

REVIEW

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Upper ocean response to tropical cyclones: a review

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Abstract

Tropical cyclones (TCs) are strong natural hazards that are important for local and global air–sea interactions. This manuscript briefly reviews the knowledge about the upper ocean responses to TCs, including the current, surface wave, temperature, salinity and biological responses. TCs usually cause upper ocean near-inertial currents, increase strong surface waves, cool the surface ocean, warm subsurface ocean, increase sea surface salinity and decrease subsurface salinity, causing plankton blooms. The upper ocean response to TCs is controlled by TC-induced mixing, advection and surface flux, which usually bias to the right (left) side of the TC track in the Northern (Southern) Hemisphere. The upper ocean response usually recovers in several days to several weeks. The characteristics of the upper ocean response mainly depend on the TC parameters (e.g. TC intensity, translation speed and size) and environmental parameters (e.g. ocean stratification and eddies). In recent decades, our knowledge of the upper ocean response to TCs has improved because of the development of observation methods and numerical models. More processes of the upper ocean response to TCs can be studied by researchers in the future.

Keywords: Tropical cyclone, Upper ocean, Dynamic response, Temperature and salinity variation, Marine biology

Introduction

Tropical cyclones (TCs) are strong natural hazards that are generated and developed in the ocean. TC wind as well as the TC-induced currents and waves usually damages offshore platforms and vessels, erode coastlines, threaten coastal areas and causes economic and personnel losses (Keim et al. 2007; Wang and Oey 2008; Han et al. 2012; Sun et al. 2013; Zhang et al. 2016b; Li et al. 2018). In recent decades, TC track forecasts have steadily improved because of the development of numerical models (Krishnamurti et al. 1999; McAdie and Lawrence 2000; Bender et al. 2007; Cangialosi and Franklin 2013; Ruf et al. 2016; Montgomery and Smith 2017), while the forecast skill of TC intensity has improved slightly (Cangialosi and Franklin 2013), partly because of inadequate observations and modelling of TC inner cores (Ruf et al.

2016) and TC–ocean interactions (Yano and Emanuel 1991; Montgomery and Smith 2014, 2017). Sea surface temperature is important for TC intensity. For example, a simple coupled model of an axisymmetric hurricane model and one-dimensional ocean model can significantly improve the intensity predictions if the negative feedback of the sea surface temperature to TCs is taken into account (Emanuel 1999). Sea surface temperature cooling greater than 2.5 °C is considered not conducive to TC strengthening (Emanuel 1999) and may even weaken TCs (Schade and Emanuel 1999; Lin et al. 2008). The sea surface cooling caused by a pre-existing TC alters the track and intensity of a subsequent TC (Baranowski et al. 2014; Wu and Li 2018). Note that salinity stratification can reduce sea surface cooling in favour of TC rapid intensification, and this effect increases significantly as the TC intensification rate increases (Balaguru et al. 2020).

TC is also important for the ocean environment, TCs import kinetic energy into surface waves, surface currents and gravitational potential energy, which contributes to ocean diapycnal diffusivity and ocean circulation

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(Liu et al. 2008). TCs can also change the local ocean heat uptake (Emanuel 2001) and contribute to global ocean heat transport (D'Asaro et al. 2007; Srivier and Huber 2007; Korty et al. 2008; Pasquero and Emanuel 2008; Hu and Meehl 2009; Fedorov et al. 2010). Some research even considers TCs is important to maintain the permanent El Niño in the early Pliocene epoch (Fedorov et al. 2010). In addition to meridional heat transport, TCs can influence west wind bursts (Lian et al. 2018, 2019), enhance eastward-propagating oceanic Kelvin waves in the tropical Pacific (Wang et al. 2019b) and modulate the occurrence and development of the El Niño–Southern Oscillation. TCs usually cause plankton blooms, which contribute to local long-term primary productivity (Mooers 1975; Foltz et al. 2015). Research shows that TCs increase 20–30% of the primary productivity in the South China Sea every year on average (Lin et al. 2003b; Tang et al. 2004a, b; Sun et al. 2010) also explains 22% of the interannual variability in seasonally averaged (June–November) chlorophyll concentration in the western subtropical North Atlantic (Foltz et al. 2015).

In brief, understanding the ocean response to a TC not only increases the TC forecast skill but also enriches our knowledge of local and global variation of ocean environment.

Ocean response to a tropical cyclone

Current response

The strong TC wind stress arouses upper ocean current response, which usually biases to the right (left) side of TC track in Northern (Southern) Hemisphere, because of better wind-current resonance (Price 1981, 1983; Price et al. 1994; Sun et al. 2015; Zhang et al. 2020b). The wind-current resonance is controlled by non-dimensional TC translation speed which is the ratio of local inertial period to the TC residence time (Zhang et al. 2020b). Generally, a TC causes internal wake in its lee when it moves faster than the first baroclinic wave speed, and the main response is almost centred under the TC and the wake is relatively inconspicuous when a TC moves slower than the speed of first baroclinic wave (Geisler 1970). The current response is a stable Ekman response with surface (bottom) cyclonic divergence (convergence) when TC is stationary (Lu and Huang 2010), and there was weak current response in the lee of the Ekman-like divergence when TC moves slowly (Zhang et al. 2020b). Because the translation speed of most TCs is greater than the local first baroclinic wave speed (e.g. Zhang et al. 2020b), we usually find a near-inertial current response after the TC passage (Pollard 1970; Maeda et al. 1996; Firing et al. 1997; Jarosz et al. 2007; Xu et al. 2019). The upper ocean near-inertial current response to a TC can be divided into mixed layer current and thermocline current (Price 1981;

Price et al. 1994; Zhang et al. 2016a, 2019). TC wind directly drives the mixed layer current, then the divergence and convergence of mixed layer current caused hydrostatic pressure anomaly that drives the thermocline current (Price et al. 1994). The phases of the mixed layer current and thermocline current are nearly uniform within themselves, while there is an angle between mixed layer current and thermocline current (Sanford et al. 2007, 2011; Prakash and Pant 2016; Zhang et al. 2016a), which depends on the ratio of TC translation speed to baroclinic wave speed (Geisler 1970). There is a transition layer between the mixed layer current and thermocline current, with the current phase turning clockwise as the depth increases (Price et al. 1986; Sanford et al. 2011; Zhang et al. 2016a). Velocity shear in transition layer is considered as the primary mechanism for deepening of upper ocean mixed layer during a TC (Glenn et al. 2016; Seroka et al. 2017; Yang et al. 2019). The TC-induced near-inertial current corresponds to upwelling and downwelling, with the transition of the upwelling (downwelling) branch to the downwelling (upwelling) branch being slow and moderate (quick and intense) (Greatbatch 1983, 1984, 1985). If the inertial period signal is removed, there is net mixed layer divergence and upwelling in the right rear quadrant of the TC, as well as net downwelling around the net upwelling zone (Zhang et al. 2018a, b). Relative to open ocean, the current response to a TC in the marginal sea is more complicated because of the secondary local circulation due to the shallow ocean bottom and coastal wall (Halliwell et al. 2011; Glenn et al. 2016; Seroka et al. 2016). TCs also have import positive vorticity into ocean, which intensifies ocean cyclonic eddies (Walker et al. 2005) and alters the three-dimensional structure of eddies (Sun et al. 2014; Lu et al. 2016) or even generates new cyclonic eddies (Chen and Tang 2012; Sun et al. 2014). On the other hand, the circulation of eddies also modulates the TC-induced convection and vertical advection in upper ocean (Jaimes and Shay 2015; Liu et al. 2017).

After TC passage, the current response decays through dispersion and propagation of near-inertial waves (Gill 1984; Park et al. 2009) with an e -folding time from days to weeks (Chen et al. 2013; Yang and Hou 2014). The decay of current is not monotonous because different orders of near-inertial baroclinic waves occasionally resonate again and re-intensify the mixed layer current (Gill 1984). Note that the current velocity to the right side of the TC track decays faster than that to the left side in the Northern Hemisphere (Zhang et al. 2016a; Wu et al. 2020a). The dispersion of waves also results in the tilting of isophasal lines of near-inertial current (Gill 1984; Zhang et al. 2016a). In general, the near-inertial waves propagate in the ocean with a vertical scale of approximately

100–300 m (Kundu 1976; Yang and Hou 2014; Alford et al. 2016) and contribute to turbulent mixing (Qi et al. 1995; Zhai et al. 2009; Jochum et al. 2013). The TC-induced current response can reach the ocean bottom, which may be a major driver of sediment dynamics of continental shelves worldwide (Larcombe and Carter 2004; Galewsky et al. 2006; Dail et al. 2007; Liu et al. 2012), impacting benthic and pelagic habitats by changing water column turbidity or modifying seabed physical characteristics (Hearn and Holloway 1990; Drost et al. 2017). The near-inertial current caused by TCs usually has a blueshift of 1–20% relative to the local inertial frequency, along with slow downward propagation of energy and upward propagation of phase (Pollard 1980; Smith 1989; Yang and Hou 2014). In addition to near-inertial waves, TCs cause inertial waves at a frequency that is two times and three times the local inertial frequency, resulting in energy cascade and dissipation (Niwa and Hibiya 1997; Meroni et al. 2017). Mesoscale ocean processes alter local relative vorticity, which changes the effective planetary vorticity; then, TC-induced near-inertial waves propagate downward more easily in an anticyclonic eddy (Zhai et al. 2005; Guan et al. 2014) and tend to be trapped in a region with more negative vorticity than its surroundings (Oey et al. 2008; Jaimes and Shay 2010) (Fig. 1).

Surface wave response

TC wind also arouses strong surface waves. Sea surface wave height is a function of radial distance from the TC centre by empirical relationships (Young 1988; Wang et al. 2005; Young and Vinoth 2013) and can reach more than 10 m (Zhang et al. 2016a; Drost et al. 2017; Zhang et al. 2018a, b). Surface wave propagation is complicated, and multiple-wave systems are frequently observed. Previous studies typically assumed that TCs impact a region more than 10 times the radius of the maximum wind (Young 2006; Beeden et al. 2015; Esquivel-Trava et al. 2015). TC-induced wave spectra rapidly evolve and vary spatially by radius away from the centre and quadrant of the TC (Moon et al. 2003; Young 2003; Fan et al. 2009; Collins et al. 2018). The TC-induced wave spectra are often bimodal, sometimes trimodal, directional wave spectra (Wright et al. 2001), the waves are asymmetrical, and the directional spectra possess unique characteristics in each quadrant (Hu and Chen 2011; Esquivel-Trava et al. 2015). In reference to the TC heading, single-wave systems propagating towards the left and left-front are usually observed in the front half of the TC coverage area, and multiple-wave systems are generally observed in the back and right quarters outside the radius of maximum wind, while the directional differences and locations of

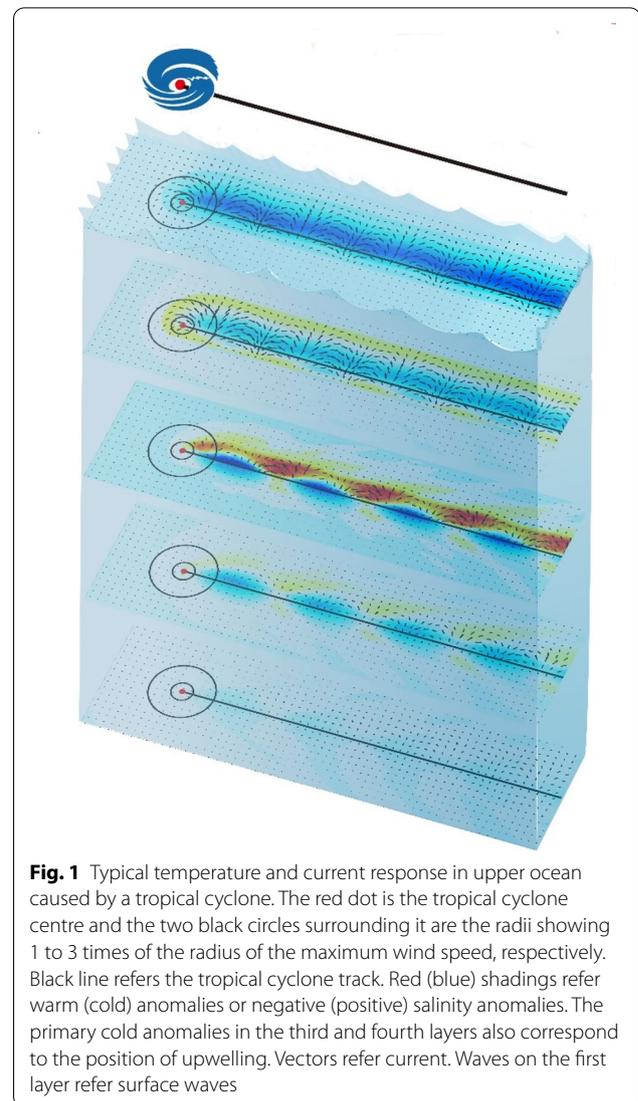


Fig. 1 Typical temperature and current response in upper ocean caused by a tropical cyclone. The red dot is the tropical cyclone centre and the two black circles surrounding it are the radii showing 1 to 3 times of the radius of the maximum wind speed, respectively. Black line refers the tropical cyclone track. Red (blue) shadings refer warm (cold) anomalies or negative (positive) salinity anomalies. The primary cold anomalies in the third and fourth layers also correspond to the position of upwelling. Vectors refer current. Waves on the first layer refer surface waves

multisystem spectra are Gaussian distributions (Hwang and Walsh 2018).

There is misalignment of wind and surface waves during a TC (Fan et al. 2009; Wang et al. 2019a). Swells dominate the surface waves at the front of and outside the central typhoon region (Xu et al. 2017b), and the wave field is more asymmetric than the corresponding TC wind field, mainly due to the “extended fetch”, which exists to the right of a translating TC in the Northern Hemisphere (Young 2003). TC-induced surface currents can reduce the fetch and inhibit the growth of surface waves (Wu et al. 2020b). Nonlinear wave–wave interactions efficiently transfer wave energy from high frequencies to low frequencies and prevent double-peak structures occurring in the frequency-based spectrum (Xu et al. 2017b).

TC-induced surface waves modulate the air–sea surface conditions and fluxes. TC-induced surface waves increase sea surface roughness (Donelan 2004; Makin 2005; Soloviev et al. 2014; Li et al. 2016; Tian et al. 2020) and reduce wind speeds (Olabarrieta et al. 2012). The wind-wave coupling deepens inflow layer, enhances boundary inflow outside the radius of maximum wind and increases the TC intensity (Lee and Chen 2012). Surface wave breaking during TCs also causes a large number of sea spray droplets (Zhang et al. 2011, 2012) in whitecaps and whipping spumes from the tips of waves, which is believed to significantly influence momentum transfer and contributes to the drag coefficient levelling off (or decreasing) at high wind speeds during a TC (Powell et al. 2003; Donelan 2004; Soloviev et al. 2014; Zhang and Song, 2018). Sea spray also influences the air–sea heat flux (Andreas and Mahrt 2016; He et al. 2018; Sun et al. 2019). The latent and sensible heat transfer coefficients are constant at low wind speeds and increase sharply when wind speed at the height of 10 m is greater than 35 m/s (Komori et al. 2018). The air–sea gas transfer also increased significantly due to the surface wave breaking (Iwano et al. 2013; Krall and Jähne 2014; Liang et al. 2020).

TC-induced breaking and unbreaking surface waves also contribute to turbulence in the upper ocean and deepening of the mixed layer (He and Chen, 2011; Toffoli et al. 2012; Aijaz et al. 2017; Stoney et al. 2017; Zhang et al. 2018a, b). The wave-breaking-induced acceleration transfers momentum from surface waves to surface currents and also contributes to sediment transport (Prakash and Pant; 2020). Non-breaking surface wave-induced mixing in numerical model improves the simulations of sea surface temperature and TC track (Guan and Zhao 2014; Li et al. 2014; Aijaz et al. 2017; Stoney et al. 2017). The Craik-Leibovich vortex force, which is the interaction between Stokes drift of surface waves and Eulerian current vorticity, causes Langmuir turbulence (Craik and Leibovich 1976), enhances turbulence entrainment and deepens mixed layer during a TC (Sullivan et al. 2012; Rabe et al. 2015; Reichl et al. 2016a,b; Zhang et al. 2018b; Wang et al. 2018, 2019). The Langmuir cell is roughly aligned with wind and Langmuir turbulence intensity is reduced by wind-wave misalignments during a TC (Wang et al. 2019a, b). Recent researches indicate that parameterization of Langmuir turbulence can improve the simulation of upper ocean temperature and current response during a TC (e.g. Sullivan et al. 2012; Reichl et al. 2016a,b; Blair et al. 2017). The Coriolis–Stokes force also increases the cold upwelling in a slow moving TC and modulates the horizontal advection of upper ocean cold wake (Reichl et al. 2016a; Zhang et al. 2018b). Langmuir circulation generates high-frequency internal waves, induces near-inertial currents at the mixed layer

bottom (the transition layer) and transports more near-inertial energy into deeper layers (e.g. Polton et al. 2008)

Temperature and salinity response

TC deepens the upper ocean mixed layer, cools the sea surface and warms the subsurface (Price 1981, 1983, 1994; Jacob et al. 2000; Zedler et al. 2009; Sanford et al. 2011; Yang et al. 2015; Chen et al., 2020), which is called the “heat pump” effect (Srifer and Huber 2007; Zhang et al. 2016a). Sea surface also lose heat through air–sea heat flux, but it is not as important as the mixing effect for the sea surface cooling (Price 1981; Zhang et al. 2016a). Sea surface cooling usually biased to the right (left) side of the TC track in the Northern (Southern) Hemisphere, and the amplitude of sea surface cooling is usually 1–6 °C (Price 1981; Zedler et al. 2002; Lin et al. 2003a; Black et al. 2007; D’Asaro et al. 2007), sometimes even reaching ~11 °C, resulting in reverse of air–sea surface sensible and latent heat flux (Glenn et al. 2016). The upwelling branch of the near-inertial pumping weakens the subsurface warm anomaly or even turns it to cold anomaly, while downwelling branch intensifies the subsurface warm anomaly. Subsurface warm anomaly caused by mixing can reach as much as ~4 °C, and usually be modulated by the TC-induced near-inertial pumping (Zhang et al. 2016a, 2019). After removing inertial period signal, TC caused a net upwelling with a net cooling in the right rear quadrant of TC, and net downwelling with net warming around the net cooling zone (Zhang et al. 2018a, 2019), which is called “cold suction” effect. During the TC relaxation stage, the air–sea heat flux dominates the upper ocean thermal response, which mainly recovers the sea surface cold anomaly through solar radiation (Price et al. 1986). Research shows that sea surface temperature usually recovers back to its original value in several days to several weeks (Hazelworth 1968; Price et al. 1986, 2008; Emanuel 2001; Hart et al. 2007; Wang et al. 2016), with an e-folding time of approximately one week (Jansen et al. 2010; Dare and McBride 2011), occasionally cooling again during recovery (Price et al. 2008). Subsurface ocean has no contact with air, so it usually recovers slower than the sea surface (Emanuel 2001; Wang et al. 2016).

The characteristics of the upper ocean temperature response to a TC are affected by the TC intensity, size and translation speed (Anthes and Chang 1978; Emanuel et al. 2004; Zhu and Zhang 2006; Samson et al. 2009; Wang et al. 2016; Lin et al. 2017). For example, a stronger TC produces more cooling up to Category 2, but TCs in Categories 3–5 produce less or approximately equal cooling (Lloyd and Vecchi 2011). Argo float observations show that the subsurface warm anomaly is comparable to the near-surface cold anomaly during strong TCs

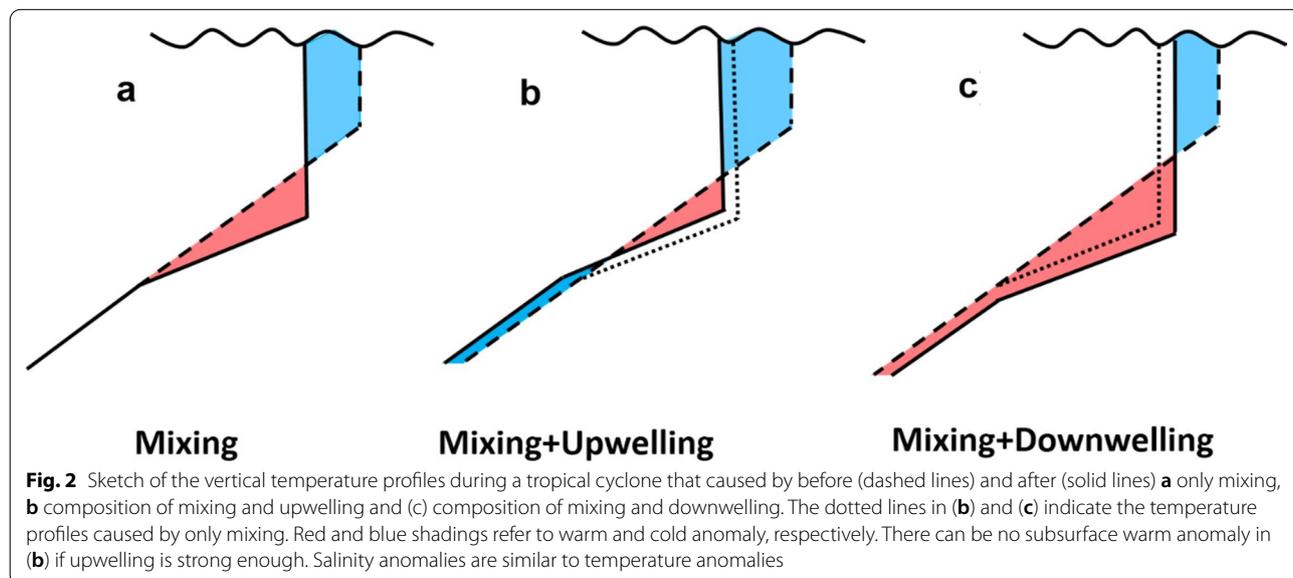
(\geq Category 4), while subsurface warming is not detectable and near-surface cooling is still significant during weak TCs (\leq Category 3), indicating that air–sea heat exchange and upwelling seem to play a somewhat greater role during weak TCs (Park et al. 2011). The sea surface cooling is quasi-symmetric for slow-moving (<6 m/s) TCs and becomes asymmetric for fast-moving TCs (Samson et al. 2009). The background ocean condition also contributes to the upper ocean temperature response during a TC. For example, there is cold upwelling (warm downwelling) in the core of an anticyclonic (cyclonic) eddy, which intensifies (weakens) sea surface cooling (Jaimes and Shay 2009; Liu et al. 2017; Wu and Li 2018; Ning et al. 2019).

Similar to temperature response, TCs usually increase sea surface salinity and decrease subsurface salinity both within 1 psu, which bias to the right (left) side of TC track in Northern (Southern) Hemisphere (Bond et al. 2011; Girishkumar et al. 2014; Domingues et al. 2015; Zhang et al. 2016a; Abernathy and Haller 2018); the positive sea surface salinity sometimes even reaches 1.5–3 psu (Chaudhuri et al. 2019). However, TC precipitation usually weakens positive sea surface salinity anomaly (Girishkumar et al. 2014; Liu et al. 2020) and causes negative sea surface salinity anomaly to the left (right) side of TC track in the Northern (Southern) Hemisphere (Grodsky et al. 2012; Liu et al. 2014). The freshness of the sea surface by precipitation increases the upper ocean stratification and weakens the TC-induced mixing (Jourdain et al. 2013; Vissa et al. 2013; Liu et al. 2015, 2020), which restricts the sea surface cooling and the negative TC–ocean feedback (Balaguru et al. 2016). There is a barrier layer if the upper ocean isosaline layer is shallower than

the isothermal layer, which also prohibits the deepening of the upper ocean mixed layer caused by a TC (Balaguru et al. 2012; Liu et al. 2015; Yan et al. 2017). Research shows that the upper ocean salinity response can persist about 10–12 days (Girishkumar et al. 2014). The background ocean condition (e.g. eddies) also contributes to the upper ocean salinity response during a TC. For example, the upwelling (downwelling) due to anticyclonic (cyclonic) eddies increases (decreases) upper ocean salinity (Jaimes and Shay 2009; Liu et al. 2017) (Fig. 2).

Biological response

TCs induce phytoplankton blooms and primary production increase, which is mainly attributed to the increased nutrient supply in the euphotic zone induced by vertical mixing (or entrainment) and upwelling during a TC (Mooers 1975; Morimoto et al. 2009; Siswanto et al. 2009; Zheng et al. 2010; Chiang et al. 2011; Shibano et al. 2011; Hung et al. 2013; Huang and Oey 2015) and ocean restratification after the TC (Huang and Oey 2015; Lin and Oey 2016). The chlorophyll increases after a TC usually ranges from 5 to 91% (Babin et al. 2004; Zhao et al. 2017; Xu et al. 2017a), while a lingered slow-moving TC (Kai-Tak in year 2000) can even triggered 30-fold of surface chlorophyll-a concentration (Lin et al. 2003b). In Northern (Southern) Hemisphere, the chlorophyll increases usually biases to the right (left) side of the TC track (Lin et al. 2003b; Babin et al. 2004; Yin et al. 2007; Hanshaw et al. 2008; Shang et al. 2008; Zhao et al. 2008; Zheng et al. 2010; Shibano et al. 2011), although the rightward (leftward) bias is not obvious or may even occur towards the left (right) side of the TC track (Zheng et al. 2010; Shibano et al. 2011). The amplitude and scope of surface



plankton blooms depend not only on the TC characteristics but also on the ocean background conditions (Lin et al. 2003b; Zhao et al. 2008; Chen and Tang 2012; Shang et al. 2015; Xu et al. 2017a). For example, weak and slow-moving TCs induce phytoplankton blooms with higher chlorophyll-a concentrations, while strong and fast-moving TCs induce blooms over a larger area (Zhao et al. 2008). A pre-existing cold core eddy plays an important role in the increase in chlorophyll-a concentration by TCs (Chen and Tang 2012; Shang et al. 2015; Xu et al. 2017a; Jin et al. 2020), and the concentration of pre-existing chlorophyll-a in cold core eddies is approximately 25–45% (8–25%) of that of the post-existing chlorophyll-a in cold core eddies for relatively high (low) TC transition speeds (Shang et al. 2015). The biological response in coastal regions is more complicated than that in the open ocean (Pan et al. 2017). TC-induced mixing, enhanced terrestrial runoff and resuspension are considered three major processes that contribute to the increased nutrient concentrations and subsequent primary production in the euphotic layer (Chen et al. 2003). The chlorophyll-a reaches its peak three days after nitrate peak after a TC (Pan et al. 2017), and TC-induced phytoplankton blooms usually last for two to three weeks (Babin et al. 2004; Chen and Tang 2012; Foltz et al. 2015; Wang 2020).

Discussion and conclusions

This work reviews the upper ocean response to tropical cyclones, including the current, surface wave, temperature, salinity and biological responses. TC usually causes

upper ocean near-inertial currents, increases strong surface waves, cools (warms) the surface (subsurface) ocean and increases (decreases) surface (subsurface) salinity, also causing plankton blooms. The upper ocean response is controlled by mixing, advection and air–sea flux (i.e. heat flux and fresh water flux). The upper ocean response usually biases to the right (left) side of the TC track because the wind-current resonance is stronger (weaker) and the corresponding mixing is stronger (weaker) on the right (left) side in the Northern (Southern) Hemisphere. The characteristics of the upper ocean response mainly depend on the TC parameters (e.g. TC intensity, translation speed and size) and environmental parameters (e.g. ocean stratification and eddies) (Fig. 3).

In recent decades, the understanding of upper ocean response to a tropical cyclone has improved because of the development of observations and modelling. Traditional observation methods such as buoys and moorings (Black and Dickey 2008; Zhang et al. 2016a, 2019; Yang et al. 2019), air-deployed drifters and floats (Black et al. 2007; D'Asaro et al. 2007; Pun et al. 2011; Sanford et al. 2011), Argo floats (Park et al. 2011; Vissa et al. 2012; Wu and Chen 2012; Fu et al. 2014; Liu et al. 2014; Lin et al., 2017; Chen et al., 2020) and satellite remote sensing (Li et al. 2018; Yue et al. 2018; Ning et al. 2019; Zhang et al. 2019), as well as new observation technology and methods such as gliders (Domingues et al. 2015; Miles et al. 2015; Hsu and Ho 2018) and wave gliders (Mitarai and McWilliams 2016), are now applied to TC–ocean observations. Regarding numerical model simulations, early

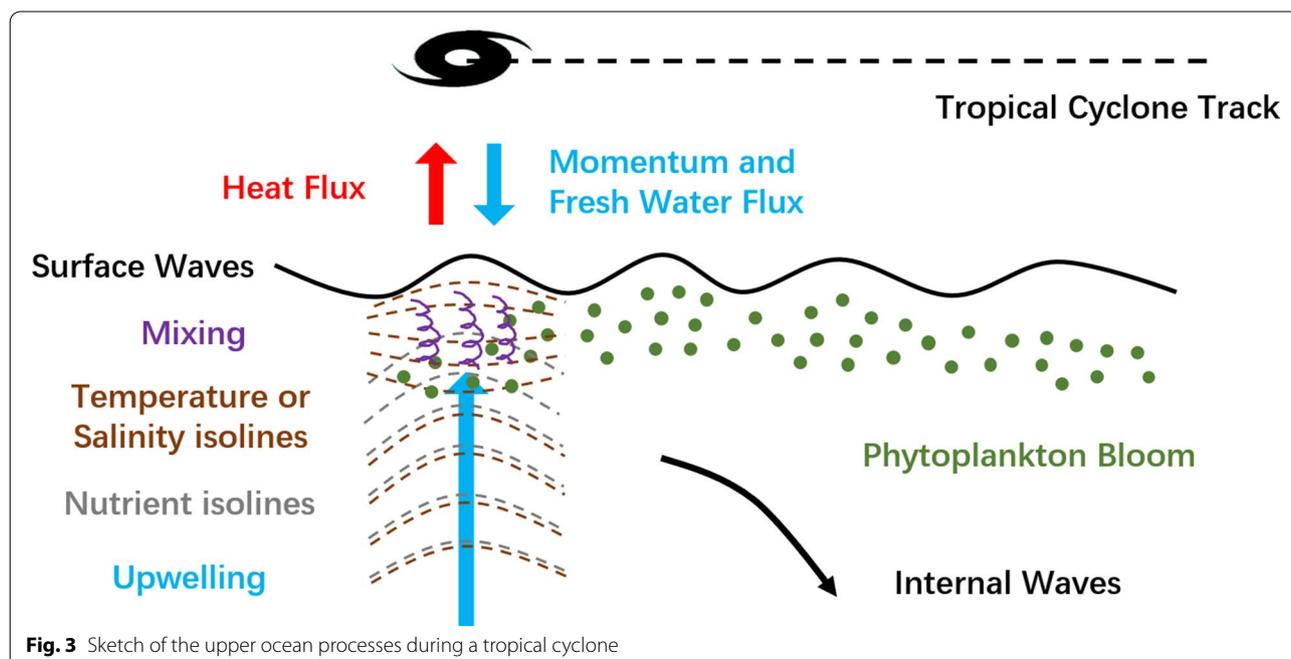


Fig. 3 Sketch of the upper ocean processes during a tropical cyclone

works use the slab ocean model to reproduce the ocean current response (Geisler 1970; Pollard and Millard 1970; Gill 1984), followed by several numerical models such as the three-dimensional Price–Weller–Pinkel model (3DPWP) (Price et al. 1994; Sanford et al. 2007; Guan et al. 2014; Zhang et al. 2016a), the regional oceanic modelling system (ROMS) (Yue et al. 2018) and the Massachusetts Institute of Technology Ocean General Circulation Model (MIT OGCM) (Zedler et al. 2009) to reproduce the three-dimensional current, temperature and salinity responses. Wave models such as the Simulating WAVes Nearshore (SWAN) (Liu et al. 2007; Huang et al. 2013) and WAVEWATCH-III (Moon et al. 2003; Xu et al. 2017b; Qiao et al. 2019) models are used to reproduce the sea surface wave response to a TC. Recently, atmosphere–ocean–wave models such as the Coupled Ocean–Atmosphere–Wave–Sediment Transport (COAWST) modelling system (Prakash and Pant 2016; Wu et al. 2018) have been gradually applied to simulate the ocean response to TCs. Note that the ocean ecological model seems to have not been widely used for the simulation of ocean biological response to a TC yet. What is more, new technology such as big data and machine learning provide a new way to study TC–ocean interaction (e.g. Wei et al. 2017, 2018; Jiang et al. 2018).

Although our understanding of the upper ocean response to a TC has increased in recent decades, some fields merit further study, such as: 1. the characteristics of the air–sea interface as well as the surface flux during TCs; 2. the effect of varied TCs on the upper ocean response, e.g. the upper ocean response during curved TC track, during intensifying (weakening) TCs or accelerating (decelerating) TCs; 3. the interaction between TCs and mesoscale or submesoscale eddies; 4. the upper ocean response to sequential TCs; 5. the effects of TCs on large ocean circulation, e.g. modulation of global ocean circulation by the kinetic energy and heat uptake caused by TCs; 6. the processes that control the recovery of the upper ocean response after TCs; and 7. the propagation of TC-induced anomalies into the ocean interior and deep ocean.

Some existing issues in observation, numerical model and technology restrict the study of upper ocean response to TCs. In situ observation of air–sea interface (i.e. air–sea flux, surface waves) and deep ocean is in shortage, which restricts our understanding of the air–sea interaction and how upper ocean anomalies propagate into ocean interior. The coupling of atmospheric and oceanic model as well as the parameterization of the processes in air–sea interface merits further study for better simulation of TC–ocean interaction. What is more, it is a common problem that surface currents simulated by model seems stronger and persists longer as well as less vertical

propagation of current signals than observation (Huang et al. 2009; Sanford et al. 2011; Zedler et al. 2009; Zhang et al. 2016a, b). In general, accurate prediction of tropical cyclone and oceanic conditions requires proper initialization of both atmospheric and oceanic components of the modelling system, as well as accurate measurements of the ocean ahead of the TC, and skillful assimilation of the ocean data into the ocean model. Besides, the practicability of the usage of big data and machine learning for the study of the upper ocean response to a TC still needs further exploration. We hope the development of science and technology in the future will uncover more processes and mechanisms of TC–ocean interactions, help improve TC forecasts and enhance our understanding of local and global air–sea interactions.

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Authors' contributions

HZ wrote the whole manuscript and HH, WZ and DT revised the manuscript. All authors read and approved the final manuscript.

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Availability of data and materials

Not applicable.

Competing interests

The authors declare that they have no competing interests.

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