17

What Are Upwelling Systems Contributing to the Ocean's Carbon and Nutrient Budgets?

J.R. TOGGWEILER and S. CARSON
GFDL/NOAA, Princeton University, P.O. Box 308, Princeton, New Jersey 08542, U.S.A.

ABSTRACT

Understanding the effect of upwelling systems on the carbon cycle requires detailed knowledge of how nutrients and carbon enter and leave these systems. In this article we use recent findings of the JGOFS equatorial Pacific process study and a detailed three-dimensional model to look specifically at the nitrate budget in the equatorial Pacific. Nitrate enters the equatorial upwelling system in the far-western Pacific via the Equatorial Undercurrent. Because the equatorial biota tend to recycle nitrogen much more effectively than they export nitrogen in sinking particles, nitrate stocks build up in the eastern Pacific. A significant fraction of the nitrate entering the upwelling system seems to be lost to denitrification in the anoxic zones off Peru and Central America. Through denitrification, the equatorial upwelling system may function as a regulator of global nitrate stocks and air–sea partitioning of CO_2.

INTRODUCTION

The distribution of dissolved CO_2 in the ocean differs from that of ordinary salts because marine organisms fix CO_2 into particulate organic matter near the surface and transport it downward in the form of sinking particles. The fixation of organic carbon in surface waters, in combination with the remineralization of organic particles in the interior, contribute to a 10% CO_2 depletion in surface waters with respect to average deep water. The 10% depletion in surface CO_2 corresponds to a nearly 100% depletion in surface nutrients.

The removal of CO_2 from surface waters and the associated buildup of CO_2 in the deep ocean maintains a relatively low partial pressure of CO_2 in the atmosphere. Broecker (1982) showed that the amount of CO_2 retained by the deep ocean is dictated primarily by the temperature and nutrient content of deep-ocean waters: the temperature

Upwelling in the Ocean: Modern Processes and Ancient Records
Edited by C.P. Summerhayes, K.-C. Emeis, M.V. Angel, R.L. Smith, and B. Zeitzschel. Published 1995 John Wiley & Sons Ltd.

determines how much CO_2 newly formed deep waters can hold, while the nutrient content determines how much carbon is exported biologically from the surface to the deep ocean every time a parcel of deep water is upwelled to the surface.

Wind-driven upwelling systems might be considered as special places where the actions of local winds concentrate the final stage of an oceanwide upwelling process. Surface waters in upwelling systems are known to have high partial pressures of CO_2 which drive large CO_2 fluxes from the ocean to the atmosphere. Of course, CO_2 outgassed in one place is reabsorbed somewhere else. In the context of Broecker's simple model, upwelling systems should not affect the partitioning of carbon between the ocean and atmosphere because details about where upwelling occurs and rates of upwelling do not affect the temperature or the nutrient content of the deep ocean.

In this chapter, we approach the relevance of upwelling systems to the global carbon budget from a different perspective. We take a close look at the way in which upwelling occurs in the equatorial Pacific with the aid of a highly resolved three-dimensional model. Nutrient and carbon dynamics in this system bear little resemblance to the nutrient and carbon dynamics in conventional box models. First, the nutrients and CO_2 upwelled to the surface in this system are derived entirely from thermocline waters. There is very little input directly from the deep ocean. Second, the upwelling system does not export much organic material to the deep sea. Instead, the equatorial upwelling system accumulates a large quantity of regenerated nitrate and resorts to denitrification to eliminate a portion of the nitrate added to the system. We suggest that denitrification in the equatorial Pacific may affect the partitioning of carbon between the ocean and atmosphere by regulating the total amount of nitrate in the ocean. Our conclusions draw on some of the most recent findings to come out of the JGOFS equatorial Pacific process study.

Much of the discussion to follow is focused on nitrogen rather than carbon. This is mainly because nitrogen is easier to budget. Surface nitrate concentrations in the equatorial Pacific are high near the equator, but generally drop to zero within 5° of the equator on the north side and 10°–15° of the equator in the south. Surface waters advecting outward from this region carry very little nitrogen (as nitrate) out of the system. Nitrogen losses from this system occur mainly through biological processes which are constrained by the new JGOFS observations.

OVERVIEW OF THE EQUATORIAL PACIFIC UPWELLING SYSTEM

Easterly winds blowing along the equator in the equatorial Pacific create a divergence in the surface flow that leads to the upwelling of subsurface water. The most obvious geochemical manifestations of the upwelling are high levels of surface nitrate and pCO_2. Surface nitrate levels in the equatorial Pacific are extremely high, falling in the range of 4–10 $\mu mol\ l^{-1}$ over a large area. A less obvious characteristic is the high nutrient content of equatorial thermocline waters. Subsurface water between 5°S and

5°N, with temperatures warmer than 20°C, typically contains some 4–15 μmol NO_3 l^{-1}. Thermocline waters with the same temperature north and south of the equatorial region typically contain no nitrate or a few micromoles per liter at most.

A map of time-averaged surface nitrate concentrations shown in Figure 17.1 illustrates another characteristic of the equatorial upwelling system. West of 150°W, maximum nitrate concentrations are found along the equator, as one would expect from the surface divergence. However, the highest nitrate levels east of 150°W tend to be south of the equator, reaching maximum values west of Peru between 5° and 10°S. Maximum pCO_2s are also consistently found between the equator and 10°S between the longitudes 95° and 110°W (Murphy et al. 1991; Murray et al. 1994). Both the nitrate and pCO_2 fields suggest that the upwelling along the equator is somehow merged or blended with upwelled water originating off the coast of Peru.

The coastal region of Peru is well known as a place where high-nutrient, high-CO_2 water upwells to the surface, but it is not obvious why geochemical properties of Peruvian water should be linked with properties derived by upwelling in the center of the basin. The equator–Peru linkage in Figure 17.1 is not a product of surface advection. The northward wind stress along the coast of Peru generates about 4 Sv of coastal upwelling between 5° and 15°S (Wyrtki 1963). Surface currents south of the equator direct the upwelled water toward the west (Pazos and Umaran-Barrueto 1989). The drift of Peruvian surface waters toward the west is about 500 km per month over 0°–10°S. In order for the nitrate and CO_2 upwelled near Peru to reach the nitrate and pCO_2 maxima at 140°W, it would have to persist in the surface layer for at least a year. This is not likely as biological uptake and gas exchange would remove much of the initial nitrate and CO_2 over this time scale.

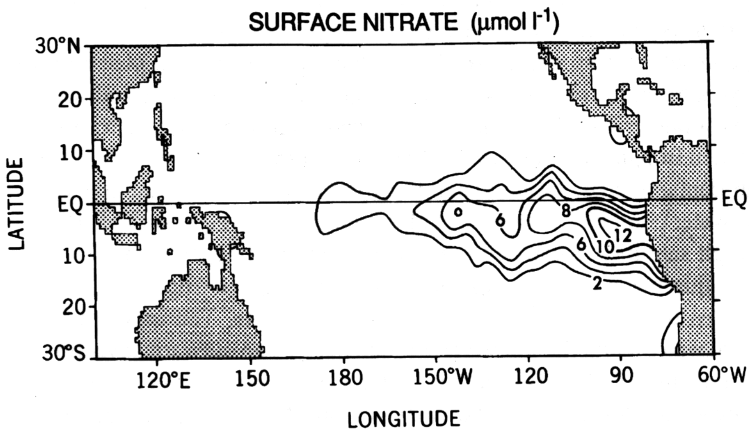


Figure 17.1 Distribution of nitrate in surface waters of the tropical Pacific Ocean (μmol l^{-1}). Data are from Levitus et al. (1993), average of 0 and 10 m.

North of the equator, the North Equatorial Countercurrent (NECC) flows into the equatorial zone from the west between 5° and 9°N. The Countercurrent is found in the latitude band of the Intertropical Convergence Zone, a region where the easterly wind stress (opposing the current) is weakest. The Countercurrent originates in the far-western Pacific where nutrient levels are near zero and salinities are low. The existence of the Countercurrent accounts for the relatively low nutrient and pCO_2 levels north of the equator.

SOURCES OF UPWELLED WATER

The divergence of water along the equator upwells water directly from the core layers of the Equatorial Undercurrent (EUC), which have temperatures between 19°–24°C and contain about 10–12 $\mu mol\ l^{-1}$ of nitrate. Water with this temperature or density outside of the equatorial region has essentially no nitrate, thus the nitrate in the core of the Undercurrent must be derived within the equatorial zone. As will be described below, nitrate reaches the core layers of the Undercurrent by a rather circuitous path from colder, deeper layers of the same current.

Ocean GCMs predict total upwelling fluxes of ~100 Sv along the equator (Philander et al. 1987). Upwelling rates are strongest in the central Pacific where the wind stress is strongest. Much of the water upwelled in the central Pacific downwells again within 5° of the equator and immediately recirculates back into the Undercurrent at depths of 50–150 m. This kind of shallow recirculation is fairly rapid with respect to biological uptake and gas exchange, so nitrate and CO_2 levels remain fairly homogeneous within the recirculation region. No nitrogen or carbon enters or leaves the upwelling system as the result of this kind of advection.

Bomb tracer observations have detected the penetration of cooler and more saline water into the equatorial upwelling system at slightly greater depths. Fine et al. (1987) show a core of tritiated water from the Northern Hemisphere penetrating into the upwelling zone with a density of about 25.0 in sigma theta. Equatorial water with this density is about 19°–20°C. Figure 17.2 shows cross-sections of $\Delta^{14}C$ and salinity at 150°W taken from Quay et al. (1983). This section shows the equatorward penetration of bomb-contaminated, saline water from both the north and south. Quay et al.'s high-^{14}C water coincides with Fine et al.'s high-tritium water north of the equator.

The high salinities associated with the tritium and bomb ^{14}C maxima point to an origin for this water in the evaporative zones of the subtropics where water with temperatures of ~20°C generally has very little nitrate. The input of high-salinity subtropical water to the equatorial zone is balanced by an outward advection of surface waters across 10°S and 10°N. Surface waters at 10°S and 10°N are depleted in nitrate. Thus, the advection of 20°C subtropical water into the equatorial zone and the outward flow balancing this input do not contribute very much to the nitrate budget. The input of subtropical water does contribute to CO_2 outgassing, as the pCO_2 of 20°C water is substantially increased when it is warmed to 24°–27°C.

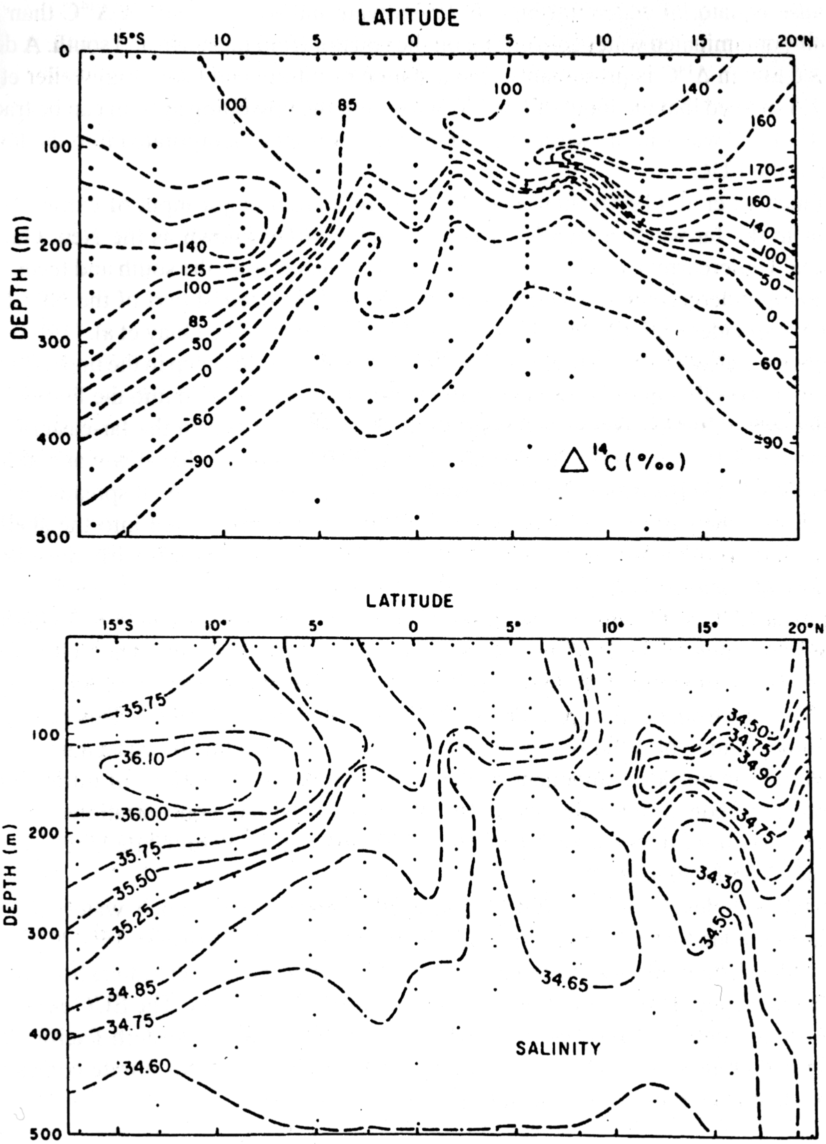


Figure 17.2 $\Delta^{14}C$ (top) and salinity (bottom) along 150°W in the equatorial Pacific showing the correlation between high salinity and bomb ^{14}C in 1979. Copied from Quay et al. (1983).

One of the major points raised by Quay et al. (1983) is that the equatorial ^{14}C budget requires an additional source of low-^{14}C water from deeper in the thermocline. This is because equatorial waters during the 1970s were uniformly lower in $\Delta^{14}C$ than the bomb-contaminated water converging on the equator from the north and south. A deep source low in $\Delta^{14}C$ is presumably also a source of nitrate and CO_2. Toggweiler et al. (1991) showed that the input of low-^{14}C water into the upwelling system can be traced to 11°–12°C water in the western Pacific which becomes incorporated into the lower part of the EUC.

Tsuchiya et al. (1990) show that the EUC receives a major input of extratropical water in the far-western Pacific. A coastal undercurrent known as the New Guinea Coastal Undercurrent, or NGCU, flows up to the equator from the south and feeds into the main undercurrent between 140°E and 155°E. The deepest part of the NGCU at 300 m coincides with the 11°–12°C water which figures in the Quay et al. (1983) and Toggweiler et al. (1991) analyses of the ^{14}C budget. Water this deep in the undercurrent does not upwell to the surface along the equator, but feeds instead into the Peru–Chile Undercurrent, which flows southward away from the equator on the far-eastern side of the basin (Lukas 1986). Toggweiler et al. (1991) suggest that this water finally upwells to the surface as 14° or 15°C water off the coast of Peru. As it spreads back to the west at the surface it is warmed up and becomes incorporated into the shallow recirculations of the main upwelling system. In this way, water from fairly deep in the thermocline makes its way into the core layers of the EUC.

Lower EUC water with temperatures of 11°–14°C and salinities of 34.8–35.0 forms a well-documented mode of the undercurrent known as "13° water." The EUC carries nearly 10 Sv of water with these properties to the east across 155°E, of which some 2–6 Sv come from outside the equatorial region (Tsuchiya et al. 1990). This is roughly the same volume of water that is removed from the thermocline by coastal upwelling off Peru. The outside component contains fairly high concentrations of nitrate when it enters the equatorial zone, about 16–20 $\mu mol\ l^{-1}$. Nitrate concentrations increase along the path of the flow such that 14°–15°C water in the Peru–Chile Undercurrent contains nearly 30 $\mu mol\ l^{-1}$.

Shallower levels of the NGCU transport about twice as much water into the equatorial zone (~8 Sv; Tsuchiya et al. 1990). This water is warmer, 15°–20°C, and contains substantially less nitrate, 7–10 $\mu mol\ l^{-1}$, than the deeper water; however, it contributes roughly the same amount of nitrate as the deeper flow. The EUC also picks up a small contribution of Northern Hemisphere water in the western Pacific. This water is high in nitrate, but it is not clear how much, if any, of this nitrate comes into the system from outside the equatorial region.

EXPORT OF NITROGEN AND CARBON IN SINKING PARTICLES

Recent JGOFS observations along 140°W show that the export of nitrogen and carbon in the form of sinking particles is rather small (Murray et al. 1994). Low particle fluxes in the equatorial Pacific are a consequence of an ecosystem dominated by small phytoplankton which are efficiently grazed by small zooplankton. Grazing holds down phytoplankton populations and limits the total amount of photosynthetic production. Zooplankton excrete dissolved nitrogen back into the water as ammonia. If ammonia is available, phytoplankton prefer ammonia as a nitrogen source rather than nitrate. This cycle — phytoplankton uptake, zooplankton grazing, ammonia excretion, and phytoplankton uptake — tends to create a nearly closed cycle, which reduces the amount of nitrate taken up from the water. It results in a fairly long residence time for upwelled nitrate within equatorial surface waters (about one year; McCarthy et al. 1995).

Table 17.1 shows an attempt to construct a nitrogen budget for the region between 5°S and 5°N. Primary production measurements show that roughly 15 mmol N m^{-2} d^{-1} are taken up by phytoplankton in the upwelling zone (Murray et al. 1994). Only 2–3 mmol N m^{-2} d^{-1} of the total N uptake is nitrate, i.e., new production (McCarthy et al. 1995; P. Wheeler, pers. comm.). The measured nitrate uptake agrees fairly well with the nitrate uptake in model simulations (see below). The loss of nitrogen in the form of particles sinking below 100–150 m is only 0.5–1.0 mmol N m^{-2} d^{-1}, i.e., only 1/6 to 1/2 of the nitrate uptake. The deficit of ^{234}Th in the upper 100 m is consistent with the N losses at the low end of the range (Buesseler et al., submitted), while direct sediment trap measurements tend toward the higher estimate (J. Newton, pers. comm.; J. Murray et al., in prep.).

The imbalance between nitrate uptake and the downward flux of sinking particles appears to be resolved through advection of dissolved organic matter (DOM; Murray et al. 1994). Production of DOM by the biota allows organic matter to advect away from the equator in fairly large quantities in the surface flow. Lateral advection of the nitrogenous components of DOM provides a sink for nitrate and explains how the nitrogen budget can be closed without a large downward flux of sinking particles.

The problem with this solution is that DOM has to be oxidized back to nitrate and CO_2 somewhere. If the lifetime of DOM is long enough, it can be exported to the adjacent subtropical and subarctic gyres before breaking down. A solution of this form has been advocated by several modeling studies (Toggweiler 1989; Bacastow and Maier-Reimer 1991; and Najjar et al. 1992). DOM produced in the equatorial zone would have to persist 5–10 years in order to be advected completely out of the tropical region in large quantities (i.e., across 10°S and 10°N).

Table 17.1 Equatorial Pacific budget, 5°S–5°N.

Primary production	15 mmol N m^{-2} d^{-1}
New production	
ρNO_3	2–3 mmol N m^{-2} d^{-1}
model	3.3 mmol N m^{-2} d^{-1}
Sinking particle flux	
^{234}Th deficit upper 100 m	0.5 mmol N m^{-2} d^{-1}
sediment traps at 150 m	1.0 mmol N m^{-2} d^{-1}
model at 144 m	2.4 mmol N m^{-2} d^{-1}

Recent studies have documented the buildup in DOC in near-surface waters over background levels. These studies suggest that newly produced DOC has a lifetime in the ocean of only several months (Copin-Montegut and Avril 1993; Amon and Benner 1994; Carlson et al. 1994). If the DOM lifetimes from these studies apply to the equatorial Pacific, then DOM cannot be advected very far. Nitrate and CO_2 remineralized from DOM will be regenerated close to the equator where regenerated nitrate will stimulate additional phytoplankton growth. Under this scenario, sinking particle fluxes are still needed to balance the input of nitrate from upwelling, but the area over which upwelling-stimulated particle fluxes might operate could be substantially larger. It is hard to tell from JGOFS observations whether such an expansion of area is realized.

A MODEL OF THE NITROGEN CYCLE IN THE EQUATORIAL PACIFIC

A nitrogen-cycle simulation derived from a coupled biological/physical model (S. Carson et al., in prep.) helps one visualize the source of nitrate in the undercurrent and the accumulation of regenerated nitrate in the eastern Pacific. The model consists of a seven-compartment ecosystem model (Fasham et al. 1990), which is embedded in a highly resolved circulation model of the tropical belt of the Pacific (Philander et al. 1987). We use the same ecosystem model and the same parameters employed by Sarmiento et al. (1993) in a nitrogen cycle model of the North Atlantic.

The ecosystem model contains compartments for nitrate, ammonia, phytoplankton, zooplankton, bacteria, DON, and detritus. Phytoplankton grow as a function of nutrient and light limitation. Phytoplankton are eaten by zooplankton, which excrete ammonia back into the water and egest sinking detrital particles. The model explicitly resolves new and regenerated production by the phytoplankton. The full set of ecosystem equations is solved over the first 12 layers of the model representing the upper 144 m. Below 144 m, a simplified set of equations is solved, providing for the remineralization of nitrate from detrital particles. For more details, see Sarmiento et al. (1993).

The DON compartment in the model represents a very short-lived food source for model bacteria. It does not in any way simulate the relatively large DON pool discussed above, which can be advected away from the equator over a period of months. This aspect of ecosystem dynamics has yet to be incorporated into the model.

The model's nitrate field is initialized with the annual mean field of Levitus et al. (1993) while other compartments are initialized with near-zero values. When the model begins to run, the nitrogen levels in the nonnitrate components (e.g., phytoplankton and zooplankton) grow in quickly. After a few years, the nitrate distribution in the upper few hundred meters evolves to a repeated seasonal cycle.

The physical model is forced with monthly mean surface winds and monthly mean surface heat fluxes. The model is quite sensitive to the drag coefficient used to convert surface winds to a stress on the ocean's surface. High values of the drag coefficient lead to stronger upwelling and higher concentrations of nitrate in near surface layers. Low values of the drag coefficient lead to weak upwelling and lower concentrations of nitrate. The illustrations below were produced with a drag coefficient of 1.4×10^{-3}. This level of drag maintains predicted nitrate levels close to those observed over 10 years of integration.

Upwelling from the Undercurrent

Figure 17.3 shows a section of current vectors (v,w) and nitrate along a meridional section at 140°W taken from the fifth year of the seasonal simulation. One can see, in Figure 17.3, how specific features of the meridional circulation are aligned with features of the nitrate distribution. Downwelling on both sides of the equator recirculates high-nitrate water between the surface and 100 m. South of the equator the shallow recirculation extends well beyond 5°S. Water converging on the equator as part of the extended recirculation advects low-nitrate water into the upwelling zone near 100 m. The extended meridional circulation actually produces a weak dilution of equatorial nitrate levels.

The main source of nitrate for the upwelling system is upwelling from below 100 m, where nitrate values are greater than 12 μmol l^{-1}. Water upwelling from these depths is coming directly from the undercurrent, which is flowing out of the page across the section. Figure 17.4 shows an east–west section of u,w vectors and nitrate concentrations along the equator. This illustrates the effect of the undercurrent more clearly. Water with progressively higher nitrate levels is upwelled from the undercurrent as the undercurrent shoals to the east.

Equatorial Nitrate Budget

In Figures 17.5–17.7, the model is used to develop a complete nitrogen budget for the equatorial zone. Figure 17.5 is a schematic diagram that illustrates how the budget will be presented in Figures 17.6 and 17.7. The equatorial zone is defined as the region bounded by 5°N and 5°S, and 90°W and the dateline. The region is divided into two boxes separated at 135°W. A budget is derived for 6 depth domains down to 400 m. The budget shows integrated nitrate transport across the faces of each box in the form of uN(x), where u, v, or w are the velocities and N(x), N(y), and N(z) are the nitrate concentrations at each face of the box. Units are 10^4 mol N s^{-1}. Nitrate removal caused by biological production and nitrate input as a result of remineralization of detrital nitrogen are also given in the same units. The budget also considers transports caused by mixing; however, these are very small in relation to the transports caused by advection.

Figure 17.6 shows nitrate transports over a series of five depth domains starting at 0–40 m and ending with 189–264 m. Labeled arrows across each face give the direction and magnitude of the transport, while the flattened numbers on the bottom of each box give the transport as a result of upwelling from the box below. The bold numbers above

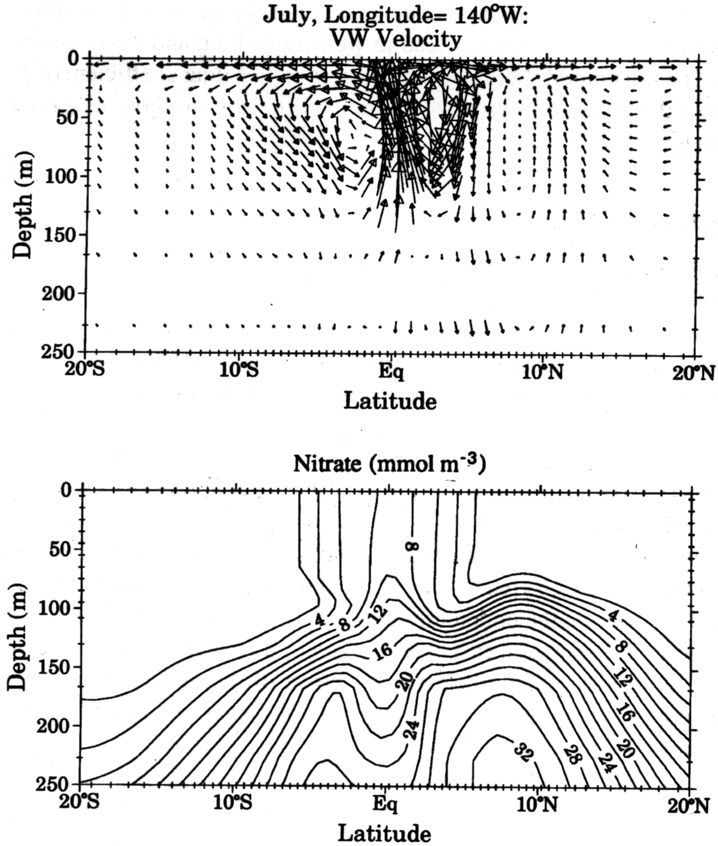


Figure 17.3 Correlation between model currents (plotted as v,w current vectors) and nitrate concentration along 140°W for climatological July conditions.

the top box indicate the amount of nitrate taken up by phytoplankton. In the eastern surface box, 32.5 units of nitrate come into the box as upwelling from the box below while 18.0 units are taken up by the phytoplankton. The difference is accounted for by zonal advection (8.3 units) into the western box and southward advection (6.1 units) across 5°S.

The main thing to note about Figure 17.6 is that the nitrate budget is dominated by zonal advection and upwelling. Meridional transports are not nearly as large. Figure 17.7 redraws Figure 17.6 as a two-dimensional figure, ignoring the meridional components. The main transport terms (those greater than 10×10^4 mol N s^{-1}) are highlighted with the bold arrows. One immediately sees how the transport is dominated by flow into the region from the west at depths below 150 m and by the shoaling of

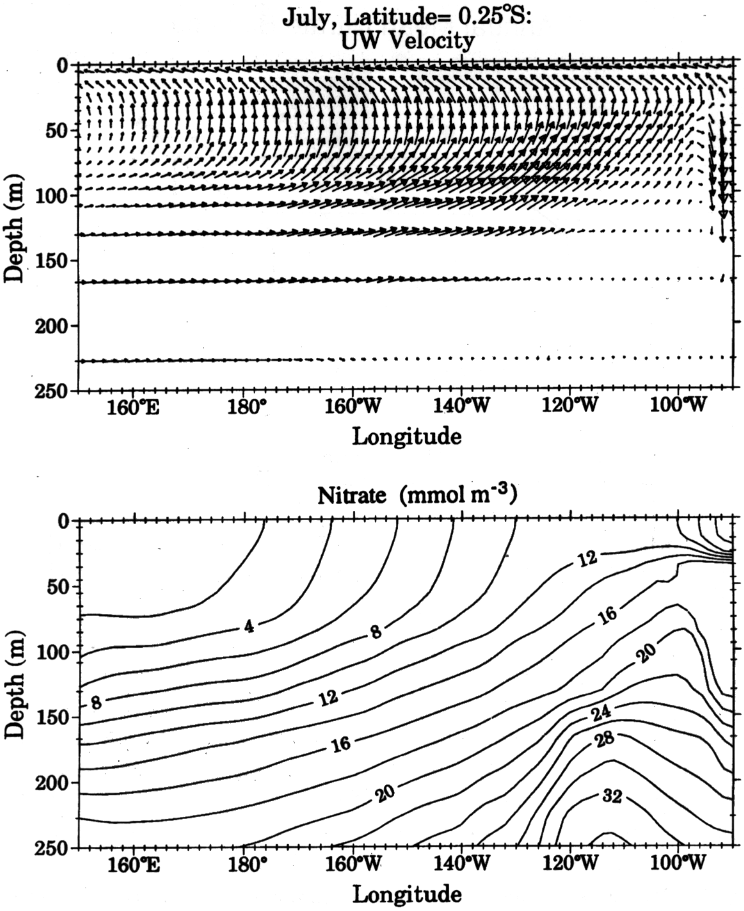


Figure 17.4 Correlation between model currents (plotted as u,w current vectors) and nitrate concentration along the equator for climatological July conditions.

that flow across the section. This illustrates in a more quantitative fashion what was shown in Figure 17.4, i.e., the main source of nitrate to the equatorial upwelling system is the lower part of the undercurrent flowing into the region from the west. The core of the undercurrent at the dateline lies between 100 m and 170 m, but the largest nitrate transport, 29.5×10^4 mol s^{-1}, is below 190 m where nitrate concentrations are higher.

Figure 17.7 includes the biological uptake/remineralization terms at all depth levels. These are the small numbers in the upper lefthand corner of each box. The biological terms are negative in the upper three boxes and become positive below as nitrate is remineralized from sinking particles. The mean nitrate concentration in each box is

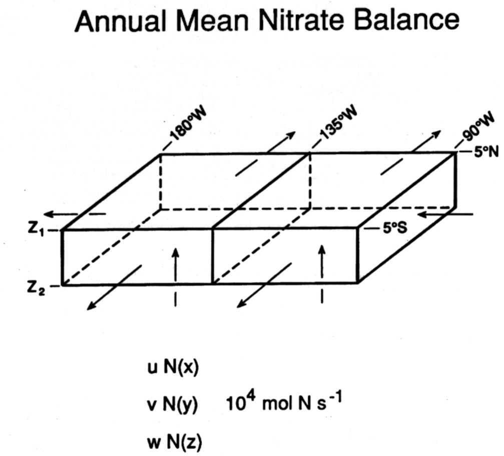


Figure 17.5 Schematic showing presentation of nitrate transports in Figure 17.6 and 17.7. Model is used to compute nitrate transports across each face of the boxes shown for the region between 5°S, 5°N, 90°W, and the dateline.

given by the italicized numbers in the lower lefthand corner of each box. The nitrate concentration of the flow crossing the dateline into the three deep western boxes is also given along the lefthand side of the figure.

An important thing to note about Figure 17.7 is that relatively low-nitrate water flows into boxes with higher nitrate concentrations. Water flowing into the western box between 189 m and 264 m, for example, has a nitrate concentration of 20 μmol l^{-1}. Water in the box has a mean concentration of 24 μmol l^{-1}. Thus the large incoming flow of 29.5×10^4 mol s^{-1} actually dilutes nitrate levels in the box. This points out a very important and ironic feature of the upwelling system: the undercurrent is clearly the main source of nitrate to the upwelling system, but it also brings in relatively low-nitrate water.

The amount of nitrate added to the euphotic zone by upwelling over 5°S to 5°N and 90° to 180°W is 59×10^4 mol s^{-1}. This is equivalent to an average input of 4.6 mmol N m^{-2} d^{-1} in the units of Table 17.1. About 30% of the upwelled nitrate is advected laterally across the edges of the budget region. The total amount of nitrate taken up by the model phytoplankton is 42×10^4 mol s^{-1}, or 3.3 mmol N m^{-2} d^{-1}. This is a bit larger than the new production measured during the JGOFS study, 2–3 mmol N m^{-2} d^{-1}.

The average sinking flux of detrital nitrogen in the model is 2.4 mmol N m^{-2} d^{-1} at 144 m. This accounts for most of the nitrate uptake, i.e., 2.4 out of 3.3 mmol N m^{-2} d^{-1}. The remainder consists of outward advection of the living components of the ecosystem plus ammonia. As pointed out above, the observed sinking flux is only

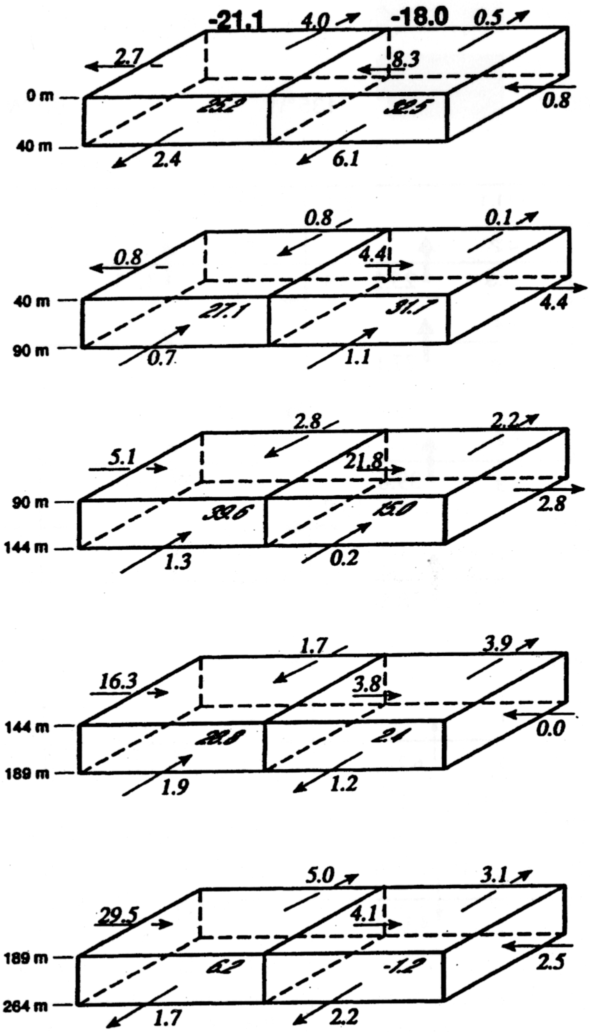


Figure 17.6 Three-dimensional representation of annual-mean nitrate transports due to advection in the upper 264 m of the tropical Pacific. Units 10^4 mol N s^{-1}. Flat numbers on bottom face of each box give the vertical nitrate flux from the box below.

0.5–1.0 mmol N m^{-2} d^{-1} along 140°W. Thus the model's sinking flux seems to be too high by a factor of 2.5x–5x.

When ammonia and/or nitrate concentrations exceed the half-saturation constant for phytoplankton N uptake (0.5 μmol l^{-1}), and light is not limiting, phytoplankton and zooplankton tend toward a quasi-stable equilibrium with relatively small populations.

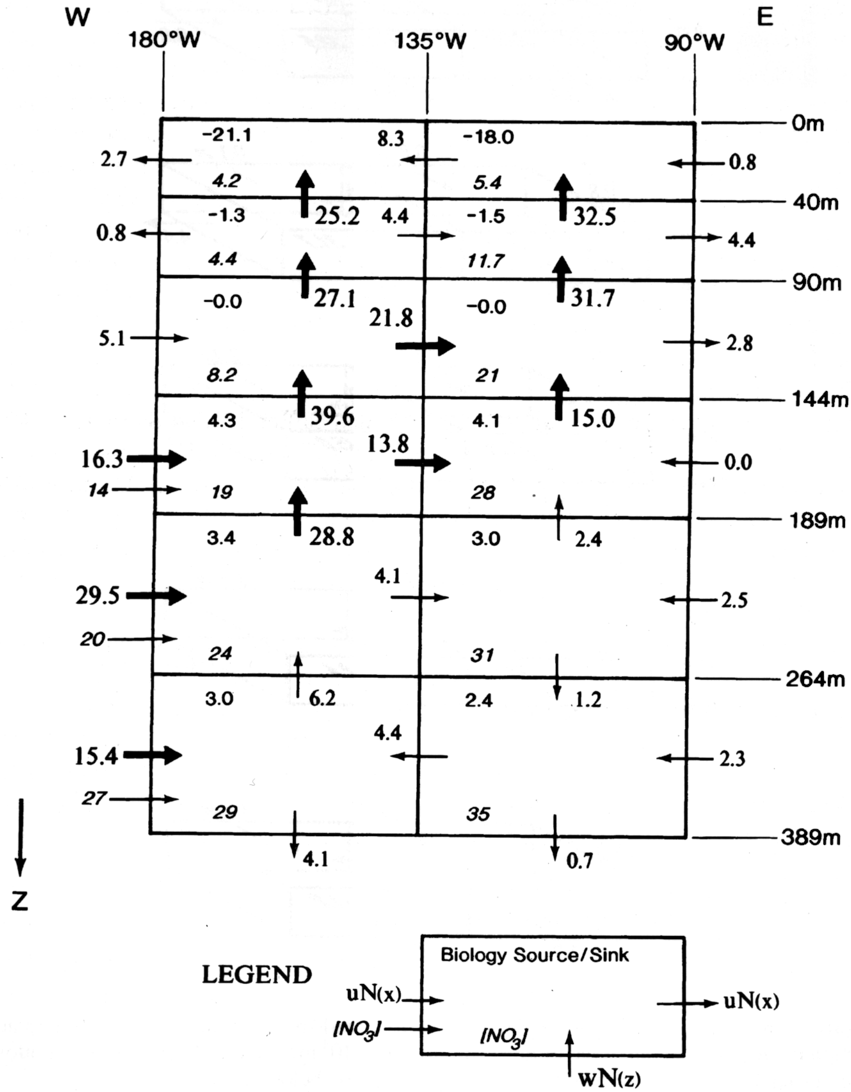


Figure 17.7 Two-dimensional representation of annual-mean nitrate transports for the upper 400 m. Meridional contributions are ignored. Small numbers in upper lefthand corner of each box give source/sink terms for nitrate caused by biological uptake or remineralization. Italicized numbers in the lower lefthand corner give nitrate concentrations averaged over the whole box.

This puts an upper limit on the potential sinking particle flux per unit area. The upper limit is much smaller than the input of nitrate right at the equator. The model maintains a balanced budget by allowing the area of high-nitrate surface waters to expand away from the equator until the areally integrated sinking flux balances the input. The model accumulates a large standing stock of nitrate in the upper 200–300 m within that area until the patch of surface nitrate is big enough to produce an adequate sinking loss.

Nitrate transports have also been analyzed in a 5°S to 5°N box, which extends from the dateline out to 135°E. This box includes the area of the western Pacific where the EUC originates (Tsuchiya et al. 1990). About 60×10^4 mol N come into this box every second and exit across the dateline, as shown in Figure 17.7. Inputs to the western box occur as roughly equal contributions from the Northern and Southern hemispheres. This equal north–south partitioning does not concur with the sense of the Tsuchiya et al. analysis, which shows that most of the outside source waters for the EUC come from the Southern Hemisphere.

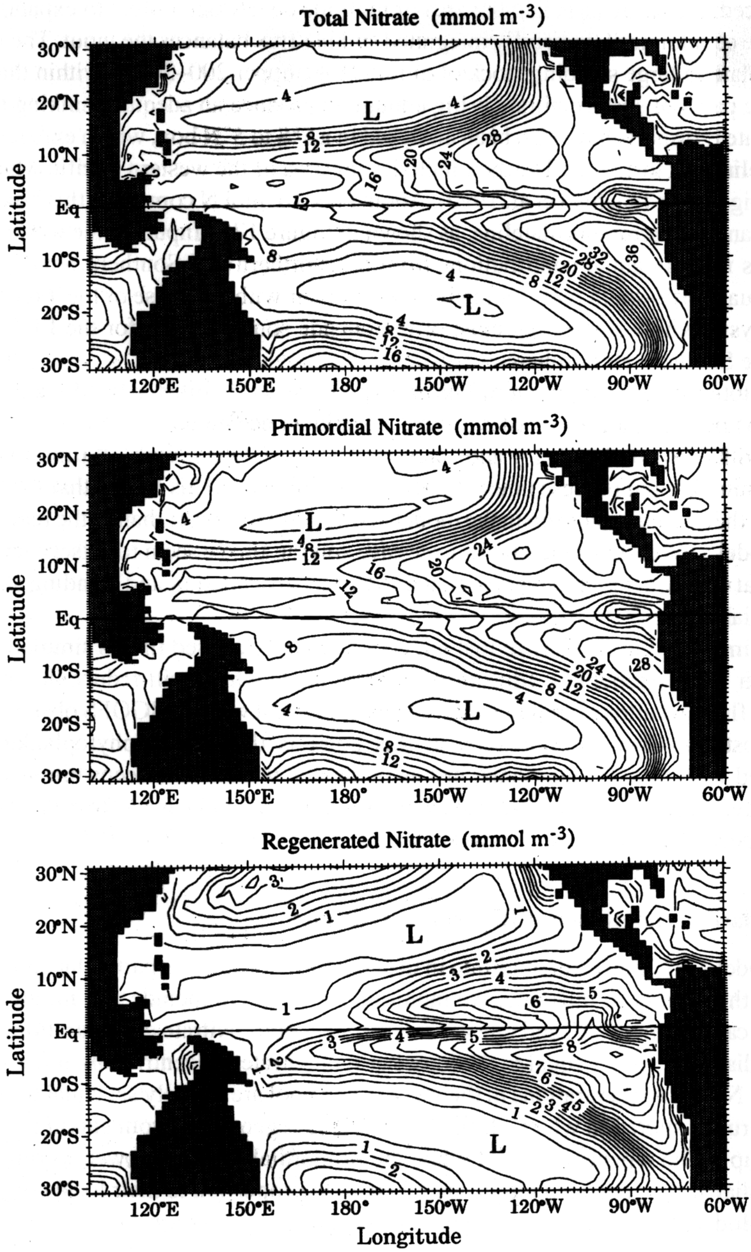
The total input to the western box is also excessive. Estimates for the northward transport of nitrate in the coastal undercurrent cited earlier, i.e., ~4 Sv of 11°–14°C water with ~18 μmol NO_3 l^{-1}, and ~8 Sv of 15°–20°C water with ~8 μmol NO_3 l^{-1}, yield a nitrate input in these units of only 14×10^4 mol s^{-1}. This means that the nitrate input to the upwelling system in the model is several times larger than in the real world. The model's excessive input and its excessive sinking fluxes seem to offset each other such that the model maintains a reasonable surface distribution and standing stock of equatorial nitrate.

It is important to emphasize that the ecosystem model used in this simulation has not been tuned. Hopefully, adjustments can be made which will reduce the model's sinking flux and nitrate input in ways that are consistent with the JGOFS observations. The most important consideration at this point is that model phytoplankton are saturated with recycled ammonia, just as in the real world. This keeps nitrate uptake relatively low and forces high-nitrate water to spread laterally away from the equator. This aspect of the simulation certainly seems realistic.

Subsurface Accumulation of Regenerated Nitrate

The model's accumulation of nitrate in the equatorial zone has a particular subsurface pattern that is easy to illustrate. We modified the model to include two nitrate pools: one we call "primordial," and one we call "regenerated." Nitrate concentrations in the primordial pool are set equal to the initial or total nitrate concentrations at the start of the run. Nitrate concentrations in the regenerated pool are initially set equal to zero. A special run is started in which all the nitrate being regenerated from sinking particles builds up in the regenerated pool while the primordial pool is allowed to run down. Primordial and regenerated nitrate are taken up in proportion to their local concentrations. Model phytoplankton are totally unaware of the separation.

Figure 17.8 shows maps of primordial and regenerated nitrate at a depth of 166 m after three years of the special run. Stocks of primordial nitrate seem only slightly


Total Nitrate (mmol m-3)
Primordial Nitrate (mmol m-3)
Regenerated Nitrate (mmol m-3)
Latitude
30°N
20°N
10°N
Eq
10°S
20°S
30°S
120°E
150°E
180°
150°W
120°W
90°W
60°W
Longitude

altered from the initial state. Regenerated nitrate builds up in two lobes that extend westward from the eastern Pacific just north and south of the equator. The buildup of regenerated nitrate is especially strong south of the equator. In only three years, up to 9 μmol l^{-1} of regenerated nitrate have accumulated at 5°S off the coast of Peru. Concentrations of regenerated nitrate in the east are steadily rising as the run continues beyond three years. An extrapolation of the trend suggests that most of the nitrate in the east will have been converted into regenerated nitrate within 15 or 20 years. In the west, concentrations of regenerated nitrate level off at 10% or 15% of total nitrate within a few years.

Along the equator, between the two lobes of regenerated nitrate in Figure 17.8, is a minimum marking the undercurrent. The undercurrent brings in water from the west that is relatively pristine, i.e., it contains very little regenerated nitrate and is dominated by primordial nitrate advected into the system from outside. Concentrations of regenerated nitrate increase strongly toward the east as the undercurrent picks up regenerated nitrate from the off-equator lobes. The undercurrent carries this nitrate back to the east where it upwells again and cycles through the biota once more.

The model's distribution of regenerated nitrate looks very similar to the observed subsurface distribution of oxygen. Tsuchiya (1968) shows that the undercurrent is a high-oxygen tongue which penetrates toward the east from the western Pacific. Just off the equator to the north and south are tongues of low oxygen in which the intensity of oxygen depletion increases toward the east. The water just off the equator accumulates the effects of organic remineralization in the form of low-oxygen levels in the same way that the model accumulates regenerated nitrate. The accumulation is especially strong in the east because the eastern Pacific is the most remote part of the basin with respect to the undercurrent oxygen source.

The distribution of regenerated nitrate in Figure 17.8 may hold the key to understanding the surface nitrate distribution in Figure 17.1. The surface nitrate maximum south of the equator is probably a surface expression of the subsurface accumulation of regenerated nitrate. The system seems primed to express this effect south of the equator: nitrate accumulation right at the equator is limited by the upwelling of relatively pristine low-nitrate water from the undercurrent, while accumulation north of the equator is limited by the presence of the NECC. Any doming of isopycnals south

---

Figure 17.8 Results of a special model run in which nitrate regenerated from sinking particles is allowed to accumulate as a separate variable, called "regenerated nitrate." Regenerated nitrate field is initially zero everywhere. "Primordial" nitrate is set equal to the initial or "total nitrate" field at the beginning of the run and is allowed to decline as nitrate is taken up by model biota. Panels show total nitrate, primordial nitrate, and regenerated nitrate three years after separation commences. Note that contour interval for regenerated nitrate is 0.5 μmol l^{-1} compared to 2.0 μmol l^{-1} for other panels.

of the equator, as a result of a negative wind-stress curl, would tend to lift this accumulated nitrate up to the surface.

DISCUSSION

Observational constraints suggest that most of the nitrate coming into the equatorial Pacific upwelling system enters the system as part of the EUC. This is clearly the case in our model simulation. The model has no input of upwelled nitrate from the deep sea and virtually no input via vertical mixing. Both observations and model results indicate that the lower part of the undercurrent in the western Pacific carries a high proportion of nitrate, which comes from outside the equatorial region.

Regenerated nitrate accumulates in the eastern half of the region, especially north and south of the equator. Nitrate accumulation is so widespread that the pool of nitrate associated with upwelling along the Peruvian margin merges with the pool of nitrate upwelling along the equator. The fact that this extensive accumulation is fed by a fairly small input of outside nitrate is testimony to the system's tendency to retain nitrogen.

Accumulation of nutrients in upwelling systems has been discussed previously by Redfield et al. (1963). According to them, nutrient accumulation is caused by an "estuarine" physical circulation in which water upwells near a coast and moves offshore as part of a surface Ekman layer. Nutrients are stripped from the offshore flow and are regenerated in a shallow onshore flow which feeds the upwelling at the coast. Nutrient accumulation is enhanced by shelves which trap sinking organic matter and allow it to be regenerated within the onshore flow.

In our model, regenerated nitrate accumulates north and south of the equator and eventually cycles back into the upwelling system, so the physical circulation contributes to the nitrate accumulation. However in a larger perspective, the accumulation of nitrate seems more properly attributed to the ecosystem itself. If organisms in the equatorial Pacific were larger and capable of exporting more material to the deep sea, then nitrate would be regenerated at depths where it could not be cycled back into the undercurrent.

Below we would like to speculate about how the accumulation of nutrients in the equatorial Pacific upwelling system fits into a global picture. There are two additional components to consider. The first concerns the ultimate source of the nitrate fed into the undercurrent in the western Pacific. The second component is denitrification.

Source of Nitrate to the Undercurrent

The deepest water feeding into the EUC in the western Pacific had no tritium or bomb ^{14}C at the time of the GEOSECS expedition in 1974. Its $\Delta^{14}C$ at the time was very similar to the $\Delta^{14}C$ in prebomb coral rings from the Galápagos Islands (Druffel 1981) and the $\Delta^{14}C$ in 9°–10°C pre-bomb thermocline water in the South Pacific. This led

Toggweiler et al. (1991) to suggest that the low-^{14}C water feeding the undercurrent has its origin in subantarctic mode waters (7°–10°C) forming in the convergence zone north of the Antarctic Circumpolar Current.

Subantarctic mode waters forming today in the South Pacific start with 10–15 μmol NO_3 l^{-1} (see Table 5 in Toggweiler et al. 1991). Subantarctic mode waters have low $\Delta^{14}C$ and relatively high nitrate levels because they are composed, in part, of Antarctic surface waters that have advected northward under the influence of the circumpolar Westerlies. Antarctic surface waters are composed of water upwelled directly from the deep sea (Toggweiler 1994). Thus, the outside nutrients coming into the equatorial upwelling system are ultimately derived from the deep sea, but they enter the system only after passing through the surface regions in the far south. Exposure of upwelled deep water to biological activity in Antarctic and subantarctic surface waters removes some of the initial nitrate while gas exchange raises the $\Delta^{14}C$. This would seem to be a critical passage: changes in the amount of deep-sea nitrate passed into the South Pacific thermocline could have a dramatic effect on the equatorial system downstream.

Denitrification

Besides being the main source of nitrate to the equatorial upwelling system, the EUC is also a major source of oxygen to the subsurface layers of the region (Tsuchiya 1968). Low-oxygen levels can be expected in areas remote from the high-oxygen waters of the undercurrent. Between 10° and 20°S off the coast of Peru, and between 10° and 15°N off Central America, oxygen levels fall to near-zero levels. These are regions where denitrifying bacteria convert nitrate (NO_3) into $N_2(g)$ in order to procure oxygen to oxidize organic matter. The production of N_2 as the end product of denitrification leads to a loss of oceanic nitrate to the atmosphere.

Figure 17.9 shows the location of the zero-oxygen water off Peru, as illustrated in Codispoti and Christensen (1985). Oxygen completely disappears along the coast between 150–400 m, i.e., between the 9° and 13°C isotherms. Denitrification is most notable in the upper part of the low-oxygen layer, where nitrate deficits in excess of 20 μmol l^{-1} are observed in water with temperatures of about 13°C. Codispoti and Christensen (1985) and Codispoti (1989) have estimated that 50×10^{12} g N in the form of nitrate are consumed in the anoxic areas off Peru and Central America every year by denitrification. This amounts to about 11×10^4 mol N s^{-1} in the units of Figures 17.6 and 17.7.

The map of regenerated nitrate in Figure 17.8 suggests that much of the nitrate being consumed off South and Central America could be nitrate that came into the equatorial upwelling system via the EUC and was cycled through the biota at least once. Earlier we showed that roughly 14×10^4 mol of outside nitrate are added to the lower EUC every second, thus the amount of nitrogen lost by denitrification in the east, 11×10^4 mol s^{-1}, is similar to the amount being added in the west. The denitrification sink can also be compared with the loss of nitrogen in particles that sink below a depth where

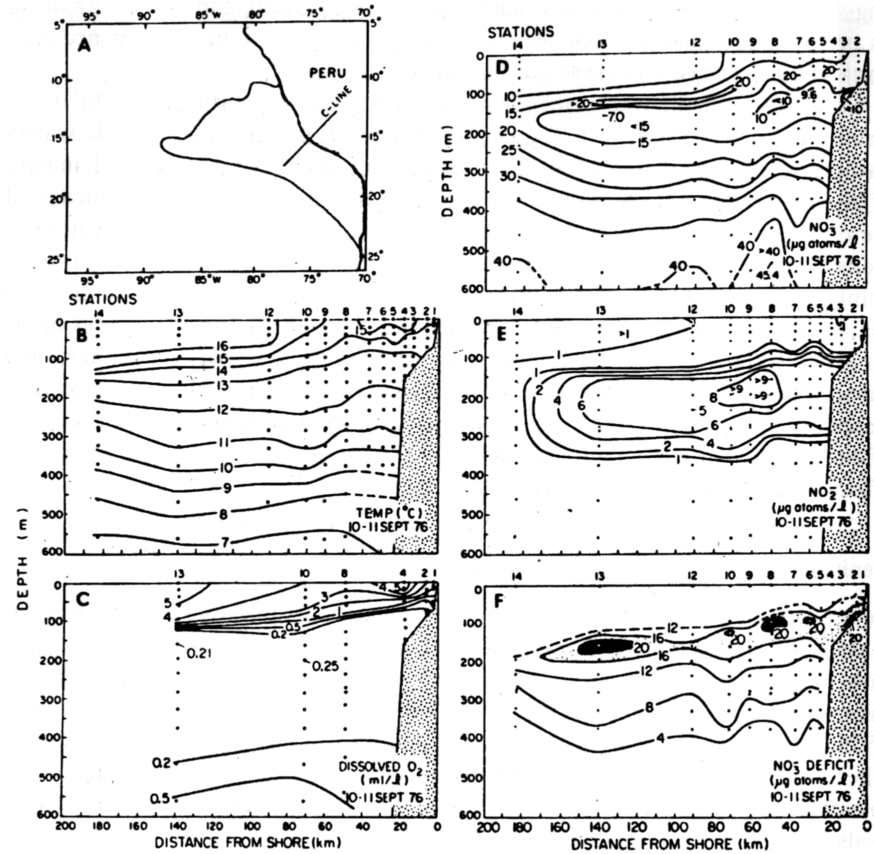


Figure 17.9 Location of denitrification region off southern Peru plotted along with water properties on a section perpendicular to the coast. Copied from Codispoti and Christensen (1985).

they no longer remain in the system. We estimate this loss very crudely as follows: we assume that nitrogen sinking below the 11°C isotherm, ~300 m, is lost from the upwelling system. We estimate this flux to be 0.3 mol N m^{-2} d^{-1} operating over an area of 3.3×10^{13} m^2, i.e., three times the area of the 5°S to 5°N boxes used in Figures 17.6 and 17.7. This amounts to a flux of 12×10^4 mol s^{-1}. Thus the particle sink is probably very similar to the denitrification sink.

Our effort to construct a nitrogen budget at this scale is very sketchy, but it shows that denitrification is probably a major loss term for the upwelling system. Denitrification occurs in relatively shallow thermocline layers which are ventilated by water from the undercurrent. The oxygen supply for this part of the thermocline is fixed by the volume of high-oxygen water brought into the undercurrent from outside the upwelling system, but the oxygen demand could be quite variable. The oxygen demand would depend on the concentration of nitrate passed into the South Pacific thermocline, i.e., the preformed nitrate, and the amount of equatorial organic matter that is remineralized in the upper few hundred meters. Thus the extent of denitrification should be linked to the nitrate input and the degree of nitrogen retention within the system.

Global Nitrogen and Global Carbon

Codispoti and Christensen's (1985) 50×10^{12} g N y^{-1} nitrogen loss to denitrification in the tropical Pacific is the ocean's largest nitrogen sink. It accounts for one-third of the ocean's total nitrogen loss. If one assumes that the rate by which new nitrate is added to the system (mainly by nitrogen fixation) is independent of the rate at which it is lost, then changes in the equatorial upwelling system could increase or decrease the total amount of nitrate in the ocean.

One way that this could happen is through a change in the nitrate content of newly formed mode waters in the South Pacific. According to Howard and Prell (1992), the fronts or convergence zones associated with the Antarctic Circumpolar Current shift position on glacial to interglacial time scales. During glacial times, these fronts were about 5° north of their present position. Shifts in frontal position are caused by shifts in the position of the circumpolar Westerlies (Klink and Smith 1993). Howard (1992) shows that subantarctic surface waters become heavier in $\delta^{13}C$ when the fronts are in a northern position. This would be consistent with stronger nitrate removal between the time when deep waters are raised to the surface around Antarctica and the time when these surface waters become subducted into the thermocline further north. Stronger nitrate removal and heavier $\delta^{13}C$ in the far south would be consistent with less nitrate being fed into the equatorial Pacific upwelling system and presumably lower denitrification rates.

If the loss of nitrate from the ocean is reduced by a northward shift of circumpolar winds during glacial time, then the nitrate content of the deep ocean would have to increase until higher denitrification rates are stimulated in some other way. According to the mechanism of Broecker (1982), higher nitrate levels in the deep sea would increase the amount of carbon lost to the deep sea each time a parcel of deep water is brought to the surface. This has the potential to change the partitioning of CO_2 between the atmosphere and the deep sea in a way that might help explain the low atmospheric pCO_2 during glacial time.

Broecker's (1982) original scenario for glacial pCO_2 invoked high deep-sea phosphate levels caused by lower sea level. Continental shelves, which today capture much of the phosphate coming into the ocean via rivers, were uncovered during glacial time. Broecker proposed that a resumption of shelf sediment deposition following a rise in sea level could cause a rapid removal of phosphate from the deep ocean and a rapid increase in atmospheric CO_2. Broecker's mechanism lost favor when it became clear that the rise in atmospheric pCO_2 during the most recent glacial termination preceded the rise in sea level.

The mystery of low glacial pCO_2 remains. We end our speculations by noting that abrupt southward movements of the Southern Ocean fronts tend to be contemporaneous with the early stage of glacial–interglacial transitions (Howard and Prell 1992; Imbrie et al. 1992). To the extent that meridional displacement of the fronts is causally related to nitrate concentrations in the thermocline, a southward movement would increase the amount of nitrate fed into the equatorial Pacific upwelling system and might produce significant reductions in global nitrate stocks over a few thousand years. Perhaps Broecker's original mechanism needs a second look with nitrate as the primary variable.

SUMMARY

The main input of external nitrate into the equatorial Pacific upwelling system occurs when lower thermocline water from the Southern Hemisphere feeds into the EUC in the far-western Pacific. The equatorial ecosystem is characteristically inefficient at converting upwelled nitrate into sinking particles. The input of external nitrate accumulates into a huge standing stock of regenerated nitrate in the shallow layers of the equatorial thermocline. The undercurrent is also the main oxygen source for the subsurface layers in which sinking particles are consumed. Particle remineralization in areas remote from the undercurrent leads to oxygen depletion and the onset of denitrification.

The large amount of nitrate lost to denitrification in the equatorial Pacific may be dictated by relatively high nitrate concentrations in the subantarctic waters which are subducted into the thermocline in the high latitudes of the South Pacific. During glacial time, the nitrate concentration of water subducted into the South Pacific thermocline seems to have been reduced. We hypothesize that this led to a reduction in denitrification rates in the equatorial Pacific, which allowed nitrate levels to increase globally. Higher nitrate levels may have contributed to lower CO_2 levels in the glacial atmosphere.

ACKNOWLEDGMENTS

We gratefully acknowledge the contributions of Rick Slater and Jorge Sarmiento, who set up the equatorial model back in 1990. We would like to thank Andy Watson, Steve Smith, Bonnie Samuels, and Rob Armstrong for their reviews of the manuscript. We would also like to thank Cathy Raphael and Jeff Varanyak for their help with Figures 17.5, 17.6, and 17.7, and Wendy Marshall for her help in preparing the text. Partial support for this work was provided by NOAA's Climate and Global Change Program. This paper is U.S. JGOFS contribution No. 153.

REFERENCES

Amon, R.M.W., and R. Benner. 1994. Rapid cycling of high-molecular-weight dissolved organic matter in the ocean. *Nature* **369:**549–552.

Bacastow, R., and E. Maier-Reimer. 1991. Dissolved organic carbon in modeling oceanic new production. *Glob. Biogeochem. Cycles* **5:**71–85.

Broecker, W.S. 1982. Ocean chemistry during glacial time. *Geochim. Cosmochim. Acta* **46:**1689–1705.

Buesseler, K.O., J.A. Andrews, M.C. Hartman, R. Belastock, and F. Chai. Regional estimates of the export flux of particulate organic carbon derived from ^{234}Th during the JGOFS EqPac program. *Deep-Sea Res.*, in press.

Carlson, C.A., H.W. Ducklow, and A.F. Michaels. 1994. Annual flux of dissolved organic carbon from the euphotic zone in the northwestern Sargasso Sea. *Nature* **371:**405–408.

Codispoti, L.A. 1989. Phosphorus vs. nitrogen limitation of new and export production, In: Productivity of the Ocean: Present and Past, ed. W.H. Berger, V.S. Smetacek, and G. Wefer, Dahlem Workshop Rept. LS 44, pp. 377–394. Chichester: Wiley.

Codispoti, L.A., and J.P. Christensen. 1985. Nitrification, denitrification, and nitrous oxide cycling in the eastern tropical South Pacific Ocean. *Mar. Chem.* **16:**277–300.

Copin-Montegut, G., and B. Avril. 1993. Vertical distribution and temporal variation of dissolved organic carbon in the northwestern Mediterranean Sea. *Deep-Sea Res.* **40:**1963–1972.

Druffel, E.R.M. 1981. Radiocarbon in annual coral rings from the eastern tropical Pacific Ocean, *Geophys. Res. Lett.* **8:**59–62.

Fasham, M.J.R., H.W. Ducklow, and S.M. McKelvie. 1990. A nitrogen-based model of plankton dynamics in the oceanic mixed layer. *J. Mar. Res.* **48:**591–639.

Fine, R.A., W.H. Peterson, and H.G. Ostlund. 1987. The penetration of tritium into the tropical Pacific. *J. Phys. Oceanogr.* **17:**553–564.

Howard, W.R. 1992. Late Quaternary paleoceanography of the Southern Ocean, Ph.D. diss. Brown University, Providence, RI, 345 pp.

Howard, W.R., and W.L. Prell. 1992. Late Quaternary surface circulation of the southern Indian Ocean and its relationship to orbital variations. *Paleoceanography* **7:**79–117.

Imbrie, J., E.A. Boyle, S.C. Clemens, A. Duffy, W.R. Howard, G. Kukla, J. Kutzbach, D.G. Martinson, A. McIntyre, A.C. Mix, B. Molfino, J.J. Morley, L. C. Peterson, N.G. Pisias, W.L. Prell, M.E. Raymo, N.J. Shackleton, and J.R. Toggweiler. 1992. On the structure and origin of major glaciation cycles 1. Linear responses to Milankovitch forcing. *Paleoceanography* **7:**701–738.

Klink, J.M., and D.A. Smith. 1993. Effect of wind changes during the last glacial maximum on the circulation in the Southern Ocean. *Paleoceanography* **8:**427–433.

Levitus, S., M.E. Conkright, J.L. Reid, R. Najjar, and A. Mantyla. 1993. Distributions of nitrate, phosphate, and silicate in the world oceans. *Prog. Oceanogr.* **31:**245–273.

Lukas, R. 1986. The termination of the Equatorial Undercurrent in the Eastern Pacific. *Prog. Oceanogr.* **16:**63–90.

McCarthy, J.J., C. Garside, and J.L. Nevins. 1995. "New" production along 140°W in the equatorial Pacific during and following the 1992 El Niño event. *Deep-Sea Res.*, in press.

Murphy, P.P., R.A. Feely, R.H. Gammon, D.E. Harrison, K.C. Kelly, and L.S. Waterman. 1991. Assessment of the air-sea exchange of CO_2 in the South Pacific during austral autumn. *J. Geophys. Res.* **96:**20,455–20,465.

Murray, J.W., R.T. Barber, M.P. Bacon, M.R. Roman, and R.A. Feely. 1994, Physical and biological controls on carbon cycling in the equatorial Pacific. *Science* **266:**58–65.

Najjar, R.G., J.L. Sarmiento, and J.R. Toggweiler. 1992. Downward transport and fate of organic matter in the ocean: Simulation with a general circulation model. *Glob. Biogeochem. Cycles* **6**:45–76.

Quay, P.D., M. Stuiver, and W.S. Broecker. 1983. Upwelling rates for the equatorial Pacific Ocean derived from the bomb ^{14}C distribution. *J. Mar. Res.* **41**:769–792.

Pazos, M.C., and J.P. Umaran-Barrueto. 1989. Drifting buoy data from the eastern equatorial Pacific for the period June 1, 1985 through June 30, 1987. NOAA Data Rept. ERL AOML16, Atlantic Oceanographic and Meteorological Laboratory, Miami, FL, 254 pp.

Philander, S. G.H., W.J. Hurlin, and A.D. Seigel. 1987. Simulation of the seasonal cycle of the tropical Pacific Ocean. *J. Phys. Oceanogr.* **17**:1986–2002.

Redfield, A.C., B.H. Ketchum, and F.A. Richards. 1963, The influence of organisms on the composition of sea-water, In: The Sea, Vol. 2., ed. M.N. Hill, pp. 26–77. New York: Wiley.

Sarmiento, J.L., R.D. Slater, M.J.R. Fasham, H.W. Ducklow, J.R. Toggweiler, and G.T. Evans. 1993. A seasonal three-dimensional ecosystem model of nitrogen cycling in the North Atlantic euphotic zone. *Glob. Biogeochem. Cycles* **7**:417–450.

Toggweiler, J.R. 1989. Is the downward dissolved organic matter (DOM) flux important in carbon transport?, In: Productivity of the Ocean: Present and Past, ed. W.H. Berger, V.S. Smetacek, and G. Wefer, Dahlem Workshop Rept. LS 44, pp. 65–83. Chichester: Wiley.

Toggweiler, J.R. 1994. The ocean's overturning circulation. *Physics Today* **47(11)**:45–50.

Toggweiler, J.R., K. Dixon, and W.S. Broecker. 1991, The Peru upwelling and the ventilation of the South Pacific thermocline. *J. Geophys. Res.* **96**:20,467–20,497.

Tsuchiya, M. 1968. Upper Waters of the Intertropical Pacific Ocean, Johns Hopkins Oceanographic Studies, 4. Baltimore: John Hopkins Univ. Press.

Tsuchiya, M., R. Lukas, R.A. Fine, E. Firing, and E. Lindstrom. 1990. Source waters of the Pacific Equatorial Undercurrent. *Prog. Oceanogr.* **23**:101–147.

Wyrtki, K. 1963. The horizontal and vertical field of motion in the Peru Current. *Bull. Scripps Inst. Oceanogr.* **8**:313–346.