Shelf sea tidal currents and mixing fronts determined from ocean glider observations

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Abstract. Tides and tidal mixing fronts are of fundamental importance to understanding shelf sea dynamics and ecosystems. We use dive-average currents from a two-month (12th October – 2nd December 2013) glider deployment along a zonal hydrographic section in the northern North Sea to determine $M_2$ and $S_2$ tidal velocities, which agree well with tidal velocities measured by current meters and extracted from a tide model. The method enhances the utility of gliders as an ocean-observing platform, particularly in regions where tide models are known to be limited. We use the glider-derived tidal velocities to investigate tidal controls on the location of a tidal mixing front. During the deployment, the front moves offshore at a rate of 0.51 km day$^{-1}$. During the first period of the deployment (i.e. until mid November), the front’s position is explained by the local balance between tidal mixing and surface heat fluxes: as heat is lost to the atmosphere, full-depth tidal mixing is able to occur in progressively deeper water. In the latter half of the deployment, the output of a simple one-dimensional model suggests that the front should have decayed. By comparing this model output to hydrographic observations from the glider, we attribute the persistence of the front beyond this period to the advection of cold, saline Atlantic-origin water across the deeper portion of the section. The glider captures the transition of the front from being one controlled by the balance between tidal mixing and surface heating, to being one controlled by advection of buoyancy. Fronts in shelf regions with oceanic influence may be geographically fixed and persist during periods of little to no thermal stratification, with implications for the thermohaline circulation of shelf seas.

1 Introduction

Tides are of fundamental importance to understanding shelf sea dynamics and ecosystems. Not only are tidal currents frequently the dominant circulation pattern in these regions (Otto et al., 1990), but the turbulence, bottom-mixing and circulation patterns to which they give rise have a profound effect on the physics, biogeochemistry and ecology of shelf seas (Simpson and Hunter, 1974; Lenhart et al., 1995; Holt and Umlauf, 2008). In shallow regions with fast tidal currents, full-depth mixing is maintained throughout the year. In deeper regions, or where tidal currents are slower, tidal mixing cannot overcome buoyancy forcing in summer and the water column stratifies seasonally. The boundaries between mixed and stratified areas are sharp (~20 km; Hill et al., 2008) and are known as tidal mixing fronts. Such fronts separate water masses with markedly different physical
and biogeochemical properties, and the density-driven jets to which they give rise are important transport pathways (Hill et al., 2008). Consequently, an accurate understanding of the processes that control the formation and location of tidal mixing fronts is necessary for effective management of economically important shelf sea ecosystems and for modelling the dispersion of tracers, contaminants and organisms.

Simpson and Hunter (1974) assume that surface heating is spatially uniform over the North West European Shelf and exclude wind mixing to propose that thermal mixing fronts may be found along a critical contour of $h/u^3$, where $h$ is the water depth and $u$ is the amplitude of the M$_2$ tidal speed. No consideration is given to the influence of salinity and of non-tidal flows. Subsequent studies have confirmed the utility of this parameter (Garrett et al., 1978; Simpson and Bowers, 1981; Bowers and Simpson, 1987); the critical value on the northwest European shelf is $\log(h/u^3) = 2.7 \pm 0.4$ (Simpson and Sharples, 1994).

Later contributions added that frontal location may be expected to move as maximum tidal speeds vary over the spring-neap cycle (Simpson and Bowers, 1981; Loder and Greenberg, 1986). The local heating-stirring balance is not, however, the only control on frontal location. Salinity has been found to influence frontal location and movement in Regions of Freshwater Influence (ROFI; Hopkins and Poulton, 2012), and ice melt water has been found to be an important component of frontal systems in the high latitudes (Schumacher et al., 1979). Furthermore, shearing of the density field by the tide can lead to a semidiurnal mixing-stratification cycle that can influence both tidal and residual circulation (Souza and Simpson, 1996; Verspecht et al., 2009; Palmer, 2010). Salinity is of clear importance in frontal dynamics, but its effect other than in ROFIs – for instance, in deeper shelf-sea regions where horizontal salinity gradients are less pronounced – has been less thoroughly investigated.

A marked meridional mixing front, co-located with the path of the Fair Isle Current (FIC; Fig. 1) is present in the northwestern North Sea to the west of 1° W (Turrell et al., 1996; Sheehan et al., 2017). The region, which is influenced by a cool (< 9 °C), saline (> 35.4 g kg$^{-1}$) water mass found to the east of the front (Sheehan et al., 2017; Hill et al., 2008), is characterised by features excluded from the $h/u^3$ theory of frontal location: there is a strong horizontal salinity gradient between fresh coastal waters and more saline, oceanic waters offshore (Turrell, 1992; Hill et al., 2008; Sheehan et al., 2017), and a generally southward flow persists throughout the year (Dooley, 1974; Turrell, 1992; Winther and Johannessen, 2006). It is thought that the location of the front is at least partially influenced by the local heating-stirring balance, with tidal stirring being responsible for maintaining fully mixed conditions west of the front (Svendsen et al., 1991). The hydrographic setting of this front means that its behaviour may be different from fronts where the influence of the open ocean is less pronounced.

We use high-resolution hydrographic and dive-average current (DAC) observations from a profiling ocean glider that repeatedly crossed this mixing front, alongside output from a one-dimensional model to: quantify tidal flows in the vicinity of the front; investigate the factors that determine frontal location; and examine the applicability of the $h/u^3$ theory to fronts in this context. The high spatial resolution of glider observations permits a more accurate estimate of frontal location than is possible from ship-based observations (e.g. Schumacher et al., 1979; Hill et al., 1997; Sheehan et al., 2017). The DAC time series is used to accurately determine the velocities of the M$_2$ and S$_2$ tidal constituents at the time and location of each glider dive without recourse to a tide model. DAC observations are known to be accurate to within a few cm s$^{-1}$ (Merckelbach et al., 2008), and a glider’s speed through water can be determined with sufficient accuracy for gliders to measure, for example,
fine-scale turbulence (e.g. Beaird et al., 2012; Fer et al., 2014). In this study, we augment this capability by demonstrating that individual DAC observations may be accurately separated into tidal and residual components. We then use these glider-derived tidal velocities, together with the model output, to investigate the influence of tidal and non-tidal processes on the location of the mixing front.

2 Glider-derived tidal velocities

2.1 Method

During a two-month deployment (12th October – 1st December 2013), the glider (Seaglider 502; Eriksen et al., 2001) completed 10 occupations of the Joint North Sea Information System (JONSIS) hydrographic section (Turrell et al., 1996). The section, between 2.23° W and the prime meridian at 59.28° N (Figure 1), crosses the combined path of the two western Atlantic inflows into the northwestern North Sea: the FIC and the East Shetland Atlantic Inflow (ESAI; Fig. 1; Turrell et al., 1996). Bathymetry along the section varies between 69 and 143 m, deepening offshore. All bathymetry data used in this study were extracted from the GEBCO dataset (GEBCO_08 grid, version 20100927, www.gebco.net; resolution 30 arc-seconds). Dives were, on average, 20 minutes and 300 m apart, and as one dive comprises two profiles, profiles are therefore an average of 150 m apart. Most dives sampled the full water column. DAC observations are obtained incidentally during a glider’s flight as the glider is advected by the flow over a dive-climb cycle (Eriksen et al., 2001; Merckelbach et al., 2008). The accuracy of DAC observations was improved post-deployment by optimising the hydrodynamic model of the glider’s flight (Frajka-Williams et al., 2011). DAC observations were visually inspected to ensure that there were no systematic errors due to the glider’s compass calibration.

DAC velocities were divided into three longitudinal bins along the JONSIS section (Fig. 2a), the boundaries being chosen such that each bin contained an approximately equal number of dives: 502 in the eastern bin, 514 in the central bin and 503 in the western bin. Binned velocities were treated as three discontinuous time series (Fig. 2b and c) located at the bin’s central point. In each bin, the amplitude and phase of the \( M_2 \) and \( S_2 \) tidal constituents were determined using harmonic analysis (Thomson and Emery, 2014). These results were used to construct tidal ellipses along the JONSIS section (Fig. 3). Combined \( M_2 \) and \( S_2 \) zonal and meridional velocities were then calculated at the time of each glider dive. Finally, tidal velocity was linearly interpolated zonally onto that dive’s location to construct time series of tidal velocity along the glider’s path, hereafter referred to as along-track velocities (Fig. 4). Nearest-neighbour extrapolation was used for dives to the east and west of the three bins; extrapolation is necessary because some dives lie to the east and west of the central points of the eastern and western bins respectively.

Two current meters were deployed on the JONSIS section for a period covering the glider deployment: an Aanderaa Seaguard single-point current meter at a depth of 40 m (1.52° W, Figure 2a), and a Nortek AWAC profiling current meter that took measurements at 4-m intervals from 9 to 89 m below the surface (0.70° W, Fig. 2a). Observations from the profiling current meter were depth-averaged for comparison with glider DAC. The amplitude and phase of the \( M_2 \) and \( S_2 \) tidal constituents were determined from the current meter records using the same harmonic analysis as for the glider data. For comparison,
estimates of M\textsubscript{2} and S\textsubscript{2} amplitude, phase and velocity, were extracted from the TPXO inverse model European shelf solution (0.1° resolution; Egbert et al., 1994; Egbert and Erofeeva, 2002; Egbert et al., 2010, volkov.oce.orst.edu/tides).

2.2 Tidal ellipses

Tidal ellipses of the glider-derived tide show a decrease in the amplitude of zonal and meridional tidal velocity with distance offshore (Fig. 3a). Meridional amplitudes are consistently larger than zonal amplitudes, and the offshore decrease in meridional amplitude is greater than the offshore decrease in zonal amplitude. The smaller rate of change in the eastern part of the section is because bathymetry gradients are smaller than in the west. Velocity amplitudes were multiplied by the mean water depth in each bin to derive ellipses of tidal transport per unit width (Fig. 3b). Compared with velocity amplitude, transport amplitude changes less markedly with distance offshore, suggesting that the greater tidal velocities observed in shallow water than in deep water are primarily a result of volume continuity rather than the exponential offshore decay of the tidal Kelvin wave.

Glider velocity ellipses compare well with velocity ellipses from the current meter observations and the TPXO model (Figure 3a). Ellipses from both sources indicate clockwise rotation of the tide. The ellipse from the western, single-point current meter observations (Fig. 2a and 3) is likely larger (i.e. indicating faster tidal velocities) than the depth-mean glider and TPXO ellipses because tidal currents at the depth of this current meter (40 m) are less influenced by bottom friction than depth-average velocities. Comparing glider ellipses with the TPXO ellipses at the same location, the difference between the M\textsubscript{2} semi-major axes is 0.10 m s\textsuperscript{-1} (25%) in the western bin and 0.01 m s\textsuperscript{-1} in the central and eastern bins (2% and 4% respectively). The difference between the M\textsubscript{2} semi-minor axes is 0.01 m s\textsuperscript{-1} in all three bins (< 1%, 8% and 15% in the western, central and eastern bins respectively). The phases of the M\textsubscript{2} ellipses differ by 10° (<1%), 7° (8%) and 11° (15%) in the western, central and eastern bins respectively. S\textsubscript{2} semi-major axes differ by 0.05 m s\textsuperscript{-1} (44%) in the western bin and by 0.01 m s\textsuperscript{-1} in the central and eastern bins (8% and 9% respectively). S\textsubscript{2} semi-minor axes differ by 0.01 m s\textsuperscript{-1} in all three bins (48%, 18% and 19% in the western, central and eastern bins respectively). The phases of the S\textsubscript{2} ellipses differ by 7° (5%), 2° (1%) and 17° (11%) in the western, central and eastern bins respectively. Percentage differences between the S\textsubscript{2} ellipses are larger in the western bin because the magnitude of the glider-derived S\textsubscript{2} tide is smaller than in the central and eastern bin. Differences between the ellipses in the western bin could be greater than in the central and eastern bins because the TPXO model is an inversion of satellite altimeter observations, which are less reliable near coastlines (Egbert and Erofeeva, 2002). The glider ellipses are potentially a more accurate characterisation of the tide in this part of the section.

The accuracy of the glider-derived ellipses is dependent on the number of dives that fall within a bin, and the number of bins determines the number of points along the section at which the tide may be resolved. There is therefore a trade off between the number of constituents that can be resolved in the harmonic analysis and the spatial resolution. For a regularly spaced, continuous time series, the Rayleigh criterion (\(\Delta f = 1/T\), where \(\Delta f\) is the difference in frequency between two constituents, and \(T\) is the length of the time series) dictates the minimum length of time series needed to separately resolve constituents of different frequencies. To resolve the M\textsubscript{2} and S\textsubscript{2} constituents from a combined signal, a time series of at least 14.8 days is needed: that is, the cycle introduced into tidal signals by the interaction of the M\textsubscript{2} and S\textsubscript{2} constituents, known as the spring-neap cycle. The Rayleigh criterion is harder to apply to a time series of irregularly spaced DAC, particularly when
temporal discontinuities are introduced by the binning process (Fig. 2). Instead, setting the limits of each bin such that an equal number of dives falls in each ensures that amplitude and phase estimates are of a comparable accuracy across the section. The cumulative length of time that the glider spends in each bin is 14.6, 17.1 and 17.7 days for the western, central and eastern bin respectively, which is approximately equal to, or greater than, the minimum length of time needed to separately resolve the M\textsubscript{2} and S\textsubscript{2} constituents in a regularly spaced, continuous time series. Separating the time series into four or more bins necessarily reduces the number of dives and the length of time the glider spends in each bin. Using four or more bins resulted in S\textsubscript{2} ellipses with unrealistic amplitudes and inclinations, suggesting an inadequate resolution of this weaker constituent.

### 2.3 Glider-derived tidal velocities

To quantify the accuracy of the glider-derived tide (Fig. 4), the along-track velocity time series are compared with the combined M\textsubscript{2} and S\textsubscript{2} along-track time series from the TPXO model sampled at full resolution. Unlike the current meter observations, this model, which is taken to be the best available estimate of tidal velocity, provides estimates of tidal velocity across the entire JONSIS section. The root-mean-square-differences (RMSD) between the glider- and the TPXO-derived tides are 0.03 and 0.02 m s\textsuperscript{-1} for the zonal and meridional velocities respectively. The smaller RMSD for the meridional than the zonal velocities is likely because the meridional velocities are larger and so, assuming the absolute differences between the two zonal and two meridional along-track velocity time series are comparable, the relative difference between the meridional time series will be the smaller.

To determine the extent to which the difference between the glider- and TPXO-derived time series may be attributed to the comparatively low resolution of the glider-derived tide, we firstly compare the fully sampled TPXO tide with the TPXO tide sampled at the same locations as the glider-derived tide (i.e. the centre of each of the three bins, Fig. 2). Velocity time series are extracted from the TPXO model at the central point of each glider bin and at the time of each glider dive. These velocities are interpolated zonally onto the location of each dive, replicating the method used to calculate the glider-derived tide. The RMSDs between the fully and sub-sampled TPXO time series are 0.04 and 0.02 m s\textsuperscript{-1} for the zonal and meridional velocities respectively. Again, the meridional velocities are more similar than the zonal velocities.

Secondly, to simulate a longer glider deployment with more dives from which it would be possible to use smaller spatial bins and so increase spatial resolution, we compare the fully sampled TPXO tide with the TPXO model sub-sampled every 0.5° longitude between 2.5° W and 0°. This decreases the zonal RMSD to 0.03 m s\textsuperscript{-1}; the meridional RMSD remains 0.02 m s\textsuperscript{-1}. Spatial resolution is therefore an important control on the accuracy of the estimated tidal velocity; the accuracy of the glider-derived tide is likely therefore dependent on the number of accurate estimates of tidal amplitude and phase that may be calculated along a section. A longer deployment that enabled the along-track DAC time series to be divided into more longitudinal bins would produce a more accurate estimate of tidal velocity. The ability to use glider DAC to estimate along-track tidal velocity to within ± 0.02 to 0.04 m s\textsuperscript{-1}, even when the tide may be resolved at only a few points along a section, could be of considerable use in regions where tide models are unreliable or unavailable, further enhancing the utility of gliders, for example in remote regions such as the Antarctic shelf.
3 Frontal location

We demonstrate the utility of the glider-derived tide by using these tidal velocities (Figure 4) to study the effect of tidal speed on the location of the tidal mixing front on the JONSIS section observed by the glider. The front is defined to be where the top-bottom temperature difference is 0.5°C. This definition has the advantage of being straightforward to calculate, both from observations and models, and follows the approach used in previous studies (Bowers and Simpson, 1987; Holt and Umlauf, 2008; O’Dea et al., 2012). In place of the amplitude of the M$_2$ tidal speed, the amplitude of tidal speed used to calculate $\log(h/u^3)$ is that of the combined M$_2$-S$_2$ constituents in order to capture changes in tidal speed over the spring-neap cycle.

The combined M$_2$-S$_2$ amplitude is derived from the along-track glider-derived tide: we extract a time series of the maximum speed achieved over each tidal cycle (Fig. 5b) and interpolate this onto the time of each glider dive.

Values of $\log(h/u^3)$ at the front vary considerably over time (Fig. 5a), from below 3 around the 21st October to over 4 around the 12th November. Often, the value of $\log(h/u^3)$ lies outside the range 2.7 ± 0.4 typically used as the critical value for the northwestern European shelf region (Simpson and Sharples, 1994). However, our modified definition of the amplitude of tidal speed (i.e. M$_2$-S$_2$) precludes direct comparison of our values of $\log(h/u^3)$ with those previously published. Using only the M$_2$ tidal speed allows comparison with previous studies, although this results in a $\log(h/u^3)$ distribution that does not account for changes in the amplitude of tidal speed over the spring-neap cycle. Values of $\log(h/u^3)$ at the front when only the M$_2$ constituent is included (not shown) fall between 3.25 and 3.75. These M$_2$-only values cover a narrower range than when the S$_2$ constituent is included and all fall outside the range 2.7 ± 0.4. Hughes (2014), using a heat flux appropriate to the northwestern North Sea, concluded from a modelling study that the critical value for frontal location in the region should be 3.4. This falls within the range of values reported in this study. The higher value of the critical contour may be attributed to the reduced heat flux and enhanced wind mixing at the latitudes of the northwestern North Sea compared with the latitudes of the Celtic Sea (Hughes, 2014), the site of much previous work on the $h/u^3$ criterion (e.g. Simpson and Hunter, 1974). Consequently, our glider observations appear to confirm model-derived predictions that the critical value of $\log(h/u^3)$ is location dependent, and that it falls within a wider range than previously thought – even on the northwestern European shelf for which the range 2.7 ± 0.4 was believed to be appropriate. Furthermore, the calculated values of $\log(h/u^3)$ point to a role for the heating-stirring balance in determining the location of the observed front, despite the front being partially the result of the confluence of the FIC and ESAI (Fig. 1).

There does not appear to be adjustment of frontal location with the spring-neap cycle, although the effects of such adjustment are much greater immediately after frontal development – i.e. in late spring and early summer (Simpson and Bowers, 1981). Instead, the dominant signal in frontal location is its offshore movement over the course of the glider deployment (Fig. 5a).

From a starting longitude of approximately 1.4° W at the start of the deployment, the front moves eastwards into deeper water, reaching approximately 1° W by the middle of November. It then widens considerably towards the end of the deployment, being spread between 1.4 and 0.8° W at the time of the final occupation of the section. A least-squares line of best fit through the frontal locations indicates that the front moves eastward at a rate of 0.53 ± 0.06 km day$^{-1}$. 
3.1 Comparison with model output

To better understand the drivers of the observed offshore frontal movement, glider observations of frontal location are compared with the output of the simple, one-dimensional heating-stirring model of Simpson and Bowers (1984) (see also Elliott and Clarke, 1991, and Simpson and Sharples, 2012). The model is forced with meteorological parameters from the NOCS Flux v2.0 data set (National Oceanography Centre, 2008; Berry and Kent, 2009, 2011) and with tidal speed; it simulates a temperature profile for a water column of a given depth. 55% of incoming heat energy is absorbed at the surface, the remaining 45% being distributed exponentially with depth. This distribution is typical for coastal waters (Ivanoff, 1977). Once heat loss to the atmosphere has been extracted from the surface layer, the additional heat is mixed downwards until the increase in potential energy equals the effective stirring energy input from the wind over the given time step; the profile is then further modified by bottom-up tidal mixing until the increase in potential energy equals the effective stirring energy input from the tide. If the net surface heat flux is negative (i.e. heat loss) the model simulates convection. The new surface temperature is then used to calculate the surface heat flux for the subsequent time step (Simpson and Bowers, 1984). The original model is modified to include the improved parameterisation of heat exchange through the sea surface implemented by Sharples et al. (2006) and Hughes (2014), and to include the additional energy available for bottom-up mixing provided by a constant background flow of 0.1 m s\(^{-1}\) following the method of Hughes (2014). The magnitude of the background flow is chosen to be representative of the values observed in the region (Turrell et al., 1990; Turrell, 1992; Turrell et al., 1996). The effects of a persistent background flow were not included in the original formulation of the \( h/u^3 \) theory, but the presence of Atlantic inflow in the study region makes it a necessary addition.

Temperature profiles are simulated at every 0.1° longitude along the JONSIS section and for every day of the glider deployment at daily resolution. The diurnal heating-cooling cycle is not resolved. A time of representative tidal velocity is selected from the along-track time series (1st November, 14:30), and the combined \( M_2-S_2 \) zonal and meridional velocities at the centre of the three bins are determined for this time. Velocities at these three locations are interpolated onto the location of each model grid point and used to calculate the tidal speed, thus ensuring that the modelling accounts for the observed offshore decay of tidal amplitude (Fig. 3a). We do not simulate the spring-neap cycle because spring-neap frontal adjustment is not observed in the glider data.

Using the same definition of frontal location as used for the glider data, the heating-stirring model places the front in a similar position to the observations during the first four weeks of the deployment (Fig. 5a), albeit approximately 0.1° further west prior to the 15th October. The coastal front that appears in the far west of the section between the 12th and 14th October is not considered here because it is outside the longitudinal range of the glider data. The speed of the background flow has little influence on frontal location while the background flow is slower than 0.1 m s\(^{-1}\). At speeds greater than 0.1 m s\(^{-1}\), increases in background flow speed have an ever greater effect on frontal location, pushing it eastwards into deeper water and causing it to decay earlier. No front forms in the heating-stirring model during the period of the glider deployment at for background flow speeds in excess of 0.31 m s\(^{-1}\).
The similarity between the heating-stirring model (with a realistic background current of 0.1 m s\(^{-1}\) and an intermediate tidal amplitude) and the observations during the first four weeks of the glider deployment suggests that the interaction between surface heat fluxes and tidal mixing explains the location of the front during this period. Specifically, the negative surface heat flux (i.e. heat loss to the atmosphere) during the period of the glider deployment (Fig. 6) means that stratification is maintained only by heat remaining in the water column after the period of summer heating (April to September; Fig. 6). As heat is progressively lost to the atmosphere, the influence of tidal stirring becomes every more dominant, pushing the front into deeper water (Fig. 5). However, the persistence of the observed front after the 17th November 2013 and its slower easterly advance compared with that of the modelled front (1.59 ± 0.08 km day\(^{-1}\); this excludes the secondary front that emerges on the 15th November 2013 around 0.1° W), suggests that the heating-stirring balance is not the primary control on frontal location in the latter period of the glider deployment (i.e. after approximately the 4th November).

### 3.2 Comparison with observations

To understand the controls on frontal location in the latter period of the glider deployment, we examine temperature and salinity observations from the glider. A thermal front, located between 0.6 and 0.9° W (Fig. 7a), is co-located with a horizontal salinity gradient between the relatively fresh coastal water in the west and the more saline water offshore (Fig. 7b). This pattern is typical for the JONSIS section in autumn (Sheehan et al., 2017). The heating-stirring model does not simulate this cool, saline water mass. It is not produced by heating-stirring interactions but instead has its temperature set when the water column is fully mixed during the preceding winter, before it is isolated under the thermocline at the onset of stratification (Svendsen et al., 1991; Turrell, 1992; Hill et al., 2008). Additionally, the southward penetration of Atlantic water from the open northern boundary of the North Sea elevates offshore salinity in the region (Hill et al., 2008). The thermal and haline buoyancy introduced by this water mass sustain the front beyond the time of year at which heating-stirring interactions would sustain the front. As the water column progressively cools throughout the late autumn and winter, removing the thermal signal of the front, the salinity gradient remains, meaning that frontal dynamics – for instance, along-frontal jets – influence the region throughout the year (Sheehan et al., 2017).

We propose that the observed front is a hybrid between a tidal mixing front and a front that forms due to the advection of two different water masses. During summer, the front’s location is principally determined by the surface heat fluxes and tidal mixing according to the mechanism originally proposed by Simpson and Hunter (1974). It is likely that this period coincides with the time during which the surface heat flux (Fig. 6) is positive. Outside of this time, i.e. during the winter and early spring, the location of the front is principally determined by the zonal extent of the cool, saline Atlantic-origin water. The glider observations presented above capture the period during which the front transitions from being a tidal mixing front governed by local heating-stirring interactions, to being governed by the location of two distinct water masses.
4 Summary

Glider DAC observations may be used to determine tidal velocities at the time and location of each dive to within $\pm 0.04$ m s$^{-1}$. The glider-derived tidal velocities compare favourably to output from current meters and the TPXO tide model sampled at the time and location of each glider dive, and particularly well to output from the model when sampled at the centre of each bin used in the calculation of the glider-derived tide. The method enhances the utility of gliders as an ocean-observing platform, particularly in regions where tide models are known to be limited.

Glider-derived tidal velocities were applied to study the location of a tidal mixing front in the northwestern North Sea. The results of a one-dimensional heating-stirring model, and comparison of these results with glider hydrographic observations, demonstrated that salinity gradients and the distribution of water masses are important controls on frontal location in the region, in addition to surface heating and tidally induced mixing. The heating-stirring balance is likely the principal control in spring and summer; salinity gradients and water mass extent are likely the principal controls in autumn and winter. An open question is how the advection of buoyancy influences the front during the summer: specifically, whether it dampens spring-neap frontal adjustment.

Water mass distribution and attendant spatial gradients of thermal and haline buoyancy are likely to be important in shelf sea where significant incursions of oceanic water are found, such as the northwestern North Sea, the South China Sea (Shaw, 1991; Su, 2004), along the eastern coast of the United States (Blanton et al., 1981) and around Antarctica (Moffat et al., 2009). Mixing fronts in such regions may persist during periods when local heating-stirring interactions would not promote frontal formation. Given the thermohaline flows commonly associated with shelf sea fronts, and given the influence that fronts have on the distribution of physical and biogeochemical properties (Turrell, 1992; Hill et al., 2008), this has important implications for the dynamics, ecology and management of shelf sea regions.

Data availability. The glider data used in this study are archived at the British Oceanographic Data Centre.

Competing interests. The authors declare that they have no competing interests.

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References


Figure 1. The location of the JONSIS hydrographic section (cyan line) in the northwestern North Sea. The approximate paths of the Fair Isle Current (FIC) and East Shetland Atlantic Inflow (ESAI) (Turrell et al., 1996) are shown. The area shown in Fig. 2a is enclosed in the orange box. The 100 m isobath is shown in grey.
Figure 2. (a) The location of glider dives along the JONSIS section (cyan line), with dives coloured by longitudinal bin. The location of the current meters is shown by the yellow dots. The 100 m isobath is shown in grey. (b) Zonal and (c) meridional dive-average current velocities, coloured by bin (same colour scheme as panel a).
Figure 3. (a) Tidal ellipses from glider observations (orange), current meter observations (blue) and the TPXO tide model (every fifth grid point, black). (b) Tidal transport ellipses, with colours as in panel a. In both panels, solid lines are for the $M_2$ constituent; dotted lines are for the $S_2$ constituent. The JONSIS section is shown in cyan, the 100 m isobath is shown in grey and land is shaded. Note that the scale of the ellipses is different in each panel.
Figure 4. (a) Zonal and (b) meridional velocity. (c) and (d) are zoomed-in excerpts of panels a and c respectively. The region shown in panels c and d is marked by the grey box in panels a and b respectively. In all panels, the solid black line is the $M_2-S_2$ tidal velocity determined from the glider observations and the dashed orange line is the $M_2-S_2$ tidal velocity from the sub-sampled TPXO model.
Figure 5. (a) Hovmöller plot of $\log(h/u^3)$ (colour scale) from glider observations. Yellow circles mark the observed location of the front. The dashed grey line is the line-of-best-fit through these points. Red circles mark the location of the front as modelled by the heating-stirring model. (b) $M_2-S_2$ tidal speed (m s$^{-1}$; grey line) at the location of each glider dive as calculated using the glider-derived tidal velocities. The red line joins up the maximum speeds and is the estimate of tidal velocity amplitude, $u$, used to calculate $\log(h/u^3)$ in panel a. (c) Bathymetry (m) along the JONSIS section – i.e. $h$ in $\log(h/u^3)$. 
Figure 6. Total surface heat flux (i.e. the sum of latent, sensible, incoming radiative and outgoing radiative fluxes; W m\(^{-2}\)) in 2013 averaged zonally across the JONSIS section. Positive fluxes indicate energy transfer into the ocean. Latent, sensible and outgoing radiative fluxes are calculated by the heating-stirring model using the method of Sharples et al. (2006) from monthly mean meteorological parameters extracted from the NOCS Flux v2.0 dataset (National Oceanography Centre, 2008). Monthly mean incoming radiative flux is extracted from NOCS Flux v2.0. The period of the glider deployment is indicated by the grey box.
Figure 7. (a) Conservative temperature (°C) and (b) absolute salinity (g kg\(^{-1}\)) from the ninth glider occupation of the JONSIS section (14th – 20th November 2013).