FATE OF NUTRIENT ENRICHMENT ON CONTINENTAL SHELVES AS INDICATED BY THE C/N CONTENT OF BOTTOM SEDIMENTS

JOHN J. WALSH, EUGENE T. PREMUZIC, AND TERRY E. WHITLEDGE

Brookhaven National Laboratory, Upton, NY 11973

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INTRODUCTION

I met a traveller from an antique land
Who said: "Two vast and trunkless legs of stone
Stand in the desert. Near them, on the sand,
Half sunk, a shattered visage lies, whose frown,
And wrinkled lip, and sneer of cold command,
Tell that its sculptor well those passions read
Which yet survive, stamped on these lifeless things,
The hand that mocked them and the heart that fed;
And on the pedestal these works appear;
'My name is Ozymandias, King of Kings;
Look on my works, ye Mighty, and despair!'
Nothing beside remains, Round the decay
Of that colossal wreck, boundless and bare
The lone and level sands stretch far away."

MASTER

(P.B. Shelley, 1817)

Phytoplankton growth processes are reasonably well-known functions of light, temperature, and nutrients; however, their loss processes are comparatively unknown. For example, the low productivity and nitrate content of most oceanic surface waters (Fig. 1) reflect the slow upward rate of nutrient input across the main thermocline to the euphotic zone. At the coastal boundaries of the ocean, e.g., on the continental shelves, daily fluxes of nitrate supply and ensuing productivity are 1 to 2 orders of magnitude larger (Walsh, 1976) as a result of locally intensified physical processes of upwelling, river runoff, and tidal mixing. ^{15}N estimates of nitrate uptake by phytoplankton suggest that only 10% of the daily nitrogen demand of photosynthesis is met by nitrate in the open ocean (Eppley and Petersen, 1979), whereas nitrate is ~50% of the daily nitrogen source for phytoplankton in coastal waters off Peru (MacIsaac and Dugdale, 1972), New York (Conway and Whitledge, 1979), California (R. Eppley, personal communication), and Alaska (J. Goering, personal communication). Nitrate uptake is considered an estimate of the "new" daily production (Dugdale and Goering, 1967) that is available for export from an ecosystem and is associated with the

Fig. 1. The vertical distribution of nitrate in the open ocean and within the eastern boundary currents.

outbreak of blooms of phytoplankton populations in coastal waters (Yentsch et al., 1977). Annual budgets, in fact, suggest that only 10% of the particulate nitrogen, fixed in the open ocean, sinks out of the euphotic zone, but that as much as 50% of the particulate nitrogen, fixed on the continental shelf, may perhaps be exported to the slope sediments (Walsh, 1980a).

The trajectory and fate of this particulate matter are poorly understood processes in a spatially heterogeneous coastal ocean. Parameterization of appropriate hydrodynamics for a quantitative description of these loss processes must thus await definition of the important biological time and space scales. Since the bottom sands tend to "record" the history of the water column, we have

selected the C/N content of shelf sediments as a possible tracer of 1) sites of nutrient introduction to the shelf by various physical mechanisms, of 2) areas of subsequent downstream utilization by the phytoplankton, and of 3) where loss of particulate matter might occur from the water column. An analysis is made of the C/N patterns of bottom surface sediments in relation to the nitrogen sources from upwelling, river runoff, and tidal mixing on the Peruvian, west African, Amazonian, Gulf of Mexico, eastern U.S., Bering, and North Sea shelves in an initial attempt to proscribe the particle trajectories of organic matter on the continental shelf.

The ratio of carbon to nitrogen (C/N) content in most marine organisms is less than 6, unlike that of land plants which use more carbohydrates for their support structures (Parsons, 1976) and have C/N ratios >15 (Muller, 1977). Detrital particles in the sea also have a C/N ratio greater than 10 as a result of the increased recycling of nitrogen compounds compared to slower decomposition of refractory carbon compounds (Degens, 1970). For example, during blooms of phytoplankton the C/N content of particulate matter in the water column is <6 (Banse, 1974), whereas the C/N content of particulate matter is >10 at low chlorophyll concentrations ($<0.5 \mu\text{g chl a l}^{-1}$), i.e., when mainly detritus is present. Recent sediment trap data, in fact, indicate that higher C/N ratios are found both after the spring bloom on the shelf (Smetacek et al., 1978) and with increasing depth in the open ocean (Honjo, 1980).

Although C/N ratios of >20 can be induced in laboratory cultures of phytoplankton under nitrogen starvation (Caperon and Meyer, 1972), individual phytoplankton cells are probably not nitrogen limited in the ocean (Walsh, 1976). At close to maximal growth rates, laboratory cultures of phytoplankton have, in fact, C/N ratios <6 , which more closely approximate the conditions of individual cell growth in both oligotrophic and eutrophic marine environments (Goldman, 1980). Dissolved organic matter has a C/N ratio of 20-30 in surface waters and as little as ~ 3 in deep waters, but this larger carbon pool has an average residence time of 3400 yr (Williams et al., 1969) and is thought to be a minor input to the sediments (Degens, 1970).

Areas of low C/N in shelf bottom sediments thus presumably reflect regions of "fresh" marine deposits, above which both the phytoplankton cell and its population had not been nitrogen limited during a relatively short travel time from the euphotic zone to the bottom. Areas of high C/N ratios could represent either marine deposits of "older" detrital particulate matter or relict terrestrial deposits not overlain by modern biogenic sediments during the Holocene (Walsh, 1980b). Measurements of natural $\delta^{13}\text{C}$ (Hunt, 1966) and $\delta^{15}\text{N}$ (Peters et al., 1978) isotopes of the sedimentary organic matter, as well as the quality of carbon, e.g. n-alkanes (Farrington and Tripp, 1977), lignin (Hedges and Parker, 1976), and fatty acids (Morris and Calvert, 1980), can then be used to infer a

marine or terrestrial origin for these >10 C/N sediments.

COASTAL UPWELLING

A. Peru

River discharge and tidal mixing of nutrients into the Peru upwelling ecosystem are minimal. A composite (Fig. 2) of the C/N distribution in these sediments (Rowe, personal communication; Suess and Mueller, 1980; ICANE, 1978) accordingly reflects the dominant physical mechanism of a supply of $20-30 \mu\text{g-at NO}_3 \text{ l}^{-1}$ by coastal upwelling (Fig. 3). Low C/N ratios are found in the sediments near the coast where phytoplankton blooms of $10-20 \mu\text{g chl a l}^{-1}$ contain a mean C/N content of 5.1 (Walsh and Howe, 1976). The band of higher sediment C/N ratios of 8-10 between the shelf and the Peru-Chile trench reflects the very low

Fig. 2. The ratio of carbon/nitrogen within surface sediments off Peru (after Suess and Muller, 1980; ICANE, 1978, and G. Rowe, personal communication).

Fig. 3. Cross-shelf distribution of nitrate on the Peru shelf during August 1976 (location shown in Fig. 2).

oxygen content of these slope waters ($\sim 0.1 \text{ ml O}_2 \text{ l}^{-1}$), which does not extend further offshore where the basin sediments again have a lower C/N ratio. Under very low oxygen conditions of the slope, the process of denitrification leads to consumption of ~ 2 moles of carbon for each mole of N_2 produced (Richards, 1965), i.e., an increase in the C/N ratio of the remaining particulate matter during early diagenesis. After further diagenesis and/or as a result of terrestrial input, Pleistocene and older sediments have C/N ratios of 20-30 as well as $\delta^{13}\text{C}$ values of -25 to -28 in contrast to -19 to -21 within recent marine sediments (Degens, 1970).

The surface sediments off Peru have an average $\delta^{13}\text{C}$ of -20.9 and n-alkane hydrocarbon fraction in the low molecular range ($< \text{C}_{24}$), indicating mainly a non-terrestrial component of the organic carbon in these muds (ICANIE, 1978). As much as 50% of the $1000 \text{ g C m}^{-2} \text{ yr}^{-1}$ primary production off Peru may now be deposited on the bottom, resulting in marine deposits of 1 to 4% carbon on the shelf and 5 to 20% on the upper slope (Walsh, 1980c). At 22 stations across the Peru shelf and slope, the mean subsurface C/N value at 50 cm depth of cores is 7.76 compared to 7.53 in surface sediments, i.e., little diagenetic change. In

contrast, the $\delta^{13}\text{C}$ carbon was 3.89 at 50 cm and almost twice as much, 4.82 at the surface (G.T. Rowe, personal communication), suggesting a recent increase in carbon flux to the sediments that could have occurred within just the last decade as a result of overfishing (Walsh, 1980c).

At a sedimentation rate of 55-140 cm 1000 yr⁻¹ off Peru (Muller and Suess, 1979), these subsurface C/N samples at 50 cm depth may be 500 to 1000 yr old. Radio-carbon dating suggests a similar age depth of recent sediments with a $\delta^{13}\text{C}$ of -19.5 to -20.5 off Southwest Africa (Morris and Calvert, 1980). A totally marine input, as also indicated by n-alkane fractions and fatty acids, as well as the apparent lack of diagenesis with depth off Southwest Africa over the last 1000 yr appear to be analogous to the Peru upwelling situation. The 8 to 14% carbon muds on the anoxic southwest African shelf, in contrast to those of 1 to 4% on the Peru shelf, do not decline in amount from 0 to 50 cm depth, however, i.e., no recent transients in carbon loading. These sediment records suggest that until recently more of the photosynthetic carbon was lost to the bottom from the pelagic food web on the Southwest African shelf than off Peru. In fact, the maximum yield of the fishery for the South African sardine was only 5×10^5 tons yr⁻¹ in contrast to 1×10^7 tons of anchovy harvested off Peru in 1970 (Walsh, 1977), before the recent decline in yield to 1×10^6 tons yr⁻¹ during 1976-79.

B. Northwest Africa

Upwelling off Northwest Africa is a seasonal phenomenon of increasing duration as a function of latitude from 15° to 25° N, unlike the more continuous input of nutrients over the same distance off Peru and Southwest Africa (Walsh, 1977). The annual primary production off Northwest Africa is thus 200-500 g C m⁻² yr⁻¹ (Schemainda et al., 1975; Huntsman and Barber, 1977), only 1 to 4% carbon is found on the upper slope (Diester-Haass, 1978), and the oxygen minimum zone still contains 1.5-2.0 ml O₂ l⁻¹. Fatty acid, n-alkane, and $\delta^{13}\text{C}$ contents of a core on the upper slope (Gaskell et al., 1975) all suggest a mixture of terrestrial and marine carbon in these sediments unlike the upwelling areas off Peru and Southwest Africa.

The sediment accumulation rate is much lower within the less productive Northwest Africa upwelling ecosystem, with an age of 1000 yr at only 5 cm depth (Gaskell et al., 1975), i.e., ~ tenfold older than off Southwest Africa (Morris and Calvert, 1980). Apparently, the detritus of marine origin has thus not diluted eolian terrestrial inputs containing as much as 16% carbon (Aston et al., 1973). For example, terrigenous suspended material is found on the inner shelf at 22°N (Milliman, 1977) where a turbidity maximum occurs in the absence of river discharge. The suspended matter of the outer shelf off Northwest Africa is of marine origin, but in small amounts compared to that on the Southwest African (Zemery et al., 1973) or Amazonian shelves (Milliman et al., 1975).

The C/N ratio of the surface sediments reflects both the terrigenous and marine origin of carbon off Northwest Africa with a ratio of >10-15, indicating terrestrial material (Diester-Haass and Muller, 1979) and found in waters less than 50 m depth on the shelf between 15-25°N; a lower ratio of 5-10 is observed in marine deposits on the outer shelf and upper slope (G.T. Rowe, personal communication; Diester-Haass and Muller, 1979). The primary productivity and sediment C/N ratios do not change with latitude off Northwest Africa, suggesting that weak upwelling now provides the nutrient supply in the northern region and river runoff in the southern area (Schemainda et al., 1975). Similar to Southwest Africa, the % carbon does not change much with depth during the 10,000 yr Holocene (0-70 cm) on the Northwest African slope (Gaskell et al., 1975) or rise (Muller and Suess, 1979), and the maximum sardine yield has been only $\sim 3 \times 10^5$ tons yr^{-1} (Gulland, 1970).

During Pleistocene glacial periods, however, the % organic carbon of the sediments increased significantly with depth (70-500 cm) at a changing sedimentation rate on the continental rise from only 3.5 to 23 cm 1000 yr^{-1} off Northwest Africa. Presumably, the rate of carbon input also increased at this time as a function of either greater river runoff or upwelling during lower sea levels (Mueller and Suess, 1979). An increase in the annual upwelling rate, similar to that off Peru and Southwest Africa, could have occurred in response to stronger trade winds (Sarnthein, 1978). For example, during the late Pleistocene, a negative temperature anomaly of 6°C has been estimated for surface waters off Northwest Africa 18,000 yr ago (McIntyre et al., 1976). The C/N content of the rise sediments appeared to have remained the same (~ 8) during the Pleistocene as in the Holocene, there is no other evidence that this sediment carbon was from a terrestrial source during glaciation, and river rafting of material at 20°N off Northwest Africa probably did not occur (Muller and Suess, 1979).

RIVER DISCHARGE

A. Amazon

Each second the Amazon River discharges 2×10^8 liters of fresh water (~ 10 times the input of the Mississippi River) containing ~ 10 $\mu\text{g-at NO}_3 \text{ l}^{-1}$ (Williams, 1968; Ryther et al., 1967) to oligotrophic surface waters (0.1 $\mu\text{g-at NO}_3 \text{ l}^{-1}$) of the adjacent Brazil shelf. The C/N ratio of particulate matter in the river is >10 (Williams, 1968), compared to <6 within the shelf euphotic zone (J. Ryther, personal communication). Phytoplankton production is negligible in the river (Richey et al., 1980), and the suspended terrestrial load drops out of the water column at the 3‰ salinity region of the estuary (Milliman et al., 1975). Export of the river thus consists mainly of dissolved organic carbon of terrestrial origin (Richey et al., 1980), with a C/N of 15-40 (Williams, 1968), and dissolved nutrients.

After uptake of nutrients by phytoplankton on the adjacent shelf, the prime components of suspended matter in the brackish Amazon plume are diatom frustules (Milliman et al., 1975) from phytoplankton populations that are two orders of magnitude larger than within outer shelf waters (Hulburt and Corwin, 1969). The C/N content (Barreto et al., 1975) of the Amazonian shelf sediments (Fig. 4) reflects terrestrial deposition with ratios >10 at the edge of the coast and marine biogenic deposition with ratios <6 on the middle and outer shelf. Little accumulation of diatom frustules is found in the surface shelf sediments (Milliman et al., 1975), however, and despite the trace amounts of marine carbon, much of these deposits are thought to be relict (Barreto et al., 1975). With the exception of the mud wedge off the Amazon mouth, which may also be relict, the highest concentration of organic matter is instead found on the upper slope. The sinking phytoplankton populations, derived from the Amazon affluent, could thus either be transported offshore and/or northward with the Guiana Current.

B. Mississippi

The annual Mississippi River discharge is fivefold that of the total freshwater influx of the Mid-Atlantic Bight (Bue, 1970), with a nitrogen content as

Fig. 4. The ratio of carbon/nitrogen within surface sediments off the Amazon River (after Barreto et al., 1975).

Fig. 5. A time series of nitrate input from the Mississippi River as measured ~50 km downplume on the Louisiana shelf (after Fucik, 1974).

much as $150 \mu\text{g-at NO}_3 \text{ l}^{-1}$ during spring flood (Everett, 1971). After dilution and phytoplankton uptake on the otherwise oligotrophic Louisiana shelf, a seasonal maximum of only $30 \mu\text{g-at NO}_3 \text{ l}^{-1}$ is found ~50 km downplume from the mouth of the river (Fig. 5). The annual primary production of the Mississippi Delta is $\sim 250\text{--}350 \text{ g C m}^{-2} \text{ yr}^{-1}$ (Thomas and Simmons, 1960; Fucik, 1974) in contrast to $\sim 25 \text{ g C m}^{-2} \text{ yr}^{-1}$ for the open Gulf of Mexico and the West Florida and Texas shelves (El-Sayed, 1972). The suspended load of the Mississippi is about 10 times the input of all U.S. east coast rivers (Emery and Uchupi, 1972), such that a turbidity maximum is found within 10 km of the coast (Manheim et al., 1972), where a carbon import of $0.5 \text{ g C m}^{-2} \text{ day}^{-1}$ is received from adjacent bays and marshes (Happ et al., 1977). Over 90% of the carbon export from the Mississippi is in the form of dissolved organic carbon (Hoffman, 1974) with a general seaward decline in concentration observed after input to the shelf (Maurer and Parker, 1972).

A particulate deposition rate of $1.6 \text{ g C m}^{-2} \text{ day}^{-1}$ to the sediments is estimated for the nearshore turbid area with a bottom carbon content of 0.4%, a delta sedimentation rate of $30,000 \text{ cm } 1000 \text{ yr}^{-1}$ (Scruton, 1960), and a primary production of only $0.3 \text{ g C m}^{-2} \text{ day}^{-1}$ (Thomas and Simmons, 1960). In contrast,

a deposition rate of only $0.4 \text{ g C m}^{-2} \text{ day}^{-1}$ may occur 20 km off the coast with bottom sediments of less carbon (0.27%), but where a greater daily primary production of $1.5 \text{ g C m}^{-2} \text{ day}^{-1}$ occurs. Like the Amazon system, there may also be seaward export of the phytoplankton carbon, derived from the Mississippi effluent, to the upper slope of the Gulf of Mexico where 1 to 2% organic carbon muds are located (Trask, 1953). In contrast to Peru, however, the near bottom oxygen content of the sedimentary organic carbon maximum is still $2\text{--}3 \text{ ml O}_2 \text{ l}^{-1}$ at depths of 150–850 m (Richards and Redfield, 1954).

The C/N ratio of particulate matter in the Mississippi River is >10 in contrast to <6 within the euphotic zone of the Louisiana shelf (Armstrong, 1974). In the surface sediments, the C/N ratio reflects the above source and deposition patterns with >10 ratios found near the coast (Fig. 6) and <6 observations found under the Mississippi River plume (Trask, 1953; Armstrong, 1974). The $\delta^{13}\text{C}$ value of particulate matter in the Mississippi River is -25.6 to -25.9 (Hoffman, 1974), whereas 10–20 km off the mouth, particulate values of -19.6 are found; the latter are representative of warm water plankton (Sackett et al., 1971). The $\delta^{13}\text{C}$

Fig. 6. The ratio of carbon/nitrogen within surface sediments of the Gulf of Mexico and the South Atlantic Bight (after Trask, 1954; Armstrong, 1974; Folger, 1972; Hathaway, 1971; and R. Iverson, personal communication).

values of the surface sediments of the Mississippi River, of adjoining bays, and of the salt marshes are similarly -22.4 to -23.6, in contrast to -25.0 of sediments in the fresh water swamps and -19.0 to -20.8 of sediments of the outer continental shelf (50-100 m) between the Mississippi Delta and the Rio Grande (Hedges and Parker, 1976).

The Holocene sediments (~0-60 cm) of the upper slope (Parker et al., 1972) also have $\delta^{13}\text{C}$ values of marine origin (-18.0 to -20.0), whereas the deeper Pleistocene sediments (~60 to 1000 cm) have $\delta^{13}\text{C}$ (-22.0 to -26.0) and C/N (~10-12) values that may be of terrestrial origin (Newman et al., 1973) rather than just reflections of diagenesis (Degens, 1970). During lower sea levels of the Pleistocene, both the Mississippi (Davies and Moore, 1970) and the Amazon (Milliman et al., 1975) instead discharged into the abyssal plains, presumably depositing terrestrial matter on the upper slope (Sackett, 1964). After glacial recessions, the sedimentary rate of input to the bottom is thought to decline by 1 to 2 orders of magnitude (Mueller and Suess, 1979), such that modern sediments are now only found underneath the Mississippi Plume on the east Texas-Louisiana shelf with the rest of the shelf sediments considered to be relict (Emery and Uchupi, 1972). The present export to the slope of the Gulf of Mexico may thus be mainly organic matter fixed on the continental shelf (Walsh, 1980a).

C. Hudson

At present, over 90% of the suspended terrestrial matter is trapped within the estuaries adjacent to the northeast continental shelf (Meade et al., 1975), where twice as much river runoff occurs as in the South Atlantic Bight (Bue, 1970). The seston and nutrient content of the water column both decline by an order of magnitude between the Hudson River and the New York Bight (F. Falkowski, personal communication). A gradient in nitrate concentration from ~40 to 5 $\mu\text{g-at NO}_3 \text{ l}^{-1}$ is observed and again the major carbon export from this estuary is in the form of dissolved organic carbon (Falkowski et al., 1980).

The suspended matter at the mouth of the estuaries is nearly all organic and on the shelf it seasonally covaries (Manheim et al., 1970; Meade et al., 1975) with changing chlorophyll distributions as a function of the nutrient supply rate mediated by the stratification of the water column (Walsh et al., 1978). During summer when the nutrient supply from local upwelling (Fig. 7) and river runoff is confined to within 10 km of the coast, the C/N content of high chlorophyll (>10 $\mu\text{g chl l}^{-1}$) patches of the Hudson River plume is <6. Offshore, where <1 $\mu\text{g chl l}^{-1}$ is found above the pycnocline, C/N values of summer particulate matter in the New York Bight are instead >10.

Most of the sediment on the Mid-Atlantic shelf is relict sand with <0.5% carbon (Emery and Uchupi, 1972), deposited during or soon after the last Wisconsin glaciation, when the shelf was dry land 15,000 years ago. Seaward of the 60 m

Fig. 7. Cross-shelf distribution of nitrate on the New York shelf during October 1978 (location shown in Fig. 8).

isobath, the sediments contain more silt (Freeland and Swift, 1978), and, beyond the shelfbreak, the organic content finally increases to 1 to 2% carbon on the upper slope, but still lower than the 5 to 20% carbon muds off the Peru coast (Walsh, 1980c). The Hudson Canyon, areas of Georges Bank (Hathaway et al., 1979), and the "mud hole" a region of $4 \times 10^3 \text{ km}^2$ southwest of Martha's Vineyard between the 50-100 m isobaths, are the exceptions where 1 to 2% carbon muds are found on the Mid-Atlantic shelf. In these areas, ^{14}C and ^{210}Pb dating suggest modern deposits with an accumulation rate of at least $50\text{--}100 \text{ cm } 1000 \text{ yr}^{-1}$ (Drake et al., 1978; Bothner et al., 1980) and perhaps $363 \text{ cm } 1000 \text{ yr}^{-1}$ on Georges Bank (Hathaway et al., 1979), in contrast to about $15 \text{ cm } 1000 \text{ yr}^{-1}$ on the slope (MacIlvaine, 1973) and $3 \text{ cm } 1000 \text{ yr}^{-1}$ on the rise (Milliman, 1973).

Fig. 8. The ratio of carbon-nitrogen within surface sediments of the Mid-Atlantic Bight (after Hathaway, 1971 and G. Rowe, personal communication).

The C/N ratio (Fig. 8) of Mid-Atlantic shelf sediments (Milliman, 1973) can be used as an index of areas of modern carbon deposition and possibly also of the fate of particles during transport in this system. At the head of Chesapeake Bay, for example, the C/N content of the sediment is >30 , and $\delta^{13}\text{C}$ values of -24.0 to -25.3 (Hunt, 1966) suggest terrestrial origin. The sediments in the Hudson River also have a $\delta^{13}\text{C}$ value of -23.3 to -24.7 . Southeast of the Hudson, Delaware, and Chesapeake estuaries, the low C/N content of the sediments suggest seaward transport of sinking marine organisms (Walsh, 1980b). Within the "mud hole," $\delta^{13}\text{C}$ measurements of -20.5 to -21.0 (Hunt, 1966) also suggest that the surface sediments with a C/N <6 south of Martha's Vineyard (Fig. 8) are of marine origin. Similar $\delta^{13}\text{C}$ measurements are not available for the >10 C/N sediments off New York and New Jersey (Fig. 8), but the n-alkane hydrocarbon fraction of the sediment carbon pool in this region suggests that this material instead had both terrestrial and petrochemical origin (Farrington and Tripp, 1977).

The higher C/N ratios of 16-30 in the shelf sediments (Fig. 8) thus suggest terrestrial or vascular plant carbon deposition, presumably when relict sands were deposited during both the last Wisconsin and previous glaciations. The depth of Pleistocene sediments on the upper slope is ~300 m in contrast to only ~100 m on the shelf (Hathaway et al., 1978), i.e., average accumulation rates of respectively 15 cm 1000 yr⁻¹ and 5 cm 1000 yr⁻¹. As the river mouths and marshes retreated towards the present shore line during the Holocene transgression, sediments of >10 C/N content were continually laid down and are still observed across the shelf in some areas, e.g., south of Montauk Point and east of Atlantic City (Fig. 8), where the annual primary production is now ~250 g C m⁻² yr⁻¹. In other areas, e.g., Georges Bank (Hathaway et al., 1979), Nantucket Shoals, the "mud hole," Hudson Canyon, and downstream of the mouth of estuaries, a surface layer of modern biogenic <6 C/N sediments may presumably now overlay the relict sediments with their >10 C/N ratios. The primary production of the Hudson River plume and Georges Bank is ~500 g C m⁻² yr⁻¹, a result of river runoff and tidal mixing, and may represent some of the upstream protein sources for the low C/N sediments and for a peak fish yield of ~1.4x10⁶ tons yr⁻¹ on this shelf.

EDDY-INDUCED UPWELLING

A. Florida Current

Classical explanations of western boundary currents (Stommel, 1948; Munk, 1950) suggest that positive (counterclockwise) vorticity is developed at the shelf-edge sides of these fast moving currents to provide a frictional balance to the negative vorticity of the basin circulation induced by the curl of the wind stress. Spin-off cyclonic eddies in the Florida Current (Lee, 1975) do, in fact, rotate counterclockwise and their occurrence is positively correlated with the curl of the wind stress (Duing et al., 1977), i.e., the greater the curl the more intense is the positive vorticity generated at the edge of the Florida-Georgia shelf. Ekman suction is created in the center of these eddies (Lee and Mayer, 1977) with an upwelling velocity of 10⁻³ cm sec⁻¹ at the edge of the shelf (Hsueh and O'Brien, 1972) in contrast to 10⁻² cm sec⁻¹ found nearshore during coastal upwelling (Walsh, 1975).

As much as 10 µg-at NO₃ l⁻¹ can be brought into the oligotrophic outer shelf euphotic zone within the center of these eddies (Lee et al., 1980), with less than 1 µg-at NO₃ l⁻¹ found at their boundaries (Fig. 9). Diatom blooms of 5 µg Chl l⁻¹ have been found associated with the nitrate input from eddy-induced upwelling at the shelf break off Florida-Georgia (Atkinson et al., 1980). Without consideration of nitrogen recycling, a minimal annual production of ~40 g C m⁻² yr⁻¹ at a C/N ratio of 5/1 (Lee et al., 1980) could be derived from such a nitrogen source. This is, in fact, fivefold the estimated nitrogen input from river

Fig. 9. Cross-shelf distribution of nitrate on the Georgia-Florida shelves (A-D) during April 1977 (location shown in Fig. 6) in relation to surface temperatures of a spin-off eddy (after Lee et al., 1980).

runoff (Haines, 1975; Dunstan and Atkinson, 1976). After nitrogen recycling, the river input eventually yields $\sim 550 \text{ g C m}^{-2} \text{ yr}^{-1}$ (Thomas, 1966) within 5-10 km of the turbid coastal zone, where a threefold higher suspended load is received in the South Atlantic Bight than in the Mid-Atlantic Bight (Emery and Uchupi, 1972).

Nutrients and presumably chlorophyll do not have a long residence time on the outer shelf of the South Atlantic Bight, however, since this area is flushed after every upwelling event of 1 to 3 week duration (Lee and Mayer, 1977). Although fast growing gelatinous zooplankton are found at the shelf break, their abundance is not correlated with the spin-off eddies (Atkinson et al., 1978). Most of the eddy-induced primary production may sink to the bottom and/or be flushed offshore rather than be incorporated within higher trophic levels. The seasonal oscillations of copepod and larval fish biomass of the Georgia-Florida shelf are instead correlated with nearshore increases of estuarine derived nutrients during late summer-early fall (Turner et al., 1979), such that both the fish biomass and yield of the South Atlantic Bight is $\sim 10\%$ that of the Mid-Atlantic Bight (Turner et al., 1979; Walsh, 1980b).

Fig. 10. Cross-shelf distribution of nitrate on the West Texas shelf (location shown in Fig. 6) during November 1969, 1977 (after Flint and Griffin, 1977; NODC).

Fig. 11. Cross-shelf distribution of nitrate on the Baja California shelf during April 1973 and on the Oregon shelf during July 1973.

The low C/N ratio of the sediments (Hathaway, 1971) at the Florida-Georgia shelf break (Fig. 6) confirms deposition of marine carbon in this region as corroborated by n-alkane fractions (Hathaway et al, 1979) and $\delta^{13}\text{C}$ sediment values of -19.4 to -21.3 (Hunt, 1966) on the outer shelf. In contrast, the seston has a $\delta^{13}\text{C}$ value of -27 in the rivers (Haines, 1979), where the sediments are similarly -25.1 to -26.1 (Hunt, 1966), implying that the high nearshore sediment C/N ratios represent terrestrial material that has dropped out of the estuarine water column. Similar to the Louisiana swamps, the Georgia salt marshes may mainly export carbon in the form of DOC at an annual rate of $\sim 22 \text{ g C m}^{-2} \text{ yr}^{-1}$ (Pomeroy et al., 1977). The ephemeral production and deposition of marine carbon from eddy-induced upwelling is restricted to the edge of the shelf, however, unlike coastal upwelling off Peru and Southwest Africa, such that most of the Florida-Georgia shelf deposits are relict sands (Emery and Uchupi, 1972).

B. Mexican Current

The presence and seasonal intensification of a western boundary current off the Mexican-Texas shelf has been correlated with changes in the wind curl

Fig. 12. A time series of nitrate input from slope water as measured at the shelf-break of the West Texas shelf (after Flint and Griffin, 1977).

(Sturges and Blaha, 1976) that might also lead to eddy-induced upwelling off this coast (Fig. 10). In contrast to coastal upwelling at the same latitude off Baja California (Fig. 11), however, high nitrate concentrations are confined to the shelf edge off the Texas coast. An annual time series of nitrate content at this shelf break (Fig. 12) further suggests that eddy-induced upwelling might be intensified during winter and summer, i.e., in phase with changes in the wind curl forcing (Sturges and Blaha, 1976). At a yearly nitrate input of half the more continuous Florida-Georgia system, at least $20 \text{ g C m}^{-2} \text{ yr}^{-1}$ might be produced on the west Texas shelf; this, in fact, is the estimated annual production (El-Sayed, 1972). The only low C/N ratios of the west Texas shelf sediments are found at the shelf break (Fig. 6), suggesting that eddy-induced upwelling leads to carbon export in this system, as well.

C. Guiana Current

The Guiana Current flows north at the edge of the Amazon Shelf as a western boundary extension of the North Equatorial Current (Metcalf, 1968). Varying at speeds of $20\text{--}200 \text{ cm sec}^{-1}$ as a function of the wind forcing (Milliman et al., 1975), the Guiana Current should shed spin-off eddies on the Amazon shelf analogous to those on the Florida, Georgia, and Texas shelves. Upwelling at the

Fig. 13. Cross-shelf distribution of nitrate on the Guiana shelf (location shown in Fig. 4) during June 1965 (after Ryther et al., 1967).

shelf-break due to an unknown mechanism was, in fact, postulated for this shelf (Ryther et al., 1967) on the basis of nitrate sections (Fig. 13), similar to those shown across the Florida and Mexican Currents (Figs. 9, 10). Away from the influence of the Amazon River discharge, the low C/N content of sediments at the shelf-break (Fig. 4) may also reflect the concomittant process of eddy-induced upwelling from this western boundary current.

D. Loop Current

Nitrate sections across the Loop Current at the edge of the West Florida shelf (O'Brien and Wroblewski, 1972) suggest shelf-break upwelling (Fig. 14). Spin-off eddies have been observed on this shelf as well (Maul, 1977). The upward slope of the nutrient isopleths in the offshore region (Fig. 14) is also typical of synoptic nutrient sections (Morrison and Nowlin, 1977), delineating the anti-cyclonic Loop Current as it penetrates farther into the Gulf of Mexico from the Yucatan Strait. During 1973-77, however, the Loop Current was found due north of 28°N only 10% of the time, and against the shelf-break less than 30% of the time north of 26°N (Vukovitch et al., 1979). Without frequent generation of positive vorticity as frictional drag of the Loop Current against the West Florida shelf, less eddy-induced upwelling would occur in response to this western boundary current as from the Florida, Mexican, and Guiana Currents. The only area of low C/N

Fig. 14. Cross-shelf distribution of nitrate on the West Florida shelf (location shown in Fig. 6) during November 1969, 1971 (after O'Brien and Wroblewski, 1972; NODC).

ratio in the sediments (Folger, 1977; Hathaway, 1971; R. Iversen, personal communication) is, in fact, found at the Key West shelf-break where the Loop Current continually passes through the Straits of Florida (Fig. 6). Marine biogenic deposition has not overlain the Mangrove detritus of the nearshore zone or the relict deposits on the rest of the West Florida shelf, while the annual fish yield km^{-2} (Heald, 1969) is 1/6 that of the comparatively impoverished South Atlantic Bight.

TIDAL MIXING

A. Georges Bank

On some shelves, tidal motion is an important factor of nutrient supply in addition to upwelling and river runoff. Large tidal velocities of 55 cm sec^{-1} maximum amplitude, for example, can lead to roughly the same vertical mixing as a 13 m sec^{-1} wind (Pingree et al., 1978), with the tidal mixing energy applied to the water column from below as opposed to the wind energy from above. Maximal tidal velocities at $\sim 30 \text{ m}$ depth are $\sim 15\text{-}25 \text{ cm sec}^{-1}$ in the New York Bight and

~ 55 – 110 cm sec^{-1} on Georges Bank. The latter velocities are similar to those of 250 cm sec^{-1} at the mouth of the Amazon (Milliman et al., 1975) and 150 cm sec^{-1} around the British Isles (Simpson and Pingree, 1978), where an index of the tidal mixing (Simpson and Hunter, 1974) has been formulated as the ratio of the depth (h) of the water column to the cube of the amplitude (u) of the tidal stream, i.e., $h u^{-3}$. The reciprocal of this ratio and a constant drag coefficient, $C h^{-1} u^3$, is the mean tidal energy dissipation rate per unit mass (Pingree et al., 1978). Because of the large range of values in both parameters, a log scale is used to estimate areas of stratification ($\log h u^{-3} \geq 2$ on $\log h^{-1} u^3 \leq -2$, transitional areas or fronts (1.5 or -1.5), and tidally well-mixed areas (1 or -1). Larger tidal velocities and/or shallower depths favor increased tidal mixing.

At depths of 40–60 m in the New York Bight, a tidal velocity of 25 cm sec^{-1} suggests values of 3.41–3.59 for $\log h u^{-3}$, and thus a stratified water column in the absence of wind forcing and thermal convection. Consequently, the annual temperature cycle at the 60 m isobath off New York (Fig. 15) is that of a typical boreal shelf in contrast to the summer upwelling regime off Oregon at the same latitude (Walsh, 1980b). The vernal heating cycle, river runoff, and a decline in wind events all lead to intense stratification and associated nutrient depletion of surface waters by late summer in the New York Bight (Fig. 7).

A tidal velocity of 110 cm sec^{-1} at similar depths on Georges Bank, however, suggests values of 1.48–1.65 for $\log h u^{-3}$, i.e., areas of tidal mixing. The seasonal temperature cycle at 60 m on Georges Bank, in fact, reflects tidal mixing with an isothermal vertical structure at all times (Fig. 15). Because of tidal and wind mixing on Georges Bank, the winter temperature minimum is about the same as that of the New York shelf but the summer temperature maximum is similar to that off Oregon during coastal upwelling (Fig. 15). The effective rate of vertical mixing on Georges Bank has been estimated (Falkowski, personal communication) to be $\sim 10^{-1} \text{ cm sec}^{-1}$ in contrast to a seasonal upwelling velocity of $\sim 10^{-2} \text{ cm sec}^{-1}$ off Oregon.

Although the source waters for Georges Bank contain less nitrate (Fig. 16) than coastal upwelling water off Oregon (Fig. 11), the relatively high and continuous flux of nutrients onto Georges Bank leads to an annual primary production of $\sim 500 \text{ g C m}^{-2} \text{ yr}^{-1}$ in contrast to ~ 200 – $250 \text{ g C m}^{-2} \text{ yr}^{-1}$ on the Oregon and New York shelves (Walsh, 1980b). Grazing on Georges Bank by copepods consumes only $\sim 30\%$ of the daily primary production (Dagg and Grill, 1980), allowing a possible input of $350 \text{ g C m}^{-2} \text{ yr}^{-1}$ to the Bank sediments, i.e., similar to the combined nearshore deposition fluxes of carbon derived from effluent of the Mississippi River (Thomas and Simmons, 1960).

With a 1.5% carbon content and a sedimentation rate of $363 \text{ cm } 1000 \text{ yr}^{-1}$ (Hathaway et al., 1979), the accumulation rate can be calculated (Muller and

Fig. 15. Seasonal temperature structure of the water column at the 60 m isobath on A) the New York, B) Oregon, and C) Georges Bank shelves (after Walsh, 1980b).

Suess, 1979) to be $27 \text{ g C m}^{-2} \text{ yr}^{-1}$. Accordingly, less than 10% of the water column production apparently remains on Georges Bank. The downstream export of phytoplankton production with a low C/N value from Georges Bank and Nantucket Shoals leads to modern sediments in the "mud hole" and on the upper slope with a $\text{C/N} < 6$ (Fig. 8) and $\delta^{13}\text{C}$ of ~ -19 to -21 (Hunt, 1966). As much as 90% of the fish yield from Cape Hatteras to Nova Scotia is taken north of New York, suggesting that export from the tidally driven production of Georges Bank may be the major supply of photosynthetic carbon to this ecosystem.

Fig. 16. Cross-shelf distribution of nitrate on the Georges Bank shelf during October 1978 (location shown in Fig. 8).

B. Bering Sea

Tidal velocities of 20-50 cm sec⁻¹ in the southeast Bering Sea generate a tidally mixed water column inshore of the 50 m isobath, 100 km off the coast (Coachman and Walsh, 1980). Without significant horizontal advection or river discharge on the southeast Bering shelf, the supply of nitrate depends upon a tidally driven "diffusion" (Csanady, 1976) shoreward over the 400 km distance from the shelf-break (Fig. 17). In contrast, the distance to the 50 m isobath from the shelf-break off Georges Bank (Fig. 16) is 40-50 km. The daily onshore flux of nitrate below the euphotic zone of the Bering sea shelf is similar to

Fig. 17. Cross-shelf distribution of nitrate on the Southeast Bering Sea shelf during April 1979 (location shown in Fig. 19).

that introduced by upwelling off Peru (Walsh, 1975). The tidal regime of the wide Bering Sea (500 km) leads to an effective vertical velocity of 10^{-3} cm sec $^{-1}$ from diffusion (Coachman and Walsh, 1980) or advection (Stigebrandt, 1980) considerations, however, in contrast to a vertical flow of 10^{-2} cm sec $^{-1}$ compressed within 20-30 km of the Peru coast. As a result of this tenfold greater vertical flux of nutrients, both Peru shelf waters and Georges Bank remain eutrophic year-round, while the surface waters of the Bering Sea become depleted of nitrogen by summer.

The phytoplankton populations of the Bering Sea have a mean C/N ratio of 5.7 (Loder, 1971) and a $\delta^{13}\text{C}$ value of -24.4 (McConnaughey and McRoy, 1979), similar to that of Antarctic phytoplankton (Sackett et al., 1965). The annual primary production is only ~ 200 g C m $^{-2}$ yr $^{-1}$ (McRoy and Goering, 1976) as a result of light and nutrient regulation. Because of different reproductive strategies, the overwintering zooplankton of the outer Bering Shelf (100-170 m) consume most of the spring bloom in this region, whereas the annual zooplankton populations of the middle shelf (50-100 m) do not crop more than 50% of the bloom in this area, similar to seasonal cycles of carbon flux on Georges Bank and in the New York Bight (Walsh, 1980b). Different carbon inputs to the bottom of the outer and middle shelf are indicated by the cross-shelf distribution of C/N ratios in the

Fig. 19. The ratio of carbon/nitrogen within surface sediments of the Southeast Bering Sea (after Lisitzin, 1966; R. Iversen, personal communication).

sediments (Lisitzin, 1966; R. Iversen, personal communication), where ratios < 6 are found on the middle shelf, with values increasing towards the shelf-break (Fig. 18).

The $\delta^{13}\text{C}$ values of the southern Bering Shelf sediments also indicate marine origin, -22.0 to -22.9 (Peters et al., 1978). The *n*-alkane hydrocarbon fractions, however, suggest a mixed terrestrial source (Venkatesan et al., 1980), similar to the sediments off northwest Africa. The origin of these terrigenous sediments, 250-500 km from the mouth of the Kuskokwim River, may be related to seasonal ice formation, rafting, and break-up in the Bering Sea rather than the result of eolian input of terrestrial carbon as off Spanish Sahara. The Pleistocene sediments are generally less than 100 m thick on the Bering Sea shelf and the near-shore holocene accumulation rate is ~ 20 cm 1000 yr^{-1} (Nelson et al., 1974), i.e., tenfold less than Georges Bank. Continuous tidal resuspension and gravitational settling along a sloping bottom may lead to a net export of particulate matter from the Bering Sea. Similar to other shelves, the gradient of sediment organic matter is from $< 0.5\%$ carbon over the mid-shelf (Lisitzin, 1966; Sharma, 1974) to the $> 1.0\%$ on the outer shelf and upper slope (Gershanovitch, 1962; Iversen, personal communication).

C. North Sea

The width and bottom topography of the continental shelf is remarkably similar in the southeast Bering Sea and the northern North Sea. A 500 km section across the North Sea from the Norwegian Deep between the Shetland Islands and Bergen, Norway to the Firth of Fourth, Scotland, traverses the rich Fladen Ground at depths of 100-150 m. A section of the same length from off Unalaska Island in the Aleutian chain to Cape Newenham on the Alaska mainland also intersects the rich pollock fishing grounds, known as the Golden Triangle, on the broad outer shelf (100-150 m depths) of Bristol Bay. Associated with changes in bottom topography of the North Sea, at least two fronts are found at both the 30-50 m and 150-200 m isobaths in relation to transitions in respectively, the physical dynamics of the tidally mixed shallow areas (Pingree et al., 1978) and of the interleaving water masses at the shelf break (Steele, 1961). Similar to the Bering Sea, changes in horizontal salinity and productivity patterns suggest a different physical regime landward of the 100 m isobath in the northern North Sea as well (Steele, 1956), reflecting increased tidal mixing in this area compared to the deeper Fladen Ground. Recent observations in the Bering Sea (Iversen et al., 1979) similarly indicate the presence of fronts at both the 50 m and 170 m isobaths, as well as a third transition area (middle front) at the 100 m isobath.

If the hydrodynamic structure is similar, the rate of cross-shelf nutrient exchange from the shelf-break of the two seas may also be similar. Furthermore, since the southeast Bering Sea and the northern North Sea are found at the same latitudes, $\sim 55-60^{\circ}\text{N}$, the annual incident radiation is about the same in both epicontinental seas. The spring bloom commences at the same time (March-April) in the southeast Bering Sea as in the northern North Sea, and the annual primary production is also similar, $\sim 100-200 \text{ g C m}^{-2} \text{ yr}^{-1}$ (Steele, 1956; McRoy and Goering, 1976). Nutrient depletion of the euphotic zone occurs by June in response to the algal growth within each shelf, and the fish yield from the Fladen Ground and the Golden Triangle is the same, $\sim 5-10 \text{ tons km}^{-2} \text{ yr}^{-1}$ (Walsh, 1980b). As might be expected, the loss process appears to be similar, such that C/N ratios < 6 are also found (Jansen et al., 1979) in the mud deposits of the Fladen Ground sediments.

In contrast to the Bering Sea, however, the southern North Sea receives large anthropogenic loadings of nutrients from sewage wastes and agricultural fertilizers in the form of river runoff. The phosphate content of the Rhine River has increased tenfold over the last 50 years and the total nitrogen content is now $380 \text{ } \mu\text{g-at N l}^{-1}$ (Postma, 1978). Within the tidally mixed area $< 30 \text{ m}$ depth, as much as $350 \text{ } \mu\text{g-at NO}_3 \text{ l}^{-1}$ has been similarly observed off the Belgian coast (Mormaerts et al., 1979); the urban input of nitrate can also be traced southward along the east coast of England (Johnston, 1973). The river discharge of nitrogen to the southern North Sea now constitutes 60% of the annual supply with the

remainder advected and/or tidally "diffused" from the English Channel and the northern North Sea (Postma, 1973). The primary production of the Waddensea has increased from $\sim 80 \text{ g C m}^{-2} \text{ yr}^{-1}$ in 1950 to $\sim 240 \text{ g C m}^{-2} \text{ yr}^{-1}$ in 1970 (Postma, 1978) with a concomitant increase in the organic content of the sediments of this area (DeGroot, 1973), as a result of such nutrient stimulation of the coastal zone.

Similar to the Amazon River, the suspended load of the Scheldt River drops out of the water column at the 1 to 2‰ salinity region (Nihoul et al., 1979) of the estuary. Despite a sedimentation rate of $400\text{--}600 \text{ cm } 1000 \text{ yr}^{-1}$ off the Sheldt estuary (McCave, 1973) a lack of diatom accumulation in the sediments is also observed at the mouth of the river (Wollast and DeBroeu, 1971). The C/N content of the Rhine River sediments similarly reflects the declining terrestrial source of carbon, with a C/N ratio of 21 found in the river, 14 in the lower estuaries, and 11 in the nearshore Waddensea (DeGroot, 1973). Presumably, the nearshore tidally mixed area prevents both high production as a result of turbidity maxima and deposition of marine organic matter in the sediments by resuspension. Where sewage outfalls discharge both dissolved and particulate material directly to aphotic zone of the continental shelf, e.g., White's Point off California, high C/N ratios of 13-20 (Word and Mearns, 1979) are also found in the adjacent 8% carbon sediments (Sweeney et al., 1978) despite the high nutrient content ($\sim 2500 \mu\text{g-at NH}_3 \text{ l}^{-1}$) of the sewage effluent.

Estimates of the net transport (Eisma, 1973) of fine sediments, originating mainly from rivers (DeGroot, 1973) in the southern North Sea, suggest a residual drift of organic matter to the northeast. From the area of high nutrient discharge, the bottom carbon increases towards the >3‰ muds of the Norwegian Deep and the Skaggerak (Van Weering, 1980), where deposition rates are $\sim 100\text{--}450 \text{ cm } 1000 \text{ yr}^{-1}$ (Eisma, 1973). If all of the shelf mud deposits are also modern, however, present mud budgets cannot account for enough source material, even at a sedimentation rate of only $\sim 10 \text{ cm } 1000 \text{ yr}^{-1}$ over most of the North Sea shelf (McCave, 1973), i.e., similar to that in the Bering Sea.

Beneath a holocene surface layer of $\sim 40 \text{ cm}$ thickness with a similar C/N ratio of 7-8 with depth in the muds of the northern North Sea, an "allochthonous" terrestrial sediment of 15-17 C/N ratio is thought to have been deposited $\sim 8700 \text{ BP}$ on the Fladen Ground (Jansen et al., 1979) after a sea level standstill of $\sim 1500 \text{ yr}$ (Jansen, 1976). Similar to Peru, the C/N ratio at a nearby station is 7.5 at 0-2 cm and 7.9 at 81-83 cm, in contrast to 0.77% carbon at the top and 0.48% at the bottom of the core (Jansen, personal communication), suggesting little diagenetic change. The higher carbon content may reflect increased primary productivity (Jansen et al., 1979) or a "recent" increase in the carbon loading as a result of the 90% decline in herring stocks over the last 40 years (Burd, 1978), with a maximum yield of 1.4×10^6 tons in 1955. If this high C/N material at the

Holocene-Pleistocene boundary is not the result of early diagenesis, but reflects a terrestrial plant origin similar to the Gulf of Mexico, an accumulation rate of 40-80 cm 8700 yr⁻¹ or 5-10 cm 1000 yr⁻¹ might have indeed occurred during the Holocene in parts of the North Sea.

In the anoxic sediments of the Santa Barbara Basin beneath California coastal waters, with a production of only ~200-300 g C m⁻² yr⁻¹ (Eppley et al., 1979), the residual terrestrial carbon is thought to be more refractory than the marine carbon. A half-life of 53,000 yr is estimated for land-derived carbon instead of 15,000 yr for diatomaceous varves (Heath et al., 1977), being deposited at a rate of 400 cm 1000 yr⁻¹ (Koide et al., 1972). The decline in total nitrogen content of this sediment column is thought to be the result of deamination of organic matter during diagenetic bacterial activity (Rittenburg et al., 1955). Similar to Northwest Africa, deposition of marine detritus on the California shelf accounts for only 50 to 67% of the sediment nitrogen, and the C/N ratios are 11.2 in the upper 8 cm of sediment, 12-14 in the 100-600 cm depth interval, and 15 in ancient rocks (Emery and Uchupi, 1972; Heath et al., 1977, Sweeney et al., 1978).

Recent $\delta^{15}\text{N}$ analyses suggest that indeed the terrestrial organic nitrogen is sufficiently bound with clay minerals to be nondegradable by sediment microflora in the Santa Barbara basin (Sweeney et al., 1978). The marine organic nitrogen content remains constant within the upper 3-4 m of the sediments, i.e., over the last 1000 yr, and then declines with depth. The total carbon also decreases with depth (Heath et al., 1977), while the concentration of methane increases (Emery and Hoggan, 1958). In contrast, the amount of terrestrial nitrogen does not change with depth, suggesting that the increase of the C/N ratio may reflect an increasing terrestrial contribution to the organic matter with depth.

Similar to the sediments off Southwest and Northwest Africa, the percent carbon in these high C/N sediments off California did not increase at the surface, despite a 90% reduction in Pacific sardine biomass over the last 75 yr (Soutar and Isaacs, 1974). The maximum total clupeid landings from these upwelling ecosystems has been 3-8x10⁵ tons yr⁻¹ in contrast to 1.4-12x10⁶ tons yr⁻¹ for the herring and anchovy fisheries of the North Sea and Peru, where the impact of overfishing may have been "recorded" in the sediments. In oxic sediments, with high kinetic energy environments (Didyk et al., 1978) such as the northern North Sea, increased diagenesis of the marine carbon might occur, however. The increase of the C/N ratio beneath 40 cm of North Sea sediment rather than beneath 4 m of California sediment may still reflect nondegraded terrestrial input to the former. If the accumulation rate of 10 cm 1000 yr⁻¹ for the North Sea shelf is correct, rather than 100 cm 1000 yr⁻¹ off Peru, most of the muds are not modern, and part of the phytoplankton carbon is exported off this European shelf as well.

CONCLUSIONS

The amount of organic matter is greater in the sediments of the upper slope than those on the shelf off Peru, Southwest Africa, Northwest Africa, California, Oregon, Brazil, Texas-Louisiana, Alaska, Europe, and in composite over 50 m depth intervals along the U.S. coast from Florida to Maine (Fig. 19). All of these sediments past the "mud line" at the shelf-break (Murray and Renard, 1891) contain $>1\%$ carbon such that the lower C/N ratios of these slope regions (Fig. 20) are not a result of sorption of nitrogen on clay minerals (Muller, 1977; Suess and Muller, 1980), but reflect marine origin of the detritus (Walsh, 1980b). In 1840 sediment samples taken between 24° and 44°N (Hathaway, 1971), mean C/N ratios <6 are only found on the slope at 400-1800 m (Fig. 20), with $>1\%$ carbon sediments extending from 800-2200 m (Fig. 19).

Although $\delta^{13}\text{C}$ values of plankton range from -19.6 to -30.6 (Sackett et al., 1965) between the tropical and polar oceans, at any given latitude of the above shelves $\delta^{13}\text{C}$ values ≥ 25 appear to indicate carbon of terrestrial origin and <21 of marine origin. The wealth of information summarized in this analysis suggests that most terrestrial material of a C/N >10 , with a $\delta^{13}\text{C}$ ≥ 25 , is now either trapped in estuaries or has been deposited on the shelf-slope during glacial

Fig. 19. The composite shelf-slope distribution of percent organic carbon within surface sediments from Florida to Maine (after Hathaway, 1971).

Fig. 20. The composite shelf-slope distribution of carbon/nitrogen within surface sediments from Florida to Maine (after Hathaway, 1971).

periods of the Pleistocene. The marine carbon, mainly ungrazed phytoplankton of a C/N <6 and a $\delta^{13}\text{C}$ $\leq$ 21, appears to transit the present shelf and be deposited on the upper slope, whether the source of nutrients is at the coast from coastal upwelling and river runoff or at the shelf-break from eddy-induced upwelling and tidal mixing.

The magnitude of shelf carbon export, its mode of transport, and the time scale of early diagenesis are unknown. Based on carbon and nitrogen budgets of the annual fluxes of these elements through food webs of the Mid-Atlantic Bight, Gulf of Mexico, Bering Sea, and the Peru Upwelling, however, about 50% (Walsh, 1980a) to as much as 75% (Schopf, 1980) of the primary production of the shelves may be exported to the continental slope. A combination of sinking phytoplankton (Walsh et al, 1978; Malone and Chervin, 1979), not consumed by coastal herbivores (Walsh, 1976), their resuspension in a bottom boundary layer by the tides, and the offshore force of gravity along a sloping bottom might lead to transport of both phytodetritus and fecal pellets to a slope depocenter of weaker tidal currents, where 1 to 25% of the shelf production might eventually be buried (Schopf, 1980; Walsh, 1980a). Organic phosphorous is presently a major constituent of the sediment pool of this element on the North Carolina upper slope, but not on the rise (Morse and Cook, 1978), suggesting that early diagenesis of this recent

marine carbon (Figs. 19,20) has apparently not yet been completed.

Future experiments and models will provide additional details on the internal coupling of the food webs and of the linkages to physical forcing of any particular continental shelf. Resolution of large-scale societal conflicts in terms of coastal zone eutrophication, overfishing, waste disposal, and control of atmospheric CO₂ levels, however, must await further definition of slope-shelf exchange processes. To avoid a marine "tragedy of the commons" (Hardin, 1968), appropriate boundary flux conditions must be established for the loss mechanisms of the shelf ecosystem.

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