

ORIGINAL ARTICLE

The maintenance of the subsurface chlorophyll maximum in the stratified western Irish Sea

Charlotte Williams,^{1,2} Jonathan Sharples,^{1,2} Mattias Green,³ Claire Mahaffey,¹ and Tom Rippeth³**Abstract**

The diapycnal flux of nitrate from the deep water provides a limit on new production in the subsurface chlorophyll maximum (SCM) during summer in stratified shelf seas. Here we estimate the diapycnal nitrate flux into the SCM in the stratified western Irish Sea (SWIS). Sampling took place immediately before neap tides when winds were light, so flux estimates reported provide a lower limit to nitrate supply to the SCM. Measurements of turbulent kinetic energy dissipation, chlorophyll *a*, and nitrate were used to estimate the flux of nitrate and chlorophyll through the SCM. Turbulent dissipation was low in the SCM (10^{-9} to 10^{-7} $m^2 s^{-3}$), driving a correspondingly low nitrate flux into the SCM (0.31 $mmol m^{-2} d^{-1}$). The thermocline was marginally stable throughout sampling, and thus the addition of shear would likely result in shear instabilities and mixing. We show that although the SWIS is documented as having an energetic internal tide at this time, there was a low level of dissipation within the thermocline. We argue that the internal tide sets up background shear, which results in marginal stability. The addition of extra shear through the passage of nonlinear internal waves and/or the wind can trigger instability and mixing. We extrapolate our flux estimate over the summer and show that the nitrate flux is insufficient to sustain the documented summer production estimates for the SWIS. This suggests that episodic events are likely to be important for nitrate fluxes, or even largely responsible for the nitrate flux that sustains the SCM.

Keywords: subsurface chlorophyll maximum, nutrient flux, turbulent mixing, internal waves, Irish Sea

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Introduction

[1] Shelf seas are areas of high biological activity, covering less than 8% of the global ocean surface area but supporting 15%–30% of all oceanic primary production (Wollast 1998). In temperate shelf regions, the onset of spring stratification leads to the development of the spring phytoplankton bloom, which rapidly draws down the nutrients in the new surface layer and typically results in nitrate limitation, with relatively low

rates of phytoplankton growth sustained by regenerated forms of nitrogen (Pingree et al. 1976; Fasham et al. 1983). New primary production, utilizing nitrate supplied from subthermocline waters, occurs within the subsurface chlorophyll maximum (SCM) (Hickman et al. 2012). Carbon (C) fixation within the SCM plays an important role in feeding the pelagic and benthic ecosystems during summer (Richardson et al. 2000). These regions have been implicated as

substantial sinks for atmospheric carbon dioxide because of the undersaturation of carbon dioxide at the sea surface (Thomas et al. 2004), the magnitude of undersaturation being related to production in the SCM (Kitidis et al. 2012).

[2] The thermocline provides a physical barrier to the transfer of nutrients and phytoplankton between the surface (euphotic) layer, where light is abundant, and the dark but nutrient-rich deeper water (Sharples et al. 2001b). Weak turbulence at the base of the thermocline drives a supply of nutrients from the nutrient replete bottom water into the euphotic zone. Nitrate concentrations in the surface water typically remain undetectable ($<0.1 \text{ mmol m}^{-3}$), which indicates that (1) the phytoplankton population in the SCM are able to utilize the entire diapycnal nutrient flux, and (2) the rate of diapycnal nutrient supply into the thermocline provides a limit to the rate of new primary production in the SCM (Sharples et al. 2001b). The source of energy to drive the diapycnal nitrate flux can be local, such as episodic wind events (Sharples and Tett 1994; Rippeth 2005) or internal tides, generated locally at steep bathymetry (Garrett 2003). A significant flux of internal tidal energy can also arrive from the shelf slope (Inall et al. 2011), potentially generating large diapycnal nitrate fluxes (Sharples et al. 2007).

[3] The stratified western Irish Sea (SWIS) provides a good location for examination of diapycnal nitrate fluxes because the water column structure and fluxes are dominated by vertical exchange (e.g., Simpson and Rippeth 1998) (Fig. 1). The region is stratified during summer months and surrounded by a tidal mixing front separating it from the vertically mixed Irish Sea (Simpson and Hunter 1974; Hill et al. 1994). The stratification occurs because of the relatively weak tidal flows in the area being unable to dissipate the large buoyancy inputs by surface heating during spring and summer. The surrounding mixed water isolates the SWIS from the shelf edge, so that sources of energy for internal mixing must be local. In addition to energy inputs from episodic wind events (Rippeth et al. 2009), there is thought to be a quasi-continuous supply of mixing energy driven by the internal tide generated close to the Isle of Man (Fig. 1) and dissipating its energy over the entire SWIS gyre (Green et al. 2010). The internal tide provides energy fluxes of $\sim 100 \text{ W m}^{-1}$ (Green et al.

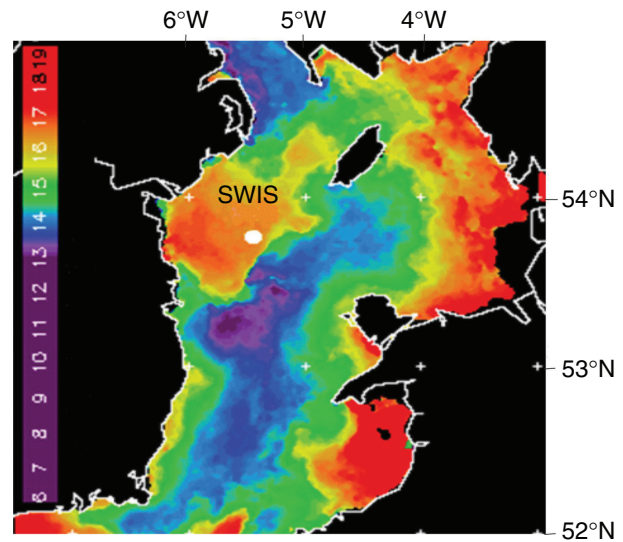


Fig. 1 Sea surface contours of temperature ($^{\circ}\text{C}$) in the Irish Sea taken from a composite of satellite data from 15 to 21 July 2005. The western Irish Sea gyre is clearly visible via sea surface temperature and the stratified western Irish Sea (SWIS) station is marked by a white dot at the edge of the SWIS gyre. Image courtesy of Plymouth Marine Laboratory. Isle of Man is northeast of the SWIS.

2010); this is quantitatively similar to the documented internal tide energy flux generated at the Celtic Sea shelf edge ($\sim 100\text{--}150 \text{ W m}^{-1}$; Green et al. 2008; Inall et al. 2011). Within the SWIS, horizontal gradients in water properties are weak, so that the advection of nitrate is small compared with vertical turbulent fluxes.

[4] In summer nitrate is generally depleted from surface waters by June and is the limiting nutrient in the SWIS (Slinn 1974; Gibson et al. 1997). Nitrate concentrations in the bottom mixed layer are typically less than 8 mmol m^{-3} in July, which are comparatively low compared with the Celtic Sea and areas closer to the shelf edge (Hydes et al. 2004). An SCM is typically observed to persist throughout the stratified period, with chlorophyll *a* (chl *a*) concentrations typically reaching 4 mg m^{-3} , which is approximately a quarter of the chl *a* concentration observed in the same region during a spring bloom (Gowen and Bloomfield 1996; Gowen and Stewart 2005). The region is important for the commercially exploited *Nephrops norvegicus* (White et al. 1988), with the postspring primary production within the SCM thought to be an important source of organic material for these benthic organisms (Trimmer et al. 1999).

[5] The diapycnal nitrate flux has previously been estimated to be 1.5 ± 0.3 (mean \pm SE) $\text{mmol m}^{-2} \text{d}^{-1}$ in the SWIS, which is capable of supporting total new production over the summer similar in magnitude to the total production generated during the spring bloom ($17 \text{ g m}^{-2} \text{C}$; Rippeth et al. 2009). This flux is comparable to estimates of nitrate fluxes calculated at the Celtic shelf edge during neap tides and in the western English Channel ($1\text{--}2 \text{ mmol m}^{-2} \text{d}^{-1}$; Sharples et al. 2001a, 2007). Gowen and Bloomfield (1996) estimate that total primary production in the western Irish Sea is $\sim 140 \text{ g m}^{-2} \text{C}$ between April to September, which is lower than summer integrated primary production observed in Irish Sea coastal waters ($\sim 194 \text{ g m}^{-2} \text{C}$; Gowen and Bloomfield 1996). In the SWIS the SCM is estimated to account for $\sim 53\%$ of the water column chl *a* standing stock (Gowen and Bloomfield 1996).

[6] The purpose of the following study is to provide a robust lower limit of nutrient fluxes into the SCM in an isolated, relatively high-energy shelf sea environment compared with nearby temperate shelf seas. We hypothesize that the documented energy flux for the internal tide in this region has a small contribution to the diapycnal nitrate flux, and thus on its own the internal tide is insufficient to sustain the documented summer primary production rates for the region. We, therefore, examined the lower limit to diapycnal nitrate fluxes in a region isolated from any significant influence of external sources of mixing energy and during a period of very low winds. This allowed us to assess the potential new primary production supported by that flux compared with other measurements made in the SCM and with the production associated with the spring bloom.

Methods

[7] Measurements were collected at the SWIS station located at $53^\circ 43.00' \text{ N}$, $5^\circ 30.00' \text{ W}$, during a research cruise in July 2006 on the *RV Prince Madog* (Fig. 1; for cruise details, see Simpson et al. 2009; Green et al. 2010). The water depth is $\sim 110 \text{ m}$, and the SWIS station is situated well within the stratified part of the Irish Sea, separated from the well-mixed region by a tidal mixing front. The field campaign lasted 10 d; here we focus on 50 h of continuous measurements taken 2 d immediately before neap tides (16–18 July). Measure-

ments included profiles of turbulent energy dissipation rate (ϵ), temperature, chl *a*, and nitrate.

Chlorophyll *a* and Inorganic Nutrients

[8] Seawater samples were collected from six to eight depths for determination of nitrate and chl *a* concentrations during eight conductivity temperature depth (CTD) profiles over 50 h. Chl *a* was extracted from 1 L of seawater filtered onto a Whatman glass-fiber filter (GF/F) by using acetone. Fluorescence was determined against a chl *a* standard by using a Turner TD-700 fluorometer (Turner Designs, USA; Evans et al. 1987). To provide high-resolution vertical profiles of chl *a*, the extracted chl *a* concentrations were used to calibrate a fluorescence sensor mounted on the CTD rosette frame. A regression of the extracted chl *a* concentrations against the fluorometer output (*V*) provided a calibration equation of $\text{chl } a_{\text{cal}} = 44.76V + 0.0529 \text{ mg m}^{-3}$ ($R^2 = 0.66$, $n = 28$, $p < 0.001$). The root-mean-square scatter of the chl *a* samples from this regression was 0.34. Sharples et al. (2001b) suggested that much of this scatter was the result of attempting to sample thin layers (i.e., 1–3 m thick) of concentrated chl *a* with a standard CTD and bottle rosette system, where the Niskin bottles are $\sim 1 \text{ m}$ long and $\sim 1 \text{ m}$ above the fluorometer.

[9] Concentrations of nitrate plus nitrite (herein nitrate, mmol m^{-3}) were determined using a LACHAT Instruments (USA) Quick-Chem 8000 autoanalyzer and standard colorimetric techniques (Hales et al. 2004; Papadimitriou et al. 2007). The analytical limits of detection for nitrate were 0.1 mmol m^{-3} .

Turbulence and Water Column Physical Structure

[10] A mooring at the SWIS station provided 8 d of continuous temperature and current measurements via the deployment of a thermistor chain with 22 VEMCO (Minilogger, VEMCO-AMIRIX, Canada) and SeaStar thermistors (Milli-T, Star-Oddi, Iceland); six of the thermistors were equipped with pressure sensors (Green et al. 2010). Sampling resolution was every 120 s with a vertical resolution between 2 and 10 m along the mooring line. A bed-frame-mounted 300-kHz acoustic Doppler current profiler (ADCP; Workhorse Sentinel, Teledyne RDI, USA) was deployed within 200 m of the thermistor mooring. The ADCP recorded current

velocity every 2 s with a 2-m vertical bin size from 4.7 m above the bed to within 15 m of the surface. All the data from both moorings were later interpolated to a common grid with 2-m vertical resolution and 2-min resolution in time (Green et al. 2010). The currents measured from the ADCP were used to calculate the bulk shear (in units of s^{-1}):

$$S^2 = \left(\frac{du}{dz}\right)^2 + \left(\frac{dv}{dz}\right)^2, \quad (1)$$

where du and dv describe the difference in velocity between two layers for the u and v components of velocity, respectively, and dz is the vertical distance (m) between the two layers.

[11] Vertical profiles of microstructure velocity shear and temperature were measured using a Fast Light Yo-yo (FLY) profiler (Dewey et al. 1987) for 50 h between 16 and 18 July 2006 (decimal days 196–198), immediately before neap tides (for full details of the FLY instrumentation used, see Simpson et al. 2009; Green et al. 2010). The FLY profiler was allowed to fall freely through the water column to the seabed, providing measurements of turbulent dissipation to within 0.15 m of the seabed at a vertical resolution of 1.5 m. The data from the upper 5 m of each FLY profile may have been degraded by the presence of the ship's wake and the acceleration of the profiler and were, therefore, removed during analysis. In the environment described here, five to six profiles could be obtained every hour. The microstructure shear and temperature vertical profiles were collected over the 50-h sampling campaign from the FLY shear profiler at ~ 1 cm vertical resolution. One ensemble of FLY dissipation data was made up of five to six consecutive vertical profiles. The profiles were averaged and calculated over 1.5 m depth bins. The microstructure shear was used to calculate the dissipation of turbulent kinetic energy, ε ($m^2 s^{-3}$), from

$$\varepsilon = 7.5\nu \overline{\left(\frac{du}{dz}\right)^2}, \quad (2)$$

where ν is the kinematic viscosity of seawater, du/dz is the (small scale) vertical shear, and the overbar denotes an average over 2 s of data. Hourly CTD casts provided independent temperature and conductivity measure-

ments for calibration of FLY temperature and conductivity sensors.

[12] Profiles of vertical eddy diffusivity, K_z ($m^2 s^{-1}$), were calculated from the profiles of the rate of dissipation of turbulent kinetic energy (ε) and the buoyancy frequency (N^2 , in s^{-2}) (e.g., Osborn 1980):

$$K_z = \Gamma \frac{\varepsilon}{N^2}, \quad (3)$$

where gamma ($\Gamma = 0.2$) represents the mixing efficiency (Osborn 1980). The mixing efficiency is defined as the ratio of potential energy gain to dissipation and is dependent on the shear, stratification, and turbulence. Our assumption of a constant mixing efficiency assumes that the stratification, shear, and turbulence are relatively stable, which is unlikely to be the case. Density (ρ , in $kg m^{-3}$) and the rate of turbulent kinetic energy (TKE) dissipation are both properties measured directly from the FLY profiler, and g is the acceleration due to gravity. The buoyancy frequency (N^2) within each 1 m bin was obtained via the vertical profile of density from the FLY profiler:

$$N^2 = \frac{-g \partial \rho}{\rho \partial z} \quad (4)$$

The vertical eddy diffusivity (Eq. 3) was used to calculate the vertical flux (J , in $mmol$ or $mg m^{-2} s^{-1}$) of a property, provided a reliable vertical gradient of that property was available. Thus, for nitrate or chl a , the vertical turbulent flux is

$$J = -K_z \frac{\partial(\text{nitrate_or_chl } a)}{\partial z}, \quad (5)$$

where $d(\text{chl } a)/dz$ and $d(\text{nitrate})/dz$ are the vertical chl a and nitrate gradients (units of $mg m^{-4}$ and $mmol m^{-4}$, respectively).

[13] In the western Irish Sea, vertical variations in salinity are small and produce a density difference over the water column of $< 0.25 kg m^{-3}$, while vertical temperature changes produce a density difference of $> 1 kg m^{-3}$, indicating that vertical stratification is controlled by temperature (Green et al. 2010). The correlation of nitrate or chl a with density or temperature within the thermocline was stronger than with depth, because the depth of the nitracline and SCM

changes according to the vertical movement of isopycnals (Sharples et al. 2007). This has important implications for flux calculations, because such movements of the SCM and isopycnals mean that selecting a single representative depth at which fluxes are calculated can lead to contamination of the turbulent diffusivity by high turbulence immediately below the thermocline in the bottom tidally mixed layer. Instead, fluxes need to be calculated across an isopycnal within the lower portion of the SCM (Gregg 1987). It is important to note that the nitrate/temperature relationship will be robust on relatively short timescales compared with the nutrient uptake rate of phytoplankton (Lucas et al. 2011). Here we are interested in calculating fluxes at the isopycnals marking the very base of the SCM and hence gaining an accurate value for the amount of nitrate being supplied to the SCM community and the amount of chl *a* being eroded into the bottom mixed layer. We used the chl *a*–density relationship for all CTD profiles to find the isopycnal marking the base of the SCM (Fig. 2).

[14] The vertical position of the SCM corresponds to the isopycnals between 1025.4 kg m^{-3} and 1026.2 kg m^{-3} . This is equivalent to a width of $> 25 \text{ m}$. We associate the base of the SCM with the 1026.3 isopycnal ($\pm 0.1 \text{ kg m}^{-3}$;

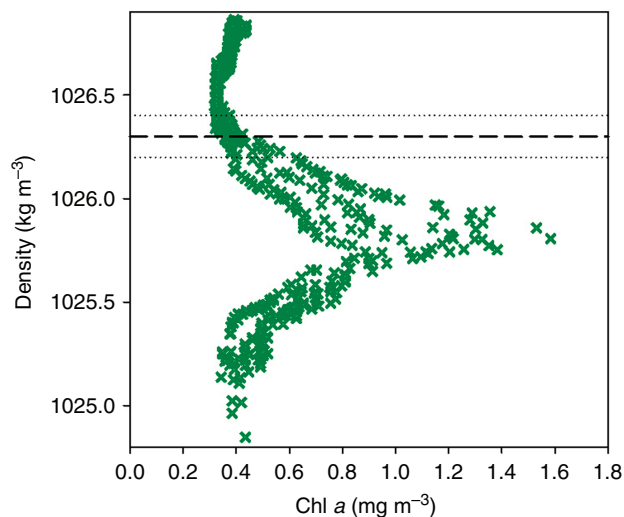


Fig. 2 Density versus chlorophyll *a* measurements during the Fast Light Yo-yo (FLY) sampling period. The dashed line marks the 1026.3 isopycnal ($\pm 0.1 \text{ kg m}^{-3}$ as marked by the dotted lines), which describes the density band marking the base of the subsurface chlorophyll maximum where fluxes were determined. Chlorophyll *a* concentrations were based on calibrated fluorescence values for each conductivity temperature depth cast during the FLY sampling campaign.

Fig. 2). As the FLY measures both density and TKE dissipation, we were able to obtain TKE dissipation rates on isopycnals. Furthermore, the density–nitrate and chl *a*–density relationships were calculated centered on the $1026.3 (\pm 0.1 \text{ kg m}^{-3})$ isopycnal (Fig. 2). K_z measurements were also calculated from TKE dissipation across the same isopycnal band, where K_z was taken as an average over the envelope of isopycnals used ($1026.2\text{--}1026.4 \text{ kg m}^{-3}$). To calculate fluxes, a nitrate–density and chl *a*–density relationship were determined by collating all nitrate and chl *a* data in the base of the SCM (Fig. 3).

[15] All chl *a*–density and nitrate–density relationships were statistically significant. The R^2 values show strong correlation between density and nitrate ($R^2 = 0.92$) or chl *a* ($R^2 = 0.94$). The chl *a* gradient at the base of the SCM is negative, indicating that any chl *a* fluxes will be out of the SCM into the bottom boundary layer, that is, loss of phytoplankton cells from the SCM.

[16] Using the nitrate–density and chl *a*–density relationships (Fig. 3) has the advantage that the scalar gradients are calculated using FLY density measurements, thus colocating in time and space the scalar gradients and the turbulent diffusivity. The nitrate (or chl *a*) flux calculation in Eq. 5 then simplifies to (Sharples et al. 2007)

$$J = m \frac{\Gamma \varepsilon \rho}{g}, \quad (6)$$

in units of $\text{mmol m}^{-2} \text{ s}^{-1}$. Here m is the slope of the density/nitrate (or chl *a*/density) linear regression as shown in Fig. 3. The mixing efficiency (Γ) is again taken as a constant of 0.2 (Osborn 1980).

[17] Because of high temporal and spatial variability of turbulence during sampling, a Gaussian “bootstrap” resampling method (Efron and Gong 1983) was used to calculate an average flux value with 95% confidence intervals (CIs) from instantaneous fluxes into the base of the SCM over the 50-h sampling period.

Results

[18] The thermocline thickness exceeded 20 m in all profiles (Fig. 4), with a density change of $\sim 1 \text{ kg m}^{-3}$. This broad thermocline separated an $\sim 20\text{-m}$ -thick surface layer with weaker density structure from an

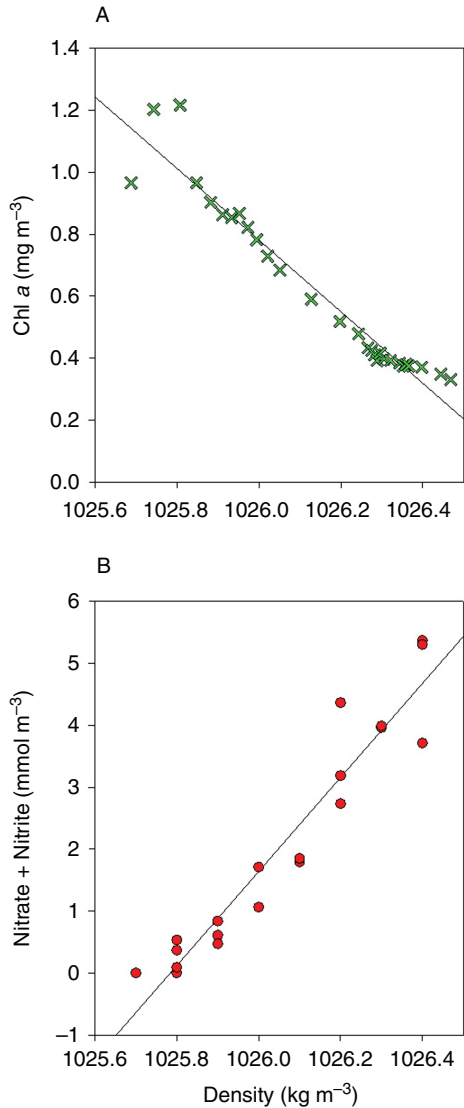


Fig. 3 Relationship between chlorophyll *a* (A) and nitrate + nitrite (B) and density for all measurements taken across the base of the subsurface chlorophyll maximum during the 50-h Fast Light Yo-yo sampling campaign. The lines of best fit for each relationship provide regression lines were chlorophyll *a* = (-1.16 density) + 1186.6 mg m⁻³ ($R^2 = 0.94$) and nitrate = (7.60 density) - 7791.80 mmol m⁻³ ($R^2 = 0.92$).

~60-m-thick bottom mixed layer. The SCM was equally broad (10–20 m), though the vertical position of the chl *a* peak varied between 20 and 30 m depth between CTD casts (Fig. 5B). The subsurface chl *a* concentration at the peak ranged from 0.7 to 1.5 mg m⁻³. Chl *a* concentrations were <0.5 mg m⁻³ in the upper 10 m of the water column during the duration of the FLY campaign. Nitrate concentrations were below the limits of detection in surface waters (0.1 mmol m⁻³).

[19] Turbulent energy dissipation (ϵ) varied by as much as 4 orders of magnitude through the water column (Fig. 5A). Enhanced turbulence was observed at the bed boundaries, as previously noted in the region (Simpson et al. 2009, Green et al. 2010), and typical for tidally dominated shelf seas. Maximum dissipation observed at the bed extended upward with an increasing phase lag; this results from the maximum shear occurring later and with decreasing intensity with increasing distance from the bed (Simpson et al.

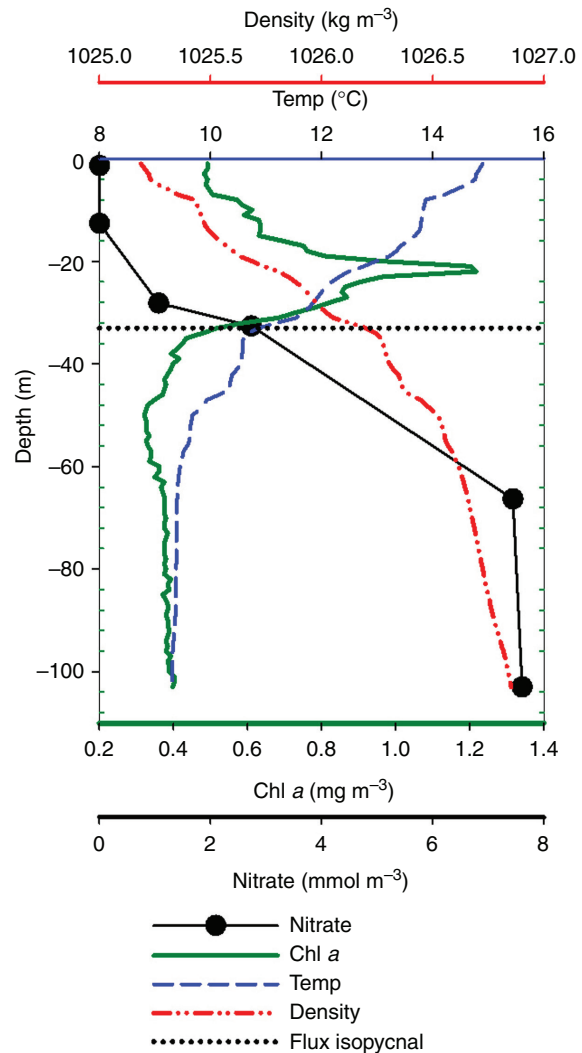


Fig. 4 Typical profiles of chlorophyll *a* and nitrate, temperature and density during the Fast Light Yo-yo sampling period for conductivity temperature depth 21 (day 196, 15 July 2006 23:16 local time). The 1026.3 isopycnal used to mark the base of the subsurface chlorophyll maximum where fluxes were determined is indicated by a dashed horizontal line.

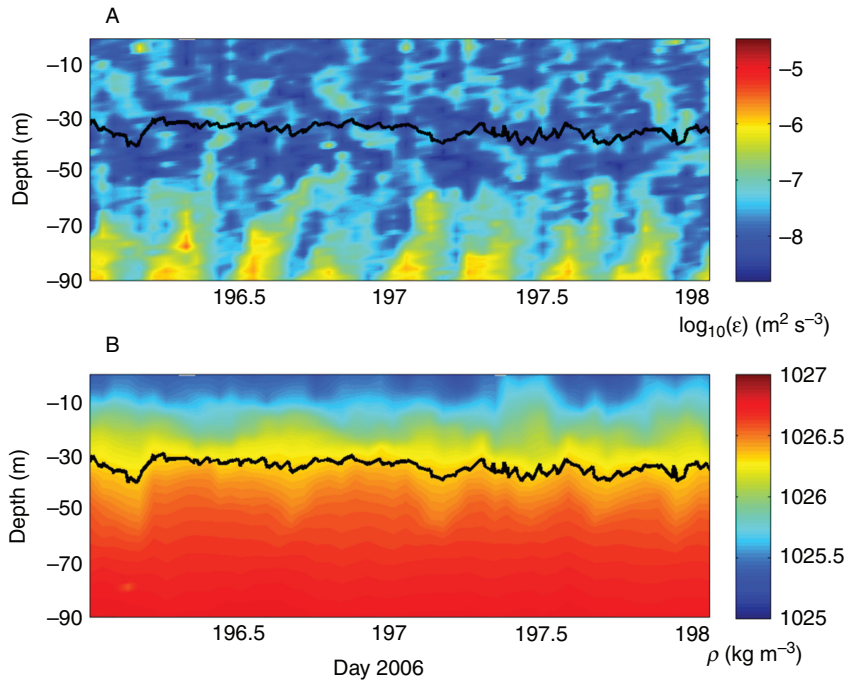


Fig. 5 A 50-h time series of turbulent kinetic energy dissipation ($\log_{10} \epsilon$; A) and density (ρ ; B) measured from the Fast Light Yo-yo profiler at the stratified western Irish Sea station in July 2006. The black line marks the 1026.3 kg m^{-3} isopycnal used to determine diapycnal fluxes of nitrate and chlorophyll *a* and marks the base of the subsurface chlorophyll maximum.

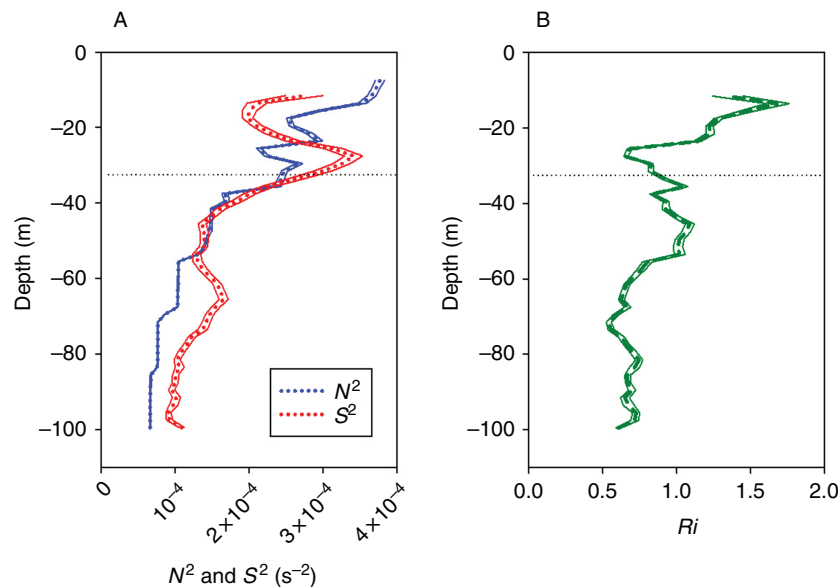


Fig. 6 A — Time-averaged vertical profiles of the shear squared, S^2 (s^{-2} ; red dashed line) and the buoyancy frequency N^2 (s^{-2} ; blue dashed line) averaged using a bootstrap method over a 9-d period between day 192 and 201 at the stratified western Irish Sea mooring station. The solid lines bordering the average indicate the 95% confidence intervals for estimates. B — Time-averaged gradient Richardson number, Ri , (dashed green line) over days 192–201, calculated from the buoyancy frequency and shear squared terms and averaged using a bootstrap resampling method. The solid lines bordering the average mark the 95% confidence limits. The dashed black horizontal line marks the approximate isopycnal depth where fluxes have been calculated.

2000). The midwater dissipation rates ranged from 10^{-9} to $10^{-7} \text{ m}^2 \text{ s}^{-3}$ and was dominated by short-lived turbulence events at the base of the thermocline (e.g., decimal day 196.5 in Fig. 5A). The lowest values of ϵ observed were within the thermocline and upper bottom mixed layer, between 20 and 60 m below the surface.

[20] To investigate the mechanisms contributing to the observed dissipation, it is necessary to consider potential shear instabilities and the probability of mixing at the thermocline. The time-averaged vertical shear squared (S^2) was calculated for each depth bin (Eq. 1) by using a bootstrap resampling method (Fig. 6A) over the 8-d period of ADCP measurements. S^2 was highest within thermocline, at a slightly shallower depth than our chosen isopycnals, which have an average S^2 value of $1.5 \pm 0.06 \times 10^{-4} \text{ s}^{-2}$. The 95% CIs for S^2 (Fig. 6A) indicate that during our measurements there was relatively small variability of shear in the water column. There was a 180° phase shift between currents in the surface and bottom mixed layer consistent with the presence of baroclinic energy driven by a mode 1 internal tide (Green et al. 2010). Additionally, the squared buoyancy frequency (N^2) was shown to have small variability throughout the water column for the duration of our measurements and is not significantly different from S^2 within our selected isopycnals ($1.48 \pm 0.04 \times 10^{-4} \text{ s}^{-2}$). Shear instabilities occur when the gradient Richardson number (Ri) is 0.25, though the water column is said to be unstable when $Ri < 1$ (Rippeth

2005; Thorpe and Liu 2009). Similar values for N^2 and S^2 would give an Ri of ~ 1 in the chosen isopycnal band, implying that the flow is marginally stable. For the duration of our campaign, Ri ranged between 0.45 and 2 within the selected isopycnals band but had a mean value of 1.1 ± 0.03 . However, it is important to note that the ADCP bin size (2 m) is significantly larger than the Ozmidov scale of shear in the midwater column. Therefore, the shear and thus Ri estimates provided here may have eliminated the detail in the small-scale shear (< 2 m) at the base of the thermocline.

[21] Turbulent dissipation was patchy within the base of the SCM. The instantaneous nitrate fluxes experienced at the base of the SCM from single profiles ranged from $\sim 9 \times 10^{-7}$ to $2 \times 10^{-5} \text{ mmol m}^{-2} \text{ s}^{-1}$. Chl a was eroded from the SCM via diapycnal mixing at a rate ranging between 10^{-8} and $3 \times 10^{-6} \text{ mg m}^{-2} \text{ s}^{-1}$ chl a .

[22] Chl a fluxes showed the same spikes as those shown by the nitrate fluxes (Fig. 7), largely because variability in scalar fluxes is dominated by the variability of the turbulence rather than changes to the scalar gradients. Table 1 summarizes the daily chl a and nitrate fluxes at the base of SCM. These fluxes are converted into milligrams C to compare C transfer associated

with phytoplankton cells being mixed out of the SCM (i.e., downward chl a fluxes) and new nitrate-driven production (i.e., fueled by upward nitrate fluxes). To convert chl a flux into a C flux, we assume a constant C:chl a ratio of 40:1 for SCM phytoplankton (Holligan et al. 1984). The Redfield ratio (Redfield 1958) was used to estimate nitrate-supported C fixation rates (units of $\text{mmol m}^{-2} \text{ d}^{-1}$, or $\text{mg m}^{-2} \text{ d}^{-1}$ when multiplied by the molecular weight of C, 12). Whereas the C to nitrogen (N) ratio in phytoplankton can vary (Tett et al. 1985), we assume that the average C:N ratio is 6.6:1 for phytoplankton growth in response to a supply of nitrate (Geider and La Roche 2002).

[23] The net rates of C fixation reported in Table 1 represent the potential net organic C accumulation in the SCM as a result of the turbulent erosion of live cells and growth fueled by diapycnal nitrate fluxes. The 95% CIs (calculated using the bootstrap method) for nitrate and chl a fluxes had a small range ($0.22\text{--}0.43 \text{ mmol m}^{-2} \text{ d}^{-1}$ and $0.036\text{--}0.072 \text{ mg m}^{-2} \text{ d}^{-1}$, respectively), indicating that we have calculated a background estimate with low variability for diapycnal fluxes across the SCM and with little evidence of sporadic spikes for the period we sampled compared with other studies (e.g., Sharples et al. 2007). The loss of C via erosion of live cells from the SCM was $\sim 1.92 \text{ mg m}^{-2} \text{ d}^{-1}$ (95% CI = $1.44\text{--}2.88 \text{ mg m}^{-2} \text{ d}^{-1}$); the potential C fixation that could be supported by the nitrate flux into the SCM was approximately 12 times more than this export via turbulent erosion ($24.7 \pm 1.3 \text{ mg m}^{-2} \text{ d}^{-1}$). It is important to note that the C loss estimate via turbulent erosion accounts for only the removal of intact cells containing chl a from the SCM and not the removal of dead cells.

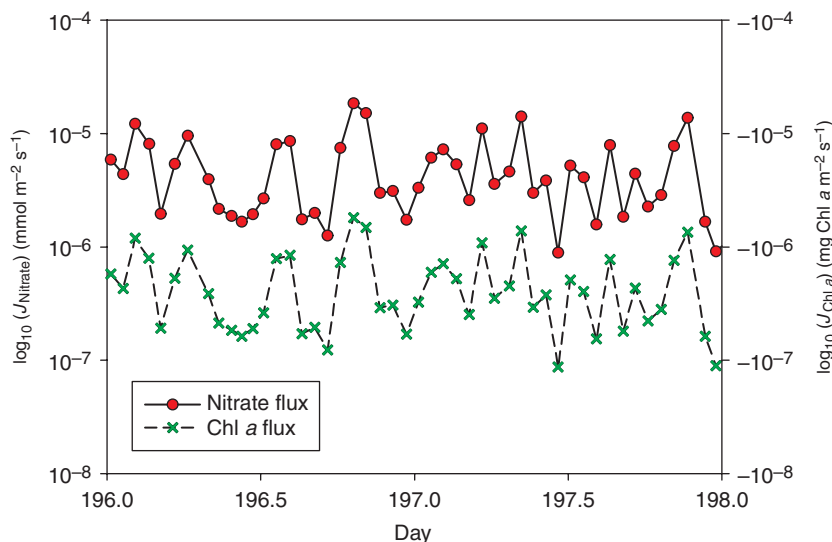


Fig. 7 Variation in nitrate fluxes and chlorophyll a instantaneous fluxes through the base of subsurface chlorophyll maximum (SCM) from the 50-h sampling period. The values shown are the average fluxes (every six Fast Light Yo-yo profiles, i.e., hourly averages). The \log_{10} nitrate fluxes shown are the nitrate flux into the base of the SCM; the chlorophyll a flux shown is the chlorophyll a flux out of the base of the SCM and hence is negative.

Discussion

[24] Observations of turbulence, nitrate, and chl a in the seasonally SWIS during a period approaching neap tides indicate that background

Table 1 Daily nitrate and chlorophyll *a* (chl *a*) fluxes and the corresponding estimates of nitrate-driven production, carbon (C) export and net C gained per day. The values for the fluxes are calculated using the bootstrap sampling method and the values in parentheses are the 95% confidence intervals. The net rates of C fixation reported represent the potential net organic C that could result from diapycnal fluxes.

Nitrate flux (mmol m ⁻² d ⁻¹)	Nitrate-driven production (mg m ⁻² d ⁻¹ C)	Chl <i>a</i> flux (mg m ⁻² d ⁻¹ C)	C export (mg m ⁻² d ⁻¹)	Potential net C accumulation (mg m ⁻² d ⁻¹)
0.31 (0.22 to 0.43)	24.7 (17.0 to 34.3)	-0.048 (-0.036 to -0.072)	-1.92 (-1.44 to -2.88)	22.8 (15.6 to 31.4)

nitrate fluxes are the lowest observed in shelf seas to date (Table 2). We have shown that the midwater turbulence in the SWIS supplies 0.22–0.43 mmol m⁻² d⁻¹ N to the SCM and erodes phytoplankton from the base of SCM at a daily rate of 0.04–0.07 mg m⁻² d⁻¹ chl *a*. By assuming the Redfield ratio and a fixed C:chl *a* ratio, we have translated the nitrate flux and chl *a* erosion rate into units of C (Table 1). During the field campaign, nitrate was consistently below the limits of detection in the surface waters. This suggests that nitrate supplied to the euphotic zone via midwater mixing is removed from the water column at the SCM, preventing transport into the surface layer. The concentrations of chl *a* observed at the SCM varied between 0.7 and 1.5 mg m⁻³ and were consistently <0.5 mg m⁻³ in the surface waters, corresponding with the very low concentrations of nitrate observed here. The lowest levels of turbulence in the SWIS were observed within the thermocline and upper bottom mixed layer, with rates of turbulent dissipation varying between 10⁻⁹ and 10⁻⁷ m² s⁻³.

[25] The SWIS has an internal tide energy flux of 100 W m⁻¹ (Green et al. 2010), which is quantitatively similar to the internal tide energy flux documented at the Celtic Sea shelf edge (~139 W m⁻¹; Green et al. 2008). The nitrate flux we have observed in the SWIS, however, is much lower than that observed at the Celtic shelf edge during neaps (Table 2; 3 mmol m⁻² d⁻¹;

Sharples et al. 2007). This suggests that although the internal tide here is almost as energetic as that observed at the shelf edge, it propagates through the region conservatively, with low levels of dissipation and, therefore, mixing within the thermocline. A comprehensive study of the physical processes during this field campaign has been provided by Green et al. (2010), who highlight that there is no clear separation of the contribution to the mixing from each of these processes. The time-averaged *Ri* calculated at the base of the SCM was 1.1 ± 0.03 and was observed to occasionally approach the critical value of < 1 for the duration of our sampling. This indicates that the base of the SCM was in a state of marginal stability (Fig. 6); similar results have also been reported for other seasonally stratified temperate shelf seas (van Haren et al. 1999; MacKinnon and Gregg 2003; Rippeth et al. 2005). In the SWIS it appears that the internal tide sets up a background state that on its own is insufficient to produce shear instabilities and significant mixing but may be modified via the addition of shear from nonlinear internal waves and wind-driven inertial oscillations. The addition of shear from these processes would most likely be enough to result in a *Ri* of ~0.25 and thus cause shear instabilities and intermittent mixing. We argue that internal tide energy fluxes are, therefore, not sufficient for estimating vertical mixing and the diapycnal fluxes of nutrients.

Table 2 Comparisons of documented turbulent kinetic energy dissipation (ϵ), daily nitrate fluxes and potential new nitrate-driven production from various shelf seas in the North Atlantic Ocean. Stratified western Irish Sea, SWIS.

Location	ϵ observed in thermocline (m ² s ⁻³)	$d(\text{nitrate})/d\rho$ (mmol m ⁻⁴)	Daily J_{nitrate} into base (mmol m ² d ⁻¹)	Nitrate-driven production (mg m ⁻² d ⁻¹ C)
SWIS (present study)	10 ⁻⁹ –10 ⁻⁷	7.6	0.2–0.4	15.8–34.1
SWIS (Rippeth et al. 2009)	10 ⁻⁸ –10 ⁻⁷	—	1.2–1.8	95–142.6
Celtic shelf edge (Sharples et al. 2007)	10 ⁻⁵ –10 ⁻⁴	10.4–11.2	1.3–9	103–712.8
Western English Channel (Sharples et al. 2001b)	10 ⁻⁸ –10 ⁻⁵	5.7	0.8–3.2	63–253

[26] The estimate for the daily nitrate flux to the base of the SCM ($0.31 \pm 0.01 \text{ mmol m}^{-2} \text{ d}^{-1}$) is approximately a fifth of the estimate given by Rippeth et al. (2009) in the same region ($1.5 \pm 0.05 \text{ mmol m}^{-2} \text{ d}^{-1}$; Table 2). The low range in the variability of the nitrate flux signal (see 95% CI in Table 1) indicates that the SWIS data were background, not event-driven fluxes. Rippeth et al. (2009) calculated the nitrate flux across the thermocline by using a fixed depth and a fixed isotherm method, this latter method is similar to the method we used to estimate the nitrate flux here. Both methods resulted in very similar flux estimates, which may suggest little vertical movement of isopycnals during their observations. By replicating the fixed depth method in our observations, we estimate a mean nitrate flux of $1.3 \pm 0.4 \text{ mmol m}^{-2} \text{ d}^{-1}$. This is significantly higher than the nitrate flux we report across a fixed isopycnal. The higher flux results from episodic contamination of thermocline turbulence with high turbulence within the upper part of the tidally mixed bottom layer. (These high-turbulence events are not relevant to the mixing of scalars through the base of the thermocline.) The large discrepancy in flux estimates, by using the fixed depth and fixed isopycnal methods, suggests that the vertical movement of isopycnals, by the internal tide was significant during our observations. Additionally, the vertical eddy diffusivity (K_z) within the thermocline during our observations ($0.77 \pm 0.25 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$) was significantly smaller than that reported by Rippeth et al. (2009) ($1.72 \pm 0.41 \text{ m}^2 \text{ s}^{-1}$) for the period of flux calculations when intermittent shear spikes were observed. This highlights the variability of the SWIS in terms of intermittent mixing events.

[27] The diapycnal nitrate flux has the potential to support new production of $25 \text{ mg m}^{-2} \text{ d}^{-1} \text{ C}$, which over a summer period of 120 d would result in $3 \text{ g m}^{-2} \text{ C}$ of new production in the SWIS. This potential daily production rate is relatively low compared with the Celtic Sea SCM, where primary production estimates during summer range from 170 to $390 \text{ mg m}^{-2} \text{ d}^{-1} \text{ C}$ (Hickman et al. 2012). Primary production estimates for the SWIS are sparse; the most comprehensive estimate taken from Gowen and Bloomfield (1996) suggests a primary production value between April and September of $140 \text{ g m}^{-2} \text{ C}$, which is only slightly higher than the

Celtic Sea (Hickman et al. 2012). By using the Redfield ratio (C:N = 6.6), we can estimate the amount of nitrate required to drive the production estimated by Gowen and Bloomfield (1996). We can assume (following Trimmer et al. 1999) that $\sim 70\%$ of this production is derived from the spring bloom (60 d); therefore, primary production during summer (120 d) is responsible for the remaining 30% ($42 \text{ mg m}^{-2} \text{ C}$). The amount of nitrate required to fuel $42 \text{ mg m}^{-2} \text{ C}$ of new production is 530 mmol m^{-2} ; our daily nitrate flux estimate extrapolated over the summer period of 120 d would provide 37.2 mmol m^{-2} . This would be 7% of the nitrate for summer production provided by the diapycnal nitrate flux. This could also be interpreted as an f ratio (the ratio of total primary production fueled by nitrate) of 0.07, which seems unlikely. Alternatively, a study by Trimmer et al. (1999) used chl a standing stock measurements to estimate total (new from nitrate and regenerated from recycled nitrogen) between February and July, giving a total annual production value of $44.8 \text{ g m}^{-2} \text{ yr}^{-1} \text{ C}$. Again, 30% ($13.44 \text{ g m}^{-2} \text{ C}$) of this total annual production occurs during summer. Approximately 170 mmol m^{-2} of nitrate is needed to drive summer production estimated by Trimmer et al. (1999). Extrapolating the nitrate flux, we found that the diapycnal nitrate flux in the SWIS provides 16%–30% of the nitrate required for total primary production during summer. Primary production estimates provided by these studies suggest that our primary production rate is very low if extrapolated over the summer period. The implication is that the diapycnal-flux-driven new production does not integrate to the mean production over the summer. Instead, the wind and/or nonlinear internal waves may be crucial in supporting phytoplankton populations during summer by increasing the average flux of nitrate substantially.

[28] At the base of the SCM, the loss of C via erosion of the SCM is negligible compared with the possible nitrate-driven production. The turbulent erosion of phytoplankton cells from the SCM via background mixing potentially exports almost $2 \text{ mg m}^{-2} \text{ d}^{-1} \text{ C}$. This corresponds to $\sim 0.2 \text{ g m}^{-2} \text{ C}$ exported from the SCM during summer in the SWIS. C loss via erosion is $\sim 7\%$ of the potential nitrate-driven production if we assume that all nitrate is assimilated

by the phytoplankton population and converted into biomass at the Redfield ratio. It is important to note, however, that the C export value accounts for only the loss of live cells transported out of the SCM during mixing and does not consider loss by grazing, viral cell lysis, or direct sinking of particles. Therefore, the remaining 93% of potential nitrate-driven new production in the SCM must be removed via the various removal processes of dead cells that occur during periods of low wind and approaching neap tides. Event-driven mixing, such as nonlinear internal waves and/or wind, would result in enhanced chl *a* erosion from the SCM, as well as increased nitrate fluxes. In order to accurately quantify C budgets in shelf seas, the export of C to the deep water via turbulent erosion of live cells (at least 7%) should be taken into account, as well as the other removal processes.

Significance to Aquatic Environments

[29] Identifying and quantifying the supply pathways of nutrients to phytoplankton in stratified seas is vital to understand primary production and hence carbon (C) use in the ocean. Such pathways would include C export from the euphotic zone, as well as the delivery of nutrients. This study identified the physical processes that drive new production in an isolated, stratified oceanic area via physical and biogeochemical measurements and estimated the export of C via turbulence.

[30] We found that although the SWIS has a significant internal tide (an internal wave caused by tides on the water surface), it propagates through the region causing only minor dissipation (energy loss to water motion) at the thermocline and hence contributes to relatively small nitrate fluxes into the surface waters. This quiescent, “background” mixing driven by the internal tide is not sufficient to sustain the C fixation observed in the region, and thus event-driven mixing, via the wind and/or nonlinear internal waves, must be responsible for the maintenance of summer production in this ecologically important environment. The low-level mixing indicates the thermocline is in a state of marginal stability, meaning that shear from the wind or nonlinear internal waves would likely “tip the balance” and generate diapycnal mixing (i.e., mixing of nutrients through the thermocline).

[31] In the past decade there has been increased interest in climate change, including increased winds in the North Atlantic. If maintenance of primary production in the SCM in temperate regions such as SWIS depends on event-driven large pulses of nutrients, then increased wind may enhance primary production during summer. This is contrary to the belief that increased winds are more likely to dampen production by deepening the euphotic zone and causing phytoplankton to experience a reduction in light.

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