An examination of the “continental shelf pump” in an open ocean general circulation model

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Abstract. In a recent study of the shelf region of the East China Sea, Tsunogai et al. [1999] estimated that a combination of air-sea exchange and biological and physical transport processes could transfer carbon from the shelf region into the open ocean at a rate of 35 g C m\(^{-2}\) yr\(^{-1}\). Contrasting with the solubility and biological pumps of the open ocean, they described this collective activity as the “continental shelf pump” and suggested that if this pump operated throughout the world’s shelf regions, it could be responsible for ocean uptake of \(~1\) Gt C yr\(^{-1}\) (\(\approx 50\%\) current ocean uptake of anthropogenic CO\(_2\)). In this work a general circulation model (GCM) is used to explore the potential strength of this pump across the world’s shelves. Since the GCM does not represent the continental shelf regions explicitly, a parameterization of the pump has been used. Results of simulations find modeled pump activity very variable between shelf regions, with the East China Sea shelf behaving very similarly to the global average. Storage of pump carbon is particularly high in the Atlantic Ocean and other regions where deep water is formed. A considerable reservoir of pump carbon becomes trapped under the Arctic ice sheet. Simple extrapolations from the results suggest that should shelf regions absorb CO\(_2\) at the rate of the East China Sea, the pump would account for a net oceanic uptake of 0.6 Gt C yr\(^{-1}\).

1. Introduction

In many regions of the world ocean the margin between the continental and oceanic tectonic plates is demarcated by the so-called continental shelf zones. These zones are commonly gently sloping regions of the continental plates which because of the contemporary sea level, lie beneath the sea surface down to a depth of \(~200\) m. They reach their greatest size off the coasts of Northern Eurasia, Eastern Asia, Oceania, and the eastern Americas. Figure 1 shows the global distribution of the Earth’s surface which lies between 0 and 200 m below contemporary sea level.

The relative shallowness of these zones (compared to the open ocean) and the input of nutrients to them from the terrestrial environment (both naturally and, recently, anthropogenically), make them considerably more biologically productive than the open ocean. Despite the fact that they make up only a relatively small fraction (\(\approx 8\%\)) of the world ocean’s surface the continental shelf zones contribute an estimated 19–28% of total global ocean production [Longhurst et al., 1995]. Some of this production is exported to the open ocean, and recently, a debate [Duarte and Agusti, 1998; Williams and Bowers, 1999; Duarte et al., 1999] has focused on whether or not the open oceans are globally net autotrophic (i.e., fix more inorganic carbon than they regenerate) or net heterotrophic (i.e., regenerate more organic carbon than they fix).

In this context and on the basis of observations collected from the East China Sea (see Figure 1), Tsunogai et al. [1999] propose a new term, the “continental shelf pump” (henceforth CSP), to collectively describe a series of processes that transport carbon from the atmosphere via the continental shelf to the deep ocean. They identify this pump as being a separate mechanism to the open ocean biological and solubility pumps traditionally recognized as routes by which carbon reaches the deep ocean. For an overview of these latter pumps, refer to Sarmiento [1993] and Raven and Falkowski [1999].

Using observations made in the surface waters of the East China Sea, Tsunogai et al. [1999] calculated a mean CO\(_2\) fugacity deficit of 55 \(\pm\) 5 ppm. This calculated deficit was found to agree well with an earlier estimate of the export flux of carbon derived in a budget for the East China Sea [Tsunogai et al., 1997]. Studying the distributions of total dissolved carbonate and alkalinity, Tsunogai et al. [1999] concluded that this large deficit could be explained as follows. Since the shallow seafloor restricts the convection of cooling water, cooling is greater for waters on the continental shelf than for waters in neighboring open ocean areas. This leads to the production of relatively cold and dense water, which in combination with biological production, results in a greater absorption of CO\(_2\) in the continental shelf zone. The absorbed CO\(_2\), both in the form of dissolved inorganic carbon (DIC) and dissolved/particulate organic carbon (DOC/POC), is transported in this denser water into the subsurface layer of the open ocean via isopycnal mixing processes. Owing to a well-developed thermocline the hydrography of the East China Sea permits this transport to occur even during the warm season. Scaling up from the East China Sea, Tsunogai et al. [1999] suggest that if the processes they have observed there operated over the entire global continental shelf zone, this pump would account for a net oceanic uptake of CO\(_2\) of \(~1\) Gt C yr\(^{-1}\) (83.3 Tmol C yr\(^{-1}\)). Given that observational [Takahashi et al., 1999] and model [Orr, 1997] estimates of the world ocean’s current (1990–2000) anthropogenic CO\(_2\) uptake are \(~2\) Gt C yr\(^{-1}\) and that these exclude or poorly represent the continental shelf zone, Tsunogai et al. [1999] work may have important implications for studies of the natural and contemporary carbon cycle. For instance, should the shelf regions be net exporters of carbon to the open ocean, observational studies that focus only on the open ocean will overestimate ocean outgassing by including outgassing CSP carbon but excluding CSP uptake.
As mentioned in the last paragraph, models are now commonly used to provide estimates of the world ocean’s uptake of CO₂ for studies of both the natural and anthropogenically forced carbon cycle. Increasingly, full ocean general circulation models (GCMs) are being used, with carbon represented simply by a single DIC tracer [Taylor, 1995] or, in more complex models, by a series of chemically or biologically based carbon tracers [Fasham et al., 1993; Six and Maier-Reimer, 1996; Palmer and Totterdell, 2001]. However, GCMs rarely represent the continental shelf zone or the processes that occur on it. This is partially related to the relatively coarse resolution of most GCMs. Grid cells are typically far greater in extent than the continental shelf. Furthermore, the physical processes that occur where the open ocean meets the shelf (e.g., shelf breaks and fronts) are not characterized well by coarse-resolution models. For these and other reasons the continental shelf zones are rarely explicitly included in model simulations. Modelers usually assume that processes occurring on the shelves have very little impact on the much larger open ocean areas of their models. Consequently, most GCMs neither resolve nor represent the continental shelf zone.

However, while GCMs may be incapable of explicitly representing the processes that Tsunogai et al. [1999] attribute to their CSP, they could represent the processes that transport carbon off the shelves and into the deep ocean (or, at least, the processes that transport carbon from the regions adjacent to the shelves into the open ocean). This study attempts to examine these latter processes by parameterizing the shelf processes that transfer carbon to denser shelf bottom water. The GCM used then follows the progress of this carbon to attempt to quantify the strength of the CSP. Additionally, the GCM’s results can suggest where CSP carbon outgasses in the open ocean, an issue difficult to resolve observationally.

2. Model Description

The model used in this study consists of two parts: the carbon cycle model itself and the world ocean GCM which it is embedded into.

2.1. Ocean GCM

This work makes use of the U.K. Meteorological Office’s HadCM3L ocean GCM. The world ocean is represented by a grid of cells each measuring 2.5° latitudinally by 3.75° longitudinally. In the vertical dimension the ocean is divided into 20 fixed level layers. These increase in thickness from 10 m at level 1 (ocean surface) to 615 m at level 20 (ocean floor; 5500 m). Within the confines of this relatively low resolution the GCM has a realistic representation of land distribution and ocean floor depths. The GCM is run with a time step of 24 hours.

The GCM is of Bryan-Cox primitive equation form, with ocean temperature and salinity modeled as active tracers. Monthly fields of heat flux [Esbensen and Kushnir, 1981], precipitation/evaporation [Jaeger, 1976; Esbensen and Kushnir, 1981], and wind stress [Hellerman and Rosenstein, 1983] force the GCM at its surface boundary. The dynamics of the GCM’s surface are also driven by a Kraus-Turner bulk mixed layer [Kraus and Turner, 1967]. This homogenizes the surface layers down to the base of the calculated mixed layer. To prevent excessive drift away from observations, temperature and salinity in the surface layer are relaxed back toward the monthly climatologies of Levitus and Boyer [1994] and Levitus et al. [1994], respectively. A relaxation timescale of 60 days (for a 50 m mixed layer) is used.

The Gent and McWilliams [1990] eddy parameterization scheme is implemented, and this permits the model to exclude explicit horizontal diffusion. Tracers are also subject to Redi’s [1982] along-isopycnal diffusion scheme. Vertical diffusion between model layers is depth dependent, with the background coefficient increasing from 0.1 cm² s⁻¹ at the ocean’s surface to 1.5 cm² s⁻¹ at 5000 m, from the observationally based estimates of Kraus [1990].

The North Atlantic overturning circulation of the GCM for the section at 24°N is 24 Sv. This compares with an observational estimate for the same section of 19 Sv by Hall and Bryden [1982]. Because of the narrowness of its representation in the model grid, flow through the Bering Straits is negligible in the GCM. Flow from the Pacific Ocean to the Indian Ocean via the Indonesian Throughflow is very strong in the GCM at 23 Sv. This is considerably greater than the 7 Sv estimate of Wijffels et al. [1996] from hydrographic sections of the region (see Banks [2000] for a thorough study of this region in a variant of the GCM). For further details regarding the GCM, refer to Gordon et al. [2000] and Palmer and Totterdell [2001].

2.2. Carbon Cycle

Although carbon in the ocean occurs in many inorganic and organic forms, for simplicity, this study uses the single DIC tracer model [Orr et al., 1999] produced for phase 2 of the international Ocean Carbon Cycle Model Intercomparison Project (OCMIP-2). The protocol describing this model and the data fields that force it
DIC is a passive tracer in the GCM, interacting solely with the model "atmosphere" at the ocean's surface. Its conservation equation is

\[
\frac{d[DIC]}{dt} = 0[DIC] + F_v + F + CSP,
\]

where [DIC] is the model's concentration (mol m\(^{-3}\)) of total dissolved inorganic carbon; \(0\) is the three-dimensional (3-D) transport operator representing changes in DIC concentration due to advection, convection, diffusion, and mixing; \(F_v\) is the "virtual" flux term for DIC representing the changes in its surface concentration due to evaporation and precipitation; \(F\) is the flux term for DIC representing air-sea exchange of CO2; and CSP is the flux term for DIC representing the addition of DIC to the GCM via the parameterized CSP (see section 2.3).

The use of a virtual flux is required as the GCM has a "rigid lid." In the real ocean, evaporation and precipitation (EP) remove and add water to the ocean, respectively, changing the volume of water above a given point of the seafloor. In the GCM the fixed vertical levels effectively prevent any change in volume, so a virtual flux is introduced to account for changes to model tracers brought about by the concentrating/diluting effects of EP. In this GCM, evaporation and precipitation are additionally modified by the relaxation term.

Air-sea exchange of CO2 is a function of the atmospheric and surface ocean concentrations of CO2, air pressure, piston velocity, and ice cover. In the model these latter three variables are specified as OCMIP-2 monthly data fields. Atmospheric CO2 is held at a globally constant value of 278 ppm (preindustrial level; see below). Surface ocean [CO2] is calculated from surface [DIC], temperature, salinity, and [Alk] using standard OCMIP-2 carbon chemistry routines, where [Alk] is alkalinity (eq m\(^{-3}\)) and is determined as a normalized linear function of model salinity.

Although this model excludes many important processes, including the biological effects on both [DIC] and [Alk] and the riverine input of carbon, it captures several aspects of the natural carbon cycle and has proved a very useful baseline model for OCMIP-2 experiments.

A preindustrial (278 ppm) level of atmospheric CO2 was chosen over a contemporary one (\(\approx 365\) ppm) for several reasons. In order to facilitate simulation analysis it is preferable to minimize model drift by initiating simulations from an "equilibrium" state, where carbon fluxes into and out of the ocean are in balance. The contemporary atmosphere-ocean system is considerably out of equilibrium because of the industrial CO2 perturbation, with a net flux into the ocean at present. In contrast, prior to the industrial period, atmospheric CO2 remained broadly constant (260–280 ppm) during the current interglacial period (\(\approx 10000\) years), suggesting a natural carbon cycle close to equilibrium or, at least, converging slowly onto one. Furthermore, the pump proposed by Tsunogai et al. [1999] should operate independently of an anthropogenic perturbation. It should play a role in both the natural and anthropogenically forced carbon cycle (although its strength in the natural state is not calculated by Tsunogai et al. [1999]).

2.3. Parameterizing the CSP

As the processes which drive the CSP cannot be properly represented within the GCM, a more direct approach was used to simulate its effects. The estimate of the pump's strength provided by Tsunogai et al.'s [1999] study (35 g C m\(^{-2}\) yr\(^{-1}\)) was assumed to be constant across all of the world's continental shelves. The shelf areas were then regredded (see Appendix A) onto the GCM's grid, and the total shelf area of any given GCM grid cell was calculated. Figure 2 shows the redistribution of shelf areas onto the GCM grid.

During simulations using the CSP, DIC was added to GCM grid cells according to the following formula:

\[
\left(\frac{\text{pump rate}(1 - \text{ice fraction})[\text{area of shelf in cell}]}{\text{area of cell}}\right) \frac{\text{cell thickness}}{1000}\]

DIC was added to layer 8 of the GCM (113–165 m) where possible. Layer 8 was chosen as the deepest model layer that did not cross the 200 m threshold (the base of layer 9 is 242 m). Where the GCM grid was shallower than layer 8, DIC was added to the deepest layer. Since a large region of the continental shelf zone lies in the Arctic Ocean (and a smaller region around the coastline of
 Antarcti(50), the pump rate was affected by ice cover in the same manner as the air-sea flux.

2.4. DOC/DIC Model Variant

The carbon model outlined above forms the basis for the bulk of the work in this study. As a secondary experiment, a model was constructed in which the carbon added to the ocean via the CSP is added as a DOC-like tracer. Unlike DIC, this tracer does not interact with the atmosphere, although it is remineralized at a fixed rate back into DIC. The conservation equations for this model are

$$\frac{d[\text{DOC}]}{dt} = \theta[\text{DOC}] + \text{CSP} - \lambda[\text{DOC}]$$  \hspace{1cm} (3)

$$\frac{d[\text{DIC}]}{dt} = \theta[\text{DOC}] + F_e + F + \lambda[\text{DOC}],$$  \hspace{1cm} (4)

where [DOC] is the model’s concentration (mol m\(^{-3}\)) of dissolved organic carbon, CSP is the flux term for DOC representing the addition of DIC to the GCM via the parameterized CSP (it is the sole input term for DOC), and \(\lambda\) is the remineralization rate of DOC into DIC. The motivation for this model is the suggestion by Tsunogai et al. [1999] that biological production acts as a mechanism in their proposed CSP. Carbon fixed by biological activity cannot escape to the atmosphere until it has been remineralized back into DIC. This model assumes that all of the carbon sequestered by the pump is of biological origin and aims to examine the significance of the delay a biological loop can introduce and its consequences on the efficiency of the CSP. However, the choice of duration for this delay is not simple. In the real ocean, the pool of DOC is composed of a wide spectrum of organic molecules, pragmatically categorized into labile (short lifespan, days), semilabile (intermediate lifespan, months to years), and refractive (long lifespan, centuries to millenia) forms [Kirchman et al., 1993; Carlson and Ducklow, 1995]. DOC lifespan is a function of many factors, including chemical composition and bonding, oxidation by UV radiation near the ocean surface, microbiological decomposition, adsorption onto particles, and even interactions between DOC molecules themselves [Anderson and Williams, 1999]. The DOC-like tracer in this model most closely resembles semilabile DOC, given that this is produced at relatively high rates (unlike refractive DOC) and survives long enough to be exported deeper into the water column (unlike labile DOC). To simplify the processes involved, a linear remineralization function is assumed, and two simulations are performed, one with a monthly decay rate (\(\lambda = 0.0333 \text{ days}^{-1}\)) and the other with an annual rate (\(\lambda = 0.0028 \text{ days}^{-1}\)).

As a side note, in reality, biological activity produces POC as well as DOC. Chen and Wang [1999] estimate that the East China Sea exports \(8.34 \pm 4.20 \text{ Mt C yr}^{-1}\) of POC, \(~40\%\) of total organic carbon exported. Should the production and export of POC be particularly high and have a size spectrum shifted toward larger particles, the sinking of this material could significantly enhance the action of the CSP. However, to simplify the model, only the effect of a delaying loop on the pump’s activity is examined here.

3. Simulations

3.1. Equilibrium Spin-up

Before any simulations involving the CSP were performed, the carbon cycle model was spun up to approach its baseline equilibrium and to minimize model drift. The OCMIP-2 protocol was followed, and the model was simulated until the globally integrated annual air-sea flux of carbon (i.e., drift) was \(<0.01 \text{ Pg C yr}^{-1}\). This took a period of \(\sim8000\) simulated model years. The end state of this spin-up period was used as the initial condition for all of the simulations performed in this study.

3.2. Experiments

Four simulations were initialized from the equilibrium state: (1) control, DIC only model and no CSP; (2) DIC only, DIC only model and CSP active; (3) DIC monthly, DOC/DIC model, CSP active, and monthly decay of DOC; (4) DIC annual, DOC/DIC model, CSP active, and annual decay of DOC. Each simulation was run for 5000 years. This period was sufficient for the outgas- sing flux of pump carbon to balance the influx parameterized by the CSP to four significant figures. The remaining drift in the simulations was considerably smaller than the equilibrium drift.
Table 1. Breakdown of the Global Shelf Into Regions of Broadly Continuous Shelf

<table>
<thead>
<tr>
<th>Shelf Region</th>
<th>Area, 10^6 km^2</th>
<th>Fraction, %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total world ocean shelf</td>
<td>20.397</td>
<td>100</td>
</tr>
<tr>
<td>Labrador Sea</td>
<td>0.382</td>
<td>1.87</td>
</tr>
<tr>
<td>Grand Banks and Eastern Seaboard</td>
<td>0.887</td>
<td>4.35</td>
</tr>
<tr>
<td>Caribbean</td>
<td>0.744</td>
<td>3.65</td>
</tr>
<tr>
<td>Amazonian shelf</td>
<td>0.661</td>
<td>3.24</td>
</tr>
<tr>
<td>Patagonian shelf</td>
<td>1.539</td>
<td>7.55</td>
</tr>
<tr>
<td>Southwest African shelf</td>
<td>0.335</td>
<td>1.64</td>
</tr>
<tr>
<td>Northwest African shelf</td>
<td>0.264</td>
<td>1.29</td>
</tr>
<tr>
<td>North Sea</td>
<td>1.015</td>
<td>4.98</td>
</tr>
<tr>
<td>Iceland and Greenland</td>
<td>0.270</td>
<td>1.32</td>
</tr>
<tr>
<td>East African shelf</td>
<td>0.243</td>
<td>1.19</td>
</tr>
<tr>
<td>Madagascar shelf</td>
<td>0.087</td>
<td>0.43</td>
</tr>
<tr>
<td>Seychelles</td>
<td>0.073</td>
<td>0.36</td>
</tr>
<tr>
<td>Indian subcontinent shelf</td>
<td>0.443</td>
<td>2.17</td>
</tr>
<tr>
<td>Indonesian archipelago</td>
<td>2.431</td>
<td>11.92</td>
</tr>
<tr>
<td>Okhotsk Sea</td>
<td>0.611</td>
<td>3.00</td>
</tr>
<tr>
<td>East China Sea</td>
<td>1.577</td>
<td>7.73</td>
</tr>
<tr>
<td>Australian shelf</td>
<td>2.445</td>
<td>11.99</td>
</tr>
<tr>
<td>New Guinea and nearby islands</td>
<td>0.120</td>
<td>0.59</td>
</tr>
<tr>
<td>New Zealand shelf</td>
<td>0.236</td>
<td>1.16</td>
</tr>
<tr>
<td>Western South America</td>
<td>0.167</td>
<td>0.82</td>
</tr>
<tr>
<td>Western Central America</td>
<td>0.048</td>
<td>0.24</td>
</tr>
<tr>
<td>Western North America</td>
<td>0.293</td>
<td>1.44</td>
</tr>
<tr>
<td>Bering Sea</td>
<td>1.123</td>
<td>5.51</td>
</tr>
<tr>
<td>Antarctic shelf</td>
<td>0.487</td>
<td>2.39</td>
</tr>
<tr>
<td>Russian Arctic shelf</td>
<td>1.503</td>
<td>7.37</td>
</tr>
<tr>
<td>Siberian Arctic shelf</td>
<td>1.825</td>
<td>8.95</td>
</tr>
<tr>
<td>North American Arctic shelf</td>
<td>0.498</td>
<td>2.44</td>
</tr>
<tr>
<td>Kerguelen Islands</td>
<td>0.057</td>
<td>0.28</td>
</tr>
<tr>
<td>Hawaii</td>
<td>0.006</td>
<td>0.03</td>
</tr>
<tr>
<td>Galapagos Islands</td>
<td>0.009</td>
<td>0.05</td>
</tr>
<tr>
<td>South Orkney</td>
<td>0.009</td>
<td>0.04</td>
</tr>
<tr>
<td>South Georgia</td>
<td>0.007</td>
<td>0.03</td>
</tr>
</tbody>
</table>

criterion. The Control simulation was used to correct the three CSP simulations for drift.

To assess the regional strength of the CSP, the global shelf was divided into 32 geographical areas. Figure 3 shows the location and extent of these regions. Table 1 lists their areas and the fractions of global shelf that they make up. They aim to characterize continuous areas of continental shelf and range in size from 2.445 10^6 km^2 (Australian shelf) to 0.006 10^6 km^2 (Hawaii).

4. Results

Plate 1 shows the actual rate of carbon addition of the parameterized CSP as per (2). Because of ice cover in the Arctic and Antarctic the pump rate is reduced in these areas below that which would be expected given the shelf areas there. This field was applied identically in all three of the simulations that included the CSP.

4.1. DIC Simulation

Figure 4 shows three global diagnostics collected from the four simulations for the full 5000 years (the results of the DOC simulations are discussed more fully below). For all three diagnostics the control simulation shows only slight, if any, drift. The three CSP simulations behave broadly similarly. As the middle panel shows, ocean surface pCO_2 increases as pump carbon returns to the surface. The lower panel shows how this is reflected in the air-sea flux. In all three cases, there is a substantial negative flux as pump carbon escapes back into the atmosphere. However, as the top panel shows, the rate of carbon escape through the simulations is insufficient to prevent a buildup of pump carbon in the oceans of the three pump simulations. At the end of 5000 years, the DIC simulation has retained 43.68 Gt of pump carbon. Given the pump rate of 0.582 Gt C yr^-1, this represents 73.3 years worth of pump carbon. As can be seen from Figure 4, after 5000 years, the simulations have broadly reached, or are close to, equilibrium.

Following on from the bottom panel of Figure 4, Plate 2 shows the global distribution of outgassing pump carbon from the DIC simulation. As can clearly be seen, some pump carbon has reached areas of the ocean far removed from the continental shelves. Of particular note is the outgassing along the northeasterly flowing Gulf Stream. This is primarily carbon entering from the Caribbean Sea and Eastern Seaboard. Pump carbon from the Patagonian shelf also outgasses along a major plume through the action of the Brazil Current and the Antarctic Circumpolar Current (ACC). In several regions though (e.g., Indonesian archipelago, North Sea, and East China Sea), there is a considerable amount of outgassing occurring on the parameterized continental shelf.

Globally, of the 0.582 Gt C annually added by the parameterized CSP, 0.276 Gt C (47.4%) outgasses within the global shelf region. To simplify analysis, this region is defined as all of those GCM cells that overlie a region of the parameterized shelf. As Plate 2 shows, the remaining 52.6% is exported to neighboring or more distant regions of the world ocean where it outgasses.

Table 2 shows a regional breakdown of outgassing and export. Table 2 is ranked by an estimate of the efficiency of export (i.e., the fraction of pump carbon added to a given shelf which is not outgassed on that shelf), and efficiencies range from ~23.7% (Iceland and Greenland) to 91.6% (Hawaii). Negative efficiencies occur where owing to import from another shelf region, more pump carbon is outgassed than added to a particular region. Since the carbon added to a given shelf region is indistinguishable from that added to another region, it is not possible to separate the fraction of outgassing pump carbon of local origin from that imported from another shelf region. This complicates the issue of export efficiency, so the figures calculated should be treated as simple estimates.

There are few obvious patterns to the export efficiencies of the various shelf regions. Although the top two regions (Hawaii and the Galapagos) are both small island shelves, other ostensibly similar regions (Seychelles and South Georgia) have considerably lower efficiencies. The four Western Pacific regions all have relatively high export efficiencies, as do those of the subpolar eastern Atlantic and Western Africa. One pattern which does appear to hold is that the least efficient regions (<30%) are all polar (or subpolar) regions in the Northern Hemisphere (with the sole exception of South Georgia). However, in contrast, high-latitude regions in the Southern Hemisphere, like the Antarctic shelf and the Kerguelen Islands, have efficiencies >60%. By coincidence, the East China Sea, the focus of Tsunogai et al.’s [1999] study, has an export efficiency extremely close to (and the closest to) that of the global average.

Pump carbon enters the modeled ocean near its surface and, as Plate 2 has shown, exits across a relatively large area of its surface. Table 3 and Plates 3 and 4 give a fuller picture of the distribution of pump carbon through the ocean interior at year 5000. Table 3 summarizes the vertical distribution of pump carbon in the world ocean. More than 94% of the total inventory is located deeper than 200 m, and >72% is deeper than 1 km. Note, however, that the actual concentration of stored pump carbon generally falls with depth.

Plate 3 shows the zonal average of pump carbon from the DIC simulation both down the Atlantic basin and back up the Pacific basin. There are considerable differences in the patterns of
storage between the basins. Because of the Atlantic’s role in deep water formation, relatively high concentrations of pump carbon have made their way into midwater (2 km depth), southward moving North Atlantic Deep Water (NADW). As this water mass reaches the ACC, its contents are distributed around the Southern Ocean. Pump carbon from the Patagonian and, particularly, Indian Ocean shelves also helps to raise pump carbon concentrations in the Southern Ocean. Because of the relatively low pump inputs directly to the waters of the Southern Ocean (the Antarctic shelf contributes only 6.7 Mt C yr\(^{-1}\)) its surface waters have considerably less pump carbon than the waters of the Atlantic Ocean.

Concentrations of pump carbon through the Pacific basin are considerably (\(\approx 1.5 \text{ mmol C m}^{-3}\)) lower than those in the Atlantic basin. This is at least partially related to the fact that per unit volume the Pacific receives less pump carbon than the Atlantic (20 versus 57 \(\mu\text{mol m}^{-3} \text{ yr}^{-1}\)) and that Pacific shelves are generally less efficient exporters than Atlantic ones (42 versus 59%). The absence of deep water formation traps most pump carbon in the surface kilometer of the ocean. Pump carbon reaches the deep Pacific partially through vertical diffusion but mostly via the ocean conveyor belt. In the real world this is perhaps most clearly illustrated by the distribution of \(^{14}\text{C}\), the carbon radioisotope. Because this isotope is replenished from the atmosphere and decays on timescales comparable to that of the conveyor, it exists at much depleted concentrations in the deep Pacific [Toggweiler et al., 1989].

Plate 4 complements Plate 3 by showing the vertical inventory of pump carbon from the DIC simulation. The carbon-rich waters of the Atlantic and carbon-poor waters of the Pacific are clearly

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**Plate 1.** Rate of carbon addition due to parameterized “continental shelf pump.” Pump rate is in g C m\(^{-2}\) yr\(^{-1}\).

**Plate 2.** Rate of “continental shelf pump” carbon outgassing from DIC simulation. Control-corrected to remove background carbon air-sea fluxes. Outgassing rate in g C m\(^{-2}\) yr\(^{-1}\).
discernable. The Indian Ocean is also now visible as a major store of pump carbon. Interestingly, the presence of ice cover over the Arctic Ocean traps some of the highest inventories of pump carbon. However, the relatively smaller area of this sea reduces the global importance of this store.

Finally, Table 4 breaks down the properties of the CSP for the global ocean and five ocean basins. The total area portion of the table details the areas of these regions, the fraction of the world ocean that they make up, and the fraction of each of them that is designated “open ocean” (i.e., GCM cells containing no continental shelf). The Indian Ocean and, especially, the Arctic Ocean have relatively greater proportions of their total area containing continental shelf, while the Pacific and Southern Oceans consist of mostly open ocean.

The total pump portion of Table 4 summarizes the CSP inputs to the various basins. Perhaps surprisingly, shelf area (and thus CSP inputs) is not evenly distributed around the globe. The Atlantic, Indian, and Arctic Oceans are comparatively shelf-rich for their sizes, while the Pacific and Southern Oceans have relatively smaller shelf areas.

The total outgassing portion of Table 4 presents summaries of pump carbon outgassing. While the Atlantic and Indian Oceans outgas at lower rates than they receive pump carbon (particularly the Indian Ocean), all of the other basins, especially the Southern Ocean, outgas more pump carbon than they receive. These latter basins are net importers of pump carbon from the Atlantic and Indian basins (which act as net exporters). In the case of the Southern Ocean, the majority of the carbon it receives comes from the Indian basin, and the majority of its imports are outgassed, primarily in open ocean areas (79.3%). A smaller fraction of the carbon reaching the Southern Ocean is transported into the Pacific basin where it outgasses there.

Note that although the value for the open ocean fraction of outgassing for the global ocean matches the global pump efficiency presented in Table 2, the corresponding values presented for the individual ocean basins are not comparable because of transport of pump carbon between ocean basins. Net transport for the global ocean is by necessity zero, but the same is not true for ocean basins.

Finally, the total inventory portion of Table 4 summarizes where stored pump carbon in the ocean is located as well. Although the largest share of pump carbon is found in the Pacific basin (33.8%), when the basin’s large volume is taken into consideration, this translates to an average concentration of pump carbon of 2.160 mmol m$^{-3}$. In contrast, while the Atlantic and Indian basins have relatively lower shares of the total inventory of pump carbon (26.6 and 20.2%, respectively), when considering their lower volumes both have average concentrations of >3.8 mmol m$^{-3}$. Similar to the Pacific Ocean, the Southern Ocean’s share of the inventory (15.4%) is diluted through the basin’s volume to 2.081 mmol m$^{-3}$. Around 4% of the total inventory is stored within 1.2% of the global ocean’s volume. As hinted at in Plate 4, the Arctic Ocean’s large covering of permanent sea ice enables this storage. Pump carbon enters the Arctic Ocean from its seasonally ice-free fringes (where the majority also outgasses) but is able to remain trapped for the longer term under the ice cap which prevents communication with the atmosphere.

4.2. DOC Simulations

In both of the simulations in which a DOC-like precursor was added instead of DIC the quantity of DIC stored at year 5000 is increased. However, the difference in DIC storage between the DOC monthly and DOC annual simulations is marked. Figure 4 shows three global diagnostics for both simulations alongside those from the DIC and control simulations. In all three cases, only the DOC annual simulation can readily be distinguished. At year 5000 the DIC simulation has a global inventory of pump carbon of 43.68 Gt C, while the DOC monthly simulation has 43.95 Gt C (+0.6%)
and the DOC annual simulation has 48.99 Gt C (+12.2%). Figure 5 shows the fractional gain delivered by the DOC phase for both simulations across the first 1000 years of the simulations. In both cases they rapidly (within 200–300 years) equilibriate to constant fractional gains over the DIC simulation. This occurs despite carbon storage increasing throughout the duration of the simulations. As Plate 5 and Table 5 show, however, this global constancy is made up of considerable regional variation. Plate 5 shows the percentage increase in pump carbon inventory resulting from the DOC annual simulation. Note that while increases in pump carbon storage in the shelf regions in Plate 5 often exceed 30%, these apparently extreme increases are usually in areas where carbon storage in the DIC only simulation is low (see Plate 4). At the large scale, while the Atlantic and, to a lesser extent, Pacific and Southern Oceans show relatively constant increases in carbon storage of ~10–15%, the Arctic and Indian Oceans lie outside this range. The Arctic ranges from ~15% to values around 22%, and these are particularly significant given that the Arctic already stores pump carbon at high concentrations. In contrast, the Indian Ocean ranges 5–10% with a strong north-south gradient. In the Arctic Ocean it is almost certainly the case that the lifespan of DOC allows more of it to get under the icecap before it decays to DIC. The variation in the Indian Ocean is more difficult to explain without a detailed analysis of pump carbon fluxes within and between basins. However, some of the variability is presumably due to regions where pump carbon is very quickly subducted or otherwise isolated from the atmosphere (i.e., pump carbon is stored whether DIC or DOC) or where pump carbon is trapped near the surface for periods greater than the lifespan of model DOC (i.e., pump carbon is outgassed whether DIC or DOC).

<table>
<thead>
<tr>
<th>Table 2. Summary of Regional Continental Shelf Pump, On-Shelf Outgassing, Off-Shelf Export, and Export Efficiency of Continental Shelf Pump</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shelf Region</td>
</tr>
<tr>
<td>Hawaii</td>
</tr>
<tr>
<td>Galapagos Islands</td>
</tr>
<tr>
<td>Amazonian shelf</td>
</tr>
<tr>
<td>Western Central America</td>
</tr>
<tr>
<td>Western North America</td>
</tr>
<tr>
<td>Patagonian shelf</td>
</tr>
<tr>
<td>Madagascar shelf</td>
</tr>
<tr>
<td>Northwest African shelf</td>
</tr>
<tr>
<td>New Guinea and nearby islands</td>
</tr>
<tr>
<td>Western South America</td>
</tr>
<tr>
<td>New Zealand shelf</td>
</tr>
<tr>
<td>Southwest African shelf</td>
</tr>
<tr>
<td>Kerguelen Islands</td>
</tr>
<tr>
<td>Antarctic shelf</td>
</tr>
<tr>
<td>Australian shelf</td>
</tr>
<tr>
<td>Indian subcontinent shelf</td>
</tr>
<tr>
<td>Carribean shelf</td>
</tr>
<tr>
<td>Grand Banks and Eastern Seaboard</td>
</tr>
<tr>
<td>East African shelf</td>
</tr>
<tr>
<td>South Orkney</td>
</tr>
<tr>
<td>Total world ocean shelf</td>
</tr>
<tr>
<td>East China Sea</td>
</tr>
<tr>
<td>Indonesian archipelago</td>
</tr>
<tr>
<td>Seychelles</td>
</tr>
<tr>
<td>Labrador Sea</td>
</tr>
<tr>
<td>North Sea</td>
</tr>
<tr>
<td>South Georgia</td>
</tr>
<tr>
<td>Russian Arctic shelf</td>
</tr>
<tr>
<td>Bering Sea</td>
</tr>
<tr>
<td>Siberian Arctic shelf</td>
</tr>
<tr>
<td>North American Arctic shelf</td>
</tr>
<tr>
<td>Okhotsk Sea</td>
</tr>
<tr>
<td>Iceland and Greenland</td>
</tr>
</tbody>
</table>

*Table entries are ranked by export efficiency.*

Table 3. Summary of the Vertical Distribution of Retained Pump Carbon From the DIC Simulation

<table>
<thead>
<tr>
<th>Depth Interval, m</th>
<th>Retained Carbon</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Gt C</td>
</tr>
<tr>
<td>0 – 200</td>
<td>2.5648</td>
</tr>
<tr>
<td>200 – 500</td>
<td>3.7101</td>
</tr>
<tr>
<td>500 – 1000</td>
<td>5.7768</td>
</tr>
<tr>
<td>1000 – 2000</td>
<td>11.0125</td>
</tr>
<tr>
<td>2000 – 5000</td>
<td>20.6151</td>
</tr>
<tr>
<td>Total</td>
<td>43.6794</td>
</tr>
</tbody>
</table>

Table 3. Summary of the Vertical Distribution of Retained Pump Carbon From the DIC Simulation

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YOOL AND FASHAM: CONTINENTAL SHELF PUMP
direction between the simulations. Similar to Table 2, there are few patterns to the changes. As a general rule, previously efficient exporting regions tend to have become less efficient, while those less efficient have become more efficient, although there are notable exceptions to this (the Patagonian shelf in the former case and the North American Arctic shelf in the latter).

Finally, for comparison with Plate 2, Plate 6 shows the difference in patterns of outgassing between the DIC only and the DOC annual simulations. Blue regions are where outgassing is reduced in the DOC simulation, while red regions are where outgassing increases. The general distribution of blue coastal areas and red open ocean areas is consistent with the increase to CSP efficiency by the addition of a DOC phase. Increases are particularly noticeable on the shelves of the eastern Americas, Australia, the East China Sea, and the Bering Strait. In most regions, a decrease in outgassing on shelf is matched by a neighboring off-shelf region of increased outgassing. In the case of the Patagonian shelf the region of decreased outgassing extends markedly into the open ocean.

5. Discussion

This study has aimed to provide a global estimate of the CSP by extrapolating up from the East China Sea study of Tsunogai et al. [1999] and using a simple carbon model and a relatively coarse
It is important to note that because the parameterized pump uses the shelf export rate calculated by Tsunogai et al. [1999], rather than some estimate of the quantity of carbon transferred to the vicinity of the shelf floor, the GCM can only predict a maximum export rate equal to that found by Tsunogai et al. [1999].

GCM-predicted export efficiency gives a more useful measure of the pump strength. If it is very low, the majority of carbon added to the shelf will remain there and outgas locally because it creates a disequilibrium. If high, the majority will be transported and outgassed elsewhere in the GCM.

In the case of the East China Sea itself the GCM predicts that 52.0% of the carbon added near the seafloor of the shelf is exported to either the open ocean or to other shelf regions (neighboring or otherwise). This corresponds to an export of 18.2 g C m$^{-2}$ yr$^{-1}$. Assuming that the GCM correctly models the hydrography around the East China Sea, the argument presented in the preceding paragraph would suggest that to correct for this lower absolute quantity of export (i.e., to bring it up to the same level as found by Tsunogai et al. [1999]), the quantity added to the seafloor should be approximately doubled (ignoring nonlinearities in air-sea exchange and transport).

Scaling up this doubling in the base rate of pump carbon addition to the whole world and assuming that all other things remain equal, this would take the total export of the world ocean’s shelves to 0.589 Gt C yr$^{-1}$ (from an addition to the shelf floor of 1.120 Gt C yr$^{-1}$). This figure is ~27% of Takahashi et al.’s [1999] estimate of the world ocean’s uptake of anthropogenic CO$_2$ during 1995, and such a flux would be significant if ignored in estimates based upon either observations or modeling studies.

Several caveats should be noted however. First, Tsunogai et al.‘s [1999] estimate of the strength of the CSP is based upon contemporary measurements. Currently the world ocean is out of equilibrium with the atmosphere because of the anthropogenic perturbation (although the surface ocean is considerably closer to

| Table 4. A Breakdown of Major Pump Carbon Statistics by Ocean Basins$^a$ |
|-------------------------|------------------|------------------|------------------|------------------|------------------|------------------|
|                        | World            | Atlantic         | Indian           | Pacific          | Southern         | Arctic           |
| Total area, 10$^6$ km$^2$ | 355.95           | 70.00            | 58.86            | 141.63           | 72.49            | 12.96            |
| Basin fraction, %       | ...              | 19.67            | 16.54            | 39.79            | 20.36            | 3.64             |
| Ocean fraction, %       | 82.07            | 77.69            | 69.63            | 89.42            | 89.64            | 39.65            |
| Total pump, Gt C yr$^{-1}$ | 0.5823           | 0.1698           | 0.1967           | 0.1264           | 0.0440           | 0.0354           |
| Basin fraction, %       | ...              | 29.16            | 33.78            | 23.42            | 7.56             | 6.08             |
| Ocean fraction, %       | 0                | 0                | 0                | 0                | 0                | 0                |
| Total outgassing, Gt C yr$^{-1}$ | 0.5823           | 0.1563           | 0.1613           | 0.1493           | 0.0780           | 0.0374           |
| Basin fraction, %       | ...              | 26.84            | 27.70            | 25.64            | 13.40            | 6.42             |
| Ocean fraction, %       | 52.62            | 57.12            | 42.56            | 52.02            | 79.25            | 24.08            |
| Total inventory, Gt C   | 43.68            | 11.62            | 8.84             | 14.76            | 6.73             | 1.73             |
| Basin fraction, %       | ...              | 26.61            | 20.23            | 33.78            | 15.41            | 3.97             |
| Ocean fraction, %       | 87.98            | 85.87            | 77.60            | 94.06            | 95.88            | 72.56            |

$^a$In addition to the four major ocean basins, the Arctic Ocean is also distinguished because of its relatively large continental shelf area. Each of the four statistics (total area, total pump, total outgassing, and total inventory) of the table is broken into three parts: the statistic itself, the fraction of the world ocean each of the basins contributes, and the fraction for each basin that occurs within that basin’s open ocean area. See text for further details.

Figure 5. Fractional gain in carbon storage due to the DOC phase. Dotted line is baseline (DIC simulation); dot-dashed line is DOC monthly simulation; dashed line is DOC annual simulation. Gain is dimensionless.
equilibrium than the deep ocean). At equilibrium the CSP may be less effective, and the premises of this study less secure. In particular, the above global estimate of CSP activity is strongly tied to the pre-industrial equilibrium approach of this study. However, note that Tsunogai et al.'s [1999] observations of the CSP's current role in the East China Sea render it important to contemporary studies irrespective of its role in the natural carbon cycle.

Second, a major assumption in this study has been that the CSP operates equally in all regions of the world shelf. This assumption is critical, and its validity is dependent on patterns of seasonal forcing (physical and biological) and hydrography. In particular, polar shelf regions (which, by area, make up 21% of the modeled shelf) have very different seasonal regimes of temperature and biological activity to those of the East China Sea. However, how common these patterns are across the world shelf cannot be ascertained until a global survey of shelf regions is undertaken with particular regard to the properties that drive the CSP. Synoptic observations such as satellite measurements of SST or surface chlorophyll may provide a qualitative estimate of the occurrence of conditions suitable for strong CSP activity.

For example, in the case of the East China Sea the formation of cool, dense water which both sinks and absorbs CO₂ is crucial to the activity of the CSP. If one restricts sites of potential CSP activity to those regions with a net negative annual heat flux (as determined from the Josey et al. [1999] climatology), this reduces the available shelf area to 13.2%. Calculating the CSP efficiency of

**Plate 5.** Percentage difference in the vertical inventories of continental shelf pump carbon between the DOC annual and DIC simulations.

**Plate 6.** Change in the rate of continental shelf pump carbon outgassing between the DIC only and DOC annual simulations. Blue areas indicate regions of reduced outgassing in the DOC annual simulation, while red areas indicate increased outgassing. Change in outgassing rate is in g C m⁻² yr⁻¹.
Table 5. Export Efficiencies of the Continental Shelf Pump for the DOC Monthly and DOC Annual simulations

<table>
<thead>
<tr>
<th>Shelf Region</th>
<th>DOC Monthly</th>
<th>DOC Annual</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Efficiency</td>
<td>Percent Gain</td>
</tr>
<tr>
<td>Hawaii</td>
<td>91.57%</td>
<td>−0.0</td>
</tr>
<tr>
<td>Galapagos Islands</td>
<td>86.62%</td>
<td>−0.0</td>
</tr>
<tr>
<td>Amazonian shelf</td>
<td>83.84%</td>
<td>−0.0</td>
</tr>
<tr>
<td>Western Central America</td>
<td>81.90%</td>
<td>−0.0</td>
</tr>
<tr>
<td>Western North America</td>
<td>75.78%</td>
<td>−0.0</td>
</tr>
<tr>
<td>Patagonian shelf</td>
<td>74.78%</td>
<td>+0.8</td>
</tr>
<tr>
<td>Madagascar shelf</td>
<td>72.37%</td>
<td>−0.0</td>
</tr>
<tr>
<td>Northwest African shelf</td>
<td>67.96%</td>
<td>+0.0</td>
</tr>
<tr>
<td>New Guinea and nearby islands</td>
<td>66.58%</td>
<td>−0.4</td>
</tr>
<tr>
<td>Western South America</td>
<td>67.58%</td>
<td>+1.2</td>
</tr>
<tr>
<td>New Zealand shelf</td>
<td>67.33%</td>
<td>+1.3</td>
</tr>
<tr>
<td>Southwest African shelf</td>
<td>65.80%</td>
<td>+0.2</td>
</tr>
<tr>
<td>Kerguelen Islands</td>
<td>68.23%</td>
<td>+5.9</td>
</tr>
<tr>
<td>Antarctic shelf</td>
<td>63.90%</td>
<td>+0.4</td>
</tr>
<tr>
<td>Australian shelf</td>
<td>63.56%</td>
<td>+0.5</td>
</tr>
<tr>
<td>Indian subcontinent shelf</td>
<td>63.20%</td>
<td>−0.0</td>
</tr>
<tr>
<td>Carribean shelf</td>
<td>61.88%</td>
<td>+0.4</td>
</tr>
<tr>
<td>Grand Banks and Eastern Seaboard</td>
<td>61.33%</td>
<td>+0.3</td>
</tr>
<tr>
<td>East African shelf</td>
<td>56.89%</td>
<td>−0.0</td>
</tr>
<tr>
<td>South Orkney</td>
<td>54.79%</td>
<td>+2.4</td>
</tr>
<tr>
<td>Total world ocean shelf</td>
<td>52.93%</td>
<td>+0.6</td>
</tr>
<tr>
<td>East China Sea</td>
<td>52.48%</td>
<td>+0.9</td>
</tr>
<tr>
<td>Indonesian archipelago</td>
<td>45.76%</td>
<td>+0.2</td>
</tr>
<tr>
<td>Seychelles</td>
<td>40.67%</td>
<td>−0.2</td>
</tr>
<tr>
<td>Labrador Sea</td>
<td>30.53%</td>
<td>+1.3</td>
</tr>
<tr>
<td>North Sea</td>
<td>29.22%</td>
<td>+1.9</td>
</tr>
<tr>
<td>South Georgia</td>
<td>31.60%</td>
<td>+10.8</td>
</tr>
<tr>
<td>Russian Arctic shelf</td>
<td>26.37%</td>
<td>+1.0</td>
</tr>
<tr>
<td>Bering Sea</td>
<td>25.91%</td>
<td>+1.1</td>
</tr>
<tr>
<td>Siberian Arctic shelf</td>
<td>24.31%</td>
<td>+1.7</td>
</tr>
<tr>
<td>North American Arctic shelf</td>
<td>20.70%</td>
<td>−2.1</td>
</tr>
<tr>
<td>Okhotsk Sea</td>
<td>13.66%</td>
<td>+5.8</td>
</tr>
<tr>
<td>Iceland and Greenland</td>
<td>−23.32%</td>
<td>−1.4</td>
</tr>
</tbody>
</table>

*The percentage changes from the DIC simulation are also shown. Entries are ranked by export efficiency of the DIC simulation.

6. Conclusions

1. Globally, a carbon addition rate of 35 g C m⁻² yr⁻¹, the GCM finds that 52.6% (0.306 Gt C yr⁻¹) is exported from shelf regions of the model.
2. The East China Sea shelf has a export efficiency of 52.0%, very similar to the global average.

3. Export efficiency from other regions of the world varies widely, from 91.6% in Hawaii to ~23.7% in the Iceland and Greenland shelf region.

4. Assuming all other things are equal, scaling pump carbon addition to reproduce Tsunogai et al.’s [1999] East China Sea export would predict a global CSP of 0.589 Gt C yr⁻¹.

5. Despite being added all over the world, pump carbon mostly enters the deep ocean (>1 km) via sites of deep water formation.

6. Because of its permanent ice cover, a large reservoir of pump carbon is located in the Arctic Ocean.

7. Modeling pump carbon as a decaying DOC-like tracer (i.e., temporarily unable to interact with the atmosphere) increases its storage in the ocean (by 12.2% in the case of annually decaying DOC).

Appendix A: Regridding the Continental Shelf

Data used to determine the location of the continental shelf was provided by the National Geophysical Data Center (Boulder, Colorado). The TerrainBase Global DTM version 1.0 database contains a matrix of land topography and ocean bathymetry for the entire world, gridded at 5 min intervals.

For ease of data handling, this matrix was first averaged down to one gridded at 30 min intervals. Then for each GCM cell a list of the overlapping database cells was identified. Weighting the cells on these lists for both area and the fraction they overlapped a given GCM cell, the total area of continental shelf within each GCM cell was calculated. Since values of topography in the database are integers, only cells with values <0 m and >-201 m were counted as continental shelf.

Because of the coarse resolution of the GCM grid, large areas of continental shelf found themselves regridded into GCM cells that are classified as land in the GCM. This is unsurprising given the necessary proximity of areas of continental shelf to land. To remedy this problem, “land-locked” areas of continental shelf were redistributed evenly to ocean GCM cells within the four cell neighborhood of land GCM cells. In principle, this permits ocean GCM cells to contain a greater area of continental shelf than the area of the cells themselves. In practice, only a minority of GCM cells were affected, and given that this study was a parameterization of the CSP, this was not considered a problem. These cells effectively “service” fringe regions of continental shelf inaccessible to the coarse resolution GCM.

Finally, areas of continental shelf occurring within land-locked regions of the GCM (e.g., Baltic Sea, Black Sea, Hudson Bay, Persian Gulf, and Red Sea) were ignored since CSP activity cannot reach the open ocean from these regions. Although a small amount of communication between the Mediterranean Sea and the North Atlantic is parameterized in the GCM, areas of continental shelf within the Mediterranean Sea were ignored because the GCM’s low resolution prevents a reasonable circulation from developing within this basin.

The result of this procedure was a 2-D field giving the total continental shelf area in each open ocean GCM cell (see Figure 2). Table A1 summarizes the changes to ocean shelf area produced by the regridding procedure. Note that although the final shelf area is only $20.4 \times 10^6$ km², this area is contained within GCM grid cells with a total sea area of $68.3 \times 10^6$ km².

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References


Table A1. Total Shelf Area From the Topographic Database and How it Changes as the Shelf is Regridded and Redistributed onto the GCM Matrix

<table>
<thead>
<tr>
<th>Process</th>
<th>Area, 10⁶ km²</th>
<th>Fraction, %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Area calculated from topology</td>
<td>24.405</td>
<td>100</td>
</tr>
<tr>
<td>Regridding to GCM matrix</td>
<td>24.395</td>
<td>99.9</td>
</tr>
<tr>
<td>Masking with GCM land</td>
<td>18.896</td>
<td>77.4</td>
</tr>
<tr>
<td>Four cell neighborhood</td>
<td>23.157</td>
<td>94.9</td>
</tr>
<tr>
<td>Remove inland seas</td>
<td>20.397</td>
<td>83.6</td>
</tr>
</tbody>
</table>


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