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Characterizing the marine iodine cycle and its relationship to ocean deoxygenation in an Earth System model

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Abstract. Iodine abundance in marine carbonates (as an elemental ratio with calcium – I:Ca) is of broad interest as a proxy for local/regional ocean redox. This connection arises because the speciation of iodine in seawater—in terms of the balance between iodate (IO_3^-) and iodide (I^-)—is sensitive to the prevalence of oxic vs. anoxic conditions. However, although I:Ca ratios are being increasingly commonly measured in ancient carbonate samples, a fully quantitative interpretation of this proxy is hindered by the scarcity of a mechanistic and quantitative framework for the marine iodine cycle and its sensitivity to the extent and intensity of ocean deoxygenation. Here we present and evaluate a representation of marine iodine cycling embedded in an Earth system model ('cGENIE') against both modern and paleo observations. In this, we account for IO_3^- uptake and reduction by primary producers, the occurrence of ambient IO_3^- reduction in the water column, plus the re-oxidation of I^- to IO_3^-. We develop and test a variety of different mechanistic relationships between IO_3^- and I^- against an updated compilation of observed dissolved IO_3^- and I^- concentrations in the present-day ocean. In optimizing the parameters controlling previously proposed mechanisms behind marine iodine cycling, we find that we can obtain broad matches to observed iodine speciation gradients in zonal surface distribution, depth profiles, and oxygen deficient zones (ODZs). We also identify alternative, equally well performing mechanisms which assume a more explicit mechanistic link between iodine transformation and environment. This mechanistic ambiguity highlights the need for more process-based studies on modern marine iodine cycling. Finally, because our ultimate motivation is to further our ability to reconstruct ocean oxygenation in the geological past, we conducted 'plausibility tests' of our various different model schemes against available I:Ca measurements made on Cretaceous carbonates – a time of substantially depleted ocean oxygen availability compared to modern and hence a strong test of our model. Overall, the simultaneous broad match we can achieve between modelled iodine speciation and modern observations, and between forward-proxy modelled I:Ca and geological elemental ratios supports the application of our Earth system modelling in simulating the marine iodine cycle to help interpret and constrain the redox evolution of past oceans.

1. Introduction

Dissolved Iodine (I) in seawater is redox sensitive and as such, a potential invaluable delineator of past ocean deoxygenation. This arises directly from: (1) observations that the oxidized iodate (IO_3^-) is reduced to iodide (I^-) under low oxygen conditions, and (2) that IO_3^- in seawater is incorporated into carbonate lattice during precipitation in proportion to its seawater abundance (whilst I^- is not) (Lu et al., 2010; Podder et al., 2017; Kerisit et al., 2018; Zhang et al., 2013; Hashim et al., 2022). Hence, past ocean IO_3^- concentrations can be recorded in coeval carbonates as I:Ca

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ratios, with the potential of carbonate I:Ca to reflect the redox variation of the ancient seawater (Lu et al., 2010). Indeed, the I:Ca ratio in marine carbonates is already widely applied as a paleoredox proxy, with studies employing it to explore the [O_2] variation throughout much of Earth history, from the Archean and through the Cenozoic (Lu et al., 2010; Hardisty et al., 2014; Zhou et al., 2015; Lu et al., 2016; Edwards et al., 2018; Lu et al., 2018; Bowman et al., 2020; Pohl et al., 2021; Wei et al., 2021; Ding et al., 2022; Shang et al., 2019; Liu et al., 2020; Fang et al., 2022; Uahengo et al., 2020; Yu et al., 2022; Tang et al., 2023). The potential for I:Ca to generate critical insights into how the oxygenation of the ocean has evolve through time, as well as the causes and biological/ecological consequences of this, requires that we have an adequate understanding, not only of carbonate IO_3^- incorporation, but of the marine iodine cycle in general.

Progress has been made towards understanding the marine iodine cycle in the past decades. Iodine has a long residence time (~300 kyr; Broecker and Peng, 1983), making its concentration among the global ocean relatively constant around 500nM (Elderfield and Truesdale, 1980; Truesdale et al., 2000; Chance et al., 2014). Although the total concentration is relatively invariant, the two most abundant species of dissolved iodine in the ocean, IO_3^- and iodide I^-, vary relative to each other depending on the environment. Today, IO_3^- is generally the dominant iodine species in oxygenated regions of the ocean, representing total iodine nearly quantitatively below the euphotic zone. Within the euphotic zone, I^- occurs and generally increases in association with release during phytoplankton growth and senescence (Hepach et al., 2020). Iodide is also more abundant in oxygen deficient zones (ODZ) – often but not always quantitatively so (Truesdale et al., 2000; Rue et al., 1997; Cutter et al., 2018; Moriyasu et al., 2020; Farrenkopf and Luther, 2002; Wong and Brewer, 1977; Truesdale et al., 2013; Rapp et al., 2020, 2019). Respectively, within the ODZ, IO_3^- is reduced to I^- and hence has low concentrations (Rue et al., 1997; Farrenkopf et al., 1997; Moriyasu et al., 2020; Rapp et al., 2019, 2020).

Although generally depleted in low-[O_2] settings, the causal spatial relationship between seawater [O_2] and [IO_3^-] is not simple and is currently not well understood. Recently published observations from global oxygen deficient zones (ODZ) reveals that the relationship between dissolved [O_2] and [IO_3^-] is not linear, but instead it is possible that there is a certain [O_2] or related redox threshold triggering IO_3^- reduction (Cutter et al., 2018; Moriyasu et al., 2020; Farrenkopf and Luther, 2002; Rue et al., 1997; Chapman, 1983). Dissimilatory IO_3^- reducing bacteria, as well as abiotic reduction with sulfide and dissolved Fe, have been identified within ODZs (Farrenkopf et al., 1997; Councell et al., 1997; Jiang et al., 2023). In addition, slow oxidation-reduction kinetics (Tsunogai, 1971; Hardisty et al., 2020; Schnur et al., 2024) imply the likelihood that *in situ* iodine signals could be integrated across large-scale physical oceanography processes – including ocean currents and mixing between water masses (Hardisty et al., 2021), and meaning that iodine species reflect regional rather than local redox conditions (Lu et al., 2020b). Non-redox related processes, such as phytoplankton-mediated IO_3^- reduction and organic matter remineralization also exerts controls on iodine speciation in the water column (Fig. 1; Elderfield and Truesdale, 1980; Wong et al., 1985; Luther and Campbell, 1991; Hepach et al., 2020). Therefore, it is difficult to infer water column redox simply based on iodine speciation without considering these physical and biological effects.

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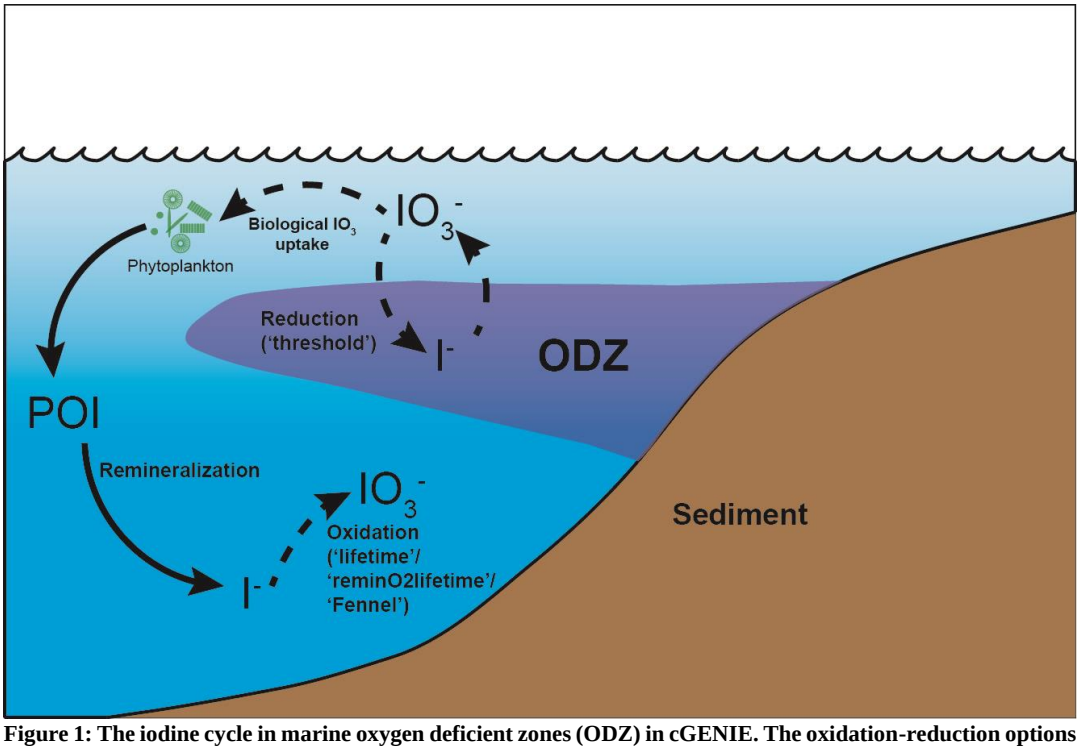

Figure 1: The iodine cycle in marine oxygen deficient zones (ODZ) in cGENIE. The oxidation-reduction options ('threshold' and 'lifetime'/ 'reminO2lifetime'/ 'Fennel') are described in Section 2.2 and Table 1. Dashed arrows indicate variable processes during ensemble simulations. Note that the POI export is controlled by temperature (TDEP).

Apart from the uncertainties associates with IO_3^- reduction, it is notable that the oxidants responsible for IO_3^- formation during the re-oxidation I^- are currently unknown, only that it is unlikely to be O_2, which is not thermodynamically favorable to oxidize iodine (Luther et al., 1995). A recent thermodynamic review indicates that the reactive oxygen species (ROS) such as hydrogen peroxide and OH radicals can fully oxidize I^- to IO_3^-. Iodide oxidation to IO_3^- is a 6-electron transfer and other ROS, such as superoxide, are only thermodynamically favorable to catalyze partial oxidation to intermediates (Luther, 2023). These ROS species have heterogenous distributions and ambient ocean concentrations that are typically relatively low compared to iodine, supporting the likelihood of temporally or spatially isolated high I^- oxidation rates despite of overall extremely slow rates (Schnur et al., 2024). Additional support for spatially or temporally heterogenous I^- oxidation rates comes from recent experimental observations of IO_3^- production from I^- in nitrifying cultures (Hughes et al., 2021). Nitrification rates vary globally, with the highest values occurring in ODZs and the dissolved chlorophyl maximum (summarized in Table 2 of Hughes et al., 2021). Regardless, nitrification or other specific mechanisms have yet to be linked directly to I^- oxidation under normal marine conditions, leaving open the question of rates and locations of I^- oxidation.

Given the prevailing uncertainty in the mechanisms governing the marine iodine cycle mentioned above, and in conjunction with I:Ca being a relatively new proxy, it is perhaps not surprising that few attempts have been made

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to model the marine iodine cycle. In a recent publication, a model was developed to simulate modern ocean surface I^- distributions, with the aim of being able to improve tropospheric ozone models (Wadley et al., 2020). This particular model was based around a relatively high horizontal ocean resolution (1° grid size) with a 3-layer vertical water column. Iodine biogeochemical cycling was coupled with the nitrogen cycle, with the surface I^- distribution sensitive to biological and hydrological factors including primary productivity, I:C ratio, oxidation, mixed layer depth, advection, and freshwater flux. Because the Wadley et al., (2020) model was specifically focused on near-surface processes within the upper 500 m, it did not consider processes occurring within ODZs and hence is not directly applicable to questions concerning the controls on I:Ca ratios. In contrast, a second model-based study deliberately targeted paleoceanographic questions and incorporated an iodine cycle including redox-controlled biogeochemical reactions into the 'cGENIE' Earth system model (Lu et al., 2018). The advantage for paleo studies afforded by this particular approach is that the cGENIE model can take into account different continental configurations, non-modern atmospheric composition (pO_2, pCO_2), and other boundary conditions that may have differed on ancient Earth relative to today (Ridgwell et al., 2007; Reinhard et al., 2016; Boscolo-Galazzo et al., 2021; Remmelzwaal et al., 2019; Pohl et al., 2022; Reinhard and Planavsky, 2022).

Despite a growing understanding of I:Ca variations through geologic time, it remains challenging to determine mechanisms driving spatiotemporal marine [IO_3^-] and the degree that these are linked to seawater oxygen. Hence, the proxy is qualitative or semi-quantitative. Here, we calibrate the iodine cycle within the cGENIE Earth System model to provide a mechanistic framework for interpreting ancient I:Ca variations. In this study, we build on the work of Lu et al., 2018 and further develop and test a series of new potential parameterizations for water column iodine oxidation, reduction, cellular uptake, and release during remineralization. We also developed 3 criteria for assessing the model: (1) Statistical evaluation using the 'model skill score' (Watterson, 1996)— a non-dimensional measure calculated using location-dependent comparisons between the model and an iodine ocean observation data compilation. (2) Graphical comparison of modeled and observed iodine across 3 illustrative iodine speciation gradients (depth profiles from multiple ocean basins, latitudinal transects of surface waters, and across transects of the Eastern Tropical North Pacific oxygen minimum zone (Moriyasu et al., 2020)). (3) Model applicability to ancient settings by comparing projections of ocean surface I:Ca with published I:Ca data from the Cretaceous (Zhou et al., 2015).

2. Model Description

2.1 The cGENIE Earth system modelling framework

cGENIE is a class of model known as an 'Earth system model with intermediate complexity' (EMIC)—a global climate-carbon cycle model that simplifies one or more (typically physical climatic) components of the Earth system. In the case of cGENIE, ocean circulation is solved for on a relatively low-resolution grid: an equal area 36×36 grid, which equates to 10° in longitude and latitude increments from 3° near the equator to 20° near the poles, and with 16 non-equally spaced vertical levels. This is coupled to a 2D energy-moisture-balance-model (EMBM) and a 2D dynamic-thermodynamic sea-ice model. The physics are described in (Marsh et al., 2011; Edwards and Marsh, 2005). We use a parameter calibration of seasonal pre-industrial climate following Cao et al., (2009).

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The primary factors controlling the oceanic iodine cycle—specifically, biological productivity, remineralization, and water column redox—are all represented in the model and described in Ridgwell et al., (2007). In that particular configuration, the rate of organic matter export from the ocean surface is calculated based on just a single nutrient (phosphate) control (together with light and sea-ice cover) and assumes a Redfield-ratio stoichiometry with carbon (Fig. 1). Organic matter is split into particulate (POM) (33% of total export) and dissolved form (DOM) (67%), with the former sinking down through the water column where it is progressively remineralized according to a prescribed fixed 'decay' curve, while the latter is physically transported by circulation and decays (is remineralized with a lifetime of 0.5 years). Here, we deviate from Ridgwell et al., (2007) (as well as the calibrated seasonal configuration of Cao et al., 2009) by adopting a calibrated temperature-dependence to both export production as well as the decay of POM in the water column (described in Crichton et al., 2021 and Boscolo-Galazzo et al., 2021).

Here, we use sulphate (SO_4^{2-}) as an electron acceptor supporting the remineralization of organic matter (both POM and DOM) is governed by a SO_4^{2-} half-saturaion limitation term as well as dissolved oxygen (O_2) inhibition, while the rate of oxic resperation of organic matter is restricted by an [O_2] half-saturaion limitation term (as described in Reinhard et al., 2020). This deviates from the framework described in Ridgwell et al., (2007). The difference is that here, SO_4^{2-} can be consumed even before dissolved oxygen can become fully depleted. Ambient temperature dictates the total fraction of POM that decays through both pathways per unit time and within each ocean depth layer (Crichton et al., 2021), with local [O_2] and [SO_4^{2-}] determining the fractional split between alternative pathways (Reinhard et al., 2020). For DOM, the decay constant determines the total fraction that decays per unit time. It should be noted that currently, there is no published nitrogen cycle in the cGENIE framework and hence we do not consider nitrate reduction as part of the redox cascade.

2.2 Marine iodine cycling in cGENIE

In the cGENIE model, iodine is present in three reservoirs: IO_3^- and I^- in the water column, and I^- incorporated in POM (and DOM). We then consider four processes that transfer iodine between these reservoirs: (1) IO_3^- reduction in the water column, I^- oxidation (also in the water column), photosynthetic IO_3^- uptake and intercellular reduction to I^-, and I^- release to seawater during the remineralization of POM (and DOM) (Fig. 1). As dissolved tracers, IO_3^- and I^- are physically transported and mixed through ocean circulation (as is I incorporated into dissolved organic matter), whereas iodine in POM settles through the water column. This is effectively the same framework as used by Lu et al., (2018). Here we re-assess this framework against an updated compilation of observed iodine speciation in the modern ocean and develop and test alternative representations of IO_3^- reduction ("threshold", "inhibition" and "reminSO4lifetime") and I^- re-oxidation ("lifetime", "Fennel", and "reminO2lifetime").

In the numerical scheme of Lu et al., (2018), when [O_2] falls below a set concentration threshold, IO_3^- is immediately and quantitatively reduced to I^- (thereafter, we term this iodate reduction parameterization "threshold").

The "inhibition" scheme links the IO_3^- reduction rate with the ambient O_2 concentration. Following the formulation for the rate of SO_4^{2-} reduction in Reinhard et al., (2020), we apply an oxygen inhibition term governed by a half-saturation constant. In devising this scheme, we note that while IO_3^- reduction rates have been determined

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experimentally, the quantitative relationship with $[O_2]$ (or other parameters) is unknown. The IO_3^- reduction under "inhibition" is mathematically described as:

$$d[IO_3^-]/dt = [IO_3^-] \times k_{red} \times \frac{k_{O_2}}{k_{O_2}+[O_2]} \quad (1)$$

in which k_{red} is the maximum first-order reduction rate of IO_3^-, and k_{O2} is the half-saturation constant of O_2.

Reduced sulfur (e.g. sulfides) is also suspected to play an important role in IO_3^- reduction in seawater, especially in the sulfidic zones (Jia-zhong and Whitfield, 1986; Luther and Campbell, 1991; Wong and Brewer, 1977; Truesdale et al., 2013). We therefore devise a scheme ("reminSO4lifetime") that scales a nominal 'lifetime' for IO_3^- with the rate of SO_4^{2-} reduction in the model. This has the effect of increasing the rate of IO_3^- reduction (a shorter lifetime) under conditions of higher sulphate reduction rates and hence lower ambient oxygen concentrations and/or higher rates of organic matter degradation:

$$d[IO_3^-]/dt = [IO_3^-] \times \frac{1}{\tau_{sul}} \times d[SO_4^{2-}]/dt \quad (2)$$

in which τ_{sul} defines the rate constant parameter linking the IO_3^- and SO_4^{2-} reduction, while the $d[SO_4]$ is amount of SO_4^{2-} reduced during each model timestep.

In Lu et al., (2018), I^- is oxidized to IO_3^- following first-order kinetics regardless of ambient O_2 (scheme "lifetime"). In this scheme, I^- oxidation follows the first-order reaction kinetics:

$$d[I^-]/dt = [I^-] \times \frac{1}{\tau} \quad (3)$$

where τ is the lifetime of I^- in seawater.

Given the potential link between I^- and nitrification, we devise an alternative "Fennel" scheme, in which I^- oxidation rates vary as a function of ambient O_2, increasing with ambient O_2 concentrations towards some hypothetical maximum value following Michaelis–Menten kinetics (Fennel et al., 2005). In Fennel et al., (2005), this parameterization was originally devised for ammonia reoxidation. The form of this response is defined by the maximum reaction rate and O_2 half-saturation constant (Fennel et al., 2005):

$$d[I^-]/dt = [I^-] \times k_{ox} \times \frac{[O_2]}{k_{fenn}+[O_2]} \quad (4)$$

in which k_{ox} defines the maximum rate constant of I^- oxidation, while k_{fenn} is the O_2 half-saturation constant.

Finally, in "reminO2lifetime", we associate I^- oxidation with O_2 consumption during remineralization. The logic behind this parameterization is the recent observation of I^- oxidation to IO_3^- catalyzed by bacteria, perhaps in association with ammonia oxidation (Hughes et al., 2021). Although the nitrogen cycle is not currently included in cGENIE, the NH_4^+ oxidation can be scaled to OM remineralization (Martin et al., 2019) and hence to O_2 consumption during remineralization. Under "reminO2lifetime", the lifetime of I^- oxidation is inversely linked to O_2 consumption so that faster remineralization—which in the ocean leads to more intensive NH_4^+ oxidation—enhances I^- oxidation. This I^- oxidation scheme follow this equation:

$$d[I^-]/dt = [I^-] \times \frac{1}{\tau_{O2}} \times d[O_2]/dt \quad (5)$$

where τ_{O2} is the rate constant parameter and $d[O_2]$ is the O_2 consumption during remineralization during a single timestep in the model.

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The final process in the marine iodine cycle framework concerns the processing of iodine directly through the biological pump. In this, IO_3^- is taken up by phytoplankton and incorporated into OM during photosynthesis (Elderfield and Truesdale, 1980) before being released as I^- during remineralization and/or cell senescence (Wong et al., 2002; Hepach et al., 2020; Wong et al., 1985). cGENIE simulates these processes as a function of a 'Redfield-ratio' of iodine to carbon (I:C ratio) in OM. We note that while I:C is tunable, it is fixed throughout the ocean. We discuss the merits of an optimized and uniform I:C compared to variable I:C (e.g., Waddley et al., 2020) in more detail in the discussion.

2.3 Model-data comparison

We used the model skill measure (M) (Watterson, 1996) to assess the performance of the marine iodine cycle in cGENIE. A major advantage of the M-score is that it is calculated through location-dependent comparison (Fig. 2; Lu et al., 2018). Another advantage of the M score is that it captures overall improvement of model performance relative to model-minus-observation maps, since it is non-dimensional, and the higher M stands for better performance. For comparison with simulated distributions of iodine speciation, we compiled oceanic iodine observation data from the literature (Fig. 2B; Table S4). This dataset includes the compiled dataset of Chance et al., (2019) and Sherwen et al., (2019), which was used to calibrate the Wadley et al., (2020) model, but includes more recent publications (referenced in Table S4) and is expanded to include the deep ocean and ODZ data. To avoid the influence of freshwater dilution and recycled iodine from the sedimentary flux, we applied a filter which only keeps the measurements with total iodine between 450nM and 550nM in the dataset. Note that the I^- measurements from the GP16 cruise in the ETSP are not included for the comparison because of potential method considerations (see Cutter et al., 2018 and Moriyasu et al., 2023). After these filtration methods, the data were re-gridded by taking the average values according to cGENIE grids. For each iodine speciation (hereby IO_3^- and I^-), a M score is calculated through comparing re-gridded observations versus model results in each corresponding grid. The synthesized M score for iodine of each model experiment is calculated through averaging those for both I^- and IO_3^-.

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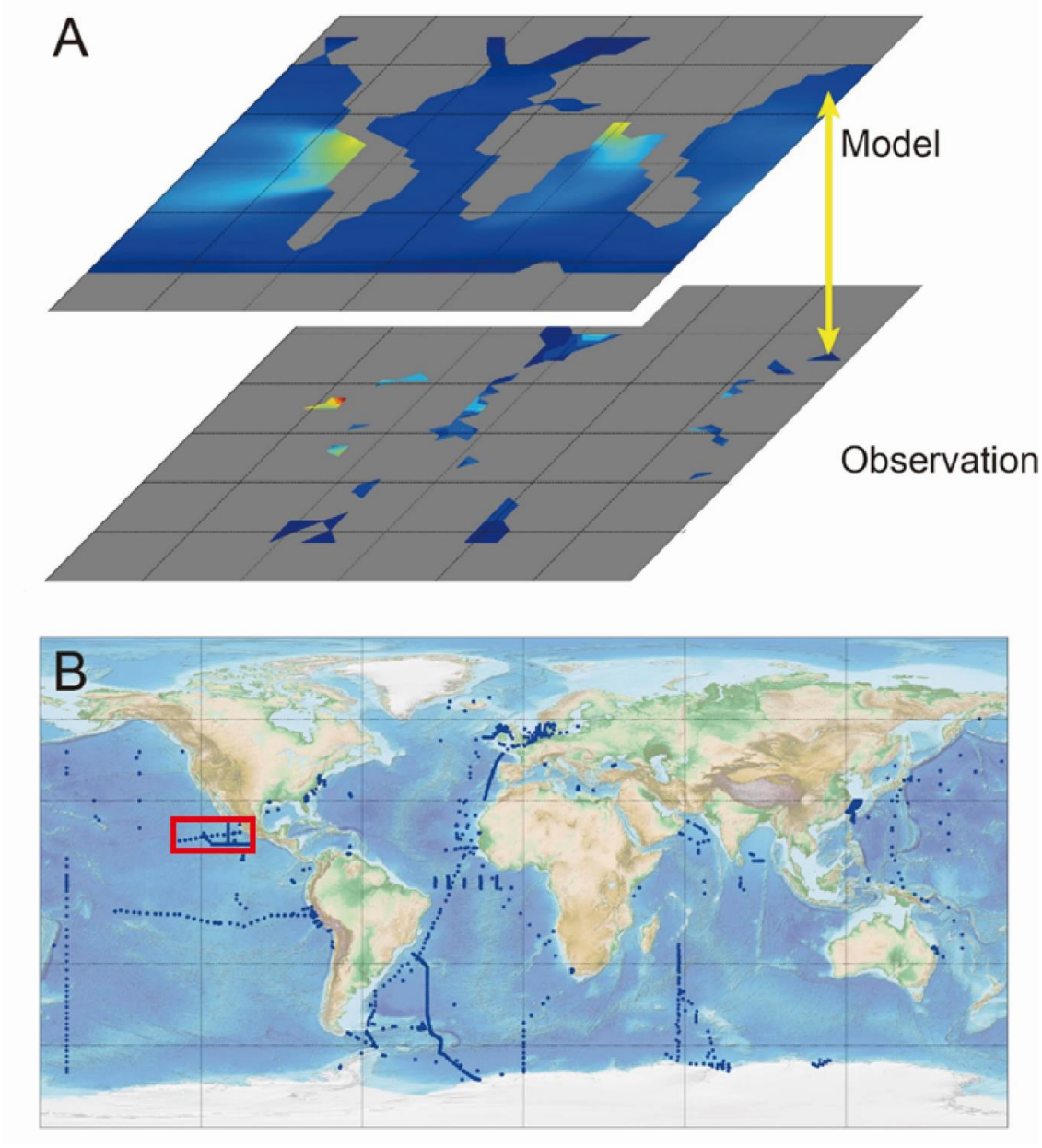

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221 **Figure 2: A). An example of location-dependent comparison between I- distributions in the cGENIE model**
222 **iodine data array and the regridded ocean observation. B). The sampling locations of iodine observation data**
223 **used for model-data comparison. Some coastal stations included in the figure are filtered out in the model-data**
224 **comparison. The ETNP transect associated with Fig.6 is labeled as red box.**

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2.4 Sensitivity analyses and model implementation

Because the relative roles of IO_3^- reduction, I^- oxidation, and IO_3^- planktonic uptake in the water column are uncertain, we calibrate the parameters controlling these processes in cGENIE by creating an ensemble of different parameter value combinations arranged in a 2D regularly-spaced grid and then repeat the same 2D parameter ensemble for different assumptions of I:C (Fig. 3). The output of each ensemble member is then statistically compared to our observational database. We assume the associated parameterization when the model reaches the best M score of replicating modern ocean iodine distribution would also be applied to simulate iodine cycling in the past. Each ensemble member was run for a total of 2,000 years and each starts from the same initial state, which was an experiment run for 10,000 years to equilibrium using a random set of iodine parameters within the ranges in Table 1. Running the models for 2000 years minimizes the CPU time but was also found to be more than sufficient to allow iodine inventories to equilibrate to new steady states. To explore whether the model simulated dissolved oxygen distribution imparted any particular bias to the tuned iodine cycle, we repeated the model ensembles, continually restoring the 3D pattern of $[O_2]$ in the model to that of the World Ocean Atlas 18 (WOA18) climatology (Garcia et al., 2018).

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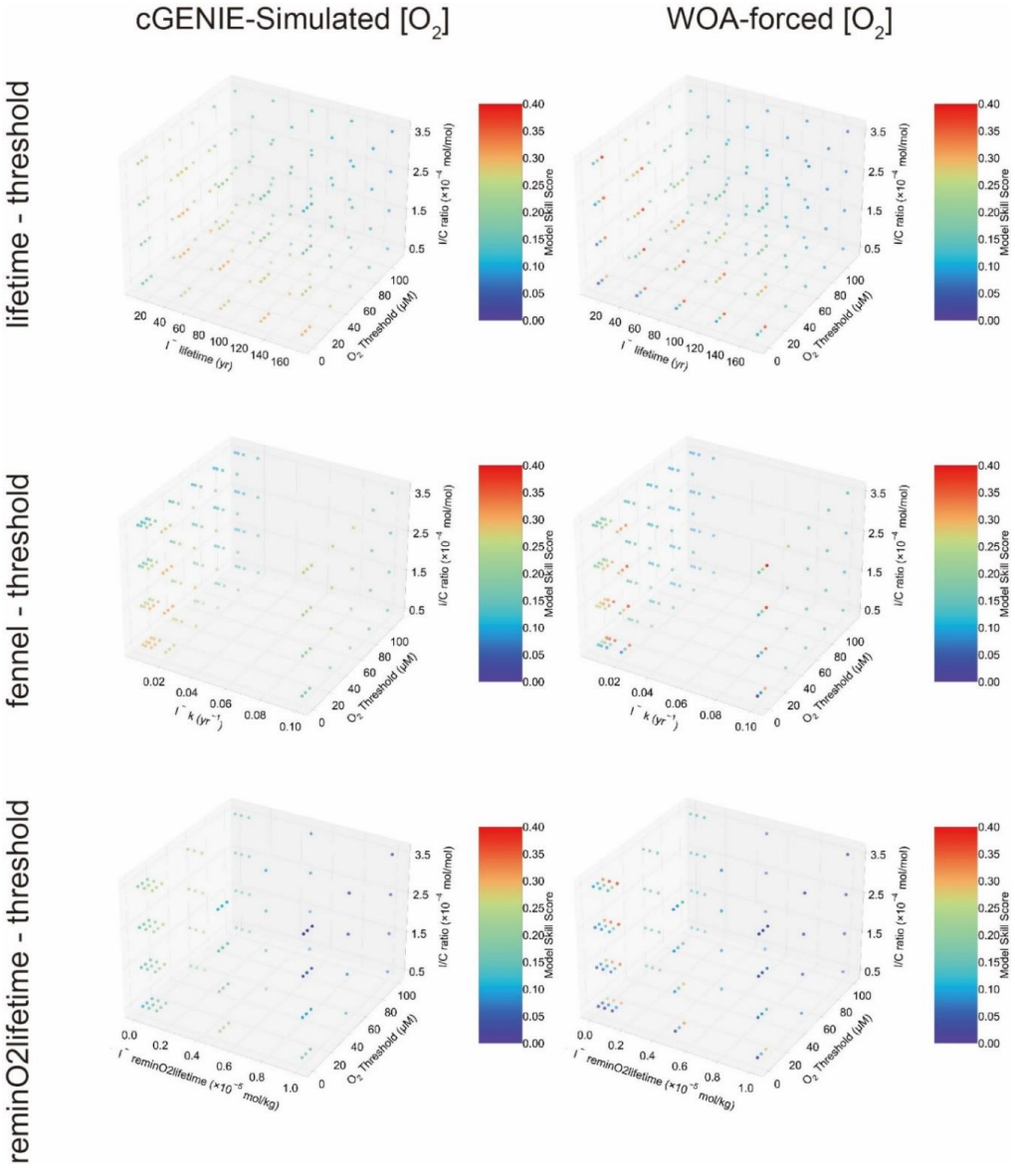

Figure 3: The three-dimensional model skill score array of the experiment ensembles.

We ran model ensembles to test five different combinations of iodine cycling parameterizations – "lifetime-threshold", "Fennel-threshold", and "reminO2lifetime-threshold", "lifetime-reminSO4lifetime", "lifetime-inhibition" (Table S1). While the results for all parameterization combinations are given in the supplement, we only focus on 3 parameterization-combinations here (Table 1) — "lifetime-threshold", "Fennel-threshold", and "reminO2lifetime-

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248 threshold". (A detailed justification and discussion for selecting these parameterization-combinations is included in
249 the Discussion section.)
250
251 **Table 1. The cGENIE iodine redox options and the associated range of parameters of these options.**
252 **The detailed introduction of each parameter is described in section 2.2.2 and the plausibility of these parameter**
253 **ranges is discussed in 4.1.1. Note that the oxidation rate constant *k* in 'Fennel' is in unit of year^{-1} in the model**
254 **configuration, which is the reciprocal of the 'lifetime'. A detailed table containing all considered**
255 **parameterization ranges can be found in Table S1.**

Parameter description		Iodine oxidation parameters			Iodine reduction parameters	I:C ratio (× 10^{-4} mol/mol)
		'lifetime' (years)	'reminO2lifetime' (× 10^{-5} mol/kg)	'Fennel' (Inhibition constant/ µM O_2)	'threshold' (µM O_2)	
Simulation 1	cGENIE O_2	10-170	\	\	1-110	0.5-3.5
	WOA	10-170	\	\	1-110	0.5-3.5
Simulation 2	cGENIE O_2 .	10-170 (1/k)	0.01-1	20	1-110	0.5-3.5
	WOA.	10-170 (1/k)	0.01-1	20	1-110	0.5-3.5
Simulation 3	cGENIE O_2	\	0.01-1	\	1-100	0.5-3.5
	WOA	\	0.01-1	\	1-100	0.5-3.5

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3. Results

In this section, we start by summarizing the overall statistical outcome of the tuning, then present a series of spatial analysis comparisons for each of the highest M-score ensemble members. The spatial analyses progressively reduce in scale, moving from global surface distributions, to global and basin-specific water column profiles, and finally to spatial comparisons for a specific ODZ region.

3.1. Model skill score

The M-score values achieved across the complete ensemble for each of the 3 main parameterization-combinations are illustrated in Fig. 3. The sensitivity test shows the M scores are sensitive to all three of the main parameters for the ensembles in Fig. 3. Higher model skill scores are usually reached when "threshold" is tuned to 10 μM [O_2] for all the ensembles, including both model-simulated [O_2] and WOA-forced [O_2]. For the ensembles, "lifetime-threshold" and "Fennel-threshold", the highest M scores are similar—0.305 and 0.308, respectively (Table 2). Both these ensembles have the highest performance when "threshold", "lifetime", and I:C ratio are tuned to 10 μM [O_2], 50 years, and 1.5 × 10^{-4} mol/mol, respectively, which is generally consistent with observations (Lu et al., 2016, 2020b; Tsunogai, 1971; Elderfield and Truesdale, 1980) (discussed in more detail later). The model performance of "reminO2lifetime-threshold" is less good than the other two combinations, with the best M score of 0.266 when "threshold", "reminO2lifetime", and I:C ratio are tuned to 10 μM O_2, 1 × 10^{-6} mol/kg, and 3.5 × 10^{-4} mol/mol, respectively (Table 2, Fig. 3). We note that for each parameterization-combination, the highest possible M score achievable by tuning improves when [O_2] is forced to that of the World Ocean Atlas 18 (WOA18) climatology (Garcia et al., 2018).

3.2. Meridional surface I^- distribution

Figure 4 shows a comparison between the observed latitudinal distribution of [I^-] at the surface and as simulated by the model for each parameterization-combination (for the respective best M-score ensemble member). Note that the observations (Section 2.3) are binned to the corresponding model grid cells and as such, reflect averages over the upper-most 80 m of the water column. This represents a reduction from 1338 to 141 surface ocean data points. We find that the surface ocean [I^-] in the model shows a trend of increasing values towards low latitudes, broadly consistent with observations (Chance et al., 2014) (Fig. 4). The "lifetime-threshold" and "Fennel-threshold" show similar latitudinal trends, but both overestimate the surface I^- in the mid-low latitudes in the southern hemisphere. The "reminO2lifetime-threshold" ensemble produces better estimation of meridional surface [I^-] trend, although overestimates [I^-] in the tropical surface ocean compared to the other two ensembles (Fig. 4).

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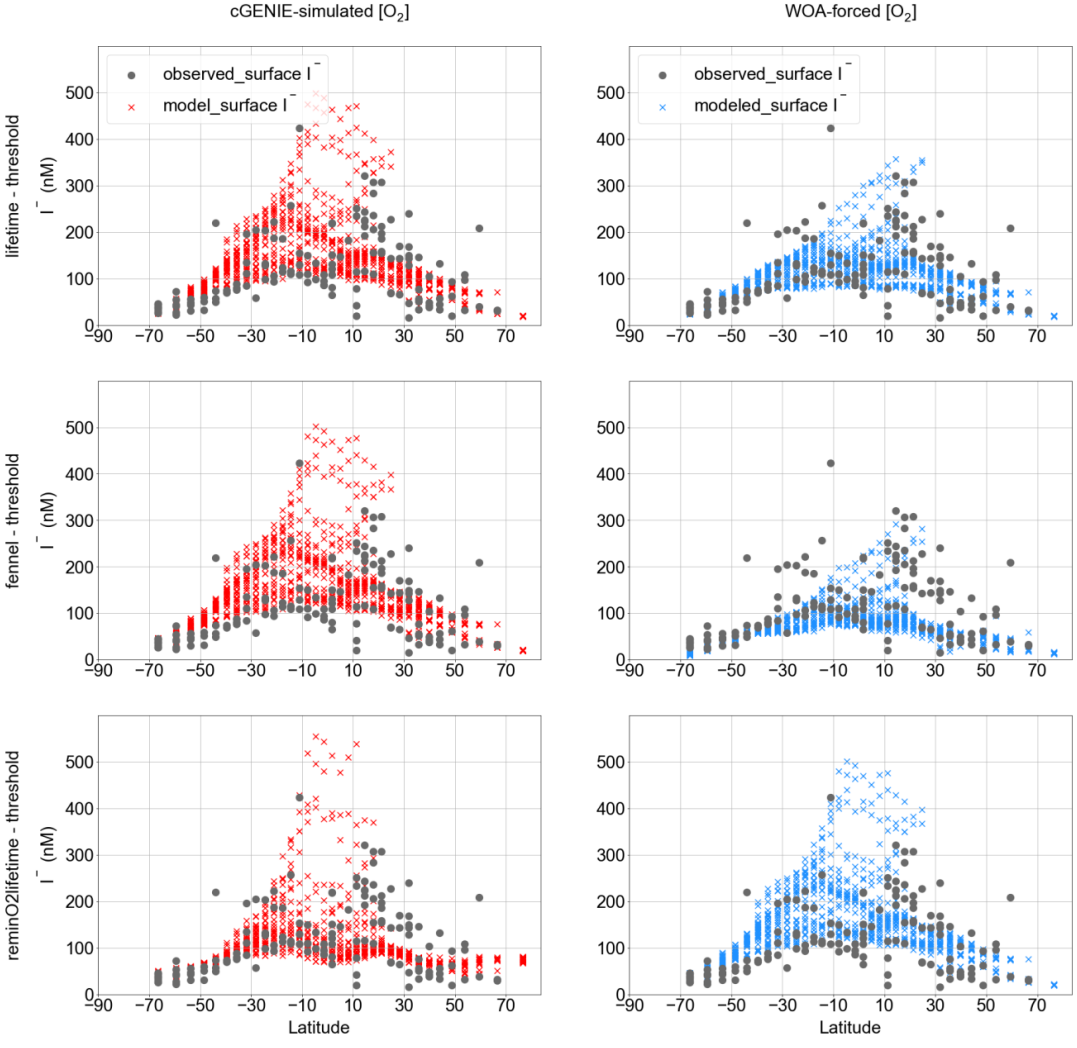

Figure 4: Modeled latitudinal surface iodide distribution compared with observation with the cGENIE simulated [O_2] and the [O_2] restoring forcing. The elevated [I^-] observed and modeled in low latitudes is the result of phytoplankton reduction in the surface ocean. Note that the I^- distribution simulated by "lifetime-threshold" and "fennel-threshold" are close but not identical.

3.3. Global and basin-specific iodine depth distributions

Comparisons between the observed distributions of I^- and IO_3^- seawater concentrations among the global ocean and the Atlantic and Pacific Oceans are presented in Fig. 5. Again, we re-gridded the iodine observations (see: Section 2.4) and selected sub-sets of the data that lay in either Atlantic or Pacific basins, contrasting with the corresponding model values at those locations. We find only relatively minor differences between the best M-score ensemble member of

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each of all three parameterization-combinations, and all show increased [IO_3^-] and decreased [I^-] with increased depth below the photic zone in the Atlantic and Pacific basins, as well as globally (Fig. 5). The modeled depth profile broadly matches with observation in the Atlantic and deep Pacific Ocean, except the underestimated subsurface peak of [I^-] observed in the Pacific and overestimated [IO_3^-] in the deep Pacific (Fig. 5). This mismatch of subsurface I^- peak is probably the result of sampling bias, with most of the Pacific iodine observations from ODZs in the Eastern Tropical North Pacific (ETNP) and the Eastern Tropical South Pacific (ETSP). For example, in model depth profiles masked to only include grid points with corresponding observations, the modelled Pacific depth profiles show a clear mid-depth ODZ feature (Fig. S8).

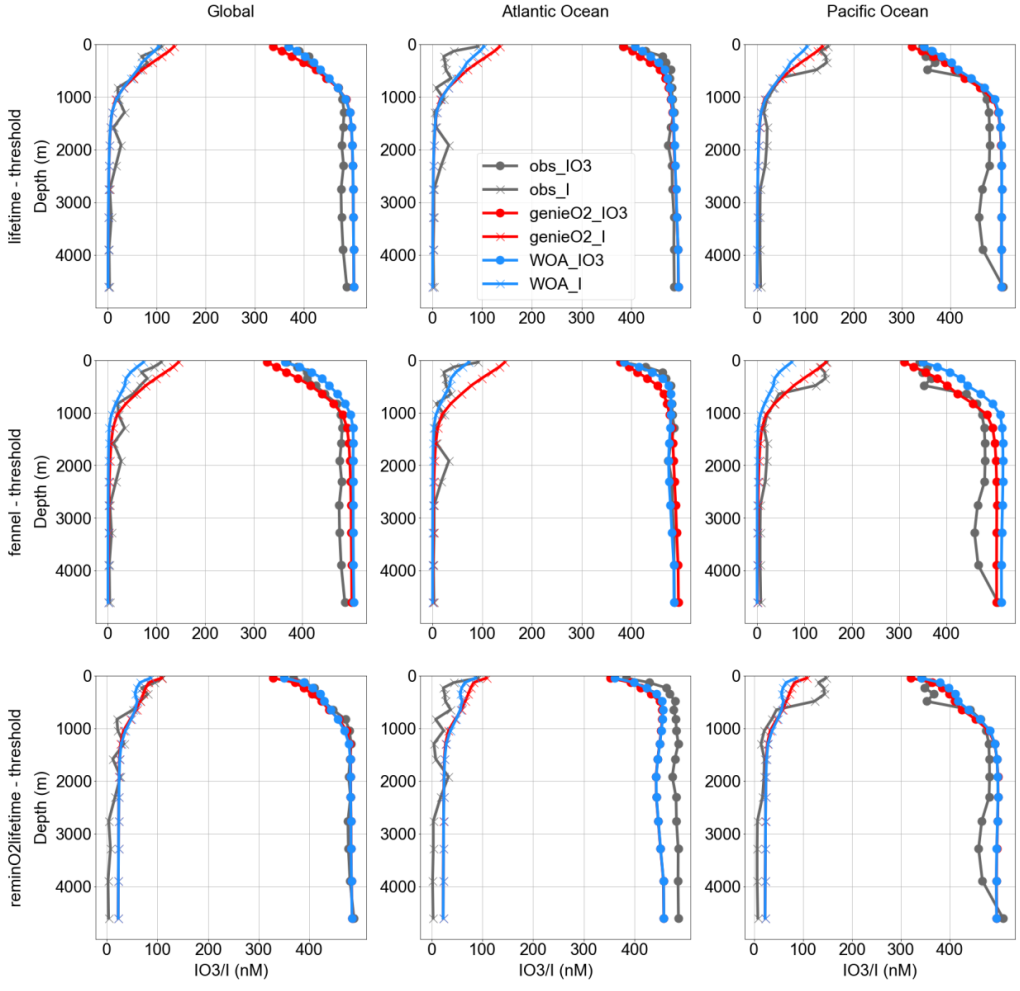

Figure 5: Modeled averaged iodine (including iodate and iodide) depth profile among global ocean, the Pacific, the Atlantic compared with observation. The surface I^- enrichment among the ocean basins is caused by phytoplankton reduction. The subsurface (~500m) I^- enrichment is the result of sampling bias that most of the observations are from the ETNP and ETSP ODZs.

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3.4. Iodine distribution within ODZs

To assess the model ability to simulate iodine cycling in marine low oxygen environments, we compared distributions of oxygen and iodine species in the ETNP (Fig. 6). The O_2 transects amongst all model simulations are the same because we only changed the parameterizations of the iodine cycle between ensembles and ensemble members (i.e., they all simulate the same biological pump in the ocean). Notably, compared to [O_2] measured in the ETNP transect, the model underestimates the extent of the ODZ. Severe deoxygenation below 50µM [O_2] was observed in relatively shallow depths between 100-200m in the ETNP, and this ODZ extends for more than 3000 km towards off-shore from Mexican coast (Fig. 6). Although cGENIE simulates the O_2-deficient pattern in the ETNP, the extent of the ODZ is underestimated. The simulated oxycline is ~200m deeper than the observation and the [O_2] variation is gradual. The ODZ below 20µM [O_2] in the model is limited to a small spatial extent within 1000km offshore, which is much smaller than that in the observation (Fig. 6).

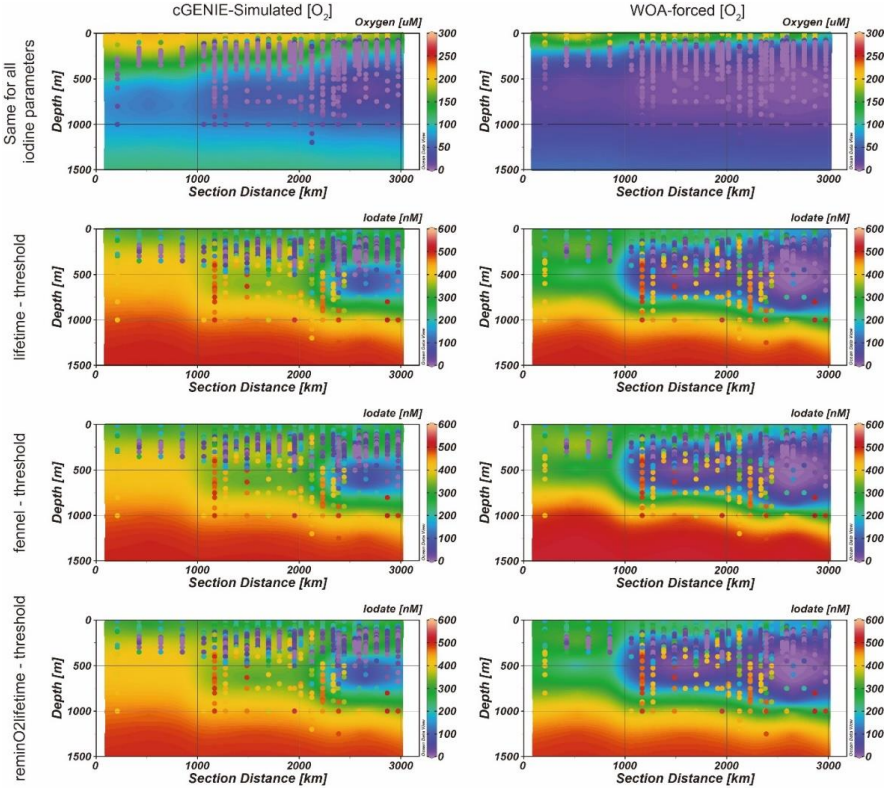

Figure 6: Modeled (contour) and observed (colored dots) west-to-east transect of IO_3^- and O_2 in the ETNP. Note that the WOA-forced [O_2] models simulate a larger extent of IO_3^- anomaly, which better matches the observation. The left-hand panel contours are model results based on cGENIE-simulated [O_2] while contours on the right are model results from WOA-forced [O_2].

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All the three chosen best-performance-experiments show similar iodine anomalies (IO_3^- depletion) in the ETNP, fitting the general feature of the observation. Other parameterizations did not replicate the ODZ (Fig. S4). However, even under the "best-fitting" parameters, compare to the observations, the ODZ feature in the model is underestimated both in intensity and in areal extent compared to the observations (Fig. 6). The observed IO_3^- depletion zone ($[IO_3^-]$ <100nM) occurs in shallower depths between 100-500m, matching the shallow ODZ and spans ~2000km offshore; however, the modeled IO_3^- depletion zones in the ETNP are located in 400-700m, and only extends within 1000km from the shore.

We also ran model ensembles forcing cGENIE to restore the modern ocean $[O_2]$ annual average climatology to that of the WOA18 (Garcia et al., 2018) (Fig. 6). Now the subsurface IO_3^- depletion zone in the ETNP ODZ in all three ensembles extends ~2000km offshore and spans across 100-1000m in depth (Fig. 6). The shallow and extended ODZ iodine distribution in the ETNP better matches the observation compared to non-O_2 restoration simulations.

Table 2. The performance of the cGENIE iodine simulations and associated parameterization when the model reaches the best global M score. Note that the oxidation rate constant *k* in 'Fennel' is in unit of year^{-1} in the model configuration, which is also the reciprocal of the 'lifetime'. The full model performance is summarized in Table S2. Note that the lifetime in 'Fennel' is parameterized as k (year^{-1}) = 1/lifetime.

Parameter description		Iodine oxidation parameters			Iodine reduction parameters	I:C ratio ($\times 10^{-4}$ mol/mol)	Model skill score
		'lifetime' (years)	'reminO2lifetime' ($\times 10^{-5}$ mol/kg)	'Fennel' (Inhibition constant/ μM O_2)	'threshold' (μM O_2)		
Simulation 1	cGENIE O_2	50	\	\	10	1.5	0.305
	WOA	50	\	\	10	1.5	0.385
Simulation 2	cGENIE O_2	50 (1/k)	\	20	10	1.5	0.308
	WOA	10 (1/k)	\	20	10	3.5	0.385
Simulation 3	cGENIE O_2	\	0.1	\	10	3.5	0.266
	WOA	\	0.1	\	10	3.5	0.365

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4. Discussion

In summary: we ran ensembles for five combinations of iodine cycling parameters (summarized in Table S2) in the cGENIE Earth system model (both with internally calculated and WOA-imposed [O_2] distribution) and presented the results for 3 of them that showed the best performance. We analyzed the performance of the ensembles via: (1) M score for the model-data match of both [I^-] and [IO_3^-] across the entire ocean, (2) qualitative model-observation comparison of latitudinal surface ocean distributions of [I^-], (3) averaged depth profiles in global and individual ocean basins for both [I^-] and [IO_3^-], (4) iodine transects across the across the Eastern Tropical North Pacific (ETNP) ODZ, and (5) M score for model and I:Ca observations for pre-OAE2.

4.1. Overall model skill score comparison

Two broad observations emerge from the M score comparison. First, the 1st-order kinetic iodine oxidation associated ensembles ("lifetime-threshold" and "Fennel-threshold") have the highest M scores (Table 2), under both cGENIE-simulated [O_2] and WOA-forced [O_2]. This is consistent with previous observations of 1st-order kinetics for I^- oxidation (Tsunogai, 1971). Second, the simulations with WOA-forced [O_2] produce significantly higher M scores than that of the cGENIE-simulated [O_2] field (at least ~0.8 of improvement; Table 2). Despite a 1st-order non-O_2 dependent oxidation mechanism providing the highest M scores, the WOA vs internally model-generated dissolved oxygen distributions comparison highlights the O_2 and related redox dependency of the iodine cycle from the perspective of IO_3^- reduction. Each of these factors are discussed in the following Section 4.1.1.

4.1.1. Parameter value plausibility

A credible representation of the marine iodine cycle requires not only that observations can be replicated, but that replication occurs when parameter values fall within real-world ranges. In this section, we discuss the validity of our best-fit (maximized M-score) parameter values. For the iodine cycle, these parameters include O_2 threshold, I:C ratio, and I^- oxidation rate.

Our model M score is highest with an [O_2] reduction threshold of 10µM (Fig. 3 and Table 2). Although it is generally well accepted that IO_3^- is reduced in low oxygen settings (Luther, 1991; Rue et al., 1997; Wong et al., 1985; Wong and Brewer, 1977; Farrenkopf and Luther, 2002), the degree of O_2 depletion that triggers IO_3^- reduction is still unclear. A relative high threshold for triggering IO_3^- reduction is proposed based on comparison between benthic foraminiferal I:Ca and ambient [O_2] (20-70 µM O_2; Lu et al., 2016, 2020a). These [O_2] thresholds are similar to that determined in a previous cGENIE-based iodine cycle study (30 µM) (Lu et al., 2018), but it is difficult to directly compare this to our results because of differences in the model representation of the ocean biological pump, the iodine observational data-set, and model-data comparison methods utilized.

Many of the studies suggesting a relatively high [O_2] threshold is based on evaluations of [IO_3^-]-[O_2] within the upper chemocline of ODZs; however, evaluation of [O_2]-[IO_3^-] from ODZs throughout the entire water column suggest the potential for IO_3^- persistence within the low oxygen cores of ODZs. Specifically, IO_3^- accumulation is observed within the ETNP and ETSP at depths where [O_2] is close to or below the detection limit of the sensors which is reported near 1 µM (Hardisty et al., 2021). In addition, it is worth noting that the kinetics of IO_3^- reduction is heterogeneous both within and between ODZs. For example, a transect evaluating IO_3^- reduction rates in the ETNP

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observed rapid rates in the upper oxycline, where [O_2] was near ~12 µM, but the potential for sluggish rates in the ODZ cores, where [O_2] was below detection. In an early study of the Arabian Sea, IO_3^- was reduced rapidly within the ODZ core. Together, these suggest IO_3^- reduction may be controlled by factors beyond O_2 (Hardisty et al., 2021; Farrenkopf and Luther, 2002). For example, IO_3^- is likely formed in high [O_2], non-ODZ water masses but can be sustained upon transport or mixing within oligotrophic, offshore ODZ regions where organic supplies are more limited (Hardisty et al., 2021). A comparison to the N cycle would also indicate a low [O_2] threshold—for example, denitrification has a sub-µM [O_2] threshold and has a similar redox potential with IO_3^- reduction (Dalsgaard et al., 2014; Thamdrup et al., 2012). A sub-µM [O_2] threshold for IO_3^- reduction could explain the [IO_3^-] variations observed in ODZ cores with [O_2] below the µM detection limits of sensors; however, iodine speciation has yet to be analyzed alongside [O_2] measurements via sensors with sub-µM detection limits, such as STOX sensors. Regardless, our 10 µM [O_2] threshold based on maximizing the M score reflects a global average value and clearly falls within the ranges of oceanographic observations.

For both our study and that of Lu et al., (2018), an I^- lifetime of 50 years maximizes model performance. However, Lu et al., (2018) chose to implement a slightly lower value of 40 years for their paleo-application because it reflected the slowest rate observed in the literature at that time (Tsunogai, 1971). Notably, though IO_3^- is the most abundant marine iodine species, its production from I^- has never been observed under normal marine conditions. This has acted as a major hinderance on providing direct constraints. That said, our cGENIE estimate is consistent with a multitude of other constraints that indicate that I^- oxidation to IO_3^- undergoes extremely slow kinetics. The I^- oxidation rates calculated through indirect methods including mass balance and seasonal iodine speciation changes (Tsunogai, 1971; Campos et al., 1996; Truesdale et al., 2001; He et al., 2013; Edwards and Truesdale, 1997; Žic et al., 2013; Moriyasu et al., 2023) or through radiogenic tracer spiked incubations (Hardisty et al., 2020; Schnur et al., 2024; Ştreangă et al., 2024) have a wide range of variation from 1.5nM/yr to 670nM/yr. The lifetime in cGENIE is 50 years, which can be approximately converted to the zeroth order rate of <9 nM/yr, falling in the lower end of the previous studies.

Our best-fit I:C ratio is 1.5×10^{-4}, and this value is in agreement with plankton measurements and mass-balance calculations (Chance et al., 2010; Elderfield and Truesdale, 1980). In the photic zone, IO_3^- is taken up by phytoplankton and incorporated into their cellular structures followed by subsequent conversion into I^- (Hepach et al., 2020). Due to this, it is assumed that IO_3^- removal in the surface layer of the ocean is a function of organic carbon fixation during primary productivity according to Redfield-like ratios (Campos et al., 1996; Chance et al., 2010). Of the parameters incorporated into the model, in theory, I:C should probably be the best constrained. However, published I:C ratios based on field and laboratory measurements vary over several orders of magnitude (10^{-5} to 10^{-3}) (Elderfield and Truesdale, 1980; Campos et al., 1996; Hepach et al., 2020; Chance et al., 2010). To limit the number of model simulations and size of the ensembles, we only varied I:C between 5×10^{-5} and 3.5×10^{-4}, increasing by 1.0×10^{-4}, which covers the range indicated by previous studies (Elderfield and Truesdale, 1980).

It is unlikely that the I:C value is constant across the global ocean due to differences in phytoplankton compositions and other factors. In cGENIE, most of the elevated surface [I^-] over 200nM is present in the ETSP and the northern Indian Ocean, representing the effect of high primary productivity and transformation of IO_3^- to I^- via the rapid

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recycling DOM 'loss term' in the representation of biological export (Fig. S6). The mismatch between the model and the observation probably hints that the I:C ratio is not constant in the ocean, as which is also hypothesized by Wadley et al., (2020), although in the absence of an explicit representation of primary productivity in the model and lack of a spatially variable f-ratio (Laws et al., 2000)(implicitly, the f-ratio is 0.33 everywhere in cGENIE). In testing a fixed, spatially uniform I:C, Wadley et al. (2020) underestimated surface [I^-] in low latitudes and overestimated concentrations in mid-latitudes. Based on their model-observation comparison, they hypothesized that the I:C ratio decreases systematically with sea surface temperature (SST) (Wadley et al., 2020). Until more constraints are developed on spatial variability and associated driving factors for I:C, a generalized approach of a globally uniform I:C remains the most parsimonious assumption, especially considering cGENIE's intended extrapolation to ancient settings.

4.1.2. Comparing alternative iodide oxidation parameterizations in cGENIE

Due to similarity in redox potentials, the iodine cycling in the ocean has been hypothesized to be linked to the cycling of nitrogen (Rue et al., 1997). Nitrification promoting I^- oxidation to IO_3^- has been hypothesized in field studies (Truesdale et al., 2001; Žic et al., 2013), and more recently has been linked via observation of I^- oxidation to IO_3^- by ammonia oxidizing bacteria in laboratory environments (Hughes et al., 2021). We further note that Wadley et al., (2020) explicitly link I^- oxidation to NH_4^+ oxidation in their surface ocean iodine cycle model.

As an alternative to the first-order lifetime oxidation parameterization used here and in Lu et al., (2018) and in the current absence of a full nitrogen cycle (and hence explicit NH_4^+ oxidation) in cGENIE, we also applied a strategy ("reminO2lifetime") which links I^- oxidation to organic carbon remineralization and the consumption rate of dissolved oxygen. The reasoning behind this is that the O_2 consumption rate in the model implicitly reflects bacterial oxidizing activity in the water column and hence the potential for I^- to be oxidized to IO_3^-.

We find that the overall model performance involved with the "reminO2lifetime" is lower than other experiments where I^- oxidation is ubiquitously oxidized according to a parametrized lifetime, or "lifetime-threshold" (maximum M score 0.266 vs. 0.305/0.308 under cGENIE simulated [O_2]) (Fig. 3 and Table 2). However, despite slightly lower M scores, the "reminO2lifetime" generally replicates the latitudinal surface [I^-] trend, the depth profiles in the ocean basins, and the ODZ transect (Figs. 4-6).

Under the default "lifetime" scheme, I^- will oxidize in the whole ocean regardless of the concentration (or even presence/absence) of ambient O_2. This scenario might hence not perform well in replicating the ocean iodine cycling at intervals during the Phanerozoic when ODZs were highly expanded, or particularly during the Precambrian when the majority of the ocean was ferruginous or euxinic and highly depleted in O_2 (for example, summarized by Lyons et al., 2014). Ideally, for application to paleoceanographic studies, an [O_2]-related I^- oxidation alternate scheme is required. Although thermodynamics theory suggests O_2 does not directly drive I^- oxidation (Luther et al., 1995), field studies in ODZs indeed observed low [O_2] inhibits this process (Farrenkopf and Luther, 2002; Moriyasu et al., 2020). We hence developed and tested variable I^- oxidation kinetics, with the ambient dissolved O_2 providing an inhibition of the rate of oxidation based on the enzymatic nitrification scheme of Fennel et al., (2005).

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Since most of the ocean is fully oxygenated today, the difference of M scores between "lifetime" and "Fennel" oxidation models are minor (0.305 vs. 0.308, Table 2). The parameters associated with the highest M score between two oxidation options are also very close to each other, except "Fennel" oxidation together with the model WOA-forcing, has higher I:C ratio (3.5×10^{-4}) and faster saturated I^- oxidation kinetics (0.1 yr^{-1} vs. 0.02 yr^{-1} in other ensembles). The parameter differences between the "Fennel"-WOA ensemble and other models make sense because the faster oxidation rate compensates the increased I^- production through the higher biotic uptake rate. The pre-OAE2 simulations are particularly illustrative of this tradeoff and are discussed in more detail in section 4.3.

In summary, all the three parameter combinations produce high and comparable M scores and similar parameters (oxidation, reduction, and I:C) associated with these M scores (Table 2). Although direct field-based evidence to evaluate the controlling parameters of "reminO2lifetime" is absent, the parameters controlling the other model scheme are consistent with previous studies.

4.2. Beyond the M-score: model-data comparison across iodine gradients

As applied here, the M score provides a quantitative measure that describes the overall model global performance and allows us to directly compare the implications of parameter value and parameterization choices. However, the M score overlooks regional gradients that may be important for both paleo- and modern oceanographic research. Indeed, amongst all the various ensembles we ran as part of this study (Table S1), only "lifetime-threshold", "lifetime-Fennel", and "reminO2lifetime-threshold" performed well in replicating the modern oceanic iodine gradients (Figs. S2-S4) and informed our decision to focus on these 3 parameterization-combinations here. We now discuss this in more detail below.

4.2.1. Meridional surface [I^-] gradient

All the parameterization-combinations summarized in Table 2, as well as the observations, show enrichment of I^- in the surface ocean at low latitudes (Fig. 4). The pathway of transforming IO_3^- into I^- in these oxidized waters is through primary productivity in the photic zone, which is temperature dependent (Chance et al., 2014). A recent north-south transect showed the highest surface I^- enrichment in the oligotrophic, permanently stratified tropical stations (Moriyasu et al., 2023). Therefore, the IO_3^- flux from deep waters through seasonal mixing may be an important balance to *in-situ* IO_3^- reduction rate by primary producers in the high latitudes (Chance et al., 2014; Moriyasu et al., 2023). The cGENIE model generates the general pattern of latitudinal surface I^- distribution pattern; however, overestimation especially in low latitudes may exist, especially in the tropical surface where [I^-] are close to 500nM among all the cGENIE-O_2 models (Fig. 4).

The cGENIE (internally generated oxygen distributions) vs WOA (imposed distributions) O_2 comparison provides evidence that I^- generated in low [O_2] settings may broadly enhance [I^-] in oxygenated photic waters, with lower and closer-to-observations [I^-] values in the WOA tunings (Fig. 4). This includes "lifetime-threshold", where O_2 only impacts the reductive portion of the iodine cycle, but also the "Fennel" and "reminO2lifetime" where rates of I^- oxidation is also [O_2] dependent. More specifically, most of the elevated (over 200nM) surface [I^-] in cGENIE, occurs in the ETSP and the northern Indian Ocean and corresponds to locations of high primary productivity (Fig. S6).

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Since the surface ocean [O_2] in the model is never below 200µM, IO_3^- reduction at the ocean surface is unlikely. Instead, *ex situ* transport from proximal regions of subsurface anoxia is the most probable source of elevated I^-. Indeed, the most prominent regions of I^- enrichment in the model occur near the Peruvian coast and in the Arabian Sea, where ODZs lie below the surface (Fig. S6). More detailed data-model comparison among these two areas is limited because the observation data are few (e.g., Farrenkopf and Luther, 2002 and Rapp et al., 2020). In contrast, the meridional trend of I^- in the surface Atlantic Ocean, where ODZs are less developed, exhibits better agreement with both the observation and the Wadley et al., (2020) model (Fig. S6). The overestimation of tropical ocean surface [I^-] by cGENIE is hence likely to be associated with deficiencies in the simulation of ODZ oxygenation.

Importantly, modeled overestimations in surface ocean [I^-] may be difficult to confirm given current observational data densities. Specifically, a comparison of observational data to model-latitudinal trends masked to only include grid points with corresponding observations show the same broad trend of increasing [I^-] in the low latitude but with fewer so called "overestimations" (Fig. S7). More observations in surface ocean [I^-] from low latitudes is required to better assess the validity of elevated modelled surface ocean [I^-] in some regions.

4.2.2. Global and basin-specific iodine depth profiles

All the models, both cGENIE-[O_2] and WOA-forced, generate a decrease in [I^-] and increase in [IO_3^-] from the euphotic zone down to the deep abyssal zone across ocean basins, matching the primary-production-driven pattern (Fig. 5). As discussed in the previous section, this surface maxima of [I^-] in the oxygenated water column is the result of biologically mediated reduction or release during cell senescence of phytoplankton. Below the photic zone, [I^-] is close to zero and IO_3^- becomes the dominant species. The deep ocean is mostly oxygenated and has longer water residence times (several millennia, Matsumoto, (2007)) compared to the I^- lifetime (<40 years, Tsunogai, (1971)), thus facilitating I^- oxidation in the absence of IO_3^- reduction in ODZs.

We note that there are multiple general discrepancies between observations and data as well as differences between WOA and cGENIE-[O_2]. In general, all models reproduce the global average better, relative to the basin-specific profiles. We suggest that the global averaged profiles are a better test of the cGENIE simulations because of sampling biases associated with individual basins. For example, the discrepancy between the model and the observation is prominent in the Pacific (Fig. 5). The observed Pacific subsurface [I^-] peak mirrors the [IO_3^-] minima that occurs at depths where ODZs are present. This ODZ feature in the averaged Pacific observation profile is likely the result of sampling bias since most of the observations from the Pacific were from the ETNP (Rue et al., 1997; Moriyasu et al., 2020), thus not reflecting the overall iodine distribution in the Pacific Ocean (Fig. S8). As for the meridional trends, sampling bias is again demonstrated in a depth profiles masked to only include modeled grid points with corresponding observation data, with a clear mid-depth ODZ feature in the modeled Pacific depth profiles (Fig. S8). A similar example is from the Indian Ocean, which we do not show, since most iodine subsurface data come from the ODZ, not normal marine, localities. All this said, while the general features of iodine speciation with depth are generally similar, our data compilation indicates the potential for some basin-specific variations which require further research to validate and mechanistically understand.

4.2.3. Iodine distribution within ODZs

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One of the major goals of calibrating the iodine cycle in cGENIE is to simulate the iodine behavior associated with ancient low oxygen settings. To assess this potential, we analyzed model performance for the ETNP (Rue et al., 1997; Moriyasu et al., 2020) where observation data are abundant and the areal extent of the ODZ is sufficiently large to be reflected in the model grid (Fig. 6). Importantly, it should still be noted that cGENIE is best applied to understanding broad scale processes and thus the scope of the ETNP ODZ transect comparison is likely too fine resolution to expect a good match. That said, the simulated reduction in IO_3^- to I^- generally overlaps with the extent of the ODZ (Fig. 6), which provides support for the use of cGENIE to understand the broad distribution of ancient [IO_3^-] and [O_2]. Other non-threshold model parameterization-combinations (shown in Fig. S4) did not replicate the ODZ feature in iodine speciation.

Across all model configurations assuming cGENIE-[O_2], the prominent discrepancy is an underestimation of the spatial extent and intensity of the IO_3^- depletion zone in the ETNP (Fig. 6) because simulated subsurface O_2 deficient area is notably narrower than that compared to WOA climatology (Fig. 6, Fig. S5). There are multiple factors that might affect the performance of simulating the O_2 cycle in cGENIE. The model might underestimate the intensity of gyres in the North and Tropical Pacific, resulting in the narrowed ODZs in these areas. Also, the pattern of upwelled nutrients into the surficial Tropical Pacific needs to be tuned to better replicate the productivity and the O_2 consumption during remineralization. Another source of uncertainty is that the short-term processes, such as seasonal or El Niño driven ODZ variations at the time of sample collection are not replicated in the cGENIE simulations.

Other data-model misfits might have arisen as a result of additional IO_3^--reduction dependencies not explicitly accounted for in the model. As discussed above, shipboard incubation and radiogenic-tracer-spiked rate calculation suggest that IO_3^- reduction is slow in the offshore ETNP ODZ (Hardisty et al., 2021). This could explain why measurable IO_3^- is present in the core of the ETNP ODZ (Fig. 6). This is further exacerbated in the Eastern Tropical South Pacific ODZ, where [IO_3^-] remains above 250 nM in some studies (Cutter et al., 2018) but near detection limits in in others (Rapp et al., 2020), suggesting extreme spatiotemporal variability related to currently unconstrained factors. Further, while the capability of microbes using IO_3^- as an electron acceptor for oxidizing organic matter has been confirmed in laboratory culture experiments (Councell et al., 1997; Reyes-Umana et al., 2021; Yamazaki et al., 2020; Amachi et al., 2007; Farrenkopf et al., 1997), no study to date has demonstrated non-O_2 dependent controls driving variable IO_3^- reduction rates.

An important factor contributing to elevated [I^-] in ODZs is benthic fluxes and reduction occurring within the uppermost layers of marine sediments (akin to denitrification). To help account for this in our M score and model calibration (see: methods section), excess iodine was filtered from our observational dataset. The excess iodine originated from the sediment flux has been observed in ODZ water columns contacting anoxic sediments (Chapman, 1983; Farrenkopf and Luther, 2002; Cutter et al., 2018; Moriyasu et al., 2020). We note that excess iodine is explicitly as I^-, reflecting the limited or lack of oxygen within the ODZ, and is a local-regional phenomenon not yet observed to persist beyond ODZ settings. As such, since our goal is a paleoceanography-focused model, cGENIE does not incorporate the benthic flux of I^- because only IO_3^- is tracked via the I:Ca paleoredox proxy.

4.3. Applicability of the cGENIE marine iodine cycle to paleo redox reconstruction

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Parameter tuning, and the ability to reproduce modern observations, does not by itself offer any guarantee that spatial patterns are being simulated for the 'correct' reason (i.e., specific set and relative importance of mechanisms). This is even more pertinent in the context of the application of a modern-tuned model to paleo redox reconstruction. To test whether our new iodine cycle had predictive power in the geological past, we carried out a deep-time plausibility test.

For the paleo plausibility test, we adopted the Cretaceous, pre-OAE2 (ca. 93 Ma) configuration (continental arrangement and ocean bathymetry, wind stress and velocity, and zonal average planetary albedo boundary conditions) from Monteiro et al., (2012). We choose this particular geological interval because the controls on ocean redox have been previously evaluated using the GENIE model (Monteiro et al., 2012; Hülse et al., 2019), the oceanic conditions are much more extensively dysoxic and anoxic than present-day and hence represent a relatively severe test of the model iodine cycle, and a number of I:Ca proxy measurements are available (Zhou et al., 2015). In order to evaluate the same configuration of the iodine cycle as optimized in this study, we also substituted the temperature-independent representation of biological export production and fixed remineralization profile of POM in the water column (i.e., Ridgwell et al., (2007)) for the temperature-dependent scheme of Crichton et al., (2021). However, in substituting the biological pump scheme in the model we change the ocean redox landscape compared with e.g., Monteiro et al., (2012). We therefore test a range of different assumptions regarding the ocean PO_4 inventory at the time as a means of generating a range of different plausible states of ocean oxygenation. In this, we test: 0.2, 0.4, 0.6, 0.8, 1.0, and 1.5 times the mean modern concentration (2.15 mM). We run the model with each of the best-fit (highest M-score) sets of parameter values associated with the 5 different parameterization-combinations (but focus on the results of the same 3 combinations we did for comparison against modern), and for each of the varying PO_4 inventory assumptions, for 10,000 years to steady-state.

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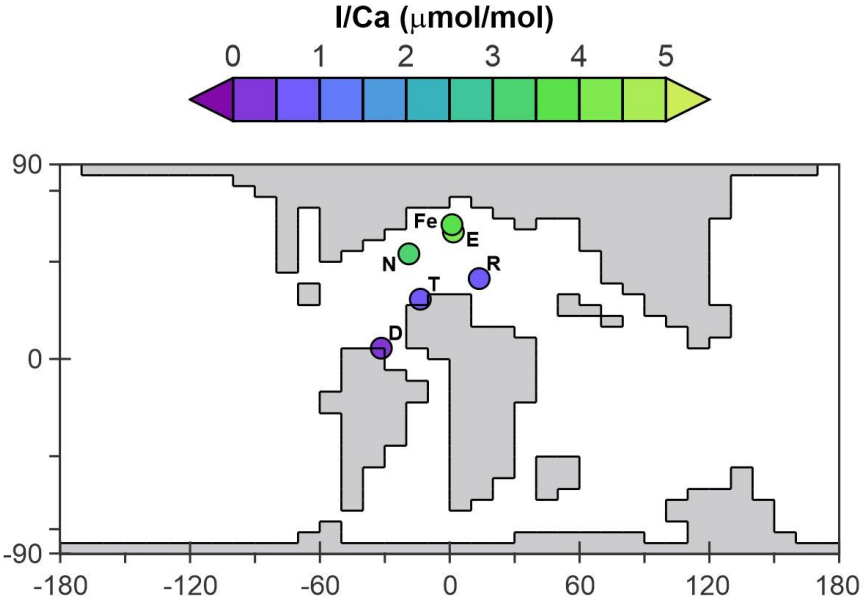

Figure 7: The continental setting during the Cretaceous OAE2 (Cenomanian - Turonian) in cGENIE. The colored dots represent averaged pre-OAE2 I:Ca measurements from each of the sections. D = Demerara Rise; E = Eastbourne; Fe = South Ferriby; N = Newfoundland; R = Raia del Pedale; T = Tarfaya.

The I:Ca data used for comparison with the model come from 6 sections (Zhou et al., 2015, Fig. 7, listed in Table S3). The pre-OAE2 I:Ca baseline value from each section is estimated through averaging the pre-CIE I:Ca measurements from Table S1 of Zhou et al., (2015). For quantitative comparison between the model and the I:Ca data, we create an empirically derived forward proxy model for I:Ca. In this, we took the simulated concentration of IO_3^- and Ca^{2+} in the ocean surface layer of the model at every ocean grid point, and applied the temperature-dependent linear incorporation relationship derived from inorganic calcite synthesis experiment of Zhou et al., (2014), to estimate I:Ca. The converted modelled I:Ca at each section were then directly compared with the mean pre-OAE2 I:Ca measurements using the M score (Fig. 8). We then extracted simulated I:Ca values from the model grid points corresponding to the sections reported by Zhou et al., (2015) and calculated the M-score. The statistical results of this comparison are illustrated in Fig. 8 for the 3 parameterizations chosen for focus in the main text and for Fig. S9 for the full parameterization-combinations and for parameter calibrations derived from internally and WOA-forced dissolved oxygen distributions.

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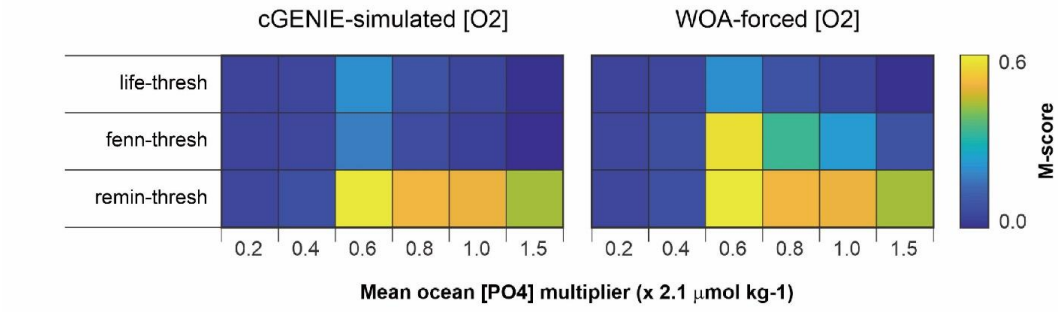

Figure 8. The model skill scores of modeled and measured I:Ca during the pre-OAE2. The iodine cycling parameters are derived from modern simulations with cGENIE-simulated [O_2] and WOA-forced [O_2], respectively. lifetime-thresh = lifetime-threshold; remin-thresh +DOC = reminO2lifetime-threshold +DOC remineralization; remin-thresh = reminO2lifetime-threshold; fenn-thresh = fennel-threshold; life-inhib = lifetime-inhibition; life-remin +DOC = lifetime-reminSO4lifetime +DOC remineralization; life-remin = lifetime-reminSO4lifetime.

Most of the parameterization-combinations tested reach their highest M scores under the assumption of 0.6-0.8 × modern [PO_4] (Fig. S9). Previous analysis using the same climatological configuration of the GENIE model indicated a PO_4 inventory of 1.0 x modern was most consistent with geological redox-related observations prior to OAE2 (Monteiro et al., 2012). However, our assumption here of temperature-dependent POM export and remineralization leads to higher export and shallower more intense ODZs compared to temperature-invariant biological scheme (see: Crichton et al., (2021)). Hence, for a similar degree of ocean anoxia, we would expect the need for a slightly lower nutrient inventory, as we indeed find.

In terms of the I:Ca M-score, we find the parameterization-combinations "reminO2lifetime-threshold" and "fennel-threshold" better replicate the geological observations compared to the "lifetime-threshold". In general, WOA-derived parameter sets perform better than those derived from cGENIE-[O_2], again hinting at the importance of reducing the uncertainties in simulating the modern O_2 cycle in cGENIE. These observations are also largely independent of the ocean PO_4 inventory assumption. Although combinations of parameterizations such as "reminO2lifetime-threshold" with DOC remineralization, "lifetime-inhibition", and "lifetime-reminSO4lifetime", also produce elevated M scores, these combinations do not perform well in replicating iodine gradients within the modern ocean (Figs. S2-4). Thus, until there is a better mechanistic understanding of IO_3^- reduction in the modern ocean, the safest choice is arguably to apply the parameterization that best reproduces modern observations and hence retain use of the "threshold" IO_3^- reduction parameterization for paleo applications.

The pre-OAE2 comparison is revealing because it encapsulates a strong gradient between high and very low I:Ca (Fig. 7), reflecting respectively, high and low surface ocean concentrations of IO_3^- in the model. All three of the parameterization-combinations we focus on here (with WOA-derived parameter values) correctly lead to very low I:Ca values in the lower latitudinal sections (Demerara Rise, Tarfaya, and Raia del Pedale; Fig. 9), although with a tendency to slightly overestimate seawater IO_3^- depletion (cross-plots in Fig. 9). Low ocean surface [IO_3^-] is due to

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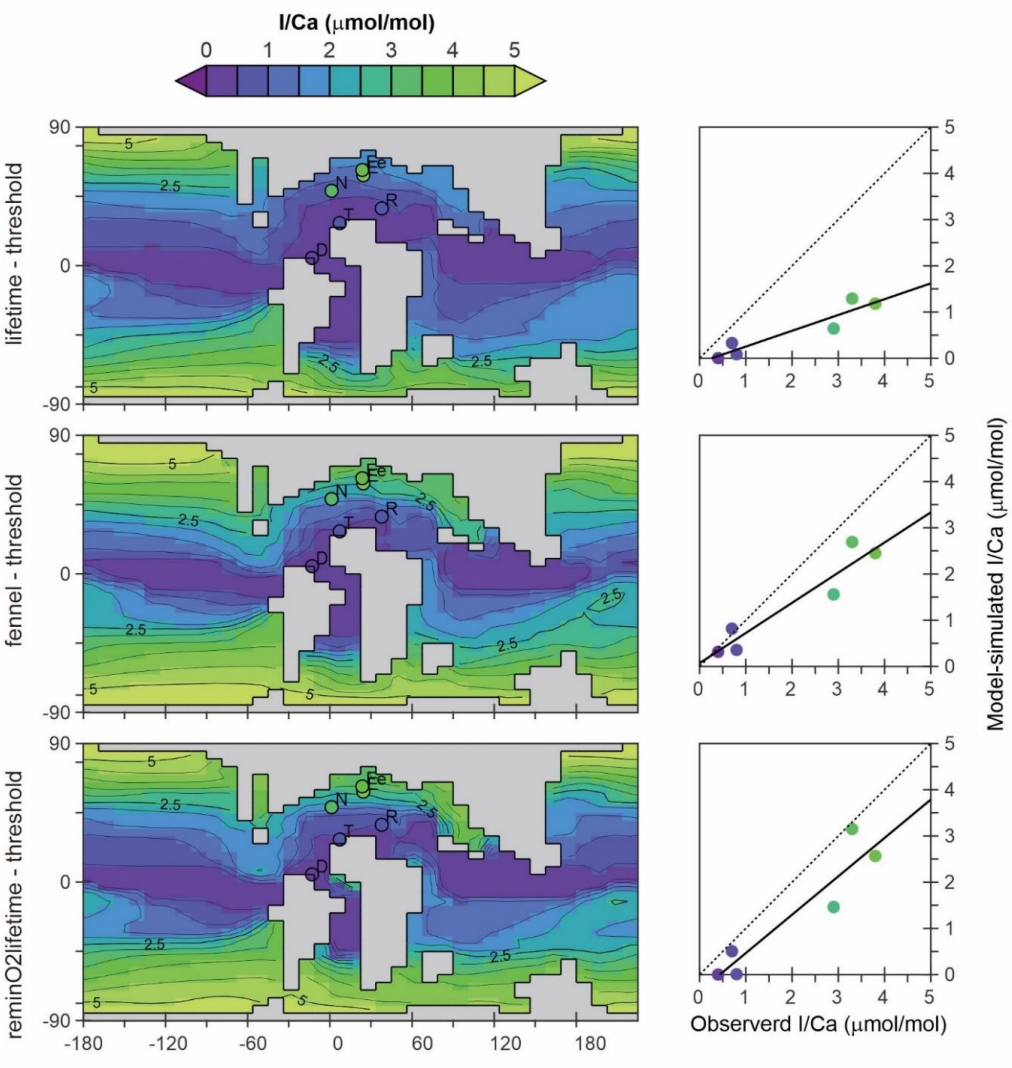
633 the existence of a circum-Equatorial band of intense sub-surface anoxia. In the higher latitudinal sections, including
634 Newfoundland, Eastbourne, and South Ferriby, I:Ca values tend to be underestimated to varying degrees (Fig. 9).
635 Compared to the "lifetime" parameterization, both "reminO2lifetime" and "fennel" oxidation simulate the I:Ca values
636 in these high latitudinal sections rather better, with the regression closer to the 1:1 line (dotted in Fig. 9). We find this
637 relationship instructional for understanding controls in the modern iodine cycle, which we discuss in more detail below.
638



639
640 **Figure 9. The pre-OAE2 I:Ca field derived from cGENIE [IO_3^-] simulations, and the comparison between**
641 **modeled and observed I:Ca from sections.**

642

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We first note that both 'lifetime' and 'fennel' iodine oxidation parameterizations in conjunction with a threshold of iodate reduction and internally generated GENIE-[O_2], give rise to identical parameter values (Table 2). We infer that this is because the modern ocean is predominately well-oxygenated and hence there is little inhibition of I^- oxidation in practice. In the Cretaceous environment, although I^- oxidation inhibition should be widespread, the M-scores are similar (Fig. 8). The rate of I^- oxidation in well oxygenated seawater is likely then critical in explaining elevated I:Ca values at higher Cretaceous latitudes. However, simply decreasing the lifetime in the modern ocean would result in an under estimation of surface ocean [I^-]. The 'fennel-threshold' combination under WOA-[O_2] reveals a trade-off that solves this—a decreased I^- lifetime compensated for by increased rates of I^- release to the ocean interior directly through the biological pump and elevated cellular I:C (3.5×10^{-4} vs. 1.5×10^{-4}). In the Cretaceous ocean this combination allows for both sharper latitudinal gradients in [IO_3^-] (and hence I:Ca) to develop, as well as steeper vertical gradients which allow for non-zero I:Ca values at low latitudes to be captured (cross-plot in Fig. 9). This slight enhancement of the upper ocean [IO_3^-] gradient is also apparent in the present-day analysis (Fig. 5). The combination of 'reminO2lifetime' with a reduction threshold works similarly—a shorter lifetime for I^- under oxic conditions offset in the modern ocean by elevated cellular I:C (Table 2). However, in this case, our gridded parameter search identifies the trade-off as producing the highest M-score for both modelled and WOA oxygen distributions.

What we learn from this is that the cGENIE iodine cycle tuned to modern observations has predictive power under a very different state of ocean oxygenation (and circulation and operation of biological pump). However, this is not true for every choice of parameterization, and the simple 'lifetime-threshold' combination, which when calibrated was statistically almost the best representation of the iodine cycle, was unable to reproduce the latitudinal I:Ca gradients in the Cretaceous ocean. Shortening the lifetime (and adding an inhibition term) together with increasing the assumed cellular I:C, could maintain modern ocean fidelity whilst much better capturing Cretaceous I:Ca. That even better representations of Cretaceous I:Ca were possible but at the expense of reproducing modern observations adequately hints that improvements in our mechanistic understanding are needed, although all of the above assumes that the simulation of the Cretaceous redox landscape is plausible.

5. Conclusions

Using the cGENIE Earth system model, we performed a series of ensemble experiment parameter searches for suitable parameterizations to represent the marine iodine cycle and identified the best performing parameter value combinations in each case. We found that the optimized parameters associated with IO_3^- planktonic uptake, water column IO_3^- reduction and I^- oxidation are within the range of field and experiment observations and hence plausible. Three iodine cycling parameter combinations, "lifetime-threshold", "reminO2lifetime-threshold", and "fennel-threshold" emerged as viable candidates following our tests of the global ocean model M score, and model-data comparison across specific iodine gradients (euphotic latitudinal distribution, depth distribution, and ODZ distribution). We further evaluate the plausibility of our parameterizations and their paleo and ocean deoxygenation applicability by contrasting forward-proxy model generated I:Ca values against observations, taking the (pre-OAE2) Cretaceous redox landscape as a case study. While some model-data discrepancies emerge for both modern and paleo, these highlight that future observational and/or experimental work is necessary to better constrain modern iodine

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cycling mechanisms and related spatiotemporal heterogeneities. While we further identified the importance of improving the simulation of dissolved oxygen distributions in models, equally, we found that our conclusions regarding preferred parameterizations and even specific parameter values, was not overly dependent on the specific details of the simulated modern OMZs. Overall, our analysis gives us a degree of confidence that carbonate I:Ca ratios can be interpreted in terms of past ocean oxygenation using models such as 'cGENIE'.

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Competing interests. *The contact author has declared that none of the authors has any competing interests.*

Author contributions. *KC, AR, and DH conceptualized the research presented in this paper. DH and AR acquired funding to support the study. AR developed iodine tracer and associated biogeochemical mechanisms in cGENIE. KC and DH designed model performance under the modern ocean configurations. KC compiled the modern ocean iodine database and the Cretaceous I:Ca data. KC ran the modern-ocean cGENIE analysis and performed model-data comparison. AR performed model-data evaluation for Cretaceous configurations. KC prepared the manuscript with contributions from all co-authors.*

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Code availability. *The code for the version of the 'muffin' release of the cGENIE Earth system model used in this paper, is tagged as v0.9.13, and is assigned a DOI: 10.5281/zenodo.3999080.*
Configuration files for the specific experiments presented in the paper can be found in the directory: genie-userconfigs/PUBS/chengetal.BG.2020. Details of the experiments, plus the command line needed to run each one, are given in the readme.txt file in that directory. All other configuration files and boundary conditions are provided as part of the code release.
A manual detailing code installation, basic model configuration, tutorials covering various aspects of model configuration, experimental design, and output, plus the processing of results, is assigned a DOI: 10.5281/zenodo.4305997.

References

Amachi, S., Kawaguchi, N., Muramatsu, Y., Tsuchiya, S., Watanabe, Y., Shinoyama, H., and Fujii, T.: Dissimilatory iodate reduction by marine Pseudomonas sp. strain SCT, Appl. Environ. Microbiol., 73, 5725–5730, https://doi.org/10.1128/AEM.00241-07, 2007.

Boscolo-Galazzo, F., Crichton, K. A., Ridgwell, A., Mawbey, E. M., Wade, B. S., and Pearson, P. N.: Temperature controls carbon cycling and biological evolution in the ocean twilight zone, Science (80-.)., 371, 1148–1152, https://doi.org/10.1126/science.abb6643, 2021.

Bowman, C. N., Lindskog, A., Kozik, N. P., Richbourg, C. G., Owens, J. D., and Young, S. A.: Integrated sedimentary, biotic, and paleoredox dynamics from multiple localities in southern Laurentia during the late Silurian (Ludfordian) extinction event, Palaeogeogr. Palaeoclimatol. Palaeoecol., 553, 109799, https://doi.org/10.1016/j.palaeo.2020.109799, 2020.

Broecker, W. S. and Peng, T. H.: Tracers in the Sea., https://doi.org/10.1016/0016-7037(83)90075-3, 1983.

Campos, M. L. A. M., Farrenkopf, A. M., Jickells, T. D., and Luther, G. W.: A comparison of dissolved iodine cycling at the Bermuda Atlantic Time-series station and Hawaii Ocean Time-series station, Deep. Res. Part II Top. Stud.

https://doi.org/10.5194/egusphere-2024-677
Preprint. Discussion started: 2 April 2024

721 Oceanogr., 43, 455–466, https://doi.org/10.1016/0967-0645(95)00100-x, 1996.
722 Cao, L., Eby, M., Ridgwell, A., Caldeira, K., Archer, D., Ishida, A., Joos, F., Matsumoto, K., and Mikolajewicz, U.:
723 The role of ocean transport in the uptake of anthropogenic CO 2, Biogeosciences, 375–390, 2009.
724 Chance, R., Weston, K., Baker, A. R., Hughes, C., Malin, G., Carpenter, L., Meredith, M. P., Clarke, A., Jickells, T.
725 D., Mann, P., and Rossetti, H.: Seasonal and interannual variation of dissolved iodine speciation at a coastal Antarctic
726 site, Mar. Chem., 118, 171–181, https://doi.org/10.1016/j.marchem.2009.11.009, 2010.
727 Chance, R., Baker, A. R., Carpenter, L., and Jickells, T. D.: The distribution of iodide at the sea surface, Environ. Sci.
728 Process. Impacts, 16, 1841–1859, https://doi.org/10.1039/c4em00139g, 2014.
729 Chance, R., Tinel, L., Sherwen, T., Baker, A., Bell, T., Brindle, J., Campos, M., Croot, P., Ducklow, H., He, P.,
730 Hoogakker, B., Hopkins, F., Hughes, C., Jickells, T., Loades, D., Reyes Macaya, D., Mahajan, A., Malin, G., Phillips,
731 D., Sinha, A., Sarkar, A., Roberts, I., Roy, R., Song, X., Winklebauer, H., Wuttig, K., Yang, M., Zhou, P., and
732 Carpenter, L.: Global sea-surface iodide observations, 1967-2018., https://doi.org/doi:10.5285/7e77d6b9-83fb-41e0-
733 e053-6c86abc069d0, 2019.
734 Chapman, P.: Changes in iodine speciation in the Benguela current upwelling system, Deep Sea Res. Part A, Oceanogr.
735 Res. Pap., 30, 1247–1259, https://doi.org/10.1016/0198-0149(83)90083-3, 1983.
736 Councell, T. B., Landa, E. R., and Lovley, D. R.: Microbial reduction of iodate, Water. Air. Soil Pollut., 100, 99–106,
737 1997.
738 Crichton, K. A., Wilson, J. D., Ridgwell, A., and Pearson, P. N. P. N.: Calibration of temperature-dependent ocean
739 microbial processes in the cGENIE.muffin (v0.9.13) Earth system model, Geosci. Model Dev., 14, 125–149,
740 https://doi.org/10.5194/gmd-14-125-2021, 2021.
741 Cutter, G. A., Moffett, J. G., Nielsdóttir, M. C., and Sanial, V.: Multiple oxidation state trace elements in suboxic
742 waters off Peru: In situ redox processes and advective/diffusive horizontal transport, Mar. Chem., 201, 77–89,
743 https://doi.org/10.1016/j.marchem.2018.01.003, 2018.
744 Dalsgaard, T., Stewart, F. J., Thamdrup, B., De Brabandere, L., Revsbech, N. P., Ulloa, O., Canfield, D. E., and
745 Delong, E. F.: Oxygen at nanomolar levels reversibly suppresses process rates and gene expression in anammox and
746 denitrification in the oxygen minimum zone off Northern Chile, MBio, 5, 1–14, https://doi.org/10.1128/mBio.01966-
747 14, 2014.
748 Ding, Y., Sun, W., Liu, S., Xie, J., Tang, D., Zhou, X., Zhou, L., Li, Z., Song, J., Li, Z., Xu, H., Tang, P., Liu, K., Li,
749 W., and Chen, D.: Low oxygen levels with high redox heterogeneity in the late Ediacaran shallow ocean: Constraints
750 from I/(Ca + Mg) and Ce/Ce* of the Dengying Formation, South China, Geobiology, 20, 790–809,
751 https://doi.org/10.1111/gbi.12520, 2022.
752 Edwards, A. and Truesdale, V. W.: Regeneration of inorganic iodine species in loch etive, a natural leaky incubator,
753 Estuar. Coast. Shelf Sci., 45, 357–366, https://doi.org/10.1006/ecss.1996.0185, 1997.
754 Edwards, C. T., Fike, D. A., Saltzman, M. R., Lu, W., and Lu, Z.: Evidence for local and global redox conditions at
755 an Early Ordovician (Tremadocian) mass extinction, Earth Planet. Sci. Lett., 481, 125–135,
756 https://doi.org/10.1016/j.epsl.2017.10.002, 2018.
757 Edwards, N. R. and Marsh, R.: Uncertainties due to transport-parameter sensitivity in an efficient 3-D ocean-climate

https://doi.org/10.5194/egusphere-2024-677
Preprint. Discussion started: 2 April 2024

model, Clim. Dyn., 415–433, https://doi.org/10.1007/s00382-004-0508-8, 2005.

Elderfield, H. and Truesdale, V. W.: On the biophilic nature of iodine in seawater, Earth Planet. Sci. Lett., 50, 105–114, https://doi.org/10.1016/0012-821X(80)90122-3, 1980.

Fang, H., Tang, D., Shi, X., Zhou, L., Zhou, X., Wu, M., Song, H., and Riding, R.: Early Mesoproterozoic Ca-carbonate precipitates record fluctuations in shallow marine oxygenation, Precambrian Res., 373, 106630, https://doi.org/10.1016/j.precamres.2022.106630, 2022.

Farrenkopf, A. M. and Luther, G. W.: Iodine chemistry reflects productivity and denitrification in the Arabian Sea: Evidence for flux of dissolved species from sediments of western India into the OMZ, Deep. Res. Part II Top. Stud. Oceanogr., 49, 2303–2318, https://doi.org/10.1016/S0967-0645(02)00038-3, 2002.

Farrenkopf, A. M., Dollhopf, M. E., Chadhain, S. N., Luther, G. W., and Nealson, K. H.: Reduction of iodate in seawater during Arabian Sea shipboard incubations and in laboratory cultures of the marine bacterium Shewanella putrefaciens strain MR-4, Mar. Chem., 57, 347–354, https://doi.org/10.1016/S0304-4203(97)00039-X, 1997.

Fennel, K., Follows, M., and Falkowski, P. G.: The co-evolution of the nitrogen, carbon and oxygen cycles in the Proterozoic ocean, Am. J. Sci., 305, 526–545, 2005.

Garcia, H. E., Weathers, K., Paver, C. R., Smolyar, I., Boyer, T. P., Locarnini, R. A., Zweng, M. M., Mishonov, A. V., Baranova, O. K., Seidov, D., and Reagan, J. R.: Volume 3: Dissolved Oxygen, Apparent Oxygen Utilization, and Oxygen Saturation. A. Mishonov Technical Ed.; NOAA Atlas NESDIS 83, World Ocean Atlas 2018, 38 pp., 2018.

Hardisty, D. S., Lu, Z., Planavsky, N. J., Bekker, A., Philippot, P., Zhou, X., and Lyons, T. W.: An iodine record of Paleoproterozoic surface ocean oxygenation, Geology, 42, 619–622, https://doi.org/10.1130/G35439.1, 2014.

Hardisty, D. S., Horner, T. J., Wankel, S. D., Blusztajn, J., and Nielsen, S. G.: Experimental observations of marine iodide oxidation using a novel sparge-interface MC-ICP-MS technique, Chem. Geol., 532, 119360, https://doi.org/10.1016/j.chemgeo.2019.119360, 2020.

Hardisty, D. S., Horner, T. J., Evans, N., Moriyasu, R., Babbin, A. R., Wankel, S. D., Moffett, J. W., and Nielsen, S. G.: Limited iodate reduction in shipboard seawater incubations from the Eastern Tropical North Pacific oxygen deficient zone, Earth Planet. Sci. Lett., 554, 116676, https://doi.org/10.1016/j.epsl.2020.116676, 2021.

Hashim, M. S., Burke, J. E., Hardisty, D. S., and Kaczmarek, S. E.: Iodine incorporation into dolomite: Experimental constraints and implications for the iodine redox proxy and Proterozoic Ocean, Geochim. Cosmochim. Acta, 338, 365–381, https://doi.org/10.1016/j.gca.2022.10.027, 2022.

He, P., Hou, X., and Aldahan, A.: Iodine isotopes species fingerprinting environmental conditions in surface water along the northeastern Atlantic Ocean, Sci. Rep., 3, 1–9, https://doi.org/10.1038/srep02685, 2013.

Hepach, H., Hughes, C., Hogg, K., Collings, S., and Chance, R.: Senescence as the main driver of iodide release from a diverse range of marine phytoplankton, Biogeosciences, 17, 2453–2471, https://doi.org/10.5194/bg-17-2453-2020, 2020.

Hughes, C., Barton, E., Hepach, H., Chance, R., Wadley, M. R., Pickering, M. D., Hogg, K., Pommerening-r, A., Stevens, D. P., and Jickells, T. D.: Oxidation of iodide to iodate by cultures of marine ammonia-oxidising bacteria, Mar. Chem., 234, https://doi.org/10.1016/j.marchem.2021.104000, 2021.

Hülse, D., Arndt, S., and Ridgwell, A.: Mitigation of Extreme Ocean Anoxic Event Conditions by Organic Matter

https://doi.org/10.5194/egusphere-2024-677
Preprint. Discussion started: 2 April 2024

795 Sulfurization, Paleoceanogr. Paleoclimatology, 34, 476–489, https://doi.org/10.1029/2018PA003470, 2019.
796 Jia-zhong, Z. and Whitfield, M.: KINETICS OF INORGANIC REDOX REACTIONS IN SEAWATER I . The
797 reduction of iodate by bisulphide Micro-organisms play a dominant role in the diagenesis of organic-rich sediments .
798 The oxidative breakdown of the organic matter , with the accom- panying reductio, Science (80-.)., 19, 121–137,
799 1986.
800 Jiang, Z., Cui, M., Qian, L., Jiang, Y., Shi, L., Dong, Y., Li, J., and Wang, Y.: Abiotic and Biotic Reduction of Iodate
801 Driven by Shewanella oneidensis MR-1 , Environ. Sci. Technol., https://doi.org/10.1021/acs.est.3c06490, 2023.
802 Kerisit, S. N., Smith, F. N., Saslow, S. A., Hoover, M. E., Lawter, A. R., and Qafoku, N. P.: Incorporation Modes of
803 Iodate in Calcite, Environ. Sci. Technol., 52, 5902–5910, https://doi.org/10.1021/acs.est.8b00339, 2018.
804 Laws, E. A., Falkowski, P. G., Smith, W. O., Ducklow, H., and James J McCarthy: Temperature effects on export
805 production in the open ocean, Global Biogeochem. Cycles, 14, 1231–1246, 2000.
806 Liu, A., Tang, D., Shi, X., Zhou, X., Zhou, L., Shang, M., Li, Y., and Fang, H.: Mesoproterozoic oxygenated deep
807 seawater recorded by early diagenetic carbonate concretions from the Member IV of the Xiamaling Formation, North
808 China, Precambrian Res., 341, 105667, https://doi.org/10.1016/j.precamres.2020.105667, 2020.
809 Lu, W., Ridgwell, A., Thomas, E., Hardisty, D. S., Luo, G., Algeo, T. J., Saltzman, M. R., Gill, B. C., Shen, Y., Ling,
810 H. F., Edwards, C. T., Whalen, M. T., Zhou, X., Gutchess, K. M., Jin, L., Rickaby, R. E. M., Jenkyns, H. C., Lyons,
811 T. W., Lenton, T. M., Kump, L. R., and Lu, Z.: Late inception of a resiliently oxygenated upper ocean, Science (80-.).,
812 361, 174–177, https://doi.org/10.1126/science.aar5372, 2018.
813 Lu, W., Rickaby, R. E. M., Hoogakker, B. A. A., Rathburn, A. E., Burkett, A. M., Dickson, A. J., Martínez-Méndez,
814 G., Hillenbrand, C. D., Zhou, X., Thomas, E., and Lu, Z.: I/Ca in epifaunal benthic foraminifera: A semi-quantitative
815 proxy for bottom water oxygen in a multi-proxy compilation for glacial ocean deoxygenation, Earth Planet. Sci. Lett.,
816 533, https://doi.org/10.1016/j.epsl.2019.116055, 2020a.
817 Lu, W., Dickson, A. J., Thomas, E., Rickaby, R. E. M., Chapman, P., and Lu, Z.: Refining the planktic foraminiferal
818 I/Ca proxy: Results from the Southeast Atlantic Ocean, Geochim. Cosmochim. Acta, 287, 318–327,
819 https://doi.org/10.1016/j.gca.2019.10.025, 2020b.
820 Lu, Z., Jenkyns, H. C., and Rickaby, R. E. M.: Iodine to calcium ratios in marine carbonate as a paleo-redox proxy
821 during oceanic anoxic events, Geology, 38, 1107–1110, https://doi.org/10.1130/G31145.1, 2010.
822 Lu, Z., Hoogakker, B. A. A., Hillenbrand, C.-D., Zhou, X., Thomas, E., Gutchess, K. M., Lu, W., Jones, L., and
823 Rickaby, R. E. M.: Oxygen depletion recorded in upper waters of the glacial Southern Ocean, Nat. Commun., 7, 1–8,
824 https://doi.org/10.1038/ncomms11146, 2016.
825 Luther, G. W.: Review on the physical chemistry of iodine transformations in the oceans, Front. Mar. Sci., 10, 1–16,
826 https://doi.org/10.3389/fmars.2023.1085618, 2023.
827 Luther, G. W. and Campbell, T.: Iodine speciation in the water column of the Black Sea, Deep Sea Res. Part A.
828 Oceanogr. Res. Pap., 38, S875–S882, https://doi.org/10.1016/s0198-0149(10)80014-7, 1991.
829 Luther, G. W., Wu, J., and Cullen, J. B.: Redox Chemistry of Iodine in Seawater, Adv. Chem., 135–155, 1995.
830 Luther, W.: Iodine Chemistry in the Water the Chesapeake Bay : Evidence Iodine Forms Column of for Organic,
831 Esruarine, Coast. Shelf Sci., 32, 267–279, 1991.

https://doi.org/10.5194/egusphere-2024-677
Preprint. Discussion started: 2 April 2024

Lyons, T. W., Reinhard, C. T., and Planavsky, N. J.: The rise of oxygen in Earth's early ocean and atmosphere, Nature, 506, 307–315, https://doi.org/10.1038/nature13068, 2014.

Marsh, R., Müller, S. A., Yool, A., and Edwards, N. R.: Incorporation of the C-GOLDSTEIN efficient climate model into the GENIE framework: "eb_go_gs" configurations of GENIE, Geosci. Model Dev., 4, 957–992, https://doi.org/10.5194/gmd-4-957-2011, 2011.

Martin, T. S., Primeau, F., and Casciotti, K. L.: Modeling oceanic nitrate and nitrite concentrations and isotopes using a 3-D inverse N cycle model, Biogeosciences, 347–367, 2019.

Matsumoto, K.: Radiocarbon-based circulation age of the world oceans, J. Geophys. Res. Ocean., 112, 1–7, https://doi.org/10.1029/2007JC004095, 2007.

Monteiro, F. M., Pancost, R. D., Ridgwell, A., and Donnadieu, Y.: Nutrients as the dominant control on the spread of anoxia and euxinia across the Cenomanian-Turonian oceanic anoxic event (OAE2): Model-data comparison, Paleoceanography, 27, https://doi.org/10.1029/2012PA002351, 2012.

Moriyasu, R., Evans, N., Bolster, K. M., Hardisty, D. S., and Moffett, J. W.: The Distribution and Redox Speciation of Iodine in the Eastern Tropical North Pacific Ocean, Global Biogeochem. Cycles, 34, 1–23, https://doi.org/10.1029/2019GB006302, 2020.

Moriyasu, R., Bolster, K. M., Hardisty, D. S., Kadko, D. C., Stephens, M. P., and Moffett, J. W.: Meridional survey of the Central Pacific reveals iodide accumulation in equatorial surface waters and benthic sources in the abyssal plain, Global Biogeochem. Cycles, https://doi.org/10.1029/2021GB007300, 2023.

Podder, J., Lin, J., Sun, W., Botis, S. M., Tse, J., Chen, N., Hu, Y., Li, D., Seaman, J., and Pan, Y.: Iodate in calcite and vaterite: Insights from synchrotron X-ray absorption spectroscopy and first-principles calculations, Geochim. Cosmochim. Acta, 198, 218–228, https://doi.org/10.1016/j.gca.2016.11.032, 2017.

Pohl, A., Lu, Z., Lu, W., Stockey, R. G., Elrick, M., Li, M., Desrochers, A., Shen, Y., He, R., Finnegan, S., and Ridgwell, A.: Vertical decoupling in Late Ordovician anoxia due to reorganization of ocean circulation, Nat. Geosci., 14, 868–873, https://doi.org/10.1038/s41561-021-00843-9, 2021.

Pohl, A., Ridgwell, A., Stockey, R. G., Thomazo, C., Keane, A., Vennin, E., and Scotese, C. R.: Continental configuration controls ocean oxygenation during the Phanerozoic, Nature, 608, 523–527, https://doi.org/10.1038/s41586-022-05018-z, 2022.

Rapp, I., Schlosser, C., Barraqueta, J. M., Wenzel, B., Lüdke, J., and Scholten, J.: Controls on redox-sensitive trace metals in the Mauritanian oxygen minimum zone, 4157–4182, 2019.

Rapp, I., Schlosser, C., Browning, T. J., Wolf, F., Le Moigne, F. A. C., Gledhill, M., and Achterberg, E. P.: El Niño-Driven Oxygenation Impacts Peruvian Shelf Iron Supply to the South Pacific Ocean, Geophys. Res. Lett., 47, https://doi.org/10.1029/2019GL086631, 2020.

Reinhard, C. T. and Planavsky, N. J.: The History of Ocean Oxygenation, Ann. Rev. Mar. Sci., 14, 331–353, https://doi.org/10.1146/annurev-marine-031721-104005, 2022.

Reinhard, C. T., Planavsky, N. J., Olson, S. L., Lyons, T. W., and Erwin, D. H.: Earth's oxygen cycle and the evolution of animal life, Proc. Natl. Acad. Sci. U. S. A., 113, 8933–8938, https://doi.org/10.1073/pnas.1521544113, 2016.

Reinhard, C. T., Olson, S. L., Turner, S. K., Pälike, C., Kanzaki, Y., and Ridgwell, A.: Oceanic and atmospheric

https://doi.org/10.5194/egusphere-2024-677
Preprint. Discussion started: 2 April 2024

methane cycling in the cGENIE Earth system model, Geosci. Model Dev., 1–45, https://doi.org/10.5194/gmd-2020-32, 2020.

Remmelzwaal, S. R. C., Dixon, S., Parkinson, I. J., Schmidt, D. N., Monteiro, F. M., Sexton, P., Fehr, M. A., Peacock, C., Donnadieu, Y., and James, R. H.: Investigating Ocean Deoxygenation During the PETM Through the Cr Isotopic Signature of Foraminifera, Paleoceanogr. Paleoclimatology, 34, 917–929, https://doi.org/10.1029/2018PA003372, 2019.

Reyes-Umana, V., Henning, Z., Lee, K., Barnum, T. P., and Coates, J. D.: Genetic and phylogenetic analysis of dissimilatory iodate-reducing bacteria identifies potential niches across the world's oceans, ISME J., https://doi.org/10.1038/s41396-021-01034-5, 2021.

Ridgwell, A., Hargreaves, J. C., Edwards, N. R., Annan, J. D., Lenton, T. M., Marsh, R., Yool, A., and Watson, A.: Marine geochemical data assimilation in an efficient Earth system model of global biogeochemical cycling, Biogeosciences, 4, 87–104, https://doi.org/10.5194/bg-4-87-2007, 2007.

Rue, E. L., Smith, G. J., Cutter, G. A., and Bruland, K. W.: The response of trace element redox couples to suboxic conditions in the water column, Deep. Res. Part I Oceanogr. Res. Pap., 44, 113–134, https://doi.org/10.1016/S0967-0637(96)00088-X, 1997.

Schnur, A. A., Sutherland, K. M., Hansel, C. M., and Hardisty, D. S.: Rates and pathways of iodine speciation transformations at the Bermuda Atlantic Time Series, Front. Mar. Sci., 1–14, https://doi.org/10.3389/fmars.2023.1272870, 2024.

Shang, M., Tang, D., Shi, X., Zhou, L., Zhou, X., Song, H., and Jiang, G.: A pulse of oxygen increase in the early Mesoproterozoic ocean at ca. 1.57–1.56 Ga, Earth Planet. Sci. Lett., 527, 115797, https://doi.org/10.1016/j.epsl.2019.115797, 2019.

Sherwen, T., Chance, R. J., Tinel, L., Ellis, D., Evans, M. J., and Carpenter, L. J.: A machine-learning-based global sea-surface iodide distribution, Earth Syst. Sci. Data, 11, 1239–1262, 2019.

Ștreangă, I.-M., Repeta, D. J., Blusztajn, J. S., and Horner, T. J.: Speciation and cycling of iodine in the subtropical North Paci fi c Ocean, Front. Mar. Sci., 1–15, https://doi.org/10.3389/fmars.2023.1272968, 2024.

Tang, D., Fang, H., Shi, X., Liang, L., Zhou, L., Xie, B., Huang, K., Zhou, X., Wu, M., and Riding, R.: Mesoproterozoic Molar Tooth Structure Related to Increased Marine Oxygenation, J. Geophys. Res. Biogeosciences, 128, 1–18, https://doi.org/10.1029/2022jg007077, 2023.

Thamdrup, B., Dalsgaard, T., and Revsbech, N. P.: Widespread functional anoxia in the oxygen minimum zone of the Eastern South Pacific, Deep. Res. Part I Oceanogr. Res. Pap., 65, 36–45, https://doi.org/10.1016/j.dsr.2012.03.001, 2012.

Truesdale, V. W., Bale, A. J., and Woodward, E. M. S.: The meridional distribution of dissolved iodine in near-surface waters of the Atlantic Ocean, Prog. Oceanogr., 45, 387–400, https://doi.org/10.1016/S0079-6611(00)00009-4, 2000.

Truesdale, V. W., Watts, S. F., and Rendell, A. R.: On the possibility of iodide oxidation in the near-surface of the Black Sea and its implications to iodine in the general ocean, Deep. Res. Part I, 48, 2397–2412, 2001.

Truesdale, V. W., Nausch, G., and Waite, T. J.: The effects of the 2001 Barotropic intrusion of bottom-water upon the vertical distribution of inorganic iodine in the Gotland Deep, Cont. Shelf Res., 55, 155–167,

https://doi.org/10.5194/egusphere-2024-677
Preprint. Discussion started: 2 April 2024

https://doi.org/10.1016/j.csr.2013.01.005, 2013.

Tsunogai, S.: Iodine in the deep water of the ocean, Deep. Res., 18, 913–919, 1971.

Uahengo, C. I., Shi, X., Jiang, G., and Vatuva, A.: Transient shallow-ocean oxidation associated with the late Ediacaran Nama skeletal fauna: Evidence from iodine contents of the Lower Nama Group, southern Namibia, Precambrian Res., 343, 105732, https://doi.org/10.1016/j.precamres.2020.105732, 2020.

Wadley, M. R., Stevens, D. P., Jickells, T. D., Hughes, C., Chance, R., Hepach, H., Tinel, L., and Carpenter, L. J.: A Global Model for Iodine Speciation in the Upper Ocean, Global Biogeochem. Cycles, 34, 0–3, https://doi.org/10.1029/2019GB006467, 2020.

Watterson, I. G.: Non-dimensional measures of climate model performance, Int. J. Climatol., 16, 379–391, https://doi.org/10.1002/(SICI)1097-0088(199604)16:4<379::AID-JOC18>3.0.CO;2-U, 1996.

Wei, B., Tang, D., Shi, X., Lechte, M., Zhou, L., Zhou, X., and Song, H.: A Pulsed Oxygenation in Terminal Paleoproterozoic Ocean: Evidence From the Transition Between the Chuanlinggou and Tuanshanzi Formations, North China, Geochemistry, Geophys. Geosystems, 22, 1–23, https://doi.org/10.1029/2020GC009612, 2021.

Wong, G. T. F. and Brewer, P. G.: The marine chemistry of iodine in anoxic basins, Geochim. Cosmochim. Acta, 41, 151–159, https://doi.org/10.1016/0016-7037(77)90195-8, 1977.

Wong, G. T. F., Takayanagi, K., and Todd, J. F.: Dissolved iodine in waters overlying and in the Orca Basin, Gulf of Mexico, Mar. Chem., 17, 177–183, https://doi.org/10.1016/0304-4203(85)90072-6, 1985.

Wong, G. T. F., Piumsomboon, A. U., Dunstan, W. M., Wong, G. T. F., Piumsomboon, A. U., and Dunstan, W. M.: The transformation of iodate to iodide in marine phytoplankton cultures, Mar. Geol. Prog. Ser., 237, 27–39, 2002.

Yamazaki, C., Kashiwa, S., Horiuchi, A., and Kasahara, Y.: A novel dimethylsulfoxide reductase family of molybdenum enzyme , Idr , is involved in iodate respiration by Pseudomonas sp . SCT, 22, 2196–2212, https://doi.org/10.1111/1462-2920.14988, 2020.

Yu, Y., Chen, Y., Li, D., and Su, J.: A transient oxygen increase in the Mesoproterozoic ocean at ~1.44 Ga: Geochemical evidence from the Tieling Formation, North China Platform, Precambrian Res., 369, 106527, https://doi.org/10.1016/j.precamres.2021.106527, 2022.

Zhang, S., Xu, C., Creeley, D., Ho, Y. F., Li, H. P., Grandbois, R., Schwehr, K. A., Kaplan, D. I., Yeager, C. M., Wellman, D., and Santschi, P. H.: Response to comment on "iodine-129 and iodine-127 speciation in groundwater at hanford site, U.S.: Iodate incorporation into calcite," Environ. Sci. Technol., 47, 13205–13206, https://doi.org/10.1021/es4046132, 2013.

Zhou, X., Thomas, E., Rickaby, R. E. M., Winguth, A. M. E., and Lu, Z.: I/Ca evidence for upper ocean deoxygenation during the PETM, Paleoceanography, 29, 964–975, https://doi.org/10.1002/2014PA002702, 2014.

Zhou, X., Jenkyns, H. C., Owens, J. D., Junium, C. K., Zheng, X. Y., Sageman, B. B., Hardisty, D. S., Lyons, T. W., Ridgwell, A., and Lu, Z.: Upper ocean oxygenation dynamics from I/Ca ratios during the Cenomanian-Turonian OAE 2, Paleoceanography, 30, 510–526, https://doi.org/10.1002/2014PA002741, 2015.

Žic, V., Carić, M., and Ciglenečki, I.: The impact of natural water column mixing on iodine and nutrient speciation in a eutrophic anchialine pond (Rogoznica Lake, Croatia), Estuar. Coast. Shelf Sci., 133, 260–272, https://doi.org/10.1016/j.ecss.2013.09.008, 2013.