Supplement of

Reviews and syntheses: The biogeochemical cycle of silicon in the modern ocean

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SUPPLEMENTAL INFORMATION

This document complements the review article. It provides detailed legends for Figures 1, 2 and 4, and a few additional comments to the main text. Annex 1 shows data for the determination of biogenic silica (bSi) production measured by isotopic techniques.

Section 1. Introduction

Detailed legend of Fig. 1 and flux abbreviations

Schematic view of the Si cycle in the modern world ocean (input, output, and biological Si fluxes), and possible balance (total Si inputs = total Si outputs = 15.6 Tmol-Si yr\(^{-1}\)) that is in reasonable agreement with the individual range of each flux (F), see Tables 1 and 2. All fluxes are in Tmol-Si yr\(^{-1}\).

-Inputs: Rivers: F_{R(dSi+aSi)} = 8.1; Aeolian inputs: F_A = 0.5; Glacial meltwater: F_{ISMW} = 0.3; Submarine groundwater: F_{GW} = 3.1; Dissolution of minerals: F_W = 1.9; Hydrothermal: F_H = 1.7

-Outputs: Burial: F_{B (net deposit)} = 9.2; Sponges: F_{SP} = 1.7; Reverse weathering: F_{RW} = 4.7.

-Biological and other fluxes: Uptake (pelagic production) = F_{P(gross)} = 255; Recycling (surface) F_{D(surface)} = 143 (D: P = 0.56 according to Tréguer and De La Rocha 2013); Export: F_E = 112; Recycling (deep) F_{D(deep)} = 28 (F_{D(deep)}/F_E=0.25 according to Tréguer and De La Rocha 2013); Rain = F_{S(rain)} = 84.0; Recycling (sediment-water interface) F_{D(benthic)} = 74.8; upwelling, diffusion: F_{upw/ed} = 102.8.

Flux abbreviations: dSi input fluxes: F_R total river net discharge, F_A dissolution of aeolian - transported siliceous dusts, F_{GW} submarine groundwater discharge, F_W dissolution of siliceous material transported from land on the continental margins, and of basalt, F_H hydrothermal activity of the oceanic ridges (axis + off axis), F_{ISMW} ice shelf melt water flux (subglacial melt water + basal melting of ice shelves + melting of icebergs).

Si output fluxes: F_{B(net deposit)} long-term burial of biogenic silica, F_SP siliceous sponges, F_{RW} reverse weathering by formation of authigenic silicate minerals.

Biological fluxes: F_{P(gross)} production of biogenic silica due to diatoms and other pelagic silicifiers, F_{E(export)} export flux of biogenic silica to the deep reservoir, F_{S(rain)} part of the export flux that reaches the sediment - water interface, F_{B(net deposit)} long-term accumulation of biogenic opal in coastal and abyssal, sediments.
Note that the dSi uptake due to benthic organisms is not represented on Fig. 1. Indeed, the total bSi production of benthic diatoms is presently unknown but it should be < 5 Tmol-Si yr\(^{-1}\) (Leynaert et al., work in progress), and that the total bSi production of sponges (6.1 (±5.9) Tmol-Si yr\(^{-1}\)) is still preliminary (Maldonado et al., work in progress).

Other fluxes: \(F_{\text{D(surface)}}\) recycling of Si by dissolution of the biogenic silica in the surface reservoir, \(F_{\text{D(deep)}}\) part of the export flux that dissolves in the deep reservoir, \(F_{\text{upw/ed}}\) transfer of dSi from the deep to the surface reservoir by upwelling or eddy diffusion. \(F_{\text{D(benthic)}}\), flux at the sediment - water interface, is according to Tréguer and De La Rocha 2013).

Section 2 - The input fluxes

2.1 Detailed legend of Figure 2.

Schematic view of the low temperature processes that control the dissolution of (either amorphous or crystallized) siliceous minerals in seawater in the coastal zone and in the deep ocean, feeding \(F_{\text{GW}}\) and \(F_{\text{W}}\). These processes (white arrows) correspond to low or medium energy flux dissipated per volume of a given siliceous particle in the coastal zone, in the continental margins, and in the abysses, and to high kinetic energy flux dissipated in the surf zone. Inputs of siliceous biogenic and lithogenic silica into the ocean are mainly due to suspended matter transferred from the continent. Rivers support the main transfer of Si to the coastal zone, either as dSi or as aSi. Abundant transfer of siliceous suspended matter into the ocean is also expected from river mouths and deltas through dissemination of suspended matter in the coastal zone, in the continental margin, and beyond to the abysses. In sandy and permeable soil zones dSi is also transferred from the continent to the coastal zone through submarine groundwater discharge processes either as net fresh water inputs or as brackish/seawater recycling due to tidal pumping. In the surf zone, low temperature dissolution of grains of lithogenic silica (quartz, feldspar, etc.) could be intense under the pressure of the intensive and continuous shaking due to waves. For the sake of clarity, we distinguish the processes at work in a quiet zone that receive river inputs from those occurring in a sand beach zone subject to strong dissipation of energy due to wave motion. Note that neither the low-temperature dissolution of wind-borne siliceous material (\(F_{\text{A}}\)) nor that occurring in high-temperature hydrothermal systems (\(F_{\text{H}}\)) are represented on this figure.

2.2 Dissolution of minerals

Regarding the marine component of \(F_{\text{W}}\) dissolution of minerals, Fabre et al. (2019) (main text reference list), focused on wave and tidal action prevailing in the intertidal surf zone (Fig. 2).

From laboratory experiment with pure quartz, they showed that quartz grains submitted to
violent agitation are capable of substantial dissolution of silica at time scale of days. According
to these authors, the flux of dissolution of siliceous material from sandy beaches is \( F_{\text{dissolution}} = k(T^\circ C) \times S_{\text{reactive}} \times (C^* - C_{\text{sw}}) \). It corresponds to a net input of dSi to the ocean. In this equation, 

\( k \) (f(T°C)) in m s\(^{-1}\) is the mass transfer coefficient, \( S_{\text{reactive}} \) in m is the reactive surface of sand grains, \( C^* \) in mol-Si L\(^{-1}\) is the temperature-dependent solubility limit of sand at thermodynamic equilibrium, and \( C_{\text{sw}} \), in mol-Si L\(^{-1}\) is the coastal seawater silicic acid concentration. Assuming that all sandy beaches is composed of quartz, Fabre et al. (2019) calculated a global flux of 3.2 (± 1.0) Tmol Si yr\(^{-1}\). However, this estimate is not well constrained.

Firstly, it is clear that the mineral composition of the world ocean beaches is not entirely composed of quartz. Indeed, the composition of sandy beaches is variable and represent 31% of the coastline of the continents at world scale (Luijendijk et al, 2018, main text reference list). Sandy beaches are composed both of non-siliceous materials (mostly calcareous), and diverse siliceous types of materials (amorphous silica, quartz, feldspars, clays, etc.). These siliceous materials have different solubility and dissolution rates in seawater (S1: Lerman et al., 1975; S2: Hurd et al., 1979), and they can be more or less coated by organic and metals, which affect silica dissolution (e.g. S3: Loucaides et al. 2010; S4: Wiley, 1975). These differing sand compositions directly affects the value of \( S_{\text{reactive}} \) as well as that of \( C^* \).

Secondly, in surf zones the mixing conditions of sand and seawater can be very variable, both over time and at local and regional scale, thus affecting the solid to liquid phase ratio, and the values of the \( C^* \).

Thirdly, Fabre et al (2019)'s value for \( C_{\text{sw}} \) (i.e. 85.4 µM), is far too high for coastal waters, irrespective of its regional context in the world ocean.

Finally, \( C_{\text{sw}} \) can be seasonally variable, particularly in temperate zones.

Also note that, by definition, this flux is already included in the marine SGD estimate according to Cho et al. (2018) (main text reference list)

**REFERENCES:**


**Section 3 - The output fluxes**
Deposition of sponge silica in marine sediments

The annual rate of sponge silica deposition to the sediments, unlike in diatoms, cannot be easily calculated from annual production rates. The longevity of sponges, which ranges from months or years to even centuries or millennia (S5: Elwood et al., 2007; S6: Mcmurray et al., 2008; Jochum et al. 2017, see main text reference list), decouples the process of skeleton production — which slowly accumulates bSi over the sponge lifespan — from the process of releasing the accumulated bSi into the sediments — which occurs within weeks to months after sponge death. The deposition of sponge bSi is also decoupled from the rain of planktonic, and therefore, from the rate of sediment deposition. The reason of this is in the benthic nature of sponges. Sponges live already attached to the bottom, so their bSi does no transit through the water column once sponges die. While the organic components of the body become rapidly degraded (S7: Rützler and Mcintyre 1978), the mineral components (i.e. the siliceous skeletal pieces of sponges, called spicules) do not. The skeletal pieces fall directly on the seafloor at the site where the sponge was growing, forming a spicule patch (S8: Laguionie-Marchais et al., 2015), which can in some cases persist for a long time (S9: Bett and Rice, 1992), being slowly disaggregated by the action of scavengers and other benthic macrofauna (S10: Katz et al., 2016), also by the action of bottom currents, turbidity currents included.

Once the spicules are delivered to the bottom, the period of time needed for them to be buried and become accumulated bSi will mostly depend on the local rates of sediment deposition, though also on the intensity of bioturbation (S10: Katz et al., 2016). In a study that has considered sediments from a variety of marine environments ranging from shallow bays to abyssal bottoms, the time required for the sponge spicules to reach the condition of permanently buried bSi ranged from 471 to 74,074 years, depending on the depositional nature of the local bottom (S11: Maldonado et al., 2019).

REFERENCES:


**Section 4 - The biological fluxes**

4.1 Biogenic silica production as measured by isotopic techniques

Annex 1 shows all bSi production data with corresponding references.

4.2 bSi pelagic production uncertainties

The bSi pelagic production estimates from satellite NPP products, ocean biogeochemical models, and empirical studies each have their own uncertainty. When NPP is extrapolated to silica production, values are multiplied by estimates of the fraction of primary productivity done by diatoms and then a Si:C ratio. Of these, the choice of a Si:C ratio for the HNLC regions is the most uncertain. Our chosen value of 0.52 is 4 fold higher than for nutrient-replete temperate diatoms (Brzezinski, 1985, see main text reference list), but field observations suggest anywhere from no effect to an 8-fold increase (S12: Marchetti & Harrison, 2007, S13: Timmermans et al., 2004). Biases in satellite NPP models also contribute to uncertainties in estimates of Si production. Particularly relevant are potential biases in Southern Ocean chlorophyll concentrations (and consequently, NPP), which may be underestimated in the Southern Ocean by as much as a factor of 3-4 (S14: Johnson et al., 2013).

For the biogeochemical models the two main sources of uncertainty are the extrapolation from silica export to gross silica production using D:P ratios (Table 1 of Tréguer & De La Rocha, 2013, main text reference list), as well as uncertainties in the parametrization of Southern Ocean physics and biology. For the seven GOBMs that report separate estimates of net silica production for the Southern Ocean, when Southern Ocean silica production is regressed against total global silica production the fitted line has an $R^2$ value of 0.96 with a slope equal to 1.05 ($\pm 0.11$). This points to model parameterizations of Southern Ocean physics and/or biology as the major determinant of differences in global silica production estimates among GOBMs.

For the global bSi pelagic production estimates derived from field data, extrapolation over both time and space is required. Few empirical studies of bSi production over an entire year are
available. The vast majority of annual estimates determined for Longhurst provinces are
extrapolated from field programs lasting a few weeks or less. Data sparsity remains a problem,
and in the analysis presented here some Longhurst zones contain only a single measurement
and nearly half of zones have no data at all (main text Fig. 3).

REFERENCES

S12. Marchetti, A., Harrison, P.J. Coupled changes in the cell morphology and elemental (C, N, and Si)
composition of the pennate diatom *Pseudo-nitzschia* due to iron deficiency. *Limnol. Oceanogr.* 52, 2270–
2284 (2007).

S13. Timmermans, K.R., van der Wagt, B., de Baar, H.J.W. Growth rates, half-saturation constants, and
silicate, nitrate, and phosphate depletion in relation to iron availability of four large, open-ocean diatoms

S14. Johnson, R., Strutton, P.G., Wright, S.W., McMinn, A., Meiners, K.M. Three improved satellite

4.3 bSi production: contribution of benthic diatoms

Determination of bSi production by microphytobenthos is still in its infancy. However, we can
already foresee its potential impact in two ways. Through intensive studies in a temperate
subtidal ecosystem (S15: Ni Longphuirt et al. 2009, S16 : Leynaert et al. 2009, and S17 :
Chatterjee et al., 2013) we conservatively estimate that benthic diatoms can produce 1 mol-Si
m⁻² yr⁻¹. Extrapolation to the photic area of the world coastal ocean gives 6.8 Tmol-Si yr⁻¹.
Cahoon (1999) (S18) proposed a global estimate of annual benthic microalgal primary
production (based on the analysis of 85 worldwide studies) of 514 × 10¹² gC yr⁻¹. Applying the
Brzezinski (1985) (main text reference list) mean Si/C molar ratio for marine diatom of 0.13 to
convert this carbon production in silica production, gives 5.4 Tmol-Si yr⁻¹. Both estimates are
in good agreement. Therefore, the biogenic silica production by the microphytobenthos might
represent about 2.5% of the global bSi production. It remains to be determined what proportion
of this flux will finally contributes to the net sink of bSi. Preliminary studies have shown that
bSi dissolution rates of benthic diatoms are 10 times slower than pelagic diatoms measured in
the same conditions.

REFERENCES

S15. Ni Longphuirt, S. *et al.* Diurnal heterogeneity in silicic acid fluxes in shallow coastal sites: causes and

S16. Leynaert, A., Longphuirt, S. N., Clauquin, P., Chauvaud, L. & Ragueneau, O. No limit? The multiphasic

Section 5 – Discussion

5.1 Introduction to Fig. 4: Depiction of a schematic Si cycle in the coastal and continental margin zone (CCMZ), excluding coastal upwelling, linked to the rest of the world ocean (« open ocean » zone including upwelling and polar zones). In principle, the CCMZ comprises proximal and distal coastal zones, as defined by Laruelle et al. (2009) (see main text reference list), which includes coastal upwelling in the distal coastal zone. However, in coastal upwelling zones both physical and biogeochemical dynamics are markedly different from those in the CCMZ as represented by Jeandel et al. 2016 (see main text reference list). Indeed, if in the CCMZ the transfer of material from land to ocean plays a major role in physical and biogeochemical processes, in coastal upwelling zones the bSi production is mainly fueled by dSi flux from below (i.e. from deeper in the water column), as it does for the global ocean sensu lato (Fig. 1).

Therefore, in our synthesis quantifying the Si cycle in the « boundary exchange » zone, coastal upwelling is discarded from the CCMZ and conceptually incorporated to the « open ocean » zone, as shown in Fig. 4. This figure represents a possible Si cycle assuming steady state in the CCMZ and in the rest of the world ocean (so called « open ocean »), that is with total inputs = total outputs = 15.6 Tmol-Si yr⁻¹.

5.2 Detailed legend of Fig. 4

Input and dSi fluxes (grey arrows), outputs and biological fluxes (black arrows). In this steady-state scenario total inputs = total outputs = 15.6 Tmol-Si yr⁻¹ (consistent with main text Figure 1). Note that (1) in the CCMZ the burial flux of bSi (3.7 Tmol-Si yr⁻¹) and the reverse weathering flux (4.7 Tmol-Si yr⁻¹) (authigenic siliceous material) are fed by the export flux of biogenic silica, and (2) the « open ocean » deficiency in dSi (4.7 Tmol-Si yr⁻¹) is made up by a transfer from the CCMZ. For details about those inputs and outputs refer to the « inputs » and « outputs » sections of the main text. For abbreviations, see detailed legend of Fig. 1 in Supplement (section 1).

Additional comments of Figure 4:

- Estimates of burial rate (planktonic bSi): in the CCMZ (excluding coastal upwelling), according to Rahman et al. 2017 (see main text reference list), F_B_CCMZ = 3.7 Tmol-Si yr⁻¹; in the « open ocean » zone, consistent with Fig. 1, F_B_openocean= 9.2 - 3.7 = 5.5 Tmol-Si yr⁻¹.
Estimates of reverse weathering flux: in Fig. 4 CCMZ $F_{RWCCMZ}$ equals 4.7 Tmol-Si yr$^{-1}$, (Rahman et al. 2017, see main text reference list). Reverse weathering in the «open ocean» remains unquantified.

Following Annex 1, bSi production in the CCMZ* is 13% of the total production (255 Tmol-Si yr$^{-1}$). Consequently, $F_{P(gross)}$ is 33 and 222 Tmol-Si yr$^{-1}$, for CCMZ and “open ocean”, respectively.

According to Tréguer and De La Rocha (2013) (see main text reference list), the pelagic production to dissolution ratio (D:P) being 0.51 and 0.57, for CCMZ and “open ocean”, respectively. $F_{D(surface)}$ for those compartments is 17 and 124 Tmol-Si yr$^{-1}$, and $F_{E}$ is 16 and 98 Tmol-Si yr$^{-1}$, for CCMZ and “open ocean”, respectively.

Note than every component of the inputs, outputs, and biological Si fluxes, although interacting between each other, is determined by independent methods, that is there is no overlap in the counting of these fluxes. Therefore, the export production (16.0 Tmol-Si yr$^{-1}$) in particular feeds both the Si burial rate (3.7 Tmol-Si yr$^{-1}$) and the reverse weathering flux (4.7 Tmol-Si yr$^{-1}$).

Figure 4 also shows that the “open ocean” bSi production is mostly fueled by dSi inputs from below (92.5 Tmol-Si yr$^{-1}$), the CCMZ only providing 4.7 Tmol-Si yr$^{-1}$ to the “open ocean”.

*Although specific estuarine sites have been studied (e.g. S19 DeMaster 1983, S20, Raimonet al. 2013), no global estimate is presently available for the estuarine bSi production. Given that most of estuarine waters are turbid we anticipate that the contribution of estuaries to the total bSi of the coastal zone should be small.

REFERENCES
### Average per province

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### Average per domain (Longhurst & Tréguer and Jacques for the OA)

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<td>Open Ocean</td>
<td>Westerlies*</td>
<td>130</td>
<td>228</td>
<td>29 647</td>
<td>10,4</td>
</tr>
<tr>
<td></td>
<td>Trades</td>
<td>140</td>
<td>360</td>
<td>50 416</td>
<td>17,7</td>
</tr>
<tr>
<td>Annual World PBSi</td>
<td></td>
<td></td>
<td></td>
<td>284 926</td>
<td></td>
</tr>
</tbody>
</table>

*Surface considered: this is the ocean surface minus the Antarctic ocean

*Polar: surface includes the Antarctic Ocean (from the polar front poleward, i.e. 44 10⁶ km²) + arctic zones as defined by Longhurst

*Westerlies: includes the subantarctic zone

*Southern Ocean: includes the Antarctic Ocean and the subantarctic zone (i.e. respectively 44 10⁶ km² + 29 10⁶ km²)

### REFERENCES


S39. Leblanc, K. et al. Silicon cycle in the Tropical South Pacific: evidence for an active pico-sized siliceous


S57. Wong, C. S. & Matear, R. J. Sporadic silicate limitation of phytoplankton productivity in the subarctic


