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Key Points:

- Seawater dimethylsulfide concentration in coral reefs is parameterized as a function of surface temperature and irradiance
- Including coral-air dimethylsulfide release at low tide in sea-air flux calculations yields a more accurate estimate of coral reef flux
- Coral reefs emit $0.03\text{--}0.04\text{ Tg yr}^{-1}$ of sulfur as dimethylsulfide, representing up to 0.2% of global sea-air flux estimates

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Parameterizing the Impact of Seawater Temperature and Irradiance on Dimethylsulfide (DMS) in the Great Barrier Reef and the Contribution of Coral Reefs to the Global Sulfur Cycle

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Abstract Biogenic emissions of dimethylsulfide (DMS) are an important source of sulfur to the atmosphere, with implications for aerosol formation and cloud albedo over the ocean. Natural aerosol sources constitute the largest uncertainty in estimates of aerosol radiative forcing and climate and thus, an improved understanding of DMS sources is needed. Coral reefs are strong point sources of DMS; however, this coral source of biogenic sulfur is not explicitly included in climatologies or in model simulations. Consequently, the role of coral reefs in local and regional climate remains uncertain. We aim to improve the representation of tropical coral reefs in DMS databases by calculating a climatology of seawater DMS concentration (DMS_w) and sea-air flux in the Great Barrier Reef (GBR), Australia. DMS_w is calculated from remotely sensed observations of sea surface temperature and photosynthetically active radiation using a multiple linear regression model derived from field observations of DMS_w in the GBR. We estimate that coral reefs and lagoon waters in the GBR ($\sim 347,000\text{ km}^2$) release $0.03\text{--}0.05\text{ Tg yr}^{-1}$ of DMS (0.02 Tg yr^{-1} of sulfur). Based on this estimate, global tropical coral reefs ($\sim 600,000\text{ km}^2$) could emit 0.08 Tg yr^{-1} of DMS (0.04 Tg yr^{-1} of sulfur), with the potential to influence the local radiative balance.

Plain Language Summary Coral reefs are important sources of biogenic sulfur through stress-induced emissions of dimethylsulfide (DMS). This volatile sulfur compound has potentially important implications for atmospheric aerosol formation and cloud microphysical properties. However, coral reefs are not accounted for in global DMS climatologies or in model simulations. Consequently, the importance of coral reef-derived DMS in local climate is uncertain. Here, a proxy for seawater DMS concentration in coral reefs is derived from field measurements taken in the Great Barrier Reef, Australia. This proxy is used to calculate a climatology of DMS flux from coral reefs to estimate the contribution to the global sulfur cycle.

1. Introduction

Aerosols and clouds play a key role in the Earth's radiative budget and climate. The influence of anthropogenic greenhouse gas (GHG) emissions on the radiative balance is relatively well understood and is estimated to have exerted an average global warming effect of 2.3 W m^{-2} over the industrial period (IPCC, 2014). Anthropogenic and natural aerosols offset this warming effect by -0.9 W m^{-2} (-1.9 to -0.1 W m^{-2}) through scattering of short-wave radiation and increasing the albedo, lifetime, and cover of clouds (Andreae, 1995; IPCC, 2014; Twomey, 1974). However, the radiative effects of natural aerosols and their role in climate is complex and less well understood. Approximately 45% of the variance in aerosol radiative forcing is derived

from uncertainties in the role of natural aerosol sources, including marine dimethylsulfide (DMS) (Carslaw et al., 2013).

DMS is a volatile sulfur compound produced in the ocean by marine planktonic and benthic algae (Stefels, 2000), giant clams (Guibert et al., 2020; Hill et al., 2000), coral endosymbiotic dinoflagellates, and corals (Broadbent et al., 2002; Jones et al., 1994; Raina et al., 2013). The marine precursor of DMS, dimethylsulfoniopropionate (DMSP) has a range of hypothesized biological and ecological functions (McParland & Levine, 2019; Stefels, 2000), including a physiological stress response (Sunda et al., 2002). DMSP catabolism is mediated by algal and microbial processes. DMSP cleavage to DMS occurs via DMSP-lyase enzymes present in free-living and coral-associated endosymbiotic algae and microbes (Bullock et al., 2017). The DMSP demethylation pathway is a microbial process by which marine microbes convert DMSP into methanethiol (Bourne et al., 2016; Bullock et al., 2017; Raina et al., 2009). A portion (~10%) of the dissolved DMS (DMS_w) pool is ventilated to the marine boundary layer where it is rapidly oxidized to various sulfate aerosol precursor compounds (Andreae & Crutzen, 1997; Berndt et al., 2019; Veres et al., 2020).

Field observations and model simulations have identified a significant influence of DMS-derived sulfates on aerosol formation and growth, cloud condensation nuclei (CCN), and cloud droplet radius over the ocean (Gabric et al., 2013, 2018; Korhonen et al., 2008; Lana et al., 2012; Sanchez et al., 2018). DMS accounts for 18–43% of the global annual mean nonsea salt sulfate (nss-SO_4) aerosol burden (Gondwe et al., 2003; Kloster et al., 2006), exerting a top-of-atmosphere radiative forcing of -1.79 W m^{-2} (Mahajan et al., 2015) to -2.03 W m^{-2} (Thomas et al., 2010). However, regional variability in the importance of DMS in aerosol and cloud properties is evident, with the greatest sensitivity occurring over remote oceans where the influence of anthropogenic aerosol emissions is minimal (Fiddes et al., 2018; Woodhouse et al., 2013).

The Great Barrier Reef (GBR), Australia, has been recognized as a significant source of DMS, with the potential to influence local climate in NE Australia (Jackson et al., 2020; Jones et al., 2018). *Acropora* spp. are dominant throughout the Indo-Pacific and produce amongst the highest reported individual concentrations of dimethylated sulfur compounds (Broadbent et al., 2002; Jones et al., 1994; Swan et al., 2017a). Early field surveys found that aerosol concentration was significantly higher in the marine atmosphere directly over the GBR, compared to the seaward side (Bigg & Turvey, 1978). More recent field surveys have observed nucleation events occurring over the GBR (Leck & Bigg, 2008), where newly nucleated aerosol particles consisted of ~40% organic matter and ~60% sulfates, which are likely derived from DMS (Modini et al., 2009). Furthermore, there is a strong seasonal increase in fine-mode aerosol loading in spring and summer throughout the GBR, that has been observed from both field (Leck & Bigg, 2008) and satellite observations (Cropp et al., 2018; Jackson et al., 2018). This increase in fine-mode aerosol implies a temperature or irradiance-dependent biogenic influence (Christiansen et al., 2019; Korhonen et al., 2008; Long et al., 2014).

In corals, DMSP biosynthesis and cleavage to DMS is upregulated in response to oxidative stress caused by exposure to high sea surface temperature (SST), irradiance (Deschaseaux et al., 2014b; Fischer & Jones, 2012; Jones et al., 2007; Swan et al., 2017b), and low salinity (Gardner et al., 2016). Oxidative stress is caused by the release of reactive oxygen compounds (ROS) by coral cell mitochondria and zooxanthellae photosystems which can damage cells, tissues, and DNA (Lesser, 2011; Weis, 2008). Irradiance stress can be exacerbated in corals during aerial exposure at low tide (Buckee et al., 2020), with DMS being released directly to the atmosphere and resulting in measured spikes in atmospheric DMS (DMS_a) of up to 45.9 nmol m^{-3} (1,122 ppt) which can persist for around 8 h (Swan et al., 2017b). This is well above typical oceanic emissions of ~4–8 nmol m^{-3} (~100–200 ppt) (Kettle et al., 1999).

The rate of photosynthesis in coral-associated zooxanthellae increases linearly with photosynthetically active radiation (PAR) until photosystems become saturated and a maximum rate is achieved (Anderson et al., 1995; Gorbunov et al., 2001; Winters et al., 2003). For PAR beyond this maximum light threshold, photoinhibition occurs and excess light energy is dissipated as heat via various photoprotective mechanisms (Gorbunov et al., 2001; Melis, 1999). Excess light energy is quantified as excess excitation energy (EEE) and when not all EEE is dissipated as heat, photodamage can occur to Photosystem II (PS II), inhibiting the transport of electrons and damaging protein structure. When the rate of photodamage exceeds that of photoprotective repair, photodamage accumulates, and ROS are released into coral tissues (Lesser, 2011;

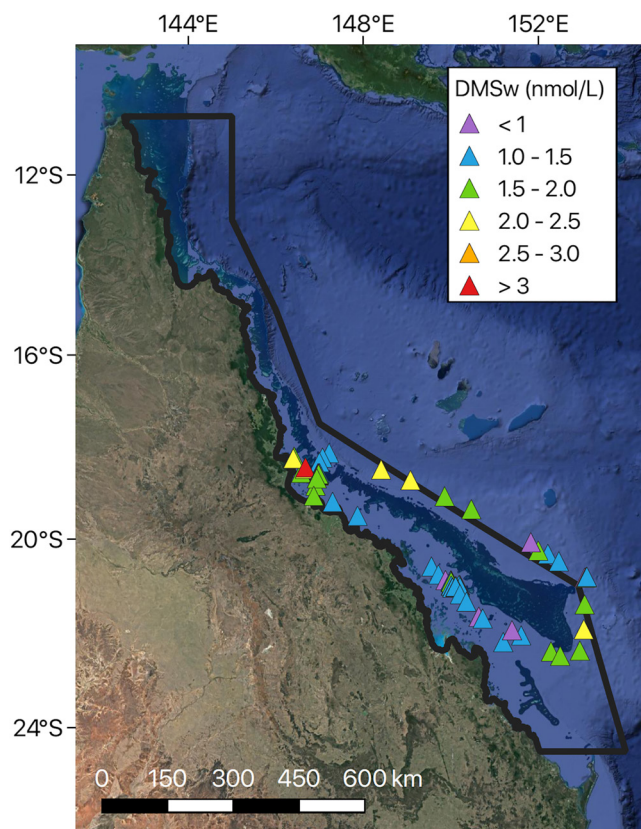


Figure 1. Map of northeast Queensland, Australia, showing the boundary of the Great Barrier Reef Marine Park (black outline) and DMS_w sampling locations and daily mean concentration during RV *Investigator* voyage IN2016_V05. Satellite imagery © Google Earth 2020. DMS_w , dimethylsulfide concentration.

Weis, 2008). If conditions persist, corals will expel their zooxanthellae and become bleached (Downs et al., 2002; Lesser, 2011; Yakovleva et al., 2009). High SST can exacerbate photodamage by further reducing the PAR absorption capacity (Jones et al., 2000, 2002).

$DMS(P)$ readily scavenge ROS, thereby raising the coral oxidative stress threshold (Jones & King, 2015). ROS scavenging by $DMS(P)$ produces dimethyl sulfoxide (DMSO), which can be reduced by bacteria and algae to DMS (Spiese et al., 2009). The concentration of DMS in coral reef waters is therefore dependent on the rate of $DMS(P)$ (O) biosynthesis, which is related to coral oxidative stress (Deschaseaux et al., 2014b; Gardner et al., 2016; Raina et al., 2013). However, when oxidative stress exceeds coral physiological tolerance thresholds, ROS scavenging, and $DMS(P)$ oxidation to DMSO increases, resulting in a decline in ambient DMS concentration (Deschaseaux et al., 2014a; Fischer & Jones, 2012). For example, Fischer and Jones (2012) found that when *Acropora* spp. were exposed to SST $\sim 2^\circ\text{C}$ above ambient temperatures (26°C), and PAR $> 6\text{ mol m}^{-2}\text{ h}^{-1}$, concentrations of DMS_a declined by $\sim 90\%$. Similarly, Jones et al. (2007) demonstrated a decline in DMS_w of $\sim 50\%$ prior to a mass coral bleaching event in the central GBR when SST was $> 30^\circ\text{C}$. The decrease in DMS_w was likely due to enhanced biochemical oxidation of $DMS(P)$ to DMSO and/or a decline in production due to coral bleaching (Fischer & Jones, 2012; Jones et al., 2007). Thus, DMS_w in coral reefs is often closely linked to coral oxidative stress.

Currently, global DMS climatologies do not explicitly account for coral-derived DMS and the local DMS climatology of the GBR is unknown (Kettle & Andreae, 2000; Lana et al., 2011; Land et al., 2014). Although global DMS_w climatologies do incorporate measurements made within or nearby to coral reef lagoons (see Kettle et al., 1999), only a small number of coral reef regions are included and interpolation techniques do not account for seasonally variable coral emissions or direct coral-atmosphere DMS flux during coral exposure to air at low tide. The sensitivity of local climate to coral reef DMS emissions cannot be accurately quantified in earth system models without an improved representation of the source strength of coral reef-derived DMS.

Here, we derive a proxy for sea surface DMS_w in coral reef waters, from field measurements of DMS_w and physical oceanic parameters taken during Marine National Facility RV *Investigator* voyage IN2016_V05 in the GBR. Jones et al. (2018) compiled DMS_w measurements from three decades of surveys in the GBR, including the RV *Investigator* data set, to estimate annual DMS sea-air flux. The same data set is used here to derive a proxy for DMS_w , which is then used to calculate a climatology of DMS_w and sea-air flux from the GBR. The results provide insight into the importance of coral reefs to the atmospheric sulfur budget; an important source which may be lost with ongoing coral bleaching and reef degradation.

2. Materials and Methods

2.1. In-Situ and Remotely Sensed Measurements

DMS_w and total DMSP ($DMSP_t$) concentrations were measured in the southern and central regions of the GBR from September 29 to October 22, 2016 during RV *Investigator* voyage IN2016_V05 (Figure 1). This data set was generously provided by Jones et al. (2018), where detailed sampling and measurement protocols can be found. Seawater samples were taken at subhourly to 5-hourly frequency at 0–5-m depth using the underway seawater system. Samples were either immediately analyzed for DMS_w or preserved with 10% hydrochloric acid for later onshore analysis of $DMSP_t$. DMS_w was measured with a purge and trap gas chromatograph (GC) fitted with a flame photometric detector (GC-FPD) up to October 14, at which point

the equipment was required for Conductivity-Temperature-Depth measurements. From October 14 to 22, DMS_w was measured with an Equilibrator Inlet-Proton Transfer Reaction Mass Spectrometer (EI-PTR-MS) (Kameyama et al., 2009; Omori et al., 2017), which yielded results in agreement with the GC-FPD (Jones et al., 2018). DMSP_t concentration in 100 ml seawater was determined using a GC-mass spectrometer operating in Single Ion Monitoring mode as described in Deschaseaux et al. (2018). DMS_w and DMSP_t measurements had a detection limit of 0.1 pmol (Jones et al., 2018).

Field and laboratory experiments have demonstrated that DMS concentration in corals and reef seawaters is affected by SST, PAR (and water clarity), salinity, and tidal height (Gardner et al., 2016; Jones et al., 2007; Raina et al., 2013). These variables were therefore used as potential predictors of DMS_w in the GBR. Wind-driven mixing and phytoplankton dynamics also influence DMS_w (Broadbent & Jones, 2006; Gabric et al., 2008; Stefels, 2000) and are, respectively, represented through the inclusion of wind speed (WS) and chlorophyll-*a* (CHL) as a proxy for phytoplankton biomass.

The DMS_w (nmol L^{-1}) and DMSP_t (nmol L^{-1}) measurements were provided by Jones et al. (2018). Corresponding underway measurements of SST ($^{\circ}\text{C}$), incident PAR ($\text{mol m}^{-2} \text{h}^{-1}$), sea surface salinity (SSS: psu) and WS (m s^{-1}) were downloaded from the Commonwealth Scientific and Industrial Research Organisation (CSIRO) Marlin online repository (<https://www.marlin.csiro.au>). Diffuse attenuation coefficient (k_{490} : m^{-1}) as a proxy for water clarity, and CHL (mg m^{-3}) were not available from the in-situ GBR data set. Daily mean k_{490} and CHL were obtained from the Moderate Resolution Imaging Spectroradiometer (MODIS) sensor aboard the Aqua and Terra satellites. Daily total PAR ($\text{mol m}^{-2} \text{d}^{-1}$) was also downloaded at 0.04-degree (~ 4 km) resolution for each survey location from NASA OceanColor (<https://oceancolor.gsfc.nasa.gov>). To account for light attenuation at the sea surface, MODIS total PAR was reduced by the corresponding k_{490} value for a depth of 5 m, which corresponds to the sampling depth of 0–5 m. Although many coral reefs reside at depths greater than 5 m, coral reef seawaters are generally well mixed due to tidal and wind-driven mixing (Davis et al., 2020; Middleton et al., 1994). Therefore, we assume that DMS_w concentrations in the upper 5 m are representative of the water column and are most important for calculating the DMS sea-air flux.

Tidal height was also provided with the DMS_w data set for 8 of the 24 survey days when the RV *Investigator* was close to the Slashers Reefs in the central GBR (18.5°S , 147°E). Jones et al. (2018) separated tidal height and corresponding DMS_w into rising and falling tides and analyzed each data set separately. Significant positive correlations were identified between DMS_w and both rising and falling tide height. These correlations reflected an increase in DMS_w as tides fell, followed by a second increase in DMS_w as tides rose (Jones et al., 2018), due to dissolution of DMS-rich coral mucous in tidal slack water (Hopkins et al., 2016). For this work, tidal data were not separated into rising and falling tides and consequently, there was no significant correlation between DMS_w and tidal height ($p > 0.05$, $n = 133$). Tidal height is more important for measurements of DMS_w in seawater directly over the shallow coral reef flat (Jones et al., 2018). Given that the RV *Investigator* sampled deeper coral reef lagoon waters (waters surrounding the coral reef platforms), tidal height was not included as a potential predictor of DMS_w in the GBR.

2.2. Correlation and Multiple Linear Regression

To identify the strongest predictors of DMS_w in the GBR, hourly ($n = 184$) and daily ($n = 24$) mean DMS_w was correlated with each variable. Graphical analysis determined DMS_w ($1.4 \pm 0.07 \text{ nmol L}^{-1}$), DMSP_t ($8.5 \pm 0.6 \text{ nmol L}^{-1}$), total daily PAR at 5 m ($49.2 \pm 0.2 \text{ mol m}^{-2} \text{d}^{-1}$) and WS ($6.6 \pm 0.4 \text{ m s}^{-1}$) to be approximately normally distributed. Values in parentheses represent the mean \pm two standard errors (SE). SST was left skewed due to the dominance of values above the mean ($25.9 \pm 0.09 \text{ }^{\circ}\text{C}$). Incident PAR was right skewed due to the inclusion of nighttime data ($2.7 \pm 0.4 \text{ mol m}^{-2} \text{d}^{-1}$), as was SSS (35.5 ± 0.02 psu) and CHL ($0.19 \pm 0.01 \text{ mg m}^{-3}$) due to the dominance of values below the mean. A conservative significance level ($\alpha = 0.01$) was used to account for nonnormality and outliers which are common in environmental data. A multiple linear regression was then derived between DMS_w and the strongest predictor variables.

2.3. Estimating DMS_w in the GBR

A climatology (2001–2019) of DMS_w was calculated from satellite observations in the GBR using the multiple linear regression derived from the RV *Investigator* IN2016_V05 data set. Daily data for each predictor variable were downloaded from MODIS Aqua and Terra at 0.04-degree resolution for the GBR (9.5°S–25°S, 142°E–154°E) from NASA OceanColor (<https://oceancolor.gsfc.nasa.gov>). Given the high number of missing observations in the high-resolution MODIS data set, a 30-days moving average of each variable was used. All data beyond the Great Barrier Reef Marine Park (GBRMP) boundary defined in Figure 1 were excluded.

2.4. DMS Sea-Air Flux

To calculate DMS sea-air flux, instantaneous wind speed at 10 m (U_{10} : m s⁻¹) was derived every 3 h from ERA5 0.25-degree u and v -wind components (Copernicus Climate Change Service, 2019). Calculated DMS_w and MODIS SST were regridded to a 0.25-degree resolution. DMS sea-air flux was then calculated from daily DMS_w, SST, and each U_{10} measurement. U_{10} was derived every 3 h to avoid underestimating DMS flux calculated from daily mean wind speed.

Sea-air gas exchange is calculated as the product of the total gas transfer velocity (K_w) and the concentration difference between the sea surface (C_w) and atmosphere (C_a) as follows: Flux = $K_w C_w - C_a \alpha$, where α is the solubility of DMS (Liss & Slater, 1974). DMS concentration is typically orders of magnitude lower in the atmosphere than at the sea surface and so, C_a is assumed to be zero. The resulting simplified equation (Equation 1) calculates DMS sea-air flux ($\mu\text{mol m}^{-2} \text{d}^{-1}$) from K_w (cm h⁻¹) and C_w (nmol L⁻¹). This common assumption can lead to a global flux overestimation of up to 5% (Johnson et al., 2011)

$$\text{Flux} = K_w C_w \quad (1)$$

For this analysis, three parameterizations for K_w are used to provide a range of DMS flux estimates. The parameterizations of Goddijn-Murphy et al. (2012) and Nightingale et al. (2000) specify that the DMS transfer velocity increases linearly with U_{10} and provide a relatively high estimate of DMS sea-air flux. These two parameterizations were derived for carbon dioxide and are normalized to a Schmidt number of 660 for DMS (Sc_{DMS}). Goddijn-Murphy et al. (2012) assume that K_w is equivalent to the water-side transfer velocity (k_w) (Equation 2), which is normalized to the SST-dependent Sc_{DMS} calculated as follows: $Sc_{DMS} = 2,674 - 147.12 \text{ SST} + 3.726 \text{ SST}^2 - 0.038 \text{ SST}^3$ (Saltzman et al., 1993)

$$K_{w,660} = k_{w,660} = (2.1 U_{10} - 2.8) (Sc_{DMS} / 600)^{-0.5} \quad (2)$$

For the parameterization of Nightingale et al. (2000), K_w is calculated as a function of both the water-side and air-side DMS transfer velocities. For this parameterization, k_w is also linearly related to U_{10} (Equation 3) and is normalized to the SST-dependent Sc_{DMS} (Saltzman et al., 1993)

$$k_{w,660} = (0.222 U_{10}^2 + 0.333 U_{10}) (Sc_{DMS} / 600)^{-0.5} \quad (3)$$

Total DMS transfer velocity is then calculated using Equation 4 (McGillis et al., 2000; Nightingale et al., 2000). The atmospheric gradient fraction (γ_a) is defined by $\gamma_a = 1 / (1 + k_a / \alpha k_{w,660})$ (McGillis et al., 2000), where α is the solubility coefficient for DMS (11.4 at 26 °C) and k_a is the air-side transfer velocity calculated as a function of U_{10} and the molecular weight of DMS and water as follows: $k_a = 659 U_{10} (62.13 / 18.02)^{-0.5}$ (Kondo, 1975)

$$K_{w,660} = k_{w,660} (1 - \gamma_a) \quad (4)$$

The third K_w parameterization suggests that the DMS gas transfer velocity decreases for wind speeds exceeding 10 m s⁻¹ and provides a more conservative estimate of DMS sea-air flux (Vlahos & Monahan, 2009). The divergence of the DMS transfer velocity at high wind speeds (>10 m s⁻¹) is accounted for by including an attenuation of the Henry's Law constant (Vlahos & Monahan, 2009). For this parameterization, K_w

is calculated using Equation 5 as a function of both $k_w = 4 \times 10^{-4} + 4 \times 10^{-5} (U_{10})^2$ and $k_a = 0.2 U_{10} + 0.3$ (Schwarzenbach et al., 1994; Vlahos & Monahan, 2009). The parameterizations for k_w and k_a are given in units of cm s^{-1} and are converted to cm h^{-1} in order to calculate the total DMS transfer velocity

$$K_{w,660} = \left(\frac{1}{k_w} + \frac{1}{k_a H_{\text{atten}}} \right)^{-1} \quad (5)$$

The attenuated Henry's Law constant for DMS (H_{atten}) is calculated using Equation 6, where H is the Henry's Law constant in seawater (0.089) (Przyjazny et al., 1983), ϕ_B is the surface area of bubbles under the sea surface given by $\phi_B = 0.09 (U_{10} / 10)^3$ and C_{mix} / C_w is the solubility enhancement of DMS, calculated to be 40 from Vlahos and Monahan (2009)

$$H_{\text{atten}} = H / (1 + \phi_B (C_{\text{mix}} / C_w)) \quad (6)$$

During coral exposure to air at low tide, DMS is directly released to the atmosphere from corals and the mucous layer which quickly forms on the coral surface upon exposure (Hopkins et al., 2016). Direct coral-atmosphere DMS exchange can result in large spikes of atmospheric DMS concentration (Hopkins et al., 2016; Jones et al., 2007; Swan et al., 2017b). However, coral aerial exposure is not currently accounted for in DMS sea-air flux calculations. In this study, we attempt to quantify total DMS emissions from the GBR by including a laboratory-based estimate of direct coral-atmosphere DMS flux. Hopkins et al. (2016) measured the rate of DMS_a production from *Acropora horrida* periodically exposed to air. Coral nubbins were obtained from parent colonies sampled in the Indo-Pacific and an estimate of coral aerial exposure at Heron Island in the southern GBR was used (Hopkins et al., 2016), where extreme low tides expose corals for about 2 h on 6 days per month (Wild et al., 2004). Assuming a similar pattern of coral exposure across the GBR, and given that *Acropora* is the dominant coral genus, it was estimated that corals in the GBR exposed to air release 9–35 $\mu\text{mol m}^{-2} \text{d}^{-1}$ (mean 22 $\mu\text{mol m}^{-2} \text{d}^{-1}$). A fraction of the mean flux estimate is added to Equation 1, depending on the percentage cover of coral reefs within each 0.25×0.25 -degree grid. Swan et al. (2017b) observed that corals have the strongest influence on DMS_a levels during winter when the background oceanic flux is relatively low. Adding a fixed estimate of coral-atmosphere DMS release to our flux estimate is consistent with these findings, as the percentage contribution of aerially exposed corals to DMS flux is higher in winter. These intermittent spikes can exceed 40 nmol m^{-3} (1,000 ppt) and persist for up to 8 h during low and rising tides (Swan et al., 2017b). Tidal range is higher in the far northern GBR (Jackson et al., 2018), which could affect the frequency and extent of coral exposure to air at low tide. Assuming direct coral-atmosphere DMS flux for only 2 h on 6 days per month is therefore a reasonable and possibly conservative estimate.

3. Results

3.1. Correlation Analysis

Hourly mean observed DMS_w positively correlated with SST and WS, and negatively with SSS and instantaneous PAR (Table 1). Hourly mean DMSP_t and CHL did not significantly correlate with DMS_w .

Correlation coefficients were also calculated using a daily mean of each variable ($n = 24$); however, no significant correlations occurred ($p > 0.01$). To reduce the variability in this limited data set, we also calculated correlation coefficients for a 3-day moving average of each variable ($n = 24$). This resulted in similar correlations as the hourly mean data set, which increased in strength due to reduced variability (Table 1). At this time scale, instantaneous PAR no longer significantly correlated with DMS_w likely due to averaging out the diel cycle. However, daily total PAR from MODIS observations was strongly positively correlated with DMS_w , reflecting a relative increase in concentration for days when total irradiance was high. Wind-driven mixing contributes to DMS sea-air exchange, and changes in SST, SSS, nutrients, and phytoplankton dynamics, which can explain the positive correlation between hourly mean DMS_w and WS. Three-day moving mean WS did not correlate with DMS_w , yet DMSP_t strongly positively correlated with DMS_w (Table 1).

Table 1
Pearson's Linear Correlation Between Hourly and 3-Day Moving Mean DMS_w and Predictor Variables

3	Hourly mean (<i>n</i> = 184 ^a)		3-days moving mean (<i>n</i> = 24)	
	<i>r</i>	<i>p</i>	<i>r</i>	<i>p</i>
SST	0.29	<0.001	0.56	<0.01
SSS	-0.28	<0.001	-0.62	<0.01
WS	0.23	<0.01	0.39	0.06
Instantaneous PAR	-0.21	<0.01	-0.19	0.37
Total PAR at 5 m ^b	0.19	<0.01	0.57	<0.01
DMSP _t	0.14	0.08	0.63	<0.01
CHL ^b	-0.08	0.26	-0.10	0.65

^aCorresponding WS and DMSP_t were not available for all DMS_w measurements. For hourly mean WS, *n* = 183 and for DMSP_t, *n* = 169. ^bData are derived from remotely sensed observations. Significant (*p* < 0.01) correlations are in bold.

Significant relationships between 3-day moving mean DMS_w and potential predictor variables (SST, SSS, PAR at 5 m, and DMSP_t) are shown in Figure 2.

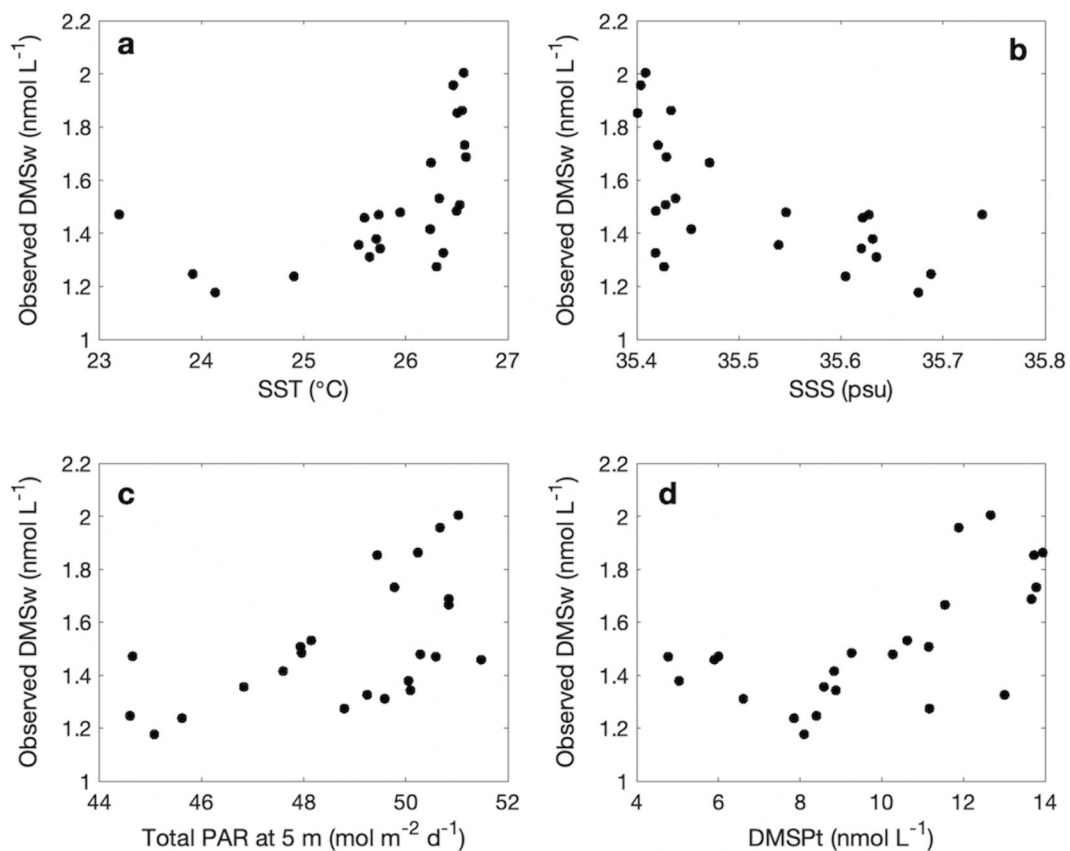


Figure 2. Relationship between 3-day moving mean observed DMS_w and (a) SST (*r* = 0.56), (b) SSS (*r* = -0.62), (c) total PAR at 5 m (*r* = 0.57), and (d) DMSP_t (*r* = 0.63). DMS_w, dimethylsulfide concentration; SST, sea surface temperature; SSS, sea surface salinity; PAR, photosynthetically active radiation; DMSP_t, total dimethylsulfoniopropionate.

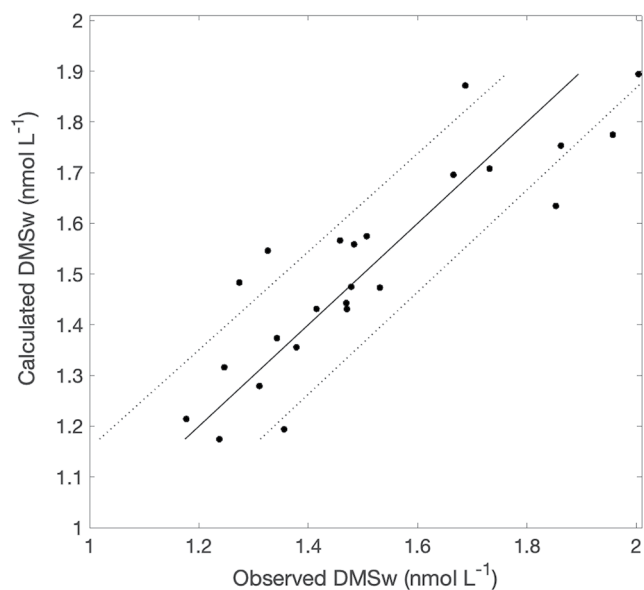


Figure 3. Observed vs. calculated DMS_w , showing the regression line (—) \pm 95% confidence interval bounds (···). DMS_w , dimethylsulfide concentration.

3.2. Multiple Linear Regression

A multiple linear regression was derived from the 3-day moving mean data set (Equation 7). The resulting regression with standardized SST and total PAR as predictor variables explained 71% of the variance in observed DMS_w (Figure 3; $r^2 = 0.76$, adjusted $r^2 = 0.71$, $p < 0.001$, $n = 24$). Quadratic terms were included to better reflect the observed relationships between DMS_w and SST (Figure 2a) and PAR (Figure 2c). This ensured that the regression residuals were normally distributed, with equal variances across the range of calculated DMS_w . Removing these terms resulted in nonrandom residuals, indicating that the regression model was not reliable. Although SSS and DMS_{PT} correlated with DMS_w (Table 1), including these as predictor variables did not improve the regression. Including either SST or SSS with PAR as predictors resulted in an $r^2 > 0.6$. However, SST and SSS are strongly inversely correlated ($r = -0.9$, $p < 0.01$) and consequently, including both variables in the regression resulted in nonrandom residuals and a negative coefficient for SST, which does not reflect the relationship shown in Figure 2a

$$DMS_w = 0.10 SST^2 + 0.34 SST + 0.14 PAR^2 + 0.12 PAR + 1.28 \quad (7)$$

The parameterization of DMS_w as a function of SST and PAR at 5 m is used to calculate DMS_w in the GBR from remotely sensed observations.

It is important to note that DMS_w in coral reefs does not linearly increase with SST when coral thermal stress thresholds are exceeded (Fischer & Jones, 2012; Jones et al., 2007). To avoid overestimating DMS_w for days when SST exceeded this threshold, SST in Equation 7 is substituted with an estimated coral thermal stress threshold.

The National Oceanic and Atmospheric Administration (NOAA) Coral Reef Watch define a coral thermal stress threshold as 1 °C above the SST of the climatologically hottest month of the year (Liu et al., 2006). This estimate is used to derive indices to predict coral bleaching (e.g., Degree Heating Weeks) and predict the extent and severity of coral bleaching events well in the GBR (Bainbridge, 2017; Hughes et al., 2018; Jackson et al., 2018; Skirving et al., 2018). These coral bleaching indices assume that living corals can tolerate summer PAR levels without bleaching (Skirving et al., 2018); however, when summer PAR combines with SST above the thermal stress threshold, coral bleaching (and a decline in DMS_w) can occur (e.g., Jones et al., 2007). Thus, for days when $SST \geq$ thermal stress threshold, DMS_w was calculated from Equation 7, substituting SST for the coral thermal stress threshold. This threshold decreased with latitude, ranging from 27.5 °C in the southern GBR to 30 °C in the northern GBR. This reduced the average calculated DMS_w in the GBR by ~ 0.01 nmol L^{-1} in late summer.

3.3. Climatology of DMS_w

A time series of DMS_w in the GBR was calculated using Equation 7 from standardized MODIS observations of SST and total PAR at 5 m from 2001 to 2019. We assume that the empirical relationship derived between DMS_w , SST, and PAR at 5 m in the southern and central GBR can be used to estimate DMS_w beyond the region and time frame for which Equation 7 was defined. It is recognized that this may not be an accurate representation of DMS_w in the northern GBR or for other times of the year; however, in the interest of estimating total DMS_w and flux, it is important to include an estimate of DMS_w for the entire GBR and time period.

Calculated annual mean DMS_w in the GBR was 1.5 nmol L^{-1} , ranging from an average of 1.2 nmol L^{-1} in winter to 1.9 nmol L^{-1} in summer, with little spatial variability (< 0.1 nmol L^{-1}) (Figure 4). SST decreased with latitude and ranged from an average of 24 °C (20–26 °C) in winter to 27 °C (25.5–29 °C) in summer. Seasonal average PAR at 5 m ranged from 40 mol $m^{-2} d^{-1}$ (35–44 mol $m^{-2} d^{-1}$) in winter to 50 mol $m^{-2} d^{-1}$ (46–53 mol $m^{-2} d^{-1}$) in summer, decreasing with latitude in winter yet increasing with latitude in summer.

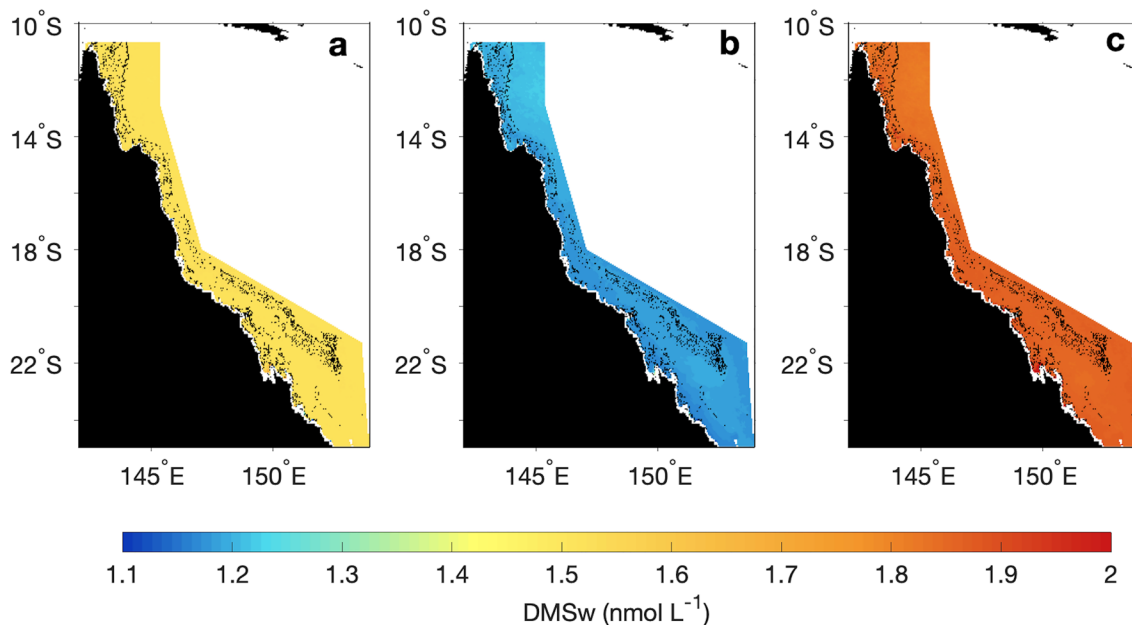


Figure 4. (a) Annual, (b) winter (May-October), and (c) summer (November-April) average calculated DMS_w in the GBR. Northeastern Australia and the location of coral reefs in the GBRMP are shown in black. DMS_w , dimethylsulfide concentration; GBR, Great Barrier Reef; GBRMP, Great Barrier Reef Marine Park.

Although a small latitudinal gradient is evident in winter, with higher DMS_w in the northern GBR due to higher SST and PAR, no clear latitudinal gradient is evident in the summer or annual mean DMS_w concentration (Figure 4). This is due to the inverse meridional trends in summer SST and PAR, which resulted in uniform calculated annual and summer average DMS_w (Figure 4). We assume that living corals and marine algae are adapted to the typical local environment and so uniform average concentrations can be expected throughout the GBR.

Calculated DMS_w area-averaged over the GBR ranged from 1 ± 0.01 (2 SE) $nmol L^{-1}$ in late winter to 2.1 ± 0.1 $nmol L^{-1}$ in summer (Figure 5). Daily mean observed DMS_w during the RV *Investigator* voyage was 1.5 ± 0.2 $nmol L^{-1}$, similar to mean calculated DMS_w for the same time period and location (1.3 ± 0.04 $nmol L^{-1}$) (Figure 5a). The climatology of calculated DMS_w shows a seasonal signal, with increased concentration in early summer (October-December), followed by a plateau in concentration in January-February when SST and PAR were highest and when coral bleaching typically occurs (Figure 5b).

3.4. Climatology of DMS Sea-Air Flux

DMS sea-air flux was calculated from MODIS observations of SST, U_{10} derived from ERA5 u and v -wind components and the DMS_w time series (Figure 5a). Three parameterizations for K_w were used to calculate DMS sea-air flux and the mean of these is mapped in Figure 6. Annual mean DMS flux averaged over the GBR was $4.9 \mu mol m^{-2} d^{-1}$ (0.9 – $11.4 \mu mol m^{-2} d^{-1}$), ranging from an average of $4.3 \mu mol m^{-2} d^{-1}$ (0.7 – $10.1 \mu mol m^{-2} d^{-1}$) in winter, to $5.4 \mu mol m^{-2} d^{-1}$ (1.2 – $12.6 \mu mol m^{-2} d^{-1}$) in summer (Figure 6). DMS flux remained consistently high over the coral reef, representing direct coral-atmosphere DMS exchange during coral exposure to air at low tide.

The climatologies of DMS sea-air flux area-averaged over the GBR show a seasonal signal, ranging from an average of $3.1 \pm 0.3 \mu mol m^{-2} d^{-1}$ in late winter to $7.3 \pm 1.6 \mu mol m^{-2} d^{-1}$ in late summer (Figure 7). Including an estimate of coral-atmosphere DMS release at low tide increased the GBR-averaged flux by $1.5 \mu mol m^{-2} d^{-1}$. Nightingale et al. (2000) and Goddijn-Murphy et al. (2012) assume that sea-air flux increases linearly with U_{10} . However, Vlahos and Monahan (2009) suggest that the DMS gas transfer velocity decreases when wind speed exceeds $10 m s^{-1}$, thereby providing a more conservative estimate of DMS sea-air flux (Figure 7). From the three DMS sea-air flux parameterizations (Figure 7), it is estimated that the

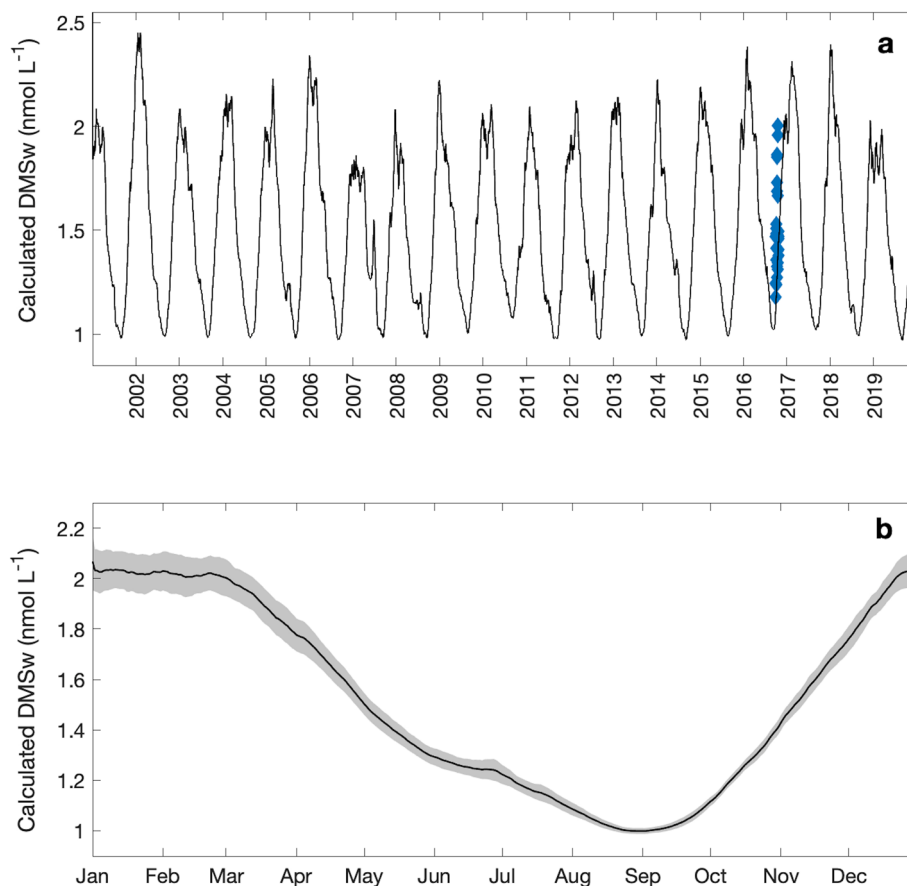


Figure 5. (a) Calculated DMS_w averaged over the entire GBR from 2001 to 2019. Blue dots represent the daily mean observed DMS_w in the southern GBR ($\sim 18^\circ S$ – $23^\circ S$). (b) Calculated DMS_w climatology ± 2 SE (shaded regions). DMS_w , dimethylsulfide concentration; GBR, Great Barrier Reef; SE, standard errors.

347,000 km² GBR and lagoon waters release 0.03–0.05 Tg yr⁻¹ of DMS (1,500–2,100 mol km² yr⁻¹), which is equivalent to 0.02 Tg yr⁻¹ of sulfur.

4. Discussion

SST and PAR synergistically impact coral physiological stress (Rosic et al., 2020) and were correlated with DMS_w concentration in the GBR. Both field and laboratory analyses have demonstrated an increase in intracellular and dissolved DMS(P) production from corals and in reef seawaters in response to increasing temperature and solar irradiance (Deschaseaux et al., 2014b; Gardner et al., 2016; Jones et al., 2007), until a coral physiological tolerance threshold is approached (Jones et al., 2007, 2018).

DMS_w negatively correlated with instantaneous PAR (Table 1), reflecting photo-oxidation of DMS(P) during the day (Gabric et al., 2008; Galí et al., 2013). Conversely, DMS_w positively correlated with daily total PAR (Table 1), reflecting a relative increase in coral and algal DMS production with solar irradiance (Deschaseaux et al., 2014b; Jones et al., 2007). DMS_w also increased with decreasing SSS, supporting previous findings that DMS(P) are involved in the coral hyposalinity stress response (Gardner et al., 2016). Hourly mean DMS_w did not correlate with $DMSP_t$ (Table 1), perhaps indicating that cleavage of DMSP is not the rate-limiting process in DMS_w production. DMS(P) removal mechanisms such as rapid photo-oxidation during the day, microbial consumption and vertical mixing in the water column can confound the relationship between DMSP and DMS at the sea surface, highlighting the complexity in DMS(P) cycling (Gabric et al., 2008; Galí et al., 2013).

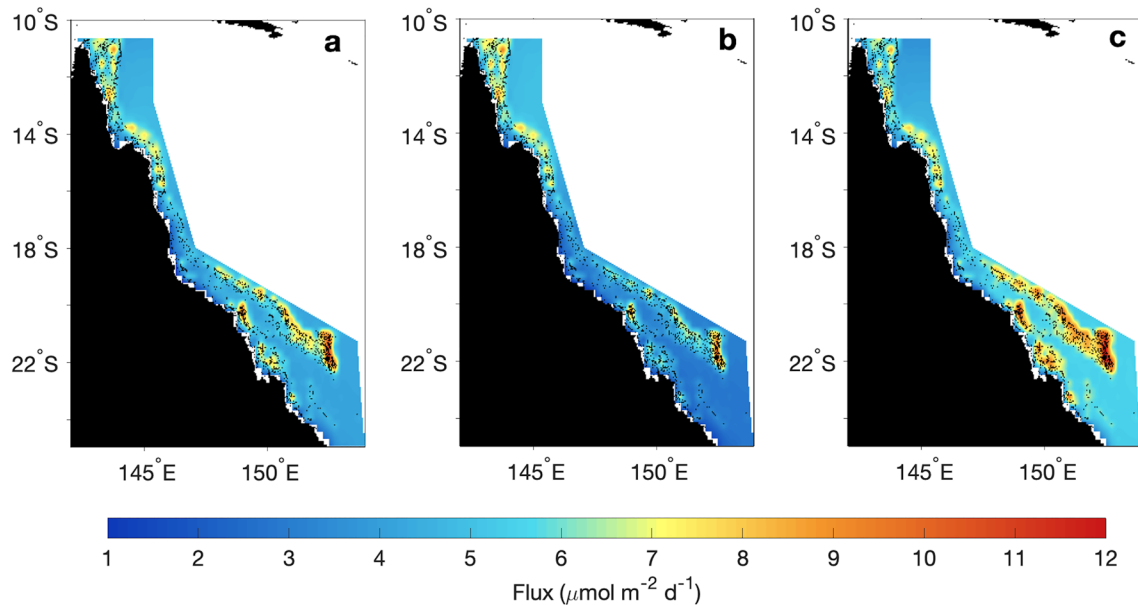


Figure 6. (a) Annual, (b) winter (May-October), and (c) summer (November-April) average DMS flux in the GBR. Values represent the average of the three DMS sea-air flux parameterizations calculated in this analysis. Northeastern Australia and the location of coral reefs in the GBRMP are shown in black. DMS, dimethylsulfide; GBR, Great Barrier Reef; GBRMP, Great Barrier Reef Marine Park.

The regression derived between DMS_w , SST, and PAR explained 71% of the variance in observed DMS_w . The resulting calculated DMS_w agrees with observed seasonal trends in the GBR (Jones et al., 2018), indicating that the regression predicts DMS_w reasonably well. The remaining variability was likely driven by nonlinear processes such as $\text{DMS}(P)$ cycling by marine microbes (Bourne et al., 2016; Bullock et al., 2017; Sun et al., 2016), photolysis and vertical mixing in the water column (Gabric et al., 2008; Galí et al., 2013; Toole et al., 2003). The relative contribution of corals and marine algae to measured DMS_w cannot be quantified from this analysis. However, *Acropora* corals are the dominant coral genus throughout the GBR and are a relatively abundant source of DMSP for liberation of DMS to the coral reef environment (Swan et al., 2017a). DMS_w is often higher in seawater sampled directly over the coral reef compared to the lagoon (Jones et al., 2018) and so, corals likely contributed a large portion of measured DMS_w .

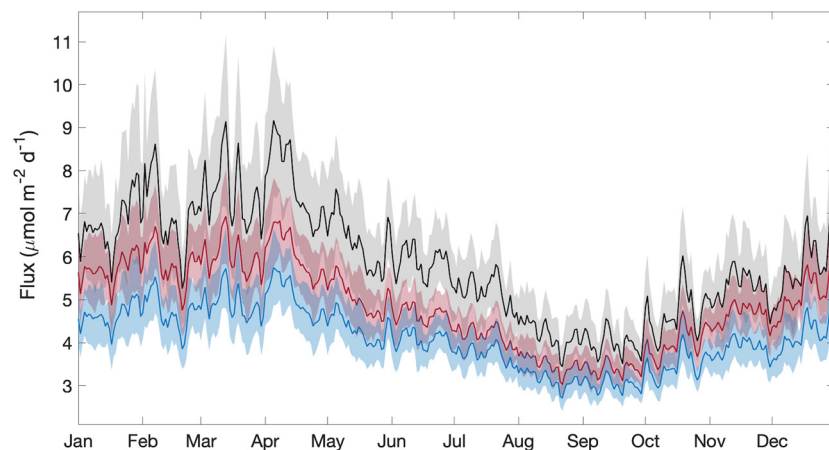


Figure 7. Climatology ± 2 SE (shaded region) of DMS sea-air flux area-averaged over the GBR, calculated from the parameterization of Nightingale et al. (2000) (black), Godrijn-Murphy et al. (2012) (red), and Vlahos and Monahan (2009) (blue). SE, standard errors; DMS, dimethylsulfide; GBR, Great Barrier Reef.

The parameterization of DMS_w as a function of PAR at 5 m agrees with that of Vallina and Simó (2007), who derived a linear regression between DMS_w and solar radiation in the upper mixed layer. The relationship with solar radiation explained >90% of the variance in a global monthly climatology of DMS_w derived from the Global Seawater DMS database (Kettle et al., 1999). While Vallina and Simó (2007) suggested that SST is not an important predictor of DMS_w , the global DMS database contains very few DMS_w surveys close to a coral reef. In the southern GBR, PAR alone explained only ~30% of the variance in observed DMS_w and required a second predictor variable of SST to explain a significant portion of the variance in observed DMS_w . The importance of SST in predicting DMS_w likely reflects the significant impact of thermal stress on DMS production by corals and zooxanthellae (Raina et al., 2013). Although the range of measured SST was narrow (3.5 °C), laboratory studies have demonstrated that small changes in SST can significantly influence DMS production by corals (Deschaseaux et al., 2014b). Further research is needed to establish the validity of the GBR observed relationship between DMS_w , SST, and PAR in other coral reef systems.

The parameterization of DMS_w in coral reef seawaters as a function of SST and PAR is a simple approach and does not account for the impacts of other factors which may influence DMS biosynthesis such as ocean acidification and changes to coral reef ecosystem structure. Mesocosm experiments in the subtropical North Atlantic revealed a decrease in algal-derived DMS with lower pH due to reduced rates of microbial catabolism of DMSP (Archer et al., 2018). The same could be true of coral-derived DMS production, which is partially driven by microbial DMSP catabolism (Raina et al., 2009). Other studies have demonstrated that temperature has a stronger influence on DMS production in algae, where increased production in response to temperature outweighed the decline in biosynthesis due to acidification (Arnold et al., 2013). Ongoing coral reef degradation, due to coral bleaching, crown of thorns outbreaks and changing water quality may also affect DMS_w concentration in reef seawaters (discussed in Jackson et al. (2020)). A decline in coral-derived DMS_w could occur with reduced coral cover; however, DMS-producing marine algae can dominate degraded coral reef ecosystems (De'ath & Fabricius, 2010; McCook & Diaz-Pulido, 2002) and counteract a decline in coral-derived DMS production. Further research is required to determine the synergistic impacts of ocean warming, ocean acidification, and coral reef degradation on DMS(P) biosynthesis in coral reefs. Additional measurements of DMS_w in coral reef seawaters are also needed and will greatly improve the parameterization presented in this analysis. Nevertheless, the results provide insight into the source strength of coral reefs to the atmospheric sulfur budget.

Calculated DMS_w area-averaged over the GBR ranged from ~1 nmol L⁻¹ in late winter to ~2.4 nmol L⁻¹ in late summer (Figure 5a). Reported mean DMS_w concentrations for the GBR lagoon (deeper waters which surround the coral reef platforms) range from 0.8 nmol L⁻¹ (0.6–1 nmol L⁻¹, $n = 4$) in winter to 1.9 nmol L⁻¹ (0.1–3.4 nmol L⁻¹, $n = 224$) in summer (Jones et al., 2018). DMS_w is typically higher in shallow waters over the coral reef, ranging from 1.7 nmol L⁻¹ (0.1–7.7 nmol L⁻¹, $n = 160$) in winter to 3.1 nmol L⁻¹ (0.1–54 nmol L⁻¹, $n = 226$) in summer (Jones et al., 2018). The range of calculated DMS_w falls within the mean range reported for the GBR lagoon and coral reef waters. This was expected given that the RV *Investigator* data set is comprised of measurements taken both in the deeper lagoon and at several stations in shallower waters close to the coral reef (Figure 1). DMS_w calculated from the multiple linear regression derived from the RV *Investigator* data set may therefore underestimate concentrations over the coral reef (Figure 4).

The range of calculated DMS_w is ~1 nmol L⁻¹ lower than that reported in the Lana et al. (2011) climatology for the Eastern Australian biogeochemical province (~2–4 nmol L⁻¹). The seasonal summer peak in DMS_w also occurs later in our climatology (February–March, rather than October), due to prescribing a maximum DMS_w value for days when SST exceeds the coral thermal stress threshold. Imposing an upper limit on DMS_w accounted for the effect of high SST and irradiance on coral and algal DMS production. High SST can reduce the PAR absorption capacity of algal photosystems, leading to excess production of ROS and oxidative damage in the coral holobiont (Lesser, 2006, 2011). DMS(P) scavenge ROS, forming DMSO, which results in a decline in ambient DMS_w concentration (Deschaseaux et al., 2014a; Fischer & Jones, 2012; Jones et al., 2007). The DMS_w climatology presented in this analysis may be closer to contemporary concentrations in the GBR after back-to-back bleaching events and the loss of up to ~40% of hard coral cover between March and November 2016 (Hughes et al., 2018). The RV *Investigator* sampled DMS_w in the GBR 8 months after the 2016 mass coral bleaching event and this may be one reason why measured DMS_w was lower than other studies have documented in the GBR (e.g., Jones et al., 2018).

Average DMS sea-air flux in the GBR ranged from $\sim 3 \mu\text{mol m}^{-2} \text{d}^{-1}$ in late winter to a maximum of $\sim 11 \mu\text{mol m}^{-2} \text{d}^{-1}$ in late summer (Figure 7). This estimate is similar in magnitude to that presented in Lana et al. (2011) for the Eastern Australian biogeochemical province ($\sim 5\text{--}15 \mu\text{mol m}^{-2} \text{d}^{-1}$). Again, the seasonal peak in sea-air flux occurs later in summer in our climatology (February–March, rather than October), likely due to prescribing an upper limit on DMS_w for SST beyond the coral thermal stress threshold. DMS flux was highest over the coral reef (Figure 6), due to the inclusion of a direct coral-atmosphere DMS flux estimate. Field-based findings have demonstrated that the southern GBR is a strong, albeit intermittent, source of DMS above the background oceanic signal (Jones et al., 2007; Swan et al., 2017b). However, there remains uncertainty in the source strength of direct coral-atmosphere DMS release.

Swan et al. (2017b) estimate that only $0.2 \mu\text{mol m}^{-2} \text{d}^{-1}$ is released from the coral reef, based on measurements of DMS_a made at a fixed location at Heron Island (southern GBR). This is significantly lower than the estimate of $\sim 22 \mu\text{mol m}^{-2} \text{d}^{-1}$ provided by Hopkins et al. (2016), possibly due to difficulties in distinguishing the coral reef and oceanic DMS_a sources under strong atmospheric mixing conditions from the prevailing southeasterly trade winds. The estimate provided by Hopkins et al. (2016) is the only direct measurement of coral-atmosphere DMS release; however, this estimate assumes that *Acropora* spp. are the sole source of direct coral-air DMS flux and does not account for variability in the extent of coral exposure or the complexity of the reef environment (Hopkins et al., 2016). Inclusion of this previously unaccounted for influence on DMS flux yields a more accurate estimate of the contribution of coral reefs to the atmospheric sulfur budget. Further research is needed to reduce the uncertainty in coral-air DMS flux and to accurately scale laboratory-derived fluxes to the natural coral reef environment.

We estimate that $0.03\text{--}0.05 \text{ Tg yr}^{-1}$ of DMS is released from the GBR (0.02 Tg yr^{-1} of sulfur). This agrees with a previous estimate of 0.03 Tg yr^{-1} of DMS for the GBR lagoon and associated coral reefs, calculated from field measurements of DMS_w over the past three decades (Jones et al., 2018). The range is slightly higher due to the inclusion of direct coral-atmosphere DMS release in our flux calculations. Assuming that DMS production and flux are constant across coral reefs, global tropical coral reefs ($\sim 600,000 \text{ km}^2$) could release $0.06\text{--}0.08 \text{ Tg yr}^{-1}$ of DMS ($0.03\text{--}0.04 \text{ Tg yr}^{-1}$ of sulfur). Global sea-air flux is estimated to be $17.6\text{--}34.4 \text{ Tg yr}^{-1}$ of sulfur in the form of DMS (Kettle & Andreae, 2000; Lana et al., 2011; Land et al., 2014). The contribution of sulfur as DMS from coral reefs represents $0.1\text{--}0.2\%$ of global sea-air flux estimates, which is a disproportionate amount of sulfur released from $\sim 0.1\%$ of the ocean surface (Spalding et al., 2001).

Global modeling studies have found that marine DMS is an important source of sulfate aerosol, influencing climate (Gabric et al., 2013; Mahajan et al., 2015; Thomas et al., 2010; Woodhouse et al., 2010) and providing a cooling effect of up to $0.45 \text{ }^\circ\text{C}$ (Fiddes et al., 2018). Given that coral reefs are strong regional sources of DMS, it is hypothesized that DMS emissions may influence the local radiative balance (Jackson et al., 2020; Jones et al., 2018; Swan et al., 2016) via a negative feedback on aerosol and cloud formation (e.g., Cropp et al., 2007). However, this coral reef source of atmospheric sulfur is not currently accounted for in global climatologies. Fiddes et al. (2020) was the first model study to explicitly account for the coral reef source of DMS by adding a laboratory-based estimate of 50 nmol to the Lana et al. (2011) DMS_w climatology, scaled to the percentage of coral cover. This approach estimated that coral reefs release 0.3 Tg yr^{-1} of DMS (Fiddes et al., 2020), an order of magnitude higher than what is estimated by this study. Despite the high coral reef DMS flux estimate, Fiddes et al. (2020) found only a small response in nucleation and Aitken mode aerosol number concentrations and mass. This implies that coral reef DMS emissions are not important in contemporary regional climate, likely due to the dominance of anthropogenic, continental, and sea spray aerosols (Chen et al., 2019; Mallet et al., 2016). However, observational studies in the GBR suggest that DMS flux could influence local aerosol processes (Modini et al., 2009; Swan et al., 2016), which are perhaps not captured by global or regional climate models. Therefore, the importance of coral reef-derived DMS on the local radiative balance remains uncertain. Future work will test if the inclusion of coral reef-derived DMS in regional climatologies and regionally focused earth system models improves model accuracy and addresses the importance of coral reefs in biogeophysical processes.

5. Conclusions

This analysis supports previous findings that coral reefs are a significant regional source of DMS. The relationship between DMS_w , SST, and PAR is successfully parameterized and used to estimate the source strength of coral reefs to the sulfur cycle. The release of DMS from corals exposed to air at low tide is an important process that is not accounted for in current DMS sea-air flux calculations. By accounting for direct coral-atmosphere DMS release, we show that the total contribution of coral reefs is higher than previous estimates. Assuming that DMS production and sea-air flux are constant across coral reefs, 0.08 Tg yr^{-1} of DMS (0.04 Tg yr^{-1} of sulfur) could be released from tropical coral reefs, with the potential to influence the local radiative balance. This is the first parameterization of seawater DMS concentration in coral reefs and provides valuable insight into the role of coral reefs in global sulfur cycling.

Our DMS_w proxy can be used to investigate how changes to SST and PAR (due to changes in cloud cover and water clarity) might affect DMS production and flux from coral reefs in future. Ongoing ocean warming, acidification, and declining water quality are adversely affecting coral reef ecosystems and may lead to an ecological regime shift as coral resilience to disturbances continues to decline. This may lead to a change in DMS production and sea-air flux from coral reefs, with potential impacts on the local radiative balance.

Conflict of Interest

The authors declare no conflicts of interest.

Data Availability Statement

Underway measurements from Marine National Facility RV *Investigator* voyage IN2016_V05 (in2016_v05u-wy5min_csv.zip) are available at url: <https://www.marlin.csiro.au>. Daily MODIS Aqua and Terra observations of SST, PAR, k490, and CHL are available from NASA's OceanColor Distributed Active Archive Centre (NASA Goddard Space Flight Center, 2018).

Acknowledgments

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