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Ocean wave buoy network in extreme tropical cyclones detects the saturation of air-sea momentum transfer

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Abstract

Tropical cyclone is one of the most destructive natural phenomena, causing tremendous disasters worldwide¹. The maximum intensity of tropical cyclones is determined by momentum and heat transfer at the air-sea interface². Momentum transfer corresponds to the momentum loss of tropical cyclones and, consequently, to the underlying ocean’s momentum gain. Air-side observations of wind profiles (top-down approach)³ and ocean-side observations of ocean subsurface currents (bottom-up approach)⁴ showed a slowdown of momentum transfer under high tropical cyclone wind speeds. Although the intensity of tropical cyclones and related disasters are directly related to momentum transfer at high wind speeds, there is still disagreement regarding the slowdown owing to lack of data. Here, we showed momentum transfer in a high-wind-speed region by observing ocean waves directly at the air-sea interface (middle-up-down approach). Although ocean wave observations are highly spatially limited, we
deployed a fleet of newly developed drifting ocean wave buoys covering the active area of tropical cyclones in the Western North Pacific. The buoys fleet captured extreme waves near the eye of the strongest category 5 tropical cyclone, showing clear saturation of momentum transfer beyond surface wind speeds of 25 m/s. Our ocean waves (air-sea interface) results fill in the missing link of quantitative momentum transfer slowdown under extreme wind speeds from both the air and ocean sides. This finding advances not only tropical cyclone modelling but also ocean waves, ocean circulation, and storm surge modelling.

**Main**

Extreme tropical cyclones, such as category 4–5 have historically caused significant damage to global coastal areas due to strong winds, destructive storm surges, and extreme ocean waves. For example, Hurricane Katrina in 2005 caused a storm surge exceeding 10 m and massive coastal inundation along the Gulf of Mexico. Typhoon Haiyan in 2013 landfalled in the Philippines with the worst class intensity of 895 hPa, and the storm surge caused 6,300 casualties. Accurate estimation of tropical cyclone intensity and the resultant storm surges and waves is critical for disaster mitigation measures, such as accurate disaster warnings and infrastructure design. However, direct observation of extreme tropical cyclones is highly challenging due to the difficulty of installing devices in extreme conditions, and air-sea interaction tropical cyclone physics is highly complex; thus, the mechanism determining tropical cyclone intensity remains uncertain.

The energy source of tropical cyclones is the heat supply from the ocean surface, and the sink is the drag at the ocean surface (energy and momentum transfer from the air to the ocean).
Air-sea momentum transfer is generally modelled using the bulk formula $u^2 = C_d U_{10}^2$ where $u_*$ is the friction velocity, $C_d$ is the drag coefficient depending on the wind speed, and $U_{10}$ is the wind speed 10 m from the ocean surface. The theoretical maximum potential intensity can be described using the drag coefficient $C_d^2$. Therefore, momentum flux is one of the main factors determining tropical cyclone intensity. Ocean-side phenomena, ocean waves, storm surges, and wind-induced ocean currents develop by momentum transfer from tropical cyclones. Therefore, the ocean waves, ocean circulation, and storm surge models are driven by wind forcing, represented by $u^2 = C_d U_{10}^2$. Air-sea momentum transfer is a key bulk formulation in atmospheric and oceanic modelling.

Historically, the drag coefficient was considered to increase monotonically with wind speed. However, in the 2000s, there was evidence of a shift in the drag coefficient regime at high wind speeds. Powell et al. used observations by Global Positioning System (GPS) sonde dropped into tropical cyclones from aircrafts and analysed the vertical wind speed profile whose shape is related to air-sea momentum transfer. As a result, the air-sea momentum transfer levelled off (drag coefficient reduction) as the tropical cyclone-induced wind speeds increased. Other studies have also shown drag coefficient reduction based on additional GPS drop sonde observations. On the other hand, Jarosz et al. took a bottom-up approach in contrast to the top-down approach of Powell et al. Momentum flux drives ocean currents. Based on the acceleration of the ocean current, the momentum flux passing from the air to the ocean was estimated, and the drag coefficient reduction was determined at high wind speeds. A subsequent study also determined the drag coefficient reduction based on ocean current observations. However, there is no consensus on drag coefficient reduction because there is contradictory evidence. Bell et al. analyzed aircraft observations considering the angular momentum budget.
of tropical cyclones and concluded that there was no dependence of the drag coefficient on wind speeds greater than 50 m/s. From laboratory experiments, the drag coefficient was estimated to be saturated and not reduced by tropical cyclone winds\textsuperscript{14,15}. Some theoretical studies have shown a continuous increase in the drag coefficient of tropical cyclone winds\textsuperscript{16,17}.

Momentum flux measurements are usually taken using vertical-horizontal wind covariance measurements from buoys near the ocean’s surface\textsuperscript{18}. These measurements have resulted in accurately formulated and widely used air-sea momentum flux parameterisations\textsuperscript{18,19}, though they can only be applied to low to moderate wind speeds (<25 m/s). Such air-sea interface-located direct measurements under high wind speeds (>25 m/s) are not possible because of this difficulty. Therefore, previous studies used indirect estimation methods based on wind profiles far from the surface or on subsurface ocean currents, which are not typically used for momentum flux estimation for low to moderate wind speeds due to inclusion of crude assumptions. The accuracy of indirect methods is generally low. At the air-sea interface, the momentum from the atmosphere consists of ocean waves and not ocean currents. Momentum is first transferred to ocean waves and then to ocean currents\textsuperscript{20}. The air–sea momentum transfer entity is the ocean wave development. Therefore, the air-sea momentum flux regime shifts in tropical cyclone winds should appear in the ocean wave features. Although ocean surface wave measurements have also been highly limited to tropical cyclones, technological advancements have enabled a more extensive capture of tropical cyclone-generated waves. Here, we firstly demonstrate that the air-sea interface of ocean waves can detect saturation of momentum transfer.
Drifting wave buoy deployment for observation of tropical cyclone-induced extreme waves

Extreme wave’s observations during tropical cyclone conditions are highly limited due to difficulty and expenses. Wave observations are typically conducted using moored buoys at fixed locations; limiting the scope of observations. On the other hand, lightweight, low-cost wave buoys have recently been developed to facilitate ocean wave observations. The drifting wave buoys can increase spatial coverage through mass deployment due to their inexpensiveness and light weight mobility. In this study, we deployed newly developed buoys, the Spotter (Sofar Ocean Technologies), which targets tropical cyclone-induced extreme waves in the Western North Pacific. We deployed 19 buoys during the summer (July to October) from 2021 to 2022.
(Extended Data Fig. 1). In addition, data from 18 buoys entering the Western North Pacific from Sofar Ocean global buoy network were provided (Extended Data Fig. 2). Fig. 1 shows the paths of all 37 drifting buoys during the summer of 2021 and 2022. In this open-water region, almost no wave observations have been made historically. The observation coverage depicted in Fig. 1 shows the significant progress made in ocean wave observation research. The buoy transmits observed data every hour using an iridium satellite. The data included bulk wave parameters (significant wave height, mean wave period, and mean wave direction, etc) and wave spectrum. In this study, significant wave heights and wave frequency spectra were analysed.

Tropical cyclones passed through the Western North Pacific domain 15 and 18 times during the summers of 2021 and 2022, respectively (Extended Data Fig. 3). Category-5 super typhoons No. 11 (HINNAMNOR) and No. 14 (NANMADOR) in particular had high intensities and passed close to the drifting buoys with extreme waves (Fig. 2). HINNAMNOR formed on 28th August and moved from east to west, rapidly intensifying on 29th. Several buoys were placed along the path (Fig. 2a), and at 16:00 UTC on 29th, one of the buoys (SPOT-1270; Extended Data Fig. 1) was 8 km away from the tropical cyclone centre, with a minimum central pressure of 955 hPa and a maximum 10-min sustained wind of 44 m/s (Fig. 2b). At that time, the eye of the tropical cyclone was clear and the buoy was located right-side close to the path (Fig. 2b). The significant wave height observed in the buoy increased and then decreased rapidly because of the small size of the tropical cyclone, reaching 11 m (Fig. 2c).
NANMADOR, developed on 13th September, moved northwest and landed in Japan (Figure 2d,g). The lifetime minimum central pressure was 910 hPa from the 16th to the 17th.
During the intensifying stage, the buoy (SPOT-1838; Extended Data Fig. 1) was located inside the tropical cyclone eye with a central pressure of 945 hPa and a maximum 10-min sustained wind of 44 m/s (Fig. 2e). The time series of observed significant wave height shows the clear drop and recovery from 7 m to 4.5 m and to 6.5 m indicating the entering of tropical cyclone eye (Fig. 2e,f). During the most intense stage (central pressure and maximum 10-min sustained wind speed reached 910 hPa and 54 m/s, respectively), one buoy (SPOT-1837; Extended Data Fig. 1) was close to the right side of the moving direction of the tropical cyclone (Fig. 2h). The shortest distance was 45 km during the most intense stage (Fig. 2h). The observed significant wave height increased slowly compared to SPOT-1270 during HINAMNOR and reached 11.5 m (Fig. 2i). Data of extreme tropical cyclone waves could be obtained during various tropical cyclones and stages. Other tropical cyclone waves have also been observed, although these two tropical cyclone observations were notable.

**Air-sea momentum flux estimation from ocean waves**

Ocean surface waves are developed by ocean surface winds, or more precisely, by the momentum flux from winds to waves. Therefore, ocean surface waves provide information on momentum flux and ocean surface wind. In fact, the high-frequency part (i.e., equilibrium range) of the ocean wave frequency spectrum \( E(f) \): Surface elevation variance density spectrum) can be represented as \( E(f) = E_0(u_\ast)f^{-4} \) where \( E_0 \) is the energy level of the equilibrium range, \( u_\ast \) is the friction velocity and \( f \) is the wave frequency\(^{24,25} \). Therefore, if \( E_0 \) is determined from the wave spectrum, \( u_\ast \) (the momentum flux) can be estimated. In addition, the ocean surface wind speed can be estimated from \( u_\ast \) assuming a wind profile based on the law of walls. Shimura et al.\(^{26} \) proposed a highly accurate algorithm for estimating ocean surface wind from the ocean
surface wave spectrum observed by buoys, which was also used in this study. The Shimura et al.\textsuperscript{26} algorithm is based on determining $E_0$ as,

$$E_o = \max E(f) f^4. \quad (1)$$

Based on the dispersion relationship in the deep ocean $(2\pi f)^2 = 2\pi g/L$, equation (1) can be

$$E_o = \max E(f)/L^2 \cdot (g/2\pi)^2$$

where $L$ is the wavelength and $g$ is the gravitational acceleration. Here, $E(f)/L^2$ corresponds to the wave slope density spectrum\textsuperscript{27}. Therefore, $E_o$ indicates the upper limit of the wave slope (i.e. steepness) density. Furthermore, $E_o$ is proportional to $u_*$. Therefore, $E_o$ corresponds to air-sea momentum flux. The method for deriving the ocean surface wind speed from $E_o$ is described in the Methods section.

The behaviour of the air-sea momentum flux caused by extreme tropical cyclone winds observed by drifting buoys is discussed. Fig. 3 shows the relationship between the air-sea momentum flux and wind speeds of up to 50 m/s over the ocean. Several wind speed products (JMA-MSM, MTCSWA, and HOLLAND; see Methods) are used as reference to account for uncertainty. The estimated wind speeds based on the observed wave spectra saturate with wind speeds over 25–30 m/s (Fig. 3a). This saturation is related to the saturation of $E_o$ as shown in Fig. 3b. Although the reference wind speed products differ, the common tendency of saturation can be observed in all of them. In Fig. 3b, the values derived from the wave model simulation (WW3, see the Methods section) have been added. The wave model (WW3) doesn’t exhibit saturation for winds of 25~30 m/s. In contrast, the wave model with a constant drag with wind speed greater than 25 m/s (denoted as WW3(DragConst); see Methods) shows saturation, and the result is closer to the observations.
The drag coefficient \( (C_d) \) was calculated as \( C_d = \frac{u^2}{U_{10}^2} \). Fig. 3c shows the relationship between the drag coefficient \( (C_d) \) and wind speed. The derived drag coefficient is consistent with the widely used drag coefficient parameterisation of up to 25 m/s (COARE3.5\textsuperscript{19}). Beyond 25 m/s, the drag coefficient decreases remarkably. Regarding the wave model (WW3), the drag coefficient is saturated at wind speeds beyond 30 m/s (Methods). Owing to this \( C_d \) saturation, the slope of \( E_o \) beyond 30 m/s is milder than that below 30 m/s, as shown in Fig. 3b. Although the wave model (WW3) shows such the \( C_d \) saturation, the observation results show a much smaller \( C_d \) beyond a wind speed of 25 m/s. The wave model with constant drag (WW3(DragConst)) shows behaviour similar to the drag coefficient reduction observed. A smaller drag (momentum flux) results in a less pronounced wave development. Consequently, the wave heights generated by tropical cyclones are reduced. Fig. 3d shows the relationship between the significant wave heights and wind speeds. The observed significant wave height at strong wind speeds is smaller than the wave model results (Fig. 3d). This is clear evidence of the smaller momentum flux and drag coefficient reduction, as shown in Fig. 3b and 3c. Based on the results of this study, the following drag coefficient parameterisation when \( U_{10} > 25 \) m/s is proposed as \( C_d = \frac{1.75}{U_{10}^2} \) and \( \min C_d = 0.001 \), as plotted in Fig. 3c.

Other wave observation platforms can support the results of the air-sea momentum flux saturation at high wind speeds. Traditional mooring wave buoy observations (Methods) have longer record lengths of four decades at several locations. The number of tropical cyclone wave observations made by mooring buoys is limited because the sparse observations which do not focus solely on tropical cyclones. Fig. 4a shows the relationship between the observed wind speeds and significant wave heights, as Fig. 3d. The mooring wave buoy observations also reveals a tendency for wave height saturation at high wind speeds, indicating—the sea momentum
flux saturation. The wave heights at wind speeds greater than 50 m/s are the same as those at a wind speed of 30 m/s. The numbers of observed extreme events with wind speeds more than 35 m/s and 40 m/s were 21 and 7, respectively. Even after four decades of observations at various locations, extreme event observations have been limited.

Altimeter satellite wind and wave observations have been available since 1985 (Methods). Although satellite wind and wave observations are indirect estimates derived from radar signals reflected at the sea surface and calibrated using mooring buoy observations, these can cover the globe. Fig. 4b shows the tendency of the wave height saturation at high wind speeds, indicating the sea momentum flux saturation. The accuracy of the calibration at high wind speeds (> 25 m/s) is uncertain because the number of mooring buoy observations is limited. However, all observation platforms, including the drifting buoy network in this study, historical mooring buoys, and satellite estimations, reached the same conclusion: the saturation of the air-sea momentum flux at tropical cyclone wind speeds, that is, drag coefficient reduction.
Fig. 3. The relationship between air-sea momentum flux and wind speed over the ocean. (a) Reference wind speed vs estimated wind speed from ocean waves. (b) Reference wind speed vs $E_o$ as momentum flux proxy. (c) Reference wind speed vs drag coefficient. (d) Reference wind speed vs significant wave heights. Three reference wind speed products (JMA-MSM, MTCSWA, HOLLAND) are plotted in (a)(b)(d). The COARE3.5 drag coefficient parameterisation and the proposed parameterisation are plotted in (c). Two wave model simulations results (WW3 and WW3(DragConst)) are plotted. The error-bars indicate the 10 to 90% tiles in each bin ($\pm 1$ m/s) and the dots are the average. If the data in the bin exceeds 10, the error-bar is plotted.
Fig. 4. The relationship between wind speeds and significant wave heights observed by (a) NDBC mooring wave buoy and (b) satellite altimeters. The error-bars indicate the 10 to 90% tiles in each bin ($\pm$ 1 m/s) and the dots are the average. If the data in the bin exceeds 10, the error-bar is plotted.

**Discussion**

The behaviour of the air-sea momentum flux under tropical cyclone high-speed wind speeds of beyond 25-30 m/s has not yet reached a consensus. Observations of wind speed profiles in the atmosphere and subsurface currents in the ocean have previously shown a drag coefficient reduction. However, the estimation accuracy is low because the observation points are far from the sea surface, where momentum transfer occurs. Therefore, we estimated the air-sea momentum flux at the air-sea interface by focusing on sea surface physics using ocean-wave information. We deployed newly developed lightweight and low-cost ocean-wave buoys and developed an ocean-wave buoy network in the Western North Pacific, targeting tropical cyclone-induced waves. We succeeded in making observations which were very close to the eye of the highest-category tropical cyclone, such as at 910 hPa. The observation results successfully captured the saturation of the air-sea momentum flux and the reduction in the drag coefficient at
wind speeds greater than 25 m/s. We proposed a new drag coefficient parameterisation with wind speeds as $C_d = 1.75/U_{10}^2$ and $\min C_d = 0.001$ ($U_{10} > 25$ m/s).

Several studies have attempted to theoretically explain the mechanism of the air-sea momentum flux regime shift (drag coefficient reduction) under strong winds. For example, under tropical cyclone wind conditions, breaking waves continuously supply sea splay droplets (water particles) that are suspended over the sea surface and form a stable wave boundary layer. This can reduce the drag coefficient. Another mechanism is explained by the nonlinearity of the wind-wave interaction when the wave steepness (slope) reaches its limit. $E_o$ shown in this study corresponds to the upper limit of the wave slope density, and the saturation of $E_o$ indicates the saturation of the wave slope. Therefore, this study supports the mechanism determined by the wave steepness.

Air–sea momentum flux is important for determining tropical cyclone intensity. In addition, it is a critical factor in ocean wave and storm surge modelling. As shown in Fig. 3d, the standard wave model with drag coefficient saturation (not momentum flux saturation) shows totally different wave heights under extreme wind speeds compared with the observations, with a difference of more than 5 m at a wind speed of 50 m/s. Furthermore, as an approximate estimation, storm surge is proportional to the magnitude of the drag coefficient, and a reduced drag coefficient directly leads to a reduced extreme storm surge height. Therefore, the findings of this study have a significant impact on coastal disaster mitigation measures and tropical cyclone modelling.
Methods

Wind estimation method

The method for deriving ocean surface wind speed from $E_o$ (Shimura et al.\textsuperscript{26}) is described here. $u_*$ is represented by $E_o$ as,

$$u_* = \frac{(2\pi)^3}{4\beta I g} E_o$$

where $\beta$ is a tuning constant, $I$ is the directional spreading function. The wind profile $U(z)$ is represented by

$$U(z) = \frac{u_*}{\kappa} \ln \left( \frac{z}{z_0} \right)$$

where $z$ is the height, $\kappa$ is the von Kalman constant, and $z_0$ is the momentum-roughness length. The momentum roughness length can be formulated using the Charnock relationship\textsuperscript{31}. Based on these equations, the estimation of $E_o$ leads to the determination of wind profile $U(z)$. Parameter $\beta$ was tuned for spotter buoys as 0.012, although it was 0.016 in Shimura et al.\textsuperscript{26}. Extended Data Fig. 4 depicts the relationship between the wave spectra and estimated wind speed. All spectra from the SPO-1837 (Extended Data Fig.1) during the target period are shown, and the colour indicates the wind speed estimated from the spectra. The dashed lines are proportional to $f^{-4}$ and correspond to specific wind speeds (5,15,25,35,45 m/s). The corresponding wind speed of $f^{-4}$ line for which a certain wave spectrum touches the estimated wind speed derived from the wave spectra.
Tropical cyclone wind data

Tropical cyclone track data were derived from the best track provided by the Regional Specialized Meteorological Center (RSMC) Tokyo-Typhoon Center. The best track data contained information on location, central pressure, maximum sustained wind speed, wind radius of 50 knots or greater, and wind radius of 30 knots or greater.

Tropical cyclone winds were obtained from hourly analysis (i.e. initial forecast conditions) and forecasts using the Japanese Meteorological Agency’s Meso Scale Model (JMA-MSM). JMA-MSM data is used for Japanese operational weather forecasts. Every 3 h of data from 00Z (03Z, 06Z, …, 21Z) are the analysis data, and the data between analyses are forecasts from the previous analysis. The spatial domain is indicated by dashed lines in Fig. 1, and the spatial resolution is 5 km. The JMA-MSM analysis is performed by incorporating observation data into the numerical weather model. However, there is uncertainty regarding the tropical cyclone winds. Therefore, two different datasets, in addition to the JMA-MSM for tropical cyclone winds, were used in this study.

One is the multi-platform tropical cyclone surface winds analysis (MTCSWA) provided by the US National Oceanic and Atmospheric Administration. The wind fields are objectively estimated using multiple satellite platforms (sounder data from six satellites and imagery data from four satellites). The data is created every three hours at 4.5 km intervals from 900 km from the tropical cyclone centre. The data has been used by a weather forecaster to predict the wind and resultant storm surge. The latest upgraded version of the MTCSWA was started on 7 September 2022 and we used this data. Data with wind speeds greater than 15 m/s were used in this study.
Other tropical cyclone wind data were obtained using the parametric tropical cyclone model\textsuperscript{33}. Based on the minimum central pressure, maximum wind speed, and radius of 50 knots winds (or 30 knots winds if 50 knots winds did not exist) from the best track data, the wind field was constructed by considering the wind speed contribution from the tropical cyclone translation speed\textsuperscript{34}. The maximum contribution was set to 7.5 m/s, here. Holland parametric winds have been widely used to estimate coastal disasters. Data with wind speeds greater than 15 m/s were used in this study. The wind data were labeled as HOLLAND.

**Numerical model simulation of ocean waves**

The spectral wave model, WAVEWATCH III\textsuperscript{35}, simulations were used for comparison with the observed results. The simulation consists of two domains using a nesting procedure. In the inner domain which is the same as that of the JMA-MSM (Extended Data Fig.3), WAVEWATCH III was forced by the JMA-MSM hourly winds. In the outer domain, the North Pacific (110E-90W and 25S-65N), wind from JRA-55\textsuperscript{36} was used as the input. The spatial resolution of the inner domain was 0.0625° longitude and 0.0500° latitude, whereas that of the outer domain was 0.25°. The data from the inner domain were used in this study. The widely used parameterisation of wind input and wave dissipation formulated by Ardhuin et al. (ST4)\textsuperscript{37} was selected in WAVEWATCH III. The directional resolution was 10°, and the frequency domain was discretised by 36 within 0.035–0.98 Hz. The drag coefficient $C_d$ of the wave model is saturated at wind speeds greater than 30 m/s. This is not the intended result of ST4 formalisation, but because an unintended limit is set in the algorithm in the ST4 calculation\textsuperscript{38}. 
In addition to the general configuration of WAVEWATCH III described above, an experiment was conducted to tune the wind drag to a constant value ($u_\ast^2 = 1.75$) at wind speeds greater than 25 m/s. The simulation period for the general configuration was from July to October in 2021 and 2022. The experimental configuration’s (WW3 (DragConst)) timeframe was set to coincide with the passage of extreme tropical cyclones passing in the domain on September 24 to 31 in 2021, August 27 to 31, and September 14 to 19 in 2022.

Ocean wind and wave observation from different platforms

Independent observation results are shown in order to support the results by drifting buoy observation deployed in this study. The National Data Buoy Center (NDBC) of the US National Oceanic and Atmospheric Administration (NOAA) provides several decades of ocean wind and wave observations using mooring buoys. The observed significant wave heights and sea-surface wind speeds were analysed. First, all data provided from 1980 to 2021 were obtained. Observation locations with water depths greater than 50 m were selected to focus on deep-water observations. The heights of the wind speed observations differed depending on the location. Therefore, the observed wind speeds were converted to $U_{10}$ based on the law of the wall and the Charnock relationship with Charnock parameter = 0.018. For some locations where the height information is missing, 4 m height was used instead. The analysis excluded the locations with a historical maximum $U_{10}$ less than 25 m/s. For quality control, observations with jumps of more than 20 m/s between single time steps were eliminated. Consequently, 12,465,890 data points from 76 locations were selected (Extended Data Fig. 5).
Other wind and wave data sources were satellite altimeter estimations. Young and Ribal\textsuperscript{28} and Ribal and Young\textsuperscript{39} compiled globally calibrated and validated significant wave height and wind speed data from 1985 to the present from 13 altimeters. Data from 1985 to 2022 were used in this study. Only the data with quality flag of wave heights “Good data” was used. The tropical cyclone-generated winds and waves data were extracted by finding the data around 1 deg from tropical cyclone centre with time difference ±1 h. The target region is the Western North Pacific. The tropical centres were derived from the best track of the RSMC Tokyo Typhoon Centre. The total data points were 128,511 (Extended Data Fig.6).

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Data availability

Drifting buoy observation data deployed by Kyoto University (Extended Data Fig.1) and wave model simulation data are available from ZENODO (but in https://www.dropbox.com/scl/fo/odn11ha01ch1nh9e39wo1/h?rlkey=uyungf189bp2x6q5cjr65084p0&dl=0 during the review process). Drifting buoy observational data from the Sofar Ocean (Extended Data Fig.2) are available for academic use upon request from Sofar Ocean. Tropical

Author contribution

T.S designed the research, deployed the ocean wave buoys, analysed the data, wrote the paper.
N.M and T.M. analysed the data and wrote the paper.

References


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Extended data

Extended Data Fig. 1. The buoy deployed by the Kyoto University. (a) The observation period for each buoy (SPOT-XXXX). (b) Drifting paths of each buoy during the analysis period.
Extended Data Fig. 2. The buoy deployed by the Sofar Ocean. (a) Observation period of each buoy. (b) Drifting paths of each buoy during the analysis period.
Extended Data Fig. 3. Tropical cyclone tracks and maximum significant wave heights by wave model simulation in the (a) 2021 summer and (b) 2022 summer.
Extended Data Fig. 4. Relationship between the wave spectrum and estimated wind speed. The colours indicate the wind speeds estimated from the wave spectra. All spectra from the SPOT-1837 were plotted. The dashed lines are proportional to $f^{-4}$ and correspond to specific wind speeds (5, 15, 25, 35, 45 m/s).
Extended Data Fig. 5. The NDBC data period used in Fig. 4a. The vertical axis represents each location (76 locations).
Extended Data Fig. 6. The amount of satellite altimeter data is shown in Fig. 4b.