Characterizing ventilation events in an anoxic coastal basin: Observed dynamics and the role of climatic drivers

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Abstract

Restricted horizontal exchange and vertical stratification restrict ventilation of bottom waters in many coastal systems, leading to stagnant basin water conditions. Dense inflows from oceanic sources can lead to dramatic changes in basin-scale properties over very short timescales. Here, we used a semi-enclosed coastal basin with deep-anoxia (Lough Furnace) as a test bed, in order to discern estuary-ocean exchange dynamics that lead to irregular and ephemeral deep-water ventilations. Measured densities and current profiler-based estimates of volume fluxes were used to compute typical renewal times of individual water masses in the stratified inner basin. Renewal timescales were primarily determined by freshwater flux and the spring-neap tidal cycle. The deployment period culminated with observations of a ventilation of the basin water which had been stagnant for over 2.5 yr. Entrainment of ambient resident water doubled the volume of the dense plume and diluted its oxygen content. Following ventilation, the volumetric extent of anoxia expanded by 20% within several months, owing to uplifting of old basin water and oxygen consumption in the new basin water. Founded on these observations, a predictive model was constructed relying only on freshwater, atmospheric pressure, and wind data and successfully recreated ventilation events over the past decade. Hindcasting the model back multiple decades implied contrasting periods of higher and lower frequency variability in ventilation occurrence. This model offers a diagnostic tool for assessing how climatic change may influence the oxygen climate in systems that experience contrasting regimes of suboxia and ventilation.

Deoxygenation is a growing global concern across coastal and marine environments (Breitburg et al. 2018; Levin 2018; Oschlies et al. 2018). In many habitats, the leading cause for this concern is anthropogenic eutrophication which increases primary productivity, with subsequent sinking of organic material amplifying respiration rates and oxygen consumption in deeper waters (Diaz and Rosenberg 2008). Climate change may compound these increased rates of oxygen consumption, as warming of near-surface waters decreases oxygen solubility and intensifies vertical stratification, reducing downward mixing of oxygenated surface waters in the process (Keeling et al. 2010). Therefore, it is usually deeper waters, isolated below a strong pycnocline, that are most susceptible to severe deoxygenation. While untenable rates of biological activity associated with nutrient loading can ultimately generate a hypoxic or anoxic environment, the natural hydrography is often the primary factor in determining how prone a system is to oxygen depletion in the first instance (Gray et al. 2002).

One step toward understanding the links between physics and oxygen dynamics is to study naturally anoxic basins. Case studies of anoxic basins have offered considerable insight into the oceanographic control of oxygen dynamics with large-scale examples including semi-enclosed seas, basins, and fjords (e.g., Skei 1988; Murray et al. 1991; Astor et al. 2003; Pawlowicz et al. 2007; Mohrholz et al. 2015; Stigebrandt et al. 2015; Thomson et al. 2017). In some of these systems, anoxia is almost exclusively a function of the natural oceanography (e.g., Anderson and Devol 1973; Skei 1988) while others show the increasingly delicate interplay between hydrodynamics and anthropogenic disturbance in controlling the severity of oxygen depletion (e.g., Capet et al. 2016; Meier et al. 2018). Current evidence suggests that different climate change variables (e.g., increased thermal stratification, altered river flow climatology) could compound the occurrence and severity of

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“dead zones” in coastal regions (Altieri and Gedan 2015). Thus there is a need to discern the precise hydrographic mechanisms that cause the formation, maintenance, and ventilation of anoxic water masses under current climate conditions.

This study used Lough Furnace, located on the west coast of Ireland, as an example of a coastal basin that experiences natural anoxia of its deeper waters (Fig. 1). The system is comprised of a distinct meromictic inner basin (maximum depth 20 m), with deoxygenated conditions typically prevalent at depths below 5–8 m (Parker 1977; Kelly et al. 2018a). The vertical saline stratification blocks direct ventilation of the deeper water by convection in winter and during storm events and strong winds; ventilation is only achieved through lateral inflows of sufficiently dense, oxygenated external water (Kelly et al. (2018a)). The inner Furnace basin communicates through a topographically constrained, shallow (< 2 m) 1-km long connecting channel with the North Atlantic coastal waters of Clew Bay, a large westerly facing bay (Fig. 1a). Tidal currents transport saltwater across this channel and through two narrow topographic constrictions either side of Nixon’s island before reaching the main inner basin (Fig. 1b). Tidal range in the inner basin during spring tides is ~0.3 m. The principal source of freshwater runoff into Furnace comes from the upstream Lough Feeagh, which drains the ~100 km² Burrishoole catchment. Two rivers flow into the northern end of the Furnace main basin (mean annual discharge of 4 m³ s⁻¹), forming a thin, seaward flowing fresh layer usually less than 2 m deep. Lough Furnace represents an ideal natural laboratory for investigating hydrodynamical processes relevant to larger-scale stratified estuarine basins. Furnace has also served as a model system for important ecologically motivated research including the population dynamics of euryhaline invertebrates (Parker and West 1979), a paleolimnological reconstruction of its evolution toward anoxic conditions, which appears to have occurred at about 3400 calendar years before present (Cassina et al. 2013), divergent evolution in fish ecotypes (Ravinet et al. 2015), biophysical interactions between benthic fish and internal waves (Kelly et al. (2018b)), and dynamics of subsurface chlorophyll maxima (de Eyto et al. 2019). Although deoxygenation in open coastal areas and well-flushed estuaries was reported as being rare in an Irish context (O’Boyle et al. 2009), documented examples of other perennially or periodically deoxygenated coastal systems along the Atlantic coast of Ireland include Ardbear Salt Lake (Henry et al. 2008) and Lough Hyne (Trowbridge et al. 2017).

The fundamental hydrographic parameter toward determining how susceptible a given waterbody is to deoxygenation is the frequency at which resident water is exchanged for new,
oxygenated water. Use of precise definitions in reference to transport time scales in waterbodies (e.g., residence time, age, flushing rate) is critical (Monsen et al. 2002). With this in mind, here we use the term renewal time (T_R) to refer to the time required for the entire volume of the basin to be replaced by new water from either freshwater or tidal sources (following Andutta et al. 2014). This renewal time is not a measure of the transport and fate of individual water parcels through the system but rather a simple integrative Eulerian measure at a fixed spatial and temporal point based on basin volumes and advective/diffusive fluxes. Thus in systems with long renewal times, the exchange rate is low and the potential for exchanging oxygen-poor with oxygen-rich water becomes limited.

A simple exchange rate assumes 100% mixing efficiency between advective fluxes of external water into the basin and the flushing out of prior resident water. This assumption becomes more problematic in stratified basins, where vertical density gradients introduce a complexity, as different water masses of specific densities will have different renewal times, with significant replacement of resident water dependent upon advective fluxes of new water of similar density. Thus a simplistic yet useful approach is to divide stratified basins into appropriate layers of uniform densities (mixed-boxes) and derive individual flux rates and renewal times for each layer. In estuarine coastal systems, this typically includes a surface freshwater layer (here denoted by L_1), an intermediate tidally exchanged saline layer (L_2) and in the case of confined basins deeper than their inflow/outflow channels, a third layer below the extent of the channel depth containing the densest resident basin water (L_3). The usefulness of variations of this layered box model approach is evident through its successful application to a variety of coastal systems that could be classified as regions of restricted exchange (e.g., Babson et al. 2004; Pawlowicz et al. 2007; Stigebrandt 2012; Gillibrand et al. 2013). A typical temperature-salinity vertical profile taken during stagnant conditions in 2016 reveals the applicability of the three-layered approach toward the hydrography of Lough Furnace (Supporting Information Fig. S1).

Renewal of the deep basin water L_3 (termed ventilation in cases of suboxic waters) is of particular interest as it can cause profound biogeochemical changes over a very short timeframe (Arneborg et al. 2004; Lehmann et al. 2015; Bergen et al. 2018). Of particular concern are the potential impacts of releasing toxic hydrogen sulfide (H_2S) gas, which has accumulated in anoxic deep water, to aquatic life in oxygenated surface waters (Luther et al. 2004) or to the atmosphere (Capet et al. 2016). In Furnace, deep-water ventilations have been linked to mass fish kills (Anon 1995) and phytoplankton population dynamics (de Eyto et al. 2019).

The parent water of L_3 will be the dense water associated with the most recent deep-water renewal, with the subsequent decreases in density occurring very slowly through diffusivity with the water in the overlying layer (e.g., Farmer and Freeland 1983; Stigebrandt 2012). This gradual reduction in density will eventually precondition the resident basin water to a replacement by an inflow carrying water of sufficient density; the slow nature of this process however means that renewal of the basin water based on this mechanism alone is typically of order years (Farmer and Freeland 1983). Understanding the mechanisms that control the temporal variability of the inflow density and volume fluxes at the entranceway of the basin is therefore crucial in determining a typical renewal time of the interior waters and has been a focus of estuarine and coastal oceanography for several decades (e.g., Geyer and Cannon 1982; Griffin and LeBlond 1990; Allen and Simpson 1998; Schinke and Matthäus 1998; Laval et al. 2003; Thomson et al. 2017). However owing to their stochastic nature, few direct observations have been made during a dense, oxygenated intrusion (e.g., Liungman et al. 2001; Arneborg et al. 2004; Holtermann et al. 2017). More generally, the dynamics associated with the intrusion of dense gravity currents into ambient basin water form a key component of overall vertical circulation and sedimentation in oceanic basins (e.g., Ivanov et al. 2004; Wåhlin and Cenedese 2006) and lakes (Fer et al. 2001).

This article is structured into two main parts. First, observations of currents, temperature, salinity, water levels, and meteorology over a 15-month period are presented and analyzed in an attempt to estimate freshwater and saltwater volume fluxes between Lough Furnace and the adjacent open coastal waters, with the objective of calculating renewal times in the main inner basin. The motivation was to develop a process-based understanding of why ventilation of the deep anoxic water occurs based on observations. The second part of the article involves the development of a predictive model informed by these observations capable of successfully predicting when a ventilation event might occur. We utilized an empirical approach rather than a numerical model to provide a simple, computationally inexpensive assessment of how climatic changes may potentially alter deep-water oxygen dynamics in Furnace and similar systems.

**Methods**

**Instrumentation and observations**

From 16th March 2016 to 3rd July 2017, the deepest (mean depth 5.8 m) section of the entrance channel to the main inner basin of Lough Furnace was instrumented with an upward looking 1-MHz Nortek Aquadopp 3-beam current profiler (Nortek AS, Rud, Norway) (MR01; Fig. 1b,c). The current profiler was mounted to a steel frame and the sensor head, with a 25° beam angle to the vertical, was fixed at 0.2 m above the bed. An L-shaped mooring configuration was used to avoid measurement interference by the surface marker buoy and line. The current profiler recorded in 0.5 m bin ensembles, with a blanking distance of 0.4 m and a velocity measure averaged over 60 s using a 2.6 Hz sampling (ping) rate. Two SBE-37 MicroCATs (Sea-Bird Scientific, Bellevue, Washington, U.S.A.) recording temperature (resolution = 0.001°C) and conductivity
(resolution = 0.001 mS) were also attached to the same mooring with one MicroCAT mounted on the bottom current profiler frame and the second MicroCAT positioned ~ 1 m below the surface on the L-shaped mooring. An additional MicroCAT, also recording pressure, was located further downstream closer to the estuary proper, on the lough side (upstream) of one of the entrance constrictions (MR02; Fig. 1b). All of these instruments had synchronized internal clocks and logged measurements every 30 min. Mooring turnaround occurred at approximately 4–6 week intervals, for data downloads, instrument cleaning, and (current profiler) battery replacement.

An automatic monitoring platform located in the central main basin (AWQMS; Fig. 1b), recorded four-daily full water column profiles (0, 6, 12, and 18 h) of temperature, conductivity, and dissolved oxygen using a Hydrolab DSS datasonde (OTT, Kempten, Germany) attached to an undulating winch profiler. The monitoring platform also included a surface meteorological station with wind speed and direction, air temperature, and solar radiation measured every 2 min. Missing gaps in the wind data were filled using a regression model with a Met Éireann station located directly on the shoreline of Furnace (MET; Fig. 1b)—this land station also recorded rainfall. A tidal gauge was located at the northern landward limit of the main Furnace basin (TG01; Fig. 1b) and was installed in the lower reaches of the Burrishoole estuary (TG02; Fig. 1b) midway through the deployment in November 2016, which provided a direct measure of tidal influence from the adjacent coastal waters.

Data processing

Profiles from the automatic monitoring platform were processed following Kelly et al. (2018a), with individual profiles extrapolated onto a standard depth grid spanning the water column from 0.6 to 12.9 m in 0.15 m depth increments. Harmonic analysis of the time series of water levels recorded at TG02 was performed to obtain tidal constituents, using the “tidein” function from the “tide” R package (Kelley and Richards 2018). Using these constituents, a tidal model was then constructed and used to predict a time series of tidal water levels from the automatic monitoring platform were then rotated into along-channel (V) and across-channel (U) components using principal components analysis (PCA) (Emery and Thomson 2001). Wind speed and direction were decomposed into zonal and meridional vectors and given the wind measurement site (AWQMS, see Fig. 1b), true north and east essentially corresponded to the along- and across-channel wind stress at MR01 without requiring rotation. In order to assess the influence of coastal winds on the density of tidal inflows into Furnace, a separate set of wind vectors were derived using the same wind parameters measured at Furnace but rotated into along-shore (U) and across-shore (V) components relative to the coastline of Clew Bay following PCA (Fig. 1a). Wind stress components \( \tau_x \) and \( \tau_y \) were computed from the equation \( \tau_{xy} = \rho_{air} c_d W_{xy}^2 \), where \( \rho_{air} \) is the density of air (kg m\(^{-3}\)), \( c_d \) is a surface drag coefficient (computed based on wind speed and atmospheric stability; Woolway et al. 2015), and \( W_{xy}^2 \) is windspeed squared at 10 m above the water surface in the x or y direction.

Volume flux and renewal time calculations

Volume fluxes in and out of the main inner basin through the deepest part of the channel were determined from the along-channel low-pass filtered horizontal velocity vectors. A volume flux \( Q \) through the entrance channel is defined as:

\[
Q = \int_A v \, dA
\]

(1)

where \( v \) is the along-channel velocity and \( A \) is the depth-varying cross-sectional area. Volume fluxes were computed through each current profiler bin based on channel width at each bin depth; an echosounder transect of the entranceway provided accurate bathymetry (Fig. 1c). While the standard current profiler bin sizes spanned 0.5 m in the vertical, the lowest bin was weighted to account for the bottom 1.1 m of the channel (i.e., 0.5 m bin size + 0.4 m blanking distance + 0.2 m frame height above the bed). For each individual current profile, the uppermost usable bin was defined as the shallowest bin below the computed depth of sidelobe interference.

The upper halocline in Furnace was often very close to the surface (< 1.5 m deep for one-third of the deployment period) and it is probable that using extrapolated velocities from below this depth would not accurately capture shallow flow dynamics during periods with a thin surface layer. Furthermore, the entrance section widens significantly at depths shallower than 2 m (Fig. 1c) and using a constant velocity from the current profiler moored in the deeper trough section to approximate volume transport through this unmonitored, shallow rocky region to the east yielded erroneous flux estimates. Thus, an alternative method used the current profiler to directly estimate volume fluxes through the lower layer \( Q_2 \) by integrating velocities up to the depth of the density interface \( \eta_1 \) that divided the fresher upper \( L_1 \) and saline lower \( L_2 \) layers such that:

\[
Q_2 = \int_{-H}^{\eta_1} u \, W \, dz
\]

(2)

where \( u \) and \( W \) are along-channel velocity and width at the entranceway at depth \( z \) and \( H \) is the depth to the channel
bottom. Using this configuration, positive (negative) values indicate flow into (out of) the inner basin through the lower layer. A timeseries of $n_1$ was inferred using daily averaged salinity profiles from the automatic monitoring platform and calculating the shallowest depth at which the salinity differed from that at the surface by a value of 2. This finite difference criteria was chosen after comparison with gradient-based approaches (e.g., depth of max buoyancy frequency or salinity gradient); the latter was often inaccurate during periods where the greatest salinity gradient occurred at the deep halocline, meaning internal density interfaces could be considered horizontal at the low-passed velocities considered here. Renewal times of the intermediary layer $T_{fl}$ were calculated by $V_2/Q_{2in}$, where $Q_{2in}$ is the inflow contribution through the lower layer during periods when the inflow density ($\rho_{in}$, measured by the bottom temperature-salinity sensor at MR01) was less than the density of the deep basin water ($\rho_3$, computed from profiles recorded by the automatic monitoring platform).

Separate renewal time estimates of the deep basin water ($L_3$) were derived depending on whether the renewal or stagnant regime prevailed. When $\rho_{in} < \rho_3$, no advective fluxes of external water occurs and density reduction in the basin water is primarily dependent upon diffusive fluxes with the overlying water (e.g., Stigebrandt 2012). A diffusive flux between $L_2$ and $L_3$ was estimated using observed data as:

$$Q_{ks} = \frac{V_3 \Delta \rho}{\frac{d \rho}{dt}}$$

where $V_3$ is the basin water volume and $\Delta \rho$ is the density difference between $L_2$ and $L_3$. If $\rho_{in} > \rho_3$, a switch was initiated in the calculation workflow and an estimate of basin water renewal time was given by $V_3/Q_{2in}$.

### Predictive model

In an effort to predict the occurrence of deep-water ventilations, and specifically to assess whether the frequency of events has changed in recent decades, a simple model was constructed. A binary categorical response variable at a weekly timestep was created to define whether ventilated or stagnant basin water conditions prevailed. Categories were based on whether the bottom water of the inner Furnace basin was anoxic (< 0.1 mg O$_2$ L$^{-1}$), used due to the oxygen sensor not being calibrated to complete anoxia and generally not reading 0 mg O$_2$ L$^{-1}$) or not (> 0.1 mg O$_2$ L$^{-1}$) when averaged over 1 week. Because the objective of the modeling exercise was to predict ventilation events over a timeframe longer than the 15-month deployment period, independent variables in the model were restricted to data sources that could be derived over the period 2009–2018, for which dissolved oxygen profiles were available to validate against model predictions (Kelly et al. 2018c). It was assumed a priori that the occurrence of dense oxygenated inflows to the inner basin would be dependent to some extent on the local magnitude of freshwater discharge and this assumption is validated in the Results section. The potential influence of coastal dynamics on deep-water ventilations were also assessed by including atmospheric pressure and along-shore wind stress relative to the Clew Bay coastline as predictors. Thus predictor variables included timeseries of mean weekly freshwater discharge, wind, and atmospheric pressure. To allow a multidecadal assessment of how deep-water oxygen dynamics may have evolved, we used
reanalyzed wind vectors and sea level pressure from the ERA5 atmospheric reanalysis, which covers the period 1979 to present (Copernicus Climate Change Service 2017), instead of local observations from the shoreline meteorological station which were only available from 2005 onward. Variables were extracted from the 0.25° × 0.25° grid containing Clew Bay; reanalyzed variables showed remarkably close agreement with local observations over the period 2009–2018 ($R^2 = 0.91 \ [p < 0.001]$ and $R^2 = 0.99 \ [p < 0.001]$ for wind vectors and surface pressure, respectively), providing justification for the use of the historic ERA5 variables. To account for possible interactions between freshwater and meteorological conditions during prior spring and neap tidal phases, a trailing 30-d running mean of freshwater discharge, wind stress, and atmospheric pressure were used.

A machine learning approach was adopted which allows a set of input values to be mapped to output values, crucially without requiring numerical interpretation of the mechanisms involved. The advantage of this method were: (1) driving variables can be limited to commonly available data sources and exclude site specific parameters (e.g., local mixing dynamics) and thus could be applied to similar systems; (2) using a traditional statistical model may have violated some assumptions of independence (e.g., freshwater discharge, atmospheric pressure, wind stress all depend on synoptic scale meteorology); and (3) such an empirical approach is relatively simple and computationally inexpensive compared to configuring and executing a full hydrodynamic model. The obvious major drawback of this method is that it omits an understanding of the underlying physics and thus will not account for changes in the overall statistics of the system.

A random forest algorithm (Breiman 2001) was implemented (following Liaw and Wiener 2002). A general overview of the random forest procedure applied to a categorical response variable (basin water ventilation or stagnation in this case) is as follows: multiple decision trees are independently constructed, each using a different bootstrap sample of the data set (bootstrap aggregating; see Breiman 1996). Each tree structure is a hierarchical classification model, where tree root splits represent decision nodes. For each tree, each decision node is optimized for only a subset of the predictor variables, randomly chosen at each node (in contrast, a standard classification tree selects an optimal split amongst all predictor variables at each node). This additional layer of randomness in the decision procedure gives random forests added robustness against overfitting compared with traditional regression and classification trees (Breiman 2001). The predictions from each tree are then aggregated and a final classification consensus is reached by selecting the most common output. An error rate is estimated for each tree based on its ability to correctly predict data not included in the bootstrap sample used to construct that tree (“out-of-bag” error rate). Out-of-bag errors from each tree are then aggregated to give a mean overall error rate.

A primary motivation to using a random forest approach was that splitting the data into training and test sets was not strictly necessary, as the method protects against overfitting.

**Fig. 2.** Probability histograms of salinity (top) and density (bottom) based on measurements made in the entrance to the inner basin at MR01 (blue—surface layer, red—lower layer) and in the center of the inner basin at the automatic monitoring platform (black solid line—full water column, dashed black line—water column only to mean depth of oxycline).
by using a bootstrap sample of the data and a randomized variable selection subset to construct each tree. Thus, it was decided not to split the observation data given the relatively small sample size ($n = 500$, for weekly oxygen data spanning January 2009–July 2018). The authors’ R code and data necessary to run the model are available at: https://github.com/IrishMarineInstitute/BurishooleLTER-Public/tree/master/PredictiveModel_oxygen.

Results

Exchange flow characteristics

A histogram of salinities recorded at the surface (blue) and bottom (red) of the entranceway for the deployment period are shown in Fig. 2. Black lines are histogram density curves for salinities recorded from the inner basin over the same timeframe (normalized for the same number of observations), with the solid black line including salinities throughout the full water column and the dashed line including salinities down to the average depth of the oxycline only (6 m). This plot illustrates that separate lower and higher salinity water masses occurred in the entranceway, which correspond to the waters above and below the upper halocline in the inner basin ($L_1$ and $L_2$). It also shows a third water mass in the inner basin below the deep halocline and oxycline depth, with salinities typically higher than those of any other water mass ($L_3$ [deep basin water]). A histogram following the same criteria for density closely resembled the salinity histogram highlighting the minimal influence of temperature on density variations in the system (Fig. 2). The appearance of water of salinity (viz. density) higher than the basin water at the entranceway is thus the necessary condition for a large-scale exchange of the stagnant deep water.

The structure and temporal variability of flow through the entranceway to the main inner basin is evidently dependent upon tidal, freshwater, and wind forcing (Fig. 3). The strongest flows occurred at the bottom and nearer the surface (visible immediately below the sidelobe cut-off depth) with generally weaker flows at intermediate depths.

The low-pass filtered mean flow often featured baroclinic exchange (Fig. 3a). This was most apparent during spring tides, with positive velocities (into Furnace inner basin) near

![Fig. 3](https://github.com/IrishMarineInstitute/BurishooleLTER-Public/tree/master/PredictiveModel_oxygen)

**Fig. 3.** Timeseries of (a) along-channel (350°T) current velocity profiles at MR01 (black line is surface water level), (b) salinity (black) at MR02 with 40-h low-passed tidal elevations at TG02 (red), (c) surface (blue) and bottom (red) salinities at MR01, (d) freshwater flux ($Q_F$), and (e) along-channel wind-stress at MR01 (positive [red] is toward inner basin and negative [blue] out of inner basin). Vertical dashed lines indicate period of deep-water ventilation.
the bottom (\sim 2 \text{ m above the bed}) sometimes exceeding 0.2–0.3 m s\(^{-1}\). At neaps, inflow nearer the bottom was attenuated or reversed completely. Currents closer to the surface (within \sim 2 \text{ m}) showed less fortnightly tidal influence and outflow (negative velocities) was the primary flow dynamic. At intermediate depths (2–4 m), weaker outflow was predominant, with occasional weak inflows during spring tides. The influence of wind was also apparent with some reversals of the surface outflow during strong up-estuary winds (e.g., early October 2016, late January 2017, late March/early April 2017; Fig. 3a,e). Similarly, surface outflow was often accelerated during strong down-estuary winds (e.g., April/May/June 2016, November 2016; Fig. 3a,e). The extent of current profiler measurements meant that some of the finer details of wind influence on surface layer reversals may not have been adequately captured.

The influence of spring-neap tidal forcing on salinity was most apparent at the shallow MR02 site, with the minimal salinity fluctuations observed during mid-neaps implying that generally nominal quantities of saltwater reached the Furnace basin during these times (Fig. 3b). During large freshwater events (e.g., September 2016, March 2017), the overall salinity at MR02 was reduced even at spring tides, indicating that tidal inflows were diluted further downstream by the flushing out of this lower salinity water from upstream or perhaps blocked by large volumes of freshwater outflow (Fig. 3b,d). Surface and bottom salinities at the entranceway (MR01) remained distinctive throughout the 15 month deployment with stratification being maintained (Fig. 3c). Lower layer salinities were more stable over the period with a mean value of 17, with short-lived decreases during freshwater events (e.g., March 2017) and gradually increasing over consecutive spring tides during dry spells (e.g., April/May 2017). Above the halocline in the freshwater-influenced surface layer, salinities were generally lower (mean value of 4.6) but during periods (weeks) of low freshwater discharge, surface salinities converged toward lower layer salinities during spring tides (June 2016, late October

### Table 1. Advective volume flux estimates in/out of the inner Furnace basin. Positive (negative) values indicate fluxes into (out of) the inner basin.

| Flux  | Mean (\pm SD) (m\(^3\) s\(^{-1}\)) | Min/max (m\(^3\) s\(^{-1}\)) |
|-------|----------------------------------|-----------------------------|
| \(Q_f\) | 3.30 (\pm 3.26)                  | 0.43/28.56                  |
| \(Q_1\) | -3.45 (\pm 4.97)                 | -27.55/22.11                |
| \(Q_2\) | 0.155 (\pm 4.28)                 | -24.46/22.63                |
| \(Q_{\text{net}}\) | -3.29 (\pm 3.28)                | -25.94/22.26                |

**Fig. 4.** Variance-preserving spectra of along-channel (350°T) velocities at MR01 recorded over the full deployment period (16\(^{th}\) March 2016–7\(^{th}\) July 2017). Depths are meters above the bed.
2016, May 2017; Fig. 3c). The general increase in bottom layer salinity from the onset of the deployment should be noted, as the 2016 spring period followed an abnormally wet winter period 2015/2016, with lingering effects on salinity levels in Furnace still apparent in early spring.

Variance-preserving spectra of the unfiltered along-channel currents for each depth section of the water column are displayed in Fig. 4. Currents at all depths of the water column show pronounced spectral energies associated with the semidiurnal ($M_2$) frequency band. Higher order harmonics ($M_4$ [3.8 cpd] and $M_6$ [5.79 cpd]) were also observed, which are associated with tidal asymmetries between the flood and ebb arising from the distortion of $M_2$ tidal currents through nonlinear interactions as the water depth shallows (Friedrichs and Aubrey 1988). The bottom (black line) portion of the water column notably showed increased spectral peaks at 0.06 cpd (spring-neap tides) and all depths showed significant spectra at 1 cpd.

### Volume fluxes

Volume fluxes in and out of Furnace showed considerable variation around mean values (Table 1.). Using $Q_2$ and $Q_3$, the mean volume transport through the surface layer $Q_1 = -3.45 \text{ m}^3 \text{s}^{-1}$. This value equated to the combined fluxes

### Table 2. Renewal time ($T_R$ in days) of each water mass in the inner Furnace basin. $T_R$ values shown here represent an average over 7-d periods ($T_{R\text{stagnant}}$ refers to time periods when renewal of basin water occurred due to diffusive fluxes only; $T_{R\text{renewal}}$ refers to periods when advective fluxes to the basin water occurred).

| Layer          | Mean (median) | Min/max       |
|----------------|---------------|---------------|
| $T_{R1}$       | 8.5 (7.3)     | 2.4/29.8      |
| $T_{R2}$       | 46.0 (31.6)   | 7.2/254.6     |
| $T_{R\text{stagnant}}$ | 1663.3 (1347.7) | 309/4789.6    |
| $T_{R\text{renewal}}$ | 13.3 (13.3)   | 14.3/12.2     |
from catchment runoff plus volumetric fluxes of precipitation and evaporation over the surface area of Furnace ($Q_p = 3.30 \text{ m}^3 \text{s}^{-1}$) and the net lower layer volume transport directly measured by the current profiler, which over the full deployment provided a small net mean inflow contribution of $Q_2 = 0.155 \text{ m}^3 \text{s}^{-1}$. Therefore, the overall mean volume flux over the full deployment period through the entranceway, $Q_{net} = -3.29 \text{ m}^3 \text{s}^{-1}$. This value was comparable to the mean volume flux estimate of $-3 \text{ m}^3 \text{s}^{-1}$ measured over summer 2010 by a current profiler positioned in shallower water further downstream (Kelly et al. 2018a).

From March 2016 to March 2017, the basin water density decreased at a rate of $1.7 \times 10^{-7} \text{ kg m}^{-3} \text{s}^{-1}$. A timeseries of density differences between $L_2$ and $L_3$ over the same period had a mean value of $\Delta \rho = 2.4 (\pm 1.55) \text{ kg m}^{-3}$, thus giving a diffusive flux to the basin water from above of $Q_K = 0.023 (\pm 0.019) \text{ m}^3 \text{s}^{-1}$. Mean volumetric fluxes associated with direct precipitation ($Q_p = 0.036 [\pm 0.05] \text{ m}^3 \text{s}^{-1}$, based on a mean rainfall rate of $3.6 \text{ mm d}^{-1}$) and evaporation ($Q_e = 0.019 [\pm 0.01] \text{ m}^3 \text{s}^{-1}$, based on a mean open water evaporation rate of $1.7 \text{ mm d}^{-1}$) over the small surface area of the inner basin provided only minor contributions to the overall water

Fig. 6. Observations of lead-up to and occurrence of deep-water renewal and ventilation on 9th May 2017. Timeseries of contours measured at the automatic monitoring platform in the central inner basin of (a) salinity, (b) temperature, and (c) dissolved oxygen. (d) Salinity (red), temperature (black), and oxygen (blue) recorded at the bottom mooring MR01. (e) Density recorded at MR02 (black), MR01 bottom (red), and mean bottom basin water density recorded at the automatic monitoring platform (blue).
Groundwater inflows were not accounted for and thus likely contribute a (small) error to the overall estimates presented in Table 1.

Timeseries of upper layer ($Q_1$; Fig. 5a, thick black line) and lower layer ($Q_2$; Fig. 5a, thin gray line) mean daily volume fluxes through the entranceway to the inner Furnace basin both showed contrasting periods of net positive and negative fluxes into and out of the inner basin. The black dashed line shows the transition between positive and negative fluxes with opposite direction volume transport between the two layers a predominant feature, which was anticipated based on the observed baroclinic flow structure previously described (Fig. 3a). The magnitude of negative volume fluxes generally exceeded positive fluxes and this is clearly observed in the timeseries of net volume transport (Fig. 5b) with only weak and rare occasions of net inflow. Significant temporal variability around the overall mean volume flux was observed (Fig. 5b, solid black horizontal line and Table 1). Large negative fluxes in Fig. 5b were associated with freshwater events and overall volumes through the mouth closely matched $Q_F$ (Fig. 3d), indicating the dominating influence of freshwater outflow on flux dynamics, specifically through the upper layer.

$Q_2$ showed spring-neap fluctuations (Fig. 5c). During large spring tides, typical volume fluxes of $\approx 10$ m$^3$ s$^{-1}$ were estimated, which were an order of magnitude larger than the mean volume transport to the inner basin over the full deployment period (Table 1.). A maximum influx of (+) 22 m$^3$ s$^{-1}$ was recorded in April 2016, which occurred during a 2 month period of persistent inflows. This period was relatively unusual with regard to the frequent southward blowing winds (Fig. 3e), which often attained windspeeds greater than the average for the region ($\sim 5$ m s$^{-1}$; Kelly et al. 2018a). These down-estuary winds appeared to reinforce the strength of the estuarine circulation. However, the salinity (density) associated with these large volume fluxes was not great enough to renew the deep basin water at this time.

During neap tides, negligible amounts of tidal water reached the entranceway to the main basin and generally negative fluxes were observed near the bottom (Fig. 5c). There was evidence suggesting that freshwater conditions increased the magnitude of these negative fluxes through the lower layer at neap tides. Larger negative fluxes were observed over autumn and winter months (September through March) when freshwater discharge was higher compared to spring and (early) summer months (April through July) when drier freshwater conditions prevailed (Fig. 3d for freshwater discharge; Fig. 5c for lower layer net fluxes).

The net volume flux through the lower layer was partitioned into contributions in and out for each 30 min timestep, using velocities measured through each individual depth bin of the current profiler (Fig. 5d). As expected, the volume transport was

![Fig. 7. Temperature-salinity diagram for inner basin water column profiles (lines with symbols) and bottom inflow water through entranceway recorded at MR01 (sole symbols) in the lead up to and following deep-water renewal event on 9th May 2017.](image-url)
primarily either into or out of the inner basin, although there were instances of simultaneous positive and negative fluxes of comparable magnitude (e.g., February and April 2017). This dynamic was highly relevant when estimating renewal times for the inner basin’s intermediate and deep saline waters as it meant that during periods of significant outflow, small positive fluxes of salt water may still occur. It also highlights the complexity of the flow through the shallow entranceway and may be related to tidal asymmetries and the differential influence of tidal forcing on currents closer to the bottom compared to currents at intermediate depths (Fig. 4).

**Renewal times**

Using a mean daily timeseries of the relevant volume flux, renewal times were calculated for each water mass and then averaged over 7 d periods. Renewal time summary statistics for $L_1$, $L_2$, and $L_3$ are displayed in Table 2. Separate renewal times are presented for $L_3$ based on whether renewal was primarily a function of advective or diffusive fluxes.

**Basin water renewal and ventilation**

On 9th May 2017, a significant renewal and ventilation of the deep basin water occurred (Fig. 6). Based on the prolonged dry weather conditions that occurred, a ventilation event was anticipated and a dissolved oxygen sensor (miniDOT) was deployed near the bed next to the MR01 mooring (Fig. 1b) to monitor oxygen of the inflowing water. The change in basin water properties from this date onward (Fig. 6a–c) indicated that the source water for the inflow was comprised of the warm, salty, and oxygenated bottom water (relative to the ambient resident water) presented at the entranceway (Fig. 6d). The necessary condition for this large-scale basin renewal was when the bottom layer inflow density (red line) equaled or exceeded the mean basin water density (blue line, $\sim 1015$ kg m$^{-3}$) during a spring tidal phase (Fig. 6e). A partial renewal, limited to the uppermost part of the deep basin water, occurred $\sim 7$ d prior during the preceding spring tides, when the inflow density briefly exceeded the mean basin density during flood tides (27th April onward, Fig. 6e). The bottom density at MR01 gradually increased over successive spring-neap tidal cycles in the lead up to the ventilation, as denser (saltier) inflows reached the inner Furnace system at flood tides and the mean density during neaps remained relatively high as denoted by the density timeseries at MR02 (black line, Fig. 6e).

Temperature-salinity (T-S) relations (Kelley and Richards 2018) were utilized to trace basin water properties at 4-d intervals relative to the external water recorded at the entranceway prior to and during the renewal event (Fig. 7; lines with symbols...
represent T-S curves from the automatic monitoring platform depth profiles [AWQMS, Fig. 1b], single symbols represent corresponding T-S relations of bottom water at the entranceway for the same date). Initially on the 6th May (prior to renewal [black, circles]), the basin water was saltier and warmer than the inflow. On the 11th May (red, stars), the inflow water was warmer and saltier (denser) and a new warmer water mass is observable along the basin water T-S curve. The salinity of this warmer water is slightly higher than the saltiest resident basin water and much lower than the water presented at the entranceway, indicating that substantial mixing occurred between resident water and the inflow as it entered the inner basin. This created a water mass of intermediate density, warmer than the ambient basin water, with slightly higher salinity and occupying roughly the same density space on the T-S plot as the old basin water (when comparing the densest portions of the black and red T-S curves). However, water warmer and saltier than the resident basin water was continually presented at the mouth over the following 4 d and on 15th May (blue, triangles) a new arm can be seen on the T-S curve for the basin water, this time containing warmer water that was salty enough to surpass the density of the old basin water, reaching densities > 1015 kg m$^{-3}$. The position of this new arm between the old resident water and the inflow implies that the new basin water was a mixture of old and external water. A second postrenewal T-S curve 4 d later on 19th May (green, diamonds) showed little change in the properties of the new basin water. A reduction of the inflow density compared to 15th May meant that the upper limit of new basin water density had already been reached. This was related to the transition from spring to neap tides (Fig. 6e) and a rainfall event on 15th May (already observable in the decreased surface salinities of the 15th May T-S profile), which together brought about a conclusion to the major renewal event which only lasted 7 d.

The mean volume flux into the basin through the lower layer calculated directly from the current profiler during the 7 d period (between the dashed vertical lines; Figs. 3, 5) was 2.02 ($\pm 0.83$) m$^3$ s$^{-1}$. This estimate was two-orders of magnitude higher than the average diffusive flux computed during the stagnant period. Using the volume fluxes calculated during the renewal, the basin water would take 13.3 d to renew completely, meaning that only a partial renewal occurred. The mean oxygen content of the volume flux from 10th–15th May was 8.5 mg L$^{-1}$ (Fig. 6d).

Some of the old basin water was displaced upward to an intermediate depth ($\sim 3.5$–5.5 m) in the aftermath of the initial deep-water inflow interleaving at a greater depth, where it remained as an isolated band of salty, cooler, oxygen-depleted...

**Fig. 9.** Validation of predictive model of deep-water ventilation in Furnace 2009–2018. (a) Probability of ventilation occurring predicted by random forest classifying ventilation or stagnation based on input variables (horizontal dashed line is 80% probability). (b) Timeseries of observed bottom water oxygen measured at the automatic monitoring platform (Fig. 1b).
water (Figs. 6a–c, 8). This water experienced slower rates of renewal than the previous intermediary water, given that its density was higher than typical inflow densities. In contrast, the newly ventilated basin water returned to a stagnant regime once the dense oxygen-rich inflows ceased (Fig. 6) and by September 2017, fully anoxic conditions had reoccurred (Fig. 8). Somewhat counterintuitively, 4 months postventilation the vertical extent of anoxia in the inner basin had increased with a shallower oxic-anoxic interface (6.6 m up to 4.2 m) and the overall volume of anoxic water had expanded from 34.5% to 53.5% (Fig. 8).

**Predictive model**

The random forest model had an average out-of-bag error estimate of 11%, that is, there was, on average, an 11% chance that a classification tree would wrongly classify a ventilated or stagnant bottom water regime using new predictor data. Freshwater discharge had the largest influence on predictive capabilities, followed by sea level pressure and along-shore wind stress in Clew Bay (Supporting Information Fig. S2). Low freshwater discharge, high atmospheric pressure, and negative (offshore) wind stress were implicated in increased likelihood of ventilation. The random forest model was used to predict the probability of a ventilation occurring in a given week and predictions were compared with actual observations of mean weekly deep-water oxygen for the period 2009–2018 (Fig. 9). Results showed that when the random forest model predicted that the probability of a deep-water ventilation was ≥ 80%, a ventilation occurred as denoted by the sudden increase in deep-water oxygen. Thus, we used this probability as cut-off criteria to predict when a ventilation would occur for periods with no observations of bottom oxygen (pre-2009).

**Discussion**

In aquatic systems, the oxidation and breakdown of organic matter creates a biological oxygen demand (BOD), which utilizes oxygen in the water column (e.g., Robinson 2019) and sediments (e.g., Livingstone and Imboden 1996). In order to maintain oxidized conditions, the rate at which oxygenated water is supplied to a basin must exceed this rate of oxygen consumption. Thus, the primary control of oxygen levels in the inner Furnace basin was renewal of (and delivery of fresh oxygen to) resident water masses, which was mainly dependent upon lateral fluxes of external water of similar density. The exchange processes observed in Furnace shared many similarities with those of coastal systems with restricted exchange (e.g., Tett et al. 2003) and have good analogies on much larger scales with fjords (e.g., Farmer and Freeland 1983) and semi-enclosed seas that are poorly ventilated (e.g., Murray et al. 1991; Schinke and Matthäus 1998; Holtermann et al. 2017).

**Renewal of surface and intermediate waters**

The primary flux pathways with mean renewal time estimates for the layered water masses in the inner Furnace basin are recapped schematically in Fig. 10. The thin surface freshwater layer $L_1$ had the shortest renewal time and was generally replaced entirely within 10 d (Table 2.). This renewal time depends upon the rate of river runoff, with shortest renewal times occurring following periods of heavy rainfall. However, as increased runoff also increased the depth of the surface layer, the renewal time decreased slowly with increasing freshwater input. Given the fast transit time of freshwater through the system, Furnace likely has a limited capacity to sequester quantities of dissolved organic carbon exported.
from the Burrishoole watershed, which drains a blanket peatland catchment (Doyle et al. 2019).

Renewal of $L_2$, the subhalocline saline water (but located above the isolated deep basin water) was dependent upon positive fluxes of saltwater. The topographically constricted connecting channel was a key determinant of the overall small positive volume fluxes (Table 2) functioning as a long, shallow sill between the basin interior and the adjacent open coastal waters, limiting tidal inflows. Spring-neap forcing modulated the pattern of exchange fluxes between the inner basin and the adjacent open coastal water, a general dynamic that has been previously documented in other semi-enclosed estuarine and coastal systems (e.g., Geyer and Cannon 1982; Griffin and LeBlond 1990). Large freshwater fluxes, opposing the general direction of landward tidal inflows, also appeared capable of limiting the magnitude of tidal intrusions, even during spring tides (e.g., September 2016, February/March 2017; Fig. 5).

The overall significance of this spring-neap volume flux asymmetry through the lower layer was that major positive fluxes to the inner basin and hence renewal of $L_2$ effectively had a fortnightly frequency. This led to a relatively long mean renewal time of 46 d despite a small volume of water. A practical timeframe for the complete exchange of this water is over approximately three spring-neap tidal cycles although small oxygenized inflows were a persistent feature. Recent observations in the Baltic sea showed that small-scale lateral intrusions of saline water such as those observed here maintain some degree of oxygenation of this intermediate layer (Holtermann et al. 2019), with the resulting hypoxic transition zone playing a key role in basin nutrient recycling. Long residence times (compared to the rapidly flushed surface layer) and the spring-neap modulation of influx may also have biological implications in systems with spring-neap modulated volume fluxes such as Furnace. The subsurface chlorophyll maxima was found to occupy this intermediate depth zone below the upper halocline (de Eyto et al. 2019) and pelagic productivity in this stable region could potentially be influenced by fortnightly variability in tidal fluxes.

Renewal of deep water

Prior to the deep-water renewal and ventilation event on the 9th May 2017, the Furnace basin water had been stagnant since September 2014 (Fig. 9b). The set of observations taken during the 7-d event allowed some description of the dynamics involved. Aside from the irregular and ephemeral nature of these events, exchange of basin water by intruding dense gravity currents forms a fundamental link in the vertical circulation of oceanic, estuarine and lake basins of all sizes. Yet detailed observations are relatively rare, with fjords and semi-enclosed basins traditionally offering the most complete recounts of the hydrodynamic processes occurring during basin water exchange (e.g., Edwards and Edelsten 1977; Liungman et al. 2001; Arneborg et al. 2004; Fer et al. 2004; Holtermann et al. 2017).

The changes in the Furnace basin water is perhaps best observed from the progression of the T-S curves in Fig. 7. A straight line traced directly from the old basin water (black curve) to the densest water presented at the entranceway during the renewal (red star) reveals that the new densest basin water (additions at the ends of the blue and green T-S curves on 15th and 19th May, which lie halfway along the straight line) was a mixture comprising ~50% old basin water and ~50% new basin water. This estimate was validated by considering that the mean oxygen concentration of the inflowing new water was 8.5 mg L$^{-1}$ (Fig. 6d), the old basin water was 0 mg L$^{-1}$ and the new basin water immediately following the ventilation was 4.5 mg L$^{-1}$ (Fig. 8). Based on oxygen content, the new basin water should therefore contain ~48% old anoxic basin water and ~52% from the oxygenated inflow.

The creation of new deep basin water with intermediate properties could have occurred due to the turbulent entrainment of ambient resident water by the dense inflow as it descended (Schmale et al. 2016). If the oxygen, salt and heat content of the Furnace inflow were reduced along its passage to the basin interior through entrainment of old water (Figs. 7, 8), then its overall volume transport should also have increased. This can be confirmed by comparing the volume fluxes estimated directly from the current profiler during the renewal, to the volume of the new deep basin water. By integrating the horizontal area of the inner basin from the upper depth limit of the new deep water (~6 m; Fig. 8, blue profile) down to the deepest point, we obtain the approximate volume of the new deep basin water, $V_{new} = 2.61 \times 10^6$ m$^3$. The mean volume influx of dense new water at the entranceway over the 7-d renewal yielded a volume, $V_{in} = 1.22 \times 10^6$ m$^3$. This was only ~47% of the new basin water volume, meaning that the remaining ~53% was entrained from the old resident water and the initial inflow volume flux doubled. This value for volume increase of the inflow through entrainment fits well with the estimates that were obtained from analysis of temperature-salinity (50%) and oxygen (48%) properties. A twofold increase in the volume of a dense inflowing plume as it enters ambient water of lower density is consistent with observational values reported for deep-water inflows into fjords (Liungman et al. 2001) and for outflows from straits into receiving oceanic waters (e.g., Baringer and Price 1997; Girton and Sanford 2003).

Significance of deep-water ventilation

Profound changes in phosphorus, nitrogen, and methane dynamics have been reported following ventilation of a deep anoxic lake (Lehmann et al. 2015) and large changes to the bacterial community of an anoxic deep basin of the Baltic were detected following oxygenation (Bergen et al. 2018). The fate of basin water H$_2$S and methane during and following ventilation also deserves further investigation and are relevant to larger-scale analogues (Capet et al. 2016).

The fate of the remaining ~50% of the old basin water not entrained into the gravity current can be seen in Fig. 6b,c,
where an uplifting of cold, anoxic basin water to shallower depths occurred in the aftermath of the renewal. A new oxygen minimum zone occupies the depth range between ~3.3 and ~5.5 m, overlying the new (relatively) well-oxygenated basin water (Fig. 8). Some mixing with shallower oxygenated water evidently occurred given the increase in oxygen concentrations above anoxic levels. However, the average oxygen content of this new intermediate low oxygen zone was only 0.7 mg L⁻¹; average oxygen values at this depth range prior to the ventilation (Fig. 8, black line) were ~7.4 mg L⁻¹. This implied that the majority (>90%) of the water constituting this oxygen minimum zone is from the remaining anoxic basin water not entrained into the dense gravity current. It also indicates that only minimal amounts of the old basin water were mixed into the shallower waters and exported. This “trapping” of old basin water below the surface layer suggests that diffusive efflux of gaseous elements accumulated in anoxic water to the atmosphere may be avoided, although this dynamic certainly warrants further investigation given that other stratified coastal systems with deep-anoxia may also maintain vertical density gradients following a deep-water inflow.

The shallower oxic-anoxic interface observed in the months following ventilation has potential consequences for resident fauna inhabiting surface oxygenated zones. In particular, the shallower depth where the oxycline now intersects the basin slope around the perimeter of the inner basin will essentially translate to a reduction in the extent of viable habitat for aerobic organisms. In Furnace (Kelly et al. 2018b) and similar deoxygenated estuaries (e.g., Sanford et al. 1990; Chikita 2000), baroclinic waves can dramatically influence dissolved oxygen conditions in nearshore areas. A shallower oxycline following ventilation will therefore mean that upwelling anoxic water could encroach further on communities of nearshore benthic organisms.

Following the initial ventilation, conditions in the deep basin water once again became stagnant and oxygen declined due to internal oxygen consumption related to BOD. BOD per unit volume of basin water is clearly of interest and the postventilation stagnant period affords us the opportunity to estimate this parameter. Using the new basin water oxygen content as a starting point, the rate of oxygen decline over time, \( dO/dt \), was estimated. However, some of the change in oxygen content may occur through diffusive fluxes of oxygen between the deep basin water and the overlying water mass. Therefore, using the mean estimate of vertical diffusive flux between the basin water and overlying water (Eq. 6), \( Q_{K1} = 0.023 \text{ m}^3\text{s}^{-1} \), the rate of internal oxygen utilization (BOD) in the new basin water is given by:

\[
\text{BOD} = \frac{Q_{K1}}{V_3} (O_2 - O_3) - \frac{dO}{dt}
\]

where \( O_2 \) and \( O_3 \) are the mean oxygen concentrations of the basin water and overlying water mass. BOD in the basin following ventilation was 0.06 mg L⁻¹ d⁻¹, with the diffusive flux accounting for less than 5% of the rate of change in oxygen content. The high oxygen consumption rate of the basin water following ventilation (e.g., compared to consumption rates in the intermediate layer of the Baltic sea; Holtermann et al. 2019) is likely due to the sedimentary oxygen uptake being much larger in the bottom basin water. In Furnace, the mean ratio of sediment area to water volume (Livingstone and Imboden 1996) at depths within the typical intermediate layer is 0.01 m⁻¹, while in the deep basin water layer (below 6.6 m; Fig. 8) this ratio increases to 0.05 m⁻¹. In the bottom 2 m, the average ratio of sediment area to water volume is 0.18 m⁻¹.

**Drivers of deep-water ventilations**

Given the overall significance of deep-water renewal and ventilation, discerning the mechanisms involved and being able to predict their occurrence has considerable pertinency to the study of physical, chemical, and biological regimes in semi-enclosed deoxygenated coastal systems. The inflow density and density gradient of the receiving water column determine the intrusion depth of inflows into the inner basin, which interleave at a depth of neutral buoyancy. Based on our profiles of the water column density in the inner basin and entranceway, this intrusion depth was typically somewhere below the upper halocline and above the depth of the densest resident water in Furnace (Fig. 2). Less than 5% of observed inflow densities were sufficiently dense (\( \geq 1016 \text{kg m}^{-3} \)) to activate a deep-water renewal, that is, two red histogram bars with density greater than solid black line in bottom panel Fig. 2. While the precise drivers of deep-water renewal and ventilation vary from system to system, a number of general processes have been shown to play a role in activating a basin water exchange and we now examine which ones may be most relevant to Furnace.

**Basin water density**

Vertical diffusion and mixing will gradually reduce the density of basin water over time, preconditioning it for a renewal by reducing the density contrast with external water (Stigebrandt 2012). From 2016 to 2017 prior to the renewal event, the Furnace basin water density reduced at a rate of 1.5 \( \times 10^{-3} \text{kg m}^{-3} \text{ d}^{-1} \) (or 0.5 kg m⁻³ yr⁻¹). Given that renewals are typically separated by at least 1 yr (Fig. 9), the rate at which the resident basin water density reduces appears not be a limiting factor in the occurrence of dense intrusions. Sources of energy for this mixing are typically derived from internal tides radiating away from the entrance sill and breaking along sloping topography (e.g., Inall 2009). In Furnace, wind may be a significant factor as previous estimates of turbulent diffusivity in the stagnant basin water indicated a mean value of 9.35 \( \times 10^{-6} \text{ m}^2\text{s}^{-1} \), which increased by an order of magnitude following wind-induced internal seiching (Kelly et al. 2018a). Another important consideration is the extent to which the properties of the dense inflowing plume...
are modified through mixing, as the rate of entrainment and density reduction will set the upper limit of the basin water density at the outset of subsequent stagnant periods (Liu et al. 2001). Our analyses showed that the density of new basin water is significantly reduced compared to the inflow source water (Fig. 7) and thus we rule out the density of the inflow as a factor that restricts future ventilations by setting the new density too high.

**Tidally modulated exchange**

The timescales of basin water exchange have often been linked to the spring-neap tidal cycle, primarily through two different mechanisms. First, larger tidal excursions during spring tides may enhance the likelihood of deep-water renewal as large volumes of dense coastal water are transported toward the basin entrance (e.g., Edwards and Edelsten 1977). Second, changes in mixing regime between springs and neaps may modify the density of tidal water as it is transported through the entrance constriction. Often, spring tides may intensify this mixing, reducing the inflow density and lessening the likelihood of deep-water inflows (e.g., Geyer and Cannone 1982; Griffin and LeBlond 1990). In such systems, basin water exchange is thus more likely to occur during neaps. In Furnace, the former mechanism is much more fitting given that significantly higher salinities and larger advective fluxes into the inner basin are recorded during spring tides (Figs. 3, 5), related to the long excursion length between the inner basin and the open coastal water.

**Freshwater flux**

Freshwater runoff may indirectly control basin water exchange dynamics by physically blocking tidal inflows and diluting inflowing coastal water especially along constricted or shallow connecting channels. Complete blockage would require the surface freshwater layer to extend down to or below sill depth and in Furnace this would likely only occur during periods of exceptional run-off, given a mean surface freshwater depth of only 1.8 m and a depth of ~ 5 m through the deeper portion of the entrance section (Fig. 1c). The role of freshwater flux on controlling basin water exchange through mixing with inflows has previously been documented in the fjords and sea-loats of Scotland, with several systems showing an inverse correlation between river runoff and inflow density (Edwards and Edelsten 1977; Gillibrand et al. 1995; Allen and Simpson 1998). Reduced freshwater run-off has been shown to play a considerable role in the large-scale ventilation of the deep Baltic sea (so-called “major Baltic inflows”), with the drastically reduced frequency of these events in recent decades being linked to changes in large-scale atmospheric circulation over the North Atlantic and Europe (Schinke and Matthäus 1998; Mohrholz et al. 2015).

Water entering an estuarine basin during a flood tide is comprised of “old” water that exited the basin during the preceding ebbs and “new” water that is transported in from the adjacent coastal region. We can assess the role that freshwater flux through the Furnace system plays in modifying inflow densities by quantifying the fraction of inflow water comprised of actual “new” coastal source water in relation to the fraction of recirculated water. In Furnace, inflowing coastal water remains relatively unchanged through the lower part of the tidal channel (Fig. 1b). A timeseries of salinities recorded in 2010 at a mooring next to TG02 revealed consistent values of ~ 31 during flood tides (Kelly et al. 2018a). As this water traverses the remainder of the connecting channel and passes Nixon’s

![Figure 11](image_url)

**Fig. 11.** (a) Monthly mean freshwater discharge (thick black line) into upper Lough Furnace from the Burrishoole catchment, relative to the long-term monthly mean discharge (dashed black line) over the period 1976–2018. (b) Monthly mean along-shore component of wind stress in Clew Bay with positive values indicating winds blowing toward 107°T (see Fig. 1a), relative to the long-term monthly mean (dashed black line) over the period 2005–2018. Black vertical arrows indicate the onset of deep-water ventilations in the inner Furnace basin.
Island, it mixes with outflowing Furnace surface water, evident in the reduced salinities of tidal water measured at MR02 and MR01 (Fig. 3b,c). Following Pawlowicz et al. (2007b) and using salinity as a tracer, the lower layer inflow water at MR01 ($S_2$) is comprised of the following mixture:

$$S_2 = \gamma S_1 + (1 - \gamma) S_0$$

where $S_1$ is the surface outflow water from Furnace measured at MR01, $S_0$ is the coastal water entering the lower estuary (taken to be 31), and $\gamma$ is the recirculating fraction. Solving for $\gamma$ gave a mean value of 0.52 ($\pm 0.08$). This indicates that on average, 52% of the water flowing into the inner Furnace basin through the lower layer is comprised of prior resident water that was flushed out (or alternatively, 48% is comprised of “new” oceanic water).

In practical terms, this means that the salinity of the inflowing tidal water should be diluted by the outflowing surface water by roughly 52% by the time it reaches the inner basin. As salinity of the outflowing surface water increases during periods of low river runoff (Fig. 3c, blue line), the salinity of the inflow will be diluted less. Given a mean surface outflow salinity of 13.5 (observed on 9th May 2017) and taking a value of 31 for incoming water at the coastal boundary of the estuary, the inflow salinity to the inner basin was 22–23 at the onset of the renewal and would go on to reach values as high as 26 at peak flood tide (Fig. 3c, red line). It was also probable that the salinities of inflows from the adjacent coastal waters were slightly higher than 31 at this time, given the very low inputs of freshwater around Clew Bay. In contrast, a surface layer outflow close to fully fresh, which occurs following heavy rainfall events (Fig. 3c, blue line), would give a bottom inflow salinity of only ~15.

A climatology of freshwater discharge from the Burrishoole watershed with the timestamp of observed deep-water ventilations over the 10 yr period 2009–2018 reveals that significant ventilations occurred only during months with negative discharge anomalies (Fig. 11a). This further highlights that in Lough Furnace, freshwater flux plays the preeminent role in determining the salinities (densities) of inflows, thereby modulating the exchange of basin water to a larger extent than any other variable. It also implies that the frequency and magnitude of ventilations will be highly sensitive to climate-induced modification of precipitation and run-off patterns.

**Density variations in coastal water**

Variations in coastal water density have long been recognized to play an important role in regulating exchange processes in semi-enclosed coastal basins (Klinck et al. 1981; Arneborg 2004; Stigebrandt 2012). Regional winds can cause variations in water density...
density at coastal locations, with wind patterns that promote offshore transport of low salinity surface water often inducing upwelling of denser water from greater depths in return (e.g., Arneborg et al. 2004; Thomson et al. 2017). In Clew Bay (Fig. 1a), the dominant mode of wind variability is related to south-westerly sea breezes coming in from the North Atlantic, which would tend to block outflow of low-salinity surface water (i.e., positive toward 107°C; Fig. 11b). A climatology of wind stress along the primary axis of wind forcing in the bay revealed that deep-water ventilations in Furnace occurred during or immediately following months with negative wind stress anomalies. The 2017 and 2018 events occurred during weaker than average southwesterly winds and the 2010, 2013, and 2014 events during or following persistent northeasterly winds (Fig. 11b). The 2013 event in particular is interesting, as the magnitude of the ventilation was particularly large (Fig. 9b) despite the freshwater conditions not being as low as other months with ventilations (Fig. 11a). Given the timing of the ventilation in 2013 (9th April), the vernal equinox tides may also have played a role in the magnitude of this renewal.

Long-term and future oxygen dynamics

An understanding of the main drivers of renewal times and deep-water ventilation in the semi-enclosed estuarine basin Lough Furnace has allowed predictions of when and how significant alterations to basin water oxygen dynamics will occur. Ventilations essentially depend on two primary parameters: (1) a prolonged reduction in local rainfall leading to very low river runoff into and subsequently out of the inner basin and (2) spring tides capable of transporting dense coastal water across the long connecting channel. Wind-induced upwelling of denser coastal water may also play a role and at the very least, it would appear that a weakening of the predominant southwesterly sea breezes may reduce blockage and backing up of lower salinity surface outflows from the adjacent coastal bay. Sufficient reduction in basin water density between renewals and entrainment of the dense plumes as they intrude into receiving basin waters appear to prevent the basin water density from prolonging the internal timescale between ventilations.

As our predictive model required only catchment freshwater discharge, atmospheric pressure, and wind data as drivers, it was hindcast back to 1979 (the current time extent of the ERA5 data set) (Fig. 12). The hindcast revealed inconsistent periodicity in the regularity of ventilations, implying a stochastic nature to deep-water renewals in Furnace. While no observational data pre-2009 can verify whether the predicted ventilations occurred, it should be noted that the model successfully recreated the only other monitored ventilation event which occurred in summer 1995 (Anon 1995).
The deep-water oxygen dynamics appeared to undergo periods where ventilations were semi-regular (~1–2 yr apart) as well as periods of more prolonged stagnation (e.g., 1996–2000) (Fig. 12). To assess whether the frequency of ventilations has changed over time, we performed wavelet analysis (Torrence and Compo 1998) of the ventilation timeseries at a monthly timestep (i.e., whether or not a ventilation was predicted for each month). Results revealed distinct time periods with frequency variability primarily in the annual and even subannual bands, with some indication that energy in these bands changed at lower frequency timescales (e.g., mid-1980s, early 1990s, and mid-2000s; Fig. 13).

A longer timeseries would reveal the deep-water renewal climatology in Furnace to a greater extent and would allow for an assessment of whether event frequency is changing over time. The ERA5 data set is intended to eventually extend back to 1950, and if combined with continued monitoring of future deep-water oxygen in Furnace to validate and further refine model predictive capabilities, could generate a modeled timeseries capable of assessing the role of changing climate in deep-water oxygen dynamics. For example, ventilation frequency appears to be lower and more irregular from 2010 onward compared to early-2000s and early 1990s. However, it remains to be seen whether frequency will increase in the coming decade as part of a lower frequency (e.g., decadal) variability or if a trend of decreased ventilation frequency is ultimately related to climatic change.

Compared to tidal and wind forcing, changes in precipitation patterns are most pertinent to deep-water oxygen dynamics in Furnace and likely to be altered by climate change in the near future. Ascertaining whether climate change will alter the frequency of deep-water ventilation events in Furnace and other specific deoxygenated estuarine systems strongly influenced by local runoff conditions is difficult, as model projections of hydrological patterns throughout Europe are typically focused at the regional scale or coarser (Lobanova et al. 2018). The most recent decade (2006–2015) of a 305-yr long rainfall analysis for Ireland was found to be the wettest on record (Murphy et al. 2018) and it would appear based on the climatology of Burrishoole discharge that the last 10 yr have experienced wetter winters on average relative to the long-term climatology of Burrishoole discharge that the last 10 yr have (Murphy et al. 2018) and it would appear based on the long rainfall analysis for Ireland was found to be the wettest on term. Predictive tools for understanding past and future oxygen dynamics in systems that experience deep-water ventilation, especially if coupled to a rainfall-runoff model that can capture individual catchment-scale response to changes in timing and intensity of future precipitation. Furthermore, it could be used as a shorter-term forecast tool for deep-water inflows using readily available meteorological forecasts in the 3–5 d range. Such a tool could allow targeted intensive monitoring of inflow dynamics and biogeochemical changes and could be useful in coastal systems used for aquaculture, where fish cages could be susceptible to shalloing deoxegenated water.

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Conflict of Interest
None declared.