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Continental Shelf Research 25 (2005) 1023-1042

CONTINENTAL SHELF RESEARCH

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Taiwan strait current in winter

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Received 4 June 2004; received in revised form 18 October 2004; accepted 13 December 2004

Abstract

Four bottom-mounted current profilers were deployed across the Taiwan Strait from September 28 to December 14 of 1999 to monitor the current velocity when the northeast monsoon was strong. Results indicate both diurnal and semidiurnal tidal currents were primarily barotropic. The barotropic diurnal tide might be explained by a single Kelvin wave propagating along the Mainland China coast from north to south. However, the barotropic semidiurnal tide manifested as a more complicated form in the Taiwan Strait.

The subtidal current generally fluctuated with the northeast winds. When the northeast wind was weak, the alongand cross-strait subtidal current flowed primarily against the wind and toward Taiwan, respectively. As the northeast wind intensified, the along-strait current flowed downwind, brought the cold China coastal water southward, and formed a baroclinic velocity front in the western portion of the Taiwan Strait. The Ekman effect forced the cross-strait current toward Mainland China in the upper water column and toward Taiwan in the lower water column, respectively. The along-strait volume transport, estimated from interpolated current velocity, varied from -5 to 2 Sv with a mean value of 0.12 ± 0.33 Sv. Similar transport was also estimated from the sea level difference across the Taiwan Strait.

Although the local wind played a dominant role for the fluctuations of current velocity and transport in the Taiwan Strait, it could be not the only important factor. The current or transport directed frequently against the wind could be related to the northward current, which was consistently observed in the Penghu Channel. © 2005 Elsevier Ltd. All rights reserved.

Keywords: Taiwan Strait; Tidal current; Subtidal current; Volume transport

1. Introduction

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The Taiwan Strait is located between Taiwan and southeast Mainland China, connecting the East China Sea and the South China Sea. It is approximately 200 km wide and 400 km long, with

an average depth of 60 m. The archipelago of Penghu is located in the southeastern sector of the strait, where the topography is complicated. The funnel shaped Penghu Channel sits between Penghu and Taiwan with Taiwan to the east. West of Penghu Channel lie the Taiwan Banks, where the water depth is generally shallower than 20 m. From south to north, the channel becomes more shallow and narrow. North of the channel, the strait is wide and shallow. It is generally believed that the Penghu Channel is important in conducting the water flow from the southern opening of the Taiwan Strait into the rest of the strait (Chuang, 1985, 1986; Wang and Chern, 1992; Jan et al., 1994).

Both tidal and subtidal currents are strong in the Taiwan Strait. The semidiurnal tide is dominant (Chuang, 1985). It is generally believed that the semidiurnal tide propagates into the Taiwan Strait through both the northern and southern openings, converging near the middle portion of the strait. However, some recent studies show different and controversial results. Lin et al. (2000) argued that the semidiurnal tide is presented as a standing wave in the Taiwan Strait, but Jan et al. (2002) claimed that the propagating and standing semidiurnal tides were co-existent, with most of the energy of the semidiurnal tide coming from the north. The diurnal tide is generally less intense than the semidiurnal tide in the Taiwan Strait, except at its southern sector. Sea level measurements revealed that the diurnal tide propagates from north to south (Hwung et al., 1986).

The subtidal current flow in the Taiwan Strait was usually northward in both winter and summer, with the summer northward current being stronger (Chuang, 1985, 1986; Wang and Chern, 1989). Chuang (1986) claimed that the subtidal current in the Taiwan Strait could be considered as two separate parts. One relates to the sea level difference between the southern and northern end. The second part relates directly to the local winds. Chuang (1985) inferred that the sea level rises at the southern end of the Taiwan Strait when the Kuroshio or South China Sea water intrudes into the strait. The local wind consists primarily of the northeast monsoon in winter and the southwest monsoon in summer. In general, the transition between monsoons occurs in April and October (Chu, 1971). The northeast monsoon has larger amplitude than the southwest monsoon (Jan et al., 2002). Chern (1982) concluded that the variations of current, sea level, and north–south wind stress were coherent and were dominated by signals of 4–8 day periods in winter. The cold China coastal water appears to flow southward along the Mainland China coast, based on remotely sensed sea surface temperature observed during the northeast monsoon (Huh, 1986).

The former studies were based upon observations at a single mooring site using 1 or 2 rotor-type current meters. The current variation in space and the volume transport through the strait remain unclear. With the support of the National Science Council of Taiwan, 4 bottom-mounted acoustic doppler current profilers were deployed to study the current across the Taiwan Strait from September 28 to December 14 of 1999. The observed current velocity has been used by Teague et al. (2003) to estimate the volume transport in the Taiwan Strait but no detail description on the current velocity variations in the Taiwan Strait. This paper presents the observed current velocity in winter when the northeast monsoon prevailed. The variability of current velocity is investigated. Using various methods, three types of volume transport are estimated and discussed. The paper proceeds as follows. Section 2 describes the fieldwork and the observations. These observations include island wind, air temperature, current velocity, sea level, and bottom water temperature at mooring sites. Section 3 examines the tidal current. Section 4 describes the subtidal current variation. The impact of local wind is investigated. The volume transport and associated error are given. Section 5 discusses the bottom frictional effect on the tidal current. The relative geostrophic transport is estimated using the sea level data. The fronts of temperature and current in the western Taiwan Strait are discussed. Section 6 gives a summary.

2. Observations

Four bottom-mounted current velocity profilers were deployed across the Taiwan Strait from

Wuchiou to Taichung in the late 1999. Fig. 1 shows the station locations and the surrounding bathymetry. From west to east, the stations are named C1, C2, C3, and C4. The water depths of the stations are 44, 64, 68, and 56 m, respectively. The distance between adjacent moorings was generally 40 km. The orientation of the array across the Taiwan Strait is 30° clockwise from the east-west axis. The bottom-mounted profilers consisted of the RD Instrument 300 kHz Workhorse with four 20° transducers at C2, C3, and C4 and the Nortek 500 kHz NDP with three 25° transducers at C1. The internal temperature sensor installed on each current profiler recorded the near sea floor water temperature. A bottom-mounted pressure gauge was installed at each station to monitor sea level variations. All of the profilers were retrieved safely, but pressure gauges at C2 and C3 were lost.

The bin sizes of all profiles were 2m, beginning at 1 bin from the transducer. The current velocities were recorded starting at 4 m above the sea floor. The near sea surface bins were contaminated by side-lobe reflection from the sea surface. The contaminated depths were about 5 m. When considered with the tidal effect, which caused a sea level variation of $3-6 \,\mathrm{m}$, the current velocities in the upper 8m were not usable. The current velocities at C1 were only recorded at depths below 12m, since only 15 bins were set. The current velocities were recorded and averaged every hour based on 120 pings for Workhorse and 300 pings for NDP. The standard deviation of nominal error of current velocity measurement is +0.5 and +1.2 cm s⁻¹ for Workhorse and NDP, respectively. Hourly water temperature data were recorded at every station. Hourly sea level data were only obtained at C1 and C4 from pressure



Fig. 1. Map of the Taiwan Strait study area showing the positions of current (solid circles) and wind (star) stations. The along-strait and cross-strait orientations are shown with arrows, with plus signs indicating positive directions. Depth contours are in meters.

gauges. The hourly air temperatures and island wind at Wuchiou were also used in this study. C1 was deployed from August 29 of 1999 to December 14 of 1999, while the other 3 stations were deployed from September 28 of 1999 to December 28 of 1999. The synchronously measured duration was 78 days from September 28 to December 14 of 1999.

Fig. 2 shows the 30° clockwise-rotated Cartesian wind velocity components and their variance density spectra. The spectra were calculated from demeaned, 10% cosine-tapered, and zero-argument time series using a fast Fourier transform. Smoothing was performed by averaging a uniform spectral window, 9 fundamental frequency bands in width. The fundamental frequency and bandwidth were 0.51×10^{-3} and 4.62×10^{-3} cycles per hour (cph), respectively. The resulting degrees of

freedom are 17. The along-strait wind velocity (v_w) was large and primarily negative. The v_w amplitude fluctuated largely from the beginning of the recorded experiment through early November. From mid-November to the end of the experiment, the v_w amplitude fluctuated less but remained strong. The v_w occasionally had positive amplitude. The variance density spectrum of v_w reveals a distribution over a wide sub-diurnal band, except for a weak peak occurring about the frequency band of 0.01 cph. The cross-strait wind velocity (u_w) was much smaller than v_w and is ignored hereafter.

Fig. 3 shows the 30° clockwise-rotated Cartesian current velocities at selected depths, 12 and 40 m. The v and u represent the along- and cross-strait current velocities, respectively. In general, the mean speed was small but had a relatively



Fig. 2. (a) Upper and lower panels display the sampled hourly time series for the along-strait wind velocity (v_w) and its variance density spectra at Wuchiou, respectively. The 90% level of confidence interval is also revealed in the spectra panel. (b) Upper and lower panels display the sampled hourly time series for the cross-strait wind velocity (u_w) and its variance density spectra at Wuchiou, respectively. The 90% level of confidence interval is also revealed in the spectra panel. (b) Upper and lower panels display the sampled hourly time series for the cross-strait wind velocity (u_w) and its variance density spectra at Wuchiou, respectively. The 90% level of confidence interval is also revealed in the spectra panel.



Fig. 3. (a) The sampled hourly time series for the along-strait current velocity (v) at selected depths 12 and 40 m. (b) The sampled hourly time series for the cross-strait current velocity (u) at selected depths 12 and 40 m. The dotted line indicates the mean speed during the measured period. From top to bottom, the four panels represent current velocities at C1, C2, C3, and C4, respectively.

complicated distribution. For example, the mean v was negative at C1 and C3 but was positive at C2 and C4. The mean u was generally negative at 12 m but it was positive at 40 m for four stations. In contrast to the mean current, the fluctuations were large. The tidal current fluctuation was largest at C1 and gradually decreased toward C4. It was larger for v than u. Fluctuations over various time scales, from days to month, were also energetic.

These sub-tidal fluctuations also had larger amplitude in v than u.

Fig. 4 shows the variance density spectra of v and u at selected depths (solid and dashed lines representing depths of 12 and 40 m, respectively). Their spectra were calculated in the same manner as that of the wind velocity. The current spectrum revealed that the high-frequency fluctuations were primarily due to the semidiurnal and diurnal tidal



Fig. 4. (a) The variance density spectra for the along-strait current velocity (v) at selected depths 12 m (solid line) and 40 m (dashed line). From top to bottom, the four panels represent current velocities at C1, C2, C3, and C4, respectively. The 90% level of confidence interval is also revealed in each panel. (b) The variance density spectra for the cross-strait current velocity (u) at selected depths 12 m (solid line) and 40 m (dashed line). From top to bottom, the four panels represent current velocities at C1, C2, C3, and C4, respectively. The 90% level of confidence interval also shows in each panel.

currents. In general, the semidiurnal tidal currents have a larger variance than the diurnal tidal currents. The semidiurnal and diurnal tidal currents were decreased from C1 to C4, but the decrease on the semidiurnal tidal current was faster than the diurnal tidal current. The variance is distributed widely over the low-frequency band. The variance peak centered at 0.01 cph, shown on the variance spectrum of v_w , was barely seen except at C2. The current at C3 had the lowest subtidal variance among the 4 stations. In general, the current variances were larger in the upper than lower water column for the low-frequency band.

Fig. 5 shows the adjusted sea level variations at C1 and C4. The sea level variation has been demeaned, and the effect of atmospheric pressure has been removed. The spectra (not shown) revealed that the fluctuations were dominated by the semidiurnal and diurnal tide. The fortnightly tidal variation was large. The spring tidal amplitude was nearly twice the neap tidal amplitude. The sea level varied 3–6 m over a single tidal period. The tidal amplitude decreased slightly from C1 to C4.

The air temperature (T_a) at Wuchiou and the near bottom water temperature (T) at C1, C2, C3,



Fig. 5. The sampled time series of sea level variation at C1 (left panel) and C4 (right panel).



Fig. 6. The sampled time series of the air temperature (dotted line) at Wuchiou and the bottom water temperature at C1 (bold line), C2 (thin line), C3 (dashed line), and C4 (dash-dotted line).

and C4 are shown in Fig. 6. The T_a revealed a continuous step-like decrease from early October through late November. Each step lasted around 2 weeks and then abruptly dropped. From late November to mid-December, the T_a remained fairly stable. The T at C1 had similar variations to the T_a , but it continuously dropped after late

November when the T_a fluctuated but ended its decline. The late drop in T at C1 can be approximated by an exponential decrease. An energetic fluctuation of T occurred in the semidiurnal tidal period just for C1. A horizontal temperature front might be near C1 moving with the semidiurnal tidal current. The T variations at C2, C3, and C4 had little step-like decrease, but had an approximately exponential drop, which were also observed at C1 after late November. There was no noticeable super-tidal fluctuation. The *T* differences among C2, C3, and C4 were generally small over the entire record. The *T* differences were also small between C1 and C2 in the beginning, but gradually increased forming a large horizontal temperature gradient between C1 and C2. The largest difference was about $3.5 \,^{\circ}$ C. The water at C1 might have different character or source than it at the other 3 stations.

3. Tidal current

A rotary spectrum was applied to the current velocities. The tidal current hodograph ellipses over the water column at 2m intervals for 4 stations are shown in Fig. 7. The central frequencies were 8.10×10^{-2} and 3.95×10^{-2} cph for semidiurnal and diurnal tides, respectively, and the frequency bandwidth was 4.62×10^{-3} cph. For the semidiurnal tide, the sense of rotation was counterclockwise and the orientation was almost parallel to the Taiwan Strait over all the ellipses. with only a few exceptions. The size of ellipses changed greatly across the Taiwan Strait, but changed only slightly with depth. Clearly, the barotropic semidiurnal tide dominated. The diurnal tidal ellipse had larger eccentricities than the semidiurnal tidal ellipse but had a similar orientation. Clockwise and counterclockwise motions were the opposite in the upper and lower water column, respectively for the diurnal ellipses. The size of ellipse changed with depth. The length of the major axis also decreased from west to east. but the rate of reduction was much less than that of the semidiurnal tide.

The barotropic tidal velocity was estimated, based on the assumption that the vertical integration of baroclinic velocity over the range of measurement of water column is zero. A similar method has been applied in a number of studies (Rosenfeld and Beardsley, 1987; Siedler and Paul, 1991; Tang and Lee, 1996). The potential error resulting from this assumption should be small since the range of measurement covered most of



Fig. 7. (a) The semidiurnal tidal current ellipses over water column at 2m interval for C1, C2, C3, and C4. With the exception of those mark with a minus sign, all of the ellipses are polarized counterclockwise. (b) The diurnal tidal current ellipses over water column at 2m interval for C1, C2, C3, and C4. With the exception of those mark with a minus sign, all of the ellipses are polarized counterclockwise.

the water column. Fig. 8 shows the estimated barotropic semidiurnal and diurnal tidal ellipses for the 4 stations. The orientations and senses of barotropic tidal ellipses were similar for both semidiurnal and diurnal tides at all 4 stations. The ellipses had major axes nearly parallel to the Taiwan Strait with counterclockwise motion. The length of axes of tidal ellipses decreased from C1 to C4. The semidiurnal tide decreased much more rapidly than the diurnal tide. The length of major



Fig. 8. (a) The barotropic semidiurnal tidal current ellipses at C1, C2, C3, and C4. All of the ellipses are polarized counterclockwise. (b) The barotropic diurnal tidal current ellipses at C1, C2, C3, and C4. All of the ellipses are polarized counterclockwise.

axis of the semidiurnal tide was nearly triple that of the diurnal tidal at C1, but similar to that at C4. The harmonic constants of 4 major tidal current components (M_2 , S_2 , K_1 , O_1) for all 4 stations are shown in Table 1. Their semi-major axes length also decreased from C1 to C4 for all 4 major tidal current components.

Assuming an exponential decay from west to east, the spatial e-folding scale of semi-major axis of tidal ellipse is $R_e = d_{i-i}/\ln(r_{ii})$, where d_{i-i} is the distance and r_{ii} is the ratio of semi-major axis of tidal ellipses between the stations i and j, respectively. Using the method of least squares, the estimated e-folding scales were 65 and 300 km for semidiurnal and diurnal tides, respectively. The latter one is close to the local barotropic Rossby deformation radius that is $R = \sqrt{gH}/f = 360$ km, where q is the acceleration of gravity, H = 50 m is the mean depth and f is the Coriolis parameter at 25 °N. Similar calculations were applied to the sea level data recorded at C1 and C4. For 2 stations, only one R_e was obtained, 840 km for the semidiurnal tide and 550 km for the diurnal tide. Without proper degrees of freedom, the estimate of e-folding scale from sea level might be less

Table 1 The 4 major harmonic constants of the barotropic tidal current at C1, C2, C3, and C4 station

	C1	C2	C3	C4
<u>M</u> ₂				
Semi-major axis (cm s^{-1})	31	22	13	5
Semi-minor axis (cm s^{-1})	8	11	7	0
Orientation of semi-major axis(°)	32	38	32	45
S ₂				
Semi-major axis (cm s^{-1})	9	9	5	3
Semi-minor axis (cm s^{-1})	1	3	2	0
Orientation of semi-major axis(°)	32	40	41	52
K ₁				
Semi-major axis (cm s^{-1})	9	9	7	7
Semi-minor axis (cm s^{-1})	0	0	1	1
Orientation of semi-major axis(°)	33	34	42	54
O ₁				
Semi-major axis (cm s^{-1})	7	7	6	4
Semi-minor axis (cm s^{-1})	0	1	0	-1
Orientation of semi-major axis(°)	35	36	41	53

The orientation of semi-major axis is in degrees with respect to east. The sign of semi-minor axis indicates the sense of rotation of the tidal current, positive values implying a counterclockwise rotation.

reliable but it suggests that the diurnal tide had a scale close to the local barotropic Rossby deformation radius, but the semidiurnal tide did not.

The phases between the sea level and depth average v at C1 and C4 were calculated. The variations of sea level and current in the diurnal tidal frequency band were almost out of phase (160°) at both C1 and C4. The phase lags between the semidiurnal tidal height and current were out of phase at C1, but had 135° phase lag at C4. Wave height and current decayed synchronously with distance offshore, with an e-folding length scale close to the local Rossby deformation radius. Wave height and current time evolutions are nearly out of phase. This indicates that the barotropic diurnal tide in the Taiwan Strait propagates as a single Kelvin wave along the Mainland China coast from north to south. The barotropic semidiurnal tide was more complicated.

The above scale analysis was also applied to the individual tidal components $(M_2, S_2, O_1 \text{ and } K_1)$, with amplitudes shown in Table 1. The e-folding scales were 65, 90, 290, and 400 km for M₂, S₂, O₁, and K₁ barotropic tidal currents, respectively. Similar to the previous results, the e-folding scales for both O₁ and K₁ diurnal tidal current were close to the local barotropic Rossby deformation radius, but M₂ and S₂ semidiurnal tidal current were not. The outcomes for the estimated phase lags between the sea level height and depth averaged current velocity for individual tidal components $(O_1, K_1, M_2 \text{ and } S_2)$ do not differ significantly from the previous results. They all reach the conclusion that the barotropic diurnal tides (or K_1 and O_1 tides) in the Taiwan Strait propagates as single Kelvin wave, but the barotropic (M_2 and S_2) semidiurnal tide could not be represented by single Kelvin wave.

By subtracting the depth average velocity from the v and u at each depth, the baroclinic tidal ellipses were obtained. Because the semidiurnal baroclinic ellipses were much smaller than the barotropic tidal ellipses, only the baroclinic diurnal tidal ellipses are shown in Fig. 9. In general, the baroclinic tidal ellipses were clockwise. Their sizes were larger near the surface and the bottom, and their sizes were over 50% of the size of the barotropic diurnal tidal ellipses. Their size



Fig. 9. The baroclinic diurnal tidal current ellipses over water column at 2m interval for C1, C2, C3, and C4. All of the ellipses are polarized clockwise.

was smallest around the mid-depth at each station. The baroclinic diurnal current velocity was decomposed into a set of orthogonal modes using the empirical orthogonal function in the frequency domain. The result indicates that the baroclinic diurnal tidal current could be dominated by the first baroclinic mode, which had nodal points at 26, 32, 36, and 26 m at C1, C2, C3, C4, respectively. However, the bottom frictional effect on the vertical distribution of tidal current velocity could have similar impact with baroclinic effect. The discussion of bottom frictional effect on the diurnal tidal current is given in Section 5.

4. Subtidal current and transport

Fig. 10 shows the 36-h low-pass filtered data for v over the water column at the 4 stations. The low-pass filtered data are obtained by using a truncated Fourier transform to remove fluctuations shorter



Fig. 10. The 36-h low-pass filtered along-strait current velocity (v) isotachs as a function of depth and time for 4 current stations, from top to bottom identifying C1, C2, C3, and C4. The contour interval is 10 cm s^{-1} . The shaded area demonstrates that the current velocity has negative value. The vertical distributions of mean and standard deviation of v appear on the right. The 36-h low-pass filtered, along-strait component of the wind velocity (v_w) is displayed at the top.

than 36h from the original hourly data. The vertical profiles of mean v with one standard deviation and 36-h low-pass filtered v_w are also displayed. At C1, v had a negative mean value over the entire water column. The vertical gradient of vwas small. The v generally fluctuated with the along-strait wind. The v was primarily negative, and was positive only occasionally when the northeast wind was weak. At C2, the mean v was nearly zero at surface, and was positive, increasing with depth below surface. Although the v had larger amplitude in the lower layer, it had a larger fluctuation (large standard deviation) in the upper layer. The v was mainly positive; the negative v was observed when the northeast wind intensified. The positive v primarily surfaced or deepened with the decrease or increase of the northeast wind. A subsurface jet of positive v was frequently observed. At C3, the mean of v was negative and positive in the upper and lower water column, respectively. The v had a smaller standard deviation than the other 3 stations. This finding agrees with the results of variance spectra shown before. The v was negative or positive when the northeast wind was large or small. The v at C4 had a positive mean and large fluctuation over the entire water column. The v was positive or negative when the northeast wind was weak or strong. Horizontally, the v had its largest mean and vertical gradient at C2, while its largest fluctuation occurred at C4. The large positive v generally occurred at C2 and C4, especially at subsurface, when the northeast wind was weak.

Fig. 11 shows the subtidal u at 4 stations. In general, the subtidal u showed less fluctuation than v. At C1, the u was weak and its vertical gradient was generally small. Its mean was negative at the uppermost depths and was near zero beneath 25 m. The u also fluctuated with along-strait wind velocity; u was negative (positive) when the northeast wind was large (small), in general. At C2, the u had its maximum mean value at 40 m, but its fluctuation decreased with depth. The u was primarily positive. The negative u was generally confined to the upper ocean, except for brief periods. The positive u generally surfaced (deepened) with the decrease (increase) of the northeast wind. A subsurface jet of positive u was

frequently observed. The u at C3 was similar to that at C2, but was weaker in both mean and fluctuation. The surfacing and deepening of positive u was generally contingent on the reduction and intensification of the northeast wind. At C4, the u was very weak in both mean and fluctuations. It demonstrated near uniformity over the water column. In general, the positive (negative) u appeared when the northeast wind was weak (large). Horizontally, the u had the largest mean and fluctuation at C2.

In general, the current velocity fluctuated with northeast wind, but it did not vary in proportion to the intensity of the wind speed (or wind stress). For example, the northeast wind velocity in early December was greater than it was in early November, but had smaller downwind current velocity. Furthermore, the current frequently flowed against the wind. These results indicate that the local wind was an important factor in the current variability in the Taiwan Strait, but it was not the only important factor.

In order to interpolate the data spatially, an optimal interpolation scheme, the same used by Jacobs et al. (2001), was applied. The spatially interpolated velocities are linear combinations of the observations at different stations, multiplied by different weights. The Gaussian distance-lagged correlation among observations is assumed to estimate the length scale. The length scale is then applied to calculate the covariance between the points of interpolation and observation. Finally, the weighted function is obtained by minimizing the expected error variance. The time-lagged correlation was not applied, so that the expected error is constant in time.

The resulting horizontal length scales are from 40 to 120 km and are less for the central portion than at the two sites in the Taiwan Strait. The resulting vertical length scale is from 20 to 100 m. The vertical length scales were smallest at C2 where the vertical gradient of the horizontal velocity was the largest among the 4 stations. The estimated interpolation errors along the mooring array line are presented as Fig. 12. Overall, the expected error is small, with the maximum value less than 10 cm s^{-1} . This finding implies that the instruments array properly re-



Fig. 11. The 36-h low-pass filtered, cross-strait current velocity (*u*) isotachs as a function of depth and time for 4 current stations, from top to bottom indicating C1, C2, C3, and C4. The contour interval is 10 cm s^{-1} . The shaded area indicates that current velocity has negative value. The vertical distributions of mean and standard deviation of *u* appear on the right. The 36-h low-pass filtered along-strait wind velocity (v_w) appears at the top.



Fig. 12. The estimated interpolation current errors along the mooring array line. The contour interval is 1 cm s^{-1} . Locations of each station are shown in the uppermost of panel.

solved the typical spatial scale in the Taiwan Strait. The error was large at surface primarily because the velocity was extrapolated from 8 m to the surface. Horizontally, the largest error was between C2 and C3, where the horizontal and vertical length scales were small.

The interpolated velocity across the strait was used to examine the evolution of current in response to the intensification and reduction of the northeast wind. Three events of change were considered. The events were chosen before the northeast monsoon became stable. For each event, 3 snapshot current distributions, before, during, and after intensification of wind (marked by dashdotted line, solid line, and dashed line, respectively), were used to illustrate the evolution.

Fig. 13 shows the snapshots of v. Each column represents one event. Three panels in each column correspond to the snapshots of v before, during, and after intensification and reduction of the northeast wind. The along-strait wind is also displayed. The duration of events varied from 3 to 7 days. In the 3 events, the v displayed similar sequential responses to the intensification and reduction of the northeast wind. Before the northeast wind intensified, the v was generally positive. Double speed cores were observed, one near the center of the strait and the other one near the Taiwan coast. The central core generally had smaller magnitude than the coastal core. The

depth of former core varied from event to event. It appeared at near bottom, subsurface, and surface. It is possible that a different strength and duration of northeast wind prior to the last event may contribute differently to a v distribution. The vbecame negative as the northeast wind intensified and reached its peak. The negative v was large at the surface and decreased with depth. Horizontally, the large downwind flow was found near the Mainland China coast. It decreased rapidly toward C2, where the water depth is greatest. A strong horizontal gradient was observed. Theoretically, the wind-driven v in the narrow strait might be nearly uniform across the strait (Chuang, 1988). The non-uniform horizontal distribution of vcould be related to the bottom topography and horizontal temperature difference. East of C2, the downwind flow increased, but the horizontal gradient was smaller than west of C2. As the northeast wind decreased, the v turned positive, resulting in a distribution similar to the one prior to wind intensification.

Fig. 14 shows the u as Fig. 13 for the v. The udisplayed similar sequential responses for the 3 events of northeast wind intensification and reduction. The *u* was generally positive, flowing toward Taiwan prior to the intensification of the northeast wind. The greatest speed occurred around C2, but at different depths for the 3 events. As the northeast wind intensified, the uturned negative in the upper water column, but remained positive in the lower water column. The magnitude of negative *u* decreased with depth. The Ekman flow had its largest amplitude at surface. Horizontally, the largest negative and positive uoccurred at mid-strait. The along-strait wind appears to have the greatest cross-strait flow in its central portion, where the Ekman flow could be developed better there than at the sides of strait (Chuang, 1988). The large negative u, which developed in the upper water column, might squeeze the positive u downward, resulting in a large positive u forming below it. After the wind decreased, the distribution of *u* generally returned to the pre-intensification pattern. A cycle was completed.

Fig. 15 shows the volume transport (T_T) , which was obtained by integrating the optimal



Fig. 13. The snapshots of the interpolated v velocity in 3 events of intensification and reduction of the northeast wind. For each event, there are 3 snapshot v distributions before (dash–dotted line), during (solid line), and after (dashed line) intensification of the northeast wind. The contour interval is 10 cm s^{-1} . The 36-h low-pass filtered along-strait wind velocity components (v_w) are also displayed at the top.

interpolated v from sea surface to bottom over the across section. The v_w is also shown. The root mean square (rms) of the expected error in the transport estimates is about 0.33 Sv. The T_T had a small positive mean, 0.12 Sv, but large fluctuations, varying from -5 to 2 Sv. In general, the transport primarily fluctuated with the wind; the

transport was upwind or downwind when the northeast wind was weak or large. The largest correlation coefficient was 0.7 between along-strait wind velocity in Taiwan Strait and volume transport with almost no time lag in this study. These suggest that the wind could play a dominant role in causing the volume transport variation.



Fig. 14. The snapshots of the interpolated *u* velocity in 3 events of intensification and reduction of the northeast wind. For each event, there are 3 snapshot *u* distributions before (dash–dotted line), during (solid line), and after (dashed line) intensification of the northeast wind. The contour interval is 10 cm s^{-1} . The 36-h low-pass filtered along-strait wind velocity components (v_w) are also displayed at the top.

However, headwind transport was observed even when the northeast wind was very strong as it was on December 3. The transport variation was also not well in proportion to the wind velocity (wind stress) variation. For example, the wind velocity was slightly larger on November 2 than November 16, but the transport on November 16 was much larger than on November 2. In addition to the local wind, another factor might be also important for the volume transport in the Taiwan Strait.

Using the same current velocity data, Teague et al. (2003) had the correlation between along-strait transport and wind stress at northeast of Taiwan was 0.54 with 1 day lag, and its mean volume transport integrated from depth under sea surface 2 m to bottom was valued 0.14 Sv. The



Fig. 15. The subtidal along-strait volume transport (T_T) , which estimates by integrating the optimal interpolated velocity (v) across of Taiwan Strait. The 36-h low-pass filtered along-strait wind velocity components (v_w) are also displayed at the top.

differences between their and present results could be related to the different wind stress that were applied to compute the correlation coefficients and the extrapolation in the upper 2 m that was used to estimate the volume transport in this study.

5. Discussion

The sizes of tidal current ellipses were decreased with depth for both diurnal and semidiurnal tides. The reduction could be accounted for by the baroclinic tidal current velocity (Section 3); it also could be related to the influence of bottom friction. The influence height (D) of bottom friction is the order of $\pi \sqrt{2Nz/\sigma}$, where Nz is the coefficient of eddy viscosity and σ is the frequency of tide constituent (Bowden, 1983). Taking Nz = 200 cm² s⁻¹, the D will be 53 and 75 m for the semidiurnal tide ($\sigma = 1.4 \times 10^{-4} s^{-1}$) and diurnal tide ($\sigma = 0.7 \times 10^{-4} s^{-1}$), respectively. Obviously, the bottom friction could have impact on the tidal current throughout the whole water column in the 4 mooring stations.

The feature that the diurnal tidal current ellipses were polarized clockwise and counterclockwise in upper and lower water column, respectively, could also be related to the bottom friction. Using a three-dimensional numerical model with constant eddy viscosity coefficient, Tee (1979) pointed out that the bottom frictional effects tend to decelerate the clockwise motion of the tidal current and even turn to counterclockwise motion in the lower column. Chern (1984) analyzed the tidal current at the northern offshore of Taiwan and made an inference that the clockwise component of tidal current is easier damped by the frictional process than the counterclockwise component of tidal current.

For a progressive tidal wave, the bottom friction damped the tidal current velocity as well as made a phase lag between the tidal elevation and current. The current reaches its maximum before the elevation reaches its peak (Bowden, 1983). This mechanism could be an explanation for the result that the phases between the sea level and depth average v in the diurnal tidal frequency were 160° both at C1 and C4, and not 180°. A similar phase lag feature was also found in the analyzed results of O₁ and K₁ tides. The Kelvin wave stated in Section 3 could be a damped one.

Without hydrographic measurement in the duration of current velocity observation, whether the baroclinic or frictional effects has larger impact on the tidal current velocity can not be determined. Further study is required. However, the uncertainty would not bias the conclusion of that the diurnal tide is a Kelvin wave propagating along the coast of Mainland China.

Using a numerical model, Ko et al. (2003) found that the transport fluctuation in the Taiwan Strait could be significantly impacted by the remote wind from Yellow and East China Seas through the coastally trapped waves. They properly explained the transport reversal, but not the total transport. During the observation, the current frequently transported the water against the wind. The result suggests that a flow from the south could continuously flow into the Strait. This inference is supported by a short-term current velocity measurement at the northern tip of Penghu Channel. A RDI 300 kHz Workhorse, with bin size of 2m, was deployed from November 19 of 1999 to April 19 of 2000. Fig. 16 shows the 36-h low-pass filtered depth-averaged current velocity vectors from November 19 to December 14 of 1999, the duration of the current velocity measurement in this study. In spite of the intense northeast wind, the current flowed primarily northward



Fig. 16. The 36-h low-pass filtered depth-averaged current velocity vectors and the mooring location. The triangle indicates the current profiler location in the map. The arrow and letter "N" identify North.

against the wind. Obviously, the northward flow, originating from the Penghu Channel, could have significant impact on both current velocity and transport in the Taiwan Strait.

The relative geostrophic along-strait transport (T_G) was estimated from the sea level differences across the strait $(\Delta \eta_x)$, which is

$$T_G = \frac{gH}{f} \frac{\Delta \eta_x}{\Delta x} L,$$

where *H* is the average depth (50 m), $\Delta x = 114$ km for the distance between C1 and C4, and *L* is strait width (155 km). The 36-h low-pass filtered sea level data at C1 and C4 were applied to get the sea level slope. Because the absolute sea level is unknown, the estimate is relative. The T_G should include the impact from the Yellow and East China Sea, because the coastally trapped wave is quasi-

geostrophic. Fig 17 shows the variations of T_G , and the differences between demeaned T_T and T_G . The variations of T_G and T_T were alike. The correlation coefficient between T_G and T_T was 0.9. However, the rms of the difference of two estimates of transport was 0.64 Sv, which was larger than the rms of the expected error (0.33 Sv.) in the transport estimates. The difference between T_G and T_T could result from the ageostrophic flow. The current, originating from the south of Taiwan Strait, sprayed out from the Penghu Channel and could have contributed to such ageostrophic flow.

The downwind current was generally weakest at C2 when the northeast wind intensified. This feature could be related to the temperature difference found between C1 and C2. The temperature difference increased as northeast wind



Fig. 17. (a) The relative geostrophic along-strait volume transport (T_G) , estimated by sea level differences across the Taiwan Strait. (b) The differences between demeaned T_T and T_G that were obtained by integration of the along-strait velocities and estimations based on sea level differences across the Taiwan Strait, respectively.

surged (seen in Fig. 6). The largest temperature difference could be up to $3.5 \,^{\circ}$ C. Assuming the salinity was 33 psu, the density difference between C1 and C2 is about $0.4 \,\mathrm{kg \,m^{-3}}$. The horizontal density difference in the front is accompanied by a relative large pressure gradient, via geostrophy. The velocity scale, $\sqrt{g'H}$, is around $40 \,\mathrm{cm \, s^{-1}}$ where g' is reduced gravity and H is depth (50 m). The along-strait currents accelerated to the downwind and headwind at C1 and C2, respectively. A large horizontal gradient of v was observed between C1 and C2 (Fig. 13) when the northeast wind was large. The current velocity at C1 and C2 could be significantly modified by the baroclinic effect, but the transport would not.

6. Summary

In summary, the current velocities and the bottom water temperature across the Taiwan Strait were measured. The sea level variation at two sites in the Taiwan Strait and the wind velocity at Wuchiou were recorded. The synchronously duration of observation was 78 days from September 28 to December 14 of 1999, and was dominated by northeasterly winds. The current

velocities in the Taiwan Strait revealed that the barotropic semidiurnal and diurnal tide were dominant. The barotropic diurnal tide could propagate as a single Kelvin wave along the coast of Mainland China, moving from north to south. The baroclinic or bottom friction has impact on vertical distribution of the tidal current velocity, but it would not change the result of the diurnal tide as a single Kelvin wave propagating along the coast of Mainland China. The along-strait subtidal current, with double speed cores, flowed generally against the wind when the northeast wind was weak. The intensification of the northeast wind forced the along-strait current to flow downwind and brought the cold China coastal water southward, forming a baroclinic velocity front in the western Taiwan Strait. The cross-strait subtidal current generally flowed toward Taiwan when the northeast wind was weak. As the northeast wind intensified, the Ekman effect made the current in the upper water column flow toward Mainland China, while the current in the lower water column continued to flow toward Taiwan. The along-strait volume transport variation primarily fluctuated with the local wind. However, the northward flow which originated from the Penghu Channel could have significant impact.

Acknowledgments

This work is a part of project of the Taiwan Strait NowCast Model Study (TSNOW), which was supported by the National Science Council, Taiwan, ROC, under Grant NSC 89-2119-M-002-002. The authors would like to thank the Taiwan Power Company for providing the data for island wind at Wuchiou and current at C1. The assistance of the crews of the R/Vs Ocean Researcher 1 and 2 are greatly appreciated. We also want to thank Dr. Yiing Jang Yang, Mr. Wen-Hwa Her in the moorings preparation.

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