

Ocean acidification and aragonite saturation compounding anthropogenic stressors on marine life

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Running head: Ocean acidification, stressors and regime-shift risk in marine water chemistry

Abstract

Aragonite saturation is a key measure of ocean acidification, but a falling index also reshapes pH, redox and zeta potential, colloidal chemistry and viscosity, which together support plankton and shell-building life. The cold, productive waters that sustain most of the marine food web now have winter aragonite saturation and pH within the marginal zone for calcifying organisms (Ω 1.0 to 1.6), and reported annual means understate this decline. Judged against zone-specific ecological reference thresholds, 37 to 52% of the world's ocean surface already sits below its winter threshold. Marine life is already stressed by exploitation, climate disruption and pollution; an interpretive synthesis of documented declines places marine animal life at about 35 to 54% of its 1850 level, most of that loss recorded before acidification became a factor. Ocean acidification should therefore be treated as an active component of cumulative marine ecosystem risk, not solely as a future concern.

1. Introduction

1.1 Aragonite saturation index and calcification

The aragonite saturation index, Ω , is a calculated index based on the measurement of several parameters: partial pressure of carbon dioxide, pH, total alkalinity, calcium concentration, salinity and temperature¹. It is a measure of how easily marine animals can build and maintain shells and skeletons of aragonite, which is a form of calcium carbonate. Aragonite dissolves below $\Omega = 1$, and a shell can be easily formed above $\Omega = 1.6$ (the marginal-zone value derived in Section 2.4); sensitive marine life starts to suffer in the marginal zone between the two values.

Policy decisions should be based on aragonite values that accurately reflect the true environmental conditions in which the animals live. They should not be average values; rather, they should be the lowest sustained values experienced during the winter months or in cold upwelling streams so that the measurement reflects the true exposure that marine life encounters.

Reports typically give index values as means, taken over a 12-month period, which will always be higher than the winter, colder-water readings. Reporting of the aragonite index

may therefore not reflect the true exposure of marine life to the most corrosive, cold-water conditions. An example is the ocean acidification planetary boundary, which is expressed as a global average of 2.75Ω , or 80% of the pre-industrial value of 3.44Ω ^{2,3}. That 80% threshold was chosen to keep high-latitude surface waters, which the framework recognizes as the first to approach undersaturation, above aragonite undersaturation, and to ensure adequate conditions for most coral systems. The limitation is not that the boundary ignores cold water, but that it is reported as a single, global, annual mean. The winter minima measured at individual high-latitude stations in this report confirm that the boundary, which was set to protect marine life, has already been passed. A recent reassessment revised the pre-industrial baseline upward to 3.51Ω , found that by 2020 the global mean had entered the uncertainty range of the 20% boundary and that four of seven ocean regions, the Arctic among them, had crossed the boundary, and argued for a stricter 10% boundary that the whole surface ocean passed by 2000⁴; the cold, productive seas crossed that line earlier still.

We examined four cold-water zones to test this: the Iceland Sea, the Irminger Sea, the subarctic North Pacific and the Southern Ocean. The Iceland Sea, the subarctic North Pacific and the Southern Ocean already sit at or below the ecological reference threshold of Ω 1.6 in winter, while the warmer Irminger Current water sits just above that threshold. The index tracks temperature, so the warmer Munida site off New Zealand reads higher.

1.2 The influence of aragonite and pH on zeta and redox potential

The sea absorbs carbon dioxide from the decomposition of organic matter and from the atmosphere through its air-water interface. The surface micro-layer (SML) can reduce the rate of gas transfer across the interface by up to about a third⁵. The SML is a biogenic layer, 1 to 1000 microns thick,⁶ comprising lipids, surfactants, bacteria, plankton, and organic and inorganic matter.

Two terms from colloid chemistry recur throughout this report: when pH changes, both zeta potential and redox potential will also change. Zeta potential is the effective surface charge on particles, and redox potential is the water's net tendency to oxidize or reduce. The zeta potential governs whether water is clear or turbid: at zero the positive and negative charges are balanced, so the particles no longer repel one another and can coagulate, settle and leave clear water, whereas a strongly negative or strongly positive zeta potential holds the particles apart as a stable colloidal suspension. As carbon dioxide diffuses through the SML into the underlying water, a series of chemical reactions takes place: pH drops, zeta potential drops (less negative), the solubility of metal ions increases (iron becomes more soluble), bicarbonate increases, carbonate decreases, and total alkalinity remains the same.

The redox potential requires more care than the other two terms. In pure water, a falling pH would raise the redox potential of proton-coupled couples, following the Nernst equation, but in natural waters, the oxidation of dissolved organic matter consumes oxygen and other electron acceptors and depresses the redox potential, an effect that overrides the thermodynamic pH term. In a sea with an active organic cycle, pH and redox potential therefore fall together, both driven by the oxidation of organic matter rather than one by

the other. Zeta potential is the primary direct control variable for particle aggregation in seawater coagulation and flocculation. The zeta potential is already low in seawater due to the high ionic strength and the presence of divalent calcium and magnesium ions, which compress the electrical double layer; therefore, while the pH-driven changes are modest in absolute terms, they still occur, and because marine particles are already close to the coagulation threshold, even a modest shift can have an effect. Redox potential acts on the same chemistry indirectly by altering the speciation of the metals and organic ligands that take part in the reactions.

The falling pH and carbonate ion concentration, together with the loss of buffer capacity ⁷, are expected to alter how organic matter coagulates and self-assembles into the calcium-bridged gels described by Verdugo ⁸, although that response has not yet been measured at open-ocean pH. A colloidal suspension then starts to develop and provides the conditions in which bacterial levels and turbidity increase; bacterial respiration of that organic matter lowers the oxygen and the pH and reduces the buffer capacity further ⁹, and in our assessment the redox and zeta potentials fall with them. It is a cascading set of chemical reactions, a self-reinforcing loop that is only restrained by the carbonate buffer or if the bacteria run out of energy. In our own operating records from large public aquaria holding 20 million litres of water, which are unpublished, these cascade reactions accelerate over a 10-hour period, with measurable declines in water quality and animal condition unless alkalinity, oxygen and redox potential are raised promptly. Aquaria are enclosed, biologically dense and intensively managed, so what they demonstrate is that the cascade chemistry is real, fast and reversible when buffering is lost; they do not establish the rate or scale at which the same reactions would proceed in open water.

The formation of colloids, as pH, redox potential, and zeta potential decrease, and their reversal through the increase of alkalinity, oxygen, and redox potential, are standard and well-characterised processes in water treatment. The novelty lies in applying that established chemistry to coastal and ocean systems, allowing for an understanding of the net effect in the same terms. The same chemistry will operate at scale in marine systems, but it has not been quantified. The net effect of increasing micro viscosity (the effective viscosity experienced at the scale of small organisms and particles, set by colloids and dissolved polymers rather than by the bulk water) reduces the swimming speed of planktonic larvae and small zooplankton and the water movement that larvae use for feeding ^{10,11}. The chemistry is also expected to control light attenuation and have the greatest impact on the smallest organisms. A change in chemistry could drive regime shifts and, under extreme conditions, such as eutrophication or a sudden rise in temperature, can trigger the same cascade reactions all the way to a dead zone.

Established. Rising carbon dioxide lowers seawater pH and carbonate ion concentration and reduces aragonite saturation; sensitive calcifiers show measurable responses once saturation falls far enough, and dissolved organic matter and inorganic colloids undergo pH-dependent charge neutralization, coagulation and flocculation. None of these reactions is in dispute in the chemical literature.

Demonstrated in engineered seawater systems. A fall in pH alters the charge state and aggregation behaviour of marine dissolved organic matter and of iron and other metal

species. The reactions are practiced daily at an industrial scale in desalination pretreatment, so the chemistry is established in seawater; what those settings cannot supply is the rate or the spatial scale at which the same reactions proceed in a mechanically mixed, effectively unbounded ocean.

We consider redox and zeta potentials to be as important as pH measurement, and in our experience of public aquaria, redox is measured continuously as a critical water quality parameter. Unfortunately, we could not locate any ocean transect surveys that have measured redox or zeta potential, so the reactions remain undemonstrated in open water because they have not been measured.

Hypothesised. That a declining pH raises the colloidal load and micro-viscosity by an amount that matters for plankton in the open unbounded ocean. It also follows that the water chemistry could impact feeding, swimming and the ability of the plankton to keep station and that these individual effects aggregate into a population-level ecological consequence. The chain as a whole has not been observed in the open ocean; its final link, that a modest rise in viscosity imposed at constant temperature slows the swimming and the feeding currents of planktonic larvae and zooplankton, has been confirmed under controlled laboratory conditions ^{10,11}, while the link from falling pH to a higher colloidal load and micro-viscosity remains to be demonstrated.

Scenario. A fourth level sits beyond the three above: the conditional projections of Sections 3.5 and 3.7, which follow from ecological theory rather than from measurement, and which are labelled as scenarios wherever they appear in this paper.

We regard these as testable predictions, set out with the observational signatures that would confirm or refute them in Section 3.4.

1.3 What are the consequences of ocean acidification?

Marine life has been decreasing due to anthropogenic factors for many decades. For the purpose of this report and to explore the ‘alignment’ of aragonite chemistry with biodiversity loss, we took 1850 as a starting point and an assumption that 100% of marine life was intact. We reviewed many reports on anthropogenic factors affecting ecosystems, and these included overexploitation, climate disruption, ecosystem destruction and pollution.

The purpose of measuring aragonite is to determine the stability of the marine water chemistry and the security of marine life. Given the impact that a change in pH exerts on water chemistry and quality, any discussion of the implications of a low aragonite index should also factor in the biology. Cold-water, high-latitude seas have passed the Ω 1.6 threshold, and the water is becoming more corrosive, which adds to the stress that marine life must endure.

The concentration of the divalent cations calcium and magnesium is central to coagulation and flocculation reactions as well as to aragonite saturation, and the records show that seawater has moved from a magnesium-to-calcium ratio of about 2, rising through the Tertiary Period to approximately 5.2 today (Section 2.4), a state it last held around the

Permian, a quarter of a billion years ago¹². Given that present global temperatures sit among the lowest of that same quarter of a billion years¹³, that the present magnesium-to-calcium ratio sits at the top of its range for the period, that the buffer capacity of surface seawater is falling⁷, and that the anthropogenic input exceeds 10 Gt of carbon per year¹⁴, we consider that the importance of the aragonite index and fragility of the marine ecosystem upon which all life on Earth depends may not have been fully appreciated.

2. Results

2.1 How aragonite saturation is measured

The aragonite saturation index can be determined by two different methods. The wet-chemistry method involves collecting a water sample and measuring its chemistry directly in the laboratory, where dissolved inorganic carbon is measured by coulometry and total alkalinity is determined by titration, with each measurement checked against certified reference materials; Ω is then calculated from these two measurements. The wet-chemistry method is the reference approach, but it is slow and yields discrete samples at set depths and times. The automated approach relies on instruments onboard ships or floats that continuously record the partial pressure of carbon dioxide and, in some cases, pH; the value of Ω is derived from these measurements. This method provides far more data, but the $p\text{CO}_2$ -based and wet-chemistry results are not fully consistent in cold water, where the dissociation constants are least certain^{15,16}, and in the records used here the $p\text{CO}_2$ -based value reads higher (Supplementary Note 1); it also samples only the surface of the water. The measurement does not reflect the exposure of marine life that migrates vertically or the change in aragonite with temperature below thermoclines in stratified waters.

The gridded products that map the whole ocean surface are built mainly from these automated records. This report takes the discrete wet-chemistry winter values as the preferred measure of the conditions marine life actually experiences, with the automated and mapped values providing spatial context and an upper bound.

2.2 The subpolar North Atlantic decline

The Iceland Sea yields the longest records. Ólafsson et al.¹⁷ observed a decrease in winter surface aragonite saturation of $0.0072 \pm 0.0007 \Omega$ per year between 1985 and 2008, from about 1.6 to about 1.5. The winter surface pH also decreased by 0.0024 ± 0.0002 units per year. The OSPAR Quality Status Report¹⁸ confirms and expands upon these discoveries, with an Iceland Sea trend of about 0.006Ω per year over 1985 to 2020, an Irminger Sea trend of about 0.012Ω per year over 1983 to 2020, and a Norwegian Sea trend of about 0.012Ω per year over 2002 to 2020 (a period that spans the 2014 change in measurement method; see Supplementary Note 1). These aragonite measurements are precise and detailed, and the downward trend is more evident when winter data are used on their own: in the Ólafsson analysis, the fitted trend explained 89% of the variance of the winter pH series, but only 38% when all seasons were combined. This decrease occurs because the change from one season to the next is as large as the underlying trend over several decades. The OSPAR assessment records Iceland Sea Ω at or around 1.5 in the 1980s, with seasonal minima reaching about 1.3 since 2015, while the Irminger Sea has only started to

reach values below 1.5 seasonally over the past decade. The 2014 change in the variable pair used to compute Ω introduces a step in the record, largest in cold water, but the trends reported here are taken from the long, internally consistent pre-2014 series, and no slope is fitted across the change, so the step does not generate the decline (Supplementary Notes 1 and 2).

The subpolar North Atlantic also holds the ocean's one persistent cold anomaly, the cold blob south of Iceland: the one region that has cooled while the rest of the ocean warmed, a pattern attributed to a weakening of the overturning circulation¹⁹. The cooling matters here because colder water holds more carbon dioxide and sits at lower aragonite saturation, so the cooling itself helps keep this region's saturation low; it is not incidental that the two longest carbonate chemistry records in this study sit inside that anomaly. The low saturation of this region therefore blends three contributions: the global anthropogenic carbon dioxide uptake that drives the long-term decline, the thermal effect of the cold anomaly itself (linked to the slowing overturning circulation, which is at least partly anthropogenic), and natural variability. We do not partition them here; the decadal downward trends track rising carbon dioxide, while the cold anomaly lowers the baseline on which those trends act. The anomaly extends well below the surface layer, and discrete casts resolve the vertical structure of aragonite saturation within it²⁰, unlike a mapped surface product. That vertical reach is relevant for the many plankton that migrate vertically each day, since they pass the daylight hours in colder, more corrosive water than no surface product records.

2.3 A global and multi-basin pattern

Ma et al.²¹ report a global surface Ω decline of 0.071 ± 0.006 per decade and a pH decline of 0.0166 ± 0.0010 per decade over 1982–2021, with the decline in aragonite saturation fastest at low latitudes. This result was achieved by reconstructing four decades of surface chemistry. Poleward of about 60° north and south, many areas have surface Ω below 1.5, while between 40°N and 40°S, the surface stays above 2.0²². The biological point is directly conveyed in Figure 1. The measured surface Ω in winter is already at or within the marginal zone and continues to decline, specifically in the Iceland Sea, the subarctic North Pacific, and the Southern Ocean. The Irminger Sea sits slightly above the marginal zone (Supplementary Note 1). The exception is the subantarctic South Pacific site at Munida, which has an Ω of approximately 2.0 in water that is about 9°C warmer than the polar sites. This reading is consistent with the relation between saturation and temperature across the neighbouring mapped cells of all four sites, which is approximately 0.08 Ω per degree Celsius (correlation 0.94, Supplementary Note 1). The Southern Ocean is the front line: a shallow aragonite saturation horizon is projected to emerge suddenly this century^{23,24}, and shell dissolution is already observable in polar and upwelling waters^{25,26}; the surface pH there is falling faster than the global average, although the saturation state itself declines more slowly than at low latitudes²¹; and live pteropods have been observed to dissolve as Ω approaches 1.0²⁷. The subtropical gyres, by contrast, remain well above the threshold: Station ALOHA, for instance, remains far above the threshold (Figure 4), so the gyres serve here as high-saturation reference sites rather than sites of undersaturation concern. The discrete sites are point measurements that together cover a negligible fraction of the ocean surface (Supplementary Note 1); rather than being extrapolated to whole basins, they

ground truth the basin-scale OceanSODA-ETHZ field, which maps the entire surface and agrees with the discrete winter casts to within about 0.01–0.12 Ω where both exist, once the discrete values are computed with the partial pressure of carbon dioxide taken at in-situ temperature (Supplementary Note 1). The regional winter picture therefore rests on the gridded product checked against the point data, not on the sites alone.

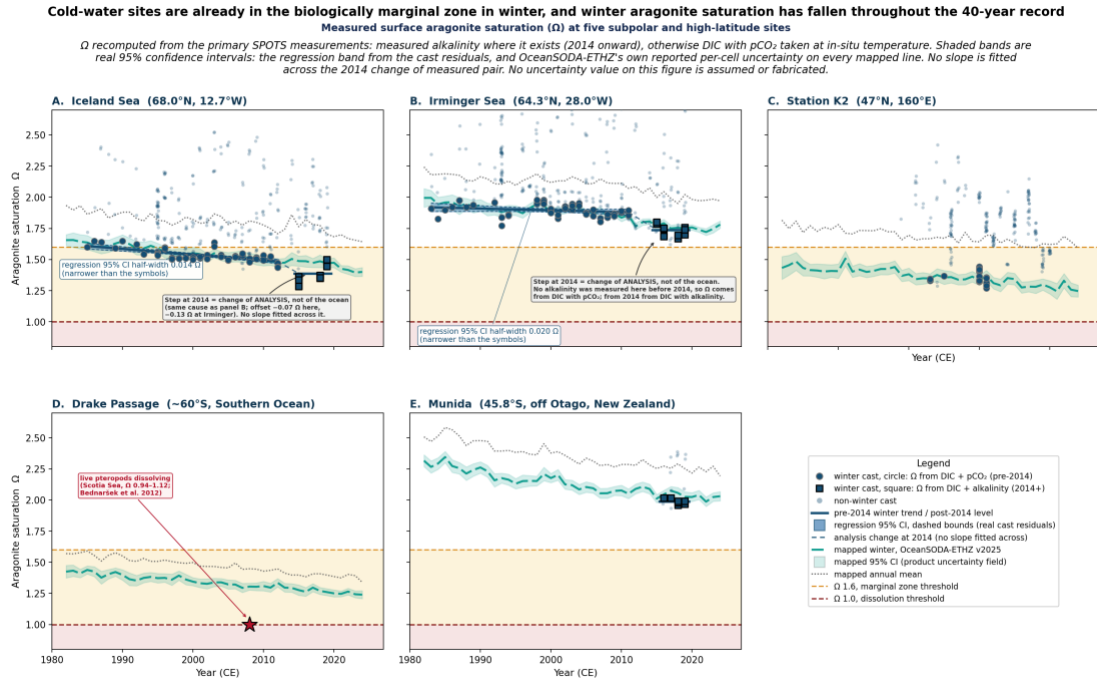


Figure 1. Measured surface aragonite saturation (Ω) at five subpolar and high-latitude sites: Iceland Sea (A), Irminger Sea (B), Station K2 (C), Drake Passage (D), and Munida (E). Symbols: discrete winter surface casts, drawn by the variable pair used to compute Ω , so that the analytical split is visible at a glance. Circles with grey edges for the pre-2014 casts computed from dissolved inorganic carbon with the partial pressure of carbon dioxide; squares with black edges for the 2014 onward casts computed from dissolved inorganic carbon with measured alkalinity (Supplementary Notes 1 and 2); dark blue solid, the pre-2014 winter regression trend with its 95% confidence band computed from the cast residuals (dashed bounds; the band half-width is 0.014 Ω at the Iceland Sea and 0.020 Ω at the Irminger Sea, narrower than the plotted symbols, as annotated on the figure) and the post-2014 mean level; teal dashed, the mapped winter mean with its 95% confidence band taken from OceanSODA-ETHZ's own per-cell reported uncertainty; grey dotted, the mapped annual mean; faint small circles, non-winter casts. Winter is the January–March mean in the Northern Hemisphere and the July–September mean in the Southern, so each panel shows its own hemisphere's cold season. Methods and cast counts, Supplementary Note 1. The mapped winter values are least reliable in these stormy, under-sampled seasons (Supplementary Note 2); the discrete winter casts shown alongside them anchor the winter picture where they exist.

Every winter cast is plotted. Ω for the discrete casts was recomputed for this revision directly from the primary SPOTS measurements, using measured total alkalinity where it exists (2014 onwards) and otherwise dissolved inorganic carbon with the partial pressure of carbon dioxide taken at in-situ temperature; Supplementary Note 1 explains why that last point matters. Fitted lines are the pre-2014 winter trend with its 95% confidence band from the cast residuals and the post-2014 casts as their level. No slope is fitted across 2014 because before that date no alkalinity was measured at these stations, and Ω must be

computed from a different pair, which introduces a small offset rather than an oceanographic change. On this basis the Iceland Sea winter record declines at $-0.0046 \pm 0.0007 \Omega$ per year ($n=57$, $p<10^{-8}$), while the Irminger Sea is statistically flat over the same period ($-0.0014 \pm 0.0009 \Omega$ per year, $n=51$, $p=0.13$), the contrast the text describes. The flat Irminger record also demonstrates that the fitting method does not create artificial declines; when the same procedure is applied to neighbouring water, it shows no trend in areas where the data do not support one, indicating that the significant decline in the Iceland Sea is due to the water itself rather than the fitting method.

The step visible in the discrete casts at 2014 is a change in the analysis and not in the ocean: Ω is computed from a different measured pair before and after that date (dissolved inorganic carbon with the partial pressure of carbon dioxide, then dissolved inorganic carbon with total alkalinity), and the two pairs yield inconsistent Ω , with the largest disagreement in cold water^{15,16}. Station K2's winter casts fall in only four distinct years between 2003 and 2011, two-thirds of them in a single year, and Munida's seven winter casts span 2016–2019 only; neither supports a fitted trend, so the long-term rate at those sites also comes from the mapped product^{28,29}. The Drake Passage has no sustained discrete DIC+TA record (only underway $p\text{CO}_2$), so only the mapped product is shown there.

One property of these thresholds matters for the comparison being made. These thresholds represent point-exposure relationships between the saturation an organism encounters and its response, which are established from animals sampled in their natural water environments rather than from annual climatologies. The correct quantity to compare them against is therefore the saturation actually experienced, which in these basins is the winter minimum, and comparing them against an annual mean understates exposure. That is the reporting error this paper identifies, and it is why crossing a threshold in winter is a statement about the conditions the animals meet rather than an artefact of choosing the lowest number available.

Figure 2 places the winter field against ecological reference thresholds rather than a single global value. Each cell of the OceanSODA-ETHZ winter field (February in the Northern Hemisphere and August in the Southern) is compared with the threshold of its own latitude zone: Ω 1.6 for polar and subpolar waters poleward of 50° , the marginal-zone value set out in Section 2.4 from the bivalve and pteropod response literature and consistent with the low saturation already measured in these waters²⁵; Ω 3.3 for tropical waters equatorward of 30° ³⁰; and an interpolated Ω 2.4 for the temperate band between them. The temperate value is the midpoint of the two anchored thresholds, a stated choice rather than a measured one: the temperate band carries a mixed calcifier community, with cold-water aragonite builders at its poleward edge and the warm-water end of the reef fauna at its equatorward edge, so no single measured threshold exists for it, and the midpoint is the parsimonious interpolation. Because it is a choice, its influence on the result is tested directly below. The tropical value of Ω 3.3 is the saturation at which net carbonate accretion on reefs approaches zero (Section 2.4) and lies within the band that the reef habitat classification of Guinotte et al.³⁰ rates as marginal for coral reef growth. Applying a reef-derived value to open-ocean cells is itself a stated choice, and it is tested in the same way: zonation scheme C restricts the reef threshold to the tropics proper, within 23.5° of the equator, and the 37–52% range reported below already contains the result of that

restriction. On this basis, about 44% of the mapped ocean surface is already below its zonal threshold in winter (mean of the 2022–2024 winters), and extending each cell's own 43-year winter trend, an estimated 49% will cross by 2035, 57% by 2045, and 65% by 2055.

Because the zone boundaries and thresholds are stated choices rather than measurements, we tested how much the result depends on them. Holding the coral threshold to the true tropics (within 23.5°) and judging the subtropics against the temperate value of Ω 2.4 lowers the present figure to 37%, because subtropical water at Ω 2.9–3.2 passes a 2.4 test but fails a 3.3 one; moving the boundaries poleward to 55° and 35° raises it to 52%, because water between the old and new boundaries is then judged against the stricter threshold of the next zone.

Under every scheme tested, the figure remains between an estimated 37% and 52%, so the qualitative conclusion, that roughly two-fifths to one-half of the winter ocean surface is at or approaching its ecological reference threshold, holds true across all zonation schemes, although the precise percentage varies within that 15-point range depending on the placement of the boundaries. The median fitted winter trend, -0.072Ω per decade, matches the published global rate ²¹. The below-threshold water is not randomly distributed: at every latitude it coincides with the productive upwelling and high-nutrient zones, the eastern equatorial Pacific, the Arabian Sea, the eastern boundary systems and the subpolar basins, while the unproductive gyre cores remain above threshold. This is one mechanism, not a coincidence: upwelled deep-water supplies both the nutrients that make these regions productive and the respired carbon dioxide that lowers their saturation, so the waters that sustain the food web are, everywhere, the closest to their biological limits. In the open equatorial upwelling, water below Ω 3.3 is an indicator of this coupling rather than direct harm to reefs, which are coastal; the direct biological pressure falls where such water meets reef and shelf habitat.

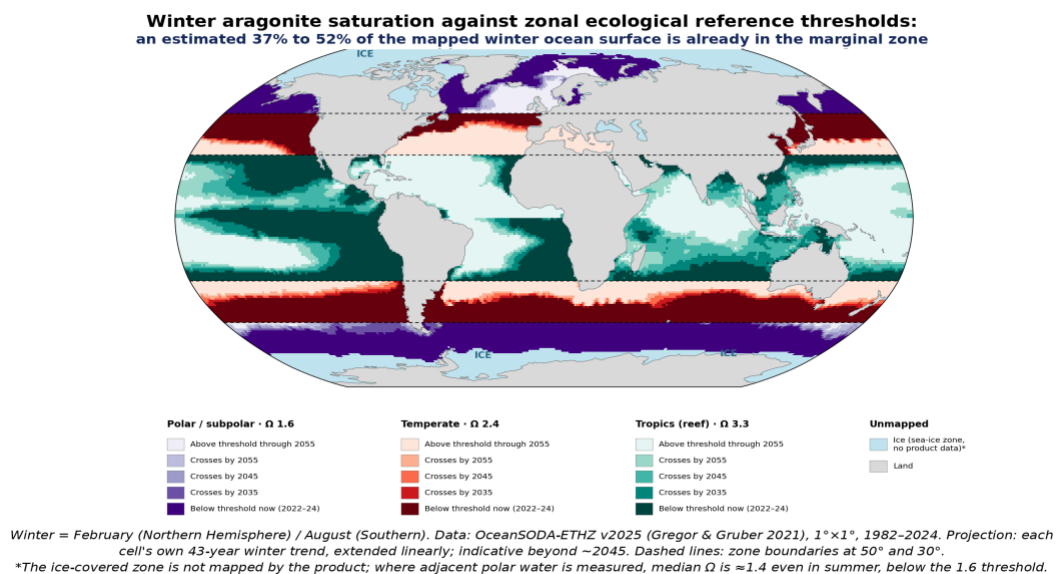


Figure 2. Winter surface aragonite saturation against zonal ecological reference thresholds. The colour family gives the zone and its threshold: purple, polar and subpolar ²⁵; red, temperate (Ω 2.4, stated scheme); and blue-green, tropical reef ³⁰. Shading gives status from above the threshold through 2055 (palest) to

below the threshold in the 2022–2024 winters (darkest), with intermediate shades marking the projected decade of crossing (2035, 2045, 2055) from each cell's own 1982–2024 winter trend³¹. In the Northern Hemisphere, winter occurs in February, while in the Southern Hemisphere, it occurs in August. Pale blue marks the sea-ice zone that the product does not map; where adjacent polar water is measured, its median Ω is about 1.4, even in summer. Projections are linear and indicative beyond about 2045.

The regions where winter saturation is lowest coincide with the ocean's main high-nutrient, low-chlorophyll (HNLC) waters. That overlap does not by itself show that either condition produces the other. The two patterns share a physical setting, and two features of that setting generate both of them at once, which is why they are found in the same places. The first of these is the temperature of the water. These waters are cold, and cold water holds more dissolved carbon dioxide and has a lower carbonate ion concentration, so its aragonite saturation is naturally low whether or not deep water is reaching the surface; the same low temperature also slows primary production, which is part of the reason why the nutrients are left unused. The second factor is the supply of water from depth, which reaches the surface through large-scale divergent upwelling in the Southern Ocean and along the equatorial Pacific, as well as through deep winter mixing instead of upwelling in the subarctic North Pacific. That deep water is rich in nutrients, and it also carries organic matter that decomposes to release carbon dioxide, so the same supply that makes these waters high in nutrients makes them low in saturation. Each factor therefore acts on both patterns, and neither pattern needs to be causing the other for the two maps to look alike.

If the cascade set out in Section 1.2 operates in these waters, a chain reaction follows: pH goes down, oxygen levels fall, redox and zeta potential fall, and the water becomes slightly more turbid; that sequence is hypothesised, not measured. The nutrients are not drawn down because primary production is limited by deficient iron³², with grazing, low temperature and, in the Southern Ocean, low winter light adding further regulation; this unused-nutrient, low-chlorophyll state is what defines an HNLC region³³. The chemistry is not passive in this pattern: falling pH further reduces the bioavailability of what little iron there is, so declining saturation is expected to deepen and extend the HNLC state (Section 3.4). Figure 3 maps the HNLC zones, defined as areas where the surface nitrate-to-chlorophyll ratio exceeds 17, over the winter saturation field: the two persistent HNLC cores, located in the Southern Ocean and the subarctic North Pacific, lie within the low saturation zone. The eastern equatorial Pacific is rich in nutrients but warm, which keeps its saturation high, while seasonally productive high-latitude seas, such as the subpolar North Atlantic, carry high nitrate levels without being classified as HNLC because their blooms consume the nitrates.

The waters that feed the marine food web are also the closest to their biological limits
 Winter aragonite saturation (colour) and the productive HNLC zones (green outline); the two overlap almost everywhere

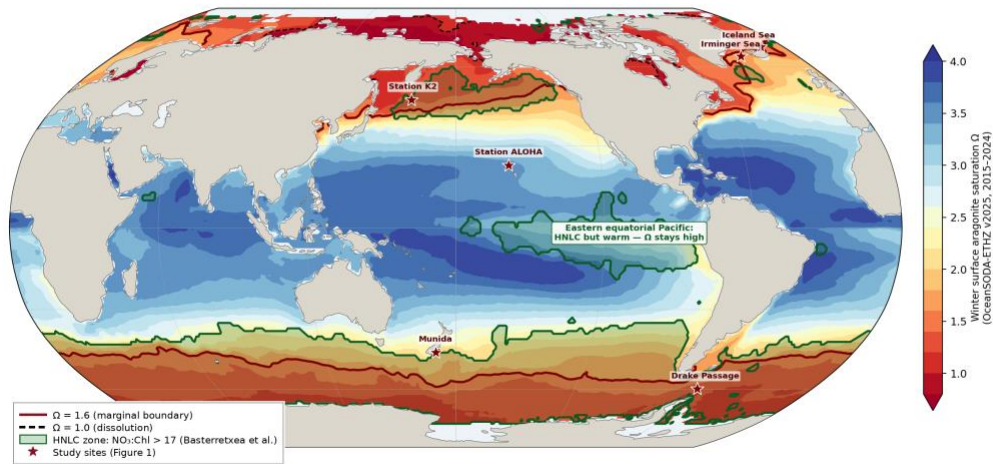
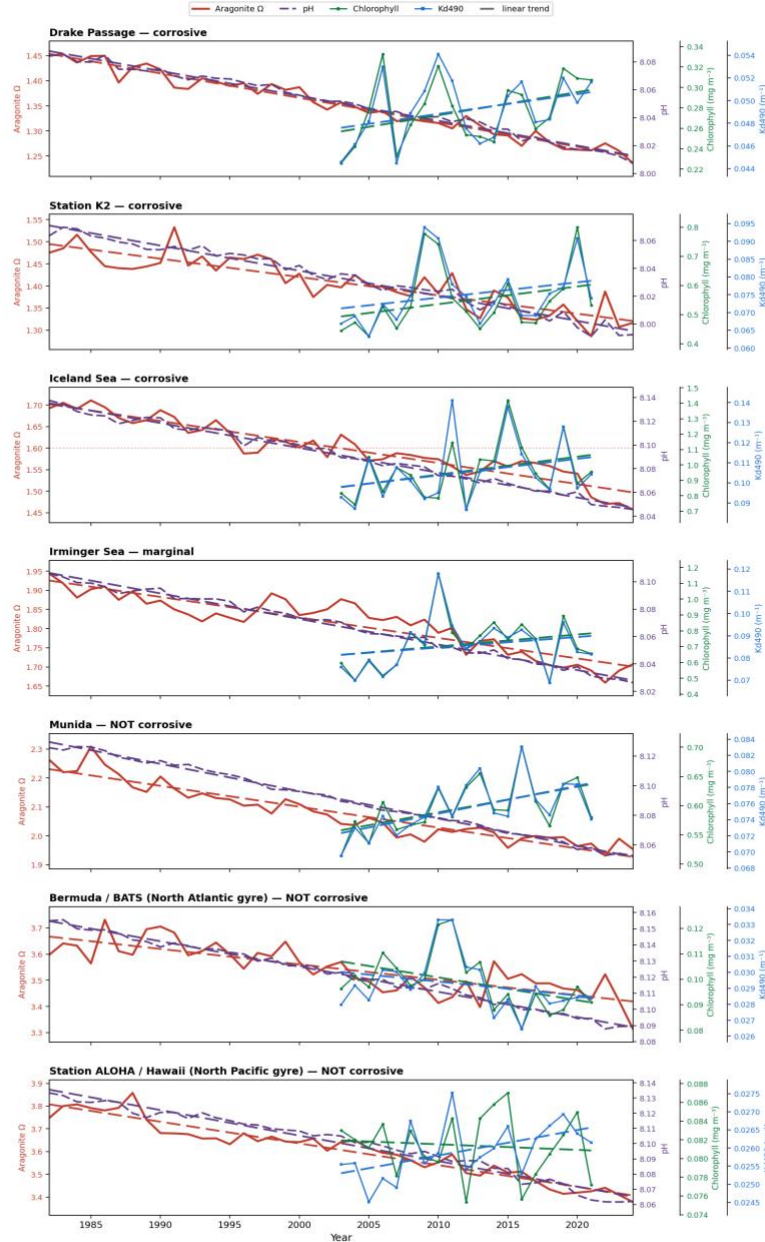


Figure 3. Winter surface aragonite saturation and high-nutrient, low-chlorophyll (HNLC) zones. Colour, winter Ω ³¹; dark red contour, $\Omega = 1.6$; dashed contour, $\Omega = 1.0$. The HNLC zones, indicated in green, are areas where the ratio of surface nitrate to chlorophyll exceeds 17, based on data from the World Ocean Atlas 2023 nitrate ³⁴ and MODIS-Aqua chlorophyll (JRC Global Marine Information System ³⁵). Nitrate is abundant there, but iron scarcity limits phytoplankton ³². The stars mark the locations of the study sites.

The aragonite decline is temporal as well as spatial, and it is general. Figure 4 compares seven sites, from the most corrosive (Drake Passage, Station K2 and the Iceland Sea, which are at or below $\Omega = 1.6$) through the marginal Irminger Sea to the non-corrosive Munida and the two subtropical-gyre stations, Bermuda in the North Atlantic and Station ALOHA off Hawaii in the North Pacific. Winter aragonite saturation and pH fall at every site from 1982 to 2024. The same multi-decadal fall in pH is recorded at the sustained ESTOC station in the eastern North Atlantic ³⁶. The regional satellite records over 2003 to 2021 show a rise in light attenuation at six of the seven sites and a fall at Bermuda, with chlorophyll moving in the same direction as attenuation at every site (Figure 4; quantified in Figure 5). At the four cold sites and at Munida, the change in chlorophyll is larger than the change in attenuation; however, none of these ten trends is statistically significant when autocorrelation is considered, indicating that at the productive sites, the observed darkening is merely the optical expression of a weak greening and does not provide any additional information. These satellite trends explain only a small part of the year-to-year variation (the coefficient of determination R^2 ranges from about 0.02 to 0.37), so they indicate direction rather than tight relationships. Both K_d490 and chlorophyll are derived from the same blue-green reflectance ratio, meaning they largely represent the same signal, and the darkening provides additional information only where it differs from chlorophyll. Station ALOHA is the one site where the two separate light attenuations rose by 3.1% over the period ($p = 0.002$) while chlorophyll did not change (+2.1%, $p = 0.61$), and the chlorophyll change would account for about one-fifth of the observed rise in attenuation. Section 3.4 examines what that pattern can and cannot show.

Do acidifying waters show the optical signature of a dissolved organic colloidal suspension?

Winter Ω and pH (1982–2024) with regional Kd490 and chlorophyll (2003–2021) at seven sites.
Station ALOHA (bottom) darkens while chlorophyll stays flat, a potential signature of such a suspension.



Each y-axis is scaled to its own series, so the amplitude of every trend is visible; absolute levels differ between panels and are not directly comparable. Bermuda and Hawaii are the two subtropical-gyre sites; both keep a high aragonite Ω and clear, oligotrophic water, and both acidify. Dashed red line marks $\Omega = 1.6$ where in range.

Figure 4. The seven sites are ordered from most to least corrosive based on winter aragonite Ω and pH³¹, along with regional darkening (Kd490) and greening (chlorophyll) data from satellite observations (2003–2021; MODIS-Aqua; NOAA CoastWatch). Red, Ω (dashed red, $\Omega = 1.6$); purple, pH; green, chlorophyll; blue, Kd490; dashed lines, linear trends. Bermuda and Station ALOHA are the two subtropical gyre sites; both hold a high aragonite Ω in clear, oligotrophic water, and both acidify. Each y-axis is scaled to its own series, so absolute levels differ between panels and are not directly comparable. Fitted trends are shown here without confidence intervals for legibility; the corresponding slopes, their 95% confidence

intervals, p-values and the product's reported per-value measurement uncertainty are tabulated for every site in Supplementary Table 3 (Methods). All fourteen Ω and pH trends are significant at $p < 10^{-10}$. *The chlorophyll and Kd490 series carry no propagated uncertainty and should be read as indicative of direction rather than as quantified rates (Methods).*

2.4 Biological thresholds and calcifier vulnerability

The link between reduced calcification and falling aragonite saturation has been recognized since Kleypas et al.³⁷, and the dissolution point, $\Omega = 1.0$, is textbook and uncontested. The biological marginal zone lies around Ω 1.2 to 1.5, where bivalve larvae exhibit a response at Ω 1.2–1.5³⁸, and pteropods are considered to be negatively impacted by Ω below 1.5³⁹. The California Current System, a strong upwelling region, shows, in a model hindcast, the exposure of shelled pteropods to damaging water (Ω below 1.5) rising from 9% to 49% between 1984 and 2019³⁹. The rising exposure is matched by measured biological response in the same system: 53% of pteropods sampled onshore show severe shell dissolution, against 24% offshore, in direct proportion to the undersaturated fraction of the water column²⁶, and when the same corrosive upwelled water entered Pacific Northwest oyster hatcheries from 2007, larval survival tracked the aragonite saturation of the intake water so closely that the industry now buffers and treats its seawater before use⁴⁰. Consequently, we regard Ω 1.6 as a marginal zone for sensitive aragonite calcifiers. Reef corals decline at a much higher Ω , with net carbonate accretion approaching zero around Ω 3.3⁴¹, in reef water that, under healthy, actively calcifying conditions, can sit as high as Ω 3.7 to 4.4⁴²; about a quarter of the sea-floor calcifiers studied already live in water undersaturated for their own skeletal mineral⁴³; and some contend that Ω is merely a proxy for the true driver, the ratio of bicarbonate to hydrogen ions⁴⁴.

The modern ocean is a high-magnesium "aragonite sea" when compared to the geological record. Its molar ratio of magnesium to calcium is approximately 5.2, which is near the highest of the past 250 million years^{45,46}. Its calcium content, at approximately 10 mmol kg⁻¹, is also less than half of the Cretaceous value⁴⁷. Calcite-building animals, as a result, produce more soluble skeletons than their counterparts did in previous low-ratio seas, and they likely expend more energy in the process of constructing them. Consequently, calcifiers today build their skeletons in seawater whose magnesium-to-calcium ratio sits at the top of its range for the last quarter of a billion years⁴⁷. The order of vulnerability is determined by mineralogy. Coralline algae and sea urchins utilise high-magnesium calcite, which dissolves more readily than aragonite, and aragonite dissolves more readily than low-magnesium calcite^{48,49}. Consequently, the initial hypothesis is that high-magnesium calcite plankton will fail first, followed by aragonite builders, with calcite-shelled plankton most resistant and expected to fail last. The order is a reasonable starting point, offered as a mechanistic expectation rather than a fitted relationship, although biology is not that tidy; solubility varies, physiology frequently overrides mineralogy, and certain species calcify more under elevated carbon dioxide⁴⁹. Physiological adaptations and species-specific responses can, however, modify this generalised mineralogical vulnerability, so the actual sequence of impacts may not strictly follow this order.

A threshold value alone cannot capture the entire biological risk. The extent of harm depends on the duration and life stage of the animal, as well as on how far Ω falls.

Bednaršek et al. (2023)²⁵ project the dissolution of pteropod shells from the saturation state encountered, using thresholds that account for both the magnitude and the duration of exposure. They make their projections for the early and middle life stages, whose demographics govern the population. Consequently, two waters that achieve the same winter minimum are not equivalent: exposure during reproduction, settlement, or moulting is more detrimental than exposure at a tolerant stage, and a month below the threshold is more detrimental than a week. This is another reason why the seasonal minimum and the duration of the corrosive season are more significant than the annual mean.

The eastern Bering Sea demonstrates that duration and season are not abstract concepts. Its shelf ranks among the most productive and corrosive seas because the decomposition of its large summer production, together with river runoff, removes carbonate from the bottom water⁵⁰. Pilcher et al.⁵¹ report, from a model hindcast of the shelf covering 1970 to 2022, that undersaturation is seasonal and that its extent is expanding. The bottom water acidifies at a rate that is approximately twice as rapid as the surface water, and the pH of the bottom water has decreased to approximately 7.8, which is the threshold at which juvenile crabs begin to experience symptoms in the laboratory. The shelf area carrying bottom water below that value grew from 5 to 10% before 2005 to more than half of the shelf by 2022. These periods are prolonged; the surface water on the inner shelf is undersaturated for approximately half of the year, and the bottom water is at or below the dissolution point for the majority of the middle and outer shelf^{52,53}. The results of the laboratory work are consistent with these field observations. A formal risk assessment finds that the fishery-dependent regions of southern Alaska carry the highest ocean acidification risk within that state⁵⁴.

Juvenile red king and Tanner crabs reared at reduced pH exhibit lower survival and growth, and red king crabs have a lower condition index⁵⁵, and juvenile snow crabs held at pH 7.5 suffer approximately 40% higher mortality than at present-day pH⁵⁶. Subsequently, the snow crab population in the Bering Sea collapsed, with the loss of over ten billion animals between 2018 and 2021. This collapse is mostly due to a marine heatwave and starvation, not acidification⁵⁷. However, the region's high acidity and low alkalinity mean that acidification is still a major and growing stressor for these populations. Nevertheless, it is a clear example of a productive cold-water fishery that has been pushed beyond a biological limit in a sea where acidification has already reached levels that have been shown to harm the same animals.

2.5 Calcifying plankton

Planktonic foraminifera provide supporting evidence; comparing shells from an 1875 Challenger station with 2011 Tara Oceans material from the eastern equatorial Pacific, Fox et al.⁵⁸ found every modern specimen thinner, by up to 76% in *Neogloboquadrina dutertrei* and by up to about 30% in *Globigerinoides ruber*. This observation is based on a single before-and-after comparison, and since foraminifera primarily build low-magnesium calcite, which is one of the most acidification-resistant materials, a signal indicating carbonate stress suggests that even the hardest plankton are affected.

Coccolithophore occurrence has risen sharply: a multi-decadal Continuous Plankton Recorder series shows North Atlantic occurrence increasing roughly tenfold, from about 2% to over 20% of samples, between 1965 and 2010, tentatively linked to rising carbon dioxide ⁵⁹. Yet where calcification itself has been measured over decades, in the Gulf of Maine GNATS series, it fell from about 0.4 to about 0.2 mg C per cubic metre per day between 2001 and 2018, the years over which the rates were sampled ^{60,61}. The authors attribute the parallel fall in primary production to a change in the water entering the gulf rather than to acidification, and they report that aragonite saturation rose over the seven years of carbonate chemistry data (2012 to 2018) while pH showed no significant trend, so we cite the record as evidence that calcification can fall steeply on decadal timescales, not as an acidification signal. One caveat should be noted, and it is marked in Table 1a: that decline happened while aragonite saturation was rising, driven by a shift in the water entering the gulf (warm, salty North Atlantic Slope Water displacing the cooler, fresher Labrador Slope Water) and by falling productivity, so it is not clear evidence of acidification harming calcifiers, and, as a single record, it establishes no global trend. Tropical coccolithophore calcification has been found to be independent of carbonate chemistry ⁶², and some strains calcify more under high carbon dioxide ⁶³, producing larger and heavier coccoliths at a lower growth rate.

2.6 The measured declines

Table 1a. Documented measured declines in marine animal life and calcifiers since 1850, in time order. Every row reports a published, referenced measurement or indicator; rows marked “indicator” are calcification or habitat-suitability measures rather than abundance counts, and “contested” figures are cited with their counter-evidence. The final column is not a property of that row’s group: it provides the whole-ocean biocurve value (Table 1b, Section 3.6) at the date the row’s record ends. Because rows are ordered by the start of each record, these values do not fall monotonically down the table.

Table 1a. Documented measured declines: every row is a published measurement.					
Species / group	Approx. location	Time frame	Documented change (loss of nature)	Biocurve value at end of record (% of 1850).†	Reference
Predatory fish (trophic level ≥3.5)	Global	1880–2007	–67% (two-thirds)	45	⁶⁴ ; cf. ⁶⁵
Planktonic foraminifera (calcite)	Global (Challenger→Tara)	1870s–2011	Shells thinner in every modern specimen: 48–76% in <i>N. dutertrei</i> , up to ~30% in <i>G. ruber</i> (4 scans, eastern equatorial Pacific, 1875 vs 2011).	43	⁵⁸
Coral reefs (coverage, biodiversity, catch per effort)	Global tropics	1950s–2015	Cover ~–50%; biodiversity –63%; CPUE –60%	40	⁶⁶
Exploited fish biomass (10 g–100 kg)	Large marine ecosystems	pre-industrial–1990s	To ~47% of pre-industrial (roughly halved)	52	⁶⁷

Table 1a. Documented measured declines: every row is a published measurement.					
Species / group	Approx. location	Time frame	Documented change (loss of nature)	Biocurve value at end of record (% of 1850).†	Reference
Antarctic krill	SW Atlantic, Southern Ocean	1976–2003	38–81% per decade reported; contested (no significant decline 1976–2016).	48	68,69; cf. 70
Krill (euphausiids)	North Atlantic (subpolar)	1958–2017	–50%	39	71
Coccolithophore calcification	Gulf of Maine	2001–2018	0.4→0.2 mg C m ^{–3} d ^{–1} , ~–50% over 2001–2018 (coincides with a water-mass change and rising Ω ; not attributed to acidification)	39	60, 61
Pteropods (habitat suitability)	California Current System	1984–2019	Modelled exposure to $\Omega < 1.5$ water rose 9% → 49% (hindcast indicator)	38	39
Wild marine mammals	Global	1850–2020	To ~30% of 1850 (~70%)	37	72
Marine vertebrates (Living Planet Index)	Global	1970–2020	–56%, 1970–2020 (contested)	37	73; cf. 74
Snow crab	E. Bering Sea	2018–2021	> 10 billion lost (starvation after marine heatwave)	37	57
Oceanic sharks & rays	Global	1970–2018	–71%	39	75
Seabirds	Global	1950–2010	–70% (monitored populations)	43	76
Coastal & estuarine systems	12 estuarine/coastal systems	historical–late 20th C	>90% of formerly important species depleted, >65% of seagrass and wetland habitat destroyed; oyster reefs –85% globally ⁷⁷ .	49	78

† The biocurve value is the weighted mean of the fraction of 1850 remaining across all documented marine animal groups (Supplementary Note 3), plotted in Figure 6. The bio curve is an interpretive synthesis of separately documented declines, not a direct census, and every value derived from it carries the uncertainty stated here. The present value ranges from approximately 35% to 54% of the 1850 level: the lower end has a greater impact on apex predators and structural and calcifying groups, while the upper end reflects total biomass, in which the dominant zooplankton pool has declined by about half rather than by the two-thirds to three-quarters recorded for apex and structural groups. Groups that held up or increased (forage fish, jellyfish) are included at a low weight. Rows marked “indicator” are calcification or habitat-suitability measures; “contested” figures are cited with their counter-evidence and discussed in Section 2.6.

Table 1a presents the individual declines in chronological order, maintaining a cautious reading in every case, although some figures are disputed. One example is the marine Living Planet Index, which fell by 56% between 1970 and 2020 in the 2024 report, against 49% for 1970–2012 in the earlier assessment ⁷³. Leung et al. ⁷⁴ show that a few fast-declining populations can pull such an index down, so trimming a few extreme values changes the result, while Loreau et al. ⁷⁹ agree that many populations are still in real decline. While some earlier estimates, such as the reported 74% drop in tuna and mackerel ⁷³, have been criticised as overstated against formal stock assessments and are not used

here, a synthesis of formal stock assessments places the decline in the adult biomass of tunas and their relatives at 52% over 1954 to 2006 ⁸⁰.

Wild marine mammals are down by about 70%, to roughly 30% of their 1850 level ⁷², against a total animal biomass on Earth of about 2 Gt of carbon, most of it marine ⁸¹. Krill in the North Atlantic have roughly halved since 1958 ⁷¹, mainly through warming and a squeezed habitat rather than acidification. A larger fall in the southwest Atlantic sector since the 1970s, reported as 38 to 81% per decade ⁶⁸, is challenged as partly a sampling artefact ^{70,82}, although a circumpolar reanalysis still finds a decline in the Atlantic sector ⁸³; a later analysis by the same group reports a southward contraction of the range together with a fall of about 70% in numerical density between the 1976 to 1995 and 1996 to 2016 sampling eras ⁶⁹. The North Atlantic figure matters here because that same subpolar water holds two of our four aragonite sites, even though its immediate cause is physical, not chemical.

The often-quoted 90% loss of large predatory fish ⁶⁵ is widely considered too high ^{84–86}. A steadier figure comes from a global biomass model: fishing has cut the biomass of exploited marine fish by roughly half since pre-industrial times, leaving about 47% of the pre-industrial biomass across large marine ecosystems by the time of peak catch ⁶⁷, without relying on disputed methods. What survives every one of these arguments is the point that matters here. Biomass models, stock assessments, mammal counts, and plankton records all indicate a clear and well-documented decline in marine animal life, even though the exact size of each loss remains disputed.

2.7 Drivers of the early decline and why they matter here

The main losses fall on the large animals, whales; big fish, sharks and rays; seabirds; and the calcifiers and coastal systems. Direct human pressure, rather than acidification, caused most of these losses. They came from industrial fishing, whaling, bottom trawling, and the destruction of coastal habitats, above all the loss of coral reefs, oyster reefs, and mangroves, together with widespread pollution, all documented at a large scale in the literature cited above. Removing the largest fish also weakens the ocean's biological carbon pump, because the carbon held in their bodies and sinking waste no longer reaches the deep sea ^{87,88}. Coral reefs support a large share of all marine species, and since the 1950s their living cover has halved; at least 63% of reef-associated biodiversity has declined, and catch per unit effort has fallen by 60%, so that the capacity of reefs to provide ecosystem services is down by about half ⁶⁶. The bio-curve weights these top and habitat-building groups most heavily, so its importance-weighted end (about 35%) sits well below its tonnage-weighted end (about 54%): the ocean holds more raw weight of life than working structure. The inclusion of coastal and estuarine declines, while important for total marine life, draws on highly degraded compartments subject to far more intense direct pressures than the open subpolar basins that are the focus of the chemical analysis, so these figures should not be read as representative of the open ocean.

Acidification is now beginning to stress marine life in the most productive oceans, especially the aragonite calcifiers. The cold Southern Ocean already shows these effects: live pteropod shells dissolve where upwelled water brings the saturation state to about 1.0

²⁷, and in the California Current a model hindcast puts the exposure of pteropods to water below Ω 1.5 at 49% by 2019, against 9% in 1984 ³⁹.

2.8 Data scarcity and a living baseline

One feature of this subject is how little it is monitored; there appears to be no sustained global census of marine animal abundance. Coccolithophore calcification has essentially one multi-decade record among a few thousand scattered spot measurements ^{60,61}. The cold-water carbonate chemistry rests on a handful of time-series stations. This scarcity is not a side detail; it is central to the argument. Judgements about the ocean's health are being made on a narrow evidence base. The few independent records that do exist, in chemistry, calcification and animal abundance, nearly all move in the same downward direction, which is easy to see, but we sometimes fail to grasp the consequences.

Human memory is one of the few direct windows onto the richer ocean of the past. People who fished or lived from shores, reefs and fishing grounds a few decades ago recall far more fish, seabirds and shellfish than are found in the same places today, which is the corrective to what Pauly ⁸⁹ called the shifting-baseline syndrome, in which each generation takes the depleted state it inherits as normal. Memory is not a substitute for measurement, but here it agrees in direction with the sparse instrument data, and our field impressions are set out separately in Supplementary Note 4 as an anecdote. Quantified evidence points the same way: in the North Sea nearly two-thirds of fish species have shifted in latitude or depth as the sea warmed⁹⁰, some populations are growing to a smaller maximum size⁹¹, and fishing pressure drove earlier maturation in the heavily exploited northern cod stock⁹².

The bio-curve is our attempt to set a baseline against which change can be measured. What prompted it was a set of field impressions gathered in the tropical North Atlantic, together with the near absence of surface microzooplankton subsequently recorded along a zonal transect ⁹³; the impressions are reported in Supplementary Note 4 and form no part of the index.

3. Discussion

3.1 The measured chemistry

Aragonite Ω and pH are falling wherever they have been observed, and in every cold, productive basin, the water has already reached the ecological reference threshold. Organisms meet that threshold sooner than the maps suggest, because the maps report annual means. The scale is quantified in Figure 2: about 44% of the mapped ocean surface is already below its zonal ecological reference threshold in winter, a figure that holds between 37% and 52% under every zonation scheme tested, and on each cell's own 43-year measured trend, an estimated 49% crosses by 2035, 57% by 2045 and 65% by 2055. More than half of the winter ocean surface would then be below the reference thresholds for the calcifying base of its own food web within two decades.

3.2 The documented declines and the acidification signal

The documented declines of Section 2.6 and the measured chemistry of Sections 2.1 to 2.3 are independent records, and the first is largely older than the second. Fishing, whaling, habitat destruction and pollution had already removed a large part of marine animal life before the carbonate chemistry began to shift, so acidification is not acting on a resilient, diverse system with capacity to spare; it is acting on communities already thinned by other pressures. The measured record can show where and how fast the chemistry is moving and which groups are known to respond at the saturation levels now reached, and the pteropod, oyster, and crab evidence in Section 2.4 shows that response in the field and in the laboratory. It cannot yet show a basin-scale population change attributable to acidification alone, and this paper does not claim one. What it does show is that the water in which the remaining calcifying base of the cold-water food web lives is already at or below the reference thresholds derived from those experiments and that its exposure is longer and deeper than the annual means report.

3.3 Hypothesis and synthesis: scope and status

Nothing in Sections 3.3 to 3.7 is a measured result of this study, except the satellite optical trends of Figure 5, which are reported here because their interpretation belongs to the hypothesis rather than to the measured chemistry. Section 3.4 sets out the chemical cascade as a hypothesis in the tiers defined in Section 1.2, along with the observational signatures that would confirm or refute the hypothesis. Sections 3.5 to 3.7 build an interpretive index of the documented declines and describe the conditional scenario that ecological theory permits. These sections are separate from the Results so that readers can evaluate the evidence independently, and they are presented as testable propositions rather than findings.

3.4 The chemical cascade in seawater: a hypothesis and its signatures

Aragonite is not just an index for shell-builders. It tracks the chemistry that sets the physical character of seawater itself, its clarity, its structure, and the ease with which a larva can move, feed, or simply hold station. The three tiers of evidence set out in Section 1.2 apply throughout what follows and are the standard against which every claim in this section should be read. One qualification governs everything that follows: the chemistry itself is established, but its magnitude in the open ocean is not, and whether these reactions proceed fast enough and far enough to matter biologically at open-ocean concentrations remains unmeasured.

The uncertain part of this argument is not the chemistry. Charge neutralisation, double-layer compression and coagulation kinetics describe how charged particles behave in any electrolyte, and seawater is an easier case than fresh water because its high ionic strength compresses the double layer, which is why ferric coagulation is run at an industrial scale in desalination pretreatment⁹⁴ and, in our own practice, in aquarium life-support systems. What is unmeasured is the magnitude: whether, at open-ocean concentrations and under continuous mixing, the reactions proceed fast enough and far enough to matter biologically. No marine figure exists because the parameters have not been collected. Water treatment

facilities log zeta potential, turbidity, particle size distribution, and coagulant demand as routine operating data, while public aquaria continuously monitor redox potential; however, no sustained ocean observing programme records any of these measurements. Adding these measurements to the existing time-series stations would be inexpensive relative to the carbonate chemistry already collected there, and it would convert the middle and lower tiers above from argument into observation.

Desalination plants use ferric chloride coagulants to strip organic matter, algae and colloids from seawater before it reaches the membranes. Ferric coagulation destabilises inorganic particles easily in seawater's compressed double layer, a result of the high ionic content⁹⁴, and the coagulation of dissolved organic matter also improves as the pH declines⁹⁵. On this evidence, we infer that every downward step in ocean pH nudges seawater towards the regime in which its organic load coagulates, while the carbonate buffering that resists the shift is consumed. What sustained acidification does to natural coagulation at a basin scale has never been measured, and that gap is what this report sets out to address.

Temperature is itself a chemical lever: warming lowers seawater pH directly through the shift of its dissociation constants by about 0.011 units per degree⁹⁶; reduces the solubility of oxygen; and accelerates the microbial respiration, so a heatwave pushes pH, oxygen and redox potential down together. When the trigger is sharp, the response can be fast: during 2025 in South Australia, in a period of anomalously warm sea surface temperatures, a bloom of *Karenia* dinoflagellates, including the brevetoxin producer *Karenia cristata*, spread across some 20,000 km², persisted for more than a year and killed on the order of a million marine animals of more than 600 taxa⁹⁷. A falling aragonite index means less capacity to buffer exactly this kind of shock.

Iron sits at the centre of all of these loops. The coupled chemistry, the loss of buffering, the calcium-bridged gels and the microbial loop together can lower oxygen and shift the redox balance, and iron, one of the most effective natural flocculants and itself acutely sensitive to pH-driven shifts in its chemical speciation⁹⁸, answers to pH, alkalinity, redox and zeta potential all at once; it also limits growth across the high-nutrient, low-chlorophyll basins that include two of our cold-water sites. Acidification lowers the iron available to phytoplankton⁹⁹, and iron cycles photochemically in the sunlit layer, where light reduces ferric iron and sustains a reactive, oxidising skin^{100,101}. The response to acidification is already measurable: falling pH speeds the reduction of ferric to ferrous iron by dissolved polyphenols, and the organic ligand slows its reoxidation, so reduced iron persists longer as the water acidifies¹⁰²; a recent review reaches a similar conclusion for the redox pathway and notes that, on this basis, acidification could raise rather than lower iron bioavailability, while stronger organic complexation could lower it, so the net direction for iron, copper and zinc depends on which pH-controlled pathway dominates¹⁰³. How the full photo-redox system responds to acidification and to the darkening reported here still deserves dedicated study.

Two kinds of place show where this sequence of reactions ends. The Sea of Marmara, after decades of pollution, produced its worst recorded mucilage outbreak from late 2020 into 2021: gelatinous 'sea snot' filled the sea for months, clogging nets and suffocating marine life, with mass mortality on the seafloor expected from the oxygen depletion of the bottom

water¹⁰⁴; the deep water was already hypoxic after years of eutrophication, so the bloom fell on a sea that had lost most of its oxygen¹⁰⁵. The world now has more than 500 dead zones, from the Baltic and the Black Sea shelf to the Gulf of Mexico, the Arabian Sea, the Benguela shelf, the Adriatic, the East China Sea and Chesapeake Bay^{106,107}. Where their carbonate chemistry has been measured, the hypoxic water is also acidified: respiration of the eutrophication-driven organic load lowers the subsurface pH in the northern Gulf of Mexico and the East China Sea⁹, and in Chesapeake Bay the bottom water reaches a pH of about 7.45 with aragonite saturation well below 1¹⁰⁸. These are not separate stories: nutrient loading has crossed its own planetary boundary, at about three times the boundary for nitrogen and about twice the boundary for phosphorus², nutrients all end up in the sea; and every dead zone is an extreme local preview of the open-ocean chemistry this report describes.

The cascade favours still water, and the ocean is becoming stiller. Layering has strengthened and ventilation weakened over the past half-century¹⁰⁹, and oxygen is falling^{106,110}, a change with direct consequences for the structure of zooplankton, fish and other pelagic communities¹¹¹. The animals themselves may be part of the circulation: vertically migrating krill and other centimetre-scale zooplankton have been observed to generate turbulence in a coastal inlet and in the laboratory, although their contribution to open-ocean mixing remains debated^{112–114}, so if that contribution is significant, their decline (Section 2.6) means less biological mixing and more stratification, and if rising micro-viscosity damps small-scale turbulence, it will slow the mixing further. Each loss makes the ocean more stratified, and the more stratified the water column, the more difficult it is for marine life.

Two separate analyses of the ocean-colour archive show the ocean changing in the way the cascade predicts. About 40% of the ocean has changed colour in two decades, becoming greener at low latitudes, a plankton-community signal assessed at approximately 95% confidence, with the authors reporting the sensitivity to that choice¹¹⁵; we use their corrected values. Over a near-identical period (2003–2022), 21% of the ocean has darkened, a change the authors describe as widespread in the major oceans rather than confined to the coasts and particularly so in polar regions, the same cold waters mapped in Figure 3, and in the Northeast Atlantic and the Northwest Pacific, with the photic zone, defined in that study by the light threshold to which *Calanus* copepods respond, shrinking by more than 50 m across about 9% of the ocean surface and by more than 100 m across 2.6% (the corresponding areas for the conventional 1% light depth are 0.39% and 0.01%); about 10% of the ocean brightened over the same interval¹¹⁶, and at the Bermuda time-series site the vertical structure of the phytoplankton community is shifting, with subsurface biomass rising over the last decade¹¹⁷. Darkening is what a rising colloidal load looks like from space; the satellite series carries no propagated per-retrieval uncertainty (Methods), and at the productive sites the darkening cannot be separated from the greening because the satellite measures of attenuation and chlorophyll share the same blue-green band ratio. Only where the two diverge can the signals be told apart, and the one site in this study where they do is Station ALOHA, which therefore carries the optical argument on its own.

At Station ALOHA, in the nutrient-poor subtropical North Pacific gyre, surface chlorophyll shows no trend in three decades of water samples¹¹⁸, yet light attenuation has risen. The rise is small, about 3% over 2003 to 2021 in the MODIS-Aqua record, but it is statistically significant after correction for autocorrelation, and the chlorophyll change over the same period accounts for about one fifth of that amount (Figure 5a). The merged European ocean colour record, which partitions the optical signal into its absorbing and scattering parts, shows no trend in total absorption at 443 nm at this site over 1998 to 2025, so coloured dissolved organic matter is not the cause of the darkening, which leaves particulate backscatter, the term that responds to small suspended particles, as the only term that rose, although that rise is not itself statistically significant.

In the same water, the dissolved organic carbon inventory climbed at about 300 mmol C m⁻² per year from 1993 to 1999¹¹⁹. A small change in attenuation at the surface becomes a large change at depth, because the loss compounds through the water column: the fitted rise at Station ALOHA removes about 4% of the light at 50 m, 8% at 100 m and 11% at 150 m (Figure 5b), in water where, on the fitted attenuation coefficient, one percent of the surface blue light reaches about 180 m depth. The cause is not settled, because the satellite record can show that the material is particulate rather than dissolved but cannot identify the particles, and mineral dust, calcite plates from coccolithophores and small non-algal particles produced by aggregation all fall within the same optical description. The station's primary production has risen, and nitrogen fixation supplies a large share of its new production¹¹⁸, so the extra particles may be biological supply rather than pH-driven aggregation.

The cascade proposed in this section predicts a rise in the small particle load as pH decreases, and while this observation aligns with that prediction, it does not confirm it, as the same observation also supports the other candidates mentioned above. The direct test is a laboratory measurement of the coagulation response of natural seawater particles to a pH change of the size already recorded, using particle sizing and zeta potential, and that measurement would settle the question for this site more directly than any satellite record. The observation should also be read for what it is, a single station in a subtropical gyre; at the cold productive sites chlorophyll rose alongside attenuation over the same period, so the two signals cannot be separated there, and one oligotrophic site cannot establish that the same process operates in the cold basins that are the subject of this paper. Testing requires in-water optical and particle measurements at the subpolar sites themselves.

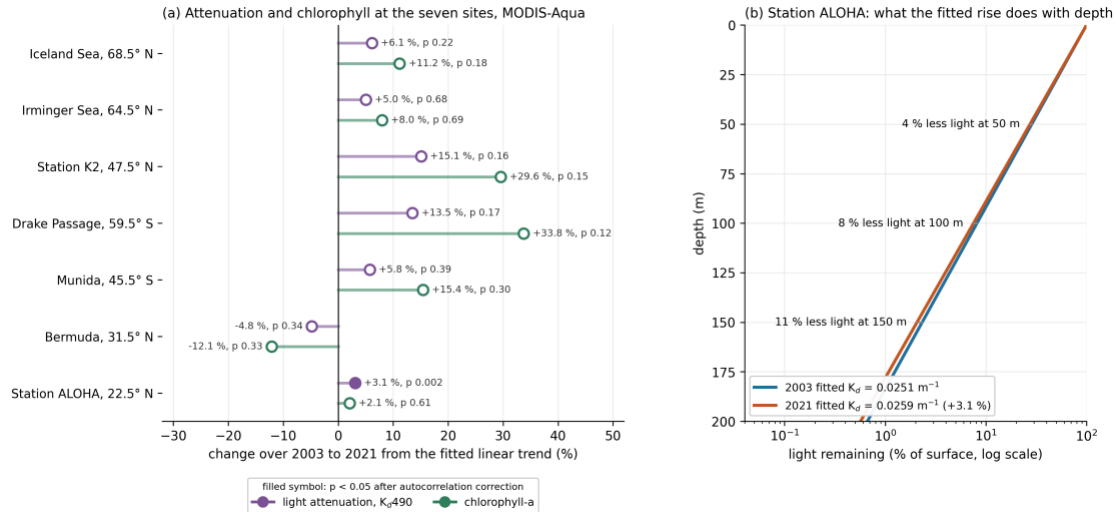


Figure 5. Satellite light attenuation and chlorophyll at the seven sites. (a) Change over 2003 to 2021 from the fitted linear trend in the diffuse attenuation coefficient K_d 490 (purple) and chlorophyll-a (green) from MODIS-Aqua alone, with p values corrected for first-order autocorrelation; filled symbols are significant at $p < 0.05$. Chlorophyll rises with attenuation at the productive sites and falls with it at Bermuda, and Station ALOHA is the one site where attenuation rises while chlorophyll does not. (b) Light remaining with depth at Station ALOHA for the fitted K_d 490 of 2003 and 2021; a rise of 3.1% at the surface removes 4% of the light at 50 m, 8% at 100 m and 11% at 150 m. Data: NASA MODIS-Aqua R2018.0 monthly 4 km products served by NOAA OceanWatch (Methods).

Two objections to this mechanism deserve a direct answer here. The first looks like a contradiction: if falling pH promotes coagulation, and coagulation makes particles clump and settle, how can the same process leave the water thicker and hold small cells in the light? The answer is that coagulation is a sequence of states rather than a single outcome. Destabilisation begins when the double layer is compressed and repulsion falls; whether the aggregates then settle depends on how far the process runs and how much the water moves. A clarifier drives the process to completion in a quiescent tank, and the water clears, but the open sea has no quiescent tank, its particle concentrations are orders of magnitude lower, and continuous mixing returns aggregates to suspension. What accumulates is the intermediate state, a partially destabilised, gel-like load of transparent exopolymer and colloidal organic material that raises the effective viscosity at small scales while removing very little mass. The second objection comes from Stokes' law: aggregation increases particle radius, and settling velocity scales with its square, so clumping should accelerate export. That is correct where aggregation runs to completion, and marine snow sinks, but the object that aggregates is the colloidal and polymeric matrix rather than the living cell, and the two pathways coexist. The position of a basin relative to the partition depends on its organic loading, mixing energy, and particle spectrum; however, this partition has not been measured in seawater. We present retention as the pathway that the cascade would favour under low mixing and rising organic load, not as the only outcome.

That partition matters less than it first appears, because both of its outcomes act against the lit layer. The retention branch raises the suspended load and attenuates light, which is the pathway described above. The export branch removes material from the surface layer altogether, and what leaves is not only carbon but also the nitrogen, phosphorus, and iron

bound up with that carbon. Surface dissolved iron is held almost entirely in organic-ligand complexes¹⁰⁰, and if that bound pool aggregates with the organic matter that carries it, accelerated flocculation would strip the element whose scarcity already holds the HNLC communities in small cells. Loss downward matters only in relation to resupply upward, and the resupply is slowing because stratification has strengthened and ventilation weakened over the same period, as described earlier in this section. Faster removal set against slower return depletes the surface layer even when the total inventory of the ocean is unchanged. There is a food-web consequence as well, since organic matter that aggregates into rapidly settling flocs falls past the grazers rather than to them.

We put this forward as a direction rather than a magnitude, and the evidence does not point one way only. High-CO₂ mesocosms have shown carbon drawdown in excess of the Redfield expectation, which is more carbon exported for a given nutrient supply¹²⁰. Modelling of an acidification-altered pump also produces extended subsurface oxygen deficits¹²¹ of the kind now observed^{106,110}, but in that model they arise because reduced carbonate ballast lets organic matter remineralise at a shallower depth, not because export increases. Against this, the organic carbon flux reaching the deep sea is tightly associated with carbonate and opal ballast¹²², so the reduced calcification that acidification brings removes ballast and would slow export rather than accelerate it, and it has separately been argued that acidification increases the buoyancy of exopolymer aggregates and drives them upward, which would reduce export directly¹²³. A large global increase in export would in any case have to be visible in the ocean carbon sink and in mesopelagic biomass, and we know of no evidence that it is visible there. We therefore state the export branch as a regional and modest possibility that would deepen surface depletion where it operates, not as a change in the global flux, and we attach no number to that possibility.

The greening waters are the HNLC zones, where iron limitation holds the community in small cells. Cell size governs sinking: by Stokes' law, the settling velocity scales with the square of the cell radius and in inverse proportion to viscosity¹²⁴, so the smallest cells are already close to neutrally buoyant, and a more colloidal, more viscous medium keeps them in the lit layer while larger cells still export. If this mechanism operates, it forms a self-reinforcing loop that feeds on itself. Small cells, with their higher surface-to-volume ratio, take up the scarce iron more efficiently and leave less for larger phytoplankton; they are also a poor food source for zooplankton, so the grazers decline, and with them go both the iron they recycle to the surface and the mixing their daily migrations provide. This proposed mechanism would act alongside the established controls of iron supply and light, and the iron limitation is partly biological in origin: whales and krill once recycled iron into the surface layer, and their depletion cut that supply¹²⁵. With each turn of the loop, the system becomes increasingly locked into a downward-spiralling small-cell state due to fewer grazers, less recycled iron, a smaller community, and reduced mixing.

Where does the cascade end when it runs in the open ocean? In the open ocean the cascade runs in slow motion, saturation and pH creeping down and turbidity creeping up, but three of the four high-latitude cold-water sites are already at or below Ω 1.6 and the fourth sits just above it, and stressed systems eventually break. One possible end state is that the cold, productive seas come to resemble the HNLC zones: small cells, few grazers, weak recycling, and falling productivity. If the cascade operates as described, it should force and propagate

this state, and HNLC-like conditions should spread as saturation declines. Our own surface sampling in the equatorial Atlantic, which found very little marine life visible above 20 microns and unusually high arsenic in the region's Sargassum, is consistent with that end state, but those observations are unpublished and not yet peer-reviewed, so we record them as motivation for the prediction rather than as evidence for it, and they are excluded from every quantitative claim in this paper. The inference from arsenic to phosphate scarcity, which rests on arsenate substituting for phosphate in phosphate-depleted water, has been demonstrated only for particular taxa, and we do not rely on it here. The satellite record is mixed so far: the unproductive gyres expanded over the first decade of the ocean-colour record ¹²⁶, while the HNLC regions have not expanded: their extent fluctuates with climate modes, and since 2010 it has contracted at the poles by up to 8% while chlorophyll there has risen ³³. In the equatorial Atlantic itself, where our own sampling was carried out, satellite chlorophyll has been low but without a significant trend across the full MODIS-Aqua record (2003–2022, geometric mean 0.15 mg m^{-3} , $+0.1\%$ per year, $p = 0.6$, over 5°N – 5°S , 35°W – 5°W ; Table 2), which is consistent with a persistently unproductive state but does not by itself demonstrate one developing. The spread of HNLC-like conditions is therefore a prediction of this framework, and one the coming decade of ocean-colour data can test.

3.5 Mechanisms of abrupt ecological change

Three well-tested ideas explain why a sudden, lasting change of state is plausible. The first is the strong Allee effect ^{127,128}: below a critical density, a population grows more slowly than it dies, so decline speeds up rather than slowing. The second idea is the link between diversity and stability. May's ¹²⁹ result that complexity can destabilise a randomly built community and the finding that diversity steadies a fluctuating system ^{130–132} fit together: a food web that loses diversity also loses its buffer. The third is the theory of critical transitions ¹³³, in which a system near a tipping point can shift abruptly, sometimes with early-warning signs that are difficult to read ¹³⁴. This does not require such a change; it indicates that a stressed food web may respond in ways other than a smooth continuation of the present trend.

Water chemistry now has to be weighed alongside the biological stability of a community. When the alkalinity of a system declines, its pH stability becomes increasingly delicate. Organic input, whether natural or from pollution, can then trigger a rapid biological and chemical cascade that could end in a dead zone or a regime shift.

3.6 The biocurve: an interpretive synthesis

Ocean acidification has been called the evil twin of climate change¹³⁵. A saturation value means little on its own; what matters is how it affects the animals that must build shells and the chemistry and physics that support the plankton and feed the food chain. It is remarkable that a marine primary-producer biomass of only about 1 Gt of carbon supports approximately 5 Gt of carbon of consumer biomass, an inverted standing pyramid held up by the rapid turnover of the producers ⁸¹; most of the roughly 2 Gt of carbon of animal biomass on Earth sits inside that consumer pool, and with it the entire marine food web, the root of the life-support system upon which we all depend, so a declining aragonite

saturation index and the chemistry that surrounds it is vitally important, not only to the stability of the marine ecosystem but to all life on Earth. This is why we built the biocurve, our attempt to assign that chemistry its biological meaning. The measured chemistry is fact, while the biocurve is a synthesis of separately documented declines, and the projection attached to it in Section 3.7 is a scenario.

A word first on why we built an index at all. There has never been a census of global marine animal life, not now and not in 1850, and none is in prospect. That leaves two options, and declining to synthesise is not the neutral one: it leaves the field with no quantified baseline against which any later measurement can be read, which is how a depleted state comes to be treated as normal. We took the other option and did it in the open. Every input is referenced to its own source, every value that is derived rather than measured is labelled as such, the interpolation is given as a formula that anyone can check (Supplementary Note 3), and the result is tested against three different weighting schemes, under all of which the present value falls between 35% and 54% of the 1850 level (Supplementary Table 5). The index therefore derives its authority from that robustness rather than from the precision of any single weighting scheme. It is offered to be replaced when better observations exist, and the fact that they do not yet exist is one of this paper's findings rather than a defence of its method.

Table 1b. Composite and interpolated biocurve values. These rows are outputs of this study, not direct measurements: the 1850 baseline is definitional, the era values are interpolations constrained by the documented declines of Table 1a, and the present value is the weighted composite with its uncertainty range.

Table 1b. Composite and interpolated biocurve values: outputs of this study					
Species / group	Approx. location	Time frame	Documented change (loss of nature)	Biocurve value (% of 1850)†	Reference
Marine animal life, baseline	Global	1850	Pre-industrial reference state	100	Definitional baseline (this study; Supplementary Note 3)
Marine animal life, composite	Global	1900	Whaling and coastal losses under way	86	This study (interpolated): era descriptions after McCauley et al. ¹³⁶
Marine animal life, composite	Global	1950	Industrial fishing and coastal and reef losses accumulate.	69	This study (interpolated): era descriptions after McCauley et al. ¹³⁶
Marine animal life, composite	Global	1985	Apex predator, seabird and reef losses advanced.	57	This study (interpolated): era descriptions after McCauley et al. ¹³⁶
Marine animal life, composite (present)	Global	~2023	~35–54% remaining (importance-weighted 35%, band 28–44%; tonnage-weighted 54%)	35–54	This study (composite; Table 1a sources)

Figure 6 plots the importance-weighted value and its uncertainty band; the tonnage weighting gives the higher end of the present range. The ocean in such a state could still hold a jelly-like layer, but the functional animal life, meaning anything that feeds fish, birds, mammals or people, would be greatly reduced.

The biocurve: an importance-weighted index of documented marine animal groups stands at 35-54% of its 1850 baseline as cold-water aragonite crosses its threshold

A weighted mean of all documented group declines (Tables 1a-1b, Appendix C), drawn to 2023. Present ~35-54% of 1850 (46-65% lost), depending on weighting. Measured aragonite Ω (right axis) has sat below the 1.6 impact threshold in cold water since the mid-1980s; the two vertical scales are independent.

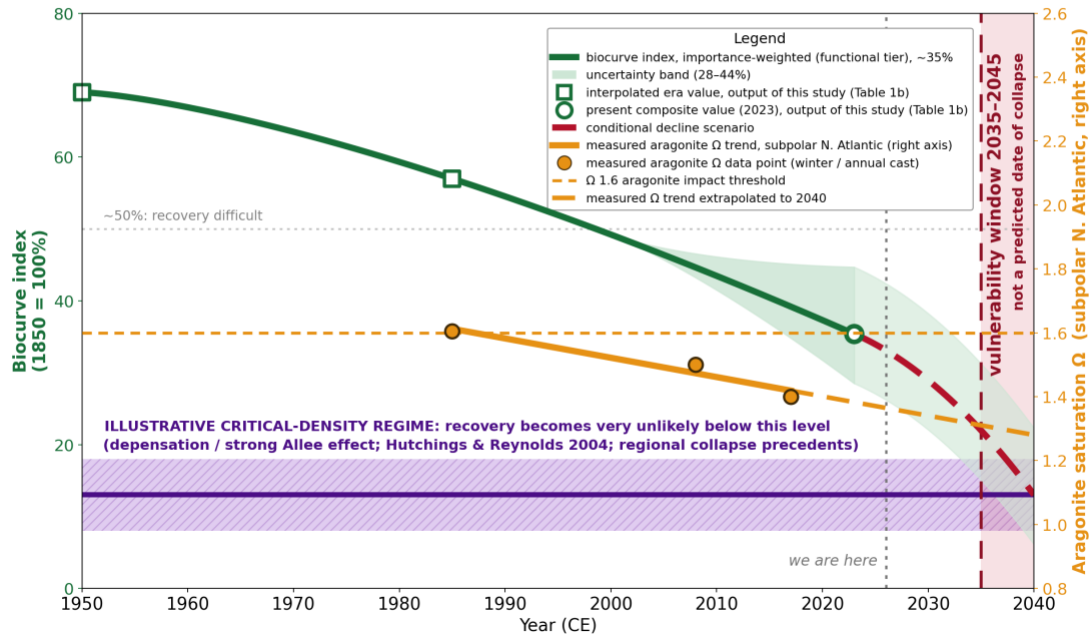


Figure 6. The biocurve: an importance-weighted index of documented group declines in marine animal life (1850 = 100%; Tables 1a-1b; Supplementary Note 3) to about 2023. Subpolar North Atlantic aragonite saturation and the Ω 1.6 threshold sit on an independent right-hand axis (Figure 1). The curve is an interpretive synthesis of separately documented declines, not a direct census; the open squares mark interpolated era values and the open circle marks the present composite (Table 1b), while the filled orange points are measured aragonite casts. The present value is about 35-54% of the 1850 level (importance and tonnage weightings; Supplementary Note 3), with the shaded band carrying the uncertainty of the importance weighting (28-44% at 2023). The red line and the shaded band from 2035 to 2045 are a scenario and not a projection: the line shows the direction that the theory of Section 3.5 permits if present pressures continue, its date of entry into the illustrative critical-density regime (purple; Sections 3.5 to 3.7) is a stated choice rather than a computed value, and the green uncertainty band continues along it at its 2023 width. The solid orange line is the least-squares trend fitted to the measured aragonite values, and the dashed orange line extends that trend linearly to 2040, reaching about Ω 1.28. The subpolar North Atlantic is shown because it holds the longest cold-water records; aragonite saturation is declining in every latitude zone at a comparable rate (Figures 2 and 4; Supplementary Table 3). The recorded decline is attributed mainly to fishing, whaling, habitat destruction and pollution (Section 2.7). The vertical shaded band marks the period from 2035 to 2045, chosen because the measured chemical trends of Figure 2 indicate that 49% of the mapped ocean will be below its winter threshold by 2035 and 57% by 2045; it is not a predicted date of collapse. By then the mechanisms of Section 3.5 could be affecting enough of the system that reversal would require rebuilding rather than easing the pressure (Section 3.7).

The extrapolation of the bio-curve and what would make up the survivor floor (Table 1c) is speculative. The measured part of the composite curve is an interpretive synthesis, a reproducible weighted mean gathered from numerous reports (Supplementary Note 3), which provides a range of about 35-54% of the 1850 level rather than a single figure. Everything beyond the present is less certain: the projected path, its dates, and its endpoint are illustrative and based on assumptions; the causes include several stressors, not just acidification; and some of the biological figures are disputed. However, the direction of

travel is not in dispute: systems are becoming more unstable, and marine life is under stress and suffering.

That reading of the evidence is not confined to the ecological literature. A United Kingdom national security assessment published in 2026 concludes, with high analytical confidence, that at current rates of biodiversity loss every critical ecosystem is on a pathway to collapse, using the same definition of collapse as this paper, passage beyond a threshold after which a system can no longer hold its essential structure or function¹³⁷. It states plainly that it is not a scientific report, that it examines a reasonable worst case and that its judgements on timing carry low confidence, so we cite it for how the risk is now treated in national planning rather than as a measurement. It still lists ocean acidification among the planetary boundaries not yet crossed, following the 2023 position while noting that a newer assessment had since appeared, whereas that reassessment, discussed in Section 1, places the global mean inside the uncertainty range of the boundary by 2020 and beyond a revised boundary since 2000⁴; that gap is the reporting lag this paper identifies.

3.7 A conditional scenario: a shortened food web, a survivor floor, and human dependence

If the chemical thresholds keep being crossed in the cold, productive waters and the calcifying base of the food web fails, one possible endpoint is what Jackson¹³⁸ called the rise of slime: a simplified, microbially dominated system with boom-and-bust cycles of toxic dinoflagellate blooms, jellyfish and disease, in which the large animals that once topped the food web are greatly reduced. In the cold productive waters considered here, our extension of that picture is a base of cyanobacteria and acid-tolerant phytoplankton feeding jelly-like zooplankton, with the commercial fish, seabirds, seals and krill-feeding baleen whales above them greatly reduced. Nor is this endpoint confined to cold water: Jackson's own examples include warm, nutrient-loaded systems such as the Gulf of Mexico, where the seasonal dead zone already runs on microbes and jellyfish. This is not just theory; it has already happened in places. A worldwide takeover by jellyfish is not supported: the available synthesis shows roughly twenty-year swings rather than a steady global rise^{139,140}. We therefore treat this endpoint as a regime shift seen regionally, not as a global trend. The high-latitude cold waters, with their low aragonite saturation and pH, are most at risk because their carbonate chemistry is the least stable. The stress on one such sea can also reach its neighbours. As the eastern Bering Sea warmed, subarctic fish, such as Pacific cod and walleye pollock, shifted northward into the northern Bering Sea^{141,142}, displacing communities and adding new predators and competitors to the Arctic systems that received them. How far such redistribution destabilises the receiving systems is still being assessed.

We do not put a precise figure on this endpoint. Below a critical density, the theory of Allee effects and critical transitions (Section 3.5) implies that a thinned community can settle into a persistent, low-diversity state rather than recover. The survivors of such a state are the disturbance-tolerant groups, above all gelatinous zooplankton, which can persist in low-oxygen water that fish cannot tolerate¹⁴³, together with cephalopods and tolerant worms; gelatinous biomass has traditionally been regarded as a trophic dead end that feeds little of what sustains the fish, seabirds, mammals and fisheries above that level, although the Benguela case shows that some of that biomass can be routed back to higher

predators ¹⁴⁴. Any percentage attached to this floor would be illustrative rather than measured, so we treat the endpoint as a conditional scenario.

The precedents below are regional, and no mechanism is established by which regional shifts propagate into a single global transition on this timescale; how that would happen, and how fast, is the least constrained part of the argument. What the 2035 to 2045 window in Figure 6 marks is therefore a shift in structural vulnerability, the point at which the Allee effects and critical transitions of Section 3.5 would be operating across enough of the system at once to make reversal a matter of rebuilding rather than of easing pressure. It is not a schedule for global biological extinction, and we do not intend it to be read as one.

Table 1c. Conditional scenario endpoint. This row is not a measurement or an interpolation; it is the illustrative endpoint of the conditional decline scenario of Section 3.7 and Figure 6, contingent on present trends continuing.

Table 1c. Conditional scenario endpoint: illustrative, contingent on present trends					
Species / group	Approx. location	Time frame	Documented change (loss of nature)	Biocurve value (% of 1850)†	Reference
Functional survivor floor (projected)	Global	~2040 (scenario)	an illustrative floor inside the 8–18% critical-density band plotted in Figure 6, not a computed estimate; jellyfish are not counted as functional animal life.	8–18 (illustrative)	This study (scenario)

The human stakes are large and well documented. Aquatic animal foods supply at least 20% of the average person's animal protein for more than 3 billion people, and around 600 million livelihoods depend at least partly on fisheries and aquaculture ¹⁴⁵. Any loss would fall hardest on the poorest coastal communities, which have the fewest alternatives.

These shifts are real; the sardine stock off Namibia, in the Benguela system, collapsed under overfishing and did not return. Jellyfish and a small fish known as the bearded goby have replaced the sardine stock, and this state of affairs has persisted for decades, even after fishing was sharply reduced in an attempt to restore this important fishing ground ^{146,147,144}. The Black Sea saw a mix of pollution, overfishing and an invading comb jelly turn the open water to jelly in the 1980s ^{148,149}; it recovered only partly in the late 1990s, after pollution fell and a second invader cut back the comb jelly ¹⁵⁰. The cod stocks in the Northwest Atlantic exhibit a similar pattern. When the stocks collapsed in the early 1990s, their place on the eastern Scotian Shelf was taken by the animals they used to eat and compete with, forage fish and macroinvertebrates such as snow crab and shrimp¹⁵¹, and, as is typical of collapsed marine fish populations, the stocks showed little or no recovery within fifteen years of the collapse^{152,153}. Recovery is sometimes possible: North Sea herring collapsed in the 1970s but rebuilt after the fishery was closed and carefully managed ^{154,153}. These are regional examples, and the available global synthesis does not yet show a single worldwide trend¹⁴⁰. But they are no longer isolated: the same end state has now appeared in different ocean basins, on different coasts and under different mixes of pressure, and regional examples repeated in many places are how a worldwide trend begins.

3.8 Conclusions

The cold, productive waters that feed most of the marine food web are acidifying at measured rates that have persisted for four decades. Every chemical stressor measured in this paper is moving in the same direction: aragonite saturation and pH are both falling, and neither trend is reversing. The redox and zeta-potential changes that follow from the same chemistry and the reduction in iron availability are expected consequences rather than measured trends, for the reasons set out in Section 3.4. Even the carbonate buffer that resists these changes is being consumed. This is not a system approaching its first threshold from a position of strength; it is a system already past several thresholds, with less and less capacity to absorb whatever comes next.

The documented record of marine animal life points the same way. Our importance-weighted index of documented declines, an interpretive synthesis rather than a census, places the loss of functional structure since 1850 at between 46% and 65% (Section 3.6; Supplementary Note 3). Acidification did not cause these issues; rather, fishing, whaling, habitat destruction, and pollution were responsible long before the chemistry began to shift. Ocean acidification is therefore not acting on a resilient, diverse system with capacity to spare; it is acting on a community already thinned and stressed to the point where the classic mechanisms of collapse are present vulnerabilities.

On that synthesis, marine life is already reduced well past the halfway point and, if present trends continue, is moving towards the 85% to 95% loss range that we have chosen to represent the critical-density regime, in which the Allee and critical-transition mechanisms of Section 3.5 would make self-reinforcing decline more likely than recovery; that range is an illustrative scenario (Section 3.7), not a value that the theory computes. On present trends, 49% of the mapped ocean surface will be below its winter aragonite ecological reference threshold by 2035 and 57% by 2045, based on a linear extrapolation of 43 years of measurement rather than on any climate model. The question now is whether the marine ecosystem will enter the critical-density regime before we take action; based on the measured chemical trends and documented biological losses, this outcome appears realistic if current trajectories continue. More than three billion people draw a significant share of their animal protein from the sea¹⁴⁵. The biological structure that is now being dismantled supports the fisheries, the carbon pump, the oxygen cycle, and the climate itself. Losing it would not be an environmental problem or an inconvenience; it would be a civilisational risk.

The decade from 2035 to 2045 is when the great majority of the world's most productive cold water will have been below its ecological reference threshold for at least ten years. That follows from the measured present state, with no extrapolation required (Methods). By this stage the conditions under which Allee effects and critical transitions operate will apply across much of the marine ecosystem, and recovery would mean rebuilding what has been lost rather than simply easing the pressure. It is no longer sufficient to just "do no harm", to stop the pollution, reduce carbon emissions and ban destructive fishing. We now have to do some good and help restore the systems. It is now time to give back to nature.

Our conclusion is that we must now intervene actively to reverse the harm already inflicted and to begin repair. An unprecedented scale of emergency response is necessary while the opportunity for recovery remains, as the alternative is not a viable option.

4. Methods

Throughout this paper, the two kinds of measurement play fixed and separate roles. The gridded OceanSODA-ETHZ product supplies the spatial and global evidence, because only the mapped field covers the whole ocean surface, while the discrete ship-based casts supply independent regional validation and direct winter measurements at selected sites. Neither is used in the other's role, and no global figure rests on the discrete sites alone.

Data sources

Table 2. Every primary dataset used in this study, with the variables taken from it, its coverage and its accession or citation. All are public.

Table 2. Data sources: every primary dataset used in this study, all public			
Dataset	Variable(s) used	Coverage	Source
OceanSODA-ETHZ v2025	omega_ar (aragonite saturation), omega_ar_uncert	1×1°, monthly, 1982–2024, global	Gregor & Gruber ³¹ ; NOAA NCEI OCADS accession 0220059, https://doi.org/10.25921/m5wx-ja34
SPOTS time-series synthesis	DIC, total alkalinity, pCO ₂ , temperature, salinity, silicate, phosphate (discrete casts)	Iceland Sea, Irminger Sea, Station K2, Munida	Lange et al. ¹⁵⁵
GLODAPv2.2022	Interior/surface DIC, total alkalinity, nutrients (quality-control cross-check for discrete casts)	Global ocean interior compilation	Lauvset et al. ¹⁵⁶
CODAP-NA	Discrete inorganic carbon, oxygen, and nutrients (North American margin cross-check)	North American ocean margins	Jiang et al. ¹⁵⁷
World Ocean Atlas 2023	Dissolved nitrate (HNLC zone definition, Figure 3)	Global climatology	Garcia et al. ³⁴
MODIS-Aqua chlorophyll-a	Chlorophyll-a (HNLC map, Figure 3; regional greening trend, Figure 4; equatorial Atlantic trend test, Section 3.4)	Global 9 km monthly climatology (Figure 3); regional 2003–2021 (Figure 4); 4 km monthly, 2003–2022 (Section 3.4)	JRC Global Marine Information System ³⁵ ; NOAA CoastWatch ERDDAP (erdMH1chlamday)
MODIS-Aqua Kd490 and chlorophyll-a (site optics)	Diffuse attenuation coefficient Kd490 and chlorophyll-a (Figure 5)	4 km monthly, 2003–2021, boxes of 6° latitude by 8° longitude around the seven Figure 4 sites	NASA OBPG R2018.0 products served by NOAA OceanWatch ERDDAP (aqua_kd490_monthly_2018_0, aqua_chla_monthly_2018_0)
ESA Ocean Colour CCI v6.0	Total absorption coefficient at 443 nm (Section 3.4, Station ALOHA)	4 km annual, 1998–2025, Station ALOHA box	Plymouth Marine Laboratory THREDDS server (oceancolour.org)

Carbonate chemistry calculations

The aragonite saturation state (Ω) for the discrete casts was computed with PyCO2SYS v1.8¹⁵⁸ from the primary SPOTS measurements¹⁵⁵, using measured dissolved inorganic carbon (DIC) with measured total alkalinity (TA) where alkalinity exists (2014 onwards) and DIC with the partial pressure of carbon dioxide ($p\text{CO}_2$) for the Iceland and Irminger Sea records before 2014, when alkalinity was not measured at those stations. Every calculation used measured in-situ temperature, salinity, silicate, phosphate and pressure, the constants of Lueker et al.¹⁵⁹, Dickson¹⁶⁰ and Uppström¹⁶¹ on the total pH scale, and samples from the shallowest 25 dbar; the full settings are given in Supplementary Table 1 and as a settings file in the deposit. Different input pairs yield Ω values that are not fully consistent, with the largest disagreement in cold water^{15,16}, so the step visible at 2014 in Figure 1 is a change in the analysis and not in the ocean, and no slope is fitted across the change. Recomputing every cast from the primary data identified a processing error in the earlier version of Figure 1: from 1995 the archived $p\text{CO}_2$ at these stations appears to be reported at a fixed laboratory equilibration temperature of 4 °C rather than at in-situ temperature (the stored-sample protocol introduced in mid-1993 is described by Ólafsson et al.¹⁶², and the 4 °C interpretation is inferred from internal evidence, Supplementary Note 1), and treating that value as the measurement temperature introduced a spurious step at 1995 of about +0.28 Ω at the Iceland Sea and -0.22 Ω at the Irminger Sea. Taking the archived value at in-situ temperature removes the step, brings the implied alkalinity into agreement with the alkalinity measured from 2014 (alkalinity-to-salinity ratio 65.9 against a measured 65.7), brings the casts within about 0.12 Ω of the independent gridded field in every period, and, when the same multivariate regression on time and temperature is applied to the 1985–2008 winter casts, reproduces the published temperature-adjusted Iceland Sea winter trend of -0.0072 ± 0.0007 Ω per year¹⁷ within the combined uncertainty (Supplementary Note 1). A simple least-squares fit against time alone over the winter (January to March) casts of the pre-2014 record gives -0.0046 ± 0.0007 Ω per year for the Iceland Sea ($n = 57$, $p < 10^{-8}$), while the Irminger Sea is statistically flat (-0.0014 ± 0.0009 Ω per year, $n = 51$, $p = 0.13$); the result is insensitive to surface cutoffs from 10 to 30 dbar. The analytical and seasonal reasons why commonly quoted saturation values exceed those that marine life encounters are set out in Supplementary Note 2.

Zonal winter threshold analysis (Figure 2)

The winter field was built from OceanSODA-ETHZ ω_{ar} ³¹ as the February value north of the equator and the August value south of it for each of the 43 years, 1982–2024, on the native $1^\circ \times 1^\circ$ grid, and the 2022–2024 mean winter field is the reference present state. A per-cell linear trend was fitted by ordinary least squares against year (43 points per cell; cells with any missing winter value excluded), and the projections to 2035, 2045 and 2055 apply that trend to the reference field over 9, 19 and 29 years counted from 2026, which omits about three years of trend and therefore understates the change; no acceleration or deceleration is assumed. Cells were classified against three ecological reference thresholds by absolute latitude: Ω 1.6 poleward of 50° , Ω 2.4 between 30° and 50° and Ω 3.3 equatorward of 30° (scheme A, Figure 2; the sources of the thresholds are given in Sections 2.3 and 2.4); scheme B moves the bands to 55° and 35° and scheme C restricts the reef

threshold to within 23.5° of the equator, and the two together give the structural range. Areas are cosine-latitude weighted. The single coldest month is used because this analysis classifies exposure, the saturation an organism meets at the seasonal minimum, rather than fitting a trend; repeating the classification on a three-month winter mean lowers the present figure by 1.1 percentage points and every projected horizon by 0.3 points or less. The site trends of Figure 4 use the three-month window instead, because there the object is a fitted slope, and the additional months improve its precision. Applied to the polar and subpolar zones alone with the same pipeline, the cosine-weighted area below Ω 1.6 in the 2022–2024 winters is 79.1% poleward of 50° and 88.0% poleward of 55°; because that water is already below the threshold, it will have been below for more than a decade by 2035 with no extrapolation required, and we report the measured present state for this zone rather than a projection, since seasonally ice-covered cells carry no fitted trend and would bias a zone-restricted projection downward.

Statistical analysis and uncertainty

Three independent sources of uncertainty were quantified for the percentage of mapped ocean below its zonal threshold. Measurement uncertainty was propagated from the product's own per-cell reported uncertainty (ω_{ar_uncert}) by Monte Carlo simulation (500 Gaussian perturbations of the 2022–2024 mean field); spatial sampling uncertainty was obtained by a block bootstrap of 10°×10° blocks (5,000 resamples), so that spatially correlated neighbouring cells are not treated as independent evidence; and structural uncertainty is the range across zonation schemes A to C. The present figure is 43.7% under scheme A, with a Monte Carlo interval of 43.7 to 43.9% and a bootstrap interval of 39.6 to 47.8%, against a structural range of 36.9 to 52.5%; the 2035, 2045 and 2055 figures are 49.0, 57.5 and 64.7%, with bootstrap half-widths of about ± 4 percentage points at every horizon (Supplementary Table 2; Supplementary Figure 1). The data-uncertainty interval is narrow relative to the other two, so the limiting factor on the estimate is where the threshold boundaries are drawn and how spatial variability is sampled, not the precision of the carbonate reconstruction at any one point, and the structural range is the bound quoted in the text. Percentages are quoted to the nearest whole point, with 57.5% given as 57% so that the projection is not overstated. For the per-cell winter trends, ordinary least squares slopes were tested against a null slope of zero, and significance was additionally assessed on a spatially thinned subsample of every fifth cell in latitude and longitude (1,329 quasi-independent cells at roughly 5° spacing): 99.8% of these cells show an individually significant negative trend ($p < 0.05$), and a Wilcoxon signed-rank test rejects a zero median trend ($p < 0.001$). The area-weighted mean trend is -0.073Ω per decade and the median is -0.072Ω per decade, consistent with the global-mean trend over 1982–2021 that Ma et al.²¹ obtained from an earlier version of the same product ($-0.071 \pm 0.006 \Omega$ per decade). The whole classification pipeline was independently re-implemented for this revision and reproduces every reported point estimate exactly.

Site trends and satellite series (Figures 4 and 5)

The winter Ω and pH trends in Figure 4 were fitted by ordinary least squares to the mapped winter mean (January to March in the Northern Hemisphere, July to September in the Southern) at the 1×1° OceanSODA-ETHZ cell nearest each site over the 43 winters

1982–2024, with two-sided t-tests against a null slope of zero and t-based 95% confidence intervals; all fourteen trends are negative and significant at $p < 10^{-10}$, and repeating the fits on the single coldest month changes each slope by less than 3%. The slopes, their confidence intervals and the product's mean reported per-value uncertainty are tabulated in Supplementary Table 3. The satellite chlorophyll and Kd490 series in Figure 4 are shown as fitted trends without confidence intervals, because the regional retrievals underlying them were compiled separately and their per-retrieval uncertainties were not propagated, so those two series should be read as indicative of direction rather than as quantified rates. The site changes in Figure 5 were computed separately from the MODIS-Aqua R2018.0 monthly 4 km Kd490 and chlorophyll-a products served by the NOAA OceanWatch ERDDAP server, as geometric means over boxes of 6° latitude by 8° longitude centred on the Supplementary Table 3 grid cells, using June to August at the three northern cold sites, December to February at Drake Passage and Munida, and all months at Bermuda and Station ALOHA, fitted by ordinary least squares over 2003 to 2021 with p values corrected for first-order autocorrelation. The total absorption statement for Station ALOHA in Section 3.4 uses the ESA Ocean Colour CCI v6.0 annual inherent optical property product over 1998 to 2025 for the same box.

Biocurve construction (Figure 6)

The biocurve of Figure 6 and the biocurve column of Tables 1a and 1b are a weighted mean of the fraction of each animal group remaining against an 1850 baseline of 100%, $\text{Nature}(t) = \Sigma(w_i \cdot r_i) / \Sigma w_i \times 100\%$, with every $r_i = 1$ in 1850. The index covers animal life in the upper 200 m and excludes the mesopelagic zone, an exclusion that runs in one direction because counting the deep layer would raise the index. Thirteen groups enter with the importance weights and documented fractions listed in Supplementary Table 4, each referenced to its source in Table 1a; the present value evaluates to 35.4% on the importance weighting with a band of 28.5 to 44.7%, to 48% with equal weights and to 54% on a production weighting (Supplementary Table 5). The 1850 value is definitional, and the present value is the only point computed directly from data; the 1900, 1950 and 1985 values are an interpolation positioned by the periods of the documented declines, drawn from 1950 as $R(t) = 69 - 33.6 \times ((t - 1950)/73)^{1.4}$, in which the exponent is a stated choice, and varying it from 1.2 to 1.8 or replacing the power law with a straight line moves the 1985 value only between 53 and 60. Removing the coastal compartment and renormalising the weights moves the present value from 35.4 to 38.5%. Because groups whose records begin after 1850 are treated as though their published decline were the whole decline, the index overstates what remains and should be read as an upper bound on surviving structure. Supplementary Note 3 provides the full construction, its caveats, and the sensitivity tests.

Use of artificial intelligence

The authors used Claude (Anthropic) as an artificial intelligence assistant in preparing this manuscript: for compiling values from the published literature, for data extraction and processing of the OceanSODA-ETHZ and SPOTS datasets, for the carbonate-system recomputation and the threshold, trend and uncertainty calculations, for figure preparation, and for checking the consistency of the text and references. Every numeric claim compiled or computed with this assistance was subsequently checked against its

primary source or re-derived from the primary data, and the corrections arising from that check are reflected in the present text. Because the carbonate-system recomputation carries the most weight, we state its status specifically: the whole pipeline, from the primary SPOTS and OceanSODA-ETHZ records through PyCO₂SYS to the threshold, trend and uncertainty results, was independently reimplemented for this revision and reproduces every reported point estimate exactly, and it was that reimplementation which identified the fixed-temperature processing error described above and in Supplementary Note 1. No reported value rests on a single unchecked pass; the reproduction package deposited at Zenodo records the checks made. The authors directed, verified, and approved all analyses, results, and conclusions, taking full responsibility for the content of this paper.

Data availability

This study synthesises published measurements and compilations; all quantitative claims are referenced in the list that follows. Every primary dataset used is public and is identified in Table 2 (Methods) with its accession or DOI: OceanSODA-ETHZ v2025 (NOAA NCEI OCADS accession 0220059, <https://doi.org/10.25921/m5wx-ja34>), the SPOTS time-series synthesis, GLODAPv2.2022, CODAP-NA, World Ocean Atlas 2023, MODIS-Aqua (JRC GMIS, NOAA CoastWatch and NOAA OceanWatch ERDDAP servers) and the ESA Ocean Colour CCI v6.0 inherent optical property product (Plymouth Marine Laboratory THREDDS server). No restricted-access data were used. The derived products generated here, namely the recomputed cast-level aragonite saturation values, the winter threshold grids and the uncertainty ensembles, are deposited at Zenodo, <https://doi.org/10.5281/zenodo.21970808>, under a CC BY 4.0 licence. The figures are built from these records and from author-derived estimates where stated (the survivor floor and the schematic response magnitude of the biocurve). The Supplementary Information contains the construction of Figure 1, the analytical and seasonal reasons why quoted saturation values read high, the full biocurve construction, the authors' field observations, and the settings, uncertainty, and site-trend tables; the deposit README contains the note for other users of the SPOTS synthesis, the depth-cutoff and neighbouring-cell checks, the spatial footprint of the discrete sites, and the record of verification of every reported value.

Code availability

Aragonite saturation values were computed with the open-source Python package PyCO₂SYS v1.8¹⁵⁸. All other analysis and figure code was written by the authors in Python using standard open-source packages (NumPy, SciPy, pandas, matplotlib, and netCDF4). The complete set of scripts, covering the carbonate-system recomputation and its depth-cutoff sensitivity, the zonal winter-threshold analysis, the polar-zone restriction, the equatorial Atlantic ocean-colour test, the seven-site attenuation and chlorophyll analysis of Figure 5 (figure6_site_optics.py, deposited with the extracted monthly box means so that it runs without server access), the uncertainty estimation, and a script that redraws each figure in this manuscript from the primary or deposited data, is deposited together with the derived grids at Zenodo, <https://doi.org/10.5281/zenodo.21970808>, under a CC BY 4.0 licence. The deposit tabulates the expected output of each script, so any divergence is

apparent at once, and we would welcome its use to reproduce any figure or number reported here.

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Author contributions

Howard Dryden and Diane Duncan designed the study. Howard Dryden compiled and analysed the published measurements and compilations; computed the aragonite saturation values with PyCO₂SYS; constructed the figures and the biocurve index; and drafted the manuscript. Diane Duncan contributed the biological interpretation and the review and revision of the manuscript. Both authors read and approved the final version.

Competing interests

The authors declare that they have no competing interests. Both authors are affiliated with the GOES Foundation, a non-profit marine science and conservation organisation; this affiliation involves no financial or commercial conflicts of interest. For transparency, Howard Dryden founded Dryden Aqua Ltd and is the inventor of AFM activated filter media, a water-treatment filtration medium manufactured from recycled coloured glass and used in water treatment systems and public aquaria. He retired from Dryden Aqua in 2022 and has since sold his entire shareholding, so he now holds no financial interest in the company. That background in water and wastewater treatment is the source of the coagulation and colloid chemistry expertise applied in this manuscript.

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