Biogeochemical impact of a model western iron source in the Pacific Equatorial Undercurrent

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ABSTRACT

Trace element distributions in the source waters of the Pacific Equatorial Undercurrent (EUC) show the existence of elevated total acid-soluble iron concentrations. This region has been suggested to contribute enough bioavailable iron to regulate interannual and interglacial variability in biological productivity downstream in the high-nitrate low-chlorophyll upwelling zone of the eastern equatorial Pacific. We investigated the advection and first-order biogeochemical impact of an imposed, data-based iron maximum in the western Pacific EUC using an ecosystem model forced by a global dynamical model. We imposed two source profiles of iron constrained by total acid-soluble iron measurements. Though the data for total acid-soluble iron included both dissolved and acid-soluble particulate iron species, we treated all of the total acid-soluble iron as if it was dissolved and bioavailable. A deeper (270 m) source was centered in the density horizon of the observed iron maximum and a shallower (180 m) source was located in the core of our model's EUC, where a dissolved iron maximum has been frequently postulated. These source runs were compared with a control run that contained no specific source of iron associated with the EUC. In the source runs elevated iron concentrations were simulated in the EUC across its entire zonal span, evident as a subsurface plume of dissolved iron slightly below the core of the EUC. In the control run there was no iron maximum associated with the EUC. Upwelling of iron-replete water in the central and eastern equatorial Pacific increased integrated primary productivity in the Wyrtki box (180°W:90°W, 5°S:5°N, 0:200 m) by 41% and 66% for the deeper and shallower iron perturbation, respectively. The source runs increased the realism of the zonal extent of HNLC conditions and the meridional distributions of biological productivity, relative to the control run. However, in the source simulations surface chlorophyll concentrations were too high by a factor of two and maximum surface nitrate concentrations were too low, relative to climatologies. The relative abundance of diatoms roughly doubled upon the input of additional iron, exceeding field observations. Though biogeochemical data are limited and we did not adjust parameters to optimize the model fits to observations, these results suggest that acid-soluble particulate iron supplied to the EUC in the western equatorial Pacific is unlikely to be entirely bioavailable.

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1. Introduction

1.1. Regulation of the equatorial Pacific high-nitrate low-chlorophyll regime

The eastern equatorial Pacific is one of the world’s largest high-nitrate low-chlorophyll (HNLC) ocean regions (Chavez and Toggweiler, 1995). Equatorial upwelling produces between 20% and 30% of global new production (Chavez and Barber, 1987; Chavez and Toggweiler, 1995). The most important findings from the US JGOFS Process Study (EqPac) and subsequent studies in the equatorial Pacific were (1) that this region has all the biological characteristics of HNLC regimes (Murray et al., 1994); (2) that iron is a major limiting nutrient for phytoplankton growth rate, biomass and new production (Martin et al., 1994; Price et al., 1994; Kolber et al., 1994; Coale et al., 1996; Behrenfeld et al., 1996; Landry et al., 1997; Aufdenkampe et al., 2002); (3) that there is a maximum of particulate and arguably of dissolved iron in or slightly below the Equatorial Undercurrent (EUC) along 140°W (Coale et al., 1996; Gordon et al., 1997); (4) that upwelling from the EUC is the primary source of iron to the euphotic zone in this region (Coale et al., 1996; Fung et al., 2000); and (5) that variations in productivity are driven by variability in upwelling and input of iron from the EUC (Chavez et al., 1999; Friedrichs and Hofmann, 2001; Le Borgne et al., 2002; Ryan et al. 2006).

The ecumenical iron hypothesis states that the two primary regulators of HNLC conditions in the equatorial Pacific are iron supply and grazing (Chisholm and Morel, 1991; Landry et al., 1997). Nanoplankton biomass is controlled primarily by rapidly growing microzooplankton (Frost and Frazen, 1992; Verity et al., 1996; Landry et al., 1997), whereas iron limitation affects a broader range of plankton size classes, particularly diatoms, but including picoplankton (Landry et al., 1997; Mann and Chisholm, 2000) and biomass response in iron fertilization experiments (Fitzwater et al., 1996).

1.2. Previous biogeochemical modeling studies

Given the apparent paradox of a highly productive, yet nutrient-limited, HNLC system, numerous modeling studies have investigated the factors controlling equatorial Pacific productivity. Toggweiler and Carson (1995) proposed that export production is kept low by efficient nutrient recycling within three-dimensional meridional overturning cells. Using an ocean circulation model coupled to a biogeochemical model with nitrate and ammonia as limiting nutrients, they showed that although new nitrate was advected eastward in the EUC, primary production in the eastern equatorial Pacific was largely fueled by regenerated nitrogen. In a similar approach, Jiang et al. (2003) investigated nutrient fluxes and included silicate as a limiting nutrient. In their study, iron was not explicitly modeled; instead, its influence was parameterized by modification of the rate constant for plankton growth. Loukos et al. (1997) examined the impact of including iron as a colimiting nutrient with nitrate and ammonia in one-dimensional biogeochemical models and showed the importance of iron in controlling both biomass and primary production. Leonard et al. (1999) found similar results in an expanded one-dimensional iron-based model that included two each of phytoplankton, zooplankton, and detrital size fractions. Christian et al. (2002a) confronted the paucity of iron field data by testing the response of various subsurface and surface (i.e. aeolian iron supply with 10% solubility) western iron sources and showed that western iron sources can propagate east in the EUC (their Fig. 7). The western-iron-source simulations of Christian et al. (2002a) showed that (1) the depth of a western iron source had significant impact on the magnitude and location of the ecosystem response, (2) iron distributions were very sensitive to the parameterization of scavenging, and (3) iron cycle parameters such as aerosol solubility and Fe:N biological uptake ratios had to be tuned to near the limits of published ranges to represent the measured extent of HNLC conditions (upper limit for solubility and lower limit for Fe:N ratios). The influence of the depth difference between the ferricline and the depth of the wind-driven upwelling in the eastern Pacific was found to be critical by Gorgues et al. (2007) in a model setup similar to the one presented here. Vichi et al. (2008) also found that the ferricline depth modulated the ecosystem response to tropical instability waves when a western Pacific shelf iron source was included.

1.3. Candidate western iron sources

It is not clear that a maximum of dissolved iron concentration exists in the EUC. The dissolved iron maximum highlighted by Gordon et al. (1997) consisted primarily of a single depth measurement located ~50 m deeper in the water column than their particulate iron maximum. Both maxima were beneath the core of the EUC as indicated by a bulge in the isotherms. Subsequent iron profiles at 140°W have been inconclusive as to the existence of a subsurface maximum in dissolved iron located in the EUC (Measures et al., 2006). Nakayama et al. (1995) presented data for dissolved iron at 150 E; however, two profiles collected 3 d apart were strikingly different. Measurements of total acid-soluble iron and manganese, which includes both dissolved metal species and particulate species soluble under pH<2 conditions, in the far western equatorial Pacific (143–156 E) by Mackey et al. (2002) and Slemons et al. (submitted) support a western origin for iron in the thermocline. They found a subsurface maximum of total acid-soluble iron slightly below the core of the EUC and elevated values in the New Guinea Coastal Undercurrent (NCCUC), which, along with contributions from the New Ireland Coastal Undercurrent (NICU), has been demonstrated to be the primary pathway for water feeding the EUC from the South Pacific (Tsuchiya et al., 1989; Fine et al., 1994; Butt and Lindstrom, 1994; Ueki et al., 2003). Despite uncertainty as to the existence, magnitude, and variability of a bioavailable iron maximum in the EUC, the hypothesis of a western thermocline iron source has been widely utilized in studies of
equatorial productivity on monthly to geological time-scales (Wells et al., 1999; Ryan et al., 2006; Vichi et al., 2008), but the biogeochemical impact of such an iron source has not been assessed using iron observations (Christian et al., 2002a).

Hydrothermal, riverine and sedimentary sources have all been proposed as sources of iron to the EUC in the western equatorial Pacific (Coale et al., 1996; Gordon et al., 1997; Wells et al., 1999; Milliman et al., 1999; Mackey et al., 2002). This region is volcanically active, and although it is not thoroughly surveyed, known hydrothermal sources in the Bismarck Sea occur either deeper than 1000 m or in shallow water, well separated from the core of the NGCUC (Gamo et al., 1993; Obata et al., 1993; Pichler et al., 1999). The coincidence of maxima in particulate Fe, Mn, and Al at 140 W on the equator are evidence of a lithogenic source of trace elements to the EUC around the New Guinea platform (Gordon et al., 1997). Neodymium isotopic signatures at 140 W indicate lithogenic input from young continental sources, such as the northern New Guinea shelf and slope (Lacan and Jeandel, 2001). Continental-shelf iron sources are increasingly recognized as important sources of iron to the open ocean (e.g. Elrod et al., 2004; Chase et al., 2007; Planquette et al., 2007; Nishioka et al., 2007; Lam and Bishop, 2008). Sediment flux to the New Guinea slope is especially high because of the large fluxes of particulate matter from the island's mountainous rivers (Milliman et al., 1999; Sholkovitz et al., 1999). A significant fraction descends the slope to the depth of the NGCUC as a hypopycnal flow (Kineke et al., 2000; Burns et al., 2008). The location of such a potentially large iron source in the formation region of the EUC suggests a significant biogeochemical role.

In this study, we utilize the existing data of the total acid-soluble iron maximum in the western equatorial Pacific to test the conditions necessary to produce the suggested EUC iron maximum at 140 W (Gordon et al., 1997) and its potential biogeochemical role.

1.4. Inclusion of a data-based candidate iron source to the EUC

In this study we evaluated the impact of a western iron source using the biogeochemical model PISCES (Aumont and Bopp, 2006). We prescribed a source based on the Mackey et al. (2002) and Slemons et al. (submitted) (later referred to as M&S) total acid-soluble iron data. Our aim was to project the resulting iron distribution, and investigate the biogeochemical impact to the eastern HNLC region. We employed three model scenarios: (1) a control with default iron sources as defined by Aumont and Bopp (2006), (2) a western source with a maximum in dissolved (bioavailable) iron matching the data of M&S with a subsurface dissolved iron maximum below the density core of EUC, and (3) a western-source iron maximum of magnitude equal to the values reported by M&S but moved upward to be located in the core density of the EUC. Our study differed from that of Christian et al. (2002a,b) in three key ways: (1) The circulation model is a global ocean, not limited to the Pacific basin. Model studies (e.g. Rodgers et al., 1999; Gorgues et al., 2007) have shown the importance of an adequate representation of the Indonesian Throughflow in achieving the correct ratio of northern and southern source waters to the EUC and depth of the EUC in the eastern Pacific. (2) Our iron sources are based on measured iron profiles from four cruises (M&S). Our western iron boundary condition with a maximum at 270 m resulted in an annually averaged flux of 2.8 mmol Fe m⁻² d⁻¹ from the model grid cells where the source iron profile was imposed, which was considerably greater than the Christian et al. (2002a) continuous injection of 8 μmol Fe m⁻² d⁻¹ at 125 m. Iron concentrations decreased rapidly downstream from our imposed iron flux. The iron source of Vichi et al. (2008) was also based on profiles from Mackey et al. (2002), but they investigated temporal variability in biological productivity whereas we examine the impact of a western iron supply in controlling annual mean nutrient fields and productivity patterns. (3) The PISCES parameterization of iron includes ligand complexation of dissolved iron, a particulate iron pool, and iron scavenging.

1.5. Bioavailability of particulate iron

Although advances in clean measurement techniques have catalyzed the categorization of iron measurements into soluble, colloidal (together making up “dissolved”) iron, small particulate and large particulate iron (e.g. Wu et al., 2001) size classes, definition and measurement of bioavailable iron has remained elusive (e.g. Wells and Trick 2004; Sunda, 2001; Castriotta et al., 2008). Free, inorganic iron is known to be readily accessible to phytoplankton, but labile particulate iron may become bioavailable through processes such as photoreduction, phagotrophy (Barbeau and Moffet, 2000), reduction at cell surfaces (Sunda and Huntsman, 1995; Kuma et al., 2000), and perhaps interaction with various iron-complexing ligands (Kustka et al., 2005; Chen and Wang, 2008). Total acid-soluble iron measurements in the western equatorial Pacific (M&S) as well as repeated observations in the central equatorial Pacific of both dissolved and particulate iron are available, but it is difficult to assess in field studies the biological significance of the observed iron, particularly nonlocal effects. By employing a global biogeochemical model with a robust iron cycle, we can test the distribution of realistic total acid-soluble iron sources across the Pacific basin and constrain the bioavailability of the iron source by evaluating the resulting climatological shifts in nutrient and biological fields. We do not aim in these experiments to optimize the fit of model fields to observations, but rather to apply the model to the following questions: (1) Would the addition of a data-based total acid-soluble iron source located in the western EUC (as dissolved iron) result in a subsurface dissolved iron concentration maximum in the central equatorial Pacific? (2) Would such a concentration maximum imposed in the western equatorial Pacific remain in the EUC as it was transported eastward? (3) What constraints on the bioavailability of this iron source...
are implied by the resulting shifts in nutrient and biological fields in the eastern equatorial Pacific?

2. Methods

2.1. Model configuration

The biogeochemical PISCES model (Aumont and Bopp, 2006) was run in its offline configuration forced by the dynamical fields from the ocean general circulation model OPA (Madec et al., 1998) in a manner similar to that of Gorgues et al. (2005). Both models were described in previous publications, and the most relevant details of this configuration are highlighted here. The OPA domain was global with 2° zonal resolution and variable meridional resolution of 1/2° near the equator and lower poleward. This global configuration maintains an open Indonesian Throughflow. There were 31 vertical levels, with 12 in the upper 120 m. Temperature and salinity fields were initialized with World Ocean Atlas 1998 (WOA98) climatologies (Antonov et al., 1998; Boyer et al., 1998). The model was forced by National Centers for Environmental Prediction (NCEP) climatological winds and daily heat and water fluxes (Kalnay et al., 1996).

PISCES includes colimitation of biological production by nitrate, ammonia, phosphate, silicate and iron, each of which are explicitly cycled. Phytoplankton, zooplankton, particulate organic carbon (POC) and particulate iron each exist in two size classes. Diatoms have higher nutrient half-saturation constants than the smaller phytoplankton, referred to hereafter as nanoplankton, although they represent all non-silicifying phytoplankton. Biogenic silica sinks at the same rate as large particulate organic carbon, both of which sink faster than small particulate organic carbon. Uptake and remineralization of C, N, and P occur with a constant Redfield ratio. PISCES was initialized using nutrient climatologies from Conkright et al. (1998) and an iron field from the model climatology employed by Aumont and Bopp (2006). PISCES was spun up for 50 years to stabilize the biogeochemical fields.

2.2. Iron cycle in PISCES

Several components of the iron cycle were explicitly modeled in PISCES (Fig. 1). Dissolved iron consists of free and ligand-complexed species, both of which were considered bioavailable. The Fe:C ratio in zooplankton was fixed, but varied in nanoplankton and diatoms according to dissolved iron availability, light and nutrient stress (Sunda et al., 1991; Marchetti et al., 2006). Particulate iron was remineralized in proportion to the particulate iron pool at the same temperature-dependent rate as POC, slowing at lower temperatures (and greater depths).

In our model, free dissolved iron was scavenged in proportion to the total concentration of large and small POC and biogenic silica. Moore et al. (2004) employed a similar parameterization, but included sinking dust flux as a scavenging surface, in part to balance greater prescribed iron solubility. Observations suggest abrupt increases in iron scavenging rates when [Fe] > ~0.6 nM, because ligand complexation keeps iron in solution up to that concentration (Johnson et al., 1997; Rue and Bruland, 1997). Concentrations in seawater are usually greater than predicted by the thermodynamic solubility of Fe(OH)_3 (~0.01 nM) (Liu and Millero, 2002). To mimic this phenomenon, our model included a ligand pool of uniform concentration throughout the oceans with a variable iron complexation equilibrium constant. The equilibrium constant was calculated based on the formulation of Liu and Millero (2002) such that the iron scavenging rate increased dramatically when dissolved iron concentrations exceeded the 0.6 nM threshold. Models without ligand complexation, employing a linear scavenging rate, rapidly deplete iron (Archer and Johnson, 2000; Christian et al., 2002a; Parekh et al., 2004). However, the single-ligand parameterization used in our model was likely an oversimplification. Deepwater iron concentrations are not globally uniform, ranging between ~0.5 and 1.5 nM although rarely exceeding 0.6 nM, suggesting a range of ligand-binding capacities (Parekh et al., 2004; Moore and Braucher, 2008).

Dissolved iron sources in PISCES include atmospheric deposition, river runoff, and coastal sediments. Atmospheric iron fluxes are based on Tegen and Fung (1995) and assume a constant 1% solubility of deposited iron. Coastal sources of iron are parameterized as an imposed 0.5 nM iron concentration in each model grid cell along the coast throughout the upper 18 vertical levels (0–275 m) with adjustment for bathymetric variability. To account for sub-model-gridscale bathymetric variations, which are especially significant near continental slopes, the OPA model grid was mapped to the

Fig. 1. Schematic of PISCES iron cycle. Prognostic model variables are shown in boxes with a grey background.
Fig. 2. Vertical profiles of iron in the western equatorial Pacific. Blue circles represent total dissolvable iron concentrations of Mackey et al. (2002) at equatorial stations between 147 and 155°E. Green squares show total acid-soluble iron concentrations of Slemons et al. (submitted) at equatorial stations between 145 and 156°E. The solid lines show dissolved iron concentrations along the equator between 154 and 158°E in the three model simulations: control (pink), 270-m source (red), and 180-m source (yellow). The purple line indicates the depth of the density of the EUC core (25.0 < σ < 26.0) during the Mackey et al. (2002) surveys. The shorter green line shows the depth of the same density surface in the model at this location. For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.

Fig. 3. Zonal sections of iron (nM) (top) and nitrate (μM) (bottom) concentrations (color shading) along the equator. The control run fields are shown on the left, and anomaly plots for the two source runs minus the control are shown in the middle and right and labeled in the panel. Red (blue) shades indicate an increase (a decrease) in the source run relative to the control. Note the nonlinear scale for the iron anomaly plots. Isopycnal surfaces contoured in black in the top panel. Zonal velocity (m/s, solid is eastward flow, dashed is westward flow) is contoured in the bottom panel for the model (black) and TAO array (1988–2006) observations (magenta).
high-resolution ETOPO5 bathymetric dataset, and the fractional ocean bottom area shallower than 275 m was calculated. The imposed iron concentration was then multiplied by this fractional area. Iron in PISCES is conserved.

2.3. Imposed western iron source scenarios

Each of our three model scenarios included the standard PISCES atmospheric, riverine, and sedimentary sources described above. The control run included only these sources. Our two source experiments included an additional western iron source based on observations (Fig. 2). We fit a curve to match the amplitude of the subsurface maximum in total acid-soluble iron (all iron solubilized after acidification: 1 ml HCl/l seawater, pH<2; 3 month storage) measured by M&S from 143 to 156°E along the equator. The imposed maxima are upper bound estimates of bioavailable iron, since the measurements are of total acid-soluble iron. The source simulations each included a dissolved iron concentration profile uniformly imposed within 0.5° of the equator from 154 to 158°E with maxima at 270 m (270 src) and 180 m (180 src). For the 270 src simulation, we located the maximum iron concentration at the depth of the iron maximum observed by M&S. For the 180 src simulation, we imposed the maximum iron concentration in the center of the model’s EUC core (defined as $\sigma = 25.5$). At the location of the imposed maximum at 180 m, the velocity core of the model’s EUC spanned 0.5°S–1°N. Both source simulations were run for 30 years.

In this study, we evaluated the downstream biogeochemical impact of a realistic EUC-associated iron source in the western equatorial Pacific. We did not attempt to locate the imposed iron source in the model NGCUC or NICU even though they are the most likely entry points for subsurface iron (e.g. Mackey et al., 2002), because our 2°-resolution configuration limits the accuracy of the model’s representation of coastal currents. By 155°E the EUC is thought to have acquired most of its western source waters (Tsuchiya et al., 1989), although the pathways of the sources from the North Pacific are poorly known. Thus, we chose our imposed source profile location where the EUC was well established in the model and in observations (Fig. 3). Starting from a fully spun-up PISCES biogeochemical state, both source profiles were implemented for 30 years.

3. Results

3.1. Iron and nitrate distributions

The additions of an elevated dissolved iron source in the western equatorial thermocline altered the pattern and quantity of model nutrients across the tropical Pacific. Zonal distributions of simulated iron and nitrate along the equator are shown in Fig. 3. Nutrient concentrations are shown for the control simulation, and nutrient concentration anomalies, relative to the control, are shown for the source simulations. In the control simulation, both iron and nitrate concentrations increased with depth with typical nutrient-like profiles and shoaled eastward along isopycnal surfaces (black contours, upper panels), especially in the nitrate field. In the source simulations, enhanced iron concentrations were greatest within several hundreds of kilometers of the imposed source and decreased exponentially eastward, primarily because of loss by scavenging (Figs. 3 and 6). The additional iron reached surface waters west of 120°W for the source originating at 270 m and west of 156°W for the 180-m
source. Model dynamics robustly represented the depth and gradient of the EUC, although the model zonal velocities were slightly lower than TAO observations (Fig. 3, lower panels). This weak anomaly resulted from the weak bias in the mean state of the trade winds found in NCEP compilations (Putnam et al., 2000).

Although the simulated iron perturbations were advected eastward, maximum source-run iron concentrations occurred below the core of the EUC. Iron concentrations at 140°W are shown alongside the dissolved iron measurements of Gordon et al. (1997) from July 1990, a non-ENSO year, in Fig. 4. During their sampling at this location, the depth of the core of the EUC was at approximately 125 m, as indicated by the doming of isotherms in their data (Gordon et al., 1997, their Fig. 1). The dissolved iron concentration “bulls-eye” maximum suggested at 200 m by Gordon et al. (1997) was well beneath both the distinct maximum in particulate iron at ~150 m (Gordon et al., 1997, their Fig. 2) and the core of the EUC. The core of the model’s EUC was at a similar depth during climatological July, indicated by the zonal velocity and the doming of the isopycnals. In contrast to the observations of Gordon et al. (1997), dissolved iron concentrations followed isopycnal surfaces in the control simulation and showed no “bulls-eye” maximum. Both source simulations produced maxima in meridional iron centered slightly north of the equator. The dissolved iron “bulls-eye” maximum highlighted by Gordon et al. (1997) was evident at this location only in the 270-m source simulation. The 180-m source simulation generated a broader iron enrichment, 2–3° across at its widest. Maximum iron concentrations in both source simulations at this location were more than double the observed dissolved iron maximum of 0.35 nM, but not significantly greater than the sum of dissolved and particulate iron measurements (0.65–0.75 nM) at 200 m (Gordon et al., 1997). In both source runs, iron anomalies relative to the control were greatest below the core of the EUC.

Both source simulations resulted in negative anomalies of surface nitrate in the central and eastern equatorial Pacific. In addition, the source simulations produced negative nitrate anomalies relative to the control below 150 m east of 100°W (Fig. 3). At this location, nitrate concentrations simulated by the control run exceeded observations from the World Ocean Atlas (Garcia et al., 2006) because of overly vigorous model denitrification and incorrect model dynamics. As a result, this area exhibited strong horizontal nitrate gradients and thus was highly sensitive to local negative nitrate anomalies. The source of these model errors (whether due to low resolution or inadequate representation of tropical overturning cells) is not clear (Cravatte et al., 2007). This nitrate depletion occurred too deep to have any immediate biological impact in the eastern equatorial Pacific and was not investigated further in our study. The source simulations resulted in positive anomalies of nitrate, relative to the control, below the zone of increased productivity in the equatorial upwelling zone.

Simulated and observed (2005 World Ocean Atlas, Garcia et al., 2006) surface nitrate concentrations are shown in Figs. 5 and 6. The magnitude of the observed maximum concentrations of surface nitrate in the eastern Pacific, located at 100°W, were most closely matched in the control simulation. The control simulation also better reproduced the meridional location of the observed surface nitrate maximum, located a few degrees south of the equator. However, the source simulations better represented surface nitrate concentrations along the equator in the central and eastern equatorial region. West of ~105°W, surface nitrate concentrations at the equator reached 10 μM in the control run and reached only 7 μM in the source runs and the WOA data. The greatest drawdown of surface macronutrients, in both magnitude and spatial extent, occurred in the 180-m source simulation.

### 3.2. Impacts on biology and carbon cycle

Both source runs resulted in increased primary production relative to the control (41% and 66% for the 270-m and 180-m source simulation, respectively) integrated across the entire Wyrtki box (180°W:90°W,
The greatest increases in vertically integrated primary production from the source simulations were generated just southeast of the maximum in surface nitrate (Fig. 5). Visual inspection indicates that primary production in the source simulations decreased where surface nitrate was drawn down (<1 µM).

The increase in primary productivity in the source simulations dramatically impacted the surface chlorophyll fields. Simulated and observed (SeaWiFS compilation, Yoder and Kennelly, 2005) surface chlorophyll concentrations are shown in Fig. 7. Our 2° zonal resolution model simulated overly broad surface chlorophyll concentrations in all simulations, but this bias was exacerbated in the source simulations.

Climatological estimates of equatorial primary production have classically been calculated over the Wyrtki box (180°W:90°W, 5°S:5°N). The vertically integrated model primary productivity of 904 mg C m⁻² d⁻¹ in the 270-m source run closely matched the non-El Niño compilation (field studies from 1990 to 1992) of 900 mg C m⁻² d⁻¹ reported by Chavez et al. (1996). Primary productivity was less than Chavez’s estimate in the control run (643 mg C m⁻² d⁻¹) and greater in the 180-m source run.
run (1066 mg C m$^{-2}$ d$^{-1}$), although we should note that this average is very sensitive to the meridional width selected. Integrated primary productivity (0–120 m) at 140°W over meridional bands of varying width from our model simulations are compared with observations from a non-ENSO season (fall 1992) in Fig. 8 (Barber et al., 1996). Observed primary productivity was greatest within 2° of the equator (131 mmol C m$^{-2}$ d$^{-1}$) and generally decreased poleward (to 59 mmol C m$^{-2}$ d$^{-1}$ averaged over 10° of the equator, a difference of 72 mmol C m$^{-2}$ d$^{-1}$). The exception to this pattern (101 mmol C m$^{-2}$ d$^{-1}$ within 1° of the equator) was likely due to a passing TIW at 2°N (see description of Fig. 9). The source simulations produced similar meridional gradients (differences of 26 and 35 mmol C m$^{-2}$ d$^{-1}$ between the 1° and 10° meridional averages for 270 and 180 src, respectively). In the control simulation, primary production was notably more homogeneous meridionally (14 mmol C m$^{-2}$ d$^{-1}$ difference between the 1° and 10° band averages). Observed primary productivity averaged across 10°S:10°N band was represented well in all three model simulation: all converged within 13 mmol C m$^{-2}$ d$^{-1}$ of the observed primary productivity along 140°W (59 mmol C m$^{-2}$ d$^{-1}$).

Among the model’s primary producers, diatoms experienced a greater response to additional iron, although nanoplanckton remained dominant in all simulations in terms of both primary productivity and biomass. Diatom biomass increased more than 3-fold for the 270-m iron source and approximately 4-fold for the 180-m source in the upper 25 m of the far eastern equatorial upwelling zone (110°W:90°W, 2°N:2°S) relative to the control simulation. The relative contribution of diatoms to...
phytoplankton biomass was within or in excess of the upper end of the range of measured values (Table 1) (Chavez et al., 1996).

All simulations generated the greatest open-ocean export flux in the eastern equatorial Pacific from 6°N to 8°S with a local minimum at the equator. Although limited measurements exist of export production for this region, a similar pattern of a near-equatorial minimum in POC export and a northern maximum in export around 6°N was observed during non-ENSO conditions (Survey II) across 140°W (Fig. 9) (Murray et al., 1996). Our simulations reproduced the zonal gradient, meridional pattern and magnitude of POC flux measurements, and the relatively low ratio of POC export to total primary productivity (the e-ratio) observed at the equator (Murray et al., 1989, 1996; Dunne et al., 2000). Observed maxima in POC export were located at 2°N and 5°S during Survey II (Murray et al., 1996), which may be associated with coincident convergence zones (Johnson, 1996, their Fig. 1; Walsh et al., 1997). The convergent front at 2°N was shown to be the result of a passing tropical instability wave (Johnson, 1996). The meridional maxima and minima in POC export were likely to have been shifted by the passing TIW and would not be expected to exactly match our model outputs, which were climatological.

4. Discussion

4.1. Constraints and fate of a western iron source

In our source runs we used measured profiles of total acid-soluble iron from ~155°E (M&S) and treated these total iron concentrations as if they were entirely bioavailable. Measurements of total acid-soluble iron included both truly soluble inorganic iron available for direct biological uptake and colloidal and labile particulate iron, some fraction of which may be available for biological uptake. We used the total acid-soluble iron profiles to define the source because the dataset is robust and there was clearly a maximum in total acid-soluble iron associated with the EUC that existed in the western and central Pacific. There is much less data for dissolved iron and a distinct maximum of dissolved iron associated with the EUC is not evident. Since all dissolved iron in the model is treated as bioavailable, whether or not it is ligand-bound, it is available for uptake by primary producers immediately upon entry into the euphotic zone. Thus, the source simulations employ an upper bound of bioavailable iron.

The prescribed western equatorial Pacific iron perturbations did not remain on their original isopycnal surfaces as they advected across the basin, but spread throughout the thermocline, particularly deeper than their original density horizon. This spreading and sinking was likely due to dilution from tropical recirculation cells (Lu et al., 1998) and sinking from scavenging (Figs. 3 and 6). Nonetheless, subsurface maxima in dissolved iron were propagated across the Pacific basin. In the central equatorial Pacific, the source simulations matched the observed distribution of dissolved and particulate iron fields (Gordon et al., 1997) better than the control simulation. The magnitudes of the maxima from the two source simulations at 140°W were double the observed dissolved iron concentration, but nearly matched (25% greater) the sum of dissolved and particulate iron values, which is consistent with our treatment of total acid-soluble iron data as dissolved iron in the model. The source simulations generated iron fields in agreement with total iron concentrations for this region.

More recent measurements of dissolved iron at 140°W reported a greater dissolved iron concentration of 0.7 nM in the core of the EUC (Measures et al., 2006). Variability in iron fields is expected, both from variations in transport of the EUC on interannual (ENSO) and decadal timescales (Chavez et al., 1999; Le Quéré et al., 2003; Le Borgne et al., 2002; Rodgers et al., 2008) and from variations in candidate sources, like the strength and depth of the NGCUC (Ryan et al. 2006). Although residence times and cycling of dissolved and particulate iron species differ, our combination of the two appears to be a sensible assumption given the high advection rates associated with the EUC. The implications of iron speciation for interior ocean iron cycling can best be addressed in future modeling studies that explicitly include cycling of both dissolved and particulate iron. The inclusion of an explicit western thermocline iron source improved the representation of the total iron field in the central equatorial Pacific.

Our model’s iron cycle included sediment, atmospheric, riverine, and desorption sources and scavenging loss dependent on dissolved iron and particle concentrations. It has been shown to be robust over a broad range of iron concentrations (e.g. Bopp et al., 2003). As demonstrated in other models (Christian et al., 2002a; Moore and Braucher, 2008), changes to model iron sources often call for optimization of iron cycle parameters. Significant uncertainties in published values of these parameters exist and preclude prognostic modeling of the iron cycle. Even given a robust distribution of total acid-soluble iron, the availability of various iron species for biological uptake remains a complicated question. In our simulations, broad shifts in nutrient and biological fields are used to evaluate the bioavailability of a realistic, data-based iron supply; no iron cycle parameters were adjusted.

4.2. High-nitrate low-chlorophyll shift

The prescribed western iron sources used in this study led to an intense localized increase in primary and new production and surface chlorophyll and a drawdown of

<table>
<thead>
<tr>
<th>Table 1</th>
<th>Relative phytoplankton biomass contribution of diatoms for the model integrated over the euphotic zone in the eastern equatorial Pacific (135°W-90°W, 5°N-5°S) and for observations compiled between 1988 and 1992 (Chavez et al., 1996, their Table 4).</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model</td>
<td>Observations</td>
</tr>
<tr>
<td>Control</td>
<td>25% 140°W 5–10%</td>
</tr>
<tr>
<td>270-m src</td>
<td>33% 125°W 15–31%</td>
</tr>
<tr>
<td>180-m src</td>
<td>36% 110°W 6–14%</td>
</tr>
</tbody>
</table>
surface nitrate. As expected from earlier studies (e.g. Christian et al., 2002a), the depth of iron injection determines the zonal extent of the biological response. A northward shift in meridional extent of the elevated nitrate region occurred in the source simulations (Fig. 5). The flow pathway of the lower EUC thermostad waters, where most of the iron enrichment in the source simulations occurred, brings nutrient-rich waters to the surface south of the equator off the South American coast (Lukas, 1986; Toggweiler et al., 1991), so this shift was expected from the localized drawdown of macronutrients south of the equator. The control simulation better represented observations of both the location and the magnitude of nutrient and biological fields in the eastern equatorial Pacific. This implies that a significant amount of the total acid-soluble iron from the western Pacific is not fully bioavailable in the eastern equatorial Pacific. We did not attempt to quantify what fraction of acid-soluble iron is bioavailable because the bioavailability of iron is not a linear function of iron supply.

Other interconnected factors besides iron bioavailability could produce excessive productivity in our simulations of equatorial Pacific productivity, including biases in equatorial dynamics or incorrect iron cycle parameterizations. We cannot rule out the possibility that our PISCES simulations may be biased in the equatorial Pacific and the additional iron simply revealed those biases. As observed in similar explicit iron cycle model studies, uncertainty and natural variability in published iron cycle parameters cause the representation of iron in ecosystem models to be an underdetermined problem (e.g. Christian et al., 2002a; Moore and Braucher, 2008). A close dataset comparison is complicated by the fact that equatorial climatologies for many biological variables do not exist. Given the global robustness of PISCES productivity and our experimental design in which only the western iron supply was changed, the most parsimonious explanation is that primary producers are not able to access a portion of the total acid-soluble iron transported eastward in the EUC.

Observations show that the dynamical (salinity) and biological fronts (new production, nutrients) are usually coincident or, in the case of suspended upwelling, elevated surface nitrate is observed east of the salinity front (Stoens et al., 1999; Radenac et al., 2001; Le Borgne et al., 2002). The zonal extent of low surface nitrate matched the dynamical warm pool best in the 180-m source simulation and worst in the control simulation (not shown). The inclusion of western iron sources better modulated the transition from oligotrophic warm-pool to HNLC cold-tongue conditions. Over-estimation of the extent and magnitude of the equatorial HNLC region is common to many coarse-resolution models and is considered to be driven by an overly shallow EUC in the eastern Pacific (Gnanadesikan et al., 2002). This bias was also present in our simulations (Figs. 3, 5 and 6), and it affects both dynamical and nutrient fields similarly.

4.3. Impacts on biology and carbon cycle

Our results were in agreement with the ecumenical iron hypothesis and demonstrate several key biological shifts in addition to increases in primary production and biomass upon relief of iron limitation. Although physical dynamics were the same for all simulations, the source simulations produced a more realistic meridional distribution of primary production at 140 W, especially in 180src (Fig. 8). In a parallel result, the meridional pattern in export production also improved in the source simulations. Recirculation cells known as Tropical Cells (Lu et al., 1998) advect settling POC poleward in surface waters, so maximum POC export coincides with maximum surface convergence off the equator where water masses are downwelling (Walsh et al., 1997). An equatorial minimum in carbon export existed in all of the simulations due to the dynamics of particle advection, but was more apparent in the source simulations. These results indicate the inclusion of a subsurface iron maximum located in the EUC improves the distribution of biological fields in the equatorial upwelling zone. Integrated nutrient fields were better represented in the control simulation, but the source simulations improved the nutrient and productivity gradients in the eastern equatorial Pacific. Nonetheless, productivity was too high and nitrate too low in the source simulations despite a better simulation of meridional and zonal gradients, respectively. Therefore, it is evident that the upper limit of bioavailable iron used here is too high.

Our model appeared to overestimate diatom productivity, although a notable shift in community structure occurred in the source simulations, favoring larger diatoms as expected from relief of iron limitation (e.g. Landry et al., 1997). Because datasets of large-scale community structure are limited, this result is not suitable to evaluate the relative fitness of our simulations, but it does suggest a shift in particle size distribution should be considered when predicting changes in export flux with iron fertilization. An increase in the relative abundance of diatoms and resultant biogenic silica ballast has been proposed as a factor to explain increases in export flux and the e-ratio (i.e. Armstrong et al., 2002; Dunne et al., 2007). Model experiments regularly produce increased export efficiency in response to an increase in diatom abundance, although with a range of sensitivities; the e-ratio of Moore et al. (2004) was less responsive than that of Aumont and Bopp (2006). In PISCES, the settling rate of POC increases with particle size, and the aggregation rate increases with POC and DOC abundance. Model diatoms produce more large (and thus faster sinking) particulate organic material, including biogenic silica. In the field, ballast production has been shown to be highly variable and dependent on nutrient conditions; eastern equatorial Pacific diatoms incorporate fivefold less silica per volume than Southern Ocean diatoms (Baines et al., 2008). Despite the increased fraction of diatoms (and their greater Si:N uptake due to light limitation, not shown) in the source simulations, the e-ratio decreased when averaged across the tropical Pacific region east of the dateline. Export efficiency decreased locally in the equatorial upwelling zone, where the greatest increase in diatom biomass and production occurred, and off the equator in zones of convergence and downwelling.

Increased particle remineralization was the immediate cause for decreased export efficiency in the source
5. Conclusions

Profiles of total acid-soluble iron in the western equatorial Pacific (~155°E) have a maximum associated with the Equatorial Undercurrent (EUC) (Mackey et al., 2002; Siemons et al., submitted). This maximum is mostly in the form of particulate iron that appears to originate from sediments along the northeastern continental margin of Papua New Guinea (Siemons et al., submitted).

We conducted model simulations in which this EUC-associated western iron source was transported to the east and we evaluated its impact on biogeochemical distributions. In our model we treated all of the total acid-soluble iron as if it was 100% bioavailable. It is important to emphasize that achieving precise model realism is limited both by the availability of biogeochemical data and by the significant uncertainty in modeling biological processes. Nonetheless, the resulting biogeochemical fields demonstrated five primary conclusions: (1) As the iron maximum in the EUC is advected from the western Pacific to the east the concentrations of iron decrease. (2) This western iron source is consistent with the maxima in total iron (dissolved plus particulate) observed at 140°W slightly below the core of the EUC by Gordon et al. (1997). (3) The control run without additional iron sources results in excessive surface nitrate and chlorophyll in the central equatorial Pacific (180°W:110°W). (4) The inclusion of a maximum of iron in the western EUC provided notable improvements to the shape of biogeochemical fields for which localized equatorial dynamics are key: concentrations and zonal gradients of surface nitrate along the equator, the narrow band of enhanced primary productivity in the equatorial upwelling zone, and meridional gradients of primary productivity and carbon export production. (5) Nonetheless, the additional iron resulted in excessive depletion of surface nitrate and excessive surface chlorophyll in the eastern equatorial Pacific, suggesting that it is incorrect to assume that all of the acid-soluble iron is bioavailable. We do not diagnose here what fraction of total acid-soluble iron is bioavailable. Future work should include both experimental and model studies of the bioavailability of different iron species besides dissolved iron.

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