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Diagnostic criteria for reconstruction of tidal continental shelf regimes: changing the paradigm

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Abstract

A biogeochemical flux paradigm is presented which provides a conceptual and numerical framework for reconstructing the dynamical and biogeochemical regimes of ancient tide-driven continental shelves. The paradigm links turbulence, primary production, and suspended particulate matter flux in the water column to microbiological and isotopic proxies in the sediment record and identifies the diagnostic signatures of sediments deposited in stratified, frontal, and mixed dynamic regimes of tidal shelves. The critical governing processes in the water column and at the sediment/water interface are temperature, particulate organic carbon supply, and benthic oxygen consumption, which exhibit strong gradients across tidal mixing fronts. The diagnostic proxies in the sediment record are microplanktonic (e.g. dinoflagellate cysts) and microbenthic (e.g. foraminifera) and their associated stable isotopic properties. Advanced numerical models are available which incorporate biophysical coupling in the water column and benthic boundary layer and which simulate biogeochemical and ecological processes and organic fluxes to the seabed. These models offer potential advances for interpreting microbiological and isotopic proxies of biogeochemical regime preserved in the fine sediment record. Quaternary shelf deposits provide the best potential validation of the biogeochemical flux paradigm since most Quaternary species are living today, but the paradigm is applicable to ancient tidal shelf deposits. The paradigm is particularly applicable to the fine sediment record which potentially preserves the most complete history of shelf evolution.

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1. Introduction

This paper proposes that a palaeoceanographic and stratigraphic model of a tidal shelf sea can be supported within a biogeochemical flux paradigm based on an ecological model of tidal shelf seas.

The essential tenet of the paradigm is that mixed and stratified regions of tide-driven continental shelves exhibit distinctive biogeochemical and ecological regimes so that there are strong gradients between them. It is proposed that these gradients are reflected in microfaunal and isotopic proxies in the seabed sediments. Such proxies, diagnostic of water column structure and hydrodynamic regime, can be used to reconstruct the temporal evolution of tidal shelf regimes in response to eustatic and climatic change.

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2. Shelf sea paradigms

2.1. *Sediment dynamics paradigm*

Application of new acoustic technologies in the 1960s and 1970s underpinned a comprehensive study of the sedimentary facies of the tide-driven shelves of NW Europe and spawned a sediment dynamics paradigm in which sediments and hydrodynamics are in dynamic equilibrium (see [Stride, 1982](#) for a synthesis). This paradigm addresses the benthic boundary shear stress and the time–shear stress asymmetry that gives rise to net sand transport. Non-linearities in the equations of motion generate higher harmonics of the fundamental tide (particularly the M_4 second harmonic) and combination of M_2 and M_4 results in an asymmetry between flood and ebb seabed stresses ([Pingree and Griffiths, 1979](#)). As a result, there is net sediment transport. Furthermore, flood and ebb currents may follow mutually evasive paths (e.g. [Caston, 1972](#)) and net sediment transport may be enhanced by other physical forcings such as wind stress currents. Transport paths, parallel to the dominant tidal currents, are characterised by diagnostic sequences of bedforms and facies ([Johnson et al., 1982](#)), which can be modelled in terms of bed shear stress ([Pingree and Griffiths, 1979](#)). The dispersal patterns represent the ‘unmixing’ ([Swift, 1970](#)) of the glacial sediments that comprise the bulk of the coarser sediments of the NW European shelf. Since the physics of sand transport and tidal bedforms are reasonably well understood, numerical models can be applied to ancient tidal shelf deposits to infer current velocity and seabed shear stress. While there is nothing inherently unique about boundary layer physics and transport in a tidal regime ([Flemming \(1980, 1981\)](#)), the sediment dynamics paradigm is a powerful tool for interpretation of ancient tidal shelf deposits.

2.2. *Biogeochemical flux paradigm*

The sediment dynamics paradigm deals with sand transport in the benthic boundary layer of the inner shelf. It contributes little to interpretation of fine sediments found at the end of trans-

port paths in water deeper than wave base ([McCave, 1971, 1972, 1985](#)). Fine sediment depocentres are regions of lowered bed shear stress so there are few textural or structural manifestations of benthic boundary layer dynamics. Our understanding of fine, cohesive sediment response to boundary layer dynamics is in any case limited. Crucially, the sediment dynamics paradigm has spawned neither conceptual nor numerical models that can be applied to the fine sediment record. This is doubly unfortunate since fine sediments constitute the majority of shelf deposits and, where they accumulate in quiescent depocentres, they have good preservation potential.

The biogeochemical flux paradigm encompasses pelagic processes and cohesive sediments. The key is the coupling of pelagic and benthic regimes, particularly with regard to seabed exchanges. The flux of cohesive material in suspension (i.e. suspended particulate matter or SPM), particularly in relation to physical and biological forcings, is important since vertical SPM flux controls benthic biogeochemical exchanges. The flux of SPM in turn depends on the interactions of water column dynamics and plankton production. We propose that proxies, diagnostic of these biogeochemical processes and exchanges, are preserved in cohesive sediments.

2.3. *Physical forcing: the tidal mixing model*

The tidal mixing model addresses the physics of seasonal thermal stratification ([Simpson and Hunter, 1974](#); [Pingree, 1975](#); [Pingree and Griffiths, 1978](#)) on tidally stirred shelves. Daily inputs of turbulent kinetic energy from tidal and wind stirring, and convective mixing, are opposed by buoyancy inputs due to solar heating. The balance between these depends on seasonal variability of the surface heat flux. In deeper regions, seasonal stratification develops whereby a thermocline separates surface and bottom mixed layers (SML and BML, mixed by wind and tide, respectively). Shallower regions remain mixed at all times (except during some neap tides and in regions of freshwater influence (ROFIs) where enhanced buoyancy is provided by river water). Between stratified and mixed regions is a tidal

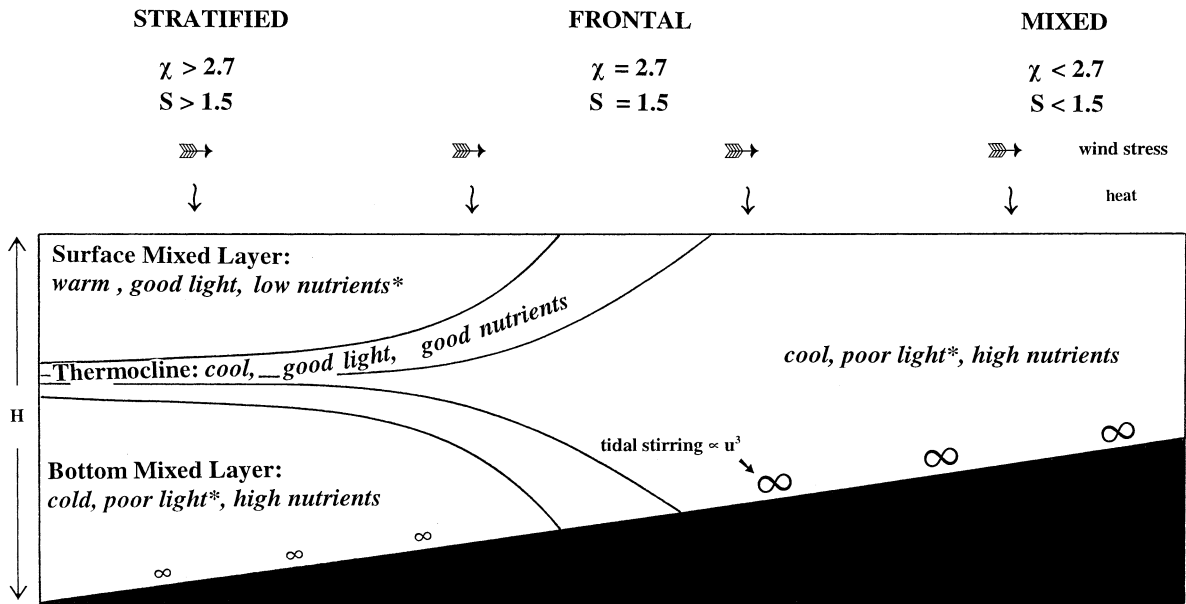


Fig. 1. Conceptual diagram of water column structure and water quality in tidal shelf seas. See text for explanation of symbols. * = factor which limits algal growth.

mixing front which is the surface expression of the thermocline (Fig. 1). Fronts break down in the autumn when tidal stirring overcomes the decreased buoyancy that results from a decrease in surface heat flux. Tidal fronts have been widely observed on the NW European shelf: Ushant front (Pingree et al., 1975, 1977, 1978; Grall et al., 1980), Celtic Sea (Savidge, 1976; Savidge and Foster, 1978), Irish Sea (Savidge, 1976; Foster et al., 1976; Beardall et al., 1978), Islay front (Pingree et al., 1978; Simpson et al., 1979), Orkney and Shetland fronts (Dooley, 1976; Pingree et al., 1978) and southern North Sea (Harding et al., 1978; Pingree et al., 1978). Fronts associated with ROFIs (e.g. the Rhine ROFI described by Simpson et al., 1993 and Simpson, 1997) differ from thermal tidal mixing fronts in that a freshwater runoff-stirring competition replaces the heating-stirring competition: ROFI-associated fronts are not considered further in this paper.

A front develops where the intensity of tidal turbulent mixing is sufficient to overcome the tendency to stratify due to surface heating. By equating turbulent mixing to the potential energy difference before and after mixing, it has been shown

that the critical determinant of water column structure is tidal stirring, predicted by h/u^3 , where u is a depth-mean average tidal current and h is water depth (Simpson and Hunter, 1974). This is defined as:

$$\frac{h}{u^3} = \frac{8C_p \rho k \varepsilon}{3\pi \alpha g Q} \quad (1)$$

where h is water depth, u is tidal current velocity, C_p is the specific heat capacity of water, ρ is water density, k is a constant in the assumed quadratic friction law, ε is the proportion of turbulent energy available for increasing the potential energy of the water column, α is the thermal expansion coefficient of water, g is the acceleration due to gravity, and Q is the rate of heat input to the water column. Since for a given tidal basin this quotient remains constant during the summer months (assuming constant surface heating), fronts in that basin remain stationary along contours of a critical value of the quotient. For a given heat flux, after correction for tidal advection (Simpson and James, 1986), the position of fronts on the NW European shelf complies with a critical transition value of h/u^3 given by:

$$\chi = \log_{10} \left(\frac{h}{\langle u_2^3 \rangle} \right) = 2.7 \pm 0.3 \quad (2)$$

where u_2 is the modulus of the depth-mean M_2 current (Simpson, 1998). The uncertainty in the critical value is equivalent to a displacement of the front by ca. ± 10 km, some of which is due to wind stirring. An alternative formulation is given by Pingree and Griffiths (1978). A combined wind and tide stirring model improves the precision of front prediction (Simpson and Bowers, 1981). Bowers and Simpson (1987) assessed several theoretical predictions with observations of frontal positions on the NW European shelf and concluded that for shallow waters those formulations which include wind mixing are the most accurate.

More sophisticated formulations of water column structure based on consideration of turbulent kinetic energy (tke) have been developed in recent years (e.g. Simpson and Sharples, 1991; Sharples and Tett, 1994; Simpson et al., 1996; Luytens et al., 1996; Sharples, 1999). In such models, a turbulence closure scheme (e.g. Mellor and Yamada, 1982) links vertical stratification, driven by surface heating, and turbulence generated by tidal friction at the seabed and wind stress at the surface; in effect, water column stability is related to the efficiency of vertical turbulent transport. Thus,

changes in the $(q^2/2)$ where q is the turbulent velocity scale) are given by:

$$\frac{\partial}{\partial t} \left(\frac{q^2}{2} \right) = \frac{\partial}{\partial z} \left[K_q \frac{\partial}{\partial z} \left(\frac{q^2}{2} \right) \right] + N_z \left[\left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial v}{\partial z} \right)^2 \right] + K_z \left(\frac{g}{\rho} \frac{\partial \rho}{\partial z} \right) - \frac{q^3}{B_1 l} \quad (3)$$

where K_q is the vertical eddy diffusivity of tke, N_z is the depth-dependent coefficient of eddy viscosity, K_z is the vertical eddy diffusivity, u and v are the x and y components of current velocity, g is the acceleration due to gravity, ρ is water density, B_1 is a constant of the turbulence closure scheme, and l is the constant turbulent length scale. This formulation contains the vertical diffusion of tke, shear production of turbulence, work done against buoyancy, and the dissipation of tke (the terms on the right hand side of Eq. 3, respectively). Distribution of heat input from solar radiation is mediated by attenuation due to SPM and algal biomass in the water column (see below). Such models have been very successful in simulating vertical mixing processes in general, and genesis of the thermocline in particular.

An important consequence of frontogenesis is the large contrast in water temperature between mixed and stratified regions, at the seabed as well as at the surface (Fig. 2). While vertical temper-

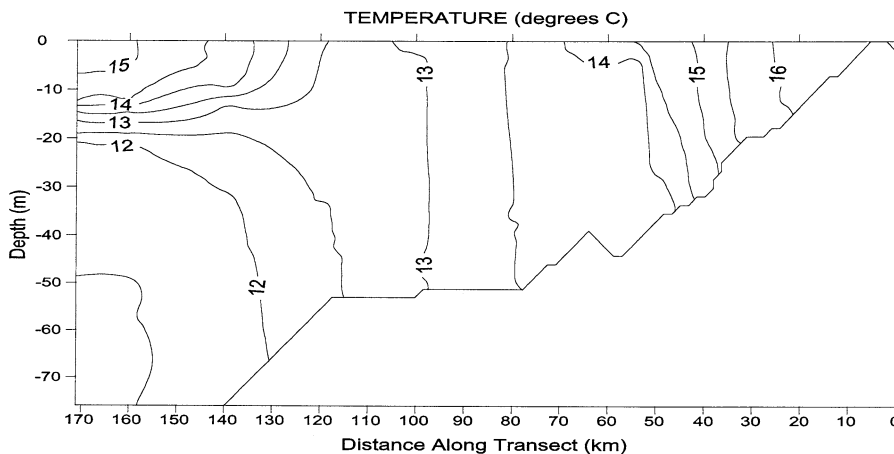


Fig. 2. Temperature section across the Irish Sea between Liverpool Bay (right) to western Irish Sea gyre (left). From Rippeth et al. (2001).

ature gradients are small in mixed waters (1–2 EC between surface and bottom), they can be large in summer stratified waters (5–6 EC). Frontal regions are characterised by large lateral surface temperature gradients of more than 1 EC km⁻¹ which can be detected by satellite imagery (e.g. Fig. 3A and Simpson and Bowers, 1979). Summer bottom temperatures may be lower in stratified than in mixed waters so there are large bottom temperature gradients: e.g. 1–2 EC and 7–9 EC across the Celtic Sea and southern North Sea fronts, respectively (Elliott et al., 1991).

A front is not an abrupt transition but the frontal zone is quite narrow (ca. 5 km wide). The frontal density structure implies a pressure gradient perpendicular to the front which drives an along-front current. The geostrophic balance is not perfect, due to friction, and a small cross-stream flow is generated down the pressure gradient. Consideration of the cross-frontal dynamical balance, including friction (James, 1978; Garrett and Loder, 1981), suggests a weak cross-frontal circulation with a surface convergence close to the region of maximum horizontal gradient (e.g. Pingree et al., 1974). The Simpson–Hunter stratification parameter predicts a lateral migration of fronts during the lunar cycle of 10–20 km. Even after removing tidal advection, Simpson and Bowers (1979) concluded that fronts migrate laterally by a few kilometers in response to the lunar cycle. Numerical models (Simpson and Bowers, 1981; Sharples and Simpson, 1996) suggest that the position of the frontal zone lags the spring-neap cycle by a few days. A major consequence of lateral migration is that water from the mixed region is incorporated into the stratified region as the front advances into shallower water on neap tides (Simpson and Hunter, 1974; Simpson and Bowers, 1979). Furthermore, infrared satellite imagery and direct observations show that baroclinic instabilities along the front grow into large eddies (typically 25–40 km) which make a large contribution to cross-frontal mixing (Pingree and Griffiths, 1978). In addition, some water from the BML may be mixed across the thermocline by spring tide currents. Finally, some diffusion of the thermocline might occur due to dissipation of internal tides.

2.4. Biological response: impact on phytoplankton

Vertical stability of the water column is important to summertime phytoplankton distributions on the NW European shelf (Pingree et al., 1976, 1978; Houghton, 1988). Primary production in shelf seas is determined by nutrient availability, grazing pressure, and growth rate. Growth rate depends on nutrient availability (Droop, 1983) and light, while peak biomass is constrained by nutrients (Tett et al., 1993). Tidal stirring controls the availability of both nutrients and light.

In mid and high latitude shelves, where plankton dynamics are dominated by transients such as the spring bloom, it is likely that growth rate is the most important regulator of production (Tett et al., 1993). The temporal evolution of phytoplankton biomass is determined by growth and vertical turbulent transport, tempered by grazing (Sharples and Tett, 1994; Sharples, 1999; Sharples et al., 2001):

$$\frac{\partial X}{\partial t} = \frac{\partial}{\partial z} \left(K_z \frac{\partial X}{\partial z} \right) + \mu X - gX \quad (4)$$

where X is biomass and μ is the specific growth rate, K_z is vertical eddy diffusivity, and g is the loss of algal biomass to grazers. This growth rate may be nutrient-limited:

$$\mu = \mu_m \left(1 - \frac{k_Q}{Q} \right) \quad (5)$$

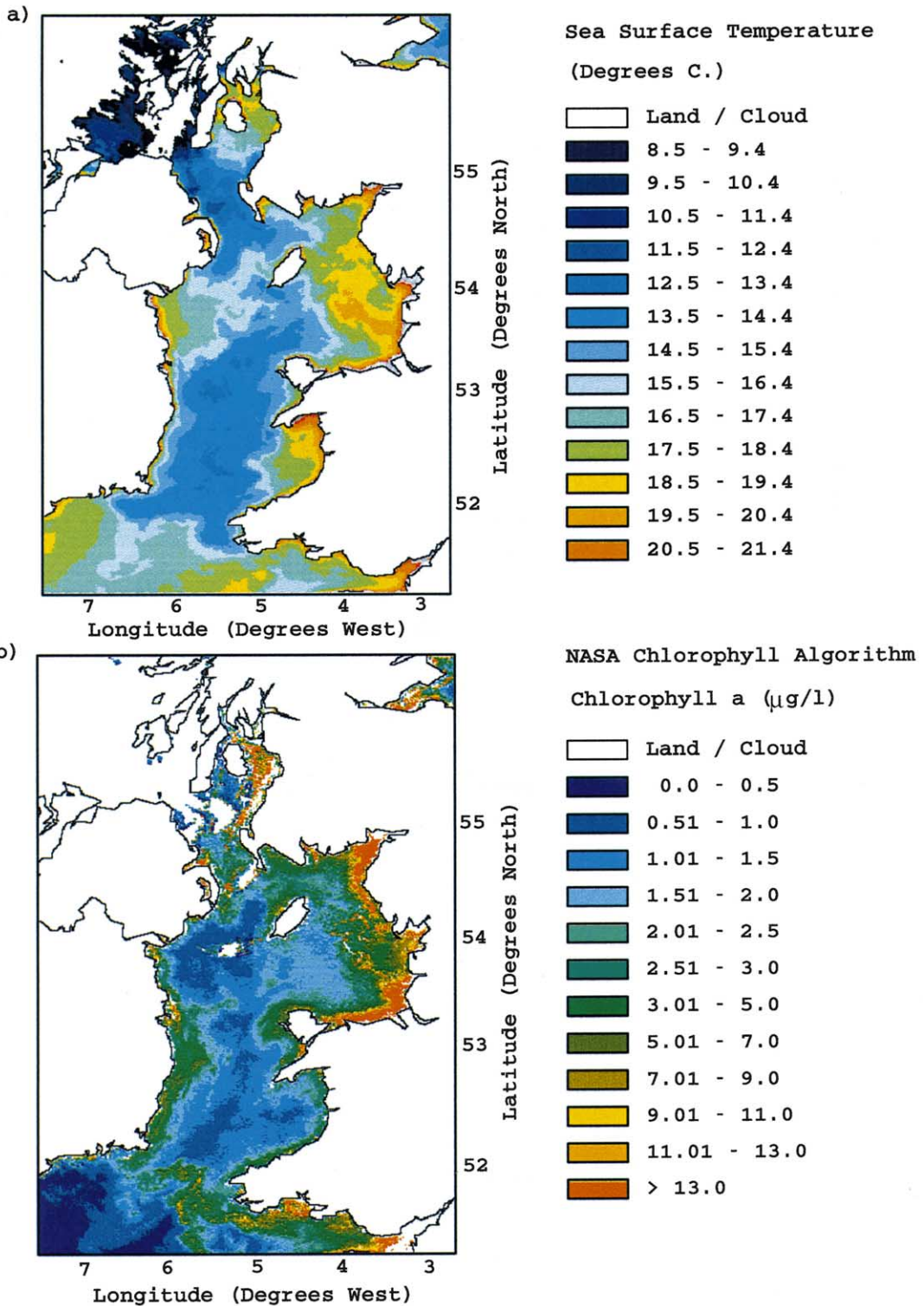
where μ_m is the maximum growth rate, k_Q is the subsistence cell nutrient quota, and Q is the cell nutrient quota (= the ratio of algal cell internal nutrient concentration to chlorophyll biomass). Nitrogen is less available for growth, and is therefore more limiting, than is phosphorus (Tett and Droop, 1988), while silicon may be more limiting than nitrogen for diatoms since these build tests with thick silica walls (Brzezinski, 1985).

Growth rate may be light-limited:

$$\mu = q^{\text{chl}} (\alpha \bar{I}_{\text{PAR}} - r^{\text{B}}) \quad (6)$$

where q^{chl} is the algal chlorophyll/carbon ratio, α is the maximum quantum yield, I_{PAR} is the profile of mean photosynthetically available radiation (PAR), and r^{B} is the respiration rate.

Primary production is highly sensitive to spatial



and temporal variations in the ‘mixed layer optical thickness’ which is λh_1 , where λ is a diffuse attenuation coefficient (a measure of turbidity) and h_1 is the thickness of the layer through which algae are transported by vertical turbulence (Tett et al., 1993). In a shallow tidal sea, the main control of λh_1 is tidal flow: tidal bottom stress resuspends sediments (changing λ) and injection of turbulent kinetic energy determines water column layering (changing h_1). Consequently, λh_1 is greater in areas of mixed waters and less in areas of stratified water; λh_1 increases in winter and decreases in summer. Algae can grow only when they receive sufficient light: this can be expressed in terms of an ‘optical critical depth’, $m_2 I_0 / I_c$, which is the ratio of PAR at the sea surface, I_0 , to the mean PAR required for net growth (exceeding losses by grazing and sinking), I_c , limited by m_2 which allows for hyper-exponential downward decay of irradiance near the sea surface (Tett, 1990). Significant algal production occurs when $m_2 I_0 / I_c > \lambda h_1$, so that production is limited by enhanced turbidity in autumn and winter. Using an annual time series of solar irradiance, wind speed, and air dewpoint temperature to drive a point model (Huthnance et al., 1993), Tett et al. (1993) show that, for the southern North Sea, frontal regions display a few months of weak stratification and mixed regions remain mixed except during some neap tides. In stratified regions the productive season roughly coincides with stratification, but it is shorter in mixed regions and longer in frontal regions (Fig. 4). This reflects the direct impact of h/u^3 on summer stratification and the indirect impact of turbidity on production: mixed waters are optically deeper than stratified waters because of increased turbidity as well as stronger mixing in summer.

Advocates of the nutrient-limiting model for algal growth (e.g. Dugdale, 1967; Fasham et al., 1990) can show that it gives a reasonable approximation for algal populations in the sea (especially when various loss processes and physical trans-

ports are taken into account). In other approaches, the maximum growth rate μ_m in Eq. 5 is made dependent on light or it is made dependent on light or nutrients depending on which is most limiting at a given time (Droop, 1983; Tett and Droop, 1988).

Winter levels of nitrate, silicate, and phosphate are 8.6, 3.7, and 0.6 $\mu\text{mol l}^{-1}$, respectively, in the central North Sea (Brockmann et al., 1988). Primary production is enhanced by nutrient supply via rivers and this is important when λh_1 is small (even in eutrophic waters). In coastal waters, higher numbers of blooms coincide with periods of snowmelt, rainfall and river runoff, causing possibly greater stability of the water column and certainly increased nutrient inputs (Gieskes and Schuab, 1990; Reid et al., 1990). Nutrient levels are supplemented by anthropogenic inputs in such waters – winter nitrogen levels may be an order of magnitude higher than oceanic levels, possibly causing eutrophication (Lancelot et al., 1987; Brockmann et al., 1988). In stratified waters, nutrients are available in the BML but ultimately limited in the SML as nutrients are consumed but not replaced from below. Nitrate levels below 0.5 $\mu\text{mol l}^{-1}$ are generally held to be insufficient to support net phytoplankton growth (Peeters and Peperzak, 1990) so production ceases in the SML once this threshold has been reached. It is likely that the spring bloom in mixed waters ends because of silicon, rather than nitrogen, depletion which curtails diatom growth (Peeters et al., 1991); a large bloom of the flagellate *Phaeocystis* follows the diatom bloom in mixed waters of the NW European shelf where high nitrogen and phosphorus concentrations endure (Lancelot et al., 1990).

It follows that there are large spatial and temporal variations of algal productivity in tidal shelf seas. Plankton are constrained in mixed regions (high nutrients but high turbidity, so growth rate according to Eq. 6); productivity seems to be greater in areas where λh_1 is small rather

Fig. 3. (a) AVHRR image showing sea surface temperature and tidal mixing fronts of the Irish Sea for 12 July 1999. (b) SeaWiFS image showing chlorophyll concentrations determined from the NASA chlorophyll algorithm for 12 July 1999; note in particular chlorophyll high at the Celtic Sea front at the southern boundary (from Binding, 1999, courtesy of NERC RSDAS, Plymouth).

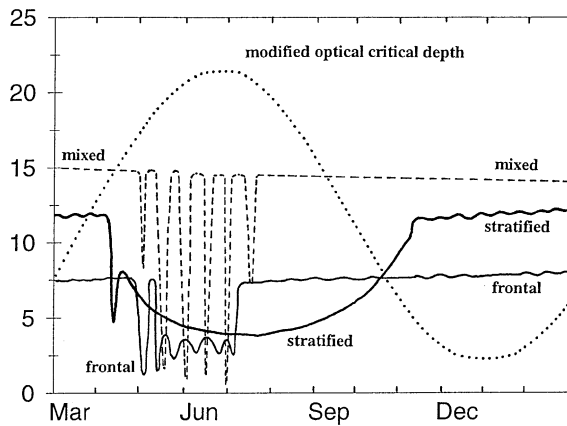


Fig. 4. Model simulation of optical thickness of the SML in mixed, frontal, and stratified regions compared to modified optical critical depth. See text for explanation of symbols. From Tett et al. (1993).

than where nutrients are greatest (Tett et al., 1993). Plankton are also constrained in the BML of stratified regions (high nutrients but low light, growth rate by Eq. 6). In the SML of stratified regions, productivity is initially high in spring (high nutrients and good light, so growth rate by Eq. 5) but rapidly diminishes during the summer as nutrients are consumed and not replaced. Grazing by herbivorous zooplankton (term gX in Eq. 4) and settling of algal cells (associated with SPM aggregates, see below) to the seabed reduce the standing stock in the SML. Subsequent predation of zooplankton reduces grazing pressure and increase of nutrients by wind-driven mixing stimulates a secondary autumn algal bloom in the SML. As the spring bloom wanes, enhanced chlorophyll concentrations (up to 100 mg m^{-3}) and maximum phytoplankton biomass are generally seen at the thermocline in seasonally stratified regions (e.g. Anderson, 1969; Cullen and Eppley, 1981; Holligan et al., 1983). Turbulent tidal mixing generates new production by periodically supplying nitrate from the BML; but turbulent tidal mixing also entrains algae into the BML where they are lost from the productive regions of the water column (Sharples et al., 2001).

It has been observed that in summer 30–80% of the total algal production in the euphotic zone takes place in the thermocline (Fransz and

Gieskes, 1984) and that the greatest production occurs near tidal mixing fronts (Pingree et al., 1975, 1978; Savidge, 1976; Simpson et al., 1979; Holligan, 1981; Holligan et al., 1983; Creutzberg, 1985; Loder and Platt, 1985; Le Fèvre, 1986; van Haren and Joordans, 1990; Tett et al., 1993; Tett and Walne, 1995). This could be a passive response to the convergent flows which have been observed at fronts but it is more likely to result from in situ growth of plankton. This is where the combination of light and nutrients is optimal (Dufour and Stretta, 1973): nutrient renewal during the summer, due to mixing by tide and wind, and surface stabilisation and reduction of λ_{h1} during fairweather and neap tides. Thus a chlorophyll maximum is observed at the thermocline and at fronts (Fig. 3B). The optimal conditions for rapid algal growth are at the front, as lateral mixing across the front is greater than vertical mixing across the thermocline (Garrett and Loder, 1981; Tett, 1981; Tett et al., 1986).

Baroclinic eddies, which develop along fronts and which inject parcels of nutrient-rich water from the mixed to stratified sides of the front, are probably an important mechanism for enhanced productivity (Pingree et al., 1978; Loder and Platt, 1985). But there is also mixing at fronts due to temporal variations over a lunar cycle. If the current speed increases by a factor of 2 from neaps to springs, the stratification parameter changes by a factor of 8; the frontal zone shifts towards deeper water. It advances towards shallower water on neaps. Since nutrient concentrations are higher in the mixed region than in the SML of the stratified region, nutrient concentrations are greater in the nascent SML near the frontal zone on neaps than the ambient levels in the rest of the SML. It is feasible that nutrients from the BML are entrained during the lunar cycle of changing tidal range. Pingree et al. (1977), Le Fèvre et al. (1983), Morin (1984) and Morin et al. (1985) have reported fortnightly pulses in phytoplankton growth at fronts. Loder and Platt (1985) estimated that 80% of the nutrients in the frontal zone could be due to a combination of this spring-neap adjustment and vertical mixing.

There are temporal and spatial variations in

algal species as well as in algal productivity (Holligan, 1981; Peeters et al., 1991). Schematics for the succession of algal types incorporate light and a single limiting nutrient in relation to vertical mixing (Margelef, 1978). Diatoms have a high ratio of photosynthetic efficiency to respiration (Tett, 1990) and are best able to grow in the turbid waters characteristic of the spring bloom and of vertically mixed regions. Dinoflagellates dominate the smaller summertime standing stock particularly where vertical mixing is low but diatoms survive below the thermocline and in the BML of stratified waters. Dinoflagellates, whose motility enables them to maintain an optimal position in the water column, are most abundant in frontal zones; as the thermocline deepens, they are replaced by flagellates (Reid et al., 1990). Coccoliths are present in the SML but not in the BML; they are not abundant in mixed waters.

2.5. Sediment response: impact on SPM

The cohesive organo-mineral particles that constitute SPM are more complex and variable than the non-cohesive mineral particles of shelf sands. The particles are not necessarily small, since they form aggregates which commonly exceed medium sand size, but they are of low density because of their high organic and water contents. Such particles have extended residence times in the water column. There is compelling evidence that SPM properties are linked to pelagic algal activity. The organic matrix of many aggregates is made up of high molecular weight, exopolymer matter (Alldredge et al., 1986) which plays an important role in agglutination of diatom blooms (Alldredge et al., 1993; Kiorboe and Hansen, 1993; Riebel-sell, 1991). The nutrient status of algal cells has been shown to affect their cohesiveness (Kiorboe et al., 1990) and enhanced post-bloom plankton agglutination has been attributed to an increase in stickiness due to nutrient depletion (Logan and Alldredge, 1989; Smetacek, 1985). Such stickiness has been linked to SPM aggregation with respect to flagellates and dinoflagellates (Passow and Wassman, 1994; Jones et al., 1998) as well as to diatoms. Strong biological mediation of SPM aggregation during the late stages of a flagellate

bloom, resulting in increased particle size and settling flux of SPM to the seabed, has been measured (Jago and Jones, 2000; Jago et al., 2002b). Links between algal activity and particulate settling flux suggest that the greatest flux per unit area of particulate matter to the seabed should occur in regions of greatest algal production per unit volume, i.e. in frontal zones.

In the aftermath of blooms, the enhanced settling flux gives rise to low density, organic-rich, benthic fluff on the bed (Jago et al., 1993). Benthic fluff has a carbon content which is an order of magnitude greater than that of surficial seabed sediment and SPM in long-term suspension (Bale and Morris, 1998). Quarter-diurnal signals in TSM (total suspended sediment matter concentration) due to resuspension of fluff, allied to semi-diurnal signals in TSM caused by longitudinal concentration gradients of SPM in long-term suspension, give rise to a distinctive ‘twin peak’ signature of TSM (Fig. 5; Weeks et al., 1992; Jago et al., 1993; Jago and Jones, 1998). This can be expressed as:

$$S(t) = S_0 + \frac{dS}{dx} \int_0^t U_x dt + k |U_x| \quad (7)$$

where S is TSM, S_0 is the concentration of SPM in long-term suspension at time $t=0$, U_x is the rectilinear current velocity and k is a pick-up function which incorporates entrainment from the seabed and vertical diffusion through the water column (Jago and Jones, 1998). The ‘twin peak’ signature is diagnostic of tidal resuspension particularly where there is a regional gradient in SPM concentration. Jones et al. (1996) and Jago and Jones (1998) described a 1-D numerical model which incorporates horizontal advection of a particulate concentration gradient and vertical diffusion and which reproduces the ‘twin peak’ signature:

$$\frac{\partial S}{\partial t} = -U_x \frac{\partial S}{\partial x} + \frac{\partial}{\partial z} \left(K_z \frac{\partial S}{\partial z} \right) + W_s \frac{\partial S}{\partial z} \quad (8)$$

where K_z is vertical eddy diffusivity and W_s is particle settling velocity. The bottom boundary condition includes fluxes due to resuspension and deposition:

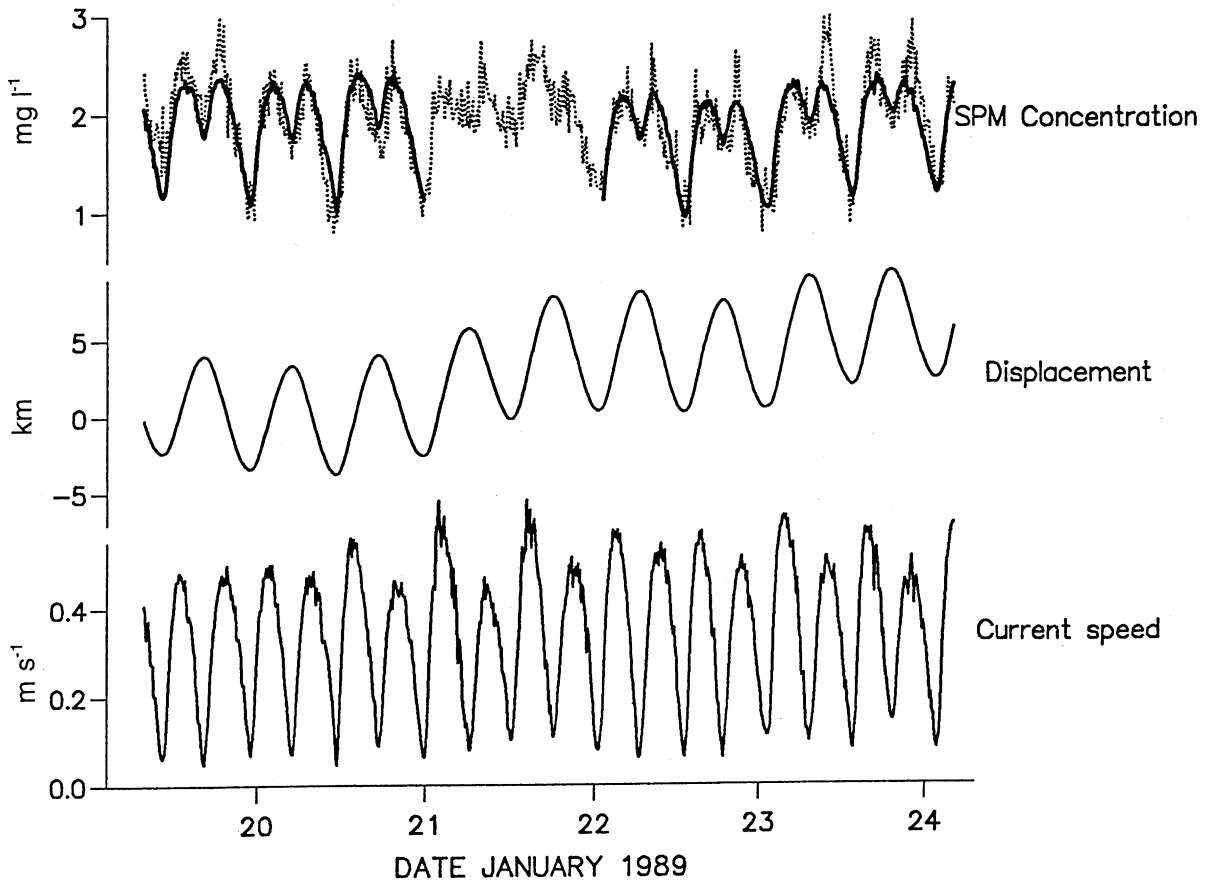


Fig. 5. Twin peak signature of SPM concentration due to combined resuspension of fluff and advection of SPM. Model simulation using Eq. 7. Twin peaks disappear between 21 and 22 January because wind waves prevented fluff deposition. Note that fluff is present even in winter. From Jago and Jones (1998).

$$-W_s S - K_z \frac{\partial S}{\partial z} = E - W_d S \quad (9)$$

E and $W_d S$ are the entrainment and deposition rates, respectively (it is assumed that $W_s = W_d$). E is a function of the bed shear stress τ_b ($E = k\tau_b^n$, where k and n are adjustable parameters).

Genesis and fate of benthic fluff depend on settling flux (depends on W_s and concentration) and resuspension flux (depends on τ and K_z). Maximum fluff sedimentation requires that settling flux exceeds resuspension flux. In mixed waters with a sandy seabed, tidal currents can resuspend sand very close to the bed; but most of the water column is characterised by resuspended fluff. The fluff settles rapidly during slack water, since its

mean settling velocity is some three orders of magnitude faster than that of SPM in long-term suspension, and temporarily resides in the troughs of ripple marks (Jago and Jones, 1998). Peak TSM precedes peak current velocity since the supply of resuspendable fluff is exhausted before maximum flow (Jago and Jones, 1998). Superimposed wind waves can suppress settling of fluff at slack water but cannot increase resuspended concentrations since the fluff on the bed is of finite supply; this is evident in the example shown in Fig. 5 where, during a 24 h period of wind waves, the 'twin peak' signal disappears but background TSM remains the same. Fluff is still present on the seabed in winter. Frequent fluff resuspension in mixed regions accelerates water column remi-

neralisation (i.e. transfer from particulate to dissolved phases) and limits transfer of organic matter to the seabed sediment.

In stratified waters, tidal stirring is weak and the mud content of seabed sediments increases with water depth. The source and supply of organic detritus varies with water depth. The SML progressively loses faster settling SPM without replenishment from below because of low mixing across the thermocline. There are fewer algae in the SML as nutrients decline. So in the aftermath of a bloom, the SML clarifies and more light penetrates the BML. There is good evidence that low tke dissipation at the base of the thermocline is associated with SPM aggregation (Jago and Jones, 2000; Jago et al., 2001; Jago et al., 2002b) which in turn increases the settling flux of SPM into the BML. There is resuspension of a finite supply of benthic fluff on spring tides and during storms (Jago et al., 2002a). Pelagic remineralisation should increase with water depth due to the longer settling times of SPM so the supply of organic matter to the bed is probably reduced progressively in deeper water.

In frontal zones, the supply of biogenic SPM increases because of enhanced productivity. Biologically mediated aggregation in the SML during the spring bloom (Jago and Jones, 2000) and physical aggregation at the base of the thermocline during the summer (Jago et al., 2001) provide optimum conditions for rapid settling flux to the seabed. Tidal currents cannot resuspend bed sediment per se unless augmented by storms, but spring tidal currents can resuspend benthic fluff (Fig. 6; Jones et al., 1998). Light penetrates both SML and BML so that plankton can remain viable in the BML even at the seabed. Spring tide resuspension and deposition of fluff comprising living diatoms and dinoflagellates has been observed in the BML in a frontal zone in the southern North Sea (Jago et al., 1993; Jones et al., 1998). In the example shown in Fig. 6, fluff resuspension begins on a critical tide during the lunar cycle; nevertheless, despite increasing peak current speeds, SSC decreases as fluff deposition exceeds supply of plankton-rich SPM from the SML. Settling flux probably exceeds resuspension flux for much of the spring

and summer so that organic matter can be transferred to the seabed sediment. The fluff layer is therefore most persistent in frontal zones.

From the foregoing, we propose that the accumulation of organic matter per unit area of the seabed is high in frontal zones and decreases towards both mixed and stratified regions.

3. Biogeochemical flux models

Coupled models have been developed which incorporate some or all of the foregoing principal elements (e.g. Sharples and Tett, 1994; Luytens et al., 1999; Sharples, 1999; Smith and Tett, 2000). Thus SEDBIOL (Smith and Tett, 2000) is a 1-D depth-resolving model which couples water column dynamics, algal production, and SPM/benthic fluff settling/resuspension (Fig. 7). The model provides seasonally varying turbulent diffusivities which drive nutrient cycles and interactions with phytoplankton (grazed by zooplankton) and SPM, including settling flux and deposition of benthic fluff. The model predicts annual net primary productivity and carbon fluxes to the seabed. COHERENS (Luytens et al., 1999) is an advanced 3-D coupled model which uses turbulence closure schemes to provide the dynamical framework for plankton cycling and SPM settling and exchange with the benthic fluff layer. It simulates plankton dynamics, settling flux and fluff deposition rate and resolves mesoscale and seasonal scale processes. Such models provide conceptual insights and numerical solution of biogeochemical fluxes in stratified and mixed regions of tide-driven shelf seas. Significantly, the models provide numerical relationships between factors such as water temperature and seabed anoxia (which are potentially recorded in the sediments by biological and geochemical proxies) and governing variables such as water depth, mixing, and turbulence.

4. Biogeochemical settling flux: impacts on seabed sediments

Stratified and mixed waters are differentiated

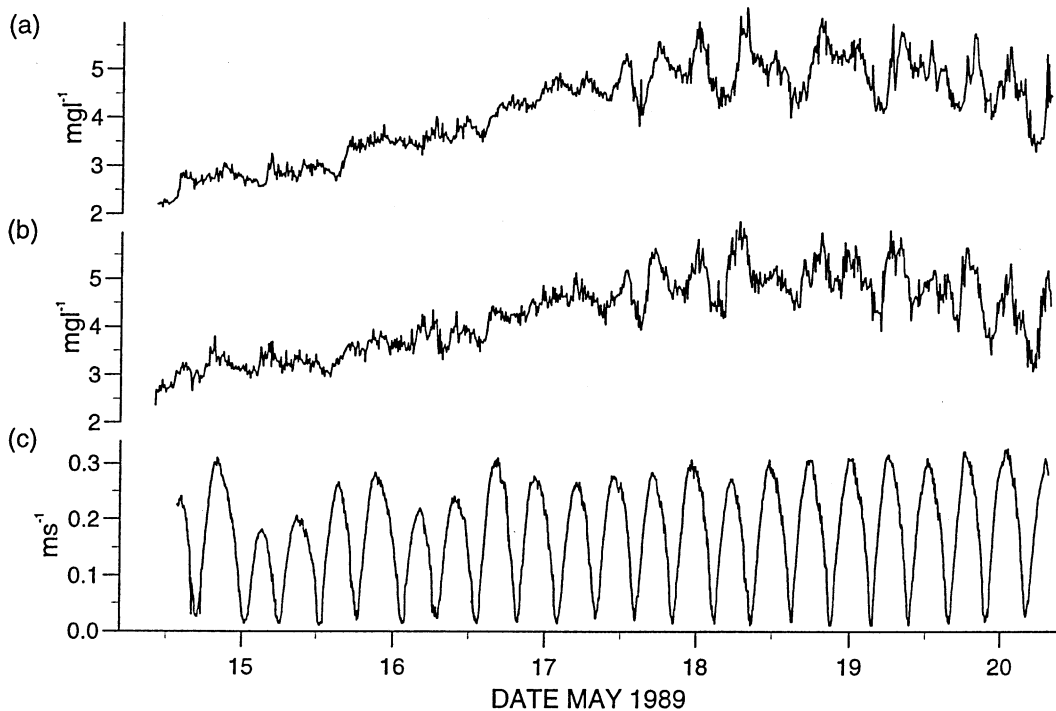


Fig. 6. SSC time series at (a) 10 m above bed and (b) 5 m above bed with (c) current speed 11 m above bed at a site on the stratified side of the southern North Sea front (47 m water depth). Twin peak signature indicative of fluff resuspension begins after 17 May. SSC then declines, despite increasing peak currents, because of net deposition of fluff (settling flux > resuspension flux). From Jones et al. (1998).

by seabed shear stress which constitutes a first-order control of seabed sediment type. There are distinct sedimentary facies associated with each region, with strong lateral gradients of sediment texture across frontal regions (Creutzberg and Postma, 1979; Creutzberg et al., 1984; Su and Wang, 1986; Su et al., 1990). Silt is constantly resuspended in the mixed, sandy regions but is deposited in the stratified, muddy regions. There are consequent gradients across fronts in benthos biomass with larger values in stratified regions (Creutzberg et al., 1984; Kunitzer et al., 1992). In addition, since water column structure controls settling flux, there should be consequent biogeochemical impacts on the seabed. The principal contributors here are particulate organic carbon (POC) settling flux and associated benthic oxygen consumption; together they mediate the biogeochemical climate at the seabed.

4.1. POC settling flux

A working hypothesis is that the seabed under frontal zones is characterised by strong lateral gradients in POC. First, the front is the region of maximum production of plankton per unit volume of seawater; second, the front is the region of maximum settling flux of POC (as aggregated SPM during blooms) per unit area of seabed. There are limited data to test the hypothesis but Creutzberg (1985) showed that POC and mud were significantly enhanced in the seabed sediments beneath a tidal mixing front in the southern North Sea (30–40 m water depth, close to the area of net fluff deposition reported by Jones et al., 1998).

4.2. Benthic oxygen consumption

Benthic fluff controls redox conditions at the seabed and determines whether biogeochemical

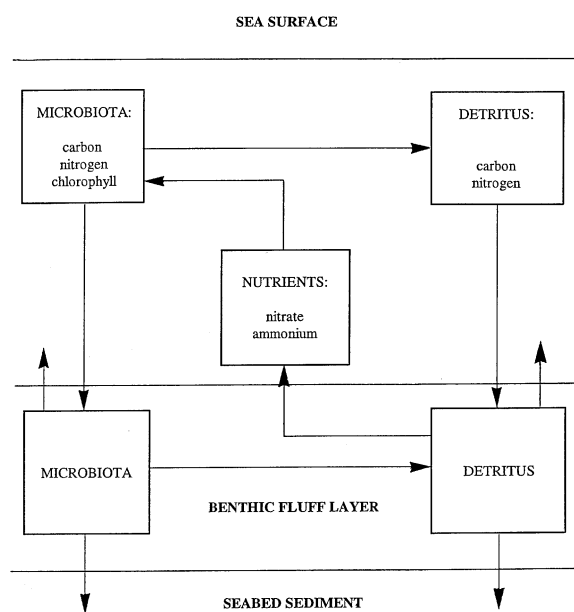


Fig. 7. Schematic representation of the numerical model SEDBIOL showing biophysical couplings in the water column and benthic fluff layer. From Smith and Tett (2000).

exchanges are oxic or anoxic. Temporary seabed anoxia due to tidal fluff deposition after a bloom occurs even in mixed regions but more ore-persistent anoxia is observed in stratified waters near fronts where settling, living aggregates may be trapped below the thermocline, which gives rise to anaerobic decay of cells and oxygen depletion near the seabed (Jago et al., 1993; Millward et al., 1998). This implies that there are significant cross-frontal gradients in benthic oxygen consumption at particular times of the year; limited data from Boon et al. (1998) provide support since they showed significantly higher benthic oxygen demand in the Frisian frontal, compared with the mixed zone of the southern North Sea, which appeared to be related to local, rather than distal, primary production.

5. Discussion

5.1. Mapping the biogeochemical flux peak onto seabed sediments

The biogeochemical flux signal at the seabed,

comprising planktonic POC production, POC flux, and oxygen demand, should be greatest in the frontal zone. But a tidal shelf mixing front is a zone some 5 km wide, rather than a sharp boundary. Even with a static mean sea level and unchanging climate, tidal mixing fronts vary in position. The biogeochemical flux peak is restricted to a short period of the year. Biogeochemical settling flux should generally coincide with the onset of stratification and there is good evidence (from organic loadings) that this is sufficient time to impart a diagnostic signal in the seabed sediments but data are limited at present. Fronts advect tens of kilometers during a summer and there is mixing of water and nutrients across fronts, so the response of the microbiological record in the sediments to the biogeochemical flux signal is likely to be smeared. Baroclinic instabilities that create eddies may also entrain and advect benthic fluff and microbiota away from frontal regions. Consideration of seabed shear stresses and turbulent energy dissipation suggests that advected material is likely to be redeposited in stratified, rather than mixed, regions so the peak signal in the sediments may be laterally displaced towards stratified regions with respect to the biogeochemical flux peak (Fig. 8). Winter storms will also advect material into deeper waters away from the front.

5.2. Diagnostic signatures in shelf sediments

From the geological perspective, the key consideration is whether this biogeochemical peak generates a diagnostic signal in the sediment record. Diagnostic proxies may be expected in the biota rather than in the sediments per se, especially those pelagic and benthic microorganisms that are sensitive to the pelagic and benthic biogeochemical regimes. The biota respond to changes in supply of organic matter and most of the particulate organic matter available to the benthos in the southern North Sea consists of decaying algal matter (Boon and Duineveld, 1996). In addition, it is likely that benthic microorganisms are sensitive to the strong temperature gradients across frontal zones. Such biological signals are particularly pertinent to Quaternary shelf

regimes as most Quaternary species are living today and their response to the biogeochemical regime can be tested.

5.2.1. Signals in the microplankton

Within the microplankton, dinoflagellates probably have the greatest diagnostic potential. Dinoflagellates regulate their position in the euphotic zone to maximise photosynthesis (Prezelin, 1987) and their motility is particularly suited to regions of strong nutrient gradients, such as the thermocline and fronts, where swimming for nutrients is biologically worthwhile (Dale, 1996). Consequently dinoflagellates occur in greatest numbers at fronts. Autotrophic rather than heterotrophic species respond directly to nutrients and light and it is known that particular species are associated with stratified, frontal, and mixed waters (Taylor, 1987). Other contributors to the microplankton are not typical of frontal zones: diatoms are more passive and thrive where physical processes bring nutrients to them (e.g. mixed waters, and stratified waters prior to nutrient depletion). Coccoliths have motile stages but do not concentrate in frontal zones.

Some dinoflagellates form resting cysts linked to their sexual cycle. Such species may be more abundant in shelf seas than in the open ocean (Dodge and Harland, 1991; Dodge, 1994). In

shelf seas, the formation of cysts appears to occur as nutrients are depleted in surface waters; nascent cysts settle rapidly to the seabed (Jago et al., 1993) where they can await a resuspension event which might take them into nutrient-rich waters. The preservation potential of dinoflagellate cysts is well documented since they fossilise under anoxic conditions (Head, 1996); they are known from the Triassic onwards. Most studies of dinocysts as environmental indicators have been in relation to deep sea sediments (e.g. Margelef, 1978; Marret, 1994; Marret and de Vernal, 1997; Zooneveld, 1997; de Vernal et al., 1998), where it has been shown that cysts are diagnostic of temperature, salinity and productivity (Dale, 1996). It has been shown that cysts may be concentrated at estuarine frontal convergence zones (Tyler et al., 1982). Marret and Scourse (2002) have shown that stratified, frontal and mixed zone dinocyst assemblages can be distinguished across the Celtic Sea tidal mixing front: mixed water sediments are dominated by *Lingulodinium machaerophorum* whereas frontal and stratified water sediments are dominated by *Spiniferites ramosus*; frontal and stratified water sediments can be distinguished by the frequencies of less abundant species (*Bitectatodinium tepikiense* and *Spiniferites elongatus*). These differences relate to sea surface temperature and a stratification index

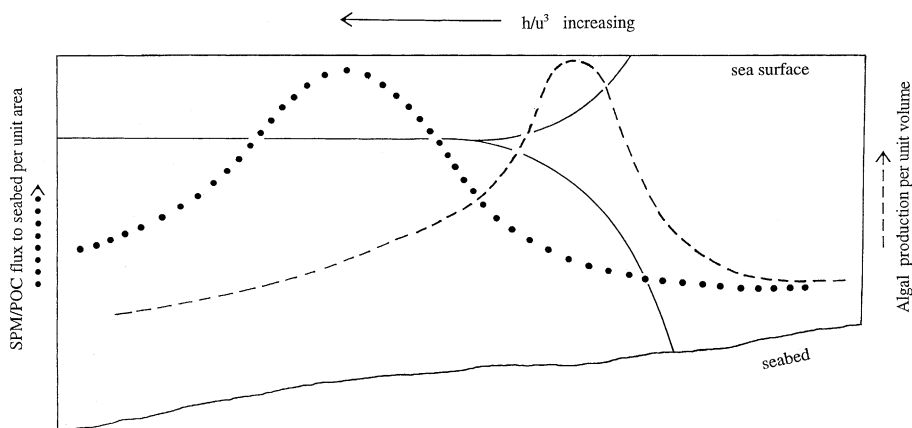


Fig. 8. Conceptual sections of depth-integrated algal production (dashed line) and POC flux to the seabed as benthic fluff (dotted line) across mixed, frontal, and stratified regimes. Vertical scales are qualitative. Note that the POC flux (benthic fluff) peak is displaced towards the stratified zone.

(the surface to bottom temperature differential). Cysts were sparse under the front and more abundant in deeper stratified waters adjacent to the front, implying advection away from the front (see above).

Many planktonic calanoid copepod species spend part of their life cycle as benthic resting eggs. Resting stages may be used as long-term survival strategy. Diapausing eggs are frequently extremely abundant (e.g. Lindley, 1990). Furthermore, the spatial and temporal distributions of copepod nauplii in, for example, the Irish Sea are associated with variations of water column dynamics and may depend on stratification and mixing, particularly during the spring plankton bloom (e.g. Burkart et al., 1995). Thus the copepod *Calanus* establishes regularly in the stratified regions of the Irish Sea but is rare in mixed regions where it is related to incursions of water from the Celtic Sea or North Atlantic. Planktonic copepods are therefore sensitive to water column dynamics and their resting eggs may provide a useful proxy for water column dynamics. However, the preservation potential of copepod resting eggs is unknown although it has been shown that they survive in sediments for at least decades (Kajastajisto, 1996); the potential of zooplankton resting eggs for shelf sea palaeoceanography has yet to be explored.

5.2.2. Signals in the microbenthos

There has been progress with microbenthic proxies of water column structure. Foraminifera are famously sensitive to environment. Scott et al. (2003) used living and dead benthic forams in the Celtic Sea to analyse the interrelationships of foram and shelf water isotopic signatures. Four foram assemblages were identified, three of them constrained by water column structure: frontal, mixed, stratified. The foram distributions were extremely sensitive to frontal dynamics and indicated both surface convergence and cross-frontal mixing. As with dinocysts, the sediments of the frontal region were impoverished in living forams, suggesting advection of tests into deeper water. The same assemblages in the stratigraphic record were used to reconstruct temporal changes of water column structure.

The stable isotope signal in the microbenthos is diagnostic of biogeochemical regime. A Holocene sequence from the Celtic Sea shows a change from sediments with an epifaunal foram assemblage characteristic of a well-stirred, oxygenated seabed to sediments with an infaunal foram assemblage characteristic of a quiescent, oxygen-limited seabed (Austin and Scourse, 1997). The stable isotope signals from *Ammonium batavus*, a shallow infaunal species, and *Quinqueloculina seminulum*, an epifaunal species free-living on plants or sediment, were diagnostic of a temporal progression of biogeochemical regimes. A positive trend in the stable O isotopic record from these species was interpreted as a 4–5 EC cooling in shelf bottom waters; a negative trend in the stable C isotopic record from the same species was consistent with increasing oxygen consumption due to seabed anoxia (Fig. 9). The change from tidally stirred mixed waters to a productive frontal zone and onset of seasonal stratification occurred between 8990 and 8440 cal. yr BP (Scott et al., 2003; Scourse et al., 2002).

5.3. Application and constraints of the new paradigm

The new paradigm offers a diagnostic tool for the interpretation of tidal shelf sedimentary sequences. Sediment proxies can signal changes of biogeochemical regime due to changes of dynamical regime: switches between mixed, frontal, and stratified conditions due to changes of sea level and/or climatic conditions. The sensitive proxies will be best preserved in the silty sands to silty muds which constitute the cohesive sediment blanket that covers much of the shelves. Since models such as COHERENS provide a numerical simulation of biogeochemical fluxes underpinned by sophisticated treatment of shelf dynamics, they offer the potential for quantitative interpretation of sediment proxies in the stratigraphic record. Combination of models and sediment proxies, calibrated by training sets, can provide information on water column structure, surface heating, mixing, and water depth. Used in concert with the sediment dynamics paradigm (for the shallow water sands poor in biogeochemical information),

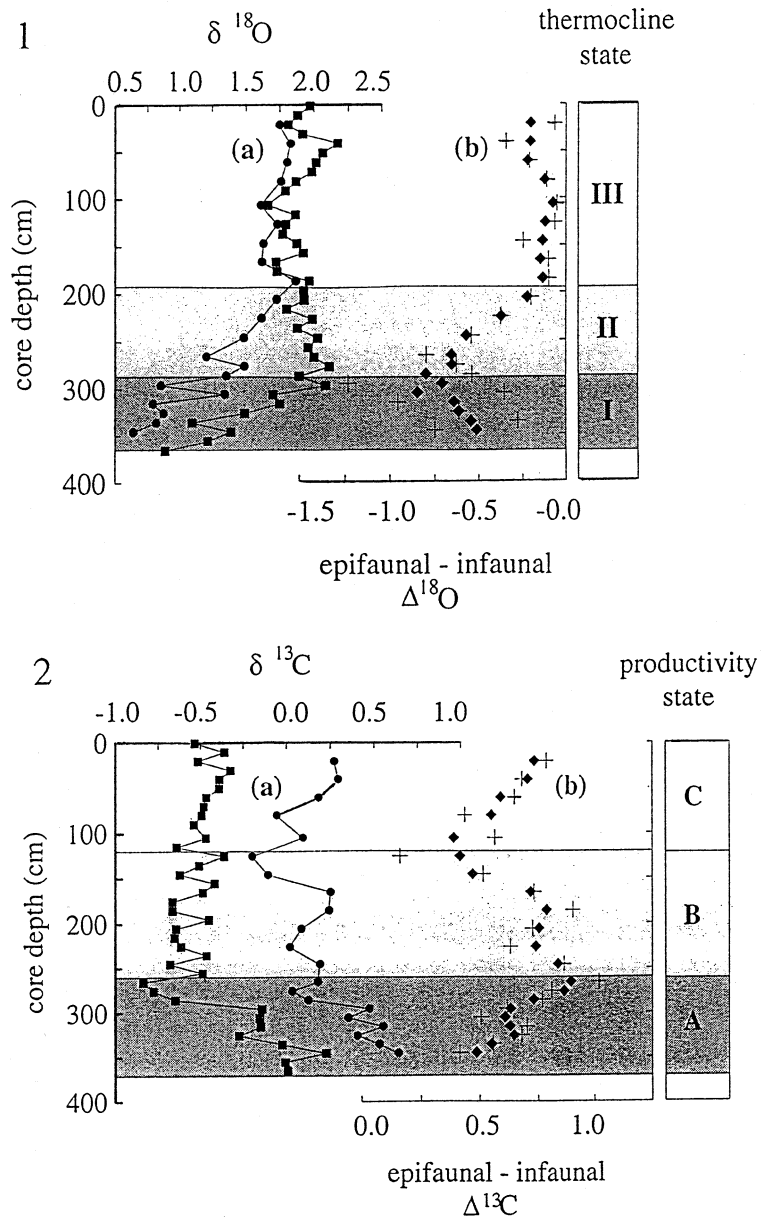


Fig. 9. (1) Stable oxygen isotope record showing: (a) *Ammonia batavus* (squares) and *Quinqueloculina seminulum* (circles); (b) epifaunal (*Q. seminulum*)–infaunal (*A. batavus*) difference (– = three-point average). Thermocline states I–III represent progressive development of seasonal stratification with deepening water. (2) Stable carbon isotope record showing: (a) Same species as 1; (b) epifaunal–infaunal difference as in 1. Passage of a frontal regime shown by strong signal between productivity states A and B. Samples from core in central Celtic Sea. From Austin and Scourse (1997).

the new paradigm provides a novel framework for reconstruction of past shelf sea regimes (Fig. 10).

There are constraints in applying the paradigm. Uncertainties in the biogeochemical models can

produce significant error in numerical calculations of pelagic processes and benthic fluxes. For example, it is unclear whether pelagic ecosystems are best modelled as self-stabilising networks (Tett

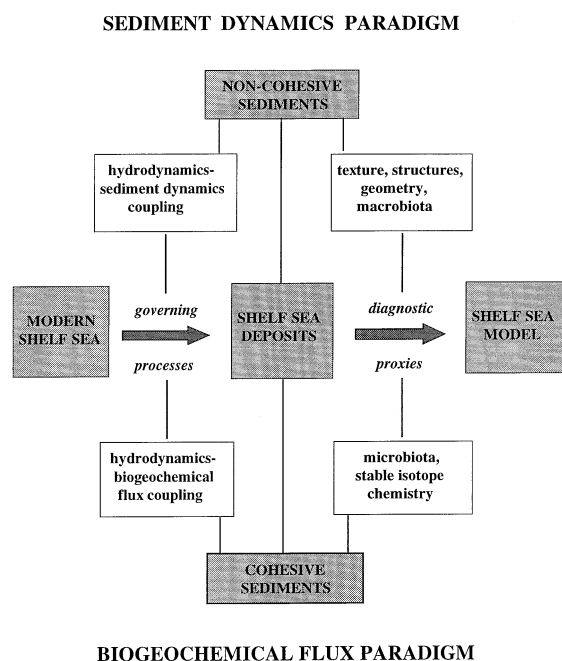


Fig. 10. Schematics of paradigms for diagnostic interpretation of shelf sea deposits.

and Wilson, 2000) or as chaotic systems (Hastings and Powell, 1991). There is limited information on the dynamics of benthic fluff and associated biogenic particles in the benthic boundary layer and on benthic/pelagic coupling with respect to both. The process pathways through which proxy gradients in sediments develop in response to biogeochemical flux gradients have not been clearly discerned by observation or modelling. Quantitative relationships between POC settling flux, benthic fluff deposition, benthic oxygen consumption, and microbiota response have yet to be established.

How the biogeochemical flux peak maps onto seabed sediments and the sediment record remains conjectural. Spring and summer spatial gradients in the water column may not directly map onto time-averaged lateral gradients in the sediment record. The magnitude of this spatial mismatch will depend on local dynamics (e.g. potential energy gradients across the front, mean flow along the front, direction of tidal streams relative to orientation of the front, tidal advection of the frontal zone, intensity of storm stirring of the

seabed) and will be very variable. The mismatch may be extreme in areas like the Celtic Sea where non-tidal frontal advection is large (Simpson and Bowers, 1981), the tidal streams are orthogonal to the front, and storm stirring is significant. Given all of the foregoing uncertainties, modelling developments are needed that conceptually and numerically rationalise the very different time scales of biogeochemical processes and sequence stratigraphy (cf. the STRATOFORM project: Nittrouer, 1999).

5.4. Modern shelves as geological models

The biogeochemical flux paradigm presented here is concerned with shelf systems in which tidal forcing is the defining control of the hydraulic regime. There has been some debate about the significance of tides in ancient shelf seas following the seminal contribution by Shaw (1964), who argued that tidal energy would be so attenuated in shallow epicontinental seas that tidal currents would be minimal and the hydrodynamic regime would be essentially wind-driven (see also Dott and Batten, 1971; Heckel, 1972; Mazzullo and Friedman, 1975). Nevertheless, we have modern epicontinental shelves around NW Europe, SE Asia and elsewhere which are dominated by tides, and tidal shelves from the past have been identified (e.g. Narayan, 1971; Swett and Smit, 1972; Banks, 1973; Anderton, 1976; Hereford, 1977; Klein and Ryer, 1978; Boyles and Scott, 1982). This debate has been illuminated by numerical models of shelf tides applied retrospectively. Thus Austin (1991) and Scourse and Austin (1995) used the M_2 constituent of the ocean tide to model the tidal regime of parts of the NW European shelf at successive stages of the Holocene transgression. Their results suggest that a lowering of water depth by up to 25 m in this area would increase the total rate of frictional dissipation by up to 20%. Seabed shear stresses would significantly decrease in some parts of the shelf, but would significantly increase in others. Tidal mixing fronts would migrate. Clearly the geometry, as well as the bathymetry, of the tidal basin controls the tidal regime so that Shaw's (1964) hypothesis is too simplistic and is certainly

invalid for epicontinental seas constricted by adjacent landmasses. The conclusion must be that except in very shallow, unconfined areas, ancient epicontinental shelves would have experienced tidal regimes with tidal forcing being the primary determinant of water column dynamics. Tidal mixing fronts would have been ubiquitous features of such regimes.

Much of the work on contemporary shelf sea sediments has been concerned with shelves starved of modern sediments. The much-documented sediments of the Middle Atlantic Bight and North Sea represent basal transgressive sand sheets rather than accumulating sedimentary prisms, and they provide limited insights on shelves as sediment depocentres. This provokes a perception that modern shelves are surfaces of sediment transport rather than of net sediment deposition (e.g. Curry, 1989), so that modern shelves make poor models of ancient shelf sedimentation. However, less-studied shelves in more favourable tectonic and climatic settings do receive substantial quantities of fine sediments. Even the sediment-starved NW European shelf, where the principal areas of erosion and deposition of fine sediments have been identified (e.g. Zuo et al., 1989; Eisma and Kalf, 1987; Dyer and Moffat, 1998), has fine sediment depocentres with good stratigraphic records of at least the Holocene (Austin, 1991; Peacock et al., 1992; Scourse and Austin, 1994); comparable depocentres occur elsewhere (e.g. Hald and Vorren, 1987; Hald et al., 1991). The biogeochemical flux paradigm should be applicable to such high resolution records so that, for example, transitions between mixed, frontal and stratified conditions could be identified. In the context of the new paradigm, modern shelves probably make excellent models of shelf sedimentation.

6. Conclusions

A new paradigm for interpretation of ancient tidal shelf sediments is proposed which has particular application for the fine sediments which constitute the extensive cohesive sediment blanket of most tidal shelves. The paradigm exploits re-

cent innovations in observation and modelling of turbulence, primary productivity, and the flux of fine particulate matter to the seabed. New coupled hydrodynamic–ecological models can reproduce the contrasting biogeochemical flux regimes in stratified, frontal and mixed regimes on tidal shelves and simulate the strong gradients of temperature, production, POC flux, and benthic oxygen consumption across frontal regions. Pelagic and benthic microbiota (e.g. dinoflagellate resting cysts, zooplankton diapausing eggs, benthic foraminifera), and associated stable isotopes, in seabed sediments provide proxies for these contrasting biogeochemical flux regimes. They are diagnostic of water column structure, surface heating, turbulent mixing and water depth and can be used to reconstruct hydrodynamic and biogeochemical regimes of tidal shelves. The new paradigm has particular application to Quaternary tide-driven shelves since most Quaternary species of microbiota are still living today and their response to the contemporary biogeochemical flux regime can be tested, but in principle the paradigm has application to ancient shelves. The paradigm is particularly applicable to the fine sediment record which potentially preserves the most complete history of shelf evolution.

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