Advective controls on primary production in the stratified western Irish Sea: An eddy-resolving model study

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The Proudman Oceanographic Laboratory Coastal Ocean Modelling System and the European Regional Seas Ecosystem Model are applied at eddy-resolving (~1.8 km) scales to the stratified region of the western Irish Sea to investigate the effects of advective transport processes on the ecosystem. We find currents can transport nutrient-rich water into the otherwise nutrient-depleted surface layer of the stratified region, fueling intermittent production throughout the summer. The currents involved fall into three classes: large-scale wind and density-driven circulation, smaller-scale eddies, and tidally mediated dispersive phenomena; all appear to play a role in this area. A model experiment without ecosystem advection does not show the intermittent surface production; summer growth only occurs at the thermocline. This experiment gives a significantly lower total annual production of 110 ± 26 g C m⁻² yr⁻¹, compared with 150 ± 40 g C m⁻² yr⁻¹ for the full model, which is in better agreement with observational estimates of 140 g C m⁻² yr⁻¹. We calculate summer averages of the terms in the scalar transport equation, which show that advective transport of all nutrients dominates over vertical diffusion above the thermocline in most of the stratified region. The transport of nitrate, ammonia, and phosphate is significantly greater than the transport of silicate. This can be attributed to the lack of silicate recycling in the pelagic ecosystem. Only limited and anecdotal observational evidence exists to support these model results, which points to a need for observations of high spatial and temporal resolution to investigate these processes in conjunction with further model studies.


1. Introduction

Our understanding of the lowest trophic levels of shelf seas ecosystems is crucial to our ability to predict and manage change in these regions. However, these ecosystems exist in a complex physical environment that is mobile in the horizontal and vertical on a wide range of space and timescales; the interaction between the mesoscale physical environment and the ecosystem is particularly apparent from satellite observations of sea surface color (see, e.g., the images in the North and Baltic Seas presented by Holligan et al. [1989] and Lavender and Groom [2001]). The aim of this modeling study is to investigate ways in which currents, on a range of scales, determine the large-scale features of the lowest trophic levels. This understanding can help direct the development of coastal-ocean management systems, e.g., to monitor eutrophication, by establishing the required level of model complexity and sophistication in a particular region. Specifically, this work provides a prototype for the real time modeling systems planned for the Proudman Oceanographic Laboratory coastal observatory in the eastern Irish Sea (http://cobs.pol.ac.uk) and for the MERSEA project under the European Union Global Monitoring for Environment and Security (GMES) program (http://www.nersc.no/~mersea/) in the southern North Sea.
The Irish Sea (Figure 1) provides an excellent location to investigate complex physics-ecosystem interaction, since it offers a wide range of physical and biogeochemical regimes in a comparatively small area, many of which have been extensively studied in the past. The Irish Sea is the shelf sea region between Great Britain and Ireland and is connected to the rest of the northwest European continental shelf by two channels, the North Channel and St. George’s Channel. The hydrodynamics of the whole shelf are dominated by tides, and these propagate into the Irish Sea through both channels leading to comparatively weak tides in the western Irish Sea. These tides, together with increased depth, result in thermal stratification during the summer in the region. This stratification leads to a mid water density-driven circulation [Hill et al., 1994] around a comparatively isolated deep water cool-pool (Figure 1 shows a typical frontal location). The modeling work of Holt and Proctor [2003] shows this pool entrains partially mixed water, particularly from the north, with a consequent effect on the temperature structure. Here we investigate how advective transport impacts the ecosystem of this stratified region of the western Irish Sea through high resolution (1.8 km) coupled physics-ecosystem model experiments using the Proudman Oceanographic Laboratory Coastal Ocean Modelling System (POLCOMS) and the European Regional Sea Ecosystem Model (ERSEM) in the western Irish Sea (Figure 1), a combination of models previously used by Allen et al. [2001].

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The phytoplankton ecosystem of the western Irish Sea has been observed in a number of surveys and can be characterized by four regions: the coastal zone, the northern and southern mixed zones, and the summer stratified (gyre) region [Gowen et al., 1995]. These surveys show annual production is highest in the coastal zone (194 g C m\(^{-2}\) yr\(^{-1}\)) with an effective growth season of around 6 months. The summer stratified region has a shorter season of around 4 months and an annual production of 140 g C m\(^{-2}\) yr\(^{-1}\). The mixed waters have the shortest production season (around 2 months) and lowest annual production (<100 g C m\(^{-2}\) yr\(^{-1}\)), [Gowen and Bloomfield, 1996]. However, high interannual variability and spatial heterogeneity are features of the region [Gowen and Bloomfield, 1996].

The physics of currents and mixing are critical controls on the ecosystems of shelf seas and can explain much of their temporal and spatial variability. These controls have been the subject of extensive investigation by models [e.g., Fennel, 2001; Franks and Chen, 1996, 2001; Sharples and Tett, 1994] and observational studies [e.g., Jones and Gowen, 1990; Sharples et al., 2001a, 2001b], and many of the governing principles are now well established. Rapid production of phytoplankton occurs when there are sufficient nutrients and light levels. In well mixed waters this growth is controlled by the euphotic depth [Joint and Pomroy, 1993]. In deeper, less energetically mixed waters, blooms occur at the onset of seasonal stratification, typically in spring, when mixing at the thermocline is sufficiently reduced to retain phytoplankton in the euphotic zone. Blooms leave the surface water depleted of nutrients, and the phytoplankton growth for the rest of the year depends critically on the mechanism by which nutrients can be resupplied to the euphotic zone. This resupply of nutrients is controlled both by in situ recycling and by the primary transport mechanisms in shelf seas. The thermal (density) structure provides two of the dominant barriers against the transport mechanisms, by suppressing mixing in the vertical and by geostrophy in the horizontal (the former by provid-
ing a potential energy barrier and the latter by turning current toward a direction normal to the internal pressure gradients). Both these barriers tend to allow intermittent fluxes of nutrients, biota, and suspended material as conditions vary, introducing a significant source of variability into the ecosystem.

In the vertical the cross-thermocline transport is controlled primarily by boundary layer (wind and tidal) mixing processes; but internal mixing processes (e.g., breaking internal tides) can also play a significant role [Sharplees et al., 2001a]. Turbulence generated by surface wind stress tends to entrain water into the upper mixed layer during strong wind events. Tidal mixing tends to entrain water into the benthic boundary layer during spring tides, raising the thermocline, which drops back during neap tides. This results in the spring neat tidal modulation of midwater production that has been observed, among others, Sharples et al. [2001b] and reproduced in a number of ecosystem model studies [e.g., Allen et al., 2004; Sharplees, 1999]. The omnipresent nature of the tides tends, through tidal stirring, to delineate the mixing regimes in shelf seas [see, e.g., Simpson and Hunter, 1974], whereas wind stress is episodic and surface mixing is characterized both by seasonal variation in the strength of mixing and the occurrence of specific mixing events (e.g., storms). Tidal currents range from ~0.3 m s⁻¹ in the north and west region of the western Irish Sea and >1 m s⁻¹ in the south and east (for a detailed description of tidal currents in this region see Davies and Jones [1992]), whereas the wind-driven circulation is generally <0.5 m s⁻¹.

The transport processes at fronts have also been extensively investigated [e.g., Chen and Beardsley, 2002; Hill et al., 1993; Houghton and Ho, 2001; Matthews et al., 1993], and there is a well-documented tendency for these regions to exhibit enhanced surface production [e.g., Pingree et al., 1982]. In the context of the western Irish Sea tidal cross-frontal transport is likely to be mediated by a complex interaction of horizontal and vertical processes including tidal straining, shear diffusion, baroclinic eddies, secondary frontal circulation, and basin-scale currents (see Elliott et al. [1986], Hill et al. [1993], Holt et al. [2001], and Souza and Simpson [1997] for descriptions of these).

As representing three-dimensional (3-D) physics-ecosystem interaction, including all these processes, over seasonal timescales with a sophisticated coupled ecosystem-hydrodynamic model requires the use of massively parallel computers. However, even with today’s computer capabilities, spatial coverage is often limited by the resolution required to model the wide range of spatial scales significant in shelf sea dynamics. To address this issue in a way that is practical with modest high performance computer resources we opt for a hierarchy of nested models (Figure 1). Preliminary results from the model system described here have been reported by Proctor et al. [2002] and Holt et al. [2003]. Such a system can be used to explore the suitability of simpler models for a particular region, such as water column models [e.g., Sharplees and Tett, 1994] or 3-D averaged box models [e.g., Baretta et al., 1995; Lenhart et al., 1997; Pätsch and Radach, 1997]. These simpler models reduce the required computer resource by several orders of magnitude providing the opportunity for a much wider range of scientific investigation, so knowledge of their limitations is important.

This paper explores the extent to which phytoplankton growth in the main body of the stratified western Irish Sea is affected by advective transport of nutrients and biota, particularly from the frontal regions. We start with brief model descriptions and present results from the physics model. Sections 2 and 3 examine the phytoplankton distributions produced by this model and assess the role of advective transport. Conclusions are drawn in section 6, along with a discussion of how these results relate to the available observations.

2. Model Descriptions

2.1. POLCOMS Model

The hydrodynamic component of POLCOMS is the three-dimensional baroclinic B-grid model described by Proctor and James [1996] and Holt and James [2001]. It is a primitive equation finite difference model, solving for velocity (u, v, w), surface elevation, ζ, potential temperature, T, salinity, S, and turbulent kinetic energy, q² in spherical polar s coordinates [Song and Haidvogel, 1994] or σ coordinates for shelf sea applications. The model uses forward-time-centered-space differencing, with time splitting between external (fast) and internal (slow) modes. It uses a sophisticated advection scheme (the “Piecewise Parabolic Method” (PPM) [James, 1996]) to minimize numerical diffusion and ensure the preservation of features even on coarse grids under oscillatory flows. This scheme is directionally split so constituents are alternatively advected in the horizontal and vertical direction by the PPM. The scheme also ensures positivity and is not subject to a vertical CFL restriction on the time step, and a flux matching scheme gives conservation of volume over both the internal and external time steps. Horizontal pressure gradients are calculated by interpolation onto horizontal planes to avoid the errors associated with calculating these on σ coordinates where the topography is steep. Turbulent viscosities and diffusivities are calculated using a Mellor-Yamada level 2.5 turbulence closure but with an algebraically specified mixing length. In shelf sea applications, such as those considered here, the model is run without (imposed) horizontal diffusion (see Holt and Proctor [2003] for a further discussion of this issue).

The POLCOMS system includes the suspended particulate material (SPM) model described by Holt and James [1999b], which simulates the deposition and resuspension of inorganic fine suspended material. This implementation uses a single fine sediment class with a settling velocity of 10⁻³ m s⁻¹. The SPM model is arbitrarily initialized with a constant suspended load of 1.5 g m⁻³ and, in water depths less than 20 m, a bed load of 200 g m⁻². Since we lack any detailed information of sources of suspended material in any region, horizontal transport of SPM is not modeled.

2.2. ERSEM Model

ERSEM is a generic ecosystem model originally developed and applied in the context of the North Sea [e.g., Allen et al., 2001; Baretta et al., 1995; Pätsch and Radach, 1997; Radach and Lenhart, 1995]. It has also been successfully applied to the Mediterranean Sea [Allen et al.,...
ERSEM (Figure 2) takes the “functional group approach”; the biota in the ecosystem are divided into three functional types: primary producers, consumers, and decomposers, which are subdivided on the basis of trophic links and/or size. Both physiological (ingestion, respiration, excretion, and egestion) and population (growth and mortality) processes are described by fluxes of carbon and nutrients between functional groups. Each functional group is defined by a number of explicitly modeled components: carbon, nitrogen, and phosphorous and, in the case of diatoms, silicon.

Three zooplankton functional groups are described: mesozooplankton, microzooplankton, and heterotrophic nanoflagellates. Grazing uptake is a function of a maximal assimilation rate, temperature, food availability, and the zooplankton biomass. An overwintering function has been applied to the mesozooplankton model in order to compensate for the omission of life cycle processes. Meso-zooplankton are assumed to become sedentary when (depth

![Small Organisms ↔ Large Organisms](image)

**Figure 2.** A schematic of the pelagic trophic links in the European Regional Seas Ecosystem Model.
integrated) food supply falls below a given threshold. In this “overwintering state” biomass is only affected by a minimal mortality and respiration rate. The food threshold is set at 300.0 mg C m\textsuperscript{−2}; winter mortality and respiration rates are both 0.0025 day\textsuperscript{−1}. An increase in food availability over the threshold in spring provides the cue for the mesozooplankton to resume their normal physiological and ecological role.

[15] There is one pelagic decomposer functional group, bacteria. Bacterial uptake is a function of potential assimilation rate, temperature, oxygen availability, and the concentration and nutrient quality of the food source (dissolved organic matter). Bacteria are considered to mediate the breakdown of particulate organic matter to dissolved organic matter according to the nutritional content of the particulate fraction.

[16] The ERSEM model is coupled as a subroutine in the hydrodynamic model. The 36 pelagic state variables are advected (by the “Piecewise Parabolic Method”) and diffused at the same time step as the physics variables (400 s), and the ERSEM equations are integrated (explicitly forward in time) every third physics time step (i.e., 20 min).

### 2.3. Model Experiments

[17] For this work we focus on a region of the western Irish Sea. Our choice of resolution is based on our expectation that the dominant hydrodynamic scales in the Irish Sea are the tidal excursion (\(R_T\)), and the external (\(R_E\)) and internal (\(R_I\)) Rossby radii. For a 0.5 m s\textsuperscript{−1} M\textsubscript{2} tide the tidal excursion is \(\sim 4\) km. On the northwest European continental shelf the external Rossby radius has typical values of 200 km, so a 12 km resolution model is sufficient to represent the large-scale features over much of the continental shelf, although complex bathymetry and coastline make this resolution inadequate for tidal models of the Irish Sea. The internal Rossby radius has a typical value in the summer stratified water of the shelf of 4 km (based on a two-layer calculation). Eddy scales are typically 2.2\(\pi R_i \sim 25\) km [Griffiths and Linden, 1982], so a 2-km resolution should be adequate to represent the dominant physical processes in the western Irish Sea. To achieve this resolution for the coupled physics-ecosystem model, in a way that allows model development and different model experiments to be run, we use a series of one-way nested models with three domains (shown in Figure 1):

1. In the shelf-wide domain, physics and ERSEM models are run with a 1/6\(^\circ\) longitude \(\times\) 1/9\(^\circ\) latitude (\(\sim 12\) km) resolution and boundary conditions of 15 tidal harmonics and climatological \(T\) and \(S\). Twenty \(s\) coordinate levels are used.
2. In the Irish Sea domain the physics model is run with a 1/40\(^\circ\) longitude \(\times\) 1/60\(^\circ\) latitude (\(\sim 1.8\) km) resolution and boundary conditions from shelf wide domain. Thirty-four \(\sigma\) coordinate levels are used.
3. In the western Irish Sea domain, physics and ERSEM models are run with a 1/40\(^\circ\) longitude \(\times\) 1/60\(^\circ\) latitude (\(\sim 1.8\) km) resolution. Physics boundary conditions come from the Irish Sea domain and ecosystem boundary conditions come from the shelf-wide domain. Twenty \(\sigma\) coordinate levels are used.

[20] Only results from the western Irish Sea domain are presented here. Results from the shelf wide domain have been reported by Holt and James [2001], Holt et al. [2001], and Allen et al. [2001] and from the Irish Sea domain by Holt and Proctor [2003].

[21] For boundary conditions between each domain we use a flux/radiation scheme for surface elevation and an advective scheme for scalars (including ecosystem variables). The advective scheme uses ramped daily scalar values and hourly depth mean fluxes from the velocity boundary conditions to supply sufficient information to calculate the advection term at the boundary grid boxes of the model. This scheme has the distinct advantage over the relaxation scheme, used at the shelf wide model boundary, in that boundary condition \(T\) and \(S\) values are not used to calculate horizontal pressure gradients and so do not generate currents to which the boundary condition \(T\) and \(S\) cannot adjust. It is not practical to save daily values of all 36 pelagic state variables in the ERSEM model, so the boundary conditions for the (phyto- and zoo-) plankton variables only use the carbon variables and the interior C:P, C:S, and C:N ratios.

[22] All three model domains use the same surface forcing: meteorological variables from 6 hourly European Centre for Medium-Range Weather Forecasting (ECMWF) model data are used in bulk formulae to calculate surface fluxes of heat and momentum [see Holt and James [1999a] for details of these]. The Irish Sea domains also use rainfall data from Bidston Observatory to calculate a surface salinity flux.

[23] Away from the river plumes, the temperature structure dominates the density field in this region [Horsburgh et al., 2000], and so temperature is our primary guide to the physical regimes in the Irish Sea. However, salinity stratification is likely to play a significant role during the spring/early summer since the spring phytoplankton bloom is sensitive to small increases in the density gradient, which can occur from a combination of weak temperature and salinity stratification. Hence salinity is included as a prognostic variable in the model simulations but is not discussed further here.

[24] In addition to a simulation with the full physics and ecosystem dynamics (experiment 1) we also consider a numerical experiment in the western Irish Sea model domain with the advection (both horizontal and vertical) of ecosystem variables omitted (experiment 2). This is equivalent to running a one-dimensional ecosystem model at each model grid point. Since there is no feed back from biology to the physics in this model, both experiments 1 and 2 see an identical physical environment, but experiment 2 is artificially decoupled from one aspect of it (the currents). This provides insights into the three-dimensional coupled physics-ecosystem dynamics.

[25] The model simulates the annual cycle of 1995 (described by Proctor et al. [2002]), starting from rest with a temperature and salinity field derived from CTD and satellite observations, and constant ecosystem initial conditions (identical conditions are used in both the shelf and Irish Sea domains). These were derived from the “15 box” runs of ERSEM in the North Sea [Baretta et al., 1995] for the pelagic variables and from the ERSEM “nd130” model [Blackford, 1997] for the benthic variables. Since we are only simulating a single year, the results of the ecosystem model will inevitably depend on the initial conditions, e.g., on the total nutrient budgets. This also means that the final
model state will not match the initial state, as there will be a readjustment among the various model compartments (e.g., between pelagic and benthic models). Moreover, we are running nonlinear equations with nonperiodic forcing (e.g., variable residual currents and meteorological forcing), and so differences between the initial and final state are inevitable.

The model simulation produces an extensive five-dimensional data set in space, time, and model variables. The results presented here focus on the phytoplankton and nutrient variables and are limited to surface fields and time series of vertical profiles. Details of the spatial variation of the vertical structure will be presented elsewhere.

3. Model Temperature and Current Structure

To assist in the interpretation of the ecosystem model results it is necessary to briefly examine the physical variables of the nested western Irish Sea model. Figure 3 shows a time series of temperature in the center of the stratified region of the western Irish Sea (site A). The vertical structure is characterized by a period of summer stratification with a shallow surface mixed layer above an almost linear gradient of temperature (compare the North Sea and Celtic Seas which tend to have a better defined two-layer structure). Compared with CTD observations the stratification (here defined as the surface to bed temperature difference, \( \Delta T \)) agrees to within 1°C RMS and the depth mean temperature tends to be too cold (by \( \sim 0.7°C \)). This time series is in good agreement with the corresponding time series from the whole Irish Sea model, except for some details in the surface variability and for the tendency to underestimate the summer time warming of near bed temperatures. Also shown on Figure 3 is the daily mean of the ECMWF model wind stress at site A. This demonstrates that the onset of stratification occurs at the first period of prolonged weak winds in the year (from about day 94). Winds increase in the autumn from about day 265, but stratification persists until around day 300. Some, but not all, of the variability in the surface temperature is correlated with the wind stress (e.g., the period between day 230 and 250).

The currents that determine the transport of nutrients and biota generally fall into three classes: large-scale residual currents, being wind, density and tidally driven;
small-scale eddies, e.g., the product of baroclinic instability or tidal rectification at headlands; and tidally driven shear diffusion, whereby vertical variations in tidal current combined with vertical mixing lead to apparently enhanced horizontal diffusion. Maps of residual currents and surface to bed temperature difference during April and July (Figure 4) show examples of the first two of these; shear diffusion is less easily depicted. The spring period shown in Figure 4 is significantly more quiescent than the summer, having weaker large-scale currents and less variability.

[31] The dominant large-scale current structure in this region is a density-driven mid-water cyclonic circulation around the stratified region [Hill et al., 1997, 1994; Horsburgh et al., 2000]. This circulation is seen in the modeled surface currents (Figure 4) as a general tendency for southward flow on the west side of the stratified region and northward flow on the east, particularly in July. However, the surface currents do not generally follow contours of $\Delta T$, and the gyre circulation is not as well defined at the surface as it is at depth. Wind forcing, baroclinic instabilities, and the absence of the stabilizing effects of bottom topography all serve to make the surface currents more variable than those at mid depth and near the sea bed. Since there is no imposed horizontal diffusion in this model the variability in these currents is freely evolving but limited by tidal friction. There is some observational evidence that this variability is realistic: The drifter tracks presented by Horsburgh et al. [2000] show small-scale recirculations are common (despite these drifters being drogued at 30 m), and satellite images of surface temperature show similar scales of variability. Another common feature in the currents shown in Figure 4 is the southward current emanating from the northern boundary. This follows the bathymetry (Figure 1) initially southeastward before turning sharply to the southwest. This current originates from the southward flow on the west side of the North Channel [Holt and Proctor, 2003], which in turn originates from the northeastward flow on

**Figure 4.** Residual (25-hour mean) currents and surface to bed temperature difference ($\Delta T$) for two 12-day periods in (top) April 1995 and (bottom) July 1995. Currents have been averaged onto every second grid point.
Evidence for this current comes from a number of observational [e.g., Brown and Gmitrowicz, 1995; Knight and Howarth, 1999] and modeling [Xing and Davies, 2001] studies.

Apart from near headlands, such as off the southeast coast of the Isle of Man on day 108, coherent eddies are not particularly common, or energetic, in these model results although weak eddies are sometimes visible in the currents near the eastern and southern frontal regions. Most of the variability in the velocity field is in the strength and course (e.g., meanders) of the larger-scale currents.

The surface to bed temperature difference (Figure 4) indicates significant variation in the sharpness of fronts around the stratified region. The western front is the sharpest, the south and east are more diffuse, and the northern front is apparent only as a gradual northward reduction in stratification. This is to be expected as the larger area Irish Sea model (grid 2 in Figure 1) shows stratification extending into the North Channel during the hot summer of 1995, as also occurred in observations in 1992 presented by Gowen et al. [1995]. Figure 4 also shows that there is significant variation in the extent and shape of the stratified region during both the April and July periods.

### 4. Model Phytoplankton Distribution

Figure 5 shows surface chlorophyll (derived from a fixed carbon to chlorophyll ratio for each phytoplankton class) for the same 6 days in April and July 1995 as shown on Figure 4. It demonstrates the formation of the spring bloom, spreading out from a patch in the middle of the domain, to form the convoluted pattern seen on day 102, then dying away, leaving an annulus shape (day 108). This patch originates in the stratified region and the annulus indicates the frontal region around the stratified water. Within the stratified region nutrients have been depleted above the thermocline by phytoplankton growth. Density stratification inhibits vertical mixing and prevents the re-
supply of nutrients from below, and the frontal region tends to isolate the water mass horizontally, inhibiting lateral supply of nutrients. The patterns in these model results resemble those seen in SeaWifs satellite observations of surface color and have a close resemblance to the current and frontal structures seen in Figure 4. However, this not a static picture for the rest of the summer. The July series in Figure 5 shows complicated patterns of chlorophyll patches particularly near the north and east frontal regions but also intruding into the body of the stratified region. The variability over the 12-day period in July illustrates that these patches of summer time surface production are intermittent and of low amplitude compared with the spring bloom; one of our aims here is to assess their overall significance.

[35] Wave number spectra of surface chlorophyll can give some insight into the origin of the phytoplankton patchiness. For each day’s model output, spatial spectra are calculated along north-south grid lines between −6°E and −4.8°E (48 lines in total) over the whole extent of the model domain (i.e., including stratified, frontal, and well-mixed regions). The resulting spectra do not show any significant peaks in horizontal scale throughout the year but have a consistent spectral slope indicating a turbulent cascade. The amplitude relationship with wave number, k, is $A(k) \sim k^\beta$, with $\beta = -1.3 \pm 0.3$, for scales between 5 km and 50 km; the range indicates the variations over the year and across this east-west band. The value of the exponent $\beta$ lies between the theoretical values of $-5/3$ for isotropic turbulence and $-1$ for two-dimensional (enstrophy conserving) turbulence [see Monin and Obzhdlov, 1985] and is significantly different from the value of $-2.5$ typical for spectra dominated by nondiffusive advection and biological (e.g., predator-prey) interactions [Abraham, 1998]. This suggests these patches arise from advective and diffusive transport of ecosystem variables rather than from nonlinearities in the biological model.

[36] Time series of phytoplankton concentration at site A (Figure 6) show the canonical picture of a strong spring bloom in the diatoms occurring at the onset of stratification (Figure 3), followed by the periodic production in mid water that has been seen in a number of previous modeling [e.g., Sharples and Tett, 1994] and observational studies [Sharples et al., 2001b]. However, in addition to this, the other phytoplankton classes also show sporadic surface production before the autumn bloom. This occurs as structures, like those seen in Figure 5, grow and are advected past the grid point at site A. The flagellate and picoplankton time series both show a background surface value of between ~5 and ~17 mg C m$^{-3}$; superimposed on this are a number of peaks of ~30 mg C m$^{-3}$. In contrast surface diatom levels are less than 6 mg C m$^{-3}$ before the autumn bloom. The time series of surface nutrients at site A (thick lines in Figure 7) show silicate is limiting for diatoms at the surface, but nitrate, ammonia, and phosphate are all intermittently resupplied at the surface to support this patchy production.

5. Role of Advevtive and Diffusive Transport

[37] The current fields and frontal structure produced by this model show a high degree of horizontal and temporal variability (as demonstrated by Figure 4), and similar variability might be expected to appear in the distribution of the ecosystem variables, as is suggested by Figures 5, 6, and 7. To investigate possible causes of such variability in more detail, the evolution through time of four of the modeled surface fields is shown in Figure 8. During this 4-day period in July, 1995 a southward current from the west side of the North Channel (discussed in section 3) pushes into the frontal region. Compared with the stratified region, the water in the partially mixed region near the northern boundary is cooler with higher nitrate and chlorophyll concentrations; this situation can occur simultaneously because of light limitation due to high SPM concentration in well mixed/partially stratified regions. Over the 4 days shown in Figure 8, the current from the north entrains this water into the stratified area, causing a localized pulse of high phytoplankton concentration. This is seen in Figure 6 as a peak in the flagellate biomass accompanying a drop in the surface temperature at day 202. The similar peak in flagellate concentration and drop in temperature at day 239 is also accompanied by a southward current similar as such as that seen in Figure 8. This transport provides two mechanisms for such pulses: It transports biota from the frontal regions of high production, and it resupplies nutrients into otherwise depleted stratified regions. The tongue of chlorophyll is generally broader than either the nitrate or the residual current structure, both of which tend to be concentrated on the northern side of the chlorophyll tongue. This suggests that the nitrate at the edge of this current is becoming depleted by phytoplankton growth as it progresses into the stratified region.

[38] To examine the spatial distributions of nutrient transport over the whole model domain, we use a similar calculation to that applied to the temperature field by Holt and Proctor [2003]. For scalar concentration $C$, the transport equation is:

$$\frac{\partial C}{\partial t} = L(C) + D(C) \quad \text{with} \quad L(C) = -\frac{\partial u C}{\partial x} - \frac{\partial v C}{\partial y} - \frac{\partial w C}{\partial z}$$

and

$$D(C) = \frac{K_z}{\frac{\partial C}{\partial z}} \frac{\partial C}{\partial z}$$

where $K_z$ is the vertical turbulent diffusivity and $x, y$, and $z$ are the eastward, northward, and vertical coordinates, respectively (written here in Cartesian coordinates for simplicity). $L(C)$ and $D(C)$ are the advective and (vertical) diffusive flux divergences (i.e., the change in concentration with time caused by each process), and since this model does not include horizontal diffusion (at this resolution), these fully describe the coupling of the ecosystem model to the currents. The calculated values of $L(C)$ and $D(C)$ for the ERSER nutrient variables (total nitrogen, phosphate and silicate) are averaged over the summer (from day 182 to day 257, i.e., from July to before the onset of the autumn diatom bloom) and integrated from the surface to the thermocline at $z = z_d$ to give $\langle L \rangle$ and $\langle D \rangle$. The thermocline is taken to be at

$$z_d = \frac{z(T - T_d)}{(T - T_d)^2}$$

for $\Delta T \geq 0.5^\circ C$ and $z_d = 0$ for $\Delta T < 0.5^\circ C$,

$$\text{with an overbar indicates a depth mean. This provides a robust definition of the depth of the midpoint of the}$$
Figure 6. Time series of phytoplankton depth profiles at site A for three of the models phytoplankton classes (For clarity, biomass less than 1.09 mg C m\(^{-3}\) has been left white).
thermocline, which is needed since the temperature profiles in the western Irish Sea can be quite complicated. This calculation gives the seasonal mean change in layer integrated nutrient concentration above the thermocline resulting from advection, \( L \), and vertical diffusion \( D \). Figure 9 shows the spatial distribution of \( L \) and \( D \) for total nitrogen (nitrate+ammonia), phosphate, and silicate. These have been scaled with the initial condition concentration (and have units \( C_0 \) m d\(^{-1}\), where \( C_0 = 10 \) mmol m\(^{-3}\) for nitrate+ammonia, 0.6 mmol m\(^{-3}\) for phosphate, and 4 mmol m\(^{-3}\) for silicate) so their relative values can be compared. The total nitrogen and phosphate show very similar patterns and scaled values; most of the stratified regions show significant advective increases in nutrient concentrations over the summer apart from in the southwest. The values are largest toward the northern boundary and in the vicinity of the \( \Delta T = 2^\circ C \) contour. In the west the diffusive increase is restricted to the frontal regions between the \( \Delta T = 2^\circ C \) and 0.5\(^\circ C \) contours. Adective loss of these nutrients is generally limited to the southern frontal region where it is coincident with a region of diffusive increase in surface nutrient concentration. This region of diffusive increase extends into the stratified region, reflecting the stronger tides at the southern front. The distribution of silicate flux divergence shows significantly reduced (scaled) values, although the overall pattern is similar. This results from silicate only being recycled in the benthic model, and this is significantly slower than the recycling of ammonia and phosphate in the pelagic model. [39] By omitting the ecosystem advection term, \( L(C) \), from the model calculation (experiment 2) we can further distinguish between the effects of advective and diffusive resupply of nutrients. Because advection mediates the main horizontal diffusive process in this region (shear diffusion) we cannot distinguish between pure advection and apparent horizontal diffusion (arising from the shear diffusion and from the numerics). Moreover, this experiment cannot distinguish between the horizontal and vertical advection, so any upwelling effects are also removed in experiment 2. Figure 10 displays similar results to those for experiment 1 shown in Figure 6 but for the experiment without ecosystem advection (experiment 2). After the spring bloom, no surface production takes place before the autumn bloom; the only growth in all the phytoplankton classes during this time is at the thermocline. This behavior bears a similarity to results from this model system when it is applied in the North Sea (e.g., the north-south sections shown by Allen et al. [2001]). The northern North Sea stratified region is mostly isolated from the frontal regions, and advective processes are less influential.

Figure 7. Time series of surface nutrients at site A. Values for both the full model (thick lines) and the experiment with ecosystem advection omitted (light lines) are shown.
Figure 8. Detailed evolution of model surface temperature, residual current, nitrate, and chlorophyll for a 4-day period during July 1995. For clarity, only a small region of the northwest corner of the model domain is shown.
The reason for the lack of surface production in experiment 2 is apparent from Figure 9, which shows that there is no diffusive supply of nutrients to the surface layer in most of the stratified region during the summer. So in the absence of an advective supply this area remains nutrient-depleted throughout the summer, and hence there is no production. This is also apparent from the nitrate at site A (Figure 7), which shows no resupply of nutrients to the surface after the initial depletion following the spring bloom in experiment 2. Figure 7 also shows that surface ammonia levels are depleted more slowly than nitrate levels (owing to recycling) and that levels in the full model (experiment 1) remain above those in experiment 2, indicating there is either an advective transport of ammonia (i.e., as a component of that shown in Figure 9) or there is enhanced nutrient recycling in experiment 1. The phosphate levels in experiment 1 do drop to the experiment 2 levels between peaks, suggesting this nutrient might be limiting. The southward current seen on day 200 in Figure 8, and discussed above, is accompanied by a drop in stratification (Figure 4). However, this does not result in a significant vertical diffusive transport of nutrients since both experiments 1 and 2 see an identical mixing regime, and there is no surface production in experiment 2 at this time. In contrast, as stratification weakens and breaks down (after about day 273), there is a significant vertical flux of nutrients (Figure 7) and an autumn bloom in both experiments 1 and 2 (seen in Figures 6 and 10).

We noted above that in omitting advection from experiment 2 we also removed upwelling as a possible mechanism of nutrient resupply (similarly the calculation of $\langle L \rangle$ is unable to distinguish between vertical and horizontal

![Figure 9](#)

Figure 9. Above thermocline layer integrated advective and diffusive change in nutrient concentrations, averaged over the summer (from day 182 to day 257). Values have been scaled with the initial condition concentration ($C_0 = 10 \text{ mmol m}^{-3}$ for nitrate+ammonia, $0.6 \text{ mmol m}^{-3}$ for phosphate, and $4 \text{ mmol m}^{-3}$ for silicate). Positive values are shaded, and negative values are contoured (interval $0.2 C_0 \text{ m d}^{-1}$). The dashed lines show the $0.5^\circ$C and $2^\circ$C $\Delta T$ contours.
Figure 10. Time series of phytoplankton depth profiles as Figure 6 but for the experiment without ecosystem advection. (For clarity, biomass less than 1.09 mg C m$^{-3}$ has been left white.)
advection). The temperature structure in experiments 1 and 2 is identical and is affected by frontal upwelling circulation, leading to the characteristic dome shape of a cyclonic gyre. This sets the level of the nutricline, if it is defined by local processes alone (experiment 2). One possibility is that in experiment 1 the upwelling circulation might further raise the nutricline above this level and there is some evidence of this occurring in these model results, which deserves further investigation. However, the nutricline is only raised by a few meters, so for this to have an effect on production at the surface it must be accompanied by a significant vertical diffusive flux of nutrients (i.e., the nutricline must be raised, relative to the thermocline, to the extent that it enters the mixing region), and this is not seen in Figure 9 for most of the stratified region. While this analysis does not definitively rule out upwelling as a contributing processes here (since we have not investigated the vertical diffusive flux of biota), these results, and the anecdotal evidence of Figure 8, do suggest that horizontal rather that vertical advective nutrient transport is responsible for the patchy surface production seen here.

To assess the overall significance of advective transport we calculate total annual production of phytoplankton in the two experiments (Figure 11). The omission of advection causes a significant decrease in production in the stratified region and northern frontal waters. The other frontal regions and well-mixed waters are less affected as these are resupplied by a diffusive transport of nutrients during the summer (Figure 9).

6. Discussion

The model experiments indicate that the surface waters of the stratified western Irish Sea are intermittently connected to the frontal regions by horizontal advective, and presumably horizontal diffusive, processes, i.e., by currents on many scales, and that the ecosystem is affected by these processes. Hence the transport processes at the fronts have a significant effect on the main body of the stratified region. We show that sporadic advective fluxes of nutrients and biota into the surface waters of the stratified region result in intermittent production during the summer that is small, but because it occurs over a period of several months, accounts for a significant proportion of the total annual production in the stratified region. The mean annual production in the stratified region has been estimated to be 140 g C m⁻² yr⁻¹ [Department of Environment Transport Regions, 2000]. If we average the model production figures over this area (taken to be where mean July and August surface to bed temperature differences are greater than 2°C) the full model gives a value of 150 ± 40 g C m⁻² yr⁻¹, but the model without ecosystem advection gives only 110 ± 26 g C m⁻² yr⁻¹. This demonstrates that advective processes account for 26% of the total annual production in this model. We have not attempted to quantify the relative importance of horizontal and vertical advection since these are intimately linked by the continuity equation. The vertical term is dominated by the barotropic component, which has little importance here but would add a significant source of uncertainty to such a calculation. Moreover, separating these terms, during the directionally split advection steps (in a way that gives a balanced budget), is not straightforward technically.

The results presented in this paper, indicating the important role of horizontal advection on production in the surface layer, are from a model, so arise from a theoretical view of reality. The accuracy of this description of the processes occurring can only be assessed by a comprehensive and detailed comparison with contemporary observations. Observations sufficient to criticize this work are not available, so we must rely on indirect and anecdotal evidence to lend weight to the conclusions drawn here. The strongest piece of indirect evidence comes from modeling the temperature structure. A comparison with a large number of CTD and satellite observations shows that horizontal transport plays a significant role in determining this structure. Holt and Proctor [2003] show that when using a turbulence closure scheme that produces good temperature structure in the North Sea in a water column model, the time evolution of the temperature structure in the stratified region of the western Irish Sea can only be accurately simulated when the advection of partially mixed
water into the stratified region is included in the model. This partially mixed water carries with it enhanced surface nutrient concentrations.

[45] For direct observations we turn to surveys made in other years and compare the surface chlorophyll levels from these with the experiments with and without advection. Surface chlorophyll measurements at site A in the summer of 1997 have a value of 1.28 mg m\(^{-3}\) (mean of three values on days 155, 185, and 221) compared with the summer mean experiment 1 value of 0.48 mg m\(^{-3}\) and the experiment 2 value of 0.15 mg m\(^{-3}\). Similar surveys made in 1992 and 1993 [Gowen and Bloomfield, 1996] show a patchy horizontal phytoplankton distribution, while (coarse temporal resolution) time series of vertical profiles [Gowen et al., 1995] show the mid water chlorophyll maximum is broad and almost reaches the surface around day 230. Again the surface chlorophyll values during the summer do not fall below 1.0 mg m\(^{-3}\) and certainly not to winter values (observed to be \(\sim 0.5\) mg m\(^{-3}\)) as would have been predicted by experiment 2. Unfortunately, satellite observations of surface color in these case 2 waters do not provide sufficiently accurate measurements of chlorophyll (given current algorithms and the levels of SPM in the Irish Sea) to detect the small concentrations important for this work against background values.

[46] The survey presented by Horsburgh et al. [2000] included some scan fish (undulating CTD) sections making fluorometer measurements. These data are not currently available, but we plan in the future to make a direct comparison, on the whole Irish Sea domain, between this data and a variable chlorophyll version of the ERSEM model currently under development and needed to address the tendency for this model, with a constant chl:C ratio, to underestimate chlorophyll levels (as seen in the above comparison).

[47] We can identify three types of current that can result in significant cross-frontal transfer: the large-scale wind and density-driven circulation, small-scale quasi-horizontal eddies, and dispersion mediated by tidal currents (e.g., shear diffusion), but we have not attempted to quantify their relative importance. However, the results presented here suggest an important role for the gyre circulation and the southward current from the North Channel. The latter implies that an extension of the domain might be necessary for a quantitative assessment of the model variability; the present boundary conditions only provide sufficient information for a reduced flux calculation (e.g., there is no momentum advection at the open boundaries). If transport from the North Channel is important on the timescales considered here, then it should be included explicitly rather than as a boundary condition. The details of the frontal exchange processes need to be examined more closely against observations, particularly to determine whether or not the high level of surface variability is realistic or if this model is underrepresenting the level of horizontal friction (momentum diffusion).

[48] Despite not being able to back up these model results with high spatial and temporal resolution observations we can still make a substantive statement on the total annual production: That the model results including advection give significantly better agreement with previous estimates than those without. This in itself encourages further investigation of horizontal transport processes on all scales in this region and their effects on the ecosystem. Holt et al. [2003] show the performance of this modeling system is scalable to over a thousand processors, making it well poised to utilized the UK's new resource of a 1280 processor IBM POWER 4 system in an extension of the domain to include the whole northwest European Continental Shelf at eddy-resolving scales. Moreover, the results presented here can be used to direct simpler modeling studies; for example, Figure 9 can be used to select a suitable site for 1-D modeling in the western Irish Sea; the need to avoid fronts and regions of strong advection limits the choice to a narrow band on the west side of the stratified region (e.g., \(-5.75^\circ\)E \(56^\circ\)N might be a suitable location).

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