test was applied (H = test statistic, df = degrees of freedom, p = level of confidence).

Relationships between halocarbons and a range of other parameters were investigated using Pearson's correlation coefficients (R), along with the associated probability (F test, $p < 0.05$ = significant). Net loss and production rates of halocarbons were derived from linear regression analyses of halocarbon concentration data as a function of time, to give the rate coefficient ($\text{pmol l}^{-1} \text{d}^{-1}$), the coefficient of determination (R^2), the standard error (SE) of the rate and the associated level of confidence (F test, $p < 0.05$ = significant).

3 Results

3.1 Halocarbon temporal dynamics

Data for chlorophyll a (chl a) and microbial plankton counts (nanoeukaryotes and picoeukaryotes, total bacteria) are shown in Fig. 1, and concentrations of halocarbons are shown in Fig. 2. The experiment was divided into three phases (PI, PII, PIII) based on the addition of nutrients and the dynamics of chl a (see Schulz et al., 2012, this issue). The divisions between phases are indicated on the figures as grey vertical lines (see Table 1 for a summary of timings). Mean concentrations of all halocarbons

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in the control mesocosms M3 and M7 were not significantly different from each other (Kruskal–Wallis ANOVA on ranks, $p > 0.05$ for all halocarbons).

3.1.1 Iodocarbons

Concentrations of CH_3I and $\text{C}_2\text{H}_5\text{I}$ (Fig. 2a, b) showed some variability over the course of the experiment, falling gradually during PI, in parallel to chl *a* concentrations and nanophytoplankton abundances. Peaks occurred following nutrient addition and in parallel with the chl *a* peak on t_{19} in PII, and during the rapid rise in chl *a* observed during PIII. Concentrations ranged from 0.04 to 10.29 pmol l^{-1} and 0.06 to 3.32 pmol l^{-1} , for CH_3I and $\text{C}_2\text{H}_5\text{I}$, respectively. Concentrations of the propyl iodides (Fig. 2c) were less variable, with concentrations varying by less than 0.5 pmol l^{-1} , and overall mean concentrations of 0.21 pmol l^{-1} (2- $\text{C}_3\text{H}_7\text{I}$) and 0.12 pmol l^{-1} (1- $\text{C}_3\text{H}_7\text{I}$). However, concentrations did show some increase that coincided with the final chl *a* maximum in PIII. 2- $\text{C}_3\text{H}_7\text{I}$ was consistently higher than 1- $\text{C}_3\text{H}_7\text{I}$ by $\sim 0.1 \text{ pmol l}^{-1}$. For all of the above and for the experiment as a whole, no significant differences in mean concentrations were detected between mesocosms and no apparent effect of $p\text{CO}_2$ were observed (Kruskal–Wallis ANOVA on ranks (df = 8): CH_3I $H = 6.06$, $p = 0.64$; $\text{C}_2\text{H}_5\text{I}$ $H = 15.03$, $p = 0.06$; 2- $\text{C}_3\text{H}_7\text{I}$ $H = 11.73$, $p = 0.11$; 1- $\text{C}_3\text{H}_7\text{I}$ $H = 10.22$, $p = 0.18$).

In contrast to all other halocarbons, CH_2I_2 concentrations (Fig. 2d) gradually increased over the course of the experiment, from below detection limit (D.L., $< 10 \text{ fmol l}^{-1}$) on t_4 , reaching 0.5–1.0 pmol l^{-1} by t_{27} . M1 displayed significantly higher concentrations over almost the entire duration of the experiment, with a maximum and seemingly anomalous value of 2.5 pmol l^{-1} on t_{19} (ANOVA $F = 2.52$, df = 8, $\sigma = 0.014$, $p < 0.05$). In PIII concentrations showed some response to $p\text{CO}_2$ treatment, with significantly higher mean CH_2I_2 concentrations as a function of mean $p\text{CO}_2$ ($R^2 = 0.451$, $n = 9$, $p < 0.05$). CH_2ClI concentrations (Fig. 2e) were generally stable (0.5–1.0 pmol l^{-1}), with the exception of a small rise to a peak on t_{14} , just prior to the nutrient-induced increase in chl *a*. Finally, concentrations of CH_2ClI did not respond

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significantly to $p\text{CO}_2$ treatment, although concentrations in M1 were significantly higher than M6, M7 and M8 (Kruskal–Wallis ANOVA on ranks $H = 22.19$, $df = 8$, $p = 0.005$, pairwise comparison with Dunn’s method – all $p < 0.05$).

3.1.2 Bromocarbons

5 The temporal development of concentrations of CHBr_3 , CH_2Br_2 and CHBr_2Cl (Fig. 2f–h) showed a high degree of similarity, with a gradual rise from t_6 , a sharp drop at the start of PII followed by a period of recovery during the nutrient-induced chl *a* peak, and falling or unchanging concentrations during PIII. For the entire experiment the concentrations of $\text{CHBr}_3 > \text{CH}_2\text{Br}_2 > \text{CHBr}_2\text{Cl}$ with mean concentrations for all mesocosms of
10 72.8 pmol l^{-1} , 12.4 pmol l^{-1} and 2.8 pmol l^{-1} , respectively. Similarly to CH_2I_2 , concentrations of CHBr_3 , CH_2Br_2 and CHBr_2Cl were almost consistently higher in M1 (significantly higher for CHBr_3 Kruskal–Wallis ANOVA on ranks $H = 27.258$, $df = 8$, $p < 0.001$), although followed similar temporal trends to the other mesocosms. Concentrations of CH_2BrCl (Fig. 2i) were low ($< 0.1 \text{ pmol l}^{-1}$) and stable, with the exception of a small
15 number of anomalous data points in PI and PII. CH_2BrI showed little variability as the experiment progressed (overall mean = 0.35 pmol l^{-1}), with the exception of some anomalous spikes in concentration during PI and II, and little response to nutrient-addition or phytoplankton growth (Fig. 2j). No significant responses to $p\text{CO}_2$ were detected (Kruskal–Wallis ANOVA on ranks ($df = 8$): CHBr_3 $H = 3.94$, $p = 0.86$; CH_2Br_2
20 $H = 2.22$, $p = 0.95$; CH_2BrCl $H = 8.94$, $p = 0.35$; CHBr_2Cl $H = 4.84$, $p = 0.68$; CH_2BrI $H = 10.67$, $p = 0.16$).

3.2 Halocarbons and biological parameters

In order to identify possible sources or sinks in the mesocosms, concentrations of halocarbons were compared with a number of biological parameters (chl *a*, nanoeukaryote
25 and picoeukaryote abundance, and total bacteria abundance). To simplify these analyses and to give an overview of general trends, the halocarbons were assigned to three

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groups based on their common biological production pathways: (1) I-monohalocarbons (CH_3I , $\text{C}_2\text{H}_5\text{I}$, $2\text{-C}_3\text{H}_7\text{I}$, $1\text{-C}_3\text{H}_7\text{I}$), potentially formed via methyl transferase activity, (2) I-polyhalocarbons (CH_2I_2 , CH_2ClI), potentially formed via iodoperoxidase activity, (3) Br-polyhalocarbons (CHBr_3 , CH_2Br_2 , CH_2BrCl , CHBr_2Cl , CH_2BrI), potentially formed via bromoperoxidase activity (compare Fig. 3). I-monohalocarbons showed the strongest correlations with biological parameters during PI (Fig. 3a–d). Significant positive correlations were identified with both chl *a* and nanophytoplankton (Fig. 3a, c, whilst significant negative correlations were observed with picoeukaryotes and total bacteria (Fig. 3b, d). No significant correlations were observed during PII and PIII. PI also revealed a number of strong relationships between I-polyhalocarbons and biological parameters (Fig. 3e–h), although the trends were consistently of an opposite nature to I-monohalocarbons. Significant negative correlations were identified with both chl *a* concentrations and nanoeukaryote abundance (Fig. 3e, g), and significant positive correlations with picoeukaryotes and total bacteria (Fig. 5f, h). No significant correlations were seen in PII. In PIII, significant positive correlations were found with chl *a* and total bacteria, and significant negative correlations were found with picoeukaryotes. No significant correlations were identified for Br-polyhalocarbons during PI and PII (Fig. 3i–l). During PIII, chl *a* and total bacteria gave significant negative correlations (Fig. 3i, l), whilst picoeukaryotes showed a significant positive relationship (Fig. 3j).

3.3 Halocarbons and $p\text{CO}_2$

In order to determine the effect of $p\text{CO}_2$ on concentrations of halocarbons, the strength of the correlation between mean concentrations and $p\text{CO}_2$ for each experimental phase was examined. A significant increase in both mean and cumulative concentrations of CH_2I_2 under increasing CO_2 was seen in PIII ($R = 0.67$, $F = 5.75$, $p < 0.05$) (Fig. 4a). No further relationships were identified between the standing stocks of halocarbons and $p\text{CO}_2$.

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Absolute concentrations of halocarbons measured in the mesocosms were comparable to two earlier studies that reported halocarbons from Kongsfjorden, and the data is summarised in Table 6 (Hughes, 2004; Schall and Heumann, 1993). Schall and Heumann (1993) (hereafter SH93) analysed seawater samples collected 1 km from the shore during September – a comparable location to the mesocosms, during a similar season. Mean concentrations of CH_3I show strong similarity, although a greater range was observed in the mesocosms, perhaps a result of the nutrient-induced phytoplankton growth. Concentrations of the remaining compounds were generally lower in the mesocosms than those measured by SH93. Similarly, mean concentrations were consistently higher in the fjord compared to the mesocosms during this study, with the greatest difference in mean concentrations seen for CH_2I_2 (78 %) and the least difference for CH_2BrCl (28 %). Whilst differences in halocarbon concentrations between the fjord and mesocosms may be a product of the temporal progression of their respective microbial communities, variations in light regimes and exclusion of benthic processes may have contributed to the variations. For instance, almost minimal ultraviolet (UV) light (< 380 nm) was transmitted through the mesocosm foil (Matthias Fischer, personal communication), and furthermore, potential macroalgal sources of halocarbons were excluded from the mesocosms.

4.1 Processes controlling halocarbon concentrations in the mesocosms

During this experiment, 11 individual halocarbon compounds were quantified, along with numerous other biological and chemical parameters. Attempts to discuss each halocarbon individually would lead to an extensive and complicated discussion. Therefore in order to rationalise the following section, the discussion will focus on one halocarbon from each of the groups detailed in Sect. 3.2, on the assumption that the remaining halocarbons of each group are subject to similar production and removal mechanisms: (1) CH_3I (I-monohalogenated), (2) CH_2I_2 (I-polyhalogenated) and

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(3) CHBr_3 (Br-polyhalogenated). These halocarbons are either the dominant gas from each group in terms of concentrations and/or are the most important in terms of their influence on atmospheric chemistry.

4.1.1 Iodomethane (CH_3I)

5 The temporal dynamics of CH_3I were characterised by periods of both net loss and net production, resulting in concentrations that ranged between below D.L. ($< 1 \text{ pmol l}^{-1}$) and $\sim 10 \text{ pmol l}^{-1}$, suggesting active turnover of this compound within the mesocosms (Fig. 2a). Numerous strong relationships to biological parameters were identified, predominantly during PI (Table 2). CH_3I concentrations gave significant positive correlations with chl *a*, nanoeukaryotes, and phytoplankton pigment concentrations (fucoxanthin, chl C1/C2, peridinin), whilst CH_3I was inversely correlated with picoeukaryotes and total bacterial abundances. Yet, despite the apparent close association with biological activity and the strong CO_2 effect on a number of biological parameters (see Brussaard et al., 2012; Schulz et al., 2012; this issue), no consistent or prolonged response to $p\text{CO}_2$ was seen in the concentrations of CH_3I .

15 In order to speculate on the lack of response of CH_3I concentrations to CO_2 , the processes controlling the production and removal of CH_3I in seawater must first be explained. Direct biological production is thought to occur via methyl transferase enzyme activity by both phytoplankton and bacteria (Amachi et al., 2001). The strong correlations with a number of biological parameters in the mesocosms provide evidence for this source. In addition, production is possible from the breakdown of higher molecular weight iodine-containing organic matter (Fenical, 1982) and through photochemical reactions between organic matter and light (Richter and Wallace, 2004), both of which may have made some contribution to the production of CH_3I in the mesocosm. In terms of removal, CH_3I undergoes nucleophilic substitution and hydrolysis in seawater (Elliott and Rowland, 1993), although the rates of reaction are minimal at the water temperatures experienced during the experiment ($0.1\text{--}0.3 \text{ \%d}^{-1}$) so it is likely that these processes made a negligible contribution to the overall loss of CH_3I . It is also probable

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that some CH_3I undergoes consumption by bacteria, and results of laboratory incubations with ^{13}C -labelled CH_3I have provided evidence of significant “biological” loss rates (Hopkins, personal communication). Seawater CH_3I is also lost via the sea-to-air flux, and this comprised a relatively small component of the total loss during this experiment. For example, during PI the mean sea-to-air flux of CH_3I was estimated at $73.1 \text{ pmol m}^{-2} \text{ d}^{-1}$. Therefore, when scaled to allow comparison with the total net loss, assuming a 12 m deep mixed water column, this flux represents $8 \text{ fmol l}^{-1} \text{ d}^{-1}$, equivalent to $< 4\%$ of the total ($0.25 \text{ pmol l}^{-1} \text{ d}^{-1}$).

Clearly, the controls on seawater concentrations of CH_3I are varied and complex. Furthermore, halocarbons occur at such low levels in seawater (picomolar) that distinguishing the underlying processes from bulk measurements is very difficult. The strongest relationships between CH_3I and biological activity were seen during PI, a period when the biological response to $p\text{CO}_2$ was minimal (See Schulz et al., 2012, this issue). Over the course of PII and PIII, the coupling between CH_3I concentrations and biological parameters such as chl *a* lessened, suggesting a decrease in the importance of direct biological production and a rise in the importance of other production processes. Consequently, a CO_2 effect on CH_3I of the kind seen on biological parameters during PII and PIII was not detectable.

4.1.2 Diiodomethane (CH_2I_2)

The main loss pathway for CH_2I_2 in seawater is photolysis at near-ultraviolet (UV) wavelengths (300–350 nm) (Martino et al., 2006). However, it is likely that this process was negligible in the mesocosms due to lack of UV transmission through the foil (Matthias Fischer, personal communication), and the solar zenith angle experienced at Ny-Ålesund in June (55–75°) would result in little UV entering the mesocosms from above. The lack of photolysis may have facilitated the gradual increase in CH_2I_2 concentrations over the course of the experiment (Fig. 2d). Furthermore, CH_2I_2 was the only halocarbon to show a significant positive response to $p\text{CO}_2$, chiefly during PIII

(Fig. 4a). The temporal data underwent linear regression analysis to reveal significant net production rates ($\text{pmol l}^{-1} \text{d}^{-1}$) in all mesocosms (Table 3). Rates ranged from $0.027 \text{ pmol l}^{-1} \text{d}^{-1}$ in M3 to $0.039 \text{ pmol l}^{-1} \text{d}^{-1}$ in M9. Next, net production rates for each mesocosm underwent correlative analysis with the associated mean $p\text{CO}_2$, revealing significant positive correlations for both the whole experiment (dashed line symbols, $R = 0.79$, $p < 0.05$) and for PIII (solid line, $R = 0.72$, $p < 0.05$) (Fig. 4b).

Furthermore, concentrations of CH_2I_2 were strongly, and often significantly, correlated with a number of biological parameters. Shown in Table 4, CH_2I_2 was closely correlated with both chl *a* and total bacteria for the whole experiment, whilst close relationships with the phytoplankton pigments fucoxanthin and peridinin were observed during PIII. The ratios of CH_2I_2 to a number of biological parameters were found to be strongly correlated with $p\text{CO}_2$ (Fig. 5). The ratio of CH_2I_2 to bacteria cell numbers ($\times 10^{-9} \text{ pmol cell}^{-1}$) significantly increased with increasing CO_2 for the entire experiment (Fig. 5a) whilst ratios of CH_2I_2 to the phytoplankton pigments (pmol ng^{-1} , fucoxanthin and chl C1/C2) showed similarly strong and significant trends during PIII (Fig. 5b, c). In summary, strong relationships between CH_2I_2 and the biological communities of the mesocosms were observed, in particular with total bacterial abundances. In addition, there appeared to be an increase in net production of CH_2I_2 in response to increasing $p\text{CO}_2$. The possible reasons for this will be explored in the following section.

The production of I-polyhalocarbons (CH_2I_2 , CH_2ClI) can be the result of iodoperoxidase enzyme activity that catalyses the destruction of H_2O_2 and stimulates iodination reactions to form polyhalogenated products (Moore et al., 1996; Leblanc et al., 2006). The exact reason for algal-mediated production of volatile halocarbons is not fully understood, although theories exist as to the function of this process (Manley, 2002; Leblanc et al., 2006). As the consequence of haloperoxidase activity is to reduce H_2O_2 concentrations, it provides an antioxidant function. Therefore, the up-regulation of CH_2I_2 production in response to increasing $p\text{CO}_2$ seen here may be indicative of an adaptive response due to perturbed cell physiology amongst the plankton community.

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The strong significant negative correlations between CH_2I_2 and total bacterial abundances over the duration of the experiment are intriguing and suggest some bacterial involvement in the turnover of this compound. There are no reported studies of the biological consumption of CH_2I_2 . However, there is direct evidence for bacterial consumption of CH_2Br_2 (Goodwin et al., 1997, 1998), so this process cannot be ruled out for CH_2I_2 . A small number of studies have investigated the involvement of bacteria in the production of I-polyhalocarbons, yielding somewhat limited and speculative information. Strains of iodine-oxidising bacteria (IOB) have been isolated from seawater, implicating species closely related to the marine bacterium *Roseovarius tolerans* (aerobic bacteriochlorophyll *a*-producer) (Fuse et al., 2003; Amachi et al., 2008). During laboratory enrichment incubations, IOB directly produced free iodine (I_2) which led to the production of abundant organic iodine species, specifically CH_2I_2 , $\text{CH}_2\text{I}\text{I}$ and CHI_3 , via an extracellular oxidase enzyme. Although enrichment incubations are far removed from processes occurring in natural seawater, Amachi (2008) speculates that IOB may be widely distributed in the marine environment, raising the possibility that given the right conditions, IOB could significantly contribute to the production of CH_2I_2 in the marine environment. The significant negative correlations between bacterial abundance and CH_2I_2 concentration as well as increasing ratios of CH_2I_2 per bacteria cell with increasing $p\text{CO}_2$ suggest either: (1) an increase in bacterial production of CH_2I_2 , or (2) a decrease in bacterial consumption of CH_2I_2 , in response to increasing $p\text{CO}_2$.

Due to its high reactivity and short photolytic lifetime, CH_2I_2 is potentially one of the most important sources of iodine atoms to the coastal marine boundary layer (Carpenter et al., 1999). Thus, changes to the sea-to-air flux of this compound could have implications for the catalytic destruction of tropospheric ozone (Chameides and Davis, 1980) and for new particle formation (O'Dowd et al., 2002). Mean fluxes ranged from -0.02 (M6) to $4.1 \text{ pmol m}^{-2} \text{ d}^{-1}$ (M4) during PI (overall mean $1.06 \text{ pmol m}^{-2} \text{ d}^{-1}$), and 7.1 (M6) to $34.4 \text{ pmol m}^{-2} \text{ d}^{-1}$ (M1) in PII (overall mean $12.3 \text{ pmol m}^{-2} \text{ d}^{-1}$). During PIII, an increase in flux was seen in all mesocosms, with an overall mean of $30.3 \text{ pmol m}^{-2} \text{ d}^{-1}$, and a range of 24.8 (M7) to 37.3 (M9) $\text{pmol m}^{-2} \text{ d}^{-1}$. There are

a number of weaknesses in the calculation of the flux – not least the atmospheric values (Schall and Heumann, 1993), so conclusions should be drawn with caution. In PI and PII, no significant differences in flux were detected between mesocosms. Figure 4c shows the estimated mean cumulative fluxes for PIII plotted as a function of ρCO_2 , showing a significant relationship ($p = 0.04$) with increasing ρCO_2 .

4.1.3 Bromoform (CHBr_3)

CHBr_3 is the most abundant form of volatile organic bromine in seawater (Carpenter and Liss, 2000; Quack et al., 2007; Hughes et al., 2009), and predictably dominated the concentrations of bromocarbons in the mesocosms (Fig. 2f–j). No relationship between CHBr_3 concentrations and ρCO_2 treatment was observed, and there was a high degree of similarity in concentrations in the majority of mesocosms.

A key feature of the CHBr_3 data were the consistently higher concentrations observed in M1, most apparent from t_{12} to t_{19} , and from t_{20} to t_{27} (Fig. 2f). The elevated concentrations occurred immediately after a period of rapid net production in all mesocosms. Significant net production rates were detected in M1 for the periods t_{10} – t_{13} ($22.3 \pm 4.1 \text{ pmold}^{-1}$, $p = 0.03$) and t_{19} – t_{21} ($33.0 \pm 1.9 \text{ pmold}^{-1}$, $p = 0.04$), significantly higher than the net production rates of the remaining mesocosms. These periods of net production were immediately followed by net loss over t_{13} – t_{16} , and t_{21} – t_{27} in all mesocosms, during which M1 displayed the greatest rates of net loss ($12.6 \pm 1.8 \text{ pmold}^{-1}$, $p = 0.02$ and $8.0 \pm 1.0 \text{ pmold}^{-1}$, $p = 0.001$, respectively). This suggests enhanced production and turnover of CHBr_3 in M1 relative to the other mesocosms.

During the final phase of the experiment significant net loss rates were observed in all mesocosms, ranging from the maximum in M1 (see above) to a minimum of $2.9 \pm 0.3 \text{ pmold}^{-1}$ in M8 over t_{21} – t_{27} (Table 5). There was found to be a no relationship between the net loss rates and ρCO_2 for this period of the experiment (Fig. 6a). This period of net loss coincided with strong negative correlations between CHBr_3 and total bacteria, with correlation coefficients (R) ranging from 0.78 to 0.98 (Table 5). Excluding

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et al., 2012, this issue). A similar affect can be seen over the period t_{19} – t_{21} . Net production rates of CHBr_3 ranged from $0.6 \text{ pmol l}^{-1} \text{ d}^{-1}$ (M3) to $33.0 \text{ pmol l}^{-1} \text{ d}^{-1}$ in M1, with those in M1 significantly higher. Again, the higher net production rates of CHBr_3 in M1 were concomitant with significantly lower bacterial biomass production on both t_{20} (M1 = $23.62 \text{ ng Cl}^{-1} \text{ h}^{-1}$, mean of all mesocosms = $87.16 \text{ ng Cl}^{-1} \text{ h}^{-1}$) and t_{22} (M1 = $75.97 \text{ ng Cl}^{-1} \text{ h}^{-1}$, mean of all mesocosms = $169.52 \text{ ng Cl}^{-1} \text{ h}^{-1}$). Lower bacterial biomass production may be indicative of a decrease in bacterial activity, leading to lower consumption rates and higher net production rates of CHBr_3 . These differences may have contributed to the range of concentrations of CHBr_3 measured on t_{21} . Similar patterns can be seen in the CH_2Br_2 and CH_2BrCl data (detail not shown here), confirming similar production and consumption pathways of these polybromocarbons, but apparently unaffected by the altered $p\text{CO}_2$ conditions.

4.2 Comparison to a previous mesocosm experiment

Concentrations of a variety of halocarbons from a CO_2 enrichment experiment performed in temperate, coastal waters off Bergen, Norway in 2006 were reported by Hopkins et al. (2010). During the 2006 experiment, maximum chl *a* concentrations of 6 – $11 \mu\text{g l}^{-1}$ were more than double of those measured in this study, and the plankton community showed a strong response to CO_2 , with significant decreases in chl *a* and microbial plankton under high CO_2 . Nevertheless, both the concentrations and the general response of the bromocarbons to biological activity and $p\text{CO}_2$ showed some similarity to the present study conducted in Arctic waters. In contrast, concentrations of iodocarbons were markedly higher during the 2006 experiment, particularly for CH_2I_2 and CH_2ClI with maximum concentrations of ~ 700 and $\sim 600 \text{ pmol l}^{-1}$, respectively. Furthermore, large, and in some cases significant, reductions in concentrations of all iodocarbons occurred at higher $p\text{CO}_2$ (CH_3I : -44% , $\text{C}_2\text{H}_5\text{I}$: -35% , CH_2I_2 : -27% , CH_2ClI : -24%). The temporal dynamics of the iodocarbons suggested a close association with the plankton communities. The lower biomass and relatively lower biological

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activity observed in this Arctic experiment may have suppressed a clear response in the iodocarbon concentrations to increasing CO₂ of the kind seen in the 2006 experiment.

5 Conclusions

Concentrations of a range of halocarbons were measured during a 5-week CO₂-perturbation mesocosm experiment in Kongsfjorden, Spitsbergen, during June and July 2010. The temporal standing stocks of the majority of halocarbons did not significantly respond to $p\text{CO}_2$ over a range from $\sim 175 \mu\text{atm}$ to $\sim 1085 \mu\text{atm}$. Halocarbon concentrations did show a large number of significant correlations with a range of biological parameters suggesting some influence of the biological communities on the production and consumption of these trace gases in Arctic waters. The temporal dynamics of CH₃I, combined with strong correlations with biological parameters, indicated a biological control on concentrations of this gas. However, despite a CO₂ effect on various components of the community, no effect of $p\text{CO}_2$ was seen on CH₃I. CH₂I₂ concentrations were closely related to chl *a* and total bacteria over the whole experiment and with the phytoplankton pigments fucoxanthin and peridinin during PIII, strongly suggesting biological production of this gas. Both the concentrations and the net production of CH₂I₂ showed some sensitivity to $p\text{CO}_2$, with a significant increase in net production rate and sea-to-air flux at higher $p\text{CO}_2$, particularly during the later stages of the experiment. The temporal dynamics of CHBr₃ indicated rapid turnover of this gas, and concentrations varied between mesocosms, although not explainable by $p\text{CO}_2$ treatment. Instead, net loss rates (corrected for loss via gas exchange) displayed a degree of concentration-dependence, and strong negative correlations with bacteria during periods of net loss suggest a degree of bacterial consumption of CHBr₃ in Arctic waters. The results of the first Arctic OA mesocosm experiment provide invaluable information on the production and cycling of halocarbons in Arctic waters, demonstrating strong associations with the biological communities. Although the effects of OA on halocarbons concentrations were in general subtle, some significant affects were observed.

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The role of halocarbons in Arctic atmospheric chemistry may increase in importance in the coming decades due to increases in open water and productivity with the loss of sea ice (Mahajan et al., 2010; Kahru et al., 2011; Stroeve et al., 2011); this work enhances our understanding of the marine production and cycling of halocarbons in a region set to experience rapid environmental change.

Acknowledgements. This work is a contribution to the European Project on Ocean Acidification (EPOCA) which received funding from the European Community's Seventh Framework Programme (FP7/2007–2013) under grant agreement no. 211384. We gratefully acknowledge the logistical support of Greenpeace International for its assistance with the transport of the mesocosm facility from Kiel to Ny-Ålesund and back to Kiel. We also thank the captains and crews of M/V ESPERANZA of Greenpeace and R/V Viking Explorer of the University Centre in Svalbard (UNIS) for assistance during mesocosm transport and during deployment and recovery in Kongsfjorden. We thank Signe Koch Klavsen for providing phytoplankton pigment data and Matthias Fischer for UV measurements through the mesocosm foil. We are grateful to the UK Natural Environmental Research Council for the accommodation and support provided through the NERC-BAS station in Ny-Ålesund. We also thank the staff of the French-German Arctic Research Base at Ny-Ålesund, in particular Marcus Schuhmacher, for on-site logistical support. Financial support was provided through the European Centre for Arctic Environmental Research (ARCFAC) (grant number ARCFAC026129-2009-140) and through Transnational Access funds by the EU project MESOAQUA under grant agreement no. 22822. Finally, we would like to thank Ulf Riebesell, Sebastian Krug and the whole of the Svalbard mesocosm team, who showed great team spirit and comradeship and helped to make the experiment both enjoyable and successful.

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Table 1. Mean $p\text{CO}_2$ (μatm) and pH (on the total scale) for the halocarbon sampling period (t_4 – t_{27}) and for each phase of the experiment referred to in this article. See Bellerby et al. (2012) (this issue) for full details of the evolution of the carbonate system within the mesocosms.

	Halocarbons sampling period		PI		PII		PIII	
	t_4 – t_{27}		t_4 – t_{12}		t_{13} – t_{21}		t_{22} – t_{27}	
	Mean $p\text{CO}_2$ (μatm)	Mean pH (Total)	Mean $p\text{CO}_2$ (μatm)	Mean pH (Total)	Mean $p\text{CO}_2$ (μatm)	Mean pH (Total)	Mean $p\text{CO}_2$ (μatm)	Mean pH (Total)
M3	175	8.34	182	8.33	177	8.33	170	8.35
M7	180	8.33	184	8.32	180	8.33	170	8.35
M2	250	8.21	269	8.18	245	8.20	234	8.24
M4	340	8.09	368	8.06	347	8.08	310	8.13
M8	425	8.01	476	7.96	426	8.00	391	8.04
M1	600	7.87	684	7.81	599	7.87	534	7.92
M6	675	7.82	842	7.73	677	7.82	579	7.89
M5	860	7.72	1064	7.63	853	7.72	747	7.78
M9	1085	7.63	1427	7.51	1062	7.63	891	7.71

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Table 2. Correlation coefficients (*R*) for CH₃I (pmol l⁻¹) and selected biological parameters during PI (*t*₄–*t*₁₂).

	<i>n</i>	CH ₃ I									mean
		M1	M2	M3	M4	M5	M6	M7	M8	M9	
Chlorophyll <i>a</i>	10	0.87*	0.70	0.74*	0.88*	0.94**	0.92*	0.90*	0.90*	0.66	0.93**
Picoeukaryotes	10	0.76*	0.60	0.71*	0.84**	0.82*	0.84**	0.86**	0.83*	0.88**	0.89*
Nano-eukaryotes	10	0.83*	0.75*	0.59	0.83*	0.94**	0.87**	0.81*	0.83*	0.67*	0.85*
Total bacteria	9	0.85*	0.95**	0.75*	0.85*	0.94**	0.90*	0.91*	0.94**	0.85*	0.92*
Fucoxanthin	6	0.99**	0.91*	0.82	0.73	0.89*	0.86*	0.69	0.73	0.25	0.81*
Peridinin	6	0.40	0.82*	0.53	0.90*	0.75	0.68	0.55	0.81*	0.38	0.89*

Asterisks indicate associated probability: * *p* < 0.05, ** *p* < 0.01.

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Table 3. CH₂I₂ net production rates and coefficient of determination (R^2) of the linear regression for period t_4 – t_{27} .

	CH ₂ I ₂ net production rate (pmol l ⁻¹ d ⁻¹) (± SE)	R^2
M1	0.037 (± 0.004)	0.82**
M2	0.033 (± 0.003)	0.89**
M3	0.027 (± 0.003)	0.83**
M4	0.030 (± 0.003)	0.84**
M5	0.035 (± 0.004)	0.83**
M6	0.029 (± 0.003)	0.84**
M7	0.025 (± 0.002)	0.87**
M8	0.028 (± 0.003)	0.78**
M9	0.039 (± 0.005)	0.77**

Asterisks indicate associated probability:

* = $p < 0.05$, ** = $p < 0.01$.

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Table 4. Correlation coefficients (*R*) for relationship between CH₂I₂ (pmol l⁻¹) and chlorophyll *a*, total bacterial abundance, and phytoplankton pigments (fucoxanthin and peridinin).

	<i>n</i>	CH ₂ I ₂									mean
		M1	M2	M3	M4	M5	M6	M7	M8	M9	
Chlorophyll <i>a</i> ¹	24	0.92*	0.97**	0.98*	0.85*	0.75	0.77	0.91*	0.88*	0.66	0.77**
Total bacteria ¹	24	0.85**	0.94**	0.91**	0.51	0.96**	0.96**	0.93**	0.91**	0.94**	0.95**
Fucoxanthin ²	6	0.97**	0.92*	0.96*	0.83*	0.92*	0.98**	0.83*	0.91*	0.83*	0.71*
Peridinin ²	6	0.61	0.80*	0.20	0.22	0.96**	0.71*	0.88*	0.77*	0.90**	0.98**

Asterisks indicate associated probability: **p* < 0.05, ***p* < 0.01. ¹ = PI–PIII (*t*_{4–*t*₂₇}), ² = PIII (*t*_{22–*t*₃₀}).

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Table 5. Correlation coefficients (R) for relationship between CHBr_3 and total bacteria, CHBr_3 net loss rates, and mean $p\text{CO}_2$ for period t_{21} – t_{27} .

t_{21} – t_{27}	R [CHBr_3] and total bacteria	CHBr_3 net loss rate ($\text{pmol l}^{-1} \text{d}^{-1}$) (\pm SE)	Mean $p\text{CO}_2$ (μatm)
M1	0.92*	8.02 (\pm 0.99)**	535.8
M2	0.98**	4.72 (\pm 0.89)**	234.6
M3	0.87*	2.26 (\pm 0.80)*	169.8
M4	0.91*	5.10 (\pm 0.95)**	311.4
M5	0.78*	3.95 (\pm 1.24)*	748.4
M6	0.92**	3.24 (\pm 0.57)**	580.2
M7	0.88**	4.31 (\pm 1.06)**	170.3
M8	0.98**	2.85 (\pm 0.33)**	390.5
M9	0.79*	5.56 (\pm 1.62)*	891.1

Asterisks indicate associated probability of rates, * = $p < 0.05$, ** = $p < 0.01$.

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Table 6. Seawater concentrations of halocarbons in Kongsfjorden reported by Schall and Heumann (1993), Hughes (2004) and measured during this study.

pmol ⁻¹	Schall and Heumann (1993)		Hughes (2004)		This study (fjord)		This study (mesocosms)	
	Mean	Range	Mean	Range	Mean	Range	Mean	Range
CH ₃ I	2.3	0.6–5.4	–	0.5–1.6	4.9	2.5–18.4	2.6	0.04–10.3
C ₂ H ₅ I	–	–	–	0.05–0.4	1.3	0.9–2.9	0.9	0.1–3.3
2-C ₃ H ₇ I	2.6	0.4–5.5	–	–	0.4	0.3–0.7	0.2	0.2–0.4
1-C ₃ H ₇ I	2.5	0.4–9.4	–	–	0.3	0.1–0.3	0.1	0.06–0.4
CH ₂ I ₂	6.2	0.9–12.7	–	–	0.9	0.0–2.7	0.2	0.01–2.5
CH ₂ ClI	1.8	0.9–2.6	–	< D.L.	1.4	0.7–2.5	0.8	0.3–1.6
CHBr ₃	77.0	34.4–157.7	–	–	207.0	122.9–358.1	84.0	35.3–151.5
CH ₂ Br ₂	15.8	7.2–30.1	–	–	23.9	14.0–44.6	12.7	6.3–33.3
CH ₂ BrCl	1.5	0.5–3.6	–	–	1.0	0.5–2.7	0.7	0.1–2.0
CHBr ₂ Cl	5.1	2.5–9.5	–	–	6.5	3.7–11.3	2.9	1.6–4.7
CH ₂ BrI	–	–	–	–	1.3	0.3–3.1	0.3	0.0–1.6

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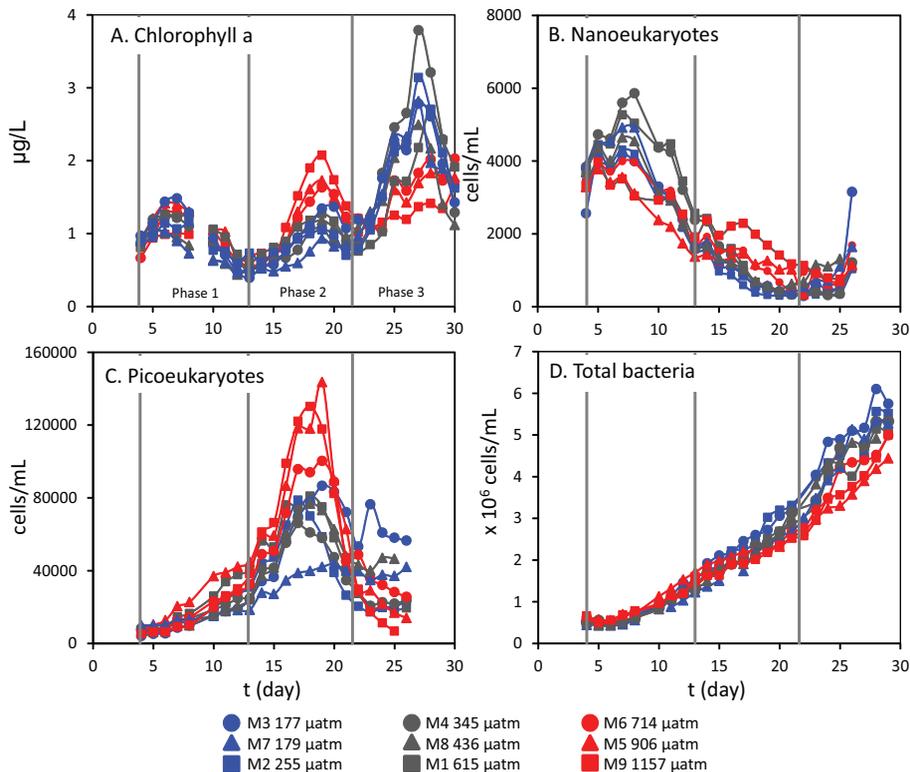


Fig. 1. Concentrations of chlorophyll *a* ($\mu\text{g l}^{-1}$) (A), nanoeukaryote (B) and picoeukaryote abundances (C) (cells ml^{-1}) and total bacteria (D) ($\times 10^6$ cells ml^{-1}) over the course of the experiment. Experimental phases are as follows: PI t_4 – t_{12} , PII Days t_{13} – t_{21} , PIII Days t_{22} – t_{30} . $p\text{CO}_2$ (μatm) shown in the legend are averages for period of halocarbon sampling (t_8 – t_{27}).

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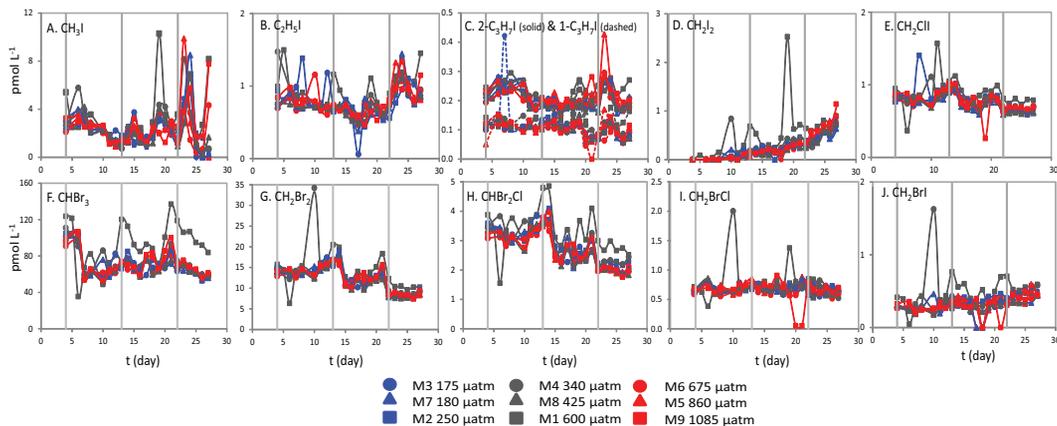


Fig. 2. Concentrations (pmol L^{-1}) of I-monohalocarbons (**A** CH_3I , **B** $\text{C}_2\text{H}_5\text{I}$ and **C** $2\text{-C}_3\text{H}_7\text{I}$ (solid) and $1\text{-C}_3\text{H}_7\text{I}$ (dashed)), I-polyhalocarbons (**D** CH_2I_2 , **E** CH_2ClI) and Br-polyhalocarbons (**F** CHBr_3 , **G** CH_2Br_2 , **H** CHBr_2Cl , **I** CH_2BrCl , **J** CH_2BrI). Grey lines indicate phases of experiment: PI t_4 – t_{12} , PII Days t_{13} – t_{21} , PIII Days t_{22} – t_{30} . $p\text{CO}_2$ (μatm) shown in the legend are averages for period t_8 – t_{27} .

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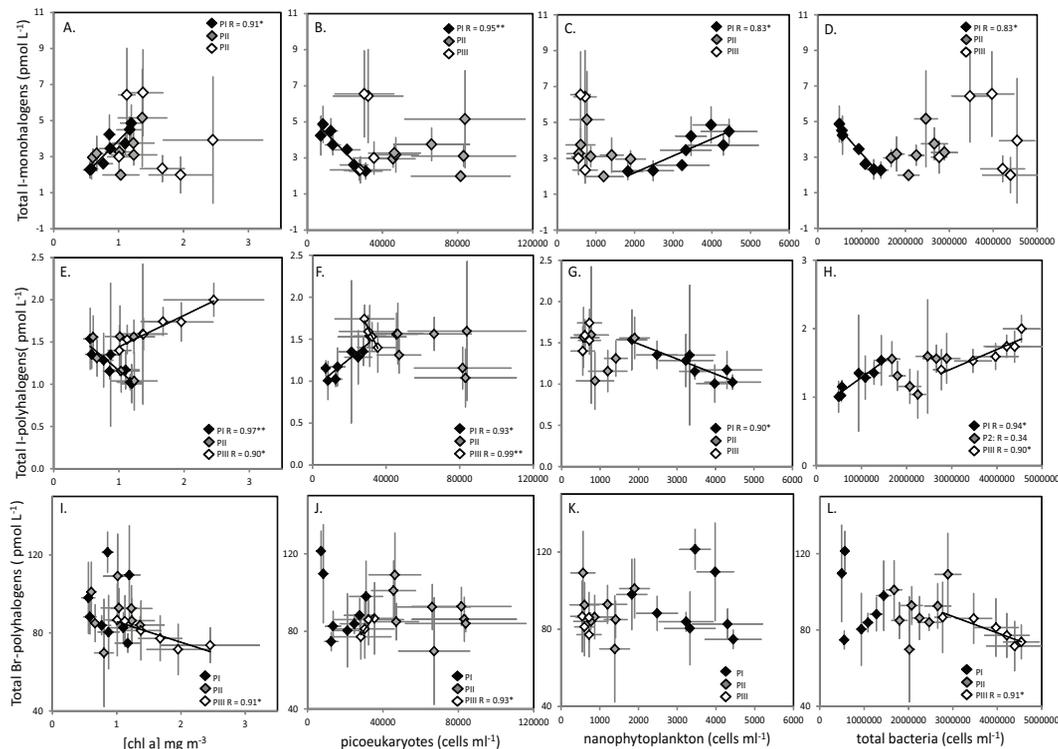


Fig. 3. Relationships between mean halocarbon concentration (pmol l^{-1}) and biological parameters (L to R, chl *a*, picoeukaryotes, nanophytoplankton, bacterial abundance). **(A–D)** total I-monohalocarbons (sum of CH_3I , $\text{C}_2\text{H}_5\text{I}$, $2\text{-C}_3\text{H}_7\text{I}$, $1\text{-C}_3\text{H}_7\text{I}$), **(E–H)** total I-polyhalocarbons (sum of CH_2I_2 and CH_2ClI), **(I–L)** total Br-polyhalocarbons (sum of CHBr_3 , CH_2Br_2 , CHBr_2Cl , CH_2BrCl , CH_2BrI). PI (black diamonds), PII (grey diamonds) and PIII (open diamonds). Error bars indicate range of all data from all mesocosms. Asterisks denote significance level of correlation (R): * = $p < 0.05$, ** = $p < 0.001$.

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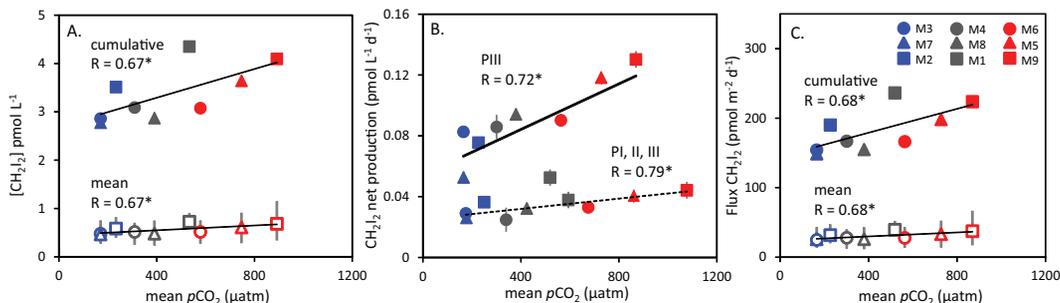


Fig. 4. Relationships in PIII between **(A)** the concentration of CH_2I_2 (pmol l^{-1}) (cumulative: solid symbols, mean: open symbols) and mean $p\text{CO}_2$ (μatm), **(B)** the mean net production rate of CH_2I_2 ($\text{pmol l}^{-1} \text{d}^{-1}$) and mean $p\text{CO}_2$ (μatm) for PI–PIII (dashed line) and PIII only (solid line), and **(C)** the sea-to-air flux of CH_2I_2 ($\text{pmol m}^{-2} \text{d}^{-1}$) (cumulative: solid symbols, mean: open symbols), all for Phase III (t_{22} – t_{27}). Asterisks denote significance level of correlation (R): * = $p < 0.05$, ** = $p < 0.01$. Error bars show range of data.

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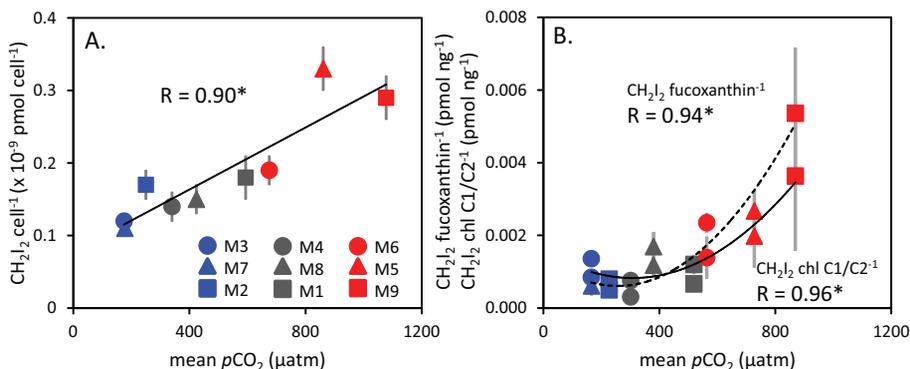


Fig. 5. Ratios of CH_2I_2 to biological parameters plotted as a function of $p\text{CO}_2$ (μatm) **(A)** CH_2I_2 bacteria cell $^{-1}$ ($\times 10^{-9}$ pmol cell $^{-1}$) for Phases I, II and III, **(B)** CH_2I_2 fucoxanthin $^{-1}$ (pmol ng $^{-1}$) (dashed line) and CH_2I_2 chl C1/C2 $^{-1}$ (pmol ng $^{-1}$) (solid line) for Phase III only. Asterisks denote significance level of correlation (R): $*$ = $p < 0.05$, $**$ = $p < 0.01$. Error bars indicate standard error. Legend shown in panel **(A)**.

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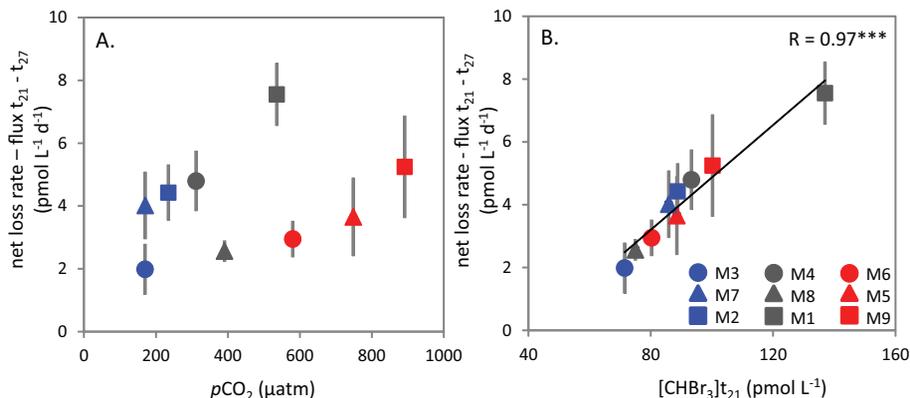


Fig. 6. Relationship between **(A)** net loss rates of CHBr_3 ($\text{pmol l}^{-1} \text{d}^{-1}$) over the period $t_{21}-t_{27}$ and mean $p\text{CO}_2$ (μatm), **(B)** net loss rates of CHBr_3 minus the sea-to-air flux ($\text{pmol l}^{-1} \text{d}^{-1}$) over the period $t_{21}-t_{27}$ and the concentrations of CHBr_3 on t_{21} . Asterisks denote significance level of correlation (R): $^{***} = p < 0.001$.

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