Middle to Late Holocene sea surface temperature and productivity changes in the northeast Pacific

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Key Points:

• Holocene temperature and productivity changes in the northeast Pacific are spatially variable
• Temperature change in proxies is amplified poleward, and productivity is decoupled from temperature
• Comparison with TraCE21ka suggests that multiple forcings and dynamical processes are responsible for these changes

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Abstract

Variations of the sea surface temperature (SST) and primary productivity in the northeast Pacific have far-reaching implications. In addition to influencing the regional and global temperature and hydroclimate, these conditions also control marine ecosystems and their services, which subsequently impact regional economies. Despite their importance, our understanding of the variability and controls of northeast Pacific SST and productivity on timescales exceeding observational records remains limited. Here, we use marine sediment records from seven locations, spanning 25.2°N to 59.6°N, in the northeast Pacific to characterize the millennial-scale variability of SST and productivity from 9000-1000 years BP. We further explore the dynamics of their spatiotemporal evolution and compare these data with transient climate model outputs to identify potential drivers. By undertaking a heat budget analysis and optimal fingerprinting analysis, we characterize the spatial pattern of forcings. We find that SST varied spatially in the northeast Pacific, with higher latitudes exhibiting greater magnitude changes than lower latitudes. This finding differs from previous work suggesting regional synchronicity and coherence. Our analysis did not find evidence for coherent variability of primary producer community nor carbon export, highlighting the difficulty of identifying the complex interactions between environmental conditions, producers, and carbon export. Model-proxy disagreement demonstrates the need for higher resolution model frameworks, but shows nonetheless that observed variability in the proxy records can be explained by a combination of greenhouse gas and orbital forcing. Based on our analyses, we suggest that future SST variations and marine ecosystems responses will be similarly complex in space and time.

1 Introduction

Sea surface temperature (SST) and primary productivity variations in the northeast (NE) Pacific have significant impacts on global and regional climate, marine ecosystems, and the economy of nearby regions. Within the NE Pacific, average SST controls the abundance of low level stratocumulus clouds, which in turn affects the global radiative balance (Wood, 2012). SST also interacts with the atmosphere and influences the hydroclimate and temperature of the western North America (Johnstone & Dawson, 2010; Swain et al., 2016). Furthermore, SST influences the distribution of marine ecosystems, species habitat and their abundances. Such influence have important implications on regional economy (Bond et al., 2015; Cavole et al., 2016). Aside from SST, the amount of primary productivity in this
region also affects the marine ecosystem and have significant consequences because of the ecosystem services they provide (Ware & Thomson, 2005). Additionally, the amount and composition of primary producers in this region influences the amount of carbon exported to the deep ocean (DeVries & Weber, 2017). All in all, these observations highlight the importance of understanding variability of SST and primary productivity in the NE Pacific.

Whereas seasonal to decadal (short term) SST and primary productivity in the NE Pacific are relatively well studied, changes on multidecadal and longer timescales that are beyond instrumental records (long term) in this region are less clear. Modern observations and modeling studies have allowed us to identify processes responsible for SST and primary productivity variations in this region and distinguish spatiotemporal patterns that correspond to these processes (e.g., Bograd et al., 2015; Di Lorenzo et al., 2008; Jacox et al., 2014, 2015; Johnstone & Mantua, 2014; Kahru et al., 2012; Pozo Buil & Di Lorenzo, 2017). However, the importance of each process on SST and primary productivity evolution have shown to be dependent on spatial and temporal scales (Cheung et al., 2019; Kahru et al., 2012; Long et al., 2014; Moore et al., 2018; Rykaczewski & Dunne, 2010; Xie et al., 2010). Therefore, we cannot infer future changes based solely upon our understanding of short term processes (e.g., Bakun, 1990; Sydeman et al., 2014). Although long climate model simulations can circumvent this discrepancy and provide a pathway to investigate long term changes of these variables (e.g., Alexander et al., 2018; Lotze et al., 2019), the robustness of these models remain unclear due to significant model structural uncertainty (Schlunegger et al., 2020). Specifically, they are unable to resolve mesoscale, submesoscale, boundary layer mixing, other processes important for upwelling (Capet et al., 2008; Renault et al., 2016; Xiu et al., 2018). There is also disagreement among parameterization of unresolved processes (Li et al., 2019), the different plankton functional groups represented, and nutrient transport dynamics (Fu et al., 2016). Alternatively, paleoclimate records can provide insights as to how SST and primary productivity in the NE Pacific could change on timescales that are not resolved by instrumental records, even though their spatial coverage is more limited.

The Holocene is an optimal time period to understand how SST and primary productivity in NE Pacific change on long timescales. Studying past climate intervals can help provide additional insights on how these variables change in response to external forcings (Harrison et al., 2015). Analyzing transient changes during the Holocene is particularly advantageous because of the abundance in proxy records (e.g., Kaufman et al., 2020a), relatively well constrained chronology (Reimer et al., 2020), a good understanding of the dominant external
forcings that are changing (primarily greenhouse gases and orbital), and the computational
ability to simulate transient changes using GCMs (e.g., Bader et al., 2020; Z. Liu et al.,
2009). These benefits allow us to quantify SST and primary productivity changes in the NE
Pacific on long timescales, identify processes that could be responsible behind these changes,
and determine the external forcings that caused these changes.

Multiple empirical studies have investigated how and why Holocene SST and primary
productivity in the NE Pacific have changed over millennial timescales. Syntheses of SST
records in the NE Pacific generally suggested an increase in SST since ca. 7000 yrs BP
(Barron & Anderson, 2011; Kim et al., 2004; Davis et al., 2020). Primary productivity
was also suggested to have increased at the same time (Addison et al., 2018; Barron et al.,
2003, 2018, 2019). These studies suggest that the increase in SST could be related to shifts
in atmospheric patterns; (i.e., Pacific North American pattern; Wallace & Gutzler, 1981),
whereas the primary productivity increase was driven by changes in the length of upwelling
season (Diffenbaugh & Ashfaq, 2007). Ultimately, all of these changes were proposed to be
related to changes in precession.

Even though these studies painted a relatively consistent image of how and why SSTs
in the NE Pacific change, whether these observed patterns and proposed mechanisms are ac-
curate remain debatable. Specifically, many temperature proxy records included in previous
studies either do not have reliable age constraints and temporal resolution (e.g., the record
with the highest resolution and best age constraints in Kim et al. (2004) contained only 3
$^{14}$C dates with an average sampling resolution of $\sim 132$ year) or are qualitative (e.g., Barron
& Anderson, 2011; Routson et al., 2020). In addition, the uniform increase in SST in the NE
Pacific contradicts temperature changes inferred from hemispheric-scale compilations and
the proposed insolation gradient change mechanism (Kaufman et al., 2020b; Routson et al.,
2019). Hence, to characterize SST evolution in the NE Pacific accurately and to identify
self-consistent mechanisms/drivers behind these changes, it is important to analyze reliable
SST records with high temporal resolution and good chronological constraints in tandem
with transient climate model simulations.

The primary productivity patterns observed, mechanisms proposed to explain those
changes, and the implications of these changes are also contentious. In particular, the
primary productivity records analyzed in those studies are dependent on spatial calibration
with modern observations (Lopes & Mix, 2018; Ren et al., 2014) or are subject to sediment
dilution (Gardner et al., 1997) and diagenesis (Ragueneau et al., 2000). Although the proposed mechanism to explain observed primary productivity changes, upwelling season length, is supported by simulations from an atmospheric regional model (Diffenbaugh & Ashfaq, 2007), it is unclear whether changes driven by this mechanism are detectable in proxy records. More importantly, recent modeling studies have suggested changes in higher trophic level marine species could be larger than primary producers in some regions due to alterations of food web dynamics (Lotze et al., 2019; Stock et al., 2014, 2017), and that carbon export efficiency of a specific region depends on the type of primary producer present (Fu et al., 2016; Jin et al., 2006). These results imply that impacts of primary productivity on marine ecosystems (via energy transfer) and the carbon cycle (via export production) may be better understood by characterizing changes in primary producer composition and each primary producer’s contribution to carbon export.

In this study, we aim to understand the millennial-scale evolution of SST and productivity in the NE Pacific during the Holocene, as well as the process and the external forcing behind the changes observed in the proxy records. Specifically, we address the following questions: 1) Do we observe similar temporal evolution of SST, primary producer composition, and each primary producer’s contribution to carbon export at these locations? 2) Do we expect similar dynamics to drive SST at these locations? and 3) Can we attribute changes observed in proxy records to specific forcings?

To address these questions, we present three new centennial-scale resolved alkenone-based SST and productivity records from Saanich Inlet (ODP1034), Santa Barbara Basin (MV0508-32JC), and Soledad Basin (a.k.a. San Lazaro Basin; MD02-2505/MD02-2506C²) in the NE Pacific. We complement these records with a number of published SST and productivity records. These proxies together cover the majority of the NE Pacific, where coarse resolution model simulations suggest a nearly uniform decrease in SST during the Holocene (Z. Liu et al., 2014a; Lohmann et al., 2013). We also supplement proxy records with outputs from transient climate model simulations to investigate physical processes and external forcing fingerprints. Furthermore, we revisit published productivity records to understand changes in the primary producer community (siliceous vs. alkenone synthesizing haptophytes) and their contributions towards carbon export (represented by total organic carbon).
2 Data and Methods

2.1 Proxy Records

We analyzed geochemical measurements in sediment cores collected from 7 different locations to infer changes in SST and phytoplankton productivity (Figure 1, Table 1). For convenience, we will refer PCM00-78, MD02-2505, MD02-2506C2, and GC41/PC14 as Soledad Basin, and MV0508-32JC, ODP893 as Santa Barbara Basin, as these cores were collected from these respective basins. Most measurements have been presented in previous studies (Addison et al., 2012, 2018; Arellano-Torres et al., 2019; Barron et al., 2003; A. S. Chang et al., 2008; Dean et al., 2006; Gardner & Dartnell, 1995; O’Mara et al., 2019; Praetorius et al., 2015; Stein & Rack, 1995). Readers are referred to those studies for detailed methodology. Here, we supplemented these published records with new alkenone measurements from three locations.

2.1.1 SST and productivity proxies

SST was inferred using the alkenone paleothermometer method based on the Müller et al. (1998) calibration. Alkenones are long chain ketones that are produced by *E. huxleyi* and *G. oceanica* in the open ocean. The unsaturation ratio of C37 alkenones (Uk′37) has been shown to correlate with temperature and has been applied extensively to reconstruct past SSTs (see Herbert, 2014, and references therein). Based on calibration studies, variations of Uk′37 downcore are often interpreted to best reflect mean annual SST change (e.g., Müller et al., 1998; Conte et al., 2006). However, some studies have argued that Uk′37 preferentially records warm season temperature, especially in high latitudes (e.g., Prahl et al., 2010; Max et al., 2020) or when Uk′37 disagrees with other paleothermometers (Schneider et al., 2010; Timmermann et al., 2014) and climate model simulation (e.g. Z. Liu et al., 2014a; Lohmann et al., 2013). While there is merit to consider seasonal bias in regions where coccolithophores are only produced during a limited period (e.g., Seki et al., 2007; Tsutsui et al., 2016), a synthesis of global sediment trap records does not show any systemic seasonality with flux weighted alkenone-based SST and observed mean annual SST in good agreement (Rosell-Melé & Prahl, 2013). In addition, a transect of core top samples from the California margin suggests Uk′37 is well correlated to mean annual SST and does not show the influence of the seasonal progression of upwelling (Herbert et al., 1998). Hence, we interpreted variations of
Figure 1. Observations and proxy locations. Observed averaged mean annual a) sea surface temperature in 1981-2020 (Reynolds et al., 2002), b) chlorophyll-a in 2003-2020 (NASA Goddard Space Flight Center, 2018), c) coccolithophores in 2003-2020 (Bracher et al., 2017; Losa et al., 2017), and d) diatoms in 2003-2020 (Bracher et al., 2017; Losa et al., 2017). Circles represent locations of a) SST, b) C\textsubscript{org}, c) C\textsubscript{37total}, and d) Si proxy records (indicated by core names) analyzed in this study.

To infer changes in productivity, we focused on C\textsubscript{37} alkenone concentration (C\textsubscript{37total}), biogenic silica (Si), and total organic carbon (C\textsubscript{org}) measurements and calculated the ratios between these variables to infer changes in productivity. All these proxies have been shown to be related primary productivity (Ragueneau et al., 2000; Raja & Rosell-Melé, 2021; Schoepfer et al., 2015) and have been applied to reconstruct past primary productivity (e.g., Addison et al., 2012, 2018; Barron et al., 2003; Bolton et al., 2010; Gardner et al., 1997; Praetorius et al., 2015).
Schubert et al., 1998). However, these proxies can also be influenced by diagenesis (Anderson et al., 2019; Ragueneau et al., 2000) and sediment dilution (e.g., Gardner et al., 1997). Even though diagenesis is not a primary concern here because of relatively high sedimentation rate and thus low oxygen exposure, concentrations of these biological measurements can be biased by changes in inputs of non-biogenic sediments. In this study, we instead focused on 
\[ \text{C}_{37\text{total}}/\text{C}_{\text{org}}, \text{Si}/\text{C}_{\text{org}}, \text{and Si}/\text{C}_{37\text{total}} \] ratios, where we do not need to consider the effects of changes in sedimentation rate, and thus can avoid the effects of sediment dilution. More importantly, we can understand how primary producers are changing, specifically changes in contributions of alkenone synthesizing haptophytes to total organic carbon (\[ \text{C}_{37\text{total}}/\text{C}_{\text{org}} \]), siliceous primary producers to total organic carbon (\[ \text{Si}/\text{C}_{\text{org}} \]), and the relative abundance of siliceous primary producers and alkenone synthesizing haptophytes (\[ \text{Si}/\text{C}_{37\text{total}} \]). These metrics, albeit different from primary productivity, are critical because the amount of carbon exported from the ocean surface depends on the type of primary producer (e.g., Fu et al., 2016; Jin et al., 2006), and each type of primary producer may respond differently to environmental changes (e.g., Xiu et al., 2018). We further compared \[ \text{Si}/\text{C}_{37\text{total}} \] with satellite estimates of diatoms and coccolithophores (Bracher et al., 2017; Losa et al., 2017) to determine if proxy records resemble patterns in observations qualitatively.

2.1.2 Chronology

The chronology of EW0408-85JC, MD02-2496, ODP1019, TN062-O550, GC41/PC14, PCM00-78C/PCM00-78K was based on \(^{14}\)C dates. Details of \(^{14}\)C ages used in these published records can be found in Davies-Walczak et al. (2014), Cosma et al. (2008), Barron et al. (2003), Barron et al. (2018), Marchitto et al. (2010), and O’Mara et al. (2019). We re-calibrated all these \(^{14}\)C ages using the latest radiocarbon age calibration curves (Table S1; Reimer et al., 2020; Heaton et al., 2020) and constructed the age-depth model using BACON (Blaauw & Christen, 2011). The chronology for ODP 893 used in this study is generated by stratigraphic correlation with an adjacent core MV0811-14JC in Santa Barbara Basin (Du et al., 2018). The chronology for MV0811-14JC is based on high-resolution \(^{14}\)C dates and variable reservoir age, so the latest radiocarbon age calibration curve would not alter it significantly.

The chronologies of sediment cores analyzed in this study (ODP1034, MV0508-32JC, MD02-2505 and MD02-2506C\(^2\)) were based on a combination of \(^{14}\)C dates and stratigraphic correlation. For ODP1034, we first spliced cores from different holes into a composite depth
Table 1. Information of sediment cores analyzed in this study. Check mark indicates the proxy is included in this study.

<table>
<thead>
<tr>
<th>Site Name</th>
<th>Lat (°)</th>
<th>Lon (°)</th>
<th>avg. sed. rate (mm/yr)</th>
<th>avg. SST resolution (yr/sample)</th>
<th>no. of 14C dates over 10 kyr BP</th>
<th>$U_{37}^{C}$</th>
<th>C$_{37}$total</th>
<th>C$_{org}$</th>
<th>Si</th>
<th>References</th>
</tr>
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<tbody>
<tr>
<td>EW0408-85JC</td>
<td>59.6</td>
<td>-144.2</td>
<td>0.62</td>
<td>166</td>
<td>15</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
<td>Addison et al. (2012) Praetorius et al. (2015)</td>
</tr>
<tr>
<td>MD02-2496</td>
<td>49.0</td>
<td>-127.0</td>
<td>1.19</td>
<td>—</td>
<td>6</td>
<td>—</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
<td>A. S. Chang et al. (2008)</td>
</tr>
<tr>
<td>ODP1034</td>
<td>48.6</td>
<td>-123.5</td>
<td>9.11</td>
<td>76</td>
<td>42</td>
<td>✓</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>This study</td>
</tr>
<tr>
<td>ODP1019</td>
<td>41.6</td>
<td>-124.9</td>
<td>0.52</td>
<td>134</td>
<td>7</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
<td>Barron et al. (2003)</td>
</tr>
<tr>
<td>TN062-O550</td>
<td>40.9</td>
<td>-124.6</td>
<td>0.99</td>
<td>—</td>
<td>10</td>
<td>—</td>
<td>✓</td>
<td>✓</td>
<td>✓</td>
<td>Addison et al. (2018)</td>
</tr>
<tr>
<td>MV0508-32JC</td>
<td>34.3</td>
<td>-120.0</td>
<td>1.00</td>
<td>57</td>
<td>—</td>
<td>✓</td>
<td>✓</td>
<td>—</td>
<td>—</td>
<td>Gardner and Dartnell (1995) Stein and Rack (1995)</td>
</tr>
<tr>
<td>ODP893</td>
<td>34.3</td>
<td>-120.0</td>
<td>0.94</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>✓</td>
<td>—</td>
<td>✓</td>
<td>O’Mara et al. (2019)</td>
</tr>
<tr>
<td>MD02-2505</td>
<td>25.2</td>
<td>-112.7</td>
<td>3.16</td>
<td>45</td>
<td>4</td>
<td>✓</td>
<td>✓</td>
<td>—</td>
<td>—</td>
<td>This study</td>
</tr>
<tr>
<td>/MD02-2506C²</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GC41/PC14</td>
<td>25.2</td>
<td>-112.7</td>
<td>1.09</td>
<td>—</td>
<td>18</td>
<td>—</td>
<td>—</td>
<td>✓</td>
<td>✓</td>
<td>Dean et al. (2006)</td>
</tr>
<tr>
<td>PCM00-78C/PCM00-78K</td>
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<td>-112.7</td>
<td>1.12</td>
<td>1.80</td>
<td>8</td>
<td>✓</td>
<td>✓</td>
<td>—</td>
<td>—</td>
<td>Arellano-Torres et al. (2019)</td>
</tr>
</tbody>
</table>


following Bornhold et al. (1998). Then, we used $^{14}$C dates provided in Bornhold et al. (1998) and Blais-Stevens et al. (2001) to construct the age-depth model using BACON. Massive layers were treated as instantaneous events and were removed in BACON (Blaauw & Christen, 2011). For MV0508-32JC, we identified gray layers using core images and used these layers as tie points to correlate with the stratigraphy presented in Du et al. (2018). For MD02-2506C$^2$, we correlated the core with PCM00-78K using Ca X-ray fluorescence measurements. We then extrapolated the PCM00-78 age model to estimate ages in MD02-2506C$^2$. For MD02-2505, we relied on $^{14}$C measurements from Rodríguez-Sanz et al. (2013), which were only available for the early Holocene portion, and extrapolated the age-depth relationship towards the core-top using BACON (Blaauw & Christen, 2011).

The number of radiocarbon dates, sedimentation rate, and sampling resolution of these cores provide a good basis to analyze millennial scale variability (Table 1). In a Bayesian age-depth model framework, the number of radiocarbon dates included strongly influences the uncertainty in the age-depth model (Blaauw et al., 2018). Even though the density of radiocarbon dates in some sediment cores used in this study remains low (e.g., 4 in MD02-2505/MD02-2506C$^2$), which results in large uncertainty (max. 95% confidence interval range = 2157 years), BACON tends to overestimate the age uncertainty (Trachsel & Telford, 2017). Nevertheless, the number of radiocarbon dates available in each core and the Bayesian age-depth model approach represent an improvement in chronological constraint compared to previous synthesis studies (e.g., Kim et al., 2004; Leduc et al., 2010a). The sampling resolution of these cores ($\sim$ 200 yr) should also minimize millennial-scale aliasing, so our results should provide an accurate representation of how SST has changed in the NE Pacific on millennial timescales.

### 2.1.3 Laboratory Method

Sediments were freeze dried and lipids from $\sim$0.1-1.5g of sediments were extracted using Dionex 200 or 350 Accelerated Solvent Extractor with 100% Methylene Chloride (DCM). Then, total lipid extracts of sediments from ODP1034 and MV0508-32JC were separated by silica gel column chromatography into non-polar, ketone, and polar fractions using Hexane, DCM, and Methanol as eluents, respectively. The ketone fraction of ODP1034 was further separated by silver nitrate column chromatography using DCM and ethyl acetate as eluents separately. Afterwards, ketone fractions from MV0508-32JC and fractions obtained using ethyl acetate from ODP1034 were dried and reconstituted with toluene and n-
hexatriacontane (C₃₆) and n-heptatriacontane (C₃₇) alkane standards. These samples were analyzed on an Agilent 6890 Gas Chromatograph coupled with a flame ionization detector (GC-FID) and a poly (trifluoropropylmethylsiloxane) stationary phase column (VF200-ms; Longo et al., 2013). We also analyzed an internal standard to determine reproducibility of our results. We defined analytical uncertainty as 1 standard deviation of replicate internal standards measurements, and found that analytical uncertainty for $U'_{37}$ was $\sim$0.002 units, whereas the relative standard deviation for C₃₇ concentration to be $\sim$5%.

### 2.1.4 Data Analysis

We calculated the $U'_{37}$ index of each sample and inferred the temperature using the calibration in Müller et al. (1998). We inferred the C₃₇ alkenone concentration (C₃₇:3+C₃₇:2) by normalizing the peak areas of C₃₇:3 and C₃₇:2 alkenone peaks against the C₃₆ and C₃₇ spikes that were co-analyzed in each sample.

To facilitate proxy-model comparison and to account for analytical and chronological uncertainties, we computed an ensemble of 200 year binned composites for each proxy record from 9000 to 1000 years before present (yrs BP) using a Monte Carlo approach. The 200 year window was selected to minimize the extrapolation needed to estimate SST changes ($U'_{37}$ temporal resolution: $< 200$ years; Table 1). We focused on changes between 9000-1000 yrs BP because some proxy records do not have measurements for the last millennium and/or prior to ca. 9000 yrs BP. To create a 200 year composite of a proxy record, we first estimated the proxy value, determined from a normal distribution (with mean = measured value, standard deviation = analytical uncertainty). Second, we selected an age-depth relationship from the posterior distribution derived from BACON. Third, we averaged the values within each 200 year bin. This was repeated 1000 times to obtain an ensemble. The analytical uncertainties of published records were obtained from the original study (Table 1). In cases where the analytical uncertainty was not reported, we used the most conservative estimate of uncertainty reported in other records.

We compared the five SST records that we focused on in this study with Holocene SST latitudinal bands computed using SST records that were synthesized in Kaufman et al. (2020a) to determine if these five records follow large scale changes. These synthesized records include inferred SSTs from $\delta^{18}O$, alkenone, and Mg/Ca (Andrews et al., 1999; Antonarakou et al., 2015; Arz et al., 2003; Bard et al., 2000; Barron et al., 2003; Benway et
al., 2006; Bolliet et al., 2011; Cacho et al., 1999, 2001; Came et al., 2007; Castañeda et al.,
2004, 2010; F. Chang et al., 2015; Doose-Rolinski et al., 2001; Elmore et al., 2015; Emeis &
Dawson, 2003; Emeis et al., 2003; Eynaud et al., 2009; Fan et al., 2018; Farmer et al.,
2008; Flower et al., 2004; Fraser et al., 2014; Harada et al., 2006; Herbert & Schuffert, 2000;
Hill et al., 2006; Hillaire-Marcel et al., 1994; Huguet et al., 2006; Ijiri et al., 2005; Isono et
al., 2009; Keigwin & Jones, 1995; Keigwin et al., 2005; Kennett et al., 2007; M. Kienast
et al., 2001; Kim et al., 2004, 2007, 2012; Kristjánsdóttir et al., 2017; Kubota et al., 2010;
Lea et al., 2003; Marchitto et al., 2010; Martrat et al., 2003, 2007, 2014; McClymont et
al., 2012; Minoshima et al., 2007; Mohtadi et al., 2014; Moossen et al., 2015; Overpeck et
al., 1996; Pelejero et al., 1999; Praetorius et al., 2015; Rühlemann et al., 1999; Riethdorf et
al., 2013; Rigual-Hernández et al., 2017; Rodrigo-Gámiz et al., 2014; Rodrigues et al., 2010;
Rosenthal et al., 2003; Sachs, 2007; Saraswat et al., 2013; Schmidt et al., 2004; Schmidt &
Lynch-Stieglitz, 2011; Schmidt et al., 2012a, 2012b; Schwab et al., 2012; Sejrup et al., 2011;
Shintani et al., 2011; Staubwasser et al., 2003; Steinke et al., 2008; Stott et al., 2007; Sun
et al., 2005; Thornalley et al., 2010; Tierney et al., 2016; Tiwari et al., 2015; Weldeab et
al., 2005, 2007; Werner et al., 2013; Yamamoto et al., 2013; Zhao et al., 1995; Ziegler et al.,
2008). We followed the approach in Routson et al. (2019) and calculated the bootstrapped
zonal average SST over four latitudinal bands: 60°–80° N, 40°–60° N, 20°–40° N, 0°–20° N.
Additionally, we calculated the Pacific and Atlantic only zonal average SST to determine if
the temporal patterns are synchronous between the two basins.

2.2 Transient Climate Model Simulations

We analyzed 9000-1000 yrs BP interval of outputs from the Transient Climate Evolution
of the last 21000 years (TraCE21ka; Z. Liu et al., 2009) experiments to understand potential
drivers of SST evolution shown in proxy records. TraCE21ka simulations were run using the
Community Climate System Model version 3 (W. D. Collins et al., 2006). In the fully coupled
experiment, greenhouse gases, orbital parameters, ice sheets, and meltwater were set as
boundary forcings and varied from last glacial maximum (LGM) to present. In single forcing
experiments, only the target forcing was allowed to evolve while the remaining forcings were
set constant at LGM conditions. Even though CCSM3 is known to have an average ~
1.5°C SST bias in ocean eastern boundaries (Large & Danabasoglu, 2006), previous studies
focusing on the Pacific have found agreement between proxies and TraCE21ka regarding
the evolution of El Nino (Z. Liu et al., 2014b) and also have used TraCE21ka to investigate
the impacts of Pacific Meridional Overturning Circulation on intertropical convergence zone shift during last deglaciation (W. Liu & Hu, 2015).

We carried out a heat budget analysis to determine the drivers of SST variation and an optimal fingerprinting analysis to understand spatial fingerprints of each forcing on SST. All analyses were done using decadally averaged annual mean data since outputs with higher temporal resolution were not available. Further, since proxy records were unevenly sampled over time, unless otherwise specified, we averaged and/or integrated model results over 200 years to yield more direct comparison with proxy records. Additionally, we focused on model simulation results between 9000-1000 yrs BP, the same time window analyzed in proxy records.

### 2.2.1 Heat Budget Analysis

Assuming a well mixed surface ocean, SST variations can be understood through a mixed layer heat budget. This budget can be written as:

\[
\frac{\partial T}{\partial t} = \frac{Q_{\text{net}}}{\rho c_p H} - \nabla \cdot u T - w \frac{\partial T}{\partial z} + \text{residual} \quad (1)
\]

where \(T\) = potential temperature, \(t\) = time, \(Q_{\text{net}}\) = net air sea heat flux, \(\rho\) = seawater density, \(c_p\) = specific heat capacity, \(H\) = pre-defined depth of the mixed layer, \(\nabla\) = horizontal divergence operator, \(u\) = horizontal velocity, \(w\) = vertical velocity, \(z\) = vertical depth, and residual = temperature tendency that could not be explained by the other three terms. \(Q_{\text{net}}\) is further estimated following:

\[
Q_{\text{net}} = Q_{\text{shf}} - Q_{\text{sw}} \left[ R \exp \left( \frac{-H}{\gamma_1} \right) + (1 - R) \exp \left( \frac{-H}{\gamma_2} \right) \right] \quad (2)
\]

where \(Q_{\text{shf}}\) = surface heat flux, \(Q_{\text{sw}}\) = shortwave heat flux, \(H\) = pre-defined depth of the upper ocean, \(R, \gamma_1, \gamma_2\) = are constants for type I water following Paulson and Simpson (1977).

Two additional steps were taken to better compare our heat budget results with proxy records. First, since proxy records are in °C and results from heat budget analysis are in temperature tendency (°C/s), we integrated the heat budget over time with a 200 year timestep and estimated the temperature change relative to 1000 yrs BP. Second, instead of comparing each proxy record with the heat budget of one grid cell, we compared each proxy record with the heat budget of a region. A grid cell was included in the region if
the correlation between its SST and the grid cell nearest to the proxy record was $> 0.85$.

We labelled these regions using the latitude of the grid cell nearest to the proxy record as opposed to the site name because grid cells and core sites are not identically collocated (see Table 1).

We carried out a singular value decomposition (SVD) on all the heat budget terms (see Eq. 1) across the five locations (H). This allowed us to determine if the heat budget was balanced similarly at each location for a given temporal pattern. This was accomplished following:

$$ H = U \cdot S \cdot V^T $$

where $U =$ principal components, $S =$ singular values, and $V =$ eigenvectors (or EOFs) of the data matrix $H$ with all the heat budget terms from the five target locations.

### 2.2.2 Optimal Fingerprinting Analysis

To understand the spatial fingerprint of each forcing onto SST fields, we carried out an optimal fingerprinting analysis on TraCE21ka simulations. Identification of spatial fingerprints from an external forcing is generally posed as:

$$ Y = Xa + u $$

where in our study, $Y =$ SST from full forcing simulation, $X =$ SST from each single forcing experiment, $a =$ scaling factors of each forcing, and $u =$ internal variability. Since SSTs from single forcing experiments contain both forced signals and internal variability, the true forced signal is not known exactly. To account for error in variables due to internal variability, a total least squares (TLS) regression approach was adopted for this analysis to solve Eq. 4. This method has been widely adopted in modern climate change detection and attribution studies (e.g. Allen & Stott, 2003; Hegerl & Zwiers, 2011, and references therein).

We followed Allen and Stott (2003) to obtain the scaling factor of each forcing, which represents the contribution of the specific forcing to total change during the analyzed time period ($a$ in Eq. 4). First, we defined a data matrix $Z$ where it includes mean centered and pre-whitened $X,Y$:

$$ Z = \begin{bmatrix} \frac{X_{anom}}{\sqrt{\text{var}(X_{anom})}} & \frac{Y_{anom}}{\sqrt{\text{var}(Y_{anom})}} \end{bmatrix} $$

where $X_{anom}, Y_{anom} =$ mean centered $X,Y$ respectively and $\text{var}(\cdot) =$ variance.
Next, we decomposed $Z$ from Eq. 5 using SVD. Since we seek to obtain the most distinct pattern of each forcing from full forcing simulations, we identified the singular vector ($V$) that corresponds to the smallest singular value (thus eigenvalues) in Eq. 3, and re-scaled the eigenvector such that the element that corresponds to $Y = -1$. The remaining elements of the vector correspond to the scaling factor of each forcing. We estimated the statistical significance of each scaling factor to 95% confidence level using a bootstrap method following DelSole et al. (2019).

3 Results

Our SST records are interpreted to reflect mean annual SST. In all the alkenone based SST records, the most recent estimates are in close agreement with modern observed mean annual SST ($\sim 1^\circ$C difference), with all but EW0408-85JC showing a slightly cooler estimate in proxies (Figure 2). If alkenones based SST were to primarily reflect warm season temperature, all of our records would be warmer than modern observation. But this is not the case. As for EW0408-85JC, even though previous studies interpreted alkenone based SST to primarily reflect summer temperature (Praetorius et al., 2015; Prahl et al., 2010), the difference between the alkenone SST and observed summer SST still exists, suggesting summer season temperature is not the only possible explanation to such difference. Furthermore, a comparison between core top $U_{37}^NC$ SST and observed mean annual SST reveals no systematic bias across latitude (Figure S1). As such, it is reasonable to interpret our proxy records as mean annual SST.

The five 200 year averaged $U_{37}^NC$ SST composite records display varying temporal patterns with different magnitudes of changes (Figure 2). The variability of Soledad Basin and MV0508-32JC records is small over the analysis period and show a small increase in SST from early to mid-Holocene to late Holocene ($\sim 0.89^\circ$ for Soledad Basin and $\sim 1.46^\circ$C for MV0508-32JC). On the other hand, EW0408-85JC, ODP1034 and ODP1019 all showed an initial decrease in SST ($\sim 1.4^\circ$C) between 9000 yrs BP and mid Holocene (ca. 7000-4000 yrs BP) before a recovery with similar magnitude towards the present (with the exception of ODP1034, where the SST change since the mid-Holocene was $< 1^\circ$C). A comparison with Pacific only, Atlantic only, and all basin zonally averaged SST suggests that these five proxy records from the NE Pacific generally follow patterns of Pacific zonal averaged SST. However, these records diverge from the all basin and Atlantic zonal average SST in $40^\circ - 60^\circ$ N and $20^\circ - 40^\circ$ N (Figure 3).
Figure 2. $U_{37}^{\delta}$ based SST reconstruction. $U_{37}^{\delta}$ based SST estimates from a) EW0408-85JC, b) ODP1034, c) ODP1019, d) Santa Barbara Basin, e) Soledad Basin. Lines with circles represent the dataset at original resolution, thick lines represent 200-year binned composites, and shaded areas represent 95% confidence interval of each 200-year composites, which accounts for chronological and analytical uncertainties. The circle indicates modern mean annual SST estimate based on OISST, and the error bar indicates 95% confidence interval based on bootstrap averaging.
Figure 3. Zonally averaged SST anomaly. Zonally averaged SST anomaly (with mean SST 9000-1000 yrs BP removed) at a) 60°–80°N, b) 40°–60°N, c) 20°–40°N, and d) 0–20°N. Left panel shows temporal evolution of SST in all ocean basins (blue), Pacific (red), Atlantic (orange), and in the three new study sites. Records that are included in these stacks are listed in Table S2. Shaded areas are uncertainties related to sampling. Right panel shows the number of marine records available (blue bars) and the number of records available in the Pacific (red) and Atlantic (orange).
Although the ratio between productivity indicators is qualitatively consistent with modern observations, the magnitude of changes in the productivity records vary by location. Observation suggests elevated diatom productivity in high latitudes relative to low latitude and high coccolithophore productivity along the western north America coast (Figure 1c-d). The most recent Si/C$_{\text{total}}$ values from the five sediment records provide a consistent image – the ratio is highest in high latitude and lowest in low latitude (Figure 4). We also observe small Si/C$_{\text{org}}$ temporal variations at all of our study sites. The absolute value of Si/C$_{\text{org}}$ appears to be highest in high latitudes and lowest in low latitudes (Figure 5). Similarly, we find small total/C$_{\text{org}}$ temporal variations at all of our study sites. The absolute value of total/C$_{\text{org}}$ is highest in low latitudes and decrease towards high latitude (Figure 6). These results suggest siliceous productivity contributes most to carbon export at high latitudes whereas calcareous productivity contributes most to carbon export at low latitudes, and that their contributions did not change significantly throughout the Holocene.

Figure 4. Si/total ratio. 200 year composites of ratio between Si and total at a) EW0408-85JC, b) ODP1019, and c) Soledad Basin. Shaded areas represent 95% confidence interval of each 200-year composite, which accounts for chronological and analytical uncertainties.
Figure 5. Molar Si/C$_{org}$ ratio. 200 year composites of molar ratio between Si and C$_{org}$ at a) EW0408-85JC, b) MD02-2496, c) ODP1019, d) TN062-O550, and e) Soledad Basin. Shaded areas represent 95% confidence interval of each 200-year composite, which accounts for chronological and analytical uncertainties.
Figure 6. total/C\textsubscript{org} ratio. 200 year composites of ratio between C\textsubscript{37total} and C\textsubscript{org} at a) EW0408-85JC, b) ODP1019, c) Santa Barbara Basin, and d) Soledad Basin. Shaded areas represent 95% confidence interval of each 200-year composite, which accounts for chronological and analytical uncertainties.

SST evolution in TraCE21ka simulations exhibit different temporal patterns from proxy records, which impedes our ability to make direct inference about mechanisms that drive SST changes shown in proxy records. Specifically, 200 year TraCE21ka based SST composites all show smaller (< 1°C) changes during the analysis window (Figure 7) compared to proxy records (Figure 2). Nevertheless, we analyzed 3 EOFs of the heat budgets to determine if different processes are responsible for driving SST variability at each location. Although the first EOF explains most of the variance across all of the heat budget components (~ 99%)
variance), including more EOFs improves the representation of variations shown in the heat budget (Figure S1).

![Image](image.png)

**Figure 7.** SST in TraCE21ka. a-e) timeseries of averaged SST of the grid nearest to the proxy site and neighboring grids with correlation > 0.85 with the nearest grid. Thick lines are 200 year composites. Right panel shows the map with the grid cell nearest to the proxy site (square) and the grids with correlation > 0.85 with that grid (contours). Background color of the map represents average of mean annual SST between 9000-1000 yrs BP.

Optimal fingerprinting analysis on TraCE21ka simulations suggests the fingerprint of each external forcing on SST is spatially heterogeneous. Specifically, greenhouse gas forcing has a positive imprint throughout the NE Pacific, but a negative one in lower latitudes, though with less certainty, as indicated by the insignificant scaling factors (Figure 9a). On the other hand, orbital forcing, primarily from precession and obliquity, is negatively related to SST at all sites in the NE Pacific, but positively related in low latitudes. The fingerprint is somewhat uncertain between 20°-40°N (Figure 9b). Only small regions display significant fingerprints of ice sheet and freshwater forcings (Figure 9c-d).
Figure 8. EOFs of heat budget at proxy locations. Shown are EOF loadings of each component at the proxy locations for a) EOF1, b) EOF2, and c) EOF3. Also shown are the principal components representing the temporal evolution of these EOFs.

Figure 9. Optimal fingerprinting analysis. Spatial fingerprints (scaling factors) of each forcing on SST fields on 200 year timescales derived from TLS regression. Gray stipplings represent statistically significant fingerprint at 95% confidence level based on bootstrapping.
4 Discussion

4.1 Changes in proxy records

The five SST records presented in this study provide new insights on the evolution of SST in the NE Pacific during the Holocene. Whereas previous studies focused on linear changes in the NE Pacific during the Holocene (e.g., Kim et al., 2004; Leduc et al., 2010a), our results highlight the additional complexity in the temporal evolution of SST in the NE Pacific, especially in mid-latitudes. These records together suggest a dynamic NE Pacific (e.g., Cheung et al., 2019) instead of a region where changes are expected to be linear and with amplitudes scale with latitude. The three new records in this study also provide better constraints on changes in SST than previously published records which are either subject to age uncertainty (S. S. Kienast & McKay, 2001), or have proxies which are influenced by other environmental factors (Kennett et al., 2007), or seasonal bias (Marchitto et al., 2010).

The temporal evolution of the SST records shown in this study also challenges previous attempts to reconcile proxy-model and proxy-proxy disagreements. Previous studies used the strong correlation between alkenone based SST records and winter/summer insolation as the primary evidence to support the idea that alkenone records exhibit a dominant seasonal bias (Bova et al., 2021; Leduc et al., 2010a; Z. Liu et al., 2014a; Schneider et al., 2010). As such, alkenone records had to be seasonally adjusted before comparing with climate models or other proxy records. Given that maximum vertical transport and temperature occurs during warm season in the NE Pacific (e.g., Jacox et al., 2018), the evidence used in previous studies would suggest that our records are likely to be biased to summer, and that they should show a negative temperature trend (with higher amplitude in higher latitude) that corresponds to summer insolation. Yet, this is not the case. With the exception of ODP1034, which shows a $\sim 1^\circ$C cooling, none of the records examined here exhibit a cooling trend (Figure 2), in contrast to many alkenone-based reconstructions from the North Atlantic (Figure 3). These results indicate that it is insufficient to use the correlation between proxy records and insolation as evidence that proxies are seasonally biased and that it may not be appropriate to correct proxy records based on their correlation with insolation in hope to reconcile proxy-model and proxy-proxy disagreements.

Our SST records further highlight the shortcomings in current synthesis and compilation studies. When compared to studies of Holocene zonally averaged temperature trends, which suggest an increasing meridional temperature gradient towards the present,
our records show the opposite trend (Figure 3; Kaufman et al., 2020a; Routson et al., 2019). Additionally, our records further confirm the contrasting temperature trends between the Pacific and the Atlantic mid latitudes (orange vs red in Figure 3). These results together point to a sampling bias towards the Atlantic in hemispheric scale temperature compilations and reconstructions, and that an increase in proxy network spatial density is needed (Judd et al., 2020).

In contrast to SST, our productivity records suggest little to no consistent changes in primary producer community and their contributions to export productivity. Even though modern studies suggest that changes in primary producer community would pose significant impact on higher trophic level species (Stock et al., 2014), our Si/total records do not display any coherent changes over time (Figure 4). There are also no coherent patterns in the contributions of coccolithophore and diatoms to carbon export (Figure 5-6). Hence, despite suggestions of an increase in productivity since the mid Holocene (Addison et al., 2018; Barron et al., 2003, 2018, 2019), our data provide no evidence for a change in primary producer community composition and subsequent export productivity.

4.2 Mechanisms behind changes in proxy records

Although proxy records allow us to capture the spatiotemporal evolution of past climates and ecosystems, climate model simulations are needed to help infer physical mechanisms behind these observed changes. Unfortunately, SST from TraCE21ka simulations do not capture the SST temporal patterns in the NE Pacific, nor is CCSM3 capable of simulating ocean biogeochemistry. Therefore, we cannot make direct inferences regarding the mechanisms driving SST and productivity changes. Nevertheless, TraCE21ka simulations remain useful in hypothesis testing.

The spatially variable SST temporal patterns shown in our records raise the question of whether similar physical processes govern the whole NE Pacific. Previous studies have suggested that changes in the persistence of Pacific North American pattern are responsible for SST changes in the NE Pacific (Kim et al., 2004; Barron & Anderson, 2011). Regardless of the cause of such a change, these studies implied a geographically uniform change in NE Pacific SST is expected. However, our EOF analysis on TraCE21ka heat budget suggest that different physical processes are responsible for different heat budget temporal patterns, and that different physical processes can be responsible even when each proxy site in the NE
Pacific shows similar temporal patterns. For instance, northern site variability is dominated by changes in surface forcing and southern site variations are driven by changes in horizontal and vertical advection (Figure 8). Together, these results paint a complex picture of SST dynamics in the NE Pacific.

While interpretations based on TraCE21ka cannot be directly applied to interpreting our proxy records, two inferences can be made based on these results. First, common interpretations of SST records in eastern boundary currents are not necessarily accurate. Many studies infer changes in SST records from eastern boundary currents to be dominated by wind driven upwelling (e.g., Abram et al., 2016; Goni et al., 2006; Leduc et al., 2010b; MARGO, 2009; McGregor et al., 2007; Vargas et al., 2007). However, our study provides proxy (different SST patterns) and modeling (different heat balance in TraCE21ka) evidence that this upwelling-only rationale is incomplete, which further confirms analysis based on decade long satellite observations (Cheung et al., 2019). Therefore, processes aside from Ekman upwelling must be responsible for the SST variations in northeast Pacific. This highlights the need for physical nuance in proxy interpretations along coastal regions. Second, the temporal disagreement between our proxy records and climate model simulations highlight the model's inability to simulate variations shown in SST proxy records. Whereas previous studies primarily highlighted proxy seasonal bias as the cause of proxy-model disagreement (Bova et al., 2021; Marsicek et al., 2018; Schneider et al., 2010), some also noted that this bias cannot fully account for proxy-model disagreement (Laepple & Huybers, 2014; Z. Liu et al., 2014a; Osman et al., 2021). In our case, a large magnitude and spatially varying changes in proxy records could simply be a result of seasonal bias. However, our heat budget analysis indicates that there were changes in process level at each proxy location, even though they were balanced out, and that the dominant process that changed is location dependent. This suggests that changes observed in proxy records are indeed possible, albeit not simulated by models, and that seasonal bias is not the only explanation that can be invoked. This lends support to the idea that errors in climate models undermine their ability to match with proxy records. Unfortunately, the cause remains disputed. Nonetheless, some studies have suggested misrepresentation of boundary forcings during the Holocene as the cause of proxy-model disagreement (Y. Liu et al., 2018; Thompson et al., 2022), whereas other studies have highlighted biases introduced due to models' inability to simulate subgrid scale variability (Constantinou & Hogg, 2021; Jüling et al., 2021; Laepple & Huybers, 2014).
4.3 External forcings

Besides identifying processes, it is also important to understand which forcings are driving changes observed in proxy records. Precession, obliquity, and greenhouse gas content have all changed throughout the Holocene. Traditionally, precession has been considered to be the external forcing that has undergone the most significant change and so, many studies have examined the relationship between proxy records and precession (Wanner et al., 2008). However, the influence of greenhouse gases and combined orbital forcing (precession and obliquity) on the climate system have also been suggested (Bova et al., 2021; Kaufman et al., 2020b; Routson et al., 2019).

The primary empirical evidence for an externally forced climate signal comes from the comparison between spatiotemporal patterns of proxy records and the temporal evolution of external forcings. Our records suggest that SST in the NE Pacific does not follow a simple linear trend, in contrast to the trend in precession. Our proxy records also showed an increase in SST towards the present instead of a decrease in mid-high latitudes, which exhibits the opposite temporal behavior compared to the zonally average temperature trend and insolation gradient (Figure 3). Furthermore, although the direction and amplitude of SST change in proxy records are consistent with greenhouse gas forcing, where there is larger amplitude change in high latitude due to polar amplification (M. Collins et al., 2013), the timing of the SST increase in mid-high latitude in the NE Pacific is inconsistent with a linear increase in greenhouse gas forcing over the Holocene. Overall, the lack of correspondence between forcing and SST proxy records suggest no single forcing dominates SST in the NE Pacific, and that the external forcing responsible is unlikely to impose a direct effect (local top of the atmosphere change) on SST, as opposed to what is commonly assumed (Bova et al., 2021; Routson et al., 2019; Wanner et al., 2008).

Although proxy records provide the most direct evidence of whether externally forcing drives changes, model simulations allow us to determine the expected fingerprints from external forcings. As noted before, results based on TraCE21ka may not be directly applicable in the real world due to discrepancies in SST variability when compared to proxy records. Nonetheless, TraCE21ka offers an environment to test hypotheses with regards to spatial fingerprint of external forcings within a physically constrained world. The detection and attribution analysis of TraCE21ka simulations suggest a spatially heterogeneous fingerprint from each forcing. For instance, in contrast to an expected uniform positive fingerprint,
greenhouse gases impose a positive fingerprint in mid to high latitude in the NE Pacific, but
a negative fingerprint in low latitudes. Orbital forcing (precession and obliquity), on the
other hand, imposes a negative fingerprint in mid-high latitude but a positive fingerprint in
low latitude in the NE Pacific, in contrast with what would be expected from the change
in insolation gradient. The fingerprint of each forcing (greenhouse gases and orbital) is also
significant in the NE Pacific. Together, these results highlight that externally driven SST
does not necessarily follow external forcing timeseries, and that it is likely that multiple
forcings have a role in driving SST changes in the NE Pacific during the Holocene. Since
TraCE21ka simulations do not include biogeochemistry, we cannot make inferences about
productivity proxies analyzed here. However, newer and planned paleoclimate simulations
that include ocean biogeochemistry (e.g., Segschneider et al., 2018) should help us better
understand how external forcings have influenced productivity in this region.

5 Summary and Conclusion

In this study, we investigated how SST and productivity in the NE Pacific changed from
9000 to 1000 yrs BP and the processes and forcings behind these observed changes. We find
that SST changes in the NE Pacific are spatially variable, with higher latitudes showing
larger changes than lower latitudes. On the other hand, we do not find significant and
coherent changes in primary producer community, nor each primary producer’s contribution
to carbon export. While model-proxy disagreement impedes our ability to directly identify
processes driving changes observed in proxies, we show that it is possible that different
processes are responsible for the observed changes. Furthermore, we suggest that it is likely
multiple external forcings are responsible for the changes observed in proxy records.

Our results have led to new insights on changes in the NE Pacific and more broadly
interpretations of Holocene climate change. First, our new SST records and analysis of
model simulations highlight the variability and the different temporal evolution of SST in
the NE Pacific, as opposed to synchronous changes proposed in previous studies (Kim et
al., 2004). Second, our productivity records underscore the difficulty of detecting changes
that have cascading effects to the foodweb and carbon export during the Holocene. Third,
we highlight the pitfalls in directly tying external forcing timeseries to Holocene proxy
records, wherein external forcing fingerprints could be complex due to radiative feedbacks
and atmospheric-oceanic processes.
More broadly, several inferences can be made with regards to future climate change in the NE Pacific. Based on our SST records, we suggest that future SST changes on long timescales are also likely to be spatially variable, with a larger magnitude of change in higher latitudes. The complex changes in productivity records also highlight the complexity of ocean ecosystem changes, and show that future changes in ocean ecosystems will likely be driven by a myriad of processes. Lastly, the model’s inability to simulate observed SST changes underscores the need to improve the model resolution of local and subgrid scale processes to improve future projections of local climate change.

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Open Research

Moderate-resolution Imaging Spectroradiometer (MODIS) Aqua Chlorophyll data was obtained from NASA Goddard Space Flight Center, Ocean Ecology Laboratory, Ocean Biology Processing Group (doi: data/10.5067/AQUA/MODIS/L3M/CHL/2018; Accessed on 2021-01-25. NOAA_OI_SST_V2 data was provided by the NOAA/OAR/ESRL PSL, Boulder, Colorado, USA and can be found: https://psl.noaa.gov/data/gridded/data.noaa.oisst.v2.html. Accessed on 2021-01-25. Paleoclimate data generated in this study (ODP1034, MV0508-32JC, MD02-2505, MD02-2506C²) can be found: https://www.ncei.noaa.gov/access/paleo-search/study/36493. TraCE-21ka simulation output was retrieved from Climate Data Gateway maintained by NCAR (https://www.earthsystemgrid.org/project/trace.html). Satellite based diatom and coccolithophore concentrations and data for PCM00-78C/PCM00-78K were downloaded from the Pangaea database (https://doi.pangaea.de/10.1594/PANGAEA.870486). https://doi.org/10.1594/PANGAEA.909404). Data for EW0408-85JC, ODP1019, TN062-O550, and GC41/PC14 were downloaded from

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