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The impact of coral reef ecosystems and upwelling events on the marine carbon dynamics of Southern Taiwan

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Running title: pCO$_2$ in a coral reef ecosystem

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ABSTRACT

The ocean is the largest carbon reservoir and plays a crucial role in regulating atmospheric CO_2 levels, especially in the face of climate change. In coral reef ecosystems, the complexity and importance of the carbonate system must be better appreciated as atmospheric CO_2 concentrations continue to rise. This study measured pCO_2 over space and time in Nanwan Bay, a coral reef ecosystem in Southern Taiwan, to identify factors that influence its variation. The results showed that mean pCO_2 values varied seasonally, with values of 394, 406, 399, and 367 µatm in spring, summer, fall, and winter, respectively. These seasonal mean differences (ΔpCO_2) relative to atmospheric pCO_2 (397, 392, 392, & 396, respectively) were -3, 14, 7, and -29 µatm, respectively. These findings suggest that the Nanwan Bay coral reef ecosystem acts as a sink for atmospheric CO_2 during the spring and winter, with an average sea-air gas flux of -1 gC m^{-2} $year^{-1}$ and a net annual uptake of -29 tC. The carbonate system parameters of the surface water in this high-biodiversity sub-tropical marine ecosystem were influenced not only by seasonal temperature variation but also by vertical mixing, intermittent upwelling, and biological effects.

Keywords: carbon sink, carbon source, coral reef, pCO_2, total alkalinity, upwelling

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1. Introduction

The concentration of atmospheric carbon dioxide (CO_2) varies significantly based on region and season. High-latitude temperate regions and coastal seas act as sinks for atmospheric CO_2, while subtropical and tropical coastal seas, estuaries, and coral reefs are generally sources (Borges et al., 2005; Cai et al., 2003; Frankignoulle et al., 1998; Frankignoulle et al., 1996; Gattuso et al., 1997; Gattuso et al., 1993; Ito et al., 2005; Ohde and Van Woesik, 1999; Wang and Cai, 2004; Yan et al., 2011; Bates et al., 2001). The hydrological characteristics of coastal waters can vary significantly, leading to differences in surface water pCO_2 even within the same continental shelf. Furthermore, upwelling areas (such as California & Oman) are sinks for CO_2, while the coasts of Galicia and Oregon are sources (Borges and Frankignoulle, 2002; Friederich et al., 2002; Goyet et al., 1998; Hales et al., 2005). Borges (2005) notes that when estuaries are included in the gas exchange process, coastal seas worldwide are sources of CO_2, becoming sinks when estuaries are excluded.

Various factors, such as temperature, tides, currents, river discharge, upwelling, vertical mixing, and biological metabolism, can influence CO_2 levels in coastal areas (e.g., Dai et al., 2009). These factors can interact and help explain why seasonal variation in CO_2 levels can differ greatly across regions. For instance, measurements taken at the Bermuda Atlantic Time-series station in the northwest Atlantic from 1996

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to 1998 showed that CO_2 levels were lowest in winter and highest in summer (Takahashi et al., 2002). Similarly, data collected from the Kyodo Western North Pacific Ocean Time-series station between 1998 and 2000 indicated that CO_2 levels were lower in summer compared to winter (Takahashi et al., 2002).

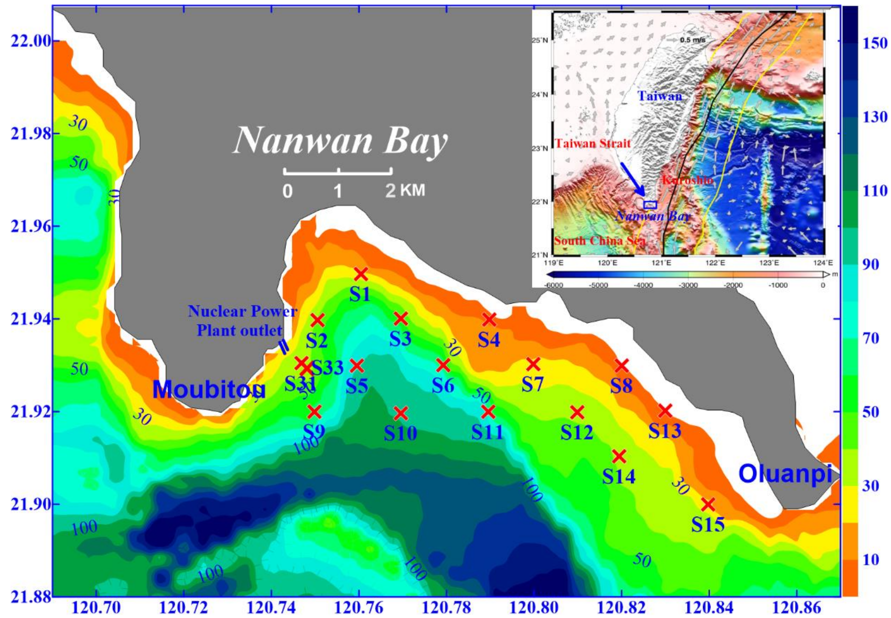

Fig. 1 Map of sampling stations in Nanwan Bay, Taiwan. The color bar indicates the bottom depth (m). The blue arrow in the upper right inset shows the sampling region.

Coral reefs, despite occupying only 0.2% of the ocean, are home to a significant portion of marine biodiversity, housing one-third of all marine species (Sheppard et al., 2017; Reaka-Kudla, 1997). These productive ecosystems boast efficient and stable carbonate sedimentation rates, contributing to 23-26% of global annual $CaCO_3$ sedimentation (Suzuki and Kawahata, 2004). However, coral reefs are highly

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vulnerable to the effects of climate change and anthropogenic pollution, which can significantly impact the organisms responsible for building the reef structure (Bellwood et al., 2004; Hughes et al., 2017; Chen, 2021). Due to the structural complexity and high biodiversity of coral reefs, their carbon dynamics may differ substantially from those of the open ocean. Currently, it is unclear whether coral reef ecosystems act as net carbon sources (Ware et al., 1992; Gattuso et al., 1993; Gattuso et al., 1999; Fagan and Mackenzie, 2007; Lønborg et al., 2019; Yan et al., 2018; Watanabe and Nakamura, 2019; Frankignoulle et al., 1998) or sinks (Kayanne et al., 1995; Mayer et al., 2018; Suzuki, 1998; Suzuki and Kawahata, 2004). Additionally, since environmental changes can result in physiological changes in resident organisms (Fabry et al., 2008), it is challenging to predict how seawater carbon levels will change in response to oceanographic anomalies.

Semi-enclosed Nanwan Bay is situated at the southernmost point of Taiwan, (Fig. 1). It is flanked by the Pacific Ocean to the east and the Taiwan Strait to the west, and faces the Luzon Strait to the south and the South China Sea (SCS) to the southwest (Lee, 1999). The bay stretches between Cape Moubitou and Cape Oluanpi and boasts a coastline primarily consisting of fringing coral reefs (Yang and Dai, 1980). Nanwan Bay is among the most diverse marine regions in Taiwan, which led to its inclusion within Kenting National Park (Meng et al., 2008). The complex seabed in

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Nanwan Bay comprises various habitats and is the first point of contact for the warm and highly saline Kuroshio Current. The water is oligotrophic, and temperatures typically range from 21 to 30°C, but with periodic upwelling events occurring (Chen et al., 2005). The bay hosts over 1,200 fish species and more than 200 species of reef-building corals, making it a significant research focus area for both the reefs and the anthropogenic stressor regime (Meng et al., 2007). Studies have shown that high levels of nutrients and suspended solids may have contributed to the decline in coral cover between 2001 and 2022 (Meng et al., 2008; Chen et al., 2022).

In previous studies of Nanwan Bay, periodic upwelling caused mixing of upper and lower seawater layers, which contributed significantly to the transfer of nutrients from the depths to shallower areas (Chen et al., 2005). In some upwelling areas, CO_2 may be released into the atmosphere, while in others, CO_2 enters the ocean from the atmosphere. Basic productivity in marine ecosystems has a potential impact on carbon cycling (Dugdale and Wilkerson, 1989; Murray et al., 1995), and upwelling brings nutrients into the photic zone, thereby stimulating the proliferation of phytoplankton and enhancing basic productivity. As a result, there is also an increased demand for carbon for CO_2 fixation at these times (Chen et al., 2004b). The average ratio of primary production to community respiration (P/R ratio) is often used to determine whether a marine system is a source or sink of CO_2 in the atmosphere (where $P/R<1$

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indicates a source, & P/R>1 indicates a sink; Smith and Hollibaugh, 1993; Robinson et al., 2002). We sought herein to determine whether Nanwan Bay is a net carbon source or sink by characterizing the marine carbonate system over time and under a range of biogeochemical processes.

2. Methods

2.1 Sampling and analysis The study was conducted across four seasons: spring (31 March 2011), summer (5 July 2011), autumn (20 October 2011), and winter (22 January 2013), in the area between Nanwan Bay's two capes, Cape Moubitou and Cape Oluanpi. A total of 17 seawater sampling stations were established, including one near the outlet of a nuclear power plant (Fig. 1). Temperature and salinity data, both with accuracy of 99.9%, were collected using an Idronaut Ocean Seven 304 CTD calibrated against an International Association for the Physical Sciences of the Ocean seawater standard. Water samples were collected using Niskin bottles with Teflon-coated inner walls. Seawater at each station was taken at two to five depths at intervals of 3 to 25 m in areas shallower than 50 m; extra samples at 65, 80 and/or 100 m were taken for stations with depths of 65-100 m. Water samples were immediately analyzed for dissolved oxygen (DO) content using YSI 52 and YSI 5905 BOD electrodes (accuracy=99.9%). Other water samples were divided into different sample bottles for additional analyses. One 300-mL amber bottle was pre-inoculated with 0.2

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mL of mercuric chloride to suppress biological activity that could affect total alkalinity (TA) and other carbonate system parameters.

Seawater pH and total TA were measured using an automated titration system consisting of a Mettler-Toledo DL53 with a DG-111 electrode. Prior to measurement, the electrode was calibrated using Merck standard buffer solution (NIST) at 25°C. The calibration ranges for pH 4, 7, and 10 were set to fall within the range of 176±30 mV, 0±30 mV, and -176±30 mV, respectively (calibration slope of -56 to -59). Measured values were expressed on the NBS scale (pH_{NBS}). The electrode was also calibrated using Tris-artificial seawater buffered at pH 8.083 and AMP artificial seawater buffered at pH 6.776, with measured values given on the total scale (pH_{tot}).

For TA measurements, 40 g of seawater were titrated with 0.1 N HCl at 25°C. Titration continued until the pH exceeded the end point (~pH 4.4), and then continued until ~pH 3.0, with the potential change and titration volume recorded. The consumed volume of HCl was calculated using the Gran (1952) function based on the linear relationship between titration volume and pH, and TA was obtained by plotting the consumed volume of HCl. The reference material for experimental quality control was obtained from Professor Andrew Dickson (Scripps Institute of Oceanography, USA), and the pH of the reference material was calculated by entering dissolved inorganic carbon (DIC) and TA data into CO_2SYS (Lewis and Wallace, 1998).

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The pH_{NBS} measurement accuracy in this study was ±0.01 units, the pH_{tot} accuracy was ±0.009 units, and the TA accuracy was ±2.7 μmol kg^{-1} (precision=0.12%). pCO_2 was also calculated with CO_2SYS from measured pH and TA. The dissociation constants of carbonic acid used were the revised K1 and K2 values from Mehrbach et al. (1973) and Dickson and Millero (1987).

2.2 Calculation of the exchange flux of CO_2 between the ocean and the atmosphere The formula for calculating the exchange flux of CO_2 between the ocean and the atmosphere was as follows:

$$F_{GAS}=k \times K_H \times (pCO_2\text{seawater}-pCO_2\text{air})$$

where k is the gas exchange rate of CO_2 (air-sea gas transfer rate), which was obtained from an empirical formula based on wind speed proposed by Wanninkhof (1992): k=0.31 × *u*2 × (*Sc*/660)-0.5, where *u* is wind speed 10 m above sea level (in m/s; data from the Central Weather Bureau's Oceanic Center-Erluanbi buoy); *Sc* (Schmidt number) is a function of temperature, which can be obtained from the *in situ* sea surface temperature (T) as follows:

$$Sc = 2073.1 - 125.62 \times T + 3.6276 \times T2 - 0.043219 \times T3$$

The solubility of CO_2 gas in seawater (K_H), expressed in moles/L·atm, was calculated using the formula developed by Weiss (1974):

$$\ln K_H = -58.0931 + 90.5069\left(\frac{100}{T}\right) + 22.2940\ln\left(\frac{T}{100}\right) + S\left[0.027766 - 0.025888 + \left(\frac{T}{100}\right) + 0.0050578\left(\frac{T}{100}\right)^2\right]$$

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Since we did not measure atmospheric CO_2, we used data from air samples collected by the United States National Oceanic and Atmospheric Administration (NOAA) at Dongsha Island: 397 µatm in March 2011, 392 µatm in July 2011, 392 µatm in October 2011, and 396 µatm in January 2013 (https://gml.noaa.gov/dv/data/index.php?category=Greenhouse%2BGases&parameter_name=Carbon%2BDioxide&site=DSI).

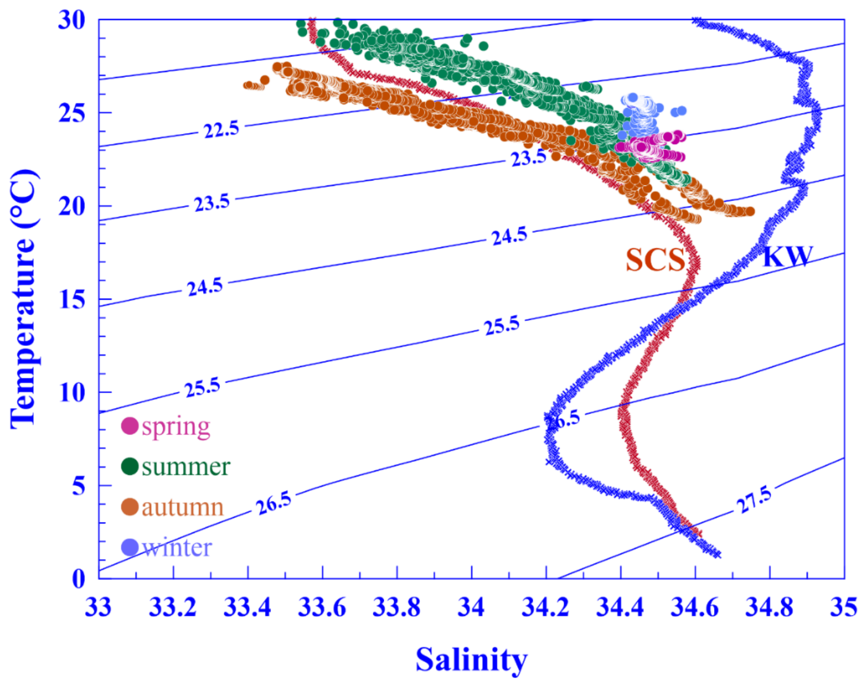

Fig. 2 Temperature vs. salinity (T-S) diagram at Nanwan Bay, Taiwan in spring, summer, autumn, and winter. SCS = South China Sea and KW = Kuroshio Current waters.

3. Results and discussion

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3.1 Variation in hydrological parameters Both temperature and salinity varied over time in Nanwan Bay (Fig. 2), with the seasonal variation likely driven by both the monsoon and SCS circulation patterns. The Kuroshio Current flows northward along Taiwan's east coast, with a portion of Western Philippine Sea (WPS) water following the Kuroshio and then flowing westward along the northern SCS shelf (Yuan et al., 2006). Nan et al. (2015) suggested that surface salinity of 34 or higher is characteristic of the Kuroshio, indicating potential inundation of the Kuroshio Current into Nanwan Bay during high-salinity spring periods. During summer, the southwest monsoon dominates, leading to a decrease in the Kuroshio's influence; the main circulation of the Kuroshio shifts westward to the Luzon Strait, limiting its intrusion into the northwestern SCS region (Liang et al., 2008). In the northern SCS region, the southwest-to-northeast circulation prevails during the monsoon, with most seawater flowing out of the SCS through the Luzon Strait and converging with the Kuroshio axis, resulting in Nanwan Bay being dominated by the SCS water mass during the summer. Analysis of temperature and salinity data from Nanwan Bay, the SCS, and the Kuroshio current indicates that Nanwan Bay mainly consists of the SCS water mass during summer and autumn, while during spring and winter, the water masses are intermediate between the two (Fig. 2). As such, Nanwan Bay is classified as a mixed water mass area, comprising both SCS and Kuroshio Current water masses.

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Table 1. Correlation matrix of variables with correlation coefficient (*r*) for spring, summer, autumn, and winter. Variables include temperature, salinity, dissolved oxygen (DO), saturation of DO (DO%), total alkalinity (TA), dissolved inorganic carbon (DIC), and pH.

Spring	Temperature	Salinity	DO	DO (%)	TA	DIC	pH
Salinity	0.35**						
DO	-0.23	-0.40**					
DO (%)	0.02	-0.32**	0.89**				
TA	0.04	0.03	0.19	0.11			
DIC	-0.17	-0.04	0.25*	0.11	0.92**		
pH	0.43**	0.13	-0.02	0.05	0.61**	0.27*	
$p\mathrm{CO}_2$	-0.28*	-0.07	0.03	-0.02	-0.45**	-0.09	-0.96**

Summer	Temperature	Salinity	DO	DO (%)	TA	DIC	pH
Salinity	-0.96**						
DO	0.65**	-0.53**					
DO (%)	0.90**	-0.81**	0.90**				
TA	-0.82**	0.81**	-0.52**	-0.72**			
DIC	-0.91**	0.84**	-0.68**	-0.87**	0.91**		
pH	0.89**	-0.79**	0.72**	0.89**	-0.72**	-0.94**	
$p\mathrm{CO}_2$	-0.22*	0.07	-0.45**	-0.38**	0.23*	0.51**	-0.64**

Autumn	Temperature	Salinity	DO	DO (%)	TA	DIC	pH
Salinity	-0.95**						
DO	0.88**	-0.85**					
DO (%)	0.95**	-0.93**	0.98**				
TA	-0.57**	0.56**	-0.51**	-0.55**			
DIC	-0.76**	0.75**	-0.66**	-0.72**	0.88**		
pH	0.79**	-0.77**	0.66**	0.73**	-0.43**	-0.80**	
$p\mathrm{CO}_2$	-0.32**	0.33**	-0.22	-0.27*	0.27*	0.63**	-0.83**

Winter	Temperature	Salinity	DO	DO (%)	TA	DIC	pH
Salinity	-0.32**						
DO	0.43**	-0.26*					
DO (%)	0.78**	-0.34**	0.70**				
TA	-0.15	0.06	-0.19	-0.17			
DIC	-0.39**	0.02	-0.34**	-0.41**	0.89**		
pH	0.59**	0.06	0.39**	0.61**	-0.12	-0.56**	
$p\mathrm{CO}_2$	-0.34**	-0.19	-0.33**	-0.44**	0.29**	0.67**	-0.94**

*: $p \leq 0.05$ and **: $p \leq 0.01$.

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During the survey period, there was a clear positive correlation between pH and temperature. Additionally, pH and TA exhibited significant correlations with salinity during summer and autumn, but not in spring and winter (Table 1). These findings suggest that vertical mixing occurs in spring and winter, whereas the positive correlation between TA and pH in the spring season indicates upwelling. It is expected that TA and salinity will covary because the charge differences between cations and anions in seawater change with salinity. Salinity generally increases with depth and is influenced by various factors such as rainfall, evaporation, and freshwater input, which can lead to changes in TA.

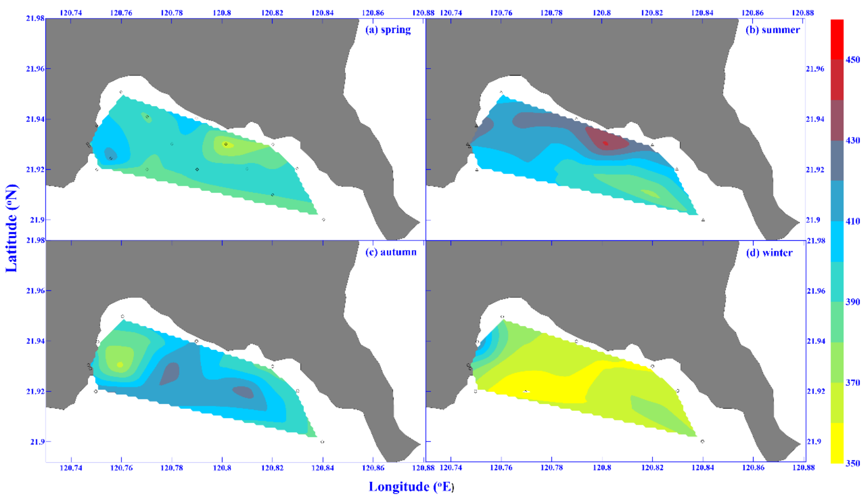

Fig. 3 Seasonal variation in sea surface pCO_2 (µatm) in Nanwan Bay, Taiwan in spring (a), summer (b), autumn (c), and winter (d). Values in legends along the right sides of panels correspond to µatm.

3.2 Changes in surface water pCO_2 pCO_2 levels in Nanwan Bay ranged from

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364–422 µatm, 362–448, 350–480, and 345–427 µatm in spring, summer, autumn, and winter, respectively (Fig. 3). The means (±SD) for these levels were 393.2 (±11.6), 411.4 (±19.0), 401.7 (±18.3), and 370 (±17.3) µatm, respectively. The mean temperatures during these seasons were 23.3, 28.5, 26.9, and 26℃, respectively. In the open ocean, $p\mathrm{CO_2}$ levels are primarily influenced by temperature, horizontal transport and vertical mixing, biological processes, and gas exchange (e.g., Dai et al., 2009).

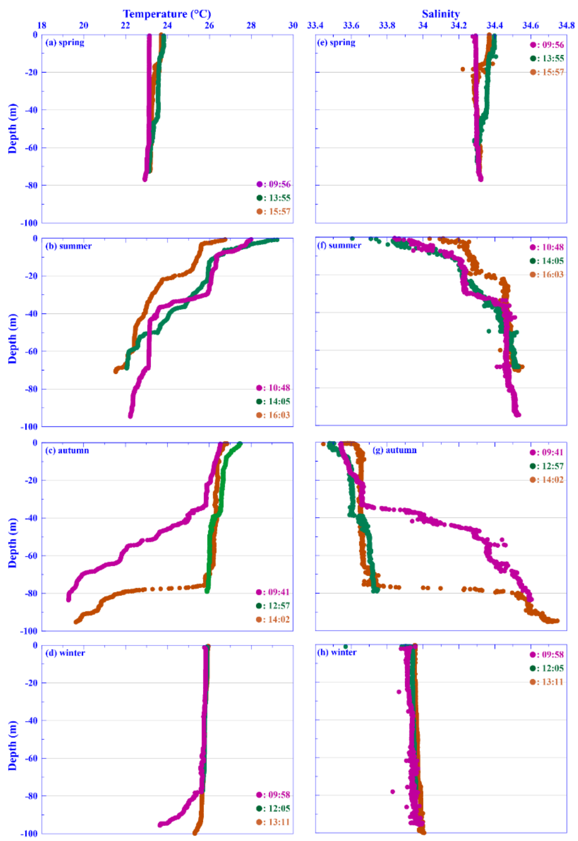

Fig. 4 Vertical profiles of temperature and salinity at station S10 in spring (a & e, respectively), summer (b & f, respectively), autumn (c & g, respectively), and winter (d & h, respectively) at three sampling times as indicated in each panel.

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Due to the mixing of different water masses by monsoons, tides, eddies, upwelling, and other ocean currents, significant gradient changes were observed at different times at station S10 (Fig. 4) and in the carbonate parameter data (Figs. 5–8). In summer and autumn, more obvious mixing occurred, as seen by the vertical variation in temperature and salinity, while in spring and winter, mixing was less evident.

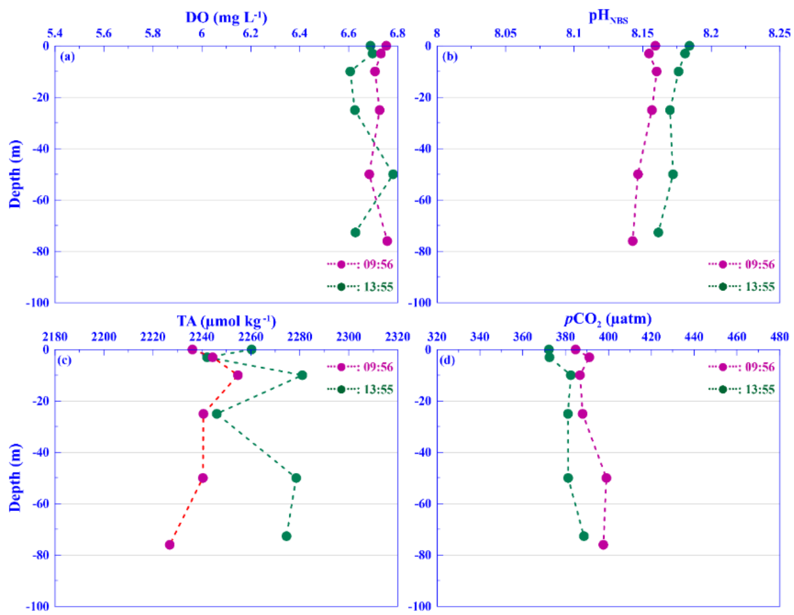

Fig. 5 Vertical profiles of dissolved oxygen (DO; a), pH (b), total alkalinity (TA; c), and pCO_2 (d) in spring at station S10 at two sampling times.

According to Lee et al. (1997; 1999a; 1999b), cold-water upwelling occurs with tidal changes in Nanwan Bay, which increases vertical mixing. The temperature-salinity-pH-DO diagram of station S1 shows that during cold-water upwelling, cold, low-DO, low-pH, and high-salinity deep-sea water intrudes the nearshore regions of Nanwan

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Bay (Fig. 9). Seawater quality profiles of S10 provide further evidence of upwelling, such as low temperatures, low DO, low pH, high salinity, and pCO_2 increases of 31 and 37 μatm in summer and autumn, respectively (Figs. 4 & 6–7).

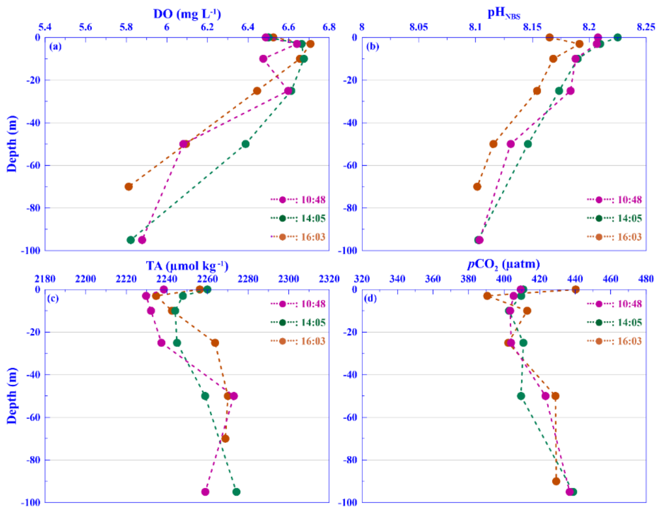

Fig. 6 Vertical profiles of dissolved oxygen (DO; a), pH (b), total alkalinity (TA; c), and pCO_2 (d) in summer at station S10 at three sampling times.

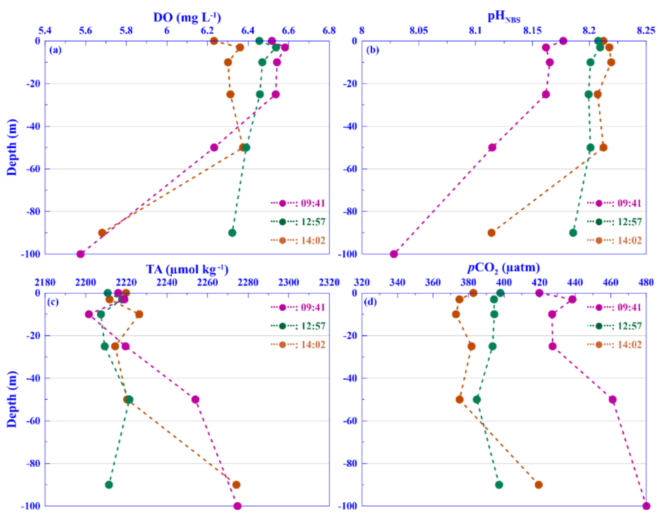

Fig. 7 Vertical profiles of dissolved oxygen (DO; a), pH (b), total alkalinity (TA; c), and pCO_2 (d) in autumn at station S10 at three sampling times.

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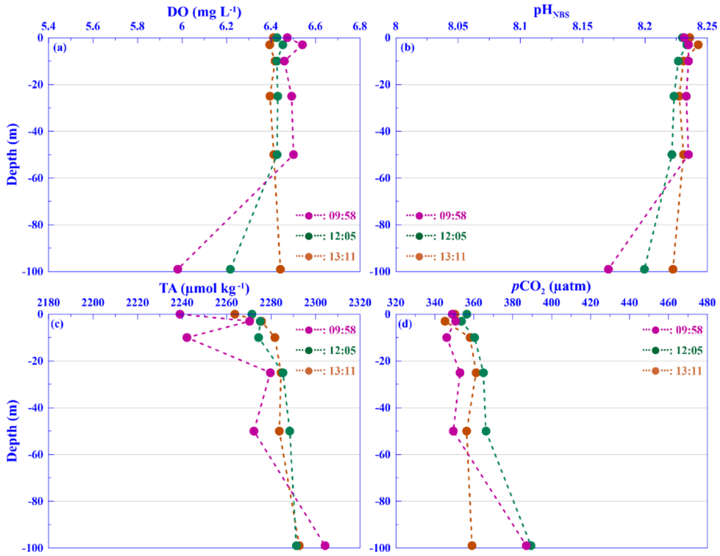

Fig. 8 Vertical profiles of dissolved oxygen (DO; a), pH (b), total alkalinity (TA; c), and pCO_2 (d) in winter at station S10 at three sampling times.

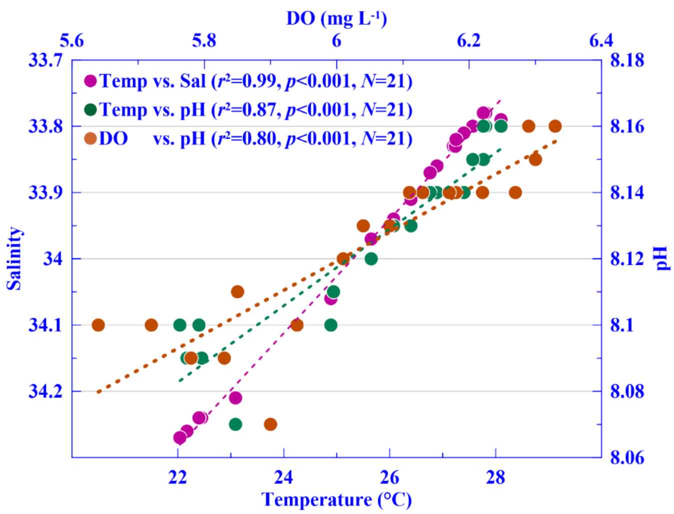

Fig. 9 Relationships between temperature and salinity or pH, as well as dissolved oxygen (DO) vs. pH during upwelling events (data from Tew et al., 2014).

As temperature increases, CO_2 solubility decreases, causing an increase in pCO_2. Takahashi et al. (2002) proposed to evaluate the relative effects of temperature and

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non-temperature effects on pCO_2 changes as follows:

$$pCO_2 \text{ at } T_{obs} = (pCO_2)_{Mean\ annual} \times \exp[0.0423(T_{obs} - T_{mean})]$$

$$pCO_2 \text{ at } T_{mean} = (pCO_2)_{obs} \times \exp[0.0423(T_{mean} - T_{obs})]$$

pCO_2 at T_{obs} is calculated using the average pCO_2 to determine the pCO_2 value at the measured temperature, assuming that the change in pCO_2 is due to temperature; pCO_2 at T_{mean} is the standardized pCO_2 value at the average temperature, assuming that the change in pCO_2 is *not* due to temperature; T_{mean} and T_{obs} are the annual average temperature and the measured temperature on-site, respectively.

The mean pCO_2 of the monitoring stations varied over time, indicating that temperature and non-temperature effects had different impacts on the average pCO_2 of each station (Fig. 10). This suggests that seasonal changes in pCO_2 are influenced by both temperature and non-temperature effects, with some stations showing larger changes than others. It is believed that the stations with larger pCO_2 changes are primarily affected by one type of effect (temperature vs. other), while the smaller changes reflect the mutual offsetting of the two effects. The variability in pCO_2 observed at S31 and S33, which are located near the Nuclear Power Plant outlet, is likely driven by temperature change, as the water temperature in this area is consistently higher than that of the surrounding area throughout the year. In fact, we expected that temperature effects on pCO_2 would be more pronounced at these sites.

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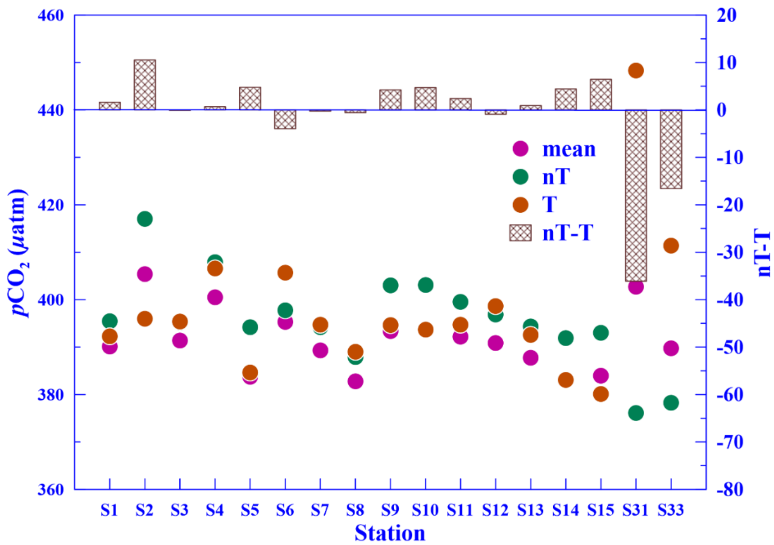

Fig. 10 Control factors and impact levels of surface water pCO_2 at each station in Nanwan Bay. "Mean" refer to average value of each station, "nT" refers to non-temperature effects, "T" refers to temperature effects, and "nT-T" represents the degree of influence.

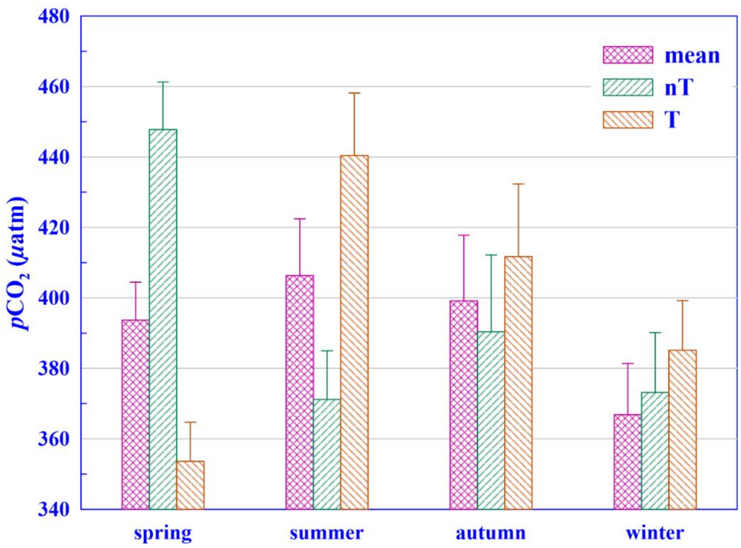

Fig. 11 Control factors of surface water pCO_2 in different seasons in Nanwan Bay. Abbreviations are as in Fig. 10. The standard deviations are shown as vertical lines.

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During the investigation period, seasonal variation in surface water pCO_2 in Nanwan Bay was mainly affected by non-temperature effects in the spring, temperature in the summer, and both effects in the autumn and winter (Fig. 11). The relationship between pCO_2, surface water temperature, and Chl *a* concentration revealed a significant correlation between pCO_2 and temperature in the summer (p<0.01; Fig. 12a), and a positive correlation between pCO_2 and Chl *a* in autumn (p<0.05; Fig. 12b). This suggests that temperature and Chl *a* are the main factors affecting pCO_2 in summer and autumn, respectively.

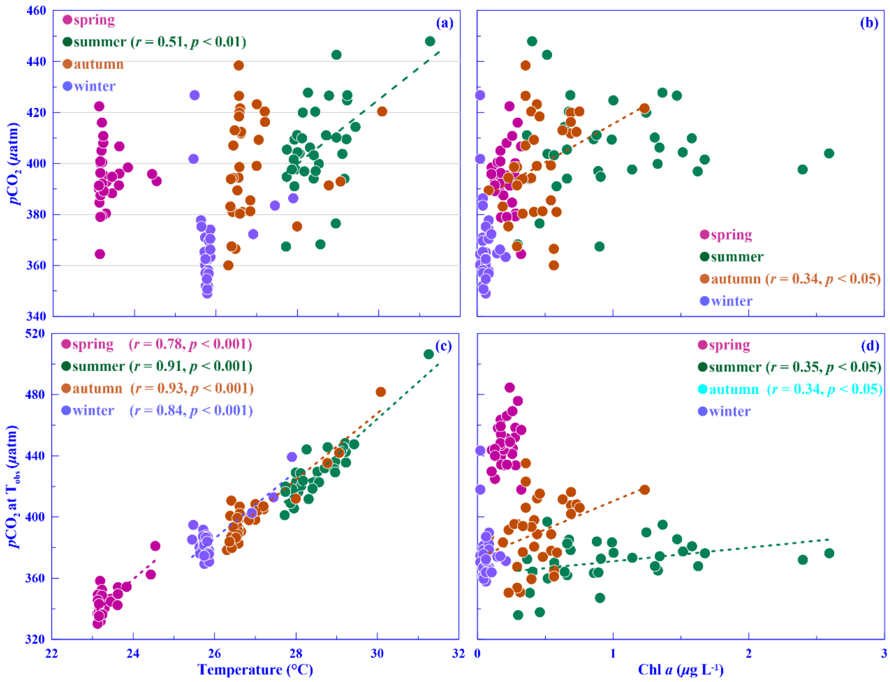

Fig. 12 Relationship between surface water pCO_2 and temperature (a), surface water pCO_2 and Chl *a* (b), surface water pCO_2 at T_{obs} and temperature (c), and surface water pCO_2 at T_{mean} and Chl *a* (d) in different seasons in Nanwan Bay.

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To evaluate the impact of temperature and non-temperature effects on seasonal variation in pCO_2, we compared the average pCO_2 under actual temperature conditions with pCO_2 values standardized to mean temperature (i.e., excluding the temperature effect). We found a significant correlation between pCO_2 and actual temperature (Fig. 12c), and a significant positive correlation between pCO_2 and Chl *a* in summer and autumn, when temperature effects were excluded (Fig. 12d; all $p<0.05$). This suggests that temperature is the main factor driving seasonal variation in pCO_2, although non-temperature effects, specifically Chl *a*, also have an influence. However, given the low *r* values, there are likely unmeasured parameters that contribute to the variation in pCO_2 over time.

Chen et al. (2004a) proposed the Degree of Nutrient Consumption (DNC) as an indicator of upwelling magnitude. This indicator is characterized by changes in temperature, DO, pH, salinity, alkalinity, DIC, nutrients, and Chl *a* during an upwelling event. These changes can alter the pCO_2 of surface waters through chemical and biological processes (Chen and Hsing, 2005).

Nanwan Bay's benthic environmental system has a regenerative effect. However, due to the high shallow water temperature and frequent stratification, regenerative nutrients cannot easily be transported to the shallows. This results in the shallow areas rarely becoming eutrophic (Leichter et al., 1996; Torréton, 1999; Wolanski and

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Pickard, 1983). Another reason for Nanwan Bay's oligotrophy is that when nutrients flow into reef areas, resident organisms quickly utilize them. Although nutrient input from outside the bay is greater than the outward flux (Su, 2009), rapid circulation of water in Nanwan Bay leads to unused nutrients being swiftly exported out of the bay. This causes oligotrophy and high benthic productivity in the area. Su (2009) reports that during spring tides, the water can be replaced in just 1.6 tidal cycles. Therefore, nutrient levels and Chl *a* may have only small influences on pCO_2 in Nanwan Bay, with temperature changes and seawater movement having a more significant impact.

3.3 Spatial distribution of $\triangle pCO_2$ and CO_2 air-sea flux The partial pressure difference between CO_2 in surface seawater and the atmosphere, denoted as $\triangle pCO_2$, indicates the direction of air-sea CO_2 exchange. When $\triangle pCO_2>0$, the seawater is supersaturated with CO_2 and releases it into the atmosphere, contributing to an increase in atmospheric CO_2 concentration (i.e., a source). On the other hand, when $\triangle pCO_2<0$, CO_2 from the atmosphere enters the seawater, acting as a sink for atmospheric CO_2. In spring, the $\triangle pCO_2$ range was between -24 and 18 µatm, with an average of -2.3 µatm (Fig. 13a). The highest value was observed near the Nuclear Power Plant outlet station. In summer, it ranged between -20 and 39 µatm, with an average of 14.3 µatm. The highest value was measured near station S7. In autumn, the range was between -29 and 37 µatm, with an average of 7.2 µatm. The highest value

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was observed near stations S10-S12. In winter, the range was between -46 and 18 µatm, with an average of -29 µatm. The lowest value occurred near stations S7-S10.

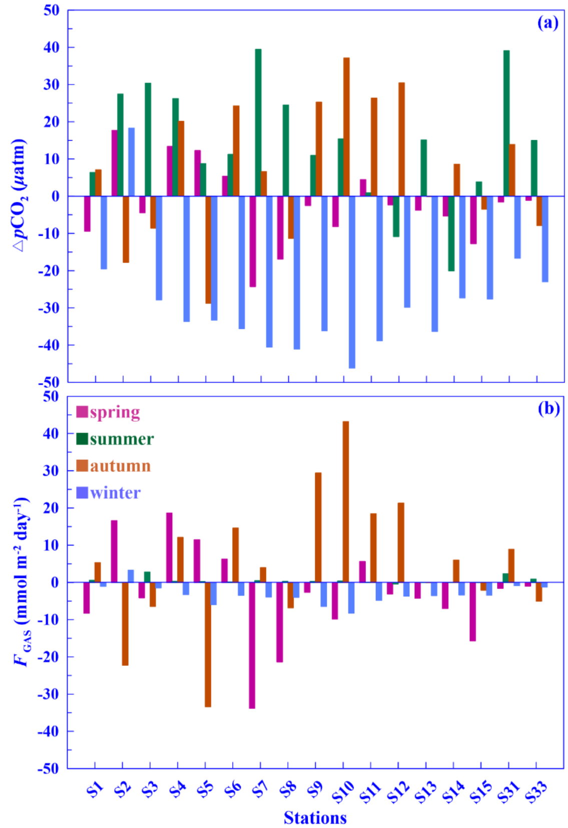

Fig. 13 Seasonal variation of (a) surface water △pCO_2 and (b) air-sea CO_2 exchange flux (F_{GAS}) at each station. Values are presented relative to the annual mean (scaled to 0).

Based on data from the Central Weather Bureau-Guanyinshan buoy, the average wind speed during the southwest monsoon season (summer) was 1.67 m s^{-1}, while during the northeast monsoon seasons (spring, autumn, & winter), it was 10, 8.1, and 2.8 m s^{-1}, respectively. Using these values, the CO_2 air-sea exchange flux in Nanwan

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Bay was calculated (Figure 13b). During spring, the CO_2 flux ranged from -33.8 to 18.7 mmol m^{-2} day^{-1} (average=-3.2 mmol m^{-2} day^{-1}). In summer, the CO_2 flux ranged from -0.4 to 2.8 mmol m^{-2} day^{-1} (average=0.5 mmol m^{-2} day^{-1}). During autumn, the CO_2 flux ranged from -33.4 to 43.2 mmol m^{-2} day^{-1}, with an average of 5.1 mmol m^{-2} day^{-1}. Finally, in winter, the CO_2 flux ranged from -8.2 to 3.3 mmol m^{-2} day^{-1}, with an average of -3.3 mmol m^{-2} day^{-1}. These findings demonstrate that wind speed is a crucial factor affecting CO_2 air-sea exchange flux.

The area between Cape Maobitou and Cape Oluanpi, covering approximately 30 km^2 (Fig. 1), showed an annual absorption of approximately -29 tC, with seasonal fluxes of -106 tC, 16 tC, 167 tC, and -107 tC in spring, summer, fall, and winter, respectively. This study's calculated value of ~-1 gC m^{-2} $year^{-1}$ as a net sink in Nanwan Bay contrasts with CO_2 sea-air flux data from other coral reef areas, such as Bermuda (14.4 gC m^{-2} $year^{-1}$; Bates et al., 2001), Okinawa (21.6 gC m^{-2} $year^{-1}$; Ohde and Van Woesik, 1999), the Great Barrier Reef (18 gC m^{-2} $year^{-1}$; Frankignoulle et al., 1996), French Polynesia (1.2 gC m^{-2} $year^{-1}$; Frankignoulle et al., 1996; Gattuso et al., 1997; Gattuso et al., 1993), and Hawaii (17.4 gC m^{-2} $year^{-1}$; Fagan and Mackenzie, 2007), all atmospheric CO_2 sources. The biogeochemistry of nearshore environments is impacted by land-based inputs, resulting in differences in $p$$CO_2$ and variations in the CO_2 sea-air flux. Prior studies showed that ship-based and satellite-

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based wind field calculations of the CO_2 sea-air flux in the East China Sea have similar trends, but there are significant differences in absolute values, with ship-based wind field calculations producing greater CO_2 exchange than satellite-based calculations (Tseng et al., 2011). Short-term and long-term wind speeds, climate conditions, and specific events can cause differences in CO_2 flux calculations due to the nonlinear empirical relationship between wind speed and gas exchange, leading to varying results (Chou et al., 2011; Evans et al., 2012; De La Paz et al., 2011).

4. Conclusions

Nanwan Bay is a region that experiences significant changes in temperature and salinity throughout the year, primarily influenced by the SCS and the Kuroshio Current. These seasonal variations have an impact on the seawater carbonate system (including pCO_2), which can also be influenced by vertical water movement and biological activity. Temperature was the most important driver of spatio-temporal differences in pCO_2, particularly at the consistently warmer outlet station. However, non-temperature effects also played a role in the spring, while the interaction between temperature and other factors was important in the autumn and winter. In terms of its impact on atmospheric CO_2 levels, Nanwan Bay acts as a sink in the spring and winter. However, in the summer and autumn, particularly during upwelling events, it becomes a source of atmospheric CO_2, releasing more CO_2 than it absorbs. Overall,

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the complex interplay of temperature, water mass origin, vertical water movement, and biological activity in Nanwan Bay has a significant impact on its carbon dioxide dynamics and its influence on atmospheric CO_2 levels.

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6. Credit author statement

This manuscript was conceptualized by PJM and CCC; CMC and HYH conducted investigations on all cruises and collected and analyzed the initial data; PJM, ABM, and CCC wrote the initial draft; all authors provided comments and edits. The authors declare that they have no conflict of interest.

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