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Emergence of Anthropogenic Signals in the Ocean Carbon Cycle

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Abstract

Attribution of anthropogenically-forced trends in the climate system requires understanding when and how such signals will emerge from natural variability. We apply time-of-emergence diagnostics to a Large Ensemble of an Earth System Model, providing both a conceptual framework for interpreting the detectability of anthropogenic impacts in the ocean carbon cycle and observational sampling strategies required to achieve detection. We find emergence timescales ranging from under a decade to over a century, a consequence of the time-lag between chemical and radiative impacts of rising atmospheric CO₂ on the ocean. Processes sensitive to carbonate-chemical changes emerge rapidly, such as impacts of acidification on the calcium-carbonate pump (10 years for the globally-integrated signal, 9–18 years regionally-integrated), and the invasion flux of anthropogenic CO₂ into the ocean (14 globally, 13–26 regionally). Processes sensitive to the ocean's physical state, such as the soft-tissue pump, which depends on nutrients supplied through circulation, emerge decades later (23 globally, 27–85 regionally).

The invasion of anthropogenic carbon and heat into the global ocean occur through a cascade of biogeochemical and physical changes which are coupled to the ocean's carbon cycle (1). The ocean carbon cycle, and in particular the ocean carbon pumps, redistribute not only carbon, but also nutrients, oxygen and organic matter between the ocean's surface and the ocean's interior, playing an important role in determining the concentration of

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Author Contributions

S.S. performed all analysis and writing, with regular feedback from K.B.R., J.L.S., T.L.F., J.P.D., M.I. and R.S. The Large Ensemble simulations were set up by K.B.R. and T.L.F. and performed and post-processed by K.B.R. The sensitivity experiments and control runs were performed by R.S.

DATA AVAILABILITY

The GFDL Large Ensemble data which supports the findings of this study is publicly available through Globus (<http://poiseidon.princeton.edu>).

INTERESTS

No competing interests to declare.

atmospheric CO₂ and the functioning of marine ecosystems (2,3). As such, anthropogenic changes in the ocean carbon pumps and the ecosystem parameters and processes to which the pumps are coupled are critical for both future climate sensitivity and marine health (1,4–6).

The characterization of ocean carbon pumps arises from a conceptual framework originally proposed by Volk and Hoffert (2) which distinguishes three pumps: a solubility pump, operating via the increased solubility of CO₂ in cold (deep) waters, and two biologically operated pumps which export organic carbon (the soft-tissue pump) and inorganic carbon (the calcium-carbonate pump) from the surface to depth. In the contemporary ocean, the sequestration or invasion of anthropogenic carbon is predominantly driven by rising atmospheric CO₂ concentrations (e.g. 7)—constituting a sub-set of the solubility pump that can be denoted the invasion flux (8).

Given the importance of the ocean carbon pumps for marine life, and of the invasion flux for the strength of the ocean carbon sink and climate, there have been significant efforts to quantify and detect changes in global carbon budgets (9–15) and project changes in the pumps (3,16–18). However, both model projections and observational records of the coupled carbon-climate system are subject to uncertainties (Fig. 1). Natural variability uncertainty, the uncertainty stemming from natural variability inherent in the climate system, is a shared uncertainty amongst modeling and observational efforts and has been shown to be a significant source of uncertainty for assessing anthropogenic changes in the ocean carbon sink (17,19,20).

Emergence denotes when a simulated anthropogenic trend “emerges,” in a statistical sense, above either natural variability (this work, 17,19,21–26) and/or model uncertainty (27, Fig. 1). Detection denotes when an observed trend exceeds the uncertainty posed by both natural variability and measurement capability of the observing system (e.g. 28). For a perfectly resolved observing system, emergence and detection times are equivalent. However, for an imperfectly resolved observing system, detection times exceed emergence times, due to additional uncertainty of mapping and/or measurement error which must be overcome for detection to be achieved. Therefore emergence timescales set a lower-bound or minimum for timescales required for detection and attribution.

In this study, we use an ESM as an observing system simulation experiment (OSSE) to quantify minimum detection time scales (i.e. time of emergence - ToE) for changes in the ocean carbon cycle. To estimate emergence timescales, we unravel natural variability from anthropogenic trends in the ocean carbon cycle using a 30-member Large Ensemble (LE) experiment conducted with GFDL’s ESM2M (29,30, see Methods section). We reference all ToE calculations to year 1990, as this is the approximate beginning of the biogeochemical ocean-observing era (e.g. 31) and therefore the start of the reference period from which contemporary anthropogenic trends can emerge.

The analysis centers around the three ocean carbon pumps, with complimentary tracers and processes to which the pumps are coupled, such as acidification, warming, oxygen, nutrients and ocean color, provided for mechanistic insight and connection to the observational record

and observing system optimization. Additional model experiments which separate the rapid carbonate-chemical versus slow climate impacts of rising atmospheric CO₂ provide attributive insight into the anthropogenic drivers of change in the ocean (Fig. S1). The following chronology of emergence provides a roadmap for when and why underway and imminent changes in important ocean biogeochemical processes and tracers might be detectable.

OVERVIEW OF EMERGENCE CHRONOLOGY

Variables which reflect the accumulated or integrated invasion of anthropogenic CO₂ into the ocean, such as surface pH and pCO₂, emerge most rapidly, with ~100% of the ocean area emerging within 15 years (Fig. 2), and the majority of regional signals emerging in under a decade (Fig. 3). The impact of acidification on the calcium carbonate (CaCO₃) pump follows closely behind, with timescales of local emergence between 20–30 years (Fig. 2), and regional emergence between 9 years (the Southern Ocean) and 18 years (the Arctic) (Fig. 3).

Next to emerge, with local ToEs between 30–50 years and regional ToEs between 10–30 years, is sea surface temperature (SST), upper ocean heat content (integration between 0 and 700 meters depth), and carbon variables that are sensitive to both the physical and carbonate-chemical state of the ocean: pCO₂ and air-sea CO₂ fluxes (Fig. 2 & Fig. 3). Upper ocean heat content emerges rapidly (4 years) on a global scale, consistent with a detection-attribution study for which data-based estimates of global upper ocean temperature increases emerged within a decade (32). Southern Ocean SST presents an outlier, with ToEs extending beyond year 2100 (23). Non-emergence results from weak anthropogenic trends in SST (Fig. S4) attributed to the dynamical effects of surface freshening stabilizing the water column and decreasing convective heating from warmer subsurface waters, thereby offsetting the surface invasion of anthropogenic heat (33).

For pCO₂ and air-sea CO₂ fluxes, global signals emerge within 17 and 14 years respectively, and all regions emerge within the range of 13 – 29 years (Fig. 3). These emergence estimates are consistent with a detection-attribution study which found reconstructed, global and regional air-sea CO₂ fluxes to be emergent sometime within the 46 year period considered (34). We interpret these to reflect emergence timescales for changes in the invasion flux, as the impact of changes in the biological pumps on surface ocean pCO₂ is small (0.5 uatm, global average) relative to that of rising atmospheric and surface ocean pCO₂ (550 uatm, global average) between 1990 and 2100 (Fig. S1 and S2).

Last to emerge are variables tied indirectly to changes in the three-dimensional physical state of the ocean (i.e. circulation, ventilation, stratification), with these changes reflected also in biological processes (Fig. 2). For these variables, emergence of globally-integrated signals are considerably shorter than local emergence. The soft-tissue pump, for example, emerges globally in 23 years, for most regions in 50 years or less, but local emergence exceeds 76 years for 50% of the global ocean area (Fig. 3). Emergence of surface chlorophyll, the primary observable currently used to monitor biological productivity and export, follows closely behind, with the global signal emerging after 25 years, however regional signals

taking up to 8 decades for emergence, exceeding previously published, biome-scale ToE estimates by up to 4 decades (26). If chlorophyll inventories over the upper 500 meters, rather surface concentrations of chlorophyll are considered, emergence times decrease by ~10 years for the global signal and by multiple decades for many regions (Fig. 3).

Emergence of globally-integrated O_2 inventories requires only 17 years, however the Arctic, the Indian Ocean, and the Equatorial and South Pacific Ocean regions require more than 100 years to emerge (Fig 3). The North Pacific stands out with only 20 years for emergence of O_2 inventories. The relatively early emergence of global and North Pacific O_2 inventories is consistent with finding of a detection-attribution study, for which observations of global and North Pacific O_2 inventories over ~20 years were found to be anthropogenically forced, whereas inventory changes in all other regions were indistinguishable from natural variability (35). The thermally-driven components of O_2 trends ($O_{2,SAT}$) emerge sooner than the full O_2 signal, locally, globally and for most regions (Fig. 2 & Fig. 3). The non-thermal component of O_2 (AOU, apparent oxygen utilization) has comparable ToEs as O_2 regionally, however the global signal for AOU is non-emergent over the 21st century, a consequence of compensating regional trends diminishing the globally-integrated signal (Fig. 2, Fig. 3 and Fig. S4).

Local emergence of over half the ocean surface area occurs this century (Fig. 2 and final column of Fig. 3), even for slowly emerging variables such as net primary production (NPP) and mixed layer depth (MLD). This highlights the value of time-series observations in climate change monitoring efforts, particularly for the fields that are not directly remotely observed (e.g. nutrients, O_2). However as a compliment to time-series locations, globally- or regionally-resolved observations which allow for integration over space require less time for detection to be achieved. This is a consequence of a noise reduction associated with averaging over compensating features of natural variability (e.g. East and West Pacific SST anomalies during ENSO events). Therefore, for nearly all variables considered, the order of emergence is global, then regional, and finally local – demonstrating the utility of observing networks with large spatial footprints for early trend detection.

THE $CaCO_3$ PUMP

The contribution of changes in $CaCO_3$ export to the sequestration of anthropogenic carbon is small (Fig. S2), however changes in the $CaCO_3$ pump still represent a change in the ocean carbon cycle and a biological impact of climate change. We present emergence timescales for the export of $CaCO_3$ rather than the indirect chemical (buffering/alkalinity) contributions that impact surface pCO_2 and air-sea CO_2 gas exchange, as these are not explicit diagnostics in ESM2M.

Decreases in the export of $CaCO_3$ emerge rapidly, with about 50% of the ocean being emergent within 30 years (Fig. 3, Fig. 4a & 4d), lagging approximately a decade behind its principal driver of declining Ω , the saturation of $CaCO_3$, which is critical for biological calcification. In ESM2M calcification rates are directly proportional to the degree of supersaturation of $CaCO_3$, calcification immediately transforms into detritus, and dissolution does not occur in the upper 100 meters where waters are supersaturated with respect to

CaCO₃. Therefore, it is changes in the production (and not dissolution) of CaCO₃ which are responsible for changes in its export (30). The decreased production and ultimately export are due entirely to the invasion of anthropogenic CO₂, and not changes in the physical climate (Fig. S1), consistent with the emergence times for the CaCO₃ pump mirroring changes in carbonate chemistry (e.g. Ω) rather than physics. The lag between declines in Ω and corresponding declines in the CaCO₃ export is due to contributions from noisier co-drivers, such as temperature and nutrient concentrations, on CaCO₃ production and ultimately export.

Emergence of CaCO₃ export occurs earliest at the high latitudes (~20 years) rather than the mid and low latitudes (30–40 years) (Fig. 4a), despite the decline in export being strongest at low-latitudes (Fig 4d). This occurs because the spatial pattern of ToE for CaCO₃ export is strongly determined by the magnitude of decadal variability, which varies strongly by latitude (Fig. S4a). In ESM2M, decadal variability of CaCO₃ export scales with the magnitude of export (i.e. lowest at the high latitudes and greatest in the tropics, Fig. S5). As a consequence, the strongest signals and earliest emergence are anti-correlated (i.e. areas with the most pronounced trend emerge the slowest).

Changes in CaCO₃ cycling result in changes in the CO₂ buffering capacity of seawater and salinity-normalized alkalinity (nALK, Fig. S3 & S4). Changes in nALK represent an accumulated or integrated effect of changes in CaCO₃ export, and therefore emerge prior to changes in CaCO₃ export (e.g. Fig. 2 and 3). In a previous study, observational and interannual uncertainty for salinity and organic-matter-cycling normalized alkalinity were combined to consider Time of Detection (ToD) of anthropogenic signals, finding local ToD of 20–30 years for the low-to-mid latitudes, and in excess of 40 years at the high latitudes (28, and consistent with our ToE estimates presented in Fig. S3). Longer ToE and ToD in the high latitudes results from the reduced exposure of upwelled waters to biogenic CaCO₃ cycling (28) and is consistent with the relatively weaker high-latitude anthropogenic trends in CaCO₃ export shown here in Fig. 4d. Estimates of ToD are only modestly longer than our estimates of ToE as contributions to uncertainty in trend detection from measurement error is of the same order of magnitude as contributions from natural variability on decadal timescales.

THE INVASION FLUX

Emergence of air-sea CO₂ fluxes occurs within high and tropical latitudes between 20–30 years, however the subtropics remain non-emergent throughout the 21st century (Fig. 4b), consistent with the ToE and spatial patterns of LE simulations with an independent ESM of similar resolution (19). The same pattern of emergence unsurprisingly holds for pCO₂, the primary driver of changes in solubility and air-sea carbon fluxes (Fig. 5a). In the subtropics, the non-emergence is a consequence of the annual mean trends in air-sea fluxes and sea-air pCO₂ being weak (Fig. 4f and 5b). However, this weak annual-mean trend results from and obscures significant increases in the amplitude of the seasonal cycle and diverging seasonal trends (Fig. 5d and 5f).

The seasonal cycle of surface ocean $p\text{CO}_2$ is driven primarily through the seasonal cycles of dissolved inorganic carbon (DIC) and SST (36), but it is *not* changes in the seasonality of the drivers which result in the seasonal amplification of $p\text{CO}_2$ (i.e. for the subtropics, DIC seasonality decreases by only 7% and SST seasonality increases by only 4% over the 21st century). Rather, amplification of $p\text{CO}_2$ seasonality is largely sustained through the cumulative effect of invading anthropogenic CO_2 upon the carbon dioxide buffering capacity of seawater, the Revelle factor (e.g. 37,38). The impacts of reduced buffering capacity on the seasonal cycle of air-sea CO_2 fluxes and $p\text{CO}_2$ finds maximum expression in summer over the subtropics (Fig. 5d), where the seasonal cycle is thermally dominated (39,40). The trend toward summer outgassing (Fig. 5d) dominates the trend toward winter uptake (Fig. 5f), producing the weak (and non-emergent) annual mean trend (Fig. 5a–b). The seasonal amplification at a subtropical location (35°N) is illustrated in SST-DIC phase-space (Fig. 5g), for which the seasonal cycle of SST and DIC remain relatively constant over time, however the resulting $p\text{CO}_2$ exhibits strong amplification during the 21st century, as evidenced by the trajectory of the seasonal cycle crossing more $p\text{CO}_2$ contours during year 2100 than year 1990.

As a result of the amplification and diverging seasonal trends, ToE in the subtropical convergence regions is significantly earlier for $p\text{CO}_2$ and fluxes considered separately for winter and summer (Fig 5c and 5e) than for the annual mean (Fig. 5a). Observationally-based products of ocean surface $p\text{CO}_2$ demonstrate enhanced seasonality over the recent decades (41), which is consistent with the ~30-year emergence timescales of seasonal trends in $p\text{CO}_2$ (Fig. 5c and 5e).

THE SOFT TISSUE PUMP

As with the CaCO_3 pump, the contribution of anthropogenic changes in the export of organic carbon to the sequestration of anthropogenic carbon is small (Fig S2), but represent modification of important ecological processes. For most of the global ocean, reductions in the soft-tissue pump emerge by the mid-to-end of century (Fig. 4c & 4f). These reductions are ultimately a consequence of the reduced supply of nutrients to the surface ocean resulting from slowly-emerging changes in ocean circulation and stratification (18, Fig. S3 & S4).

The soft-tissue pump and surface chlorophyll broadly agree in both the pattern and timing of emergence (Fig. 4 & 6). For the two fields, ToEs agree (within 20% of each other, i.e. $|\text{ToE}_{\text{soft}} - \text{ToE}_{\text{chlorophyll}}|/\text{mean}(\text{ToE}_{\text{soft}}, \text{ToE}_{\text{chlorophyll}}) < 0.2$) for 66% of the ocean area, and the underlying signal direction agrees for 87% of ocean area (i.e. decreased chlorophyll corresponding to decreased export). The agreement in timing and direction of changes in surface chlorophyll concentrations and the soft-tissue pump in GFDL's ESM2M is supports the underlying assumption of field campaigns such as NASA's EXPORTS (42), namely that anthropogenic signals in ocean color correspond to the strength of biological pump on decadal to centennial timescales.

An exception to coupling between surface chlorophyll and the soft-tissue pump occurs over the Southern Ocean, where export decreases despite increasing surface chlorophyll

concentrations (Fig. 6). The disagreement between trends in surface chlorophyll and the soft-tissue pump arises as a consequence of increased iron limitation with depth producing divergent surface and subsurface trends in both productivity and chlorophyll concentration (Fig. 6, Fig. S6). Considering instead chlorophyll integrated over the upper 500 meters (Fig. 6d), which decreases in the Southern Ocean, provides better agreement with the reduction in upper ocean net primary production (NPP; Fig. S3, S4) and the soft-tissue pump (Fig. 4c & 4f).

Disagreement in anthropogenic trends between surface and subsurface chlorophyll increases uncertainty for data-based estimates of NPP and the soft-tissue pump, which are derived primarily through observations of surface ocean color. One solution to mitigate this uncertainty, and to reduce emergence times, is increasing coverage by observing platforms that are not limited only to surface measurements, for example biogeochemically equipped profiling floats (43) and water-column-profiling satellite lidar (44).

EMERGENCE AS A LOWER BOUND ON DETECTION

Our emergence timescales provide a lower bound for detection timescales, for the following five reasons. Firstly, overcoming measurement and sampling error extends the duration of observational time-series needed for detection. Secondly, uncertainty in the methodology of emergence calculation, and the indication that alternative methods can produce longer ToEs in the case of surface chlorophyll (26) and ToEs differing by >20% with the use pre-industrial rather than contemporary noise for a variety of ocean variables (Fig. S7).

Thirdly, we consider a high-emissions scenario, which would act to shorten ToEs relative to a lower-emission scenario, but only for slowly emerging variables like the soft-tissue pump. In contrast, ToEs for more rapidly emerging variables like pH, CaCO_3 export and SST are scenario-insensitive, as these variables generally emerge prior to the separation between future scenarios (year 2006 formally, but an additional 2–3 decades for the impact of differential emissions to be evident in upper-ocean temperature (45), indicating change induced from committed surface warming and acidification is sufficient for emergence.

Fourthly, uncertainty in model representation of natural variability could extend detection times. Model inter-comparison indicates ESMs show significant differences in natural variability estimates (27) and model-observation comparison indicate models such as ESM2M under-predict natural decadal variability in the ocean's physical state relative to the natural variability estimated from observational products (e.g. 46,47). Such insufficiencies in simulated variability have been shown to arise on interannual timescales and shorter timescales due to insufficient resolution to permit eddies (e.g. 48) and there continues to be open discussion in the literature as to why coupled ocean-atmosphere decadal variability is anemic (46). Under-represented natural variability in models implies emergence calculations would be biased early.

Fifthly, uncertainty in model response to anthropogenic forcing poses an additional scientific uncertainty for estimating ToE (i.e. other ESMs have different forced responses, internal variability and potentially emergence timescales). For variables like O_2 and NPP, inclusion

of model uncertainty in the framework for calculating emergence timescales extends emergence estimates by decades (27) in comparison to the values presented here. Thus for the carbon pumps and drivers, its inclusion could also significantly extend the ToEs presented here.

CONCLUSIONS

The three ocean carbon pumps considered have distinct spatial patterns of emergence, including rapid emergence of CaCO_3 export at high-latitudes and non-emergence of the annual-mean invasion flux in the sub-tropics. The three pumps have disparate ToEs, ranging from under a decade to over a century. This disparity reflects slower emergence for physical upper-ocean properties which determine emergence timescales for the soft-tissue pump, and more rapid emergence for the invasion of anthropogenic CO_2 and its biological impacts on calcification. The primary observables tied to each flux can emerge before (alkalinity preceding CaCO_3 pump), in tandem (pCO_2 and the invasion pump), or after (ocean color lagging soft-tissue pump), further widening the gap between detection timescales for changes in the ocean carbon pumps.

Our results highlight the considerable observing system requirements for trend detection including high temporal and spatial resolution and multidecadal length sampling. For example, the analysis presented in this work shows that full seasonal resolution of surface pCO_2 and depth resolution of ocean color is critical to optimal observing system design. This LE OSSE is best understood as complementary to parallel OSSE efforts that consider constraints of observing platforms and address optimal spatiotemporal sampling strategy (e.g. 49,50).

Another important challenge will be to apply the results derived here to better constrain mechanistic controls on marine feedbacks to the climate system, an important source of uncertainty in climate projection over the coming decades to centuries (51). We present the emergence times for changes in the ocean carbon cycle induced by the summation of direct anthropogenic forcings and climate-carbon feedbacks. Distinguishing between the two, within an emergence framework, could provide timescales over which the magnitude of the ocean's climate-carbon feedback could be observationally constrained, and contribute to the mechanistically-based framework for interpreting emergence timescales presented here.

METHODS

I. SIMULATIONS

All simulations are conducted with the coupled Earth system model GFDL-ESM2M developed at the Geophysical Fluid Dynamics Laboratory (29,30) for which fidelity of the biogeochemical model (TOPAZ) has been documented for preindustrial (30), historical (52) and future (53) boundary conditions. We use a 30-member ensemble simulation over the period 1950–2100 (23). Each simulation follows a historical (1950–2005) and RCP8.5 concentration pathway (2006–2100) boundary condition with prescribed atmospheric CO_2 concentrations. The initial conditions of ensemble members 2–30 are modestly perturbed through using the climate state (ocean, atmosphere, land, sea-ice) from January 2nd-30th

from year 1950 from the first ensemble member as the January 1st 1950 condition for ensemble members 2–30, respectively. Initial condition perturbation results in a rapid (within ~5 years) randomization of internal modes of variability across ensemble members. Differences between ensemble members are solely due to natural internal variability, and similarities in the evolution of ensemble members over time are due to anthropogenic forcing.

Additionally, we conduct 2 sensitivity experiments and a 1600 yearlong pre-industrial control run. The first sensitivity experiment excludes the radiative impacts of rising atmospheric CO₂ on the climate system (i.e. no warming), while maintaining the carbon chemistry impacts of rising atmospheric CO₂ on air-sea gas exchange (labeled as BGC-only in Fig. S1). The second sensitivity experiment has precisely the opposite configuration, in which the radiative impacts of rising atmospheric CO₂ are included (i.e. warming), however air-sea gas exchange is not affected by rising atmospheric CO₂ (labeled as Rad-only in Fig. S1).

II. TIME OF EMERGENCE METHOD

We refer to anthropogenically forced trends as “signals” and natural trends as “noise”. The signal is computed by averaging across the 30 trends (linear, least-squares) given by the Large Ensemble (LE). The noise is computed by taking the standard deviation of these 30 trends. All trend calculations are performed on annual means (unless otherwise stated) and started in year 1990, as this is the approximate beginning of the ocean biogeochemical observing era. Decadal trends in a stationary climate system (the noise) are approximately normally distributed about zero, therefore we can use the standard 2-sided Student’s *t* test. The null-hypothesis (signal is due to natural variability) is rejected with >95% confidence when the magnitude of the signal is twice the magnitude of the noise, i.e. when the signal-to-noise (SNR) equals or exceeds two. The Time of Emergence (ToE) is the first year at which $SNR \geq 2$.

All trend calculations are performed on annual means (unless otherwise stated) and started in year 1990, as this is the approximate beginning of the data-rich ocean biogeochemistry observing era (31). ToE calculations are performed at the grid-cell level (1°x1°), regionally, and globally. At each grid-cell, there are 30 individual time-series. At each domain (either global or regional), first a single timeseries of the domain-averaged or domain-integrated quantity is computed, providing 30 individual time-series. From these individual time-series, the trends, signal, noise and ToE are computed.

The regional bounds from the Regional Carbon Cycle Assessment Project (RECCAP) protocol (<http://www.globalcarbonproject.org/reccap/protocol.htm>), are used. The Southern Ocean is defined as lying south of 45°S. The Arctic is defined as the region north of 65°N. For the Pacific and Atlantic basins, North is defined as 18°N–65°N, Equatorial is defined as 18°N–18°S and South is defined as 18°S to 44°S. For the Indian basin, North is defined as lying north of 0°N, and South is defined as 44°S–0°S. The color schemes for Figures 2–6 and S2–S7 were created with tools available at davidjohnstone.net.

Supplementary Material

Refer to Web version on PubMed Central for supplementary material.

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Figure 1. Venn Diagram schematic of sources of uncertainty in simulating (using Earth-System Modeling approach) and observing changes in the Earth system.

For emergence, detection or attribution of an observed or simulated signal, the signal must overcome the sources of uncertainty in their respective brackets.

Figure 2. Fractional Area of Emergence for Anthropogenic Trends in Ocean Biogeochemical and Physical variables

A time series of the percent of the global ocean area with locally-emergent anthropogenic trends illustrates the disparity of emergence timescales for anthropogenic changes in the ocean carbon cycle. Emergence is defined as the point in time when the LE's signal-to-noise ratio for a linear trend referenced to year 1990 first exceeds a magnitude of 2, representing 95% confidence of the identification of an anthropogenic trend in the LE. Omega applies to the saturation of both aragonite and calcite forms of calcium carbonate, for which emergence times are approximately equivalent. The CaCO_3 and soft-tissue pump are calculated as the export flux at 100 meters (m) depth of CaCO_3 and particulate organic carbon, respectively. The heat content is calculated as an integral over 0–700 meters, whereas the O_2 inventories consider the integral 200–600m, and chlorophyll inventories are considered over 0–500m. NPP represents an integral over 0–100m. All other variables represent sea surface properties.

Figure 3. Global and regional time of emergence

for globally- and regionally-integrated anthropogenic signals and 50% of local anthropogenic signals for the given biogeochemical variables. Globally- and regionally-integrated signals emerge considerably sooner than local trends for most variables considered. For each domain a single domain-averaged or domain-integrated timeseries for each variable is used to compute ToE. The color of each cell corresponds to the emergence year. Dashed boxes around the 3 ocean carbon pumps. Variables same as defined in caption of Figure 2.

Figure 4. Time of Emergence and Signal Maps for the three carbon pumps.

The three pumps have local emergence times which span the 21st century, ranging in most grid-cells from (a.) 2–3 decades for the CaCO_3 pump, to (b.) 3–5 decades for the air-sea CO_2 flux, to (c.) over a century for the soft-tissue pump. The underlying, emerging signal or anthropogenic trend associated with each flux is shown in the panels d-f. Signals are defined as the mean of the 30 ensemble member's linear trends between 1990 and the ToE for each grid-cell. The corresponding maximum on the colorbar for panels d-f. is an order of magnitude smaller for (d.) than for (e.) and (f.). The maximum is $[0.02 \text{ gC m}^{-2} \text{ yr}^{-1} \text{ yr}^{-1}]$ for d., the CaCO_3 pump, and $[0.2 \text{ gC m}^{-2} \text{ yr}^{-1} \text{ yr}^{-1}]$ for e. and f., the air-sea CO_2 flux and soft-tissue pump, respectively. $[\text{NO}_3] = 0.5 \mu\text{mol kg}^{-1}$ contours imposed on panel f. to show the concurrence of counter-trends (anthropogenic increases in the soft-tissue pump) at the boundaries of surface NO_3 limitation.

Figure 5. Amplified Seasonality of $p\text{CO}_2$ ToE and Signals.

Time of Emergence for (a.) annual (c.) local summer and (e.) local winter trends in sea-air $p\text{CO}_2$ and the corresponding trend (signals) in same order (b, d, f). July-September and January-March define summer and winter for their respective hemispheres. Signals are the linear trend between 1990 and the ToE for each grid-cell. Summer and winter trends in $p\text{CO}_2$ are stronger and emerge sooner than the annual mean trend. Panel g. shows the ensemble-mean seasonal cycle of SST and DIC at 3 locations along 160°W (marked in b.) for year 1990 (solid) and year 2100 (dashed). Isolines of $p\text{CO}_2$ are superimposed under the assumption of constant salinity and alkalinity. The decreased spacing between constant $p\text{CO}_2$ isolines with increasing SST and DIC reflects the non-linearity in the buffering capacity of CO_2 in seawater. The seasonal cycle over the sub-tropical gyres (e.g. 35°N) amplifies (i.e. crosses more $p\text{CO}_2$ contours in a given year) between year 1990 and 2100, creating divergent anthropogenic trends in summer and winter $p\text{CO}_2$.

Figure 6. ToE and Signal Maps for surface vs. depth-integrated chlorophyll.

Emergence of surface chlorophyll (a.) is generally later than of depth-integrated (0–500m) chlorophyll (c.) due to the noise reduction that occurs with depth integration. For the Southern Ocean, the signal for surface (b.) and depth integrated chlorophyll (d.) are of opposite sign, with the depth-integrated signal being representative of upper-ocean declines in productivity and the soft-tissue pump. The corresponding maximum on the colorbar for panels b. and d. is two orders of magnitude larger for (b.) than for (d.). The maximum is $[2.3 \times 10^{-3} \text{ mmol m}^{-3} \text{ yr}^{-1}]$ for surface chlorophyll and $[4 \times 10^{-5} \text{ mmol m}^{-3} \text{ yr}^{-1}]$ for depth-integrated chlorophyll.