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Alkenone isotopes show evidence of active carbon concentrating mechanisms in coccolithophores as aqueous carbon dioxide concentrations fall below 7 µmol L$^{-1}$

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Abstract. Coccolithophores and other haptophyte algae acquire the carbon required for metabolic processes from the water in which they live. Whether carbon is actively moved across the cell membrane via a carbon concentrating mechanism, or passively through diffusion, is important for haptophyte biochemistry. The possible utilization of carbon concentrating mechanisms also has the potential to over-print one proxy method by which ancient atmospheric CO$_2$ concentration is reconstructed using alkenone isotopes. Here I show that carbon concentrating mechanisms are likely used when aqueous carbon dioxide concentrations are below 7 µmol L$^{-1}$. I compile published alkenone-based CO$_2$ reconstructions from multiple sites over the Pleistocene and recalculate them using a common methodology, which allows comparison to be made with ice core CO$_2$ records. Interrogating these records reveals that the relationship between proxy CO$_2$ and ice core CO$_2$ breaks down when local aqueous CO$_2$ concentration falls below 7 µmol L$^{-1}$. The recognition of this threshold explains why many alkenone-based CO$_2$ records fail to accurately replicate ice core CO$_2$ records, and it suggests the alkenone proxy is likely robust for much of the Cenozoic when this threshold was unlikely to be reached in much of the global ocean.

1 Introduction

Alkenones are long-chain (C$_{37-39}$) ethyl and methyl ketones (Fig. 1; Brassell et al., 1986; Rechka and Maxwell, 1987) produced by a restricted group of photosynthetic haptophyte algae (Conte et al., 1994). Produced by a narrow group of organisms which live exclusively in the photic zone, alkenones allow probing of algal biogeochemistry, and as alkenones are often preserved in the sedimentary record, alkenones can also provide information about past environmental conditions.

Two main proxy systems based on alkenone geochemistry exist: one allows reconstruction of sea surface temperature (SST) and relies on the changing degree of unsaturation of the C$_{37}$ alkenone (U$_{37}'$) (Brassell et al., 1986), whilst a second for atmospheric CO$_2$ concentration is based on reconstructing the isotopic fractionation which takes place during photosynthesis ($\varepsilon_p$) (Laws et al., 1995; Bidigare et al., 1997). It is the second system using the stable carbon isotopic composition of the preserved alkenones for reconstructing atmospheric CO$_2$ concentration (referred to throughout as CO$_2$(U$_{37}'$-alk)) which is the focus of this study.

In the modern ocean, alkenones are produced primarily by two dominant coccolithophore species: *Emiliania huxleyi* and *Gephyrocapsa oceanica*. *E. huxleyi* first appeared 290 kyr ago and began to dominate over *G. oceanica* around 82 kyr ago (Gradstein et al., 2012; Raffi et al., 2006). However alkenones are commonly found in sediments throughout the Cenozoic, with the oldest reported detections from mid-Albian-aged black shales (Farrimond et al., 1986). Prior to the evolution of *G. oceanica*, alkenones were most likely produced by other closely related species from the Noelaerhabdaceae family (Marlowe et al., 1990; Volkman, 2000). Micropalaeontological and molecular data split the coccolith-bearing haptophytes into two distinct phylogenetic clades: the Isochrysidales and Coccolithales. The Isochrysidales contain the modern alkenone-producing taxa, including *E. huxleyi* and *G. oceanica*, and fossil reticulofenestrids. Meanwhile the non-alkenone-producers are separated into
the order Coccolithales, which includes *Coccolithus pelagicus* and *Calcidiscus leptopus* along with most other coccolithophores.

Proxies for atmospheric CO$_2$ concentration – including CO$_2$(ε$_p$-alk), those based on the δ$^{13}$B of planktic foraminifera, geochemical modelling, and stomatal density – broadly agree that over the Cenozoic atmospheric pCO$_2$ declined from high levels (> 1000 µatm) in the "greenhouse" worlds of the Palaeocene and Eocene to close to modern-day values (around 400 µatm) in the Pliocene (Pagani et al., 2005, 2011; Pearson et al., 2009; Anagnostou et al., 2016; Foster et al., 2017; Sosdian et al., 2018; Super et al., 2018; Zhang et al., 2013; Beerling and Royer, 2011). However, recent discrepancies have emerged between CO$_2$(ε$_p$-alk) and other CO$_2$ proxies at the < 400 µatm atmospheric CO$_2$ concentrations of the Pleistocene (i.e. Badger et al., 2019, 2013a, and compare Badger et al., 2013b, and Pagani et al., 2009, with Martínez-Botí et al., 2015). Whilst the long-standing differences between alkenone (Pagani et al., 1999), δ$^{13}$B (Foster et al., 2012), and stomatal proxies (Kürschner et al., 2008) in the Miocene CO$_2$ reconstructions have been partially resolved with new SST records (Super et al., 2018), differences remain in the Pliocene (Pagani et al., 2009; Badger et al., 2013b; Martínez-Botí et al., 2015) and Pleistocene (Badger et al., 2019).

Carbon concentrating mechanisms

One plausible reason for the discrepancies between CO$_2$(ε$_p$-alk) and other proxies for atmospheric CO$_2$ is the operation of active carbon concentrating mechanisms (CCMs) in haptophytes. These are potentially important as CO$_2$(ε$_p$-alk) assumes purely passive uptake of carbon into the haptophyte cell purely via diffusion (Laws et al., 1995, 1997; Bidigare et al., 1997). The potential for CCMs to affect the haptophyte cell purely via diffusion (Laws et al., 1995, 2002) allows increased CO$_2$ uptake into the cell passively, and DIC acquisition (Riebesell et al., 2000; Zondervan et al., 2003), which may indicate a less-well-developed CCM in CO$_2$(ε$_p$-alk) proxy, firstly by changing the effective [CO$_2$]$_{aq}$ within the cell and secondly by imparting another carbon isotope fractionation during CA catalysis which is not considered by the CO$_2$(ε$_p$-alk) proxy system. However CA activity in coccolithophores does not appear to be regulated by CO$_2$ as it is in diatoms and *Phaeocystis* (Rost et al., 2003), which may indicate a less-well-developed CCM in coccolithophores.

Calcifying coccolithophores (which include alkenone producers *E. huxleyi* and *G. oceanica*) may be able to utilize HCO$_3^-$ directly as a carbon source (Trimborn et al., 2007), with precipitation of CaCO$_3$ providing an acid for the dehydration of HCO$_3^-$, but this still requires sufficient HCO$_3^-$ entering the cell, and it is unclear whether calcification aids DIC acquisition (Riebesell et al., 2000; Zondervan et al., 2002). The light-dependent leak of carbon (as CO$_2$ and DIC) back from haptophyte cells (including the coccolithophore *E. huxleyi*) to seawater (Tchernov et al., 2003) suggests that CCMs are energy intensive and can concentrate DIC within the cell. Even with active CCMs, it appears that in the ocean through water and the slow kinetics of the bicarbonate-to-[CO$_2$]$_{aq}$ transformation, surface water [CO$_2$]$_{aq}$ can still be depleted by photosynthetic activity. This can become particularly problematic in species which form blooms and at the cell boundary of species with limited motility. It should be no surprise therefore that many marine photosynthetic organisms have evolved with mechanisms to concentrate carbon within the cell.

The enzyme carbonic anhydrase (CA) can catalyse the dehydration of HCO$_3^-$ to [CO$_2$]$_{aq}$ to speed up availability of carbon if the [CO$_2$]$_{aq}$ reservoir is depleted and has been observed in several haptophytes, including coccolithophores (Rost et al., 2003; Riebesell et al., 2007). The exact contribution of CA remains unclear, but two possible mechanisms for CCMs have been postulated (Reinfelder, 2011): (1) CA catalyses dehydration of HCO$_3^-$ at the cell surface, which then allows increased CO$_2$ to diffuse into the cell passively, and (2) HCO$_3^-$ is transported into the cell and then converted by CA. Both of these options will likely impact the CO$_2$(ε$_p$-alk) proxy, firstly by changing the effective [CO$_2$]$_{aq}$ within the cell (and so impacting ε$_p$) and secondly by imparting another carbon isotope fractionation during CA catalysis which is not considered by the CO$_2$(ε$_p$-alk) proxy system. However CA activity in coccolithophores does not appear to be regulated by CO$_2$ as it is in diatoms and *Phaeocystis* (Rost et al., 2003), which may indicate a less-well-developed CCM in coccolithophores.

**Figure 1.** Alkenones are C$_{37}$ unsaturated methyl ketones (Brassell et al., 1986; Rechka and Maxwell, 1987).
cocolithophores are CO₂ limited under some circumstances (Riebesell et al., 2007).

2 Materials and methods

Calculating CO₂ from alkenone δ¹³C values: the CO₂(ε₉₋ₐlk) proxy

In this study I use the now large number of published CO₂(ε₉₋ₐlk) records which overlap with ice core records of atmospheric CO₂ concentration (Tables 1 and 2) to explore the relationship between CO₂(ε₉₋ₐlk) and CCMs in the Pleistocene, where our understanding of atmospheric CO₂ concentration is best.

Multiple records of CO₂(ε₉₋ₐlk) have been published for the Pleistocene (Fig. 2, Table 1), allowing direct comparison with ice-core-based CO₂ records (Table 2). These records are globally distributed in longitude but are concentrated at low-latitude sites, largely as there is a general preference for sites which have (in the modern ocean) surface waters close to equilibrium with the atmosphere (Fig. 2, Table 1). In longer-term palaeoclimate studies there has also been a preference for low-latitude gyre sites in the belief that these sites are more likely to be oceanographically stable over long time intervals (Pagani et al., 1999). Most of the records included here (Table 1, Fig. 2) were generated with the aim to reconstruct atmospheric CO₂ concentration; however one, the MANOP Site C of Jasper et al. (1994), was used to explicitly reconstruct changing disequilibrium due to oceanographic frontal changes over time and so is excluded from the following analysis.

Whilst these sites do only span a relatively small latitudinal extent, the diversity of settings do allow for investigation of any secondary controls on alkenone δ¹³C values (δ¹³Calkenone) – in particular, differences in oceanographic setting and SST to test the hypothesis that low [CO₂]ₐq) breaks the relationship between δ¹³Calkenone and atmospheric CO₂ concentration, as might be expected if haptophytes are able to actively take up carbon from seawater to meet metabolic demand (i.e. activate CCMs).

To facilitate fair comparison between sites and consistent comparison with the ice core records, all CO₂(ε₉₋ₐlk) records were recalculated using a consistent approach. The approach is based on Bidigare et al. (1997), which updated the initial approach of Jasper and Hayes (1990) to CO₂(ε₉₋ₐlk). This approach removes some additional corrections used in the original publication of the records (such as growth rate adjustment for NIOP 464; Palmer et al., 2010) but does allow for direct comparison to be made. For all sites the “b” term was estimated using modern-day surface [PO₄⁴⁻] (Bidigare et al., 1997; Pagani et al., 2009).

An overview of how CO₂(ε₉₋ₐlk) data are typically generated is given in Badger et al. (2013b). Briefly, to calculate εₗ requires the stable carbon isotopic composition of the dissolved CO₂ (δ¹³CCO₂(aw)) and haptophyte biomass (δ¹³Corg). The isotopic fractionation between δ¹³Calkenone and δ¹³Corg is first corrected assuming a constant fractionation (εalkenone) of 4.2‰ (Garcia et al., 2013; Popp et al., 1998; Bidigare et al., 1997):

\[
\varepsilon_{\text{alkenone}} = \frac{\delta^{13}\text{C}_{\text{alkenone}} + 1000}{\delta^{13}\text{C}_{\text{org}} + 1000} - 1.
\] (1)

The isotopic composition of DIC is estimated using (ideally) the δ¹³C value of planktic foraminifera and the temperature-dependent fractionation between calcite and [CO₂]ₐq) experimentally determined by Romanek et al. (1992), where T is sea surface temperature in degrees Celsius (SST):

\[
\varepsilon_{\text{calcite}-\text{CO}_2(g)} = 11.98 - 0.12T.
\] (2)

The value of the carbon isotopic composition of CO₂(ε₉₋ₐlk) (δ¹³CCO₂(ε₉₋ₐlk)) can then be calculated:

\[
\delta^{13}\text{C}_{\text{CO}_2(g)} = \frac{\delta^{13}\text{C}_{\text{carbonate}} + 1000}{\varepsilon_{\text{calcite}-\text{CO}_2(g)} + 1000} - 1000.
\] (3)

From this δ¹³CCO₂(aw) can be calculated using the relationship experimentally determined by Mook et al. (1974),

\[
\varepsilon_{\text{CO}_2(aw)-\text{CO}_2(g)} = \frac{-373}{T + 273.15} + 0.19.
\] (4)
Table 1. Sites with Pleistocene CO$_2$($\varepsilon_p$–alk) records. Note that the MANOP Site C record was generated to track changes in surface water–atmosphere equilibrium, not atmospheric $p$CO$_2$, so, although it is included here for completeness, it is not included in the analysis. Distance from the coast is calculated from the intermediate-resolution version of GSHHG and computed using Generic Mapping Tools (Wessel and Smith, 1996; Wessel et al., 2019).

| Site            | Age interval (kyr) | Latitude       | Longitude       | Water depth (m) | Distance from coast (km) | Reference          |
|-----------------|--------------------|----------------|-----------------|-----------------|--------------------------|--------------------|
| 05PC-21         | 0.5–188            | 38°24′N        | 131°23′E        | 1721            | 108                      | Bae et al. (2015)  |
| DSDP 619        | 3–92               | 27°11.61′N     | 91°24.54′W      | 2259            | 489                      | Jasper and Hayes (1990) |
| NIOP 464        | 7.8–29             | 22°9′N         | 63°21′E         | 1470            | 333                      | Palmer et al. (2010) |
| ODP 999         | 111–258            | 12°44.639′N    | 78°44.360′W     | 2839            | 249                      | Badger et al. (2019) |
| ODP 925         | 20–580             | 4°12.249′N     | 43°29.334′W     | 3042            | 626                      | Zhang et al. (2013) |
| MANOP Site C    | 0.8–253            | 0°57.2′N       | 138°57.3′W      | 4287            | 998                      | Jasper et al. (1994) |
| GeoB 1016-3     | 1.3–196            | 11°46.2′S      | 11°40.9′E       | 3410            | 185                      | Andersen et al. (1999) |

The full record for ODP Site 925 extends to 38.62 Ma.

Table 2. Sources of ice core data, as compiled by Bereiter et al. (2015). WAIS – West Antarctic Ice Sheet; TALDICE – TALos Dome Ice CorE; and EDML – EPICA Dronning Maud Land. Age given as gas age relative to 1950.

| Age interval (kyr) | Ice core location | Reference                  |
|--------------------|-------------------|----------------------------|
| −0.051 to 1.8      | Law Dome          | Rubino et al. (2013)       |
| 1.8–2              | Law Dome          | MacFarling Meure et al. (2006) |
| 2–11               | Dome C            | Monnin et al. (2001, 2004) |
| 11–22              | WAIS              | Marcott et al. (2014)      |
| 22–40              | Siple Dome        | Ahn and Brook (2014)       |
| 40–60              | TALDICE           | Bereiter et al. (2012)     |
| 60–115             | EDML              | Bereiter et al. (2012)     |
| 105–155            | Dome C            | Schneider et al. (2013)    |
| 155–393            | Vostok            | Petit et al. (1999)        |

and

$$\delta^{13}C_{CO_2(aq)} = \left( \frac{\epsilon_{CO_2(aq)} - \epsilon_{CO_2(g)}}{1000} + 1 \right) \times (\delta^{13}C_{CO_2(g)} + 1000) - 1000. \tag{5}$$

Finally $\varepsilon_p$ can be calculated:

$$\varepsilon_p = \left( \frac{\delta^{13}C_{CO_2(aq)} + 1000}{\delta^{13}C_{CO_3^{\text{alk}}}} + 1000 \right) \times 1000, \tag{6}$$

and from that $[CO_2]_{(aq)}$ is calculated using the isotopic fractionation during carbon fixation ($\epsilon_f$) and $b$, which represents the summation of physiological factors:

$$[CO_2]_{(aq)} = \frac{b}{\epsilon_f - \epsilon_p}. \tag{7}$$

Here $\epsilon_f$ is assumed to be a constant 25% ($\varepsilon_p$) (Bidigare et al., 1997). In the modern ocean the $b$ term, which accounts for physiological factors such as cell size and growth rate, shows a close correlation with $[PO_4^{3-}]$ (Bidigare et al., 1997; Pagani et al., 2009). However, the relationship between $b$, growth rate, and $[PO_4^{3-}]$ has recently been questioned (Zhang et al., 2019, 2020) but for the purposes of this analysis is assumed to hold. This is discussed further below. Values for SST, $\delta^{13}C_{\text{alkenone}}, \delta^{13}C_{\text{carbonate}},$ salinity, and $[PO_4^{3-}]$ are either taken from the original publications or estimated from modern ocean estimates (Takahashi et al., 2009; Antonov et al., 2010; Garcia et al., 2013; Locarnini et al., 2013).

Providing that the atmosphere is in equilibrium with surface water, the concentration of atmospheric CO$_2$ can be calculated from $[CO_2]_{(aq)}$ (and vice versa if atmospheric CO$_2$ concentration is known) using Henry’s law:

$$pCO_2 = \frac{[CO_2]_{(aq)}}{K_H}. \tag{8}$$

The solubility coefficient ($K_H$) is dependent on salinity and SST, and here it is calculated following the parameterization of Weiss (1970, 1974).
To produce time-equivalent estimates of atmospheric CO$_2$ concentration for comparison with the ice core records, a simple linear interpolation of the Bereiter et al. (2015) compilation was initially used (Fig. 4). This assumes that both the age model of the ice core and the published age models of the sites are correct and equivalent. This is almost certainly not the case, and so for the calculations below, a ±3000 year uncertainty is included for ages of both the ice core and CO$_2$(ε$_{p-alk}$) values. Figure 4 shows that CO$_2$(ε$_{p-alk}$)-based atmospheric CO$_2$ concentration agree with ice core CO$_2$ at some sites and at some times, but not throughout. Sites 05-PC21 (Baе et al., 2015) and DSDP Site 619 (Jasper and Hayes, 1990) perform quite well throughout, whilst ODPSite 999 (Badger et al., 2019) and NIOP 464 (Palmer et al., 2010) only appear to agree at higher values of CO$_2$, and at ODPSite 925 (Zhang et al., 2013) and GeoB 1016-3 (Andersen et al., 1999) there is very little overlap between the two methods of reconstructing atmospheric CO$_2$ concentration.

To explore whether [CO$_2$(aq)] is an important influence on CO$_2$(ε$_{p-alk}$) I calculate predicted [CO$_2$(aq)] ([CO$_2$(aq)]−predicted) for each of the samples. To calculate [CO$_2$(aq)]−predicted, the time-equivalent value of atmospheric CO$_2$ concentration from the ice core record is used in combination with Eq. (8) to calculate [CO$_2$(aq)] at the time of alkenone production for each sample. Reconstructed estimates of SST and salinity are used as for CO$_2$(ε$_{p-alk}$) above, along with any estimated surface water–atmosphere disequilibrium. Points in Fig. 4 are then coloured by [CO$_2$(aq)]−predicted.

Inspection of Fig. 4 suggests a connection between ([CO$_2$(aq)]−predicted) and the skill of CO$_2$(ε$_{p-alk}$) to reconstruct atmospheric CO$_2$ concentration. The points clustering around the 1 : 1 line are lighter in colour (so with higher [CO$_2$(aq)]−predicted), whilst points falling away from the 1 : 1 line have lower [CO$_2$(aq)]−predicted. To explore this relationship, I progressively restricted the included samples on the basis of [CO$_2$(aq)]−predicted and at each stage calculated a Pearson correlation coefficient ($r$) and coefficient of determination ($r^2$) for each subset. Under this analysis the correlation progressively increased as more of the low [CO$_2$(aq)]−predicted samples were excluded (Fig. 5). All analyses were performed in R (R Core Team, 2020) using RStudio (RStudio Team, 2020). This suggests that the fidelity of the CO$_2$(ε$_{p-alk}$) depends on the concentration of [CO$_2$(aq)], improving at higher levels of [CO$_2$(aq)].

To further investigate this potential relationship, I progressively exclude samples based on [CO$_2$(aq)]−predicted with a step size of 0.05 μmol L$^{-1}$, again calculating Pearson correlation coefficients and coefficients of determination between ice core and CO$_2$(ε$_{p-alk}$) for each subsample of the population. The result is shown in Fig. 6. Here the analysis shows, similar to Fig. 5, that, as the samples with lowest [CO$_2$(aq)]−predicted are progressively removed, the correlation between ice core and CO$_2$(ε$_{p-alk}$) increases. Furthermore, this continues only up until [CO$_2$(aq)]−predicted reaches
Figure 4. Crossplots of $\text{CO}_2(\varepsilon_{p-alk})$-based atmospheric $\text{CO}_2$ concentration (y axes) vs. the time-equivalent estimate from ice core records (x axes; Bereiter et al., 2015; Table 2). The large panel compiles all sites, with the exception of MANOP Site C, as explained in the text. Symbols are coloured by predicted $[\text{CO}_2]_{aq}$ for each site and time as explained in the text. Full sources for alkenone data are shown in Table 1. A 1 : 1 line is included in all plots for comparison.

7 µmol L$^{-1}$. Above this, the coefficient of determination plateaus, until the subsample reaches such a small size that spurious correlations become important (Fig. 6b).

3.2 Sensitivity and uncertainty tests

It is possible that the pattern seen in Fig. 6b could emerge from a dataset shaped with increasing density surrounding the 1 : 1 correlation line without being driven by changes in $[\text{CO}_2]_{aq}$-predicted. To explore this possibility, I ran a series of sensitivity experiments. In these, rather than reducing the sample by filtering by $[\text{CO}_2]_{aq}$-predicted, the whole dataset (Table 1) was randomly ordered and then stepwise subsampled. To make this equivalent to the $[\text{CO}_2]_{aq}$-predicted analysis above, I set the size of each subsample to be equal to each step in the original analysis. This produces a randomly selected but same-sized subsample such that the size of the subsample reduces in the same way as shown in Fig. 6b. Pearson correlation coefficients and coefficients of determination were calculated for each subsample as above, and I repeated this 1000 times, with the order of each sample randomized each time.

To allow for possible age model uncertainties, a 3000-year ($1\sigma$) uncertainty was also applied to each sample. This uncertainty was applied to the age of each sample prior to sampling of the ice core record, and it is applied as a normally distributed uncertainty. Uncertainty in $\text{CO}_2(\varepsilon_{p-alk})$ measurements is typically calculated using Monte Carlo modelling of all the parameters (i.e. Pagani et al., 1999; Badger et al., 2013a, b); however this was not done in all the published work (Table 1), and some differences in approach were found across the published work. Therefore to create $\text{CO}_2(\varepsilon_{p-alk})$ uncertainty estimates for each value in this study, I emulate the uncertainties based on the $\text{CO}_2(\varepsilon_{p-alk})$ value. I built a simple emulator (Fig. 7) by running Monte Carlo uncertainty estimates for all of the included datasets (Table 1) using the same estimates of uncertainty for each variable in the $\text{CO}_2(\varepsilon_{p-alk})$ calculation as applied in Badger et al. (2013a, b). This then allows the uncertainty to be included in the $[\text{CO}_2]_{aq}$-predicted calculation as well as $\text{CO}_2(\varepsilon_{p-alk})$, and it allowed for uncertainty estimates to be site-ambivalent.

The result is shown in Fig. 6c and d, and it suggests that the 4 µmol L$^{-1}$ break point remains valid. The absolute value of $r^2$ is reduced, even at higher $[\text{CO}_2]_{aq}$-predicted, but this would be expected given the addition of uncertainty in the age model, as the published age is most likely to align with the ice core. Given the rapid rate of change at deglaciations, this effect is likely to be particularly pronounced in this dataset as many records have high temporal resolution around deglaciations in order to attempt to resolve them.
Any small age model offset introduced by the error modelling in these intervals also clearly has the potential to induce large differences between the CO$_2$(p$_\text{alk}$) and ice core values. Figure 6c and d clearly demonstrate that it is the filtering by [CO$_2$_(aq)]$\text{predicted}$ rather than any spurious correlations which determines the shape of the data in Fig. 6a.

4 Discussion

The plateau in $r^2$ in Fig. 6a and c suggests that below a [CO$_2$_(aq)]$\text{predicted}$ of $\sim 7$ µmol L$^{-1}$ CO$_2$(p$_\text{alk}$) is no longer as good a predictor of ice core CO$_2$ as when [CO$_2$_(aq)]$\text{predicted}$ $> 7$ µmol L$^{-1}$. This is clear from comparing the relationship between samples where [CO$_2$_(aq)]$\text{predicted}$ $< 7$ µmol L$^{-1}$ with those where [CO$_2$_(aq)]$\text{predicted}$ $> 7$ µmol L$^{-1}$ in Fig. 8. Here the $r^2$ for the former of 0.15 is substantially less than the latter of 0.55. I suggest that this is because below this threshold the fundamental assumption of CO$_2$(p$_\text{alk}$), that carbon is passively taken up by haptophytes, no longer holds true. One obvious explanation for why this would be the case is that at low levels of [CO$_2$_(aq)] haptophytes have to rely more on active uptake of carbon via CCMs in order to satisfy metabolic demand. Similar behaviour has been recognized in some culture studies (Laws et al., 1997, 2002; Cassar et al., 2006), with some evidence that the diatom Phaeodactylum tricornutum has a similar CCM threshold of 7 µmol L$^{-1}$ (Laws et al., 1997). Whilst the evidence for the mechanism of CCM is poorer for coccolithophores than it is for diatoms, any CCM would be expected to compromise the CO$_2$(p$_\text{alk}$) proxy, either by increased supply of [CO$_2$_(aq)] or by further carbon isotopic fractionation effects during carbon transport, or both (Stoll et al., 2019). By applying a threshold value for [CO$_2$_(aq)$\text{predicted}$ of 7 µmol L$^{-1}$ to the published records (Table 1), values of CO$_2$(p$_\text{alk}$) which are influenced by active CCMs can be eliminated. Recognition of this new threshold value of [CO$_2$_(aq)$\text{predicted}$ allows for a new record of Pleistocene CO$_2$(p$_\text{alk}$) to be compiled. This compilation then much better replicates the glacial–interglacial pattern of CO$_2$ change over the last 260 kyr (Fig. 9). Whilst this present compilation does rely on ice core CO$_2$ records to estimate [CO$_2$_(aq)$\text{predicted}$, and therefore has little direct utility as a CO$_2$ record, it does demonstrate that recognition of a threshold response allows accurate CO$_2$ reconstruction using CO$_2$(p$_\text{alk}$). This may represent the point at which isotopic effects of CCMs (plausibly through increased CA activity or HCO$_3^-$ dehydration to meet C demand) overwhelm the assumptions of the CO$_2$(p$_\text{alk}$) proxy. This, as well as

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Figure 6. Coefficient of determination (a) of a reducing sample of all compiled \(\text{CO}_2(\varepsilon_{p-\text{alk}})\) (Table 1) vs. the time-equivalent estimate from ice core records (Bereiter et al., 2015; Table 2). The sample reduces stepwise by 0.05 \(\mu\text{mol L}^{-1}\), and the number of records in each subsample is shown in panel (b). Panel (c) shows a 1000-member Monte Carlo analysis, whereby uncertainty in \(\text{CO}_2(\varepsilon_{p-\text{alk}})\) and age is considered, as detailed in the text. Panel (d) shows a similar 1000-member Monte Carlo analysis, but with random sampling of the whole \(\text{CO}_2(\varepsilon_{p-\text{alk}})\) population so that the number of samples is equivalent to the dataset shown in panel (c); i.e. the size of the sample follows that shown in panel (b). Means and 1\(\sigma\) uncertainties are shown as the bold lines.

Figure 7. Emulated uncertainty in \(\text{CO}_2(\varepsilon_{p-\text{alk}})\), generated by running Monte Carlo uncertainty models for all sites in Table 1, applying the same approach to uncertainty as Badger et al. (2013a, b). Estimates used in this study are highlighted in blue.

Figure 8. Correlations between \(\text{CO}_2(\varepsilon_{p-\text{alk}})\) and ice core \(\text{CO}_2\), where \([\text{CO}_2(\text{aq})-\text{predicted}] > 7 \mu\text{mol L}^{-1}\) (black symbols) and \([\text{CO}_2(\text{aq})-\text{predicted}] < 7 \mu\text{mol L}^{-1}\) (red symbols).

the behaviour shown in Fig. 6a and c, suggests that from the standpoint of the \(\text{CO}_2(\varepsilon_{p-\text{alk}})\) proxy CCMs may effectively be considered either active or not, and that when \([\text{CO}_2(\text{aq})]\) is plentiful passive uptake dominates, at least sufficiently in most oceanographic settings that \(\text{CO}_2(\varepsilon_{p-\text{alk}})\) can accurately record atmospheric \(\text{CO}_2\) concentration. This implies that, if areas of the ocean (or intervals of time) with low \([\text{CO}_2(\text{aq})]\) can be avoided, accurate reconstructions of atmospheric \(\text{CO}_2\) concentration can be acquired using \(\text{CO}_2(\varepsilon_{p-\text{alk}})\).

As \([\text{CO}_2(\text{aq})]\) is affected by both SST via the temperature dependance of the Henry’s law constant and atmospheric \(\text{CO}_2\) concentration, for \(\text{CO}_2(\varepsilon_{p-\text{alk}})\) to be effective in reconstructing atmospheric \(\text{CO}_2\) concentration, areas of warm wa-
for CCMs in the rest of the Cenozoic, when the assumption out the late Pliocene and Pleistocene, it is plausible that cor-

gested by the analyses presented here, CCMs only act at low 
concentrations (Zhang et al., 2019; Stoll et al., 2019; Zhang 
(Bolton et al., 2012; Bolton and Stoll, 2013). 

earliest, and likely not widespread until the Plio-Pleistocene 
providing the most difficulty (Badger et al., 2019). This find-
ing aligns well with evidence that CCMs developed in coc-
clostrophores at low CO2

Figure 9. Revised compilation of Pleistocene CO2(CHCl-C-alk) vs. ice core records. The compiled published records (Table 1) are shown as circles, coloured red where [CO2]aq-predicted is below a thresh-
old of 7 µmol−1 and blue where [CO2]aq-predicted > 7 µmol−1. The solid blue line is a loess filter (span 0.1) through the [CO2]aq-predicted > 7 µmol−1 values, with 95% confidence inter-
vals (dashed blue line). The black line is the ice core compilation of Bereiter et al. (2015) (Table 2).

ter (i.e. tropical or shallow shelf regions) under relatively low 
atmospheric CO2 concentration must be avoided. However, 
as the atmospheric CO2 control renders the global surface 
ocean sufficiently replete with [CO2]aq at Pliocene-like lev-
elleS of atmospheric CO2 concentration and above (Martínez-
Botí et al., 2015) at all but the warmest surface ocean temper-
atures, CO2(CHCl-C-alk) is likely to be a reliable system for most 
of the Cenozoic. It is only in the Pleistocene that atmospheric 
CO2 concentration is low enough for CCMs to be widely 
active across the surface ocean, with the low-CO2 glacials 
providing the most difficulty (Badger et al., 2019). This find-
ing aligns well with evidence that CCMs developed in cocc-
lolithophores as a response to declining atmospheric CO2 
concentration through the Cenozoic and were developing in 
[CO2]aq-limited parts of the ocean in the late Miocene at the 
earliest, and likely not widespread until the Plio-Pleistocene 
(Bolton et al., 2012; Bolton and Stoll, 2013).

There have been recent attempts to correct for CCMs in 
CO2(CHCl-C-alk)-based reconstructions of atmospheric CO2 
concentrations (Zhang et al., 2019; Stoll et al., 2019; Zhang 
et al., 2020). However, these assume that CCMs are al-
ways active and crucially do not fundamentally break the 
relationship between CHCl-C and atmospheric CO2 con-
centration. However if this is not the case, and the relation-
ship between CHCl-C and atmospheric CO2 concentration 
fails at Pleistocene levels of atmospheric CO2, then 
Pleistocene records cannot be used to develop corrections of 
CO2(CHCl-C-alk) to be applied throughout the Cenozoic. If, as sug-
gested by the analyses presented here, CCMs only act at low 
[CO2]aq, and largely only in conditions prevalent through-
out the late Pliocene and Pleistocene, it is plausible that cor-
rections based on Pleistocene records could overcompensate 
for CCMs in the rest of the Cenozoic, when the assumption 
of passive carbon uptake inherent in CO2(CHCl-C-alk) as tradition-
ally applied may still be valid.

5 Conclusions

Reconstructions of past atmospheric CO2 concentration with 
proxy tools like CO2(CHCl-C-alk) are critical for understanding 
how the Earth’s climate system operates, as long as the tools 
used can be relied upon to be accurate and precise. This re-
analysis of existing Pleistocene CO2(CHCl-C-alk) records reveals 
that below a critical threshold of [CO2]aq of 7 µmol−1 the 
relationship between 13Calkenone and atmospheric CO2 con-
centration breaks down, plausibly because below this thresh-
old haptophytes are able to actively take up carbon using 
CCMs in order to satisfy metabolic demand.

Although reconstructing the low levels of atmospheric 
CO2 concentration in the Pleistocene glacials and areas of 
the global ocean where [CO2]aq is less than 7 µmol−1 will be 
impossible, for much of the Cenozoic the CO2(CHCl-C-alk) 
proxy retains utility. If care is taken to avoid regions and 
oceanographic settings where [CO2]aq is expected to be ab-
normally low, CO2(CHCl-C-alk) remains an important and useful 
proxy to understand the Earth system.

Code and data availability. This paper relies exclusively on previ-
ously published data, available with the original papers and in pub-
licly available repositories. An R notebook supplement is available 
alongside this paper, along with data files, which allow full replica-
tion of all analyses performed.

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