Chapter 1

Influence of Tropical Cyclones in the Western North Pacific

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Abstract

The Western North Pacific (WNP) is the most favorable area in the world for the generation of tropical cyclones (TCs). As the most intense weather system, TCs play an important role in the change of ocean environment in the WNP. Based on many investigations published in the literature, we obtained a collective and systematic understanding of the influence of TCs on ocean components in the WNP, including sea temperature, ocean currents, mesoscale eddies, storm surges, phytoplankton (indicated by chlorophyll *a*). Some ocean responses to TCs are unique in the WNP because of the existence of the Kuroshio and special geographical configurations such as the South China Sea.

Keywords: tropical cyclone, typhoon, influence, ocean response, Western North Pacific

1. Introduction

The Western North Pacific (WNP), including its marginal seas (similarly hereafter), is the most favorable area in the world for the generation of tropical cyclones (TCs) since more TCs are born in this area than in any other region every year. The TCs in the WNP account for about one-third of TCs born in the world oceans [1, 2]. According to the data during the period of 1971–2000, the annual generation rate of TCs is 27.2 in the WNP [2]. For the activity of TCs, the South China Sea (SCS) is the most important marginal sea of the WNP where many TCs pass through and some locally form every year. Averaged from 1968 to 1988, the annual mean number of TCs in the WNP is 25.7 among which the numbers of the TCs passing over the SCS and the TCs forming in the SCS are 10.3 and 3.5, respectively [3]. The TCs in the WNP can occur in any month of the year but most of them are generated in the boreal summer (June-August) and autumn (September-November). Strong TCs are traditionally called typhoons in the WNP.

As the most intense weather system which usually causes drastic air-sea interaction, TCs play an important role in the change of ocean environment in the WNP and attract many researches focusing on this topic. By reviewing these researches, we aim to get a collective and systematic understanding of the influence of TCs in the WNP.

2. Temperature response to TCs

2.1. In the open ocean

Sea surface temperature (SST) response to TCs has been widely noticed and extensively investigated in the world oceans (e.g., [4, 5]). TCs usually cause SST to drop (SST cooling) via vertical mixing, air-sea heat flux, and advection. Vertical mixing is regarded as the most important mechanism of the cooling within the initial mixed layer and meanwhile it is responsible for the warming below the initial mixed layer [4]. Owing to the asymmetry of both wind stress and its rotation with time, the strongest SST cooling is shifted to the right of TC track in the Northern Hemisphere [4, 6]. In addition, this rightward shift or bias depends on the TC translation speed [4, 7]. Upwelling (vertical advection) enhances the SST cooling associated with TCs, especially slowly moving ones, and tends to reduce its rightward shift [4]. The exclusion of upwelling may result in the underestimation of the SST cooling as shown in Chiang et al. [8]. The cooling caused by upwelling may overwhelm the warming due to vertical mixing, resulting in a subsurface cooling, in the subsurface layer where the vertical gradient of temperature is large [6].

Mei et al. [9] demonstrated that stronger, slower-moving, and higher-latitude TCs usually cause a larger magnitude of SST anomaly. Slowly moving typhoons tend to induce larger SST drop than fast moving typhoons for the former have relatively longer residence time, producing a stronger vertical mixing and more heat loss from ocean to air. The magnitude of the SST cooling depends not only on the intensity and translation speed of typhoon, but also on the preceding thermal structure (i.e., mixed layer depth and upper-ocean stratification) of the upper ocean. Lin et al. [5] showed that a typhoon causes quite smaller SST cooling under the condition with a thicker warm upper layer than it does under the climatological condition. The scope of the SST cooling is related to the typhoon size. D’Asaro et al. [7] observed that among several TC-induced cold wakes, the smallest typhoon, Megi in Philippine Sea, produced a narrowest wake, indicating that the width of the cold wake depends on the TC size.

The SST cooling reaches a peak during the day following a TC passage and the surface cold wake restores to normal in an e-folding time of 5–20 days [9, 10]. In the WNP, Mrvaljevic et al. [11] found that a cold wake induced by Typhoon Fanapi (2010) extended to more than 80 m depth and four days later a thin and warm surface layer was formed above the cold wake. The capped wake (defined as the layer of 26–27°C) returned to normal after an e-folding time of 23 days, almost twice that of the corresponding SST cooling. Similar phenomenon has also been observed in the WNP by D'Asaro et al. [7] and their observations showed that the subsurface wake commonly occurred after the passage of several typhoons.

2.2. In the Kuroshio region

As a famous western boundary current originating from the North Equatorial Current, the Kuroshio has unique properties, characterized by high temperature, high salinity, and large velocity. In the Kuroshio region, the mixed layer as well as the thermocline is usually deeper than that in the neighboring ocean water [12, 13], which restrains the typhoon-induced SST cooling. Based on satellite remote sensing data, Wu et al. [12] found that Typhoon Nari (2001) did not induce a significant SST cooling near the Kuroshio axis where the thermocline depth is 80–100 m, while a significant SST cooling occurred in the shelf region north of the Kuroshio where the thermocline depth is 20–30 m. In addition to the deep thermocline, they suggested that the strong advection of heat along the Kuroshio also contributed to weak SST cooling near the Kuroshio axis. Analyzing the SST responses to 22 typhoons which went across the Kuroshio in the East China Sea (ECS) from 2001 to 2010, Liu and Wei [13] demonstrated that the SST cooling (averaged within 150 km on the right of each typhoon track) in the Kuroshio region ranged from 0.61°C to 4.93°C with a mean of 2.09°C while the mean SST cooling in the neighboring ocean region was 2.68°C. Wei et al. [14] reported that a SST cooling of about 3°C in the Kuroshio was produced by Typhoon Megi (2004). Their results indicated that vertical mixing is mainly responsible for the SST cooling and the kinetic energy drawn from the Kuroshio baroclinic potential energy may contribute to the cooling by enhancing local vertical mixing.

There are two special cases of temperature cooling associated with typhoons in the Kuroshio region. First, a significant temperature drop of 4°C in the upper layer from the sea surface to about 100 m was observed in the Kuroshio region, near the southeast tip of Taiwan, before Typhoon Morakot (2009) passed over [15]. This was caused by an offshore cool jet which was generated by persistent westerly wind and stretched along the Kuroshio. Secondly, typhoon-induced inshore transport of Kuroshio subsurface water in the ECS can produce upwelling and cause severe drop in the SST on the shelf beside the Kuroshio. This happened off northeastern Taiwan Island during Typhoons Gerald (1987) and Hai-Tang (2005) [16–18].

Using 10-year satellite remote sensing SST data and Argo temperature profiles, Liu and Wei [13] investigated the temperature change (warming) in the surface and subsurface of Kuroshio when the temperature recovered from typhoon-induced temperature cooling. They found that the surface temperature change (1.24°C) in the Kuroshio region is slightly smaller than that (1.39°C) in the general ocean while the subsurface temperature change (3.52°C) in the former is much larger than that (1.52°C) in the latter. Their numerical simulations indicated that the subsurface temperature change is mostly caused by downwelling related to typhoon-induced

Ekman pumping. The warm water is extracted into the subsurface from the surface by the downwelling and subsequently moves downstream with Kuroshio current.

2.3. In the SCS

The SCS is the largest semi-enclosed marginal sea of the WNP. Many TCs pass through the SCS and some are born in this sea every year [3]. Compared with the open ocean of the WNP, the SCS displays a similar but some different temperature response to typhoons due to its unique hydrological environment and complex topography.

Using the Princeton Ocean Model, Chu et al. [19] showed that Typhoon Ernie (1996) induced a significant SST cooling with a rightward bias in the SCS, similar to that in the open ocean. But it also caused some unique responses such as the SST warming in the region from southwest of Taiwan Island to northwest of Luzon Island because of the convergence between the northward coastal current west of Luzon and the Kuroshio intrusion current through the Luzon Strait.

Owing to the shallow pre-typhoon mixed layer and thermocline, Typhoons Kai-Tak (2000), Lingling (2001) and Megi (2010) generated a very large SST drop of 10.8°C, 11°C and 8°C, respectively, in the SCS [8, 20, 21]. Chiang et al. [8] suggested that upwelling (account for 62%) dominated vertical mixing (31%) in producing the SST cooling under the influence of Kai-Tak, a weak and slowly moving typhoon. Megi's translation speed was 5.5–6.9 m/s over the ocean east of the Philippines, faster than 1.4–2.8 m/s over the SCS, and the pre-typhoon mixed layer depth in these two regions was about 40 m and 20 m, respectively [22]. As a result, the SST cooling in the former was only 1–2°C, quite smaller than that in the SCS. Based on the mooring observations in the northern SCS during Megi, Guan et al. [23] showed that the temperature cooling occurred in the entire observed water column (60–360 m), which was mainly caused by typhoon-induced upwelling.

After comparing the temperature responses to TCs in the SCS and in the tropical ocean of the NWP, Mei et al. [24] found that under the influence of TCs with an identical intensity, the SST cooling in the SCS is more than 1.5 times that in the tropical ocean, which could be attributed to the shallower mixed layer and stronger subsurface thermal stratification in the former. Numerical simulations showed that Typhoon Nuri (2008) induced a stronger SST cooling in the SCS than in the open ocean of the WNP when it travelled northwestward from the open ocean to the SCS [25]. Sun et al. [25] indicated that three processes are responsible for the different regional responses. Firstly, the SCS has a thinner mixed layer, which makes it easier to entrain cooler subsurface water into the surface layer. Secondly, the cyclonic background vorticity in SCS allows stronger current shears and turbulent eddy diffusivity to be generated, however, the background vorticity in the open ocean is anticyclonic. Finally, as the typhoon moved to higher latitude in SCS, the larger Coriolis frequency in the SCS is more favorable for producing stronger wind-current resonance and then stronger inertial amplitudes and turbulence.

3. Upper ocean current response to TCs

3.1. In the open ocean

Geisler [26] used a two-layer ocean model to investigate the linear response of ocean to a moving hurricane. He concluded that inertial-gravity waves are the dominant feature of the upper ocean response if TC's translation speed exceeds the phase speed of the first baroclinic mode, and inversely the oceanic response is a barotropic, geostrophical, and cyclonic gyre with upwelling in the storm's center. In the upper ocean, a TC typically generates near-inertial currents on the right side of the TC track (looking in the moving direction) in the Northern Hemisphere and causes the resonance between wind stress vectors and currents [4]. The wind stress rotates clockwise with time on the right of the track, which is homodromous with the near-inertial currents. In contrast, the wind stress rotates anticlockwise, against the near-inertial currents, on the left. The shear of near-inertial current across the mixed layer base can deepen the mixed layer. The near-inertial currents decay rapidly within a few inertial cycles, propagating downward into the thermocline and even deep ocean as near-inertial internal waves [27, 28].

Based on the observations from drifters during 1985–2009, Chang et al. [29] illustrated the composited near-surface ageostrophic currents under all recorded TCs with various intensity levels in the WNP. Strongest current is shifted to the right of the TC track. On average, the maximum velocity increases with the intensity of the TC and that for category 4 and 5 TCs may exceed 2.0 m/s. Moreover, they found that the near-surface current responses depend on the TC translation speed as presented by Geisler [26]. Hormann et al. [30] also observed a rightward shift of maximum near-inertial currents associated with Typhoon Fanapi (2010). Observed peak current magnitudes are up to 0.6 m/s and the e-folding decay time of the strong near-inertial currents within the cold wake is about 4 days.

Analyzing the observations of the upper ocean currents induced by four category-5 typhoons [Chaba (2004), Maon (2004), Saomai (2006) and Jangmi (2008)] in the WNP, Chang et al. [31] found that besides the rightward shift, the maximum mixed layer current velocity increased with the decreasing translation speed of the four typhoons. The maximum current velocities varied from 1.2 m/s to 2.6 m/s when the translation speed of typhoons changed from 8.1 m/s to 2.9 m/s. Additionally, the maximum current velocity shows a proportional relationship with the Saffir–Simpson hurricane scale of typhoons.

3.2. Influence on the Kuroshio current

The Kuroshio is a strong western boundary current in the WNP, similar to the Gulf Stream in the North Atlantic. Under the influence of Typhoon Hai-Tang (2005), the Kuroshio axis northeast of Taiwan Island moved onto the shelf [16, 18]. The average speed of the Kuroshio surface current increased by 18 cm/s after the typhoon's passage. The northward wind in the eastern part of the typhoon caused a coastal upwelling along the east coast of Taiwan Island and an east-west sea level slope was set up. This sea level slope generated a northward geostrophic current which enhanced the Kuroshio and pushed it onto the shelf northeast of Taiwan Island [18].

Zheng et al. [15] simulated and described the response of the Kuroshio to Typhoon Morakot (2009) in the region east of Taiwan Island. When Morakot came to the Kurosho, the southward wind before the typhoon center forced the surface flow of the Kuroshio to slow down to zero. Subsequently, the surface flow strengthened as Morakot got closer to the Kuroshio. There was a sudden speedup in the surface flow due to the disappearance of strong southward wind and the release of accumulated potential energy. Before Morakot approached the Kuroshio, the Kuroshio main stream was shifted eastward for more than 1.5°. The shifted main stream returned to its original location when the typhoon center went through the Kuroshio. As the surface flow slowed down, the Kuroshio main core shifted from the surface to the depth of 50–100 m and its maximum speed decreased from more than 1.3 m/s to less than 1.1 m/s.

Using three ADCPs deployed in the area east of Taiwan Island, Yang et al. [32] revealed that during the period of 2014–2015, the volume transport of the Kuroshio was reduced by six typhoons which moved almost from southeast to northwest in the region east of the island, but intensified by two typhoons which travelled northward.

3.3. In the SCS

As frequently appearing strong weather systems, typhoons have potential influence on local currents and large-scale circulations in the semi-enclosed SCS. The large-scale circulations in the SCS are controlled mainly by strong northeast monsoon wind in winter and by weak southwest monsoon wind in summer. As a result, a basin-wide cyclonic gyre appears in winter and it is replaced by a large diploe structure (a cyclonic gyre in the north and an anticyclonic gyre in the south) in summer. Under the influence of typhoons, both the cyclonic and anticyclonic gyres are intensified in summer while the northern and southern parts of the cyclonic gyre are intensified and weakened, respectively, in winter except October and November when both are intensified [33]. Additionally, accumulative effect of typhoons can affect mean mesoscale structures: weakening the cyclonic eddy northeast of Luzon Island and enhancing the cyclonic and anticyclonic eddies off Vietnam central coast [33].

Typhoons often trigger near-inertial waves or near-inertial oscillations (NIOs) in the SCS (e.g., [23, 34, 35]). Based on ADCP observations in the northern continental shelf of the SCS, Sun et al. [35] showed that Strong NIOs were generated by Typhoon Fengshen (2008) and lasted for about 15 days after the passage of the typhoon. A similar phenomenon was induced by Typhoon Chanchu (2004) in the west of the SCS [36]. Using the observations from an ADCP mooring deployed in the northern SCS, Chen et al. [34] found that Typhoon Nangka (2009) triggered an intensive NIOs while Typhoon Linfa (2009) did not, although they both passed by the mooring in the same month of June. This is because the mooring was located in the right of Nangka's moving track but in the left of Linfa's track. As a result, the wind stress affecting the mooring location rotated clockwise during Nangka but counterclockwise during Linfa.

Yang et al. [37] demonstrated that the second baroclinic mode dominated in the NIOs appearing after the passage of Typhoon Nesat (2011) in the northern SCS.

In the SCS, the signals of internal solitary waves (ISWs) often appear in current observations during the influence periods of typhoons. Xu et al. [38] observed a series of ISWs excited by a tropical storm Washi (2005) in the Northwestern SCS. The response of the ISWs was related to direct wind forcing and remote forcing from the inertial internal waves generated by Washi. Such ISWs were also observed in the northern SCS after the passage of Typhoon Neast (2011) [39] and in other marginal seas of the WNP after a typhoon passed over [40].

3.4. In the Taiwan Strait

A strait is a special sea area that connects two different sea waters. Thus, it is of great importance for material exchange and energy transfer between the two sea waters. Here we take the Taiwan Strait joining the ECS and the SCS as an example of strait in the WNP. The Taiwan Strait is a wide and long channel with an average depth of about 60 m, bounded by the Chinese Mainland to the west and Taiwan Island to the east.

Several typhoons pass over or pass by the Taiwan Strait every year. A typhoon can induce strong southward currents and reduce or reverse northward transport through the Taiwan Strait temporarily [41–44]. Chen et al. [41] observed that the northward currents in the northern end of the strait were reversed after the passage of Typhoons Rusa and Sinlaku in 2002. Based on buoy observations and numerical model simulations, Zhang et al. [42] found that five typhoons reversed northeastward current in the middle of the Taiwan Strait and induced five southward transport events through the strait during the period of 27 August to 5 October 2005. These southward transport events were directly forced by wind stress and/or along-strait water level gradient associated with the typhoons [42]. A similar southward transport event during Typhoon Krosa (2007) was simulated by Lin et al. [44]. The observations of the drifters deployed in the Taiwan Strait revealed current reversal when Typhoon Hai-Tang (2005) traversed the strait [43].

However, the southward transport event does not always occur under the effect of typhoons. Zhang et al. [45] identified four typhoons in 2005–2009 that enhanced northward transport through the Taiwan Strait. These typhoons travelled westward in the area south of the strait or moved northward from the south to the north along special tracks, resulting in a weak southward atmospheric forcing in the early stage and a strong northward atmospheric forcing in the later stage. Meanwhile, the effect of ageostrophic process generated by the atmospheric forcing also contributed to the enhanced northward transport.

The accumulative effect of all typhoons can modify monthly mean transports during the typhoon season and even annual mean transport through the Taiwan Strait. Based on numerically simulated results, Zhang et al. [46] demonstrated that if the effects of typhoons are considered, the monthly mean transport and annual mean transport are reduced by up to 0.45 Sv and 0.09 Sv (more than 10%), respectively, compared with those without typhoons.

4. Influence on mesoscale eddies

A mesoscale eddy is one vortex with its core surrounded by closed circulations. In the ocean, it has a time scale of 10–100 days and a dimension of 10–100 km in horizontal and 100–1000 m in vertical. It can travel a distance of 100–1000 km during its lifetime of several months. According to the rotation direction of closed circulations, mesoscale eddies are classified into cyclonic (counterclockwise) and anticyclonic (clockwise) eddies. Because the temperature inside a cyclonic eddy is usually lower than that of surrounding water, the cyclonic eddy is also called cold eddy. Inversely, an anticyclonic eddy is also called warm eddy. Sea level anomaly with respect to local mean sea level is negative in a cyclonic eddy whereas it is positive in an anticyclonic eddy.

Mesoscale eddies are ubiquitous in the open ocean and marginal seas (e.g., the SCS) of the WNP. There are two eddy-rich regions in the open ocean of the WNP [47–49]: the Kuroshio Extension region (30°–40°N, 140°E–180°W) and the North Pacific Subtropical Countercurrent region (18°–25°N, 122°E–160°E). The SCS is a well-known eddy-rich marginal sea with a large area of 3,500,000 km^2 and an average depth of over 2000 m [50, 51]. In these regions, typhoons appear frequently, which may influence the structure and evolution of some eddies.

Pre-existing cyclonic eddies can be intensified after a typhoon passes over. Zheng et al. [52, 53] revealed that the SST in pre-existing cyclonic eddies markedly decreased after the passage of Typhoon Hai-Tang (2005) because cold deep water was easily brought up due to uplifted thermocline, large current shear just below the mixed layer, and small thermal inertia within the cyclonic eddies. Chiang et al. [8] showed that when Typhoon Kai-Tak (2000) passed over the West Luzon Eddy (a cyclonic eddy), it produced an unusually intense SST drop of about 10.8°C and high chlorophyll *a* (Chl-*a*) concentration in the eddy. Lai et al. [54] found that a cold eddy was intensified when Typhoon Morakot (2009) passed by the eddy, resulting in a significant cooling. Nam et al. [55] pointed out that during Typhoon Man-Yi (2007), more distinct cooling of the SST and deepening of the mixed layer occurred within a cyclonic eddy with a thin mixed layer than within an anticyclonic eddy with a thick mixed layer. Sun et al. [56] demonstrated that four typhoons in 2004 successively enhanced and enlarged a cyclonic eddy in the Kuroshio meandering region, south of Japan, due to strong air-sea interaction and typhoon-induced upwelling. Although 49 super typhoons passed over 192 cyclonic eddies in the WNP during the period of 2000–2008, only about 10% of these eddies were intensified by the typhoons [57]. Thus, Sun et al. [57] concluded that the impact of typhoons on the strength of cyclonic eddies is inefficient.

Aside from the impact on the intensity of cyclonic eddies, typhoons can directly generate cyclonic eddies via strong wind stress curls and long-time forcing. Hu and Kawamura [58] found that three TCs separately generated a cyclonic eddy in the northern SCS where they had a loop track and provided sufficient forcing time for the eddy generation. Yang et al. [59] also reported that two mesoscale cyclonic eddies were separately induced in the SCS and the WNP by long time forcing of strong wind stress curls associated with Typhoons Hagibis and Mitag in 2007, respectively. Similar phenomenon has been revealed in other studies (e.g., [57, 60]). It is interesting that these cases happened in the SCS or in the west of the WNP where typhoons

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often travel slowly and easily change their moving direction. The main reason may be that this area is close to or located at the western edge of the North Pacific Subtropical High around which clockwise atmospheric circulation steers the movement of typhoons in the WNP to a great degree.

Compared with cyclonic eddies, the mixed layer is usually thicker and the thermocline is deeper within anticyclonic eddies. As a result, anticyclonic eddies are not significantly impacted by typhoons. Lin et al. [49] demonstrated that Super-typhoon Maemi (2003) caused a weak sea surface cooling of only about 0.5°C within an anticyclonic eddy, quite smaller than that (2°C) outside the eddy, while the typhoon was rapidly intensified due to the presence of the warm eddy. They reckoned that a thick mixed layer in the anticyclonic eddy prevented deep, cold water from being entrained into the surface layer and at the same time the warm water in the thick layer contributed to the intensification of the typhoon and sustained the typhoon's intensity due to a reduced negative feedback from typhoon-induced weak sea surface cooling. The results obtained by Nam et al. [55] showed that under the influence of Typhoon Man-Yi (2007), the entrainment within an anticyclonic eddy was weak because of the thick mixed layer, resulting in small changes of temperature profiles.

5. Storm surges and large waves

A typhoon usually induces low frequency abnormal variations in sea level (namely storm surges) and high frequency large waves (commonly called typhoon waves or storm waves). The main disaster causes accompanying with typhoons encompass storm surge, huge wave, strong wind and heavy rainfall. In low-lying coastal areas and islands, storm surge together with high spring tide is most destructive.

In the open ocean, the abnormal sea level is mainly controlled by pressure drop in the center of a TC according to the inverted atmospheric pressure effect. Although the cyclonic wind tends to reduce sea level height by Ekman transport, a round water bulge is often produced in response to low pressure in the center of the cyclone. At the same time, some long waves are induced and propagate forward faster than the cyclone itself. When these free waves arrive at the shallow and wide shelf waters near Mainland China, their wave height rises because of shallow depth and then their energy is gradually dissipated by bottom friction.

When a TC comes close to the coastal area, strong wind forcing is dominant in the generation of storm surges. The combination of onshore wind and low pressure results in a maximum storm surge near the TC center. The system of topography, bathymetry and coastline plays an important role in the distribution of storm surges. The inertial oscillations in storm surges generated by a strong typhoon would disappear quickly due to overdamping in the ultra-shallow water of the Bohai Sea. Surge waves can be induced by a typhoon and propagate counterclockwise or from north to south along the coasts in the China Seas, including the Bohai Sea, Yellow Sea, ECS, and SCS [61, 62]. Edge waves may appear when a typhoon moves in parallel with the coastline. The dissipation effect of bottom friction on these coastal trapped waves is significant in the wide shelfs of the China Seas. After a typhoon makes landfall and moves away, Ekman setup can generate surges along the coasts such as Tottori coasts of the East Sea [63].

Tides affect storm surges via tide-surge interaction in the shallow waters, which is significant in the coastal areas in the WNP [64–67]. In the Taiwan Strait, the tide-surge interaction is intensified because bottom friction is enhanced by strong tidal currents and storm-induced currents in the along-strait direction [68]. As a result of the tide-surge interaction, obvious oscillations appear in storm surges. The waves accompanying with a typhoon modulate storm surges by wave radiation stress, wave-dependent bottom shear stress and surface wind stress [69, 70]. Wave setup due to the breaking of wind waves associated with a typhoon can directly contribute to storm surges in the nearshore.

Typhoon-induced storm surges and huge waves often lead to loss of lives and damage to property in coastal regions and islands with a large population. Typhoon Saomai (2006) with a maximum wind of 75.8 m/s [71] and a lowest pressure of 915 hPa was the strongest one of typhoons making landfall on Mainland China in the past 50 years. When it made landfall at the coast of Zhejiang Province, the central pressure was still 920 hPa and a momentary maximum wind of 68 m/s was recorded. During its influence period, the highest wave recorded by a buoy (27.5°N,122.53°E) in the ECS was 8.6 m. Saomai induced devastating storm surges and a maximum storm surge of 4.01 m was observed at Aojiang station, Zhejiang Province. As reported in the Bulletin of Marine Disaster of China, the storm surges superposing on a high spring tide caused 230 persons dead and 96 missing, and a damage of more than 7 billion yuan to property. Another devastating typhoon, Haiyan, hit Philippines islands on 8 November 2013. Before its landfall, the maximum sustained wind was nearly 88 m/s and the momentary maximum wind was about 105 m/s. Storm surges induced by Typhoon Haiyan inundated a large coastal area in Philippines and caused a catastrophic damage [72, 73]. The inundation depth was up to more than 7 m in Tacloban city. During the typhoon, 6300 persons were killed and 1061 were still missing [72].

Storm surges, together with tides, mainly determine extreme sea levels and their spatial pattern along the coasts of the WNP [66, 74]. For some places where tidal range is small and tide-surge interaction is weak, storm surges play a more important role in the extreme sea levels [75]. Storm surges also contribute to the local increasing trend of annual maximum water level at coastal stations [76].

6. Biogeochemical and biological responses

TCs definitely have significant impacts on biogeochemical and biological processes in the oceans. Biological responses, including associated biogeochemical aspects, to typhoons have been observed in the WNP by ship surveys, buoys and satellite remote sensing data (e.g., [77–79, 20]), which will be discussed in the following text. Although typhoons may have direct and indirect effects on the carbon cycling in the WNP (e.g., [80–82]) and potentially contribute to global climate change, detailed contents of these effects are not included here. Many investigations show that various typical waters (open ocean waters, marginal seas, shelf waters) display some different biological responses to typhoons with unique mechanisms.

6.1. In open ocean water

In the deep ocean water of the WNP, Merritt-Takeuchi and Chiao [83] found an obvious growth of biological substances after the passage of Typhoon Xangsane (2006) which brought nutrients from the depths to the surface layer. There is a negative correlation between SST and Chl-*a* with the correlation coefficient of –0.67. Salyuk et al. [84] demonstrated that 81% of 123 TCs increased Chl-*a* concentration after the TC passage, which could last about 2 weeks. Based on multiple satellite observations and numerical experiments, however, Lin [85] showed that only two cases (18%) among 11 typhoons in 2003 induced phytoplankton blooms in the WNP.

The biological response of the upper ocean depends on the translation speed, spatial size, moving track and intensity of a typhoon and pre-existing ocean conditions. There is a positive correlation between Chl-*a* concentration and wind speed [83, 86]. Lin [85] found that 9 of 11 typhoons in 2003 did not induce phytoplankton blooms in the WNP. Among the 9 typhoons, eight typhoons had relatively small size, fast translation speed and insufficient wind intensity, and then they only caused weak responses in the ocean with a deep nutricline/mixed layer. Owing to the presence of warm eddy, the other typhoon, Maemi, was not able to induce phytoplankton bloom although it was very strong with a maximum wind of about 77 m/s. Typhoon Haitang (2005) did not enhance Chl-*a* in the sea area east of Taiwan because of high translation and a shallower MLD than the nutricline [87].

6.2. In the SCS

Typhoons are very active in the SCS. They often trigger phytoplankton blooms in this oligotrophic sea. Sun et al. [86] found that Typhoon Hagibis (2007) with a steep turn track had a significant effect on the surface Chl-*a* concentration in the SCS. Its long forcing time is favorable for the enhancement of Chl-*a* concentration although it is just a category 1 typhoon. Typhoon Kai-Tak (2000) increased surface Chl-*a* concentration by 30-fold on average in the SCS during three days [49]. As a result, Kai-Tak alone contributed 2–4% of the SCS's annual new primary production. A similar enhancement of Chl-*a* concentration appeared separately after typhoons Lingling (2001) and Damrey (2005) in the SCS [20, 88]. Zhao et al. [89] conservatively estimated that typhoons accounted for 3.5% of annual primary production in the SCS during the period of 1945–2005. Chen et al. [90] estimated that 5–15% of annual new primary production in the SCS was attributed to typhoons during 2003–2012.

Eddies are ubiquitous in the SCS. Some investigations have shown that Chl-*a* concentration enhanced by typhoons may be associated with cold core eddies in this area (e.g., [22, 91]). Chen and Tang [92] found that in the SCS a cold eddy was generated where Typhoon Linfa (2009) hovered, and subsequently an eddy-feature phytoplankton bloom was induced. The consistence between the bloom pattern and the cold eddy suggested that the typhoon-induced eddy potentially brought nutrients upward to the surface water, which contributed to the bloom.

Chen et al. [90] compared typhoon-enhanced primary production in the SCS with that in the subtropical ocean water of the WNP during the period of 2003–2012. Their results showed that the annual mean carbon fixation induced by typhoons was more in the SCS than in the ocean water. This is because the mixed layer is thicker and the nutricline depth is deeper in the latter in spite of its larger area and more super typhoons appearing there.

The biological response to typhoons can happen not only in the surface layer but also in the subsurface. Ye et al. [93] found that a Chl-*a* bloom appeared in the subsurface layer (20–100 m depth) of the SCS after the passage of Typhoon Nuri (2008) and lasted for three weeks. This subsurface bloom was stronger and its life was longer than the synchronous surface Chl-*a* bloom. Previous estimates of the contribution from typhoons to annual primary production in the SCS were mostly based on the results in the surface layer using remote sensing data. On this aspect, these estimates probably underestimate the actual contribution of typhoons.

6.3. In continental shelf waters

There are many wide and shallow continental shelf regions in the WNP, mostly located in the China Seas. The ecosystem in these shelf regions can become more productive after typhoons pass through [77, 78, 87].

The continental shelf of the southern ECS, northwest of Taiwan, is a typical oligotrophic and strong stratification area during summer. However, Shiah et al. [78] found that all chemical and biological parameters measured in a survey after the passage of Typhoon Herb (1996) were much larger than normal summer conditions. The typhoon caused primary production, particulate organic carbon (POC) concentration, bacterial production, and biomass to increase by at least two-fold. Their results indicated that wind mixing, re-suspension and terrestrial runoff associated with the typhoon were responsible for these responses. Based on multi-satellite observations, Chang et al. [16] showed that Typhoon Hai-Tang (2005) induced upwelling and increased Chl-*a* concentration in this shelf region, persisting for more than 10 days. The upwelling was likely caused by Ekman pumping due to strong typhoon wind and wind-driven shoreward intrusion of Kuroshio water along the shelf break. Siswanto et al. [94] also demonstrated that long-lasting southerly winds accompanying with typhoons can force Kuroshio current axis to move toward the shelf of the southern ECS, inducing upwelling. This process uplifts nutrients and then increases new productivity, which contributes 0.6–11.8% of the summer-fall new productivity in the ECS. In addition to the effects on primary production represented by Chl-*a*, typhoons may have influence on phytoplankton composition. Chung et al. [95] observed that a diatom bloom was induced in the southern ECS after the passage of Typhoon Morakot (2009) and the species composition was changed also.

Zhang et al. [96] observed that the surface Chl-*a* concentration in the continental shelf southeast of Hainan Island was increased by 38.5% after the passage of tropical storm Washi (2005). Chen et al. [97] showed that the primary productivity and nitrate-uptake-based new production in the upstream Kuroshio close to southern Taiwan were enhanced after the passage of three typhoons in 2007 by riverine mixing associated with the typhoons. After Typhoon Malou (2010) passed over Sagami Bay in the central part of Japan, both Chl-*a* concentration and bacterial abundance increased at an inshore station due to terrestrial runoff and sediment resuspension while only Chl-*a* concentration rose at an offshore station due to terrestrial runoff [98].

There are some contrary cases. Zhou et al. [99] found that Typhoon Fengshen (2008) destroyed the pre-existing upwelling and meanwhile caused freshwater plume to spread in the continental shelf of the northeastern SCS, which led to nutrient-limited conditions. As a result, a negative phytoplankton growth rate appeared a week after the typhoon passage. Zhao et al. [100] also found that a sharp decrease of Chl-*a* concentration was caused by Typhoon Matsa (2005) in the nearshore area of the ECS.

Therefore, biological response to typhoons in the continental shelfs, especially nearshore areas, is very complex. It changes with different sea areas and different time periods, depending on specific circumstances such as geographical configuration, hydrological environment as well as the existence of rivers or not.

7. Summary

TCs usually change temperature and salinity vertical profiles by vertical mixing, upwelling and heat flux. Generally, the vertical mixing causes surface temperature cooling and subsurface warming. The upwelling brings cold deep water upward and makes the water temperature in the upper layer to decrease, which suppresses the subsurface warming and enhances the surface cooling. Surface sea water loses heat into air by heat flux and its temperature further decreases.

TCs typically generate energetic transient currents in the upper ocean, which modulates mean circulations such as the Kuroshio. The upper layer flow of the Kuroshio can be slowed or shut down temporarily just before a TC approaches. It can shift the location of the Kuroshio main stream. In the SCS, TCs affect both large scale and mesoscale circulations, and generate NIOs and ISWs. They change the instantaneous volume transport through a strait like the Taiwan Strait and then accumulatively modify its seasonal and annual mean transports. TCs not only influence pre-existing oceanic mesoscale eddies but also generate cyclonic eddies when they travel slowly.

TCs produce storm surges and huge waves which can flood coastal low-lying areas and threaten coastal structures. Because of wide continental shelves and flexural coastlines, the combination of onshore wind and low pressure accompanying with typhoons easily causes devastating storm surges along the coasts of the continent and islands in the WNP. Additionally, Ekman setup and wave setup also contribute to the storm surges in some shallow waters. In the continental shelf regions of the China Seas, coastal trapped waves are often induced and tide-surge interaction is significant in the shallow coastal sea areas and in the Taiwan Strait.

As the most intense case of air-sea interaction, TCs can cause biogeochemical and biological responses in the ocean. The vertical mixing and upwelling associated with TCs transport nutrients from the depths to the surface layer. As a result, phytoplankton blooms are often triggered and contribute to the primary productivity in the WNP, especially in the SCS. In continental shelf waters, the biological responses are more complex because of pre-existing upwelling, TC-enhanced freshwater plume, riverine mixing, terrestrial runoff, and sediment resuspension.

Since we focus on the influence of TCs in the WNP, some important aspects are not included here, such as the feedback of ocean to TCs. The accumulative effects of TCs on climate are not considered, either, because they are global and indirect, not limited to the WNP, in a long time scale. Regarding the complexity and extensiveness of ocean responses to TCs, some questions remain open and more observations and investigations are necessary to explore their answers in the future.

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