

1 **Linear trends of temperature, salinity, and oxygen in the North Pacific based on**
2 **GOBAI-O₂ over 2 decades and their controlling factors**

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14

15 **Abstract**

16 Oxygen concentrations in the ocean are believed to be declining due to global warming. However, our
17 understanding of its variability remains limited compared to physical parameters such as temperature
18 and salinity, because oxygen is difficult to observe with high spatial and temporal resolution. In this
19 study, we analyzed linear trends in potential temperature, salinity, and dissolved oxygen in the North
20 Pacific over the past two decades (2004–2023), using the GOBAI-O₂ data. We then compared oxygen
21 trends with physical parameters to investigate the spatial pattern of linear changes across the region.
22 The oxygen trends derived from GOBAI-O₂ were consistent with those observed along hydrographic
23 lines that have been relatively frequently and continuously surveyed by ship-based observations.
24 Although an overall declining trend in dissolved oxygen was evident, localized increases were
25 observed in certain density layers. By examining the associated physical conditions, we found that the
26 spatial heterogeneity of the oxygen trends can be attributed to known oceanographic processes, such

27 as the southward retreat of the oxygen minimum layer and the northward migration of a recently
28 identified front separating the subtropical and subarctic gyres. Our findings highlight the utility of
29 GOBAI-O₂ data in linking physical changes to previously unrecognized biological and biogeochemical
30 patterns in the ocean.

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32

33 **Plain Language Summary**

- 34 1. We examined the linear trends in potential water temperature, salinity, and dissolved oxygen in the
35 North Pacific over the past 20 years using GOBAI-O₂ data and explored the mechanisms behind
36 these trends.
- 37 2. Our findings confirmed that the observed trends are consistent with findings from previous studies
38 and further revealed spatial trend linkages that had only been partially recognized.
- 39 3. The results showed an overall decreasing trend in oxygen, although some localized regions
40 exhibited an increasing trend, indicating the fluctuations were not uniform in the North Pacific.
41 The spatial non-uniformity was explained by the background of physical phenomena. In particular,
42 the oxygen increase in the northern North Pacific was attributed to the northern meridional shift of
43 the front between the subtropical and subarctic gyres. Meanwhile, the retreat of the westward
44 penetration of the oxygen minimum layer in the equatorial area was associated with a weakening
45 and poleward shift of the NECC on interdecadal variability.
- 46 4. The result of this study provides one of the pieces of evidence linking physical changes to
47 previously unclear changes in biological observations.

48

49 **Keywords:** dissolved oxygen, 20-year linear trends, Bio-Argo float data, global warming

50

51 **1. Introduction**

52 Recent decades have witnessed a reduction in the global ocean's dissolved oxygen inventory, a trend
53 that is projected to continue over the course of the 21th century [Keeling et al., 2010; Breitburg et al.,
54 2018; Stramma and Schmidtko, 2021; Limburg et al., 2020; Ito et al. 2017, 2024; Kolodziejczyk et al.
55 2024]. This decline is attributed to the release of oxygen from the ocean due to oxygen solubility
56 resulting from rising surface temperatures. Additionally, enhanced stratification and stagnation of
57 ocean circulation caused by global warming are also contributing factors [Keeling et al., 2010, Bopp
58 et al. 2013; Ito et al. 2017]. Oxygen outgassing can have harmful effects on aerobic marine organisms
59 [Pörtner and Farrell, 2008; Sampaio et al., 2021] and may alter biogeochemical cycles and trigger
60 important climatological feedback [Berman-Frank et al., 2008]. Historical deoxygenation has been
61 inferred from analyses of globally distributed observation data [Helm et al., 2011; Schmidtko et al.,
62 2017; Ito et al., 2017; Takatani et al., 2012; Sasano et al., 2015; Lauvset et al., 2022b]. Earth system
63 models have been used to simulate historical and projected future deoxygenations of the oceans [Bopp
64 et al., 2013; Kwiatkowski et al., 2020; Li et al. 2020].

65

66 Oxygen trends have been observed and discussed using the discrete measurements of dissolved oxygen
67 concentration (O_2). These measurements are usually made by Winkler titration [Winkler, 1988]. These
68 titrations are also used to calibrate the readings of electrode (or more recently optical) dissolved oxygen
69 sensors attached to conductivity-water temperature-depth (CTD) profilers [Helm et al. 2011;
70 Schmidtko et al., 2017; Lauvset et al., 2022b]. **In the WOCE, CLIVAR, and GO-SHIP programs,**
71 **oxygen data have been collected globally using this approach, but their temporal resolution is on the**
72 **order of ten years**, which does not allow for a robust analysis of annual or seasonal variations in
73 **dissolved oxygen** (O_2). Some **dissolved** O_2 data are relatively frequently observed in a specific local
74 area during a year [Takatani et al. 2012, Sasano et al., 2015], but they are localized, and so it is
75 difficult to know how the phenomena are connected or disconnected each other, although some

76 research did the trials to understand the spatial variability of dissolved O₂ [Ito et al. 2017; Stramma et
77 al. 2020; Kolodziejczyk et al. 2024]. Therefore, we must have a very limited spatial-temporal
78 understanding of the observed data in dissolved oxygen.

79

80 Oxygen sensors were first installed on Argo profiling floats in the mid-2000s, and since then,
81 approximately 1,800 of these floats, equipped with oxygen sensors, have been deployed in the ocean
82 around the world, advancing efforts to deepen our understanding of trends and variability in oceanic
83 oxygen levels [Sharp et al., 2023]. The effort to establish a global biogeochemical Argo network has
84 led to the deployment of oxygen-sensor-equipped Argo floats in regions of the ocean that were
85 previously less sampled [Claustre et al., 2020]. As the number of deployed floats has grown,
86 significant progress has been made in sensor calibration, data adjustments, and quality control.
87 Noteworthy improvements include pre-deployment drift corrections [D'Asaro and McNeil, 2013;
88 Johnson et al., 2015; Bittig and Körtzinger, 2015; Bushinsky et al., 2016; Drucker and Riser, 2016;
89 Nicholson and Feen, 2017], climatology-based calibrations [Takeshita et al., 2013], in-air oxygen
90 measurement calibrations [Körtzinger et al., 2005; Bittig and Körtzinger, 2015; Johnson et al., 2015;
91 Bushinsky et al., 2016], post-deployment drift corrections [Johnson et al., 2017; Bittig et al., 2018a,
92 b], and the development of procedures for delayed-mode quality control [Maurer et al., 2021]. These
93 advancements have significantly reduced measurement uncertainty and improved the consistency of
94 optode-based [O₂] measurements from Argo floats.

95

96 Measurement data from oxygen sensors installed on the Argo float have primarily been used for
97 localized process studies. Specially, these studies have focused on air-sea oxygen exchange [Wolf et
98 al., 2018], upper-ocean primary production [Alkire et al., 2012; Estapa et al., 2019], the efficiency
99 of the biological pump [Johnson and Bif, 2021], and the dynamics of the oxygen minimum zone

100 [Udaya Bhaskar et al., 2021].

101

102 Sharp et al. [2023] recently produced a four-dimensional gridded [O₂] product, called GOBAI-O₂
103 (Gridded Ocean Biogeochemistry from Artificial Intelligence (AI) – Oxygen). GOBAI-O₂ is
104 developed using machine learning (ML), a subfield of AI that specializes in training, understanding,
105 and applying algorithms that learn and reproduce patterns artificially using data. Their analysis **shows**
106 **the possibility** that the GOBAI-O₂ data **may** reveal spatial patterns, seasonal cycles, and decadal
107 variability, with particular emphasis on the potential for these data to serve as valuable new
108 observational assets in areas where observational data gaps and high background O₂ variability overlap.

109

110 Regarding the O₂ trends in the North Pacific, Ito et al. [2017] used the objectively mapped monthly
111 climatology of O₂ based on the World Ocean Database 2013 (WOD13) [Boyer et al., 2013] at standard
112 depths from 1958 to 2015 and found multidecadal trends and variability in dissolved O₂ in the surface
113 layer. Sasano et al. [2015] detected significant warming trends and circulation changes by using the
114 JMA's high-frequency shipboard observational sections along the 137°E and 165°E lines from 1987
115 to 2011 in the western North Pacific. Their results showed that the O₂ declines in the northern
116 subtropical to subtropical-subarctic transition zones averaged $-0.45 \pm 0.16 \mu\text{mol/kg/yr}$ at $25.3 \sigma_{\theta}$ and -
117 $0.45 \pm 0.14 \mu\text{mol/kg/yr}$ at $26.8 \sigma_{\theta}$, respectively. **A significant positive trend in dissolved O₂ was**
118 **detected in the tropical Oxygen Minimum Layer (OML), with an increase of $+0.36 \pm 0.004 \mu\text{mol/kg/yr}$.**
119 Their analysis revealed the spatial heterogeneity in the observed O₂ trends. Significant and extensive
120 O₂ declines were observed in the subpolar North Pacific and off the California coast, while a modest
121 O₂ increase occurred in the mid-subtropical and western tropical North Pacific.

122

123 For horizontal broad-scale trends in the North Pacific, Stramma et al. [2020] analyzed horizontal data
124 using data from the Hydro base and the World Ocean Database bottle, and found linear trends and the
125 Pacific Decadal Oscillation (PDO) in the tropical and eastern Pacific Ocean, In particular, they found
126 that climate signals such as the North Pacific Gyre Oscillation (NPGO) in the North Pacific, and the
127 18.6-year oscillation at depths between 50 and 300 m in the equatorial and North Pacific are found, as
128 are physical fields such as water temperature and salinity. Although they collected as much data as
129 possible and analyzed them separately by region, it was difficult to ascertain the physical significance
130 of how these data trends are connected or not. **Several historical studies have revealed the existence of**
131 **O₂ changes with stronger amplitudes and non-uniform local regional patterns in specific areas [Sasano**
132 **et al. 2015; 2018; Ito et al. 2017; 2024; Stramma et al., 2020; Kolodzijejczyk et al. 2024]. These**
133 **phenomena should be linked to physical factors such as changes in circulation and stratification, as**
134 **well as to changes in biological activity that accompany physical changes in dissolved O₂, to some**
135 **extent.**

136

137 Opportunities to understand specific trends in dissolved O₂, including those associated with trends in
138 ocean temperature and salinity, are currently limited and will remain so in the future due to constrained
139 observational resources. Data such as those from GOBAI-O₂ will therefore become increasingly
140 valuable for understanding trends at basin scales. Understanding how dissolved O₂ is changing is also
141 crucial for assessing the impact on marine organisms [Breitburg et al. 2018]. Argo floats equipped
142 with oxygen sensors will continue to be deployed in the future. In this study, we present the linear
143 trends of temperature, salinity, and **dissolved O₂** using the GOBAI-O₂ data set from 2004 to 2023 in
144 the North Pacific, and how their trends connect to each other. In the discussion section, we show how
145 oxygen changes are influenced by **physical** factors such as global warming and the physical oceanic
146 **circulation** in the North Pacific.

147

148 2. Data and Methods

149 The four-dimensional gridded product (GOBAI-O₂) data on ocean interior oxygen are generated using
150 machine learning algorithms trained on dissolved O₂ data from Argo float sensors and ship-based
151 discrete observations. These data are applied to temperature and salinity distributions constructed from
152 the global Argo array [Roemmich and Gilson, 2009]. The GOBAI-O₂ data originate from the
153 combined datasets of ship-based surveys (GLODAP version 2022) and Argo float data (Argo Global
154 Data Assembly Centres), which are processed after quality control [Sharp et al., 2023]
155 (<https://doi.org/10.25921/z72m-yz67>). The combined dataset served as training input for machine
156 learning algorithms to predict dissolved O₂ [O₂], based on variables such as absolute salinity,
157 conservative temperature, potential density anomaly, hydrostatic pressure, bottom depth, and
158 additional spatiotemporal information representing geographic, seasonal, and interannual variability.
159 While biological processes are not directly incorporated into the machine learning algorithms, Giglio
160 et al. [2018] showed how incorporating spatiotemporal variables can effectively capture their
161 influence implicitly for biological processes.

162

163 The development of GOBAI-O₂ involved the use of these two types of machine learning algorithms—
164 random forest regressions and feed-forward neural networks: feed-forward neural networks (FNNs)
165 and random forest regressions (RFRs) [Breiman, 2001]. The average of the FNN and RFR estimates
166 was used as the [O₂] estimate for a given input data point. The data cover 86% of the global ocean
167 area on a 1° × 1° (latitude × longitude) grid, spanning the years 2004–2023 with a monthly resolution.
168 The vertical levels range from the ocean surface to a depth of 2 km, with 58 levels. Sharp et al. [2023]
169 evaluated 0.79 ± 0.04% per decade decrease in the oxygen inventory of the upper 2 km of the global
170 ocean over the period 2004–2022. More details of their data sources, processing, algorithm training,
171 evaluation, and uncertainty estimation are provided in Sharp et al. [2023].

172

173 The uncertainty of GOBAI-O₂ datasets [Sharp et al., 2023] is estimated by uncertainty estimates, in
174 which the individual uncertainty components are combined in quadrature (assuming independence) to
175 assess the uncertainty of dissolved O₂ data for each gridded [O₂] value:

$$176 \quad u([O_2])_{tot.} = \sqrt{u([O_2])_{meas.}^2 + u([O_2])_{grid.}^2 + u([O_2])_{alg.}^2} \quad (1),$$

177 where $u([O_2])_{meas.}^2$ represents the measurement uncertainty of the [O₂] observation themselves,
178 $u([O_2])_{grid.}^2$ reflects the gridding uncertainty arising from a single [O₂] value to represent a four-
179 dimensional box that is coarser in time and space than the resolution of many processes influencing
180 [O₂], and $u([O_2])_{alg.}^2$ corresponds to the algorithmic uncertainty from the machine-learning
181 methods used to estimate [O₂] on the grid of Roemmich and Gilson (2009) data. We use this
182 uncertainty estimate $u([O_2])_{tot.}$ and evaluate their uncertainty for estimating oxygen trends (Figs. 1–
183 4). The mean uncertainty is added when calculating the linear trend in oxygen for almost all figures in
184 this study.

185

186 The GOBAI-O₂ data are the oxygen data on 1° × 1° horizontal grid with 58 vertical levels at the
187 following depths: 2.5, 10, 30, 40, 50, 60, 70, 80, 90, 100, 110, 120, 130, 140, 150, 160, 170, 182.5,
188 200, 220, 240, 260, 280, 300, 320, 340, 360, 380, 400, 420, 440, 462.5, 500, 550, 600, 650, 700, 750,
189 800, 850, 900, 950, 1000, 1050, 1100, 1150, 1200, 1250, 1300, 1350, 1412.5, 1500, 1600, 1700, 1800,
190 1900 and 1975 m). High vertical resolution near the surface is essential for accurately capturing strong
191 temperature, salinity, density, and oxygen gradients in the mixed layer [Kara et al., 2000]. In our
192 analysis relating to the isopycnal surfaces, we interpolated the original GOBAI-O₂ data to 1-m vertical
193 resolution using cubic spline interpolation and gridded them onto a 1° × 1° × 1 m grid, because
194 interpolation makes us evaluate linear trends in potential temperature, salinity, and dissolved O₂ for
195 each latitudinal band at 1° intervals in latitude and at 0.1σ_θ intervals (Figs. 4–7). To evaluate whether

196 this interpolation procedure is appropriate or not, we also conducted sensitivity tests using multiple
197 interpolation methods (linear, shape-preserving cubic (PCHIP)) and coarser vertical grids (2m and 5m).
198 The results showed no clear differences among the three interpolation methods (Figures S1 (a, b) and
199 S2 (a, b)). A 5-m grid fails to resolve densities lighter than $24.0\sigma_\theta$ at some latitudes, but we confirmed
200 that the main features of the fields are preserved across all tested resolutions.

201

202 In Section 3.3.2, we have utilized the model outputs from the eddy-resolving OGCM for the Earth
203 Simulator (OFES) [Masumoto et al., 2004; 2010; Sasaki et al., 2008] to examine a relevant physical
204 process. The OFES model is based on the Modular Ocean Model (MOM3) [Pacanowski and Griffies,
205 2000]. The quasi-global configuration of the OFES covers 75° S– 75° N with a horizontal resolution
206 of $0.1^\circ \times 0.1^\circ$ and 54 vertical levels. The model is initialized from a state of rest using the World Ocean
207 Atlas 1998 (WOA98) [Boyer and Levitus, 1997], and spun up for 50 years with the climatological
208 forcing data of the National Centers for Environmental Prediction/National Center for Atmospheric
209 Research (NCEP-NCAR) reanalysis [Kalnay et al., 1996]. After the 50-year spin-up, a hindcast
210 experiment is conducted from 1950 to 2024 using daily atmospheric forcing from the NCEP-NCAR
211 reanalysis. In this study, we used the OFES data from 1950 to 2023.

212

213 **3. Results**

214 **3.1 Horizontal distributions of linear trends**

215 Figure 1 shows the horizontal and vertical distributions of linear trends in potential temperature,
216 salinity, dissolved O_2 , from 2004 to 2024. Positive trends in potential temperature are primarily
217 distributed in the surface layer above 200 m depth (Fig. 1a–c). Trends are relatively strong in the higher
218 latitudes. Negative trends appeared below the surface from the eastern tropical area (180° – 120° W, 5° –
219 15° N) (Fig. 1b). This area extends westward and deepens with increasing depth (Fig. 1d–f). Positive

220 and negative trends are distributed differently in the subarctic and subtropical gyres below 400 m depth.

221

222 Negative trends in salinity are generally observed throughout the surface layer (Fig. 1h–i), with
223 localized positive trends in the Kuroshio–Oyashio transition area and the northwest Pacific (140°–
224 180°E, 20°–50°N), as well as in the tropical region (120°–170°E, 0°–10°N). Some positive trends are
225 also detected in the eastern California coastal area (130°–199°W, 20°–40°N). Below 200 m depth,
226 weak positive and negative trends are evident, mirroring the trends in potential temperature (Fig. 1j–
227 k). Positive and negative trends tend to be distributed differently in the subarctic and subtropical gyres.
228 The negative trends are detected around the Alaska gyre (170°–130°W, 40°–55°N) (Fig. 1j–l), which
229 differs from the distribution of potential temperature.

230

231 Negative trends in dissolved O₂ are distributed both horizontally and zonally throughout the depth
232 range (Fig. 1o–u). Large negative trends are observed in the high latitudes and are weaker with latitude
233 near the surface (Fig. 2). The location of these trends shifts with depth. Notably, large negative trends
234 occur along the northeast coastal region (140°–130°W, 40°–50°N) and the southern region (10°–25°N)
235 within the density range of 25.2–26.8σ_θ, from 200 to 600 m depth (Fig. 1q–s). Weak positive trends
236 are partially detected around the Kuroshio–Oyashio transition area (130°–150°E, 30°–40°N) below
237 200 m depth. These positive trends extend into deeper layers and spread northeastward across the
238 North Pacific (Fig. 1r–u). Positive trends are only observed in specific regions: the tropical region at
239 100 m depth (Fig. 1p); the Alaska Gyre at 200–400 m depth (Fig. 1q–r); the western tropical region
240 at 400–600 m depth (Fig. 1r–s); and the Kuroshio–Oyashio transition region at 400–600 m depth (Fig.
241 1r–s). If we see the linear trends in dissolved oxygen by latitude, the magnitudes of the negative
242 latitudes not necessary rely on the latitudes only, because the magnitudes in the surface layer (above
243 200m depth) is shown in the higher latitude, but in case of the middle depth layer (400–600m), the

244 magnitude of negative linear trends in the middle latitude (30–40°N) is the largest. This pattern
245 completely implies the remote transports with the circulation. The mechanism of this pattern is
246 discussed more in the later sections (Section 3.3).

247

248 The total uncertainty in dissolved O₂, $u([O_2])_{tot.}$, in these layers (Fig. 1v–bb) indicates that large
249 values exist in the northern Pacific above 50°N and gradually become smaller to the south. The
250 relatively large uncertainty values are shown in the surface layer, and the steeper gradients of density
251 in the tropical eastern region [150°–120°W, 10–30°N] at 100–200m depth (Fig. 1w–x). These large
252 uncertainties typically exhibit a maximum of 100 m and tend to decrease with depth (Fig. 2 and Figure
253 A14 in Sharp et al. [2023]). Regionally large uncertainty has been shown to depend primarily on
254 algorithmic uncertainty in Figure 8 in Sharp et al. [2023] among the three components (measurement
255 uncertainty, gridding uncertainty, and algorithm uncertainty) contributing to the total uncertainty (Eq.
256 1). Algorithmic uncertainty generally decreases with increasing spatio-temporal coverage of the
257 available training data. The fact that greater regional algorithmic uncertainty is observed in the
258 northern Pacific above 50°N and in the western and eastern Pacific coastal regions below 20°N
259 suggests that the sampling data are sparse in those areas, and actually, their data sampling is low
260 (Figure 1 in Sharp et al. 2023).

261

262 Compared with the historical horizontal distributions of dissolved O₂ in the North Pacific reported by
263 Ito et al. [2017] (Figure 3 in Ito et al. 2017), our negative trend extends over a broader area. The
264 positive trend detected in the Kuroshio-Oyashio Transition Zone and the northeastern region, with a
265 density range of 26.8–27.0σ_θ (Figure 1r), was not evident in the distribution presented by Ito et al.
266 [2017]. Similarly, the positive trend observed in the tropical western Pacific below 400 m depth (Figure
267 1r–t) is relatively stronger than that identified in their analysis (Figure 3 in Ito et al. [2017]). The

268 positive trend in dissolved O₂ in the Kuroshio–Oyashio Transition Zone coincides with a region of low
269 uncertainty values (Fig. 1p–s and 1w–y), suggesting this signal may result from the high observation
270 frequency by Argo profiling floats. Other regions exhibiting positive signals—the northeastern region
271 with a density range of 26.8–27.0σ_θ (170°E–150°W, 45–55°N, Fig. 1r) and the tropical western Pacific
272 (130°–170°E, 0°–10°N, Figure 1r–t)—also correspond to areas of low uncertainty (Fig. 1y–aa).
273 Therefore, these pronounced positive trends likely represent highly reliable increases in dissolved O₂,
274 implying a robust deoxygenation signal.

275

276 Some artificially induced or unnatural expansions of the trend are also observed (e.g. 170–130°W, 0–
277 20°N), particularly in dissolved O₂ (Fig. 1q–s; Fig. 4i). The uncertainty estimates (Fig. 1v–bb) clearly
278 indicate certain locations and depths with large values, sometimes exceeding 15 μmol/kg.
279 Uncertainties are generally within 10 μmol/kg in the North Pacific. Such large uncertainties arise from
280 limited observational data and high background variability [Sharp et al. 2023]. Areas with strong
281 variability and steep background O₂ gradients tend to exhibit larger uncertainties than the global mean.
282 An additional potential source of errors is a bias in sensor calibration for Argo-O₂ observations, which
283 may arise from the finite response time of optode sensors and can cause systematic bias in the oxycline
284 regions [Bittig et al., 2014; 2018a,b]. Despite these uncertainties, our analysis shows that the observed
285 trends are spatially coherent and connected smoothly each other, and then we can relate them to well-
286 known physical mechanisms in the North Pacific. The GOBAI-O₂ dataset is, therefore, expected to
287 provide improved insights into other oceanic biogeochemical processes and physical phenomena from
288 an aerial perspective.

289

290 **3.2 Vertical sections and isopycnal density analysis of liner trends in 137°E and 165°E lines**

291 We picked up the vertical sections and isopycnal density distributions of linear trends in potential
292 temperature, salinity, and dissolved O₂ along the 137°E and 165°E lines to compare with historical
293 observational results reported by Ogata and Nonaka [2020] and Sasano et al. [2015] (Fig. 3). Ogata
294 and Nonaka [2020] used temperature and salinity data from 40 years of cruises along the 137°E line
295 between 1967 and 2009, while Sasano et al. [2015] analyzed temperature, salinity, and dissolved O₂
296 data from 25 years of cruises along the 165°E line between 1987 and 2011.) Large negative trends in
297 temperature and salinity were detected along the 25.0–26.0σ_θ isopycnal, corresponding to a potential
298 temperature of 10–12°C and salinity of 34.4–34.5. In contrast, large negative trends in dissolved O₂
299 were observed along the 26.0–27.0σ_θ isopycnal, indicating that the areas showing significant trends in
300 dissolved O₂ differ from those of temperature and salinity. Additionally, large positive trends in
301 dissolved O₂ were observed south of 15°N below 200 m depth along the 137°E line (Fig. 3c),
302 corresponding to the upper boundary of the OML. Compared with the trends in dissolved O₂ (Fig. 3c,
303 g) and associated uncertainties (Fig. 3d, h), the regions exhibiting positive or negative oxygen trends
304 do not overlap with areas of large uncertainty, which means that such oxygen trends are highly reliable
305 in this dataset.

306

307 The distributions of trends in temperature and salinity on the isopycnal surfaces show distinct
308 differences for temperatures, salinity, and dissolved O₂ (Fig. 4). The linear trends in temperature and
309 salinity are closely aligned, indicating that warming occurs with increasing salinity, while cooling
310 occurs with decreasing salinity (Fig. 4a–b, d–e). Distinct positive trends in temperature and salinity
311 are observed in the density range of 22.0–26.0σ_θ in the tropical region (5°S–5°N). No noticeable
312 trends are found in the salinity minimum region (S = 34–34.1) within the density range of 26.5–
313 27.0σ_θ. At higher latitudes (40°–50°N), large positive trends are evident in the density range of 26.0–
314 27.0σ_θ (Fig. 4e). For dissolved O₂, negative trends dominate overall; however, weak positive trends
315 are observed throughout the density range of 23.0–26.0σ_θ in low-latitude regions (5°S–5°N).

316 Additionally, large positive trends are present in the deeper density range of $26.0\text{--}27.0\sigma_\theta$ in the 5°--
317 10°N band. Weak positive trends are also detected between $10^\circ\text{--}20^\circ\text{N}$ in the density range of 23.0--
318 $25.0\sigma_\theta$ along both the 137°E and 165°E lines.

319

320 The temperature and salinity of GOBAI- O_2 data [Roemmich and Gilson, 2009; Sharp et al., 2023]
321 reveal a much wider area with negative trends in salinity in the density range of $22.0\text{--}24.0\sigma_\theta$ along the
322 137°E line compared to the results reported by Ogata and Nonaka [2020], who examined the OFES
323 data. The general characteristics of the linear trends in potential temperature and salinity align closely
324 with the findings of Sasano et al. [2015], although the current results are notably smoother,
325 particularly for dissolved oxygen. This smooth distribution may be due in large part to the smoothing
326 process applied to the data. The shipboard observations by Sasano et al. [2015] identified patchy
327 positive trends in oxygen within the density range of $24.5\text{--}27.5\sigma_\theta$ in the regions of $5^\circ\text{--}15^\circ\text{N}$ and $6^\circ\text{S}\text{--}$
328 1°N . In contrast, the current data depicts a broader, smoother, and more consistent positive trend
329 spreading in the tropical region ($-6^\circ\text{S}\text{--}5^\circ\text{N}$). The repeated shipboard observations also identified
330 localized positive trends in deeper layers within the density range of $26.5\text{--}27.5\sigma_\theta$. However, the current
331 data more clearly highlights the core of negative trends in oxygen within $5^\circ\text{--}15^\circ\text{N}$ (Fig. 3c and f).
332 These trends represent the negative signals extending along the lower isopycnals, which are
333 characteristic of the subtropical gyre.

334

335 **3.3 Horizontal distribution of linear trends along isopycnal surfaces**

336 **3.3.1 Potential temperature and salinity**

337 The horizontal distributions of linear trends in potential temperature, salinity, and dissolved oxygen on
338 specific isopycnal surfaces at 25.0 , 26.0 , and $26.8\sigma_\theta$ (Fig. 5) are illustrated to explore how these trends
339 occur and connect each other. These density surfaces correspond to the shallower densities of

340 Subtropical Mode Water (STMW), the shallower densities of Central Mode Water (CMW) [Suga
341 et al., 1997; 2004], and the representative density of North Pacific Intermediate Water (NPIW)
342 [Nakamura et al., 2000a, b; Nakamura and Awaji, 2003; Yasuda, 2004], respectively. STMW is
343 formed south of the Kuroshio Extension, between 30–35°N and 130–170°E, and penetrates to depths
344 of about 400 m in late winter. It then spreads nearly to the subtropical front through advection across
345 the Kuroshio recirculation area. CMW is formed in the transition area of the central North Pacific and
346 spreads eastward along the North Pacific Current before turning southward and westward in the
347 subtropical gyre [Suga et al., 1997; 2004]. In contrast, NPIW does not outcrop during its formation
348 process in the North Pacific. Its origin lies in Okhotsk Sea Mode Water, which is formed through
349 overturning driven by diapycnal upwelling and tidal mixing around the Kuril Islands [Nakamura et
350 al., 2000a, b; Nakamura and Awaji, 2003; You, 2003; Yasuda, 2004] as well as double diffusions in
351 the North Pacific [You, 2003].

352

353 The distributions of linear trends on the 25.0, 26.0, and 26.8 σ_θ isopycnal surfaces indicate that positive
354 and negative trends are linked to specific locations. These trends are generally distributed according
355 to the geostrophic streamlines (Fig. 5a–b, d–e, g–h). Although there are exceptions, such as the region
356 showing weak positive (150–175°E, 20–30°N) (Fig. 5a–b), negative trends in temperature and
357 salinity are predominantly observed in the western and central North Pacific on the 25.0 and 26.0 σ_θ
358 isopycnal surfaces (Fig. 5a–b, d–e). Conversely, positive trends in temperature and salinity are
359 primarily distributed in the northeastern and/or eastern side of the North Pacific along the geostrophic
360 streamlines (Fig. 5a–b, d–e). These results mean that the waters subducted in the frontal region with
361 reduced temperature and salinity originate primarily in the northeastern North Pacific and are
362 transported south along geostrophic currents (Figs 5a–b, d–e). There are exceptional areas with
363 warmer and saline waters in the northeast North Pacific (170–130°W, 40–60°N). The exceptional
364 waters affect waters that sink near the Alaska gyre and transport outside the subtropical gyre, and along

365 the California coast.

366

367 As for the water in the density range of $26.8\sigma_\theta$ (Fig. 5g-h), it exhibits large positive trends in potential
368 temperature and salinity along the Kuril Islands and moderate positive trends on the eastern side of the
369 North Pacific, respectively. Water within this density range ($26.8\sigma_\theta$) is not directly ventilated but is
370 formed through diapycnal mixing processes [Nakamura et al., 2000a, b; Nakamura and Awaji, 2003;
371 You, 2003; Yasuda, 2004] or double diffusion like a salt finger [You, 2003]. Therefore, the positive
372 trends in the North Pacific likely reflect influences from the oxygen change in the upper layer of $26.8\sigma_\theta$
373 (Fig. 5d–e and g–h).

374

375 A meridional northward shift of the outcrop line in the North Pacific due to recent climate changes
376 was detected in the analysis of the OFES data [Ogata and Nonaka, 2020]. Similar northward
377 meridional shifts of the front contour lines at 25.0 and $26.0\sigma_\theta$ are evident in the current data (Fig. 6).
378 These shifts in outcrop densities are therefore attributed to global warming and can explain the
379 observed negative trends, as less saline water from the subarctic region of the North Pacific is
380 subducted and subsequently transported southward via the subtropical circulation. The positive trends
381 in temperature and salinity occurring in the Alaska region [160 – 130°W , 30 – 60°N] (Fig. 5a–b and d–
382 e) may also be attributed to the direct influence of global warming. The $26.0\sigma_\theta$ fronts shift horizontally
383 rather than vertically between 2004 and 2024 (Fig. 6). Therefore, it should be directly affected by the
384 warming and salinity decrease due to global warming, rather than by the water level-directed spreading
385 of the water temperature-salinity distribution.

386

387 3.3.2 Dissolved oxygen

388 The linear trends in dissolved oxygen on the isopycnal surfaces at 25.0 , 26.0 , and $26.8\sigma_\theta$ exhibit

389 predominantly negative trends across the North Pacific (Fig. 5c, f, i), but the spatial distributions of
 390 these trends are not uniform. Large negative trends are concentrated in the northeastern and eastern
 391 regions, gradually decreasing toward the west (Fig. 5q, f, and i). However, there are exceptions,
 392 particularly in the tropical regions, with notable positive trends observed in the western tropical areas
 393 on the 26.0 and 26.8 σ_θ isopycnal surfaces.

394

395 The change of dissolved O₂ was decomposed with the same method of Sasano et al. [2015]. The
 396 processes and derivation underlying the equations (Eqs. 2, 3) for the change of dissolved oxygen are
 397 as follows. We tried to calculate each factor below and discussed the contributing factors for the
 398 dissolved oxygen.

399
$$\frac{\partial O_2}{\partial t} = \left(\frac{\partial O_2}{\partial z} \frac{\partial z}{\partial t} \right) + \left(\frac{\partial O_2^{sat}}{\partial t} \right)_{net} - \left(\frac{\partial(AOU)}{\partial t} \right)_{net}, \quad (2)$$

400 and consequently, they proposed the following equation:

401
$$\frac{\partial O_2}{\partial t} = \underbrace{\left(\frac{\partial O_2}{\partial z} \frac{\partial z}{\partial t} \right)}_{(i)} + \underbrace{\left(\frac{\partial O_2^{sat}}{\partial t} - \frac{\partial O_2^{sat}}{\partial z} \frac{\partial z}{\partial t} \right)}_{(ii)} + \underbrace{\left(-\frac{\partial(AOU)}{\partial t} + \frac{\partial(AOU)}{\partial z} \frac{\partial z}{\partial t} \right)}_{(iii)} \quad (3)$$

402 (i) (ii) (iii) (iv) (v) (vi)

403 $\partial O_2 / \partial t$ ($X = O_2, O_2^{sat}, AOU$ (Apparent Oxygen Utilization)) and $\partial z / \partial t$ denote the temporal
 404 rate of change in the depth of the isopycnal horizon (z), while $\partial X / \partial z$ indicates the vertical gradient
 405 of the variable X at that horizon., The latter was averaged over the past 20 years in this analysis. The
 406 term ($\partial X / \partial t$)_{net} represents the net change in a variable X. By applying Eq. (3) to the time-series
 407 data of the reconstructed O₂ data estimated from the linear regression analysis, the rate of O₂ change
 408 on each isopycnal surface (i) was attributed to various factors: (ii) the apparent effect of the
 409 deepening of isopycnal surface due to warming/freshening; (iii) the effect of O₂^{sat} change due to
 410 temperature/salinity changes; (iv) the effect of the deepening/shalowness of isopycnal surfaces with
 411 temperature/salinity changes; (v) the changes in AOU resulting from factors such as disequilibrium

412 with atmosphere, biological activities with lateral advection and/or circulation (the elapsed time after
413 the water lost contact with the atmosphere); (vi) the effect of deepening/shallowing of AOU changes
414 unrelated to O_2^{sat} . (The schematic illustrating the decomposition of oxygen change following
415 equations (2, 3) is shown in Supplementary Figure S5 after Sasano et al. [2015].)

416

417 Figure 7 shows the horizontal distributions of the magnitude of each factor contributing to the rate of
418 O_2 change on $25.0\sigma_\theta$, $26.0\sigma_\theta$, and $26.8\sigma_\theta$ surfaces. The results showed that the prominent O_2 declines
419 (Fig. 5c, f, i) are driven by multiple contributing negative and positive factors. The dominant factors
420 vary by latitude. Around the Alaska Gyre in higher latitudes ($170\text{--}130^\circ\text{W}$, $40\text{--}60^\circ\text{N}$), the primary
421 contributors to O_2 decline are the deepening of isopycnal surfaces (ii) and the shallowing of AOU
422 changes (vi) (Fig. 7f, j, k, o). The effect of O_2^{sat} change due to temperature/salinity (iii) and the AOU
423 changes (v) are affected oppositely at that time. The negative trends in the Bering Sea ($150^\circ\text{E}\text{--}170^\circ\text{W}$,
424 $50\text{--}60^\circ\text{N}$) (Figs. 5o) can be judged to be mainly affected by the AOU changes (v) (Fig. 7n).

425

426 In contrast, the O_2 decline in the subtropical and mid-latitudes ($10\text{--}40^\circ\text{N}$) is largely driven by AOU
427 changes (v) (Fig. 7d, i, and n). The relative weakening of the total O_2 decline in the western North
428 Pacific (Fig. 5c, f, i) is influenced by positive O_2 changes (Fig. 7f and k). Such positive trends are
429 driven by O_2^{sat} changes due to temperature and salinity (Fig. 7b), and the vertical movement of AOU
430 changes (Fig. 7j and o, Fig. 8b–c).

431

432 Slightly in the mid-ocean between 170°E and 160°W , the positive trend pattern shows a negative trend,
433 but in the North Pacific Ocean between 30°N and 50°N , there is a pronounced positive trend

434 continuously extending from east to west (Fig. 7i and n). This phenomenon may be related to the
435 northward meridional shift of the fronts between the subtropical and subarctic regions due to global
436 warming [Ogata and Nonaka, 2020]. With deeper winter convection, more nutrients are introduced
437 into the surface layer, potentially enhancing biological activity and leading to an increase in AOU.
438 One area of NPIW formation near the Kuril Islands exhibits negative trends in factor (iii) (Fig. 7l).
439 This result suggests weaker vertical mixing during the observational period, likely related to enhanced
440 surface-layer stratification. Supporting this, the current data show positive trends in both temperature
441 and salinity in this region (Fig. 5g–h).

442

443 The increases in dissolved O₂ are particularly prominent in the western tropical area at the density
444 range of 26.8–27.2 σ_θ (Fig. 3c and g; Fig. 4c and g; Fig. 5c, f, and i). This area overlaps with the
445 OML [Reid, 1997]. A similar tendency has also been reported by Sasano et al. [2015] and Takatani
446 et al. [2012]. In this area, the physical variability of the North Equatorial Counter Current (NECC)
447 is likely relevant. Following Chen et al. [2016], the NECC exhibits two distinct modes of variability –
448 interannual and interdecadal– based on the outputs of a multidecadal (1960–2014) ocean general
449 circulation model for the Earth Simulator (OFES). The first mode fluctuates on an interannual
450 timescale, strengthening with a southward migration. The second mode occurs on an interdecadal
451 timescale and is characterized by a progressive weakening in strength, poleward migration, and
452 broadening in width. Their analysis further suggested that, for the second mode, a decrease in wind-
453 curl intensity near the equator was responsible for the weakening, poleward migration, and widening
454 of the current.

455

456 The time-varying signals in the western tropical Pacific in the OFES data are validated by Chen et al.
457 [2016]. We then examined the OFES data (1950–2023), as well, for poleward, eastward velocities,
458 as well as potential temperature and salinity here. The potential temperature anomalies in the OFES
459 data (Fig. 9c, g) indicate that positive anomalies in 0° – 5° N occur above 250 m depth, whereas negative
460 anomalies appear along the potential density along $26.0\sigma_{\theta}$ between 5° – 20° N. A similar pattern is also
461 shown in the GOBAI-O₂ data (Fig. 3a). There is a discrepancy: Negative trends in salinity along $26.0\sigma_{\theta}$
462 are captured in GOBAI-O₂ data (Fig. 3 b), whereas positive trends in salinity along $26.0\sigma_{\theta}$ are detected
463 in the OFES model (Figs. 9b, f). This discrepancy is likely due to the higher salinity present between
464 200–600 m depth in 0 – 7° N in OFES, which differs from GOBAI-O₂ data (Fig. 10b, d).

465

466 Anomalies of the poleward and eastward velocities from 1950 to 2023 and from 2024 to 2023 (Fig.
467 9a–b, e–f) indicate that poleward velocities increase around 5° N above 200m depth, while eastward
468 velocities decrease in 0 – 4° N and increase in 5 – 7° N. The main axis of eastward velocity certainly shifts
469 poleward in 137° E lines (Fig. 11a), consistent with changes in the second mode variability of the
470 NECC detected by Chen et al. [2016]. The maximum of the poleward velocity of the NECC is also
471 shifted northward from 1964 to 2023 continuously (Fig. 11b). The broadening of NECC width was not
472 apparent in this analysis, differing from Chen et al. [2016]. This could be because we analyzed the
473 original data rather than isolating the second EOF mode of the velocities. Examination of the wind-
474 curl intensity in the equatorial region from 1950 to 2023 confirmed that the decreases in wind-curl
475 intensity along the 0° – 10° N band and the poleward broadening of the negative wind stress curl have
476 persisted to 2023 (Fig. 12).

477

478 The westward penetration of the OML is slow, occurring between two eastward-extending tongues of
479 high O₂ water originating from the western boundary around the equator area [Reid, 1997]. The O₂
480 increase at the density range of 26.8–27.2 σ_θ (Fig. 3c, g and Fig. 4c, g) could therefore be explained
481 from weakening and north-poleward shifts of the second mode of the NECC with interdecadal
482 variability. The O₂ increase (Fig. 1o-u) follows with such physical backgrounds, penetrating larger
483 dissolved O₂ from westward. The depth of the isopycnal surface becomes shallower along the equator
484 during the observational period in the boundary between the subtropical gyre and the tropical region,
485 meaning the boundary shifts to the north along 137°E line.

486

487 **4. Conclusion**

488 The oxygen variabilities in the North Pacific include the influences of global warming and climate
489 variabilities. This study utilizes the four-dimensional gridded [O₂] product [GOBAI-O₂] created
490 using artificial intelligence technology [Sharp et al. 2023] based on past potential temperature, salinity,
491 and dissolved O₂ and profiling float data to examine linear trends in temperature, salinity, and dissolved
492 O₂ over the past 20 years (2004–2023) in the North Pacific and the underlying mechanisms. The
493 linear trends in potential temperature, salinity, and dissolved O₂ are consistent with findings from
494 previous studies [Takatani et al. 2012; Sasano et al. 2015; Ogata and Nonaka, 2020], and we explained
495 how they change spatially along 137°E and 165°E (Figs. 3 and 4).

496

497 A major achievement of this study is that the GOBAI-O₂ data, being globally gridded, allowed us to
498 show that the trends are smoothly connected, spreading horizontally and vertically to the west and east,
499 presenting a reasonably reliable picture. The horizontal trend mapping over the North Pacific area
500 along the isopycnals was also clarified (Fig. 5). Notably, for the oxygen data, which has fewer
501 observation and a more fragment understanding compared to potential temperature and salinity, we

502 observed an overall decreasing trend in oxygen. However, in some areas, we also identified locally
503 increasing oxygen trends associated with warming and the northward meridional shift of subtropical
504 and subpolar fronts (Fig. 4). The increasing oxygen trends in the western equatorial region were
505 explained to be linked to the weakening of the second mode of the NECC variability. The
506 decomposition analysis of dissolved O₂ evaluated possible origins of oxygen trends at each isopycnal
507 surface (Fig. 7).

508

509 Of course, we need to keep in mind that these datasets are estimated from the available information in
510 temperature and salinity data, and available oxygen data using machine learning methods. So, caution
511 is warranted regarding the results. Future work should compare and validate these results against
512 various reanalysis datasets and other sources. Ito et al. [2024] created a similar O₂ dataset by combining
513 historical shipboard (bottle and CTD-O₂) profiles from 1965 to 2020 and biogeochemical Argo profiles
514 after 2005 using different machine learning approaches, as with the GOBAI-O₂ dataset. Cross-
515 comparisons between their dataset and ours are expected to be beneficial.

516

517 The monthly mean climatological GOBAI-O₂ data set should include the Pacific Decadal Oscillation
518 (PDO; Stramma et al., 2020; Pozo Buil and Di Lorenzo, 2017) and the North Pacific Gyre Oscillation
519 (NPGO; Stramma et al., 2020). This dataset is expected to provide a valuable basis for investigating
520 how such climate variation influences dissolved O₂ through physical driving mechanisms. These
521 challenges remain topics for future research.

522

523 **Data availability:**

524 GOBAI-O₂ data is available at [https://www.ncei.noaa.gov/access/metadata/landing-](https://www.ncei.noaa.gov/access/metadata/landing-page/bin/iso?id=gov.noaa.nodc:0259304)
525 [page/bin/iso?id=gov.noaa.nodc:0259304](https://www.ncei.noaa.gov/access/metadata/landing-page/bin/iso?id=gov.noaa.nodc:0259304). Temperature and salinity are from Roemmich and Gilson

526 (2009) Argo climatology (https://sio-argo.ucsd.edu/RG_Climatology.html). The OFES and NCEP-
527 NCAR data used in our study are obtained from APDRC, University of Hawaii ([http://apdrc.
528 soest.hawaii.edu](http://apdrc.soest.hawaii.edu)).

529

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534

535 **Author contributions:**

536 MI designed the study, and wrote the initial manuscript draft, and produced all figures. All authors
537 contributed to checking and improving the manuscript.

538

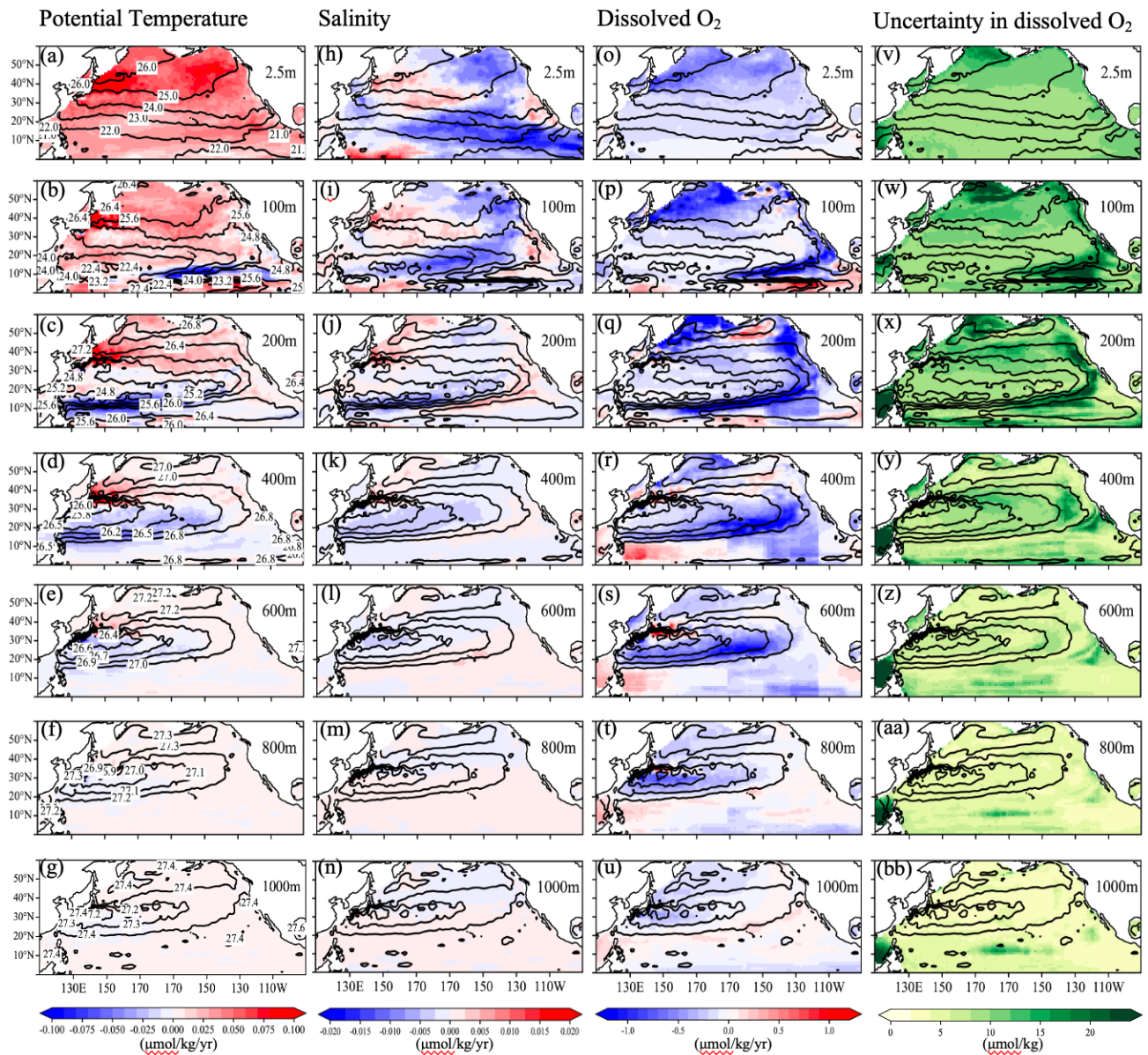
539 **Financial support:**

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542 JP22H00176.

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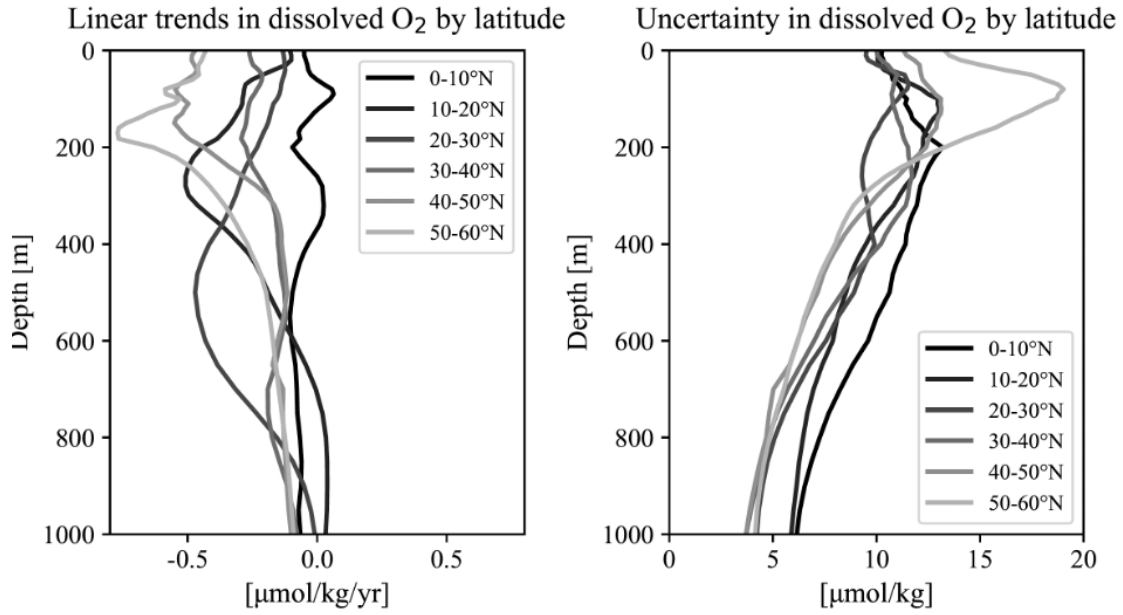
545 **Figure captions:**



546

547 **Figure 1** Horizontal distributions of linear trends ($\mu\text{mol/kg/yr}$) in (a–g) potential temperature, (h–n)
 548 salinity, and (o–u) dissolved oxygen (O_2) during the observational period at depths of 0, 100, 200, 400,
 549 600, 800, and 1000 m, respectively. Contours denote potential density at each depth. Labels for the
 550 potential density are shown only in the potential temperature sections. **Corresponding distributions of**
 551 **the mean uncertainty in dissolved O_2 ($\mu\text{mol/kg}$) are presented in panels (v–bb).**

552

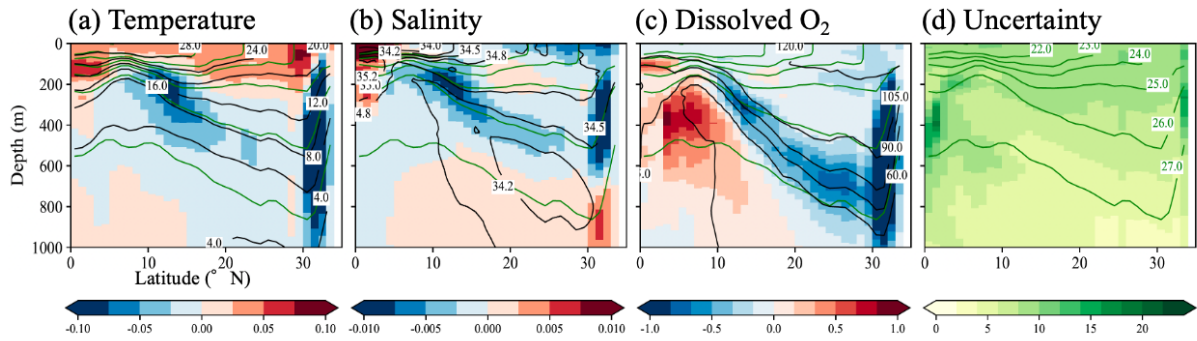


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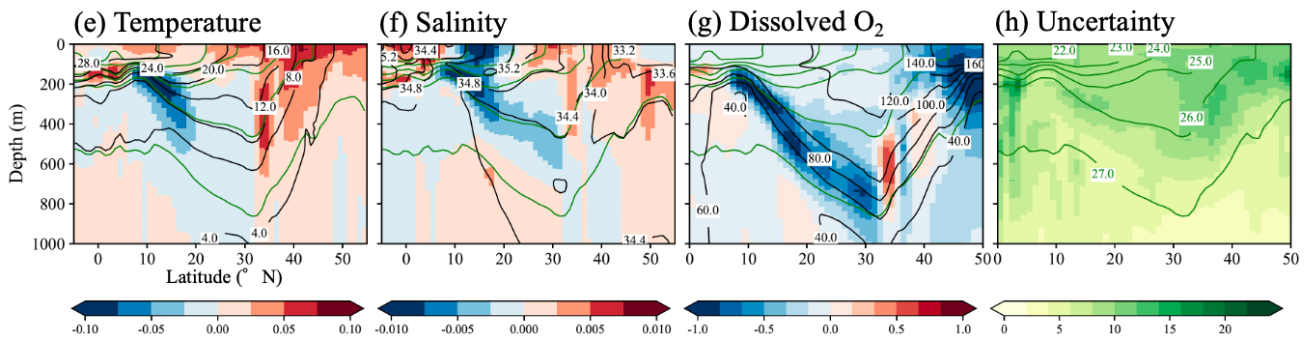
554 **Figure 2** Vertical profiles of linear trends and uncertainty in dissolved O₂ by latitude.

555

137° E line



165° E line

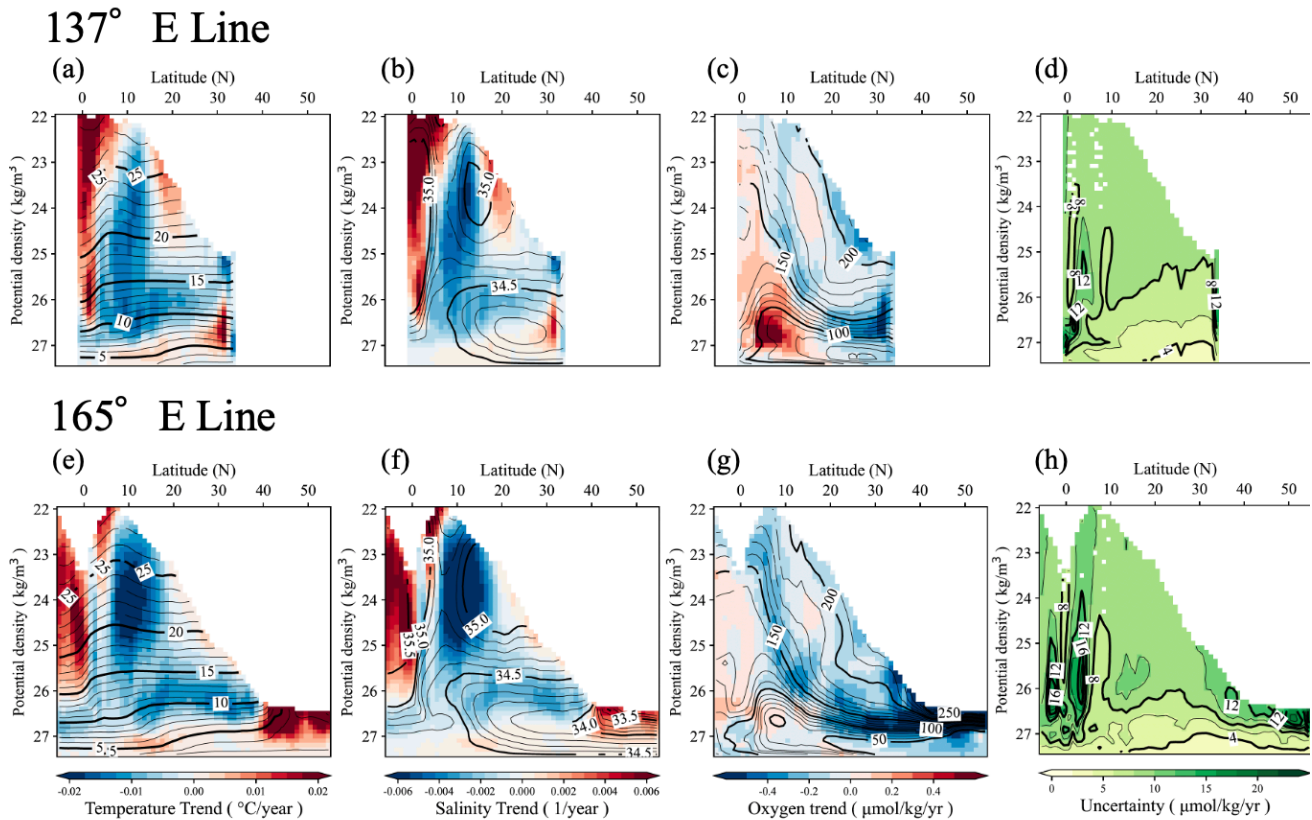


556

557 **Figure 3** Vertical sections showing linear trends in potential temperature (a, e), salinity (b, f), and
 558 dissolved O₂ (c, g) along the 137°E and 165°E meridians, respectively. **Corresponding vertical sections**
 559 **of the mean uncertainty are presented in panels (d) and (h).** Black contour lines indicate the mean

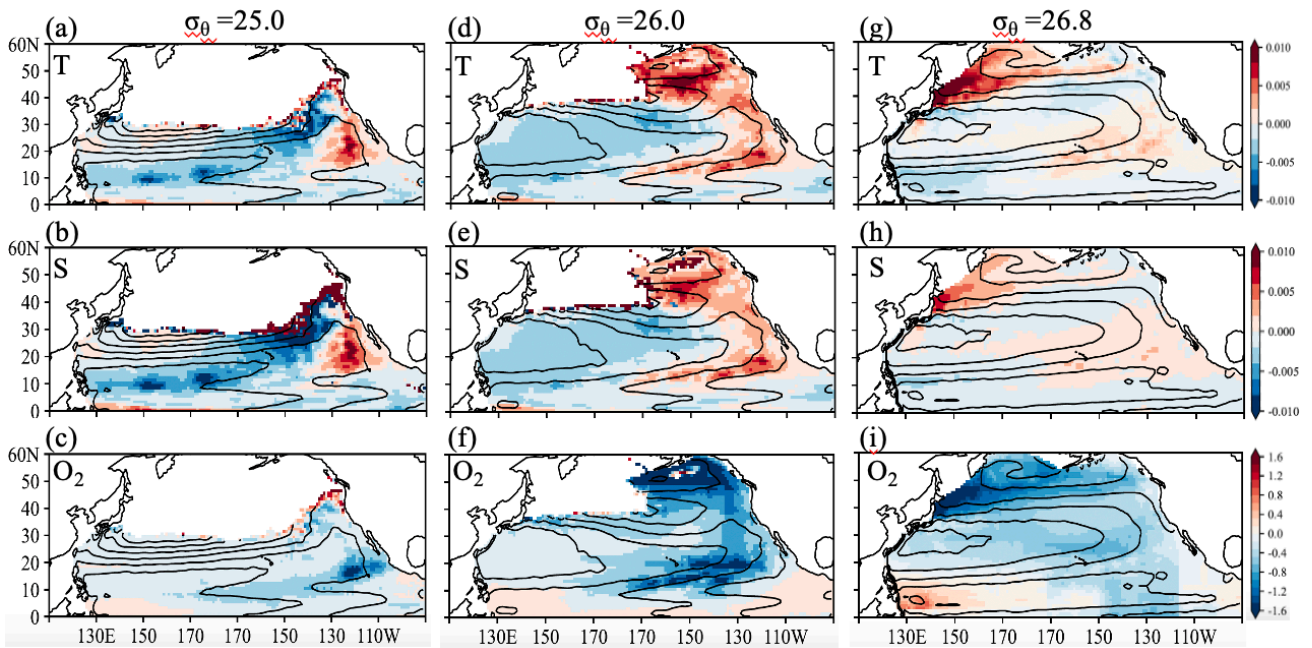
560 potential temperature (a, e), salinity (b, f), and dissolved oxygen (c, g) over the period 2004–2023,
 561 while green contour lines represent the mean potential density. Labels for the potential density are
 562 shown only in the uncertainty sections.

563



564

565 **Figure 4** Linear trends in (a, e) potential temperature, (b, f) salinity, (c, g), and dissolved O₂ on each
 566 isopycnal surface at intervals of $0.1\sigma_\theta$, calculated at every 1.0 deg of latitude in 137 °E and 165 °E
 567 lines, respectively. Contour lines represent the mean values during the target observation periods,
 568 plotted at intervals of $0.1\sigma_\theta$ for each 1 deg. of latitude. Fig. 4(d) and 4(h) are the same but show the
 569 averaged uncertainty during the target observation periods, on each isopycnal surface.

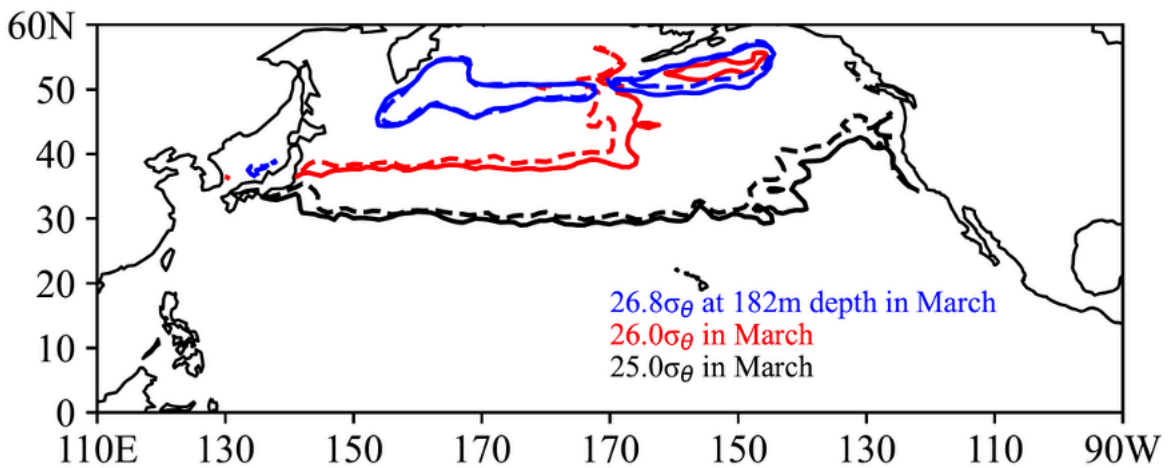


570

571 **Figure 5** Linear trends in (a) potential temperature, (b) salinity, and (c) dissolved O₂ on each isopycnal
 572 surface at 25.0, 26.0, and 26.8 σ_θ . Contour lines represent geostrophic flow streamlines on 26.0 and
 573 26.8 σ_θ surface, relative to 2000 m.

574

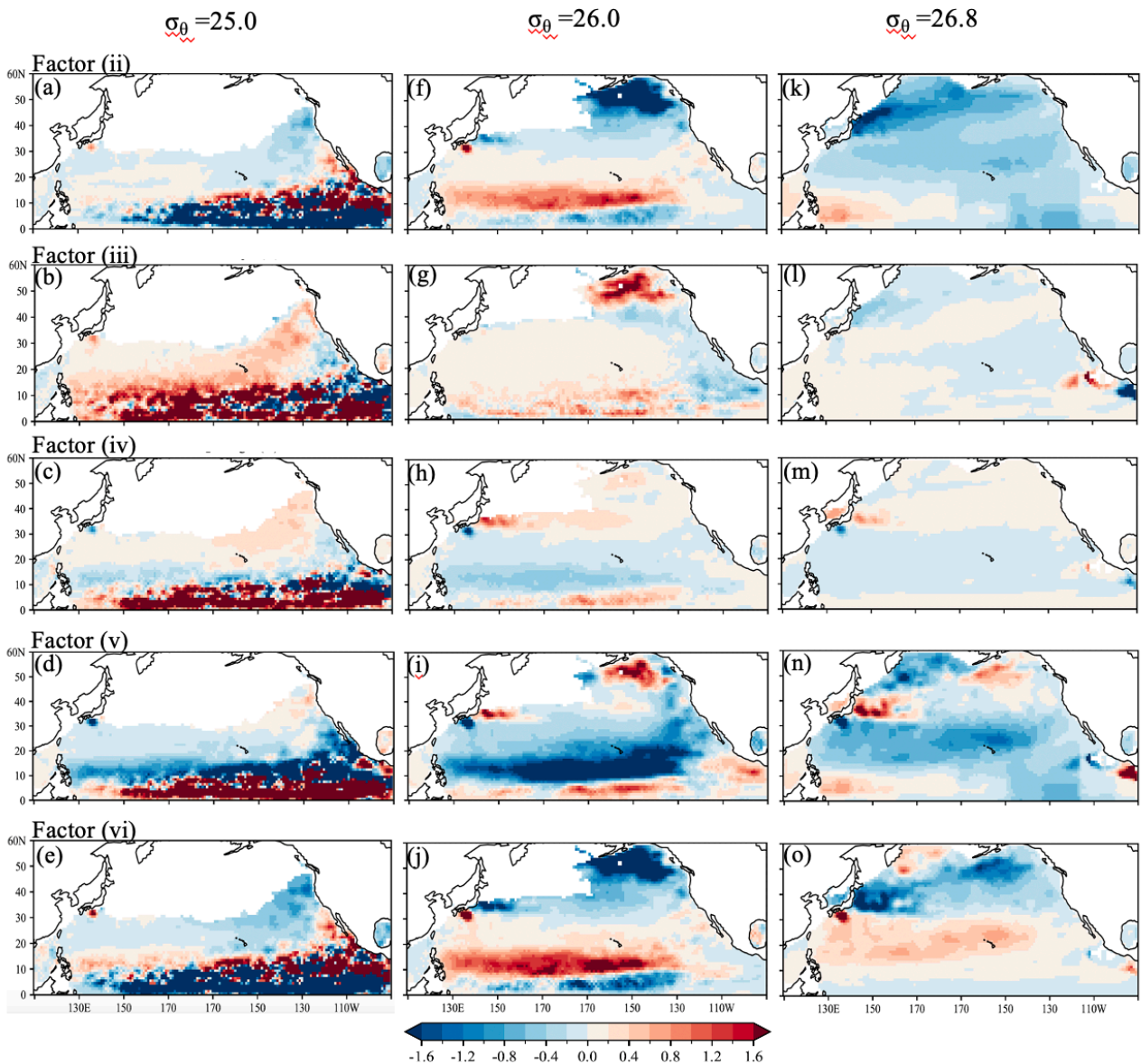
575



576

577 **Figure 6** Density contours of 25.0 σ_θ (black), 26.0 σ_θ (red), and 26.8 σ_θ (blue). Solid lines indicate
 578 the average density contours for March 2004–2009, while dashed lines represent 2019–2023.

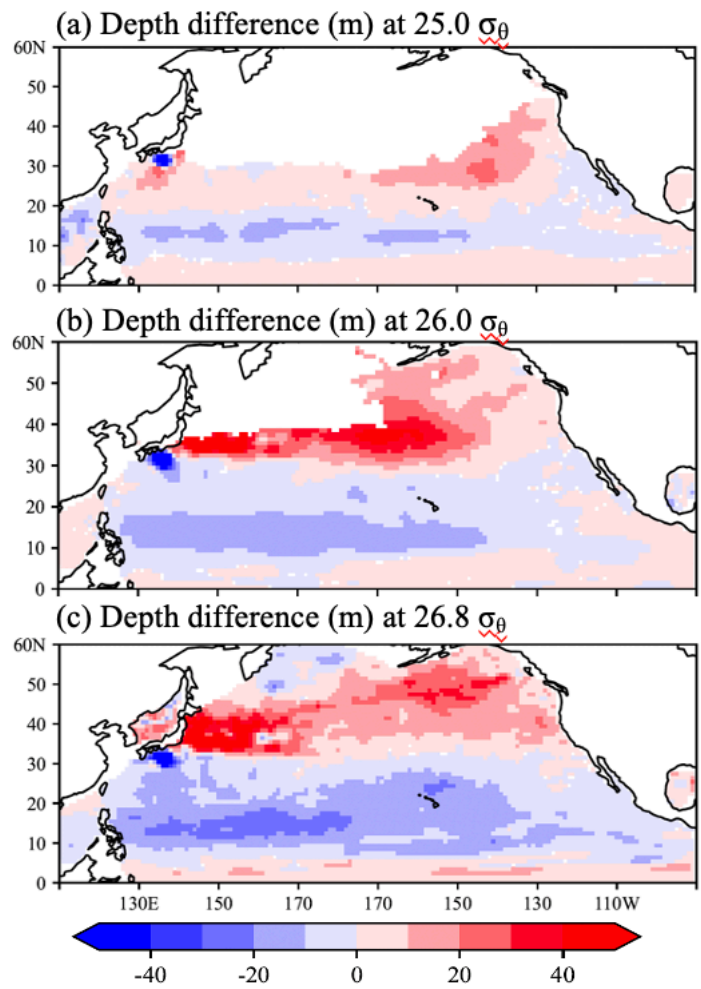
579



580

581 **Figure 7** Horizontal distributions of the magnitude of each factor contributing to the rate of O₂ change
 582 on 25.0, 26.0, and 26.8 σ_θ in Eq. (1). The rate of O₂ change on each isopycnal surface is attributed to
 583 these factors: (ii) the apparent effect of the deepening of the isopycnal surface due to warming/
 584 refreshing, (iii) the effect of O₂^{sat} change due to temperature/salinity changes, (iv) the effect of the
 585 deepening/shalowness of isopycnal surfaces with temperature/salinity changes, (v) the changes in
 586 AOU resulting from factors such as disequilibrium with atmosphere, biological activities with lateral
 587 advection and/or circulation, (vi) the effect of **vertical movement** (deepening/shallowing) of AOU
 588 changes unrelated to O₂^{sat}. This analysis was calculated with the reconstructed O₂ data estimated from
 589 the linear regression analysis.

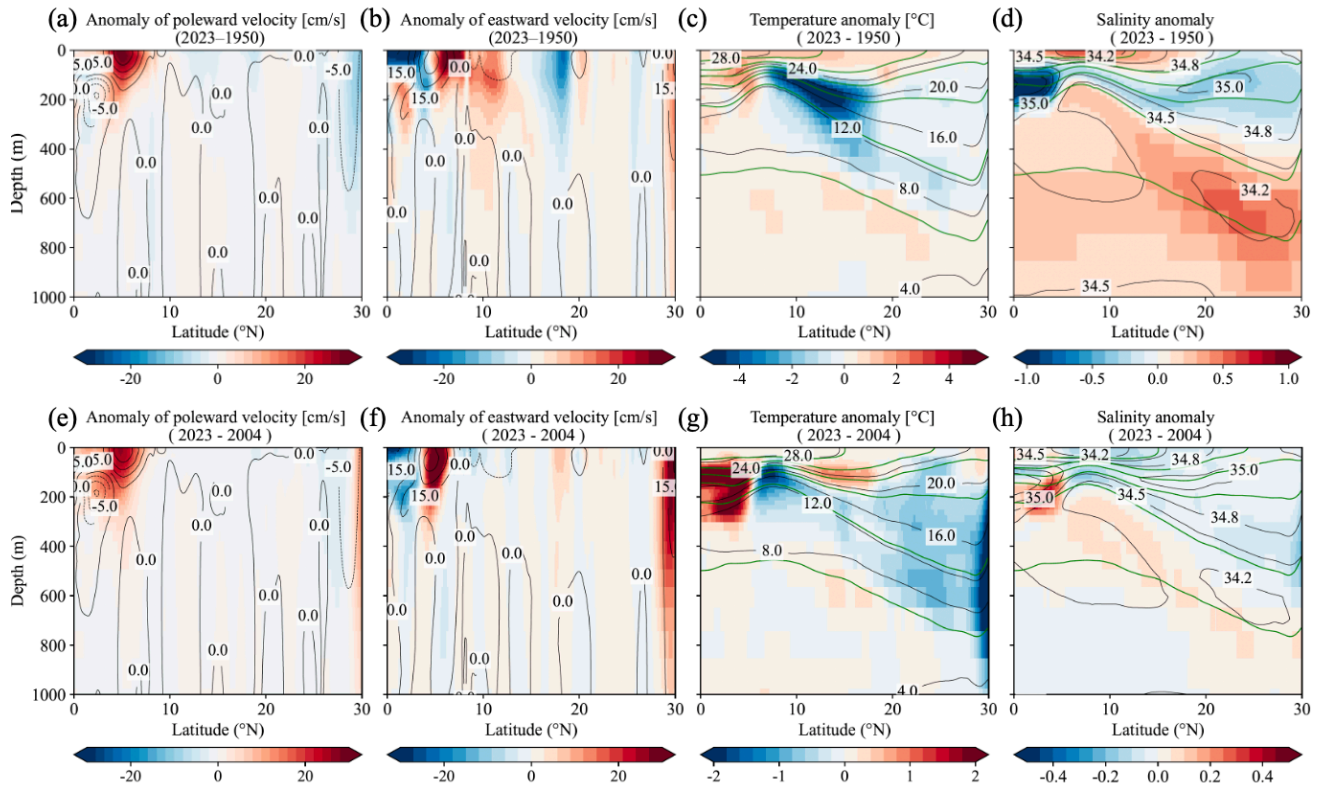
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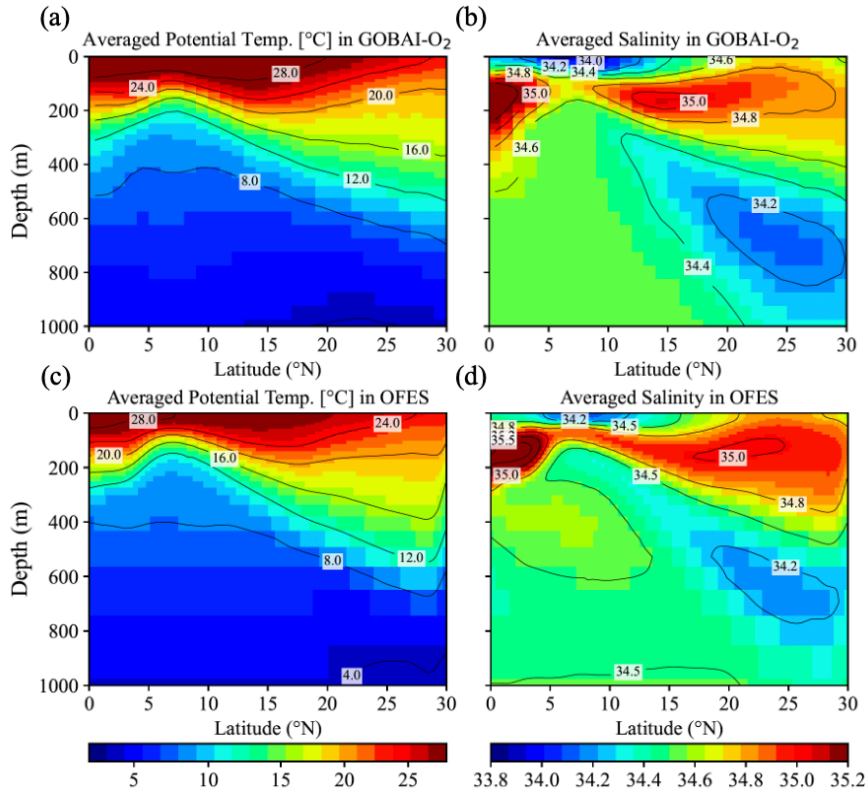
592 **Figure 8** Depth difference (m) between the 5-year averaged data in March, 2004–2009 and 2018–2023
 593 at 25.0, 26.0, and 26.8 σ_{θ} . The reconstructed O₂ data estimated from the linear regression analysis were
 594 used in this calculation. Positive and negative values indicate the deepening and shallowing,
 595 respectively, from the depth of each density in 2018–2023.

596



597

598 **Figure 9.** Anomaly of poleward and eastward velocity, potential temperature, and salinity in the OFES
 599 model outputs from 1950 to 2023 (a–d) and from 2004 to 2023 (e–h), respectively, in the 137°E line.
 600 Contours of averaged values of poleward and eastward velocity, potential temperature, and salinity
 601 during the target period are also shown in each figure. Green contour lines in (c–d, g–h) indicate the
 602 average potential density of 22, 23, 24, 25, 26, and 27 σ_{θ} , during the target periods.



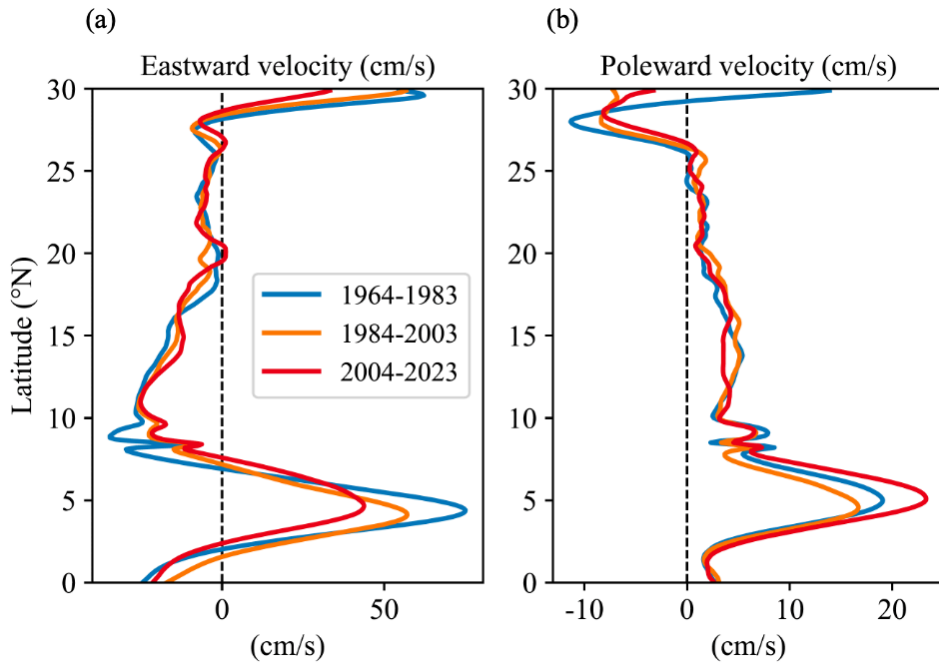
603

604

Figure 10. Averaged Potential Temperature (a, c) and salinity (b, d) in GOBAI-O₂ from 2004 to 2023

605

and OFES data from 1950 to 2023, respectively, in the 137°E line.



606

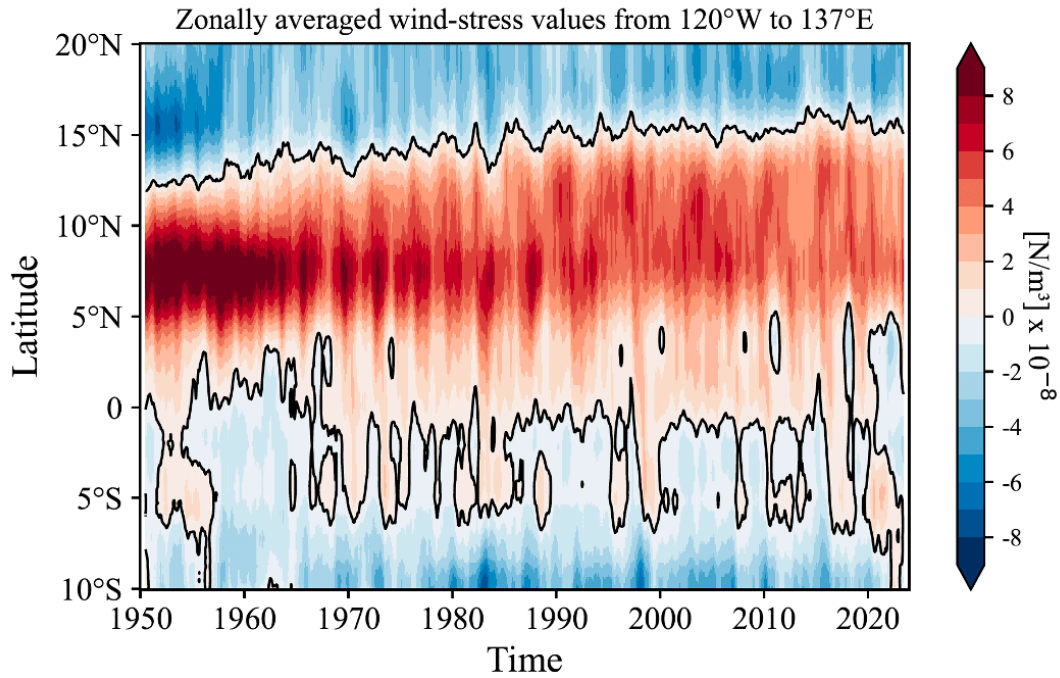
607

Figure 11. Latitudinal distribution of averaged eastward (a) and poleward velocity (b) in the OFES

608

data from 1964 to 1983, from 1984 to 2003, and from 2004 to 2023, respectively, in the 137°E line.

609



610

611 **Figure 12.** NCEP-NCAR wind-stress curl values zonally averaged from 137°E to 120° W from 1950
 612 to 2023. A 13-month running-mean filter has been applied in time.

613

614 **Appendix: Essential concepts and derivations for Equation (2-3)**

615 The essential concepts and derivation for equations (2) and (3) were originally proposed by Takatani
 616 et al. [2012] and described in detail by Sasano et al. [2015]. We follow their derivation here.

617

618 When the temperature at a depth z_A increases from θ_A to θ_A' as a consequence of ocean heat content
 619 increase, the density at that depth decreases from σ_A to σ_A' . For simplicity, the vertical profile of
 620 salinity is assumed to remain unchanged with time. As a result, the isopycnal surface of σ_A deepens
 621 from z_A to z_B (Figure S5). If surface freshening occurs simultaneously due to a net freshwater input,
 622 both the density decreases at z_A (from σ_A to σ_A') and the deepening of the isopycnal (from z_A to z_B)
 623 are enhanced. Because density is a function of temperature and salinity ($\sigma = f(\theta, S)$), the density of
 624 an isopycnal surface σ_A can be expressed as

625
$$\sigma_A = f(\theta_A, S_A) \quad (\text{before warming}) \quad (\text{C1})$$

626
$$= f(\theta_B', S_B). \quad (\text{after warming}) \quad (\text{C2})$$

627 Here, S_A and S_B denote salinity at depth z_A and z_B , respectively, and θ_B' represents the temperature at
 628 σ_A at z_B after warming. To satisfy (C1) and (C2), z_B is determined. In the region where salinity
 629 decreases with depth (e.g., above the salinity minimum layer of NPIW), $S_A > S_B$, and thus $\theta_A > \theta_B'$.
 630 This means that the potential temperature on an isopycnal surface effectively decreases as a result of
 631 warming, implying that biogeochemical properties on an isopycnal surface are also expected to change.

632

633 For a tracer X whose vertical profile with respect to depth does not change with time (e.g., salinity; see
 634 Figure S5(c)), the temporal change of X on potential density σ_A is attributed to the apparent change
 635 caused by the deepening of the isopycnal surface from z_A to z_B :

636
$$\frac{\partial X}{\partial t} = \left(\frac{\partial X}{\partial z} \cdot \frac{\partial z}{\partial t} \right) \quad (\text{C3})$$

637 Here, $\partial X/\partial t$ represents the temporal change of X observed on σ_A (gray arrows in Figure S5), z the
 638 depth at σ_A , $\partial X/\partial z$ the vertical gradient of X with respect to the depth (assumed to be time-invariant),
 639 and $\partial z/\partial t$ the rate of increase in the depth of density σ_A . $\partial X/\partial z \cdot \partial z/\partial t$ denotes the effect of
 640 deepening (illustrated as white arrows in Figure S5, corresponding to the difference between the filled
 641 square and filled circle).

642

643 Alternatively, for a variable Y whose vertical profile evolves with time and warming occurs
 644 simultaneously, the temporal change of Y at density σ_A can be expressed as the sum of two
 645 components: the temporal change associated with the deepening of the isopycnal surface from z_A to z_B
 646 and the net temporal change of Y , $(\partial Y/\partial t)_{net}$ between the time before and after warming:

647
$$\frac{\partial Y}{\partial t} = \left(\frac{\partial Y}{\partial z} \cdot \frac{\partial z}{\partial t} \right) + \left(\frac{\partial Y}{\partial t} \right)_{net} \quad (\text{C4})$$

648 To evaluate the net change $(\partial Y/\partial z)_{net}$ (illustrated by the black, blue, or pink arrows of a difference in
 649 symbols between filled square and open square in Figure S5), it is necessary to evaluate the
 650 contribution of the temporal change of Y due to the deepening of the isopycnal surface and to subtract
 651 it from the change of Y observed at density σ_A . For instance, the change of O_2^{sat} in Figure S5(f) is
 652 observed as a gray isopycnal surface (white arrow). Therefore, the net change (blue arrow) is obtained
 653 as a difference between the observed change and the deepening effect.

654

655 The dissolved oxygen concentration O_2 can be expressed as:

$$656 \quad O_2 = O_2^{sat} - AOU \quad (C5)$$

657 where O_2^{sat} is the oxygen saturation concentration (a function of temperature and salinity) and AOU
 658 is “apparent oxygen utilization”. AOU represents the oxygen consumed by biological activity since
 659 subduction. However, these contributions are typically negligible because $AOU \approx 0$ in surface waters.

660

661 According to (C4), the temporal change of O_2 at a given density for a fixed station is:

$$662 \quad \frac{\partial O_2}{\partial t} = \left(\frac{\partial O_2}{\partial z} \cdot \frac{\partial z}{\partial t} \right) + \left(\frac{\partial O_2}{\partial t} \right)_{net}. \quad (C6)$$

663 Similarly,

$$664 \quad \frac{\partial O_2^{sat}}{\partial t} = \left(\frac{\partial O_2^{sat}}{\partial z} \cdot \frac{\partial z}{\partial t} \right) + \left(\frac{\partial O_2^{sat}}{\partial t} \right)_{net}, \quad (C7)$$

665 and

$$666 \quad \frac{\partial AOU}{\partial t} = \left(\frac{\partial(AOU)}{\partial z} \cdot \frac{\partial z}{\partial t} \right) + \left(\frac{\partial(AOU)}{\partial t} \right)_{net}. \quad (C8)$$

667 The term $(\partial O_2^{sat}/\partial t)_{net}$ is directly related to a certain value with respect to warming as O_2^{sat} changes
 668 as a function of temperature. If AOU does not change with time, i.e., the change in O_2 concentration
 669 is attributed only to the change in O_2^{sat} , $(\partial(AOU)/\partial t)_{net} = 0$ and $\partial(AOU)/\partial t$ follow equation

670 (C3). On the other hand, if AOU varies with time, $\partial(AOU)/\partial t$ follows equation (C4) and
 671 $(\partial(AOU)/\partial t)_{net} \neq 0$, as shown by dashed gray line in Figure S5(g).

672

673 Since O_2 is expressed as equation (C5), the net temporal change of O_2 on an isopycnal surface is given
 674 by:

$$675 \quad \left(\frac{\partial O_2}{\partial t}\right)_{net} = \left(\frac{\partial O_2^{sat}}{\partial t}\right)_{net} - \left(\frac{\partial(AOU)}{\partial t}\right)_{net}. \quad (C9)$$

676 According to equations (C6) and (C9), the temporal change of O_2 on an isopycnal surface can be
 677 expressed:

$$678 \quad \left(\frac{\partial O_2}{\partial t}\right) = \left(\frac{\partial O_2}{\partial z} \cdot \frac{\partial z}{\partial t}\right) + \left(\frac{\partial O_2^{sat}}{\partial t}\right)_{net} - \left(\frac{\partial(AOU)}{\partial t}\right)_{net}, \quad (C10)$$

679 as described in equation (1) in the main text. Each term in equation (C10) corresponds to an arrow in
 680 Figure S5(e): From left to right, these are represented by the gray arrow, white, blue and pink arrows,
 681 respectively. The blue arrow is identical to Figure S5(f), while pink arrow corresponds to Figure S5(g),
 682 but with its direction reversed. Finally, substituting equations (C7) and (C8) into (C10) yields (C11)
 683 (corresponding to equation (2) in the main text):

$$684 \quad \frac{\partial O_2}{\partial t} = \left(\frac{\partial O_2}{\partial z} \frac{\partial z}{\partial t}\right) + \left(\frac{\partial O_2^{sat}}{\partial t} - \frac{\partial O_2^{sat}}{\partial z} \frac{\partial z}{\partial t}\right) - \left(\frac{\partial(AOU)}{\partial t} - \frac{\partial(AOU)}{\partial z} \frac{\partial z}{\partial t}\right). \quad (C11)$$

$$685 \quad \quad \quad (i) \quad \quad (ii). \quad \quad (iii). \quad \quad (iv) \quad \quad (v) \quad \quad (vi)$$

686 Note: The signs in Eq. 3(v) and (vi) are reversed from those in Equations C11(v) and (vi) for
 687 convenience.

688

689 **Data availability:**

690 GOBAI- O_2 [Sharp et al. 2022; 2023] can be accessed from NOAA's National Centers for
 691 Environmental Information at <https://accession.nodc.noaa.gov/0259304> (doi: 10.25921/z72m-yz67).

692

693 **References:**

694 Alkire, M. B., D'Asaro, E., Lee, C., Jane Perry, M., Gray, A., Cetinić, I., et al. (2012). Estimates of
695 net community production and export using high-resolution, Lagrangian measurements of O₂, NO₃⁻,
696 and POC through the evolution of a spring diatom bloom in the North Atlantic. *Deep Sea Research*
697 *Part I: Oceanographic Research Papers*, 64, 157–174. <https://doi.org/10.1016/j.dsr.2012.01.012>

698 Berman-Frank, I., Lundgren, P., & Falkowski, P. (2003). Nitrogen fixation and photosynthetic oxygen
699 evolution in cyanobacteria. *Research in Microbiology*, 154(3), 157–164.
700 [https://doi.org/10.1016/S0923-2508\(03\)00029-9](https://doi.org/10.1016/S0923-2508(03)00029-9)

701 Bittig, H. C., Fiedler, B., Scholz, R., Krahnemann, G., & Körtzinger, A. (2014). Time response of oxygen
702 optodes on profiling platforms and its dependence on flow speed and temperature. *Limnology and*
703 *Oceanography: Methods*, 12(8), 617–636. <https://doi.org/10.4319/lom.2014.12.617>

704 Bittig, H. C., & Körtzinger, A. (2015). Tackling Oxygen Optode Drift: Near-Surface and In-Air
705 Oxygen Optode Measurements on a Float Provide an Accurate in Situ Reference.
706 <https://doi.org/10.1175/JTECH-D-14-00162.1>

707 Bittig, H. C., Körtzinger, A., Neill, C., van Ooijen, E., Plant, J. N., Hahn, J., et al. (2018a). Oxygen
708 Optode Sensors: Principle, Characterization, Calibration, and Application in the Ocean. *Frontiers in*
709 *Marine Science*, 4. <https://doi.org/10.3389/fmars.2017.00429>

710 Bittig, H. C., Steinhoff, T., Claustre, H., Fiedler, B., Williams, N. L., Sauzède, R., et al. (2018b). An
711 Alternative to Static Climatologies: Robust Estimation of Open Ocean CO₂ Variables and Nutrient
712 Concentrations From T, S, and O₂ Data Using Bayesian Neural Networks. *Frontiers in Marine*
713 *Science*, 5. <https://doi.org/10.3389/fmars.2018.00328>

714 Bopp, L., Resplandy, L., Orr, J. C., Doney, S. C., Dunne, J. P., Gehlen, M., et al. (2013). Multiple
715 stressors of ocean ecosystems in the 21st century: projections with CMIP5 models. *Biogeosciences*,
716 *10*(10), 6225–6245. <https://doi.org/10.5194/bg-10-6225-2013>

717 Boyer, T. P., and S. Levitus (1997), *Objective Analyses of Temperature and Salinity for the World*
718 *Ocean on a 1/48 Grid*, vol. 11, NOAA Atlas NESDIS, Natl. Oceanic and Atmos. Admin., Silver Spring,
719 Md.

720 Boyer, T. P., Antonov, J. I., Baranova, O. K., Coleman, C., Garcia, H. E., Grodsky, A., et al. (2013).
721 *World Ocean Database 2013*. Retrieved from
722 <https://repository.oceanbestpractices.org/handle/11329/357>

723 Breiman, L. (2001). Random Forests. *Machine Learning*, *45*(1), 5–32.
724 <https://doi.org/10.1023/A:1010933404324>

725 Breitburg, D., Levin, L. A., Oschlies, A., Grégoire, M., Chavez, F. P., Conley, D. J., et al. (2018).
726 Declining oxygen in the global ocean and coastal waters. *Science*, *359*(6371), eaam7240.
727 <https://doi.org/10.1126/science.aam7240>

728 Bushinsky, S. M., Emerson, S. R., Riser, S. C., & Swift, D. D. (2016). Accurate oxygen measurements
729 on modified Argo floats using in situ air calibrations. *Limnology and Oceanography: Methods*, *14*(8),
730 491–505. <https://doi.org/10.1002/lom3.10107>

731 Chen, X., Qiu, B., Du, Y., Chen, S., & Qi, Y. (2016). Interannual and interdecadal variability of the
732 North Equatorial Countercurrent in the Western Pacific. *Journal of Geophysical Research: Oceans*,
733 *121*(10), 7743–7758. <https://doi.org/10.1002/2016JC012190>

734 Claustre, H., Johnson, K. S., & Takeshita, Y. (2020). Observing the Global Ocean with
735 Biogeochemical-Argo. *Annual Review of Marine Science*, *12*(Volume 12, 2020), 23–48.
736 <https://doi.org/10.1146/annurev-marine-010419-010956>

737 D'Asaro, E. A., & McNeil, C. (2013). Calibration and Stability of Oxygen Sensors on Autonomous
738 Floats. <https://doi.org/10.1175/JTECH-D-12-00222.1>

739 Drucker, R., & Riser, S. C. (2016). In situ phase-domain calibration of oxygen Optodes on profiling
740 floats. *Methods in Oceanography*, 17, 296–318. <https://doi.org/10.1016/j.mio.2016.09.007>

741 Estapa, M. L., Feen, M. L., & Breves, E. (2019). Direct Observations of Biological Carbon Export
742 From Profiling Floats in the Subtropical North Atlantic. *Global Biogeochemical Cycles*, 33(3), 282–
743 300. <https://doi.org/10.1029/2018GB006098>

744 Giglio, D., Lyubchich, V., & Mazloff, M. R. (2018). Estimating Oxygen in the Southern Ocean Using
745 Argo Temperature and Salinity. *Journal of Geophysical Research: Oceans*, 123(6), 4280–4297.
746 <https://doi.org/10.1029/2017JC013404>

747 Helm, K. P., Bindoff, N. L., & Church, J. A. (2011). Observed decreases in oxygen content of the
748 global ocean. *Geophysical Research Letters*, 38(23). <https://doi.org/10.1029/2011GL049513>

749 Ito, T., Minobe, S., Long, M. C., & Deutsch, C. (2017). Upper ocean O₂ trends: 1958–2015.
750 *Geophysical Research Letters*, 44(9), 4214–4223. <https://doi.org/10.1002/2017GL073613>

751 Ito, T., Cervania, A., Cross, K., Ainchwar, S., & Delawalla, S. (2024). Mapping Dissolved Oxygen
752 Concentrations by Combining Shipboard and Argo Observations Using Machine Learning Algorithms.
753 *Journal of Geophysical Research: Machine Learning and Computation*, 1(3), e2024JH000272.
754 <https://doi.org/10.1029/2024JH000272>

755 Johnson, K. S., Plant, J. N., Riser, S. C., & Gilbert, D. (2015). Air Oxygen Calibration of Oxygen
756 Optodes on a Profiling Float Array. <https://doi.org/10.1175/JTECH-D-15-0101.1>

757 Johnson, K. S., Plant, J. N., Coletti, L. J., Jannasch, H. W., Sakamoto, C. M., Riser, S. C., et al. (2017).
758 Biogeochemical sensor performance in the SOCCOM profiling float array. *Journal of Geophysical*
759 *Research: Oceans*, 122(8), 6416–6436. <https://doi.org/10.1002/2017JC012838>

760 Johnson, K. S., & Bif, M. B. (2021). Constraint on net primary productivity of the global ocean by
761 Argo oxygen measurements. *Nature Geoscience*, 14(10), 769–774. [https://doi.org/10.1038/s41561-](https://doi.org/10.1038/s41561-021-00807-z)
762 [021-00807-z](https://doi.org/10.1038/s41561-021-00807-z)

763 Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., et al. (1996). The
764 NCEP/NCAR 40-Year Reanalysis Project. Retrieved from
765 [https://journals.ametsoc.org/view/journals/bams/77/3/1520-
766 0477_1996_077_0437_tnyrp_2_0_co_2.xml](https://journals.ametsoc.org/view/journals/bams/77/3/1520-0477_1996_077_0437_tnyrp_2_0_co_2.xml)

767 Kara, A. B., Rochford, P. A., & Hurlburt, H. E. (2000). An optimal definition for ocean mixed layer
768 depth. *Journal of Geophysical Research: Oceans*, 105(C7), 16803–16821.
769 <https://doi.org/10.1029/2000JC900072>

770 Keeling, R. F., Körtzinger, A., & Gruber, N. (2010). Ocean Deoxygenation in a Warming World.
771 *Annual Review of Marine Science*, 2 (Volume 2, 2010), 199–229.
772 <https://doi.org/10.1146/annurev.marine.010908.163855>

773 Kolodziejczyk, N., Portela, E., Thierry, V., & Prigent, A. (2024). ISASO2: recent trends and regional
774 patterns of ocean dissolved oxygen change. *Earth System Science Data*, 16(11), 5191–5206.
775 <https://doi.org/10.5194/essd-16-5191-2024>

776 Körtzinger, A., Schimanski, J., & Send, U. (2005). High Quality Oxygen Measurements from Profiling
777 Floats: A Promising New Technique. <https://doi.org/10.1175/JTECH1701.1>

778 Kwiatkowski, L., Torres, O., Bopp, L., Aumont, O., Chamberlain, M., Christian, J. R., et al. (2020).
779 Twenty-first century ocean warming, acidification, deoxygenation, and upper-ocean nutrient and
780 primary production decline from CMIP6 model projections. *Biogeosciences*, 17(13), 3439–3470.
781 <https://doi.org/10.5194/bg-17-3439-2020>

782 Lauvset, S. K., Lange, N., Tanhua, T., Bittig, H. C., Olsen, A., Kozyr, A., et al. (2022b).
783 GLODAPv2.2022: the latest version of the global interior ocean biogeochemical data product. *Earth
784 System Science Data*, 14(12), 5543–5572. <https://doi.org/10.5194/essd-14-5543-2022>

785 Li, C., Huang, Jianping, Ding, L., Liu, X., Yu, H., Huang, Jiping, 2020. Increasing Escape of Oxygen
786 From Oceans Under Climate Change. *Geophysical Research Letters* 47, e2019GL086345.
787 <https://doi.org/10.1029/2019GL086345>

788 Limburg, K. E., Breitburg, D., Swaney, D. P., & Jacinto, G. (2020). Ocean Deoxygenation: A Primer.
789 *One Earth*, 2(1), 24–29. <https://doi.org/10.1016/j.oneear.2020.01.001>

790 Masumoto, Y., Sasaki, H., Kagimoto, T., Komori, N., Ishida, A., Sasai, Y., et al. (2004). A fifty-year
791 eddy-resolving simulation of the world ocean: Preliminary outcomes of OFES (OGCM for the Earth
792 Simulator). *Journal of the Earth Simulator*, 1, 35–56. [https://www.jamstec.
793 go.jp/ceist/j/publication/journal/jes_vol.1/pdf/JES1-3.2-masumoto.pdf](https://www.jamstec.go.jp/ceist/j/publication/journal/jes_vol.1/pdf/JES1-3.2-masumoto.pdf)

794 Masumoto, Y. (2010). Sharing the results of a high-resolution ocean general circulation model under
795 a multi-discipline framework—a review of OFES activities. *Ocean Dynamics*, 60(3), 633–652.
796 <https://doi.org/10.1007/s10236-010-0297-z>

797 Maurer, T. L., Plant, J. N., & Johnson, K. S. (2021). Delayed-Mode Quality Control of Oxygen, Nitrate,
798 and pH Data on SOCCOM Biogeochemical Profiling Floats. *Frontiers in Marine Science*, 8.
799 <https://doi.org/10.3389/fmars.2021.683207>

800 Nakamura, T., & Awaji, T. (2004). Tidally induced diapycnal mixing in the Kuril Straits and its role
801 in water transformation and transport: A three-dimensional nonhydrostatic model experiment. *Journal*
802 *of Geophysical Research: Oceans*, 109(C9). <https://doi.org/10.1029/2003JC001850>

803 Nakamura, T., Awaji, T., Hatayama, T., Akitomo, K., Takizawa, T., Kono, T., et al. (2000a). The
804 Generation of Large-Amplitude Unsteady Lee Waves by Subinertial K1 Tidal Flow: A Possible
805 Vertical Mixing Mechanism in the Kuril Straits. Retrieved from
806 [https://journals.ametsoc.org/view/journals/phoc/30/7/1520-
807 0485_2000_030_1601_tgolau_2.0.co_2.xml](https://journals.ametsoc.org/view/journals/phoc/30/7/1520-0485_2000_030_1601_tgolau_2.0.co_2.xml)

808 Nakamura, T., Awaji, T., Hatayama, T., Akitomo, K., & Takizawa, T. (2000b). Tidal Exchange
809 through the Kuril Straits. Retrieved from [https://journals.ametsoc.org/view/journals/phoc/30/7/1520-
810 0485_2000_030_1622_tetts_2.0.co_2.xml](https://journals.ametsoc.org/view/journals/phoc/30/7/1520-0485_2000_030_1622_tetts_2.0.co_2.xml)

811 Nicholson, D. P., & Feen, M. L. (2017). Air calibration of an oxygen optode on an underwater glider.
812 *Limnology and Oceanography: Methods*, 15(5), 495–502. <https://doi.org/10.1002/lom3.10177>

813 Ogata, T., & Nonaka, M. (2020). Mechanisms of Long-Term Variability and Recent Trend of Salinity
814 Along 137°E. *Journal of Geophysical Research: Oceans*, 125(2), e2019JC015290.
815 <https://doi.org/10.1029/2019JC015290>

816 Pacanowski, R. C., and S. M. Griffies (2000), MOM 3.0 Manual, Technical Report 4, 680 pp., Geophys.
817 Fluid Dyn. Lab., Princeton, N. J. Philander, S. G. H. (1990), El Niño, La Niña, and the Southern
818 Oscillation, pp. 293, Academic Press, San Diego, Calif.

819 Pörtner, H. O., & Farrell, A. P. (2008). Physiology and Climate Change. *Science*, 322(5902), 690–692.
820 <https://doi.org/10.1126/science.1163156>

821 Pozo Buil, M., & Di Lorenzo, E. (2017). Decadal dynamics and predictability of oxygen and
822 subsurface tracers in the California Current System. *Geophysical Research Letters*, 44(9), 4204–4213.
823 <https://doi.org/10.1002/2017GL072931>

824 Reid, J. L. (1997). On the total geostrophic circulation of the pacific ocean: flow patterns, tracers, and
825 transports. *Progress in Oceanography*, 39(4), 263–352. [https://doi.org/10.1016/S0079-](https://doi.org/10.1016/S0079-6611(97)00012-8)
826 [6611\(97\)00012-8](https://doi.org/10.1016/S0079-6611(97)00012-8)

827 Roemmich, D., & Gilson, J. (2009). The 2004–2008 mean and annual cycle of temperature, salinity,
828 and steric height in the global ocean from the Argo Program. *Progress in Oceanography*, 82(2), 81–
829 100. <https://doi.org/10.1016/j.pocean.2009.03.004>

830 Sampaio, E., Santos, C., Rosa, I. C., Ferreira, V., Pörtner, H.-O., Duarte, C. M., et al. (2021). Impacts
831 of hypoxic events surpass those of future ocean warming and acidification. *Nature Ecology &*
832 *Evolution*, 5(3), 311–321. <https://doi.org/10.1038/s41559-020-01370-3>

833 Sasaki, H., Nonaka, M., Masumoto, Y., Sasai, Y., Uehara, H., & Sakuma, H. (2008). An Eddy-
834 Resolving Hindcast Simulation of the Quasiglobal Ocean from 1950 to 2003 on the Earth Simulator.
835 In K. Hamilton & W. Ohfuchi (Eds.), *High Resolution Numerical Modelling of the Atmosphere and*
836 *Ocean* (pp. 157–185). New York, NY: Springer. https://doi.org/10.1007/978-0-387-49791-4_10

837 Sasano, D., Takatani, Y., Kosugi, N., Nakano, T., Midorikawa, T., & Ishii, M. (2015). Multidecadal
838 trends of oxygen and their controlling factors in the western North Pacific. *Global Biogeochemical*
839 *Cycles*, 29(7), 935–956. <https://doi.org/10.1002/2014GB005065>

840 Sasano, D., Takatani, Y., Kosugi, N., Nakano, T., Midorikawa, T., & Ishii, M. (2018). Decline and
841 Bidecadal Oscillations of Dissolved Oxygen in the Oyashio Region and Their Propagation to the
842 Western North Pacific. *Global Biogeochemical Cycles*, 32(6), 909–931.
843 <https://doi.org/10.1029/2017GB005876>

844 Schmidtko, S., Stramma, L., & Visbeck, M. (2017). Decline in global oceanic oxygen content during
845 the past five decades. *Nature*, 542(7641), 335–339. <https://doi.org/10.1038/nature21399>

846 Sharp, J. D., Fassbender, A. J., Carter, B. R., Johnson, G. C., Schultz, C., Dunne, J. P. (2022). GOBAI-
847 O₂: A Global Gridded Monthly Dataset of Ocean Interior Dissolved Oxygen Concentrations Based on
848 Shipboard and Autonomous Observations (NCEI Accession 0259304). NOAA National Centers for
849 Environmental Information. Dataset. <https://doi.org/10.25921/z72m-yz67>.

850 Sharp, J. D., Fassbender, A. J., Carter, B. R., Johnson, G. C., Schultz, C., & Dunne, J. P. (2023).
851 GOBAI-O₂: temporally and spatially resolved fields of ocean interior dissolved oxygen over nearly 2
852 decades. *Earth System Science Data*, 15(10), 4481–4518. <https://doi.org/10.5194/essd-15-4481-2023>

853 Stramma, L., Schmidtko, S., Bograd, S. J., Ono, T., Ross, T., Sasano, D., & Whitney, F. A. (2020).
854 Trends and decadal oscillations of oxygen and nutrients at 50 to 300 m depth in the equatorial
855 and North Pacific. *Biogeosciences*, 17(3), 813–831. <https://doi.org/10.5194/bg-17-813-2020>

856 Stramma, L., & Schmidtko, S. (2021). Tropical deoxygenation sites revisited to investigate oxygen
857 and nutrient trends. *Ocean Science*, 17(3), 833–847. <https://doi.org/10.5194/os-17-833-2021>

858 Suga, T., Takei, Y., & Hanawa, K. (1997). Thermostad Distribution in the North Pacific Subtropical
859 Gyre: The Central Mode Water and the Subtropical Mode Water. Retrieved from
860 [https://journals.ametsoc.org/view/journals/phoc/27/1/1520-](https://journals.ametsoc.org/view/journals/phoc/27/1/1520-0485_1997_027_0140_tditnp_2.0.co_2.xml)
861 [0485_1997_027_0140_tditnp_2.0.co_2.xml](https://journals.ametsoc.org/view/journals/phoc/27/1/1520-0485_1997_027_0140_tditnp_2.0.co_2.xml)

862 Suga, T., Motoki, K., Aoki, Y., & Macdonald, A. M. (2004). The North Pacific Climatology of Winter
863 Mixed Layer and Mode Waters. Retrieved from
864 [https://journals.ametsoc.org/view/journals/phoc/34/1/1520-
865 0485_2004_034_0003_tnpcow_2.0.co_2.xml](https://journals.ametsoc.org/view/journals/phoc/34/1/1520-0485_2004_034_0003_tnpcow_2.0.co_2.xml)

866 Takatani, Y., Sasano, D., Nakano, T., Midorikawa, T., & Ishii, M. (2012). Decrease of dissolved
867 oxygen after the mid-1980s in the western North Pacific subtropical gyre along the 137°E repeat
868 section. *Global Biogeochemical Cycles*, 26(2). <https://doi.org/10.1029/2011GB004227>

869 Takeshita, Y., Martz, T. R., Johnson, K. S., Plant, J. N., Gilbert, D., Riser, S. C., et al. (2013). A
870 climatology-based quality control procedure for profiling float oxygen data. *Journal of Geophysical
871 Research: Oceans*, 118(10), 5640–5650. <https://doi.org/10.1002/jgrc.20399>

872 Udaya Bhaskar, T. V. S., Sarma, V. V. S. S., & Pavan Kumar, J. (2021). Potential Mechanisms
873 Responsible for Spatial Variability in Intensity and Thickness of Oxygen Minimum Zone in the Bay
874 of Bengal. *Journal of Geophysical Research: Biogeosciences*, 126(6), e2021JG006341.
875 <https://doi.org/10.1029/2021JG006341>

876 Winkler, L. W. (1888). Die Bestimmung des im Wasser gelösten Sauerstoffes. *Berichte Der Deutschen
877 Chemischen Gesellschaft*, 21(2), 2843–2854. <https://doi.org/10.1002/cber.188802102122>

878 Wolf, M. K., Hamme, R. C., Gilbert, D., Yashayaev, I., & Thierry, V. (2018). Oxygen Saturation
879 Surrounding Deep Water Formation Events in the Labrador Sea From Argo-O₂ Data. *Global
880 Biogeochemical Cycles*, 32(4), 635–653. <https://doi.org/10.1002/2017GB005829>

881 Yasuda, I. (2004). North Pacific Intermediate Water: Progress in SAGE (SubArctic Gyre Experiment)
882 and Related Projects. *Journal of Oceanography*, 60(2), 385–395.
883 <https://doi.org/10.1023/B:JOCE.0000038344.25081.42>

884 You, Y. (2003). The pathway and circulation of North Pacific Intermediate Water. *Geophysical
885 Research Letters*, 30(24). <https://doi.org/10.1029/2003GL018561>

886

887

888