Introduction

The supporting information gives more details on materials and methods, including the coral samples, the potential factors affecting coral skeletal $\delta^{13}C$ values, the calculation of shelf flooding, the calculation of reef carbonate volume and mass accumulation, the carbon cycle modelling framework, and the complete description of a 3-box-model to understand $\delta^{13}C$ recorded in corals of the Great Barrier Reef during the last deglaciation.
Text S1.

Materials and Methods

Coral samples

Fossil, massive and robust branching/columnar *Isopora palifera/cuneata* corals were recovered in 2010 by IODP Expedition 325 in the central GBR (Webster et al., 2011; Felis et al., 2014a). Corals were recovered at depths between 56 and 126 m below modern sea level at Noggin Pass (NOG; 146.6° E, 17.1° S; IODP sites M0053 and M0057) and Hydrographer’s Passage (HYD; 150.3° E, 19.7° S; IODP sites M0031, M0033, M0035, M0036, and M0039) using the Mission Specific Platform ‘Greatship Maya’ (Webster et al., 2011; Felis et al., 2014a) (Figure 2). The corals were screened for possible diagenetic alteration of their skeletons using X-radiography, powder X-ray diffraction, thin section petrography and Mg/Ca screening, and were demonstrated to be well preserved (Felis et al., 2014a). U-Th dating yielded coral ages spanning 25.0 to 11.7 ka (Felis et al., 2014a). Modern *Isopora palifera/cuneata* corals were collected at Heron Island (HER; 151.9° E, 23.4° S) in the southern GBR from 1974 to 1979 at depths between 0 and 14 m, and at Myrmidon Reef (MYR; 147.4° E, 18.3° S) and Magnetic Island (MAG; 146.8° E, 19.1° S) in the central GBR in 1982 at depths of <15 m and <0.5 m, respectively (Felis et al., 2014a) (Figure 2). The coral specimens were analysed for Sr/Ca, $\delta^{18}$O, Mg/Ca, and $\delta^{13}$C along their major growth orientation (Felis et al., 2014a). A total of 881 samples were analysed at subseasonal resolution in seven fossil *Isopora* corals (Felis et al., 2014a). An additional 18 fossil and 10 modern *Isopora* were analysed for bulk geochemical composition (Felis et al., 2014a), as the major aim of a previous study was the reconstruction of mean temperature changes from geochemical averages of corals (Felis et al., 2014a). For more details see Felis et al. (2014a).

Potential factors affecting coral skeletal $\delta^{13}$C values

The GBR anomaly of coral $\delta^{13}$C values between 12.8 and 11.7 ka could also be seen as a substantial widening of the range of coral $\delta^{13}$C values, or intercolony variability, approaching values broadly comparable to those observed in modern corals along the GBR today (Figure 1c) (Swart & Coleman, 1980). The pronounced decrease in coral $\delta^{13}$C values could therefore reflect a time interval when the GBR had reached, for the first time during the last deglaciation, a coral community level with a wide range of shallow-water habitats on the reef in terms of differences in water depth and location, comparable to today. Importantly, a potential influence of changes in light availability on the overall temporal evolution of the GBR record of coral $\delta^{13}$C values, due to a larger water depth of coral growth and/or increased reef water turbidity during the course of the last deglaciation, especially after 13.0 ka, can be largely ruled out. This is because *Isopora* corals, when associated with other multiproxy indicators (growth form, other corals, algae, vermetids), are limited to very shallow water depths and considered to be highly sediment intolerant (Webster et al., 2018; Humblet et al., 2019). Thus, the paleoecological constraint of a shallow-water (<5-10 m), high-energy reef crest environment indicated by the *Isopora*-dominated coralgial assemblage after 13.0 ka (Webster et al., 2018; Humblet...
et al., 2019) strongly suggests the GBR anomaly of coral δ¹³C values between 12.8 and 11.7 ka is likely not a result of more diverse shallow-water habitats on the reef.

We note that the pronounced decrease in GBR coral δ¹³C values between 12.8 and 11.7 ka could also, at least partly, reflect a decrease in light availability (Swart, 1983; Swart et al., 1996c; Felis et al., 1998; Swart et al., 2005; Swart et al., 2010; Al-Rousan & Felis, 2013; Gagan et al., 2015; Linsley et al., 2019), caused by a potentially larger water depth of coral growth or increased reef water turbidity after the 13.0 ka shelf flooding. However, a significant increase in water depth can be largely ruled out, because the GBR coral record is based on shallow-water (0-10 m) *Isopora* colonies (Felis et al., 2014a) recovered predominantly from multiproxy coralgal assemblages with estimated paleowater depth of <10 m (Webster et al., 2018; Humblet et al., 2019). In particular, the records' 12.8 to 11.7 ka interval that reveals the pronounced decrease in coral δ¹³C values is derived from an *Isopora*-dominated assemblage with even shallower paleowater depth estimates of <5-10 m (Webster et al., 2018; Humblet et al., 2019). The individual paleowater depth estimates of the corals that make up the GBR record of δ¹³C values clearly rule out a deeper shallow-water habitat after the 13.0 ka shelf flooding compared to the preceding period (Webster et al., 2018; Humblet et al., 2019) (Table S2). Moreover, no clear relationship between skeletal δ¹³C values and water depth is observed in our modern GBR *Isopora* corals (Figure S6).

Similarly, an increase in turbidity of reef waters can be largely ruled out, as the *Isopora*-dominated communities are highly sediment intolerant (Kojis & Quinn, 1984), and the very rapid initial accretion rates of this particular assemblage (Webster et al., 2018) (9.6 mm yr⁻¹) point to favourable conditions for photosynthesis and coral calcification at that time. Furthermore, the regional paleogeography characterized by coastal lagoons and estuaries may have served as sediment traps and does not support excessive turbidity during this time interval (Hinestrosa et al., 2016), and the developing mangrove forests (Grindrod et al., 1999) might have acted as additional sediment traps. In any case, the increase in regional sediment flux, as recorded in distal slope and basin sediment cores (Dunbar & Dickens, 2003; Page & Dickens, 2005), may have ultimately decreased water quality and led to the final demise of the reef around 10 ka (Webster et al., 2018), but this occurred well after the pronounced decrease in coral δ¹³C values that occurred between 12.8 and 11.7 ka.

For modern *Isopora* corals, an intercolony variability in δ¹³C values among corals of the same reef reaching differences of up to 1.51‰ (HER) as well as differences in δ¹³C values among corals from different GBR reef sites of up to 1.72‰ (HER, MAG, MYR) are observed (Figure 1c, 2, S6, Table S2). These differences are similar to those reported for *Porites* corals at the northern Gulf of Aqaba (1.63‰) (Al-Rousan & Felis, 2013) and between islands of the Fiji region (1.4‰) (Dassié et al., 2013). These intercolony variability and site-specific differences are larger than the pronounced decrease in GBR coral δ¹³C values (1.25‰) between 12.8 and 11.7 ka (Figure 1c). However, the large modern intercolony variability in δ¹³C values at HER is based on corals from a wide range of shallow-water habitats (lagoon, reef flat, crest, slope, channel) (Felis et al., 2014) and therefore may not be representative of the environmental conditions under which the
last deglacial *Isopora* corals grew. The fossil *Isopora* corals have a very narrow paleoecological constraint of a shallow-water, high-energy reef crest environment as indicated by the corresponding *Isopora*-dominated coralgal assemblage, especially after 13.0 ka (Webster et al., 2018; Humblet et al., 2019). Further, the last deglacial reefs were growing in quite different physiographic, oceanographic and ecologic settings compared to the modern, being narrow fringing reefs and/or close to the coastline (Hinestrosa et al., 2014, 2016; Webster et al., 2018). Importantly, we demonstrate a coherent temporal trend in fossil coral $\delta^{13}C$ values during the last deglaciation from two GBR sites (NOG and HYD) that are ~500 km apart (Figure 1c, 2), that clearly overwhelms any putative environmental differences between the NOG and HYD sites or at either site. The coral $\delta^{13}C$ values of the two sites show a remarkable similarity, with differences of only 0.37‰ to 0.39‰ for corals of similar age from a single site (e.g., ~21 ka at NOG, ~15 ka at HYD) and 0.55‰ for corals of similar age from two different sites (e.g., ~19 ka at NOG and HYD) (Figure 1c, Table S2). This is substantially less than the amplitude of the decrease in coral $\delta^{13}C$ values (1.25‰) after 13.0 ka.

A potential role for growth rate related effects in the decrease in coral $\delta^{13}C$ values can be largely ruled out, as these should have affected coral $\delta^{18}O$ values in a similar way (Felis et al., 1998; Swart et al., 2010; Al-Rousan & Felis, 2013; Gagan et al., 2015; Linsley et al., 2019), which is not observed. Importantly, the GBR record of coral $\delta^{13}C$ values reveals a rather different spatio-temporal evolution throughout the last deglaciation compared to the corresponding Sr/Ca and $\delta^{18}O$ temperature proxy records derived from the same corals (Felis et al., 2014a) (Figure S5a,b,c), which further supports its significance as a recorder of independent environmental information such as changes in GBR surface water $\delta^{13}C_{DIC}$. More specifically, whereas both coral Sr/Ca and $\delta^{18}O$ values form two separate groups for the northern (NOG) and southern sites (HYD), interpreted as steeper meridional temperature gradient in the GBR between 17° and 20°S during the deglaciation (Felis et al., 2014a), the coral $\delta^{13}C$ values do not show this site-dependent difference but appear to follow a common forcing. Further, the Sr/Ca and $\delta^{18}O$ values after 13 ka do not show a pattern comparable to that of the pronounced decrease in coral $\delta^{13}C$. Finally, because of the different pattern of GBR coral $\delta^{13}C$ values and the corresponding Sr/Ca and $\delta^{18}O$ temperature proxies during the anomalous 12.8 to 11.7 ka interval, a substantial influence of temperature dependent equilibrium fractionation on coral $\delta^{13}C$ values at that time (Lynch-Stieglitz et al., 2019) can be largely ruled out.

A potential role for changes in incoming solar radiation (insolation) on orbital timescales in modulating light availability and therefore the overall evolution of the GBR coral $\delta^{13}C$ record during the last deglaciation can be excluded. The insolation changes during the 11-25 ka time interval are minor (<4 W m$^{-2}$), and of the same magnitude as the difference in absolute insolation between the northern (NOG, 17°S) and the southern latitude IODP site (HYD, 20°S) (Berger, 1978) (Figure S5e). If there would be any influence, the ~6 W m$^{-2}$ lower absolute insolation at the southern site (HYD, 20°S) should be reflected as consistent offset towards more negative coral $\delta^{13}C$ values relative to corals from the northern site (NOG, 17°S), which is not observed (Figure S5c,e).
An increased advection of $^{13}$C-depleted waters from the tropics, driven by strengthening of the southward-flowing East Australian Current at that time (Felis et al., 2014a), cannot be totally ruled out as a cause for the decrease in coral $\delta^{13}$C values between 12.8 and 11.7 ka. However, recent findings challenging the role of “oceanic tunnelling” for the global distribution of $^{13}$C-depleted carbon released from the deep ocean (Lynch-Stieglitz et al., 2019; Shao et al., 2021) and, consequently, its transfer from the Southern Ocean towards the tropics by mode and intermediate waters, as well as the magnitude of change in GBR coral $\delta^{13}$C values, argue against such a potential mechanism.

Calculation of shelf flooding

The calculation of the postglacial flooding of the GBR shelf (Hinestrosa et al., 2019) used a 0.001° (~100 m) resolution bathymetric grid (Beaman, 2010) (Figure 2) divided into 33 latitudinal zones at 50 km intervals (Hinestrosa et al., 2019). Flooding was calculated for each of the latitudinal zones with pre-defined sea level inundation values ranging from 130 to 0 m at 5 m step intervals (Hinestrosa et al., 2019). For each sea level increment within each latitudinal zone, the polygons representing the marine-flooded areas were extracted and their surface area calculated in km², and the increase in marine-flooded area within each sea level increment was calculated (Hinestrosa et al., 2019). These calculations were also performed for a bathymetry subset corresponding to the shelf edge area, defined as the total area comprised between the modern outer GBR reef front and the 130 m isobath on the high-resolution bathymetry (Beaman, 2010). Therefore, the change in flooded area represents the total area flooded in the arbitrary period between two past sea level marks (Hinestrosa et al., 2019). The flooded area for a time interval of interest was calculated from the total (cumulative) marine-flooded area curve (Figure 1d), by using the slope between adjacent sea level marks of known age. In contrast to a previous study (Hinestrosa et al., 2019), the GBR sea level reconstructions (Webster et al., 2018; Yokoyama et al., 2018) are used here for calculating the shelf flooding of the time interval >10 ka, whereas the sea level curve of Lambeck et al. (2014) is used for the time interval <10 ka (not shown) where no corresponding GBR sea level reconstruction is available. The GBR sea level curve used here is the average of the maximum relative sea level reconstructions at Noggin Pass (NOG) and Hydrographer’s Passage (HYD) (Webster et al., 2018; Yokoyama et al., 2018) (Figure 1e) interpolated to 250-year time steps and re-sampled at 5 m sea level step intervals. A similar interpolation and re-sampling were applied to the sea level curve of Lambeck et al. (2014).

Calculation of reef carbonate volume and mass accumulation

Flooded area (km²): We used marine flooded area values estimated for the GBR shelf edge (Hinestrosa et al., 2019) by the calculations described above. These represent areas flooded since the last glacial maximum, calculated every 5 m of sea level increment, and tied to geological time according to the GBR sea level curve (Webster et al., 2018; Yokoyama et al., 2018) and the sea level curve of Lambeck et al. (2014) (for the time interval <10 ka, not shown).
Postglacial vertical accretion rate, VA (m kyr\(^{-1}\)) and VA\(_{SL}\) (m [sea level increment]\(^{-1}\)): The glacial-postglacial boundary recognized in the drillcores and the radiometric ages (Webster et al., 2011; Webster et al., 2018; Yokoyama et al., 2018) allowed us to extract a vertical accretion gradient (m kyr\(^{-1}\)) for the postglacial period (VA). This accretion rate was then transformed into an equivalent thickness per sea-level increment gradient (VA\(_{SL}\)).

Reef cover (%): The percentage of reef cover compared to total shelf edge area was estimated at the control sites (Noggin Pass and Hydrographer’s Passage) (Webster et al., 2011) by digitising the area occupied by reef banks and comparing it to the total area comprised between the modern outer GBR reef front and the 130 m isobath on the high-resolution bathymetry (Beaman, 2010). The identification of the banks was constrained by the bathymetry, backscatter, seismic and geographic information system data (Abbey et al., 2011; Harris et al., 2012; Hinestrosa et al., 2014; Hinestrosa et al., 2016). The results were averaged and rounded to the nearest ten to obtain a value of 20%.

Maximum cumulative reef thickness (m): This is the maximum postglacial thickness observed in drill cores in the shelf-edge reefs at the control sites (Noggin Pass and Hydrographer’s Passage) (Webster et al., 2011). We estimated a maximum possible reef thickness for different stages of shelf-edge reef development. This was necessary to represent the observed reef drowning and to avoid unrealistic cumulative thickness values as sea level progressed.

Porosity (%): Formation porosity as per drill core samples of 35% (Webster et al., 2011).

Density (kg m\(^{-3}\)): Aragonite mineral density of 2,930 kg m\(^{-3}\).

The postglacial-thickness method relies on the assumption that reef cover, vertical accretion rate and maximum cumulative thickness values observed in the control sites (Noggin Pass and Hydrographer’s Passage) can be extended to other locations along the GBR shelf margin. Firstly, the vertical accretion rate (VA) for each reef development episode was converted to an equivalent rate relative to past sea level increments, VA\(_{SL}\). This allowed us to associate an incremental reef thickness to each of the 5 m sea level increments considered. Conversion from geological age to equivalent sea level was performed using the GBR sea level curve (Webster et al., 2018; Yokoyama et al., 2018) and the sea level curve of Lambeck et al. (2014) (for the time interval <10 ka, not shown).

To obtain an estimate of reef thickness at each sea level increment, we multiplied the transformed postglacial vertical accretion rate (VA\(_{SL}\)) by each sea level increment (5 m). At each location, the cumulative thickness had an upper bound, representing reef drowning, defined by the maximum cumulative reef thickness (Figure S7). Here, we used a maximum cumulative reef thickness of 20 m for all periods as the likely scenario. The resulting reef thickness was then multiplied by the area in km\(^2\) for each flooded area polygon, from 130 m to 0 m in 5 m increments. The resulting formation volume in km\(^3\) was scaled down by the reef cover values to account for the lack of reef accretion in the inter-reef areas. This consideration made the estimations consistent with observations at the shelf-edge control sites. The scaled volume was then multiplied by the non-porous fraction of the formation (1 - porosity) and by the density to obtain CaCO\(_3\) accumulated mass. The
CaCO₃ accumulation for a time interval of interest was calculated from the total (cumulative) reef carbonate (CaCO₃) accumulation curve (Figure 1d), by using the slope between adjacent sea level marks of known age.

For more details on the calculations of reef carbonate volume and mass accumulation see Hinestrosa et al. (2022).

**Carbon cycle modelling framework**

To be able to simulate the rather large anomaly in δ¹³C values recorded in the GBR corals (Felis et al., 2014a) we developed a 3-box carbon cycle model specifically for this purpose (Figure S1). It consists of a box surrounding the flooded shelf where the GBR was growing, a box for the rest of the surface ocean, and a box for the atmosphere. Fluxes of carbon, alkalinity and δ¹³C values are considered, leading to 8 differential equations, which have been solved within the Matlab or Octave software environment. A full description of the model is found in the next section. The model code is available with the simulation results (see open research section of main paper).

Due to the rather local feature of the GBR growth its effects on surface water δ¹³CDIC is difficult to be simulated successfully with established models. However, for the interpretation of the overall effect of the GBR growth on atmospheric CO₂ and δ¹³CO₂ values the global carbon cycle box model BICYCLE (Köhler & Munhoven, 2020) is still a useful tool. BICYCLE consists of global average atmosphere, a 10-box-ocean and a 7-box terrestrial land biosphere. It is here used in a revised so-called Solid Earth version (BICYCLE-SE), which has been applied for investigating the carbon cycle of the late Pleistocene (Köhler & Munhoven, 2020) (see sketch of the model showing all relevant carbon fluxes in Figure S2). In this version a process-based sediment module is implemented, which simulates in each of the three ocean basins (Atlantic, Southern Ocean, Indo-Pacific) the pressure-dependent carbonate system in steps of 100 m water depth. Carbonate is thus either accumulated in the sediments or dissolved, depending on the deep ocean over- or undersaturation of the carbonate ion. This approach led to glacial/interglacial dynamics in CO₃²⁻ well comparable with reconstructions (Köhler & Munhoven, 2020). Other solid Earth fluxes considered in this model version are not only shallow water accumulation of carbonate in corals, but also the constant riverine influx of bicarbonate from both silicate and carbonate weathering (12 TmolC yr⁻¹ each) as well as volcanic outgassing as function of sea level change (8-15 TmolC yr⁻¹). In the scenarios analysed here, we use all time-dependent changes as used in are standard setup before (scenario SE in Köhler & Munhoven (2020)), but in our control run we set the global coral reef carbonate sink to zero. We then investigate carbon flux anomalies obtained with the model when the GBR coral reef growth as calculated here (Figure 4a) are implemented as additional sink fluxes of CaCO₃ in the equatorial surface box of the Indo-Pacific. Simulation results obtained with BICYCLE (Figure 4) use the carbon fluxes and δ¹³C values from the best-guess scenario of the 3-box-model (Figure 3b). The respired land carbon is in BICYCLE directly released into the atmosphere, since any release into the large surface ocean boxes has, due to water exchange processes, only minimal impact on the atmosphere. The plotted anomalies in the simulated atmospheric carbon are obtained by
adding the shallow water carbonate accumulation in the GBR and the respired carbon from the flooded shelf to a control run without coral reef growth (scenario C0 in Köhler & Munhoven (2020)) that covers the last glacial cycle.

For our simulation experiments, we consider the time interval from 15.9 ka, thus 2 kyr before the onset of the of rapid sea-level rise (Webster et al., 2018; Yokoyama et al., 2018), until 11.7 ka when the last deglacial GBR record of coral $\delta^{13}C$ values (Felis et al., 2014a) ends. For these simulation experiments one needs to be aware that the GBR record of coral $\delta^{13}C$ values (Felis et al., 2014a) is not measured on corals from a specific site, but is a composite from corals that grew at different sites along two transects (NOG and HYD) during the flooding of the GBR shelf (Felis et al., 2014a; Webster et al., 2018; Yokoyama et al., 2018; Hinestrosa et al., 2019). The marine-flooded area and reef carbonate (CaCO3) accumulation values used in our simulations for the individual time intervals were calculated as described above, and are given here:

**Time interval: 15.9-13.9 ka (slow sea-level rise on steep slope), absolute time: 2.0 kyr.**
Flooded area (all GBR incl. shelf edge): 4,290 km²
CaCO3 accumulation (GBR shelf edge only): 11.47 Gt (114.7 Tmol-CO3²⁻)
\(=\) (57.4 Tmol-CO3²⁻ kyr⁻¹)
Flooded area (GBR shelf edge only): 2,390 km²

**Time interval: 13.9-13.0 ka (rapid sea-level rise on steep slope), absolute time: 0.9 kyr.**
Flooded area (all GBR incl. shelf edge): 25,160 km²
CaCO3 accumulation (GBR shelf edge only): 29.34 Gt (293.4 Tmol-CO3²⁻)
\(=\) (326.0 Tmol-CO3²⁻ kyr⁻¹)
Flooded area (GBR shelf edge only): 8,360 km²

**Time interval: 13.0-11.7 ka (slow sea-level rise on flat platform), absolute time: 1.3 kyr.**
Flooded area (all GBR incl. shelf edge): 24,590 km²
CaCO3 accumulation (GBR shelf edge only): 18.12 Gt (181.2 Tmol-CO3²⁻)
\(=\) (139.4 Tmol-CO3²⁻ kyr⁻¹)
Flooded area (GBR shelf edge only): 4,330 km²

**Complete description of a 3-box-model to understand $\delta^{13}C$ recorded in corals of the Great Barrier Reef during the last deglaciation**

We built a 3-box model (Figure S1) tailored for this specific application consisting of:

- **b**: box around the Great Barrier Reef (GBR)
- **r**: rest of the surface ocean
- **a**: atmosphere

Fluxes of carbon (C), alkalinity (ALK) and $\delta^{13}C$ are considered, leading to 8 time-dependent differential equations, which have been solved within the Matlab or Octave software environment. The box b considers changes in the water box, that has just been
flooded. Since the GBR record of coral $\delta^{13}$C values was not measured on corals from a fixed location, but is a composite of corals that grew at different sites along two transects during the flooding of the GBR shelf (Felis et al., 2014a), the area of $b$ varies according to Table S1.

If respired land carbon from flooded shelves $C_{\text{resp}}$ is considered, it is assumed that the carbon is released constantly over the entire time window. For example, for 13.9-13.0 ka it is assumed that the 30 kgC m$^{-2}$ of land carbon over the area of 25.16 $10^9$ m$^2$ is released with a constant annual flux of $30 \times 10^3 \times 25.16 \times 10^9 / 900 = 0.84$ TgC yr$^{-1}$.

It is important to have the concentrations of dissolved inorganic carbon (DIC) and alkalinity in $r$ to be representative of the rest of the surface ocean (not the whole ocean), otherwise the gas exchange fluxes will be biased.

A crucial detail is the size of the GBR box $b$. Here, exchange fluxes, water depth, and corals’ carbonate production have to take each other into account to get meaningful results. Otherwise, the concentrations of DIC or alkalinity in $b$ might drop to 0. This box size of $b$ is essential to be able to simulate such large drops in $\delta^{13}$C of 1.25‰, and is also the reason, why such a signal cannot be produced with the global carbon cycle model BICYCLE (Köhler & Munhoven, 2020), since the surface ocean boxes are much too large there. On the other hand, simulated atmospheric signals are rather poor within this 3-box-approach, since we fix DIC and alkalinity in $r$ (the rest of the surface ocean) to constant values to avoid drifts. Thus, the mass of carbon is not preserved in the 3-box model. Atmospheric variables are best discussed in the output of the well-tested BICYCLE model (Köhler & Munhoven, 2020).

Due to the necessary balance between ingoing and outgoing fluxes in the small box $b$ we also restrict sensitivity runs to changes in the parameter values to ±25% only and use this relative change throughout most sensitivity tests.

Variables are initialized as given in Table S3. The simulation time is restricted to 15.9-11.7 ka. The end of this time window at 11.7 ka is determined by the last data point of the last deglacial record of GBR coral $\delta^{13}$C values (Felis et al., 2014a). For the beginning of this simulation period we choose to start at 15.9 ka, 2 kyr before significant changes in the GBR shelf flooding have been detected (13.9 ka) (Webster et al., 2018; Yokoyama et al., 2018) in order to be free of initial effects in the simulated carbon signals.

All parameter values used to solve the set of differential equations are contained in Table S4. All simulated scenarios performed for plotting the results seen in Figure 3 and in Figures S3 and S4 are described in Table S5.
Differential equations:

\[
\frac{\delta C_b}{\delta t} = r2b - b2r + a2b - b2a - C_{coral} + C_{resp}
\]

\[
\frac{\delta C_a}{\delta t} = b2a - a2b + r2a - a2r
\]

\[
\frac{\delta C_r}{\delta t} = b2r - r2b + a2r - r2a
\]

\[
\frac{\delta}{\delta t} (\delta^{13}C_b) = (r2b \cdot \delta^{13}C_a - b2r \cdot \delta^{13}C_b + a2b \cdot (\delta^{13}C_a + e_{2a}) - b2a \cdot (\delta^{13}C_b + e_{2b}) - C_{coral} \cdot \delta^{13}C_{coral} + C_{resp} \cdot \delta^{13}C_{resp} - \frac{\delta}{\delta t} C_b \cdot \delta^{13}C_b)/C_b
\]

\[
\frac{\delta}{\delta t} (\delta^{13}C_a) = (b2a \cdot (\delta^{13}C_b + e_{2a}) - a2b \cdot (\delta^{13}C_a + e_{2a}) + r2a \cdot (\delta^{13}C_r + e_{2a}) - a2r \cdot (\delta^{13}C_a + e_{2a}) - \frac{\delta}{\delta t} C_a \cdot \delta^{13}C_a)/C_a
\]

\[
\frac{\delta}{\delta t} (\delta^{13}C_r) = (b2r \cdot \delta^{13}C_b - r2b \cdot \delta^{13}C_r + a2r \cdot (\delta^{13}C_a + e_{2a}) - r2a \cdot (\delta^{13}C_r + e_{2a}) - \frac{\delta}{\delta t} C_r \cdot \delta^{13}C_r)/C_r
\]

\[
\frac{\delta}{\delta t} ALK_b = r2b_{ALK} - b2r_{ALK} - 2 \cdot C_{coral}
\]

\[
\frac{\delta}{\delta t} ALK_r = b2r_{ALK} - r2b_{ALK}
\]

Applied boundary conditions, which mimics the unrestricted supply of DIC and alkalinity in the global ocean:

\[
\frac{\delta}{\delta t} C_r = 0
\]

\[
\frac{\delta}{\delta t} ALK_r = 0
\]

In our carbon cycle models, we assume an \( \varepsilon_{cor} \) of -2‰, but test widely how this value and that of other parameters influence our results (Figure S3). Note, that if a largely different \( \varepsilon_{cor} \) is chosen, the simulated coral \( \delta^{13}C \) values also change dramatically, shifting outside the range of measured coral \( \delta^{13}C \) values. For example, if \( \varepsilon_{cor} \) of -0.5 or -3.5‰ is assumed, our 3-box model produces before 13.9 ka a \( \delta^{13}C_{coral} \) of 2.2 or 0.0‰ instead of 1.0‰ in our control run (Figure S3a).

At 13.0 ka the carbonate accumulation rate per whole flooded area dropped more than two-fold from 13.0 to 5.7 mol yr\(^{-1}\) m\(^{-2}\) (Table S1). If implemented in the 3-box carbon cycle model this process alone can already explain a drop in seawater \( \delta^{13}C_{DIC} \) values of ~1.0‰, thus the dominant part of what is recorded in the corals between 12.8 and 11.7 ka (scenario 100 in Figure 3a). In our sensitivity test, in which we varied the most important parameters of our model by ±25% the amplitude of this \( \delta^{13}C \) drop at 12.8 ka ranged from -0.6‰ to -1.3‰ (Figures S3 and S4a), but note that for the most extreme cases the GBR box became undersaturated with respect to aragonite, which would prevent reef calcification. Thus, the upper end of this range is geochemically unrealistic. We note that the aragonite saturation state of seawater in the 3-box model although being slightly oversaturated is substantially lower than values reported for the modern GBR (Mongin et al., 2016). However, the model design primarily aims at identifying processes responsible for the coral \( \delta^{13}C \) dynamics observed in the last deglacial GBR, and we note that no undersaturation occurs in the global carbon cycle model BICYCLE (Köhler & Munhoven, 2020) due to a larger volume and exchange fluxes of the ocean boxes.
Around 13.9 ka, the about 6-fold increase in flooded area (Table S1), and therefore the higher decomposition rate of organic land carbon, leads to a drop in $\delta^{13}$C$_{\text{DIC}}$ by 0.5‰ (scenario 110 in Figure 3a and Figure S4b). Around 13.0 ka, this land carbon decomposition slightly dampens the drop in $\delta^{13}$C$_{\text{DIC}}$ caused by the reduced carbonate accumulation in corals by 0.2‰ (scenario 110 versus 100 in Figure 3a and Figure S4a,b). This is a rather unexpected change, since the flux of respired land carbon (directly related to the shelf area flooded by sea level rise, Figure 4a) has hardly changed at that point in time. The $\delta^{13}$C$_{\text{DIC}}$ nevertheless changes because of the nonlinear processes in the marine carbonate system (Zeebe & Wolf-Gladrow, 2001): dissolved CO$_2$ concentration in the surface waters, that via gas exchange might enter the atmosphere, is a function of DIC and alkalinity, which are both differently altered by the coral’s carbonate production and the decomposition of land carbon. The subsequent atmosphere-ocean gas exchange itself contains an isotopic fractionation (Mook, 1986) that leads to the well-known net depletion of atmospheric $\delta^{13}$CO$_2$ with respect to surface ocean $\delta^{13}$C$_{\text{DIC}}$ of about -8‰.

If implemented in our 3-box-model, a decrease in water exchange rates between the GBR and the open ocean at 13.0 ka leads to a reduction of the drop in $\delta^{13}$C$_{\text{DIC}}$ caused by the previously implemented processes (scenario 101 in Figure 3a and Figure S4c). Again, this can in detail only be understood by the nonlinearity of the shallow marine carbonate system (Zeebe & Wolf-Gladrow, 2001).

Our modelling framework does not account for changes in the interplay of photosynthesis and respiration during diurnal carbon cycling on the reef (Geyman & Maloof, 2019). We note that minor changes towards reduced diurnal $\delta^{13}$C amplitudes, e.g., due to a potential shift in reef metabolism from net autotrophy towards net heterotrophy, could have also contributed to the pronounced decrease in coral skeletal $\delta^{13}$C values after 13 ka (Geyman & Maloof, 2019). However, the very rapid initial accretion rates of the highly sediment-intolerant Isopora-dominated coral assemblage after 13.0 ka (Webster et al., 2018) point to favourable conditions for photosynthesis compared to the preceding period, which broadly argues against a potential shift towards less overall photosynthesis at that time.
Figure S1. Tailored 3-box carbon cycle model for interpretation of Great Barrier Reef coral $\delta^{13}$C values. In addition to the exchange fluxes between the three boxes $a$ (atmosphere), $b$ (Great Barrier Reef box), and $r$ (rest of the surface ocean), the production of carbonate by corals, and the release of respired land carbon on the flooded shelf is considered. In $b$ and $r$, dissolved inorganic carbon and alkalinity are traced, from which the dissolved CO$_2$ concentration is calculated following the marine carbonate chemistry (Zeebe & Wolf-Gladrow, 2001), that can then exchange with the atmosphere. More details are found in Text S1.
Figure S2. Sketch of the BICYCLE model, adapted from Köhler & Munhoven (2020). Top: Model version including solid Earth fluxes (BICYCLE-SE). V: outgassing of CO$_2$ from volcanoes on land potentially and temporally overlain by land ice and from hot spot island volcanoes (and mid-ocean ridges, not shown) influenced by changing sea level; C: shallow water carbonate deposition due to coral reef growth; Si-W: silicate weathering and Ca-W: carbonate weathering with different sources of C, but both delivering HCO$_3^-$ ions into the ocean; P: PO$_3^{2-}$ riverine input and sedimentary burial; S: CaCO$_3$ sedimentation and dissolution. A2B: atmosphere-biosphere exchange of CO$_2$; A2O: atmosphere-ocean exchange of CO$_2$. Cyan-colored broken circles mimic the two overturning cells in the Atlantic and Indo-Pacific Ocean. The previous model version has been restricted to fluxes A2O, A2B, and some simplified representation of carbonate compensation between ocean and sediment. Bottom: Details of atmosphere-ocean-biosphere subsystem (Köhler et al., 2010) with the prescribed pre-industrial water fluxes in the ocean (blue, numbers in Sv). Black arrows denote carbon fluxes. Terrestrial biosphere contains: C4: C$_4$ ground vegetation; C3: C$_3$ ground vegetation; NW: non-woody parts of trees; W: woody parts of trees; D: detritus or above-ground litter; FS: fast decomposing soil or below-ground litter; SS: slow decomposing soil.
Figure S3. Results of 3-box carbon cycle model of Great Barrier Reef growth. Simulated $\delta^{13}$C$_{\text{DIC}}$ (left y-axes) and coral-recorded $\delta^{13}$C (right y-axes) values for the Great Barrier Reef (GBR) box. Sensitivity runs for the scenario in which only coral carbonate accumulation is considered (100). Parameter variation in (a) is the isotopic fractionation factor ($\varepsilon_{\text{cor}}$) during carbonate buildup; (b) the depth of the GBR box; (c) the water exchange between GBR box and the open ocean ($b_{2r}$); and (d) the gas exchange between atmosphere and surface ocean ($a_{2s}$). In (a) the labels of right y-axis depend on $\varepsilon_{\text{cor}}$. Thus, two additional sets of labels in dark/light grey are included for $\varepsilon_{\text{cor}} = -3.5(-0.5)$ ‰. In two sensitivity cases (broken lines, labels in brackets) aragonite becomes undersaturated ($\Omega_{\text{ar}} < 1$), which would hinder coral reef growth.
Figure S4. Results of 3-box carbon cycle model of Great Barrier Reef growth. Simulated $\delta^{13}C_{\text{DIC}}$ (left y-axes) and coral-recorded $\delta^{13}C$ (right y-axes) values for the Great Barrier Reef (GBR) box. Prescribed scenarios follow a three-digit code showing 1 (on) or 0 (off), where the first digit states if coral carbonate accumulation, second digit if land carbon release after shelf flooding, and the third digit if changes in the water mixing between GBR box and open ocean at 13.0 ka is considered. Sensitivity runs with changes in (a) coral carbonate accumulation rate; (b) carbon released from land flooding; (c) lifetime ($\tau$) of GBR water masses with respect to the exchange with the open ocean. (d) CTRL of individual contributions and all processes combined (111). In one sensitivity case (a) indicated with broken line and label in brackets aragonite becomes undersaturated ($\Omega_{Ar} < 1$), which would hinder coral reef growth.
Figure S5. Great Barrier Reef (GBR) coral Sr/Ca temperature, $\delta^{18}$O and $\delta^{13}$C records, ice core records of atmospheric carbon, and insolation at the latitude of the GBR. The GBR record of coral $\delta^{13}$C values reveals a rather different spatio-temporal evolution throughout the last deglaciation compared to the corresponding Sr/Ca and $\delta^{18}$O temperature proxy records derived from the same corals. a Mean skeletal Sr/Ca of GBR shallow-water corals (Felis et al., 2014a). Central GBR sites: Noggin Pass (NOG, red circle), Hydrographer’s Passage (HYD, blue square). Southern GBR site: Heron Island (HER, dark green triangle). Weighted least-squares regression lines shown for NOG and HYD (Felis et al., 2014a). Coral-based sea surface temperature (SST) anomalies not adjusted for seawater Sr/Ca changes, thus provide upper estimates of magnitude of cooling. Fossil coral-based Sr/Ca-SST anomalies plotted relative to average Sr/Ca at HER (dashed green line). Modern mean SST (Locarnini et al., 2010) at NOG (26.6 °C) and HYD (26.0 °C) shown.
relative to SST at HER (24.5 °C), scaled using the mean coral Sr/Ca-SST relationships of –0.084 mmol mol⁻¹ per °C (Gagan et al., 2012) (solid red and blue lines) and –0.140 mmol mol⁻¹ per °C (Felis et al., 2009) (dashed red and blue lines). b As in (a), but for mean skeletal δ¹⁸O corrected for changes in ice volume (Felis et al., 2014a). Coral δ¹⁸O-SST anomalies shown relative to average δ¹⁸O at HER (dashed green line) using the average of three mean coral δ¹⁸O-SST relationships (–0.22‰ per °C, Felis et al., 2009; DeLong et al., 2010; Gagan et al., 2012)). Modern mean SST (Locarnini et al., 2010) at NOG and HYD shown relative to HER (solid red and blue lines). c As in (a), but for mean skeletal δ¹³C (Felis et al., 2014a), with anomalous 12.8 to 11.7 ka interval indicated (grey bar). Central GBR sites: Myrmidon Reef (MYR, light green rhomb), Magnetic Island (MAG, light green triangle). Antarctic ice core reconstructions of δ¹³C of atmospheric CO₂ (dark brown (Schmitt et al., 2012), Monte Carlo average; light brown (Bauska et al., 2016), smoothing spline). d Antarctic ice core reconstructions of atmospheric CO₂ (dark brown (Schmitt et al., 2012), light blue (Marcott et al., 2014), light brown (Bauska et al., 2016)). e Annual mean insolation at the latitude of NOG (17°S) and HYD (20°S) (Berger, 1978). For details on the reconstructed intensification of the meridional temperature gradient in the GBR following the Last Glacial Maximum from coral Sr/Ca and δ¹⁸O see Felis et al. (2014a). For associated uncertainties see original publications and Table S2.
Figure S6. Comparison of skeletal $\delta^{13}$C and water depth for modern Great Barrier Reef (GBR) corals. Mean skeletal $\delta^{13}$C of modern GBR corals (Isopora palifera/cuneata) and the corresponding approximate water depth and year of collection (Felis et al., 2014a). Corals are from Heron Island (HER) (southern GBR) and Myrmidon Reef (MYR) and Magnetic Island (MAG) (central GBR). No clear relationship between skeletal $\delta^{13}$C and water depth is observed. If the water depth of collection is only known as a range, the midpoint of this range is plotted here as best approximation. The modern coral $\delta^{13}$C data are the same as shown in Figure 1c and Figure S5c. For associated uncertainties see original publication (Felis et al., 2014a) and Table S2.
Figure S7. Schematic of the postglacial-thickness method (from Hinestrosa et al., *Scientific Reports*, 2022; Image licensed under a Creative Commons Attribution 4.0 International License: [http://creativecommons.org/licenses/by/4.0/](http://creativecommons.org/licenses/by/4.0/)). For details on the calculations of reef carbonate volume and mass accumulation by using the postglacial-thickness method see Text S1.
<table>
<thead>
<tr>
<th></th>
<th>Time (ka)</th>
<th>duration (y)</th>
<th>area_a(i) (m²)</th>
<th>volume_b(i) (m³)</th>
<th>CaCO₃ or C_corals(i) (mol yr⁻¹)</th>
<th>C_corals(i) (mol yr⁻¹ m⁻³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>15.9-13.9</td>
<td>2000</td>
<td>4.29 · 10⁹</td>
<td>25.74 · 10⁹</td>
<td>57.4 · 10⁹</td>
<td>13.4</td>
</tr>
<tr>
<td>2</td>
<td>13.9-13.0</td>
<td>900</td>
<td>25.16 · 10⁹</td>
<td>150.96 · 10⁹</td>
<td>326.4 · 10⁹</td>
<td>13.0</td>
</tr>
<tr>
<td>3</td>
<td>13.0-11.7</td>
<td>1300</td>
<td>24.59 · 10⁹</td>
<td>147.54 · 10⁹</td>
<td>139.4 · 10⁹</td>
<td>5.7</td>
</tr>
</tbody>
</table>

**Table S1.** Change in flooded area and coral reef carbonate production in the Great Barrier Reef as function of time (this study). Here, $C_{corals(i)}$ in mol yr⁻¹ m⁻³ is calculated for a box $b$ with a water depth of 6 m (Text S1).
The grey shading indicates the anomalous 12.8 to 11.7 ka interval of the Great Barrier Reef coral record that reveals the pronounced decrease in skeletal $\delta^{13}C$ values. For this time interval, a larger water depth of coral growth relative to the preceding period cannot be inferred. Coral paleowater depth estimates are from Webster et al. (2018) and Humblet et al. (2019). A few relatively large paleowater depth ranges (0-20 m, 0-30 m, 0-40 m) are samples that may lack the necessary multiproxy paleoenvironmental indicators or are indicative for possible transport from more shallower habitats (Webster et al., 2018; Humblet et al., 2019).

<table>
<thead>
<tr>
<th>Site</th>
<th>Sample ID Palaeo*</th>
<th>U-Th Age [kyr BP]**</th>
<th>± 2SD</th>
<th>Palaeowater depth [m]</th>
<th>Coral $\delta^{13}C$ ± 1SD</th>
<th>n</th>
<th>Microsampling</th>
</tr>
</thead>
<tbody>
<tr>
<td>NOG</td>
<td>325-57A-4R-1 (75-88)</td>
<td>11.655</td>
<td>0.059</td>
<td></td>
<td>0 - 5</td>
<td>-0.68</td>
<td>0.06</td>
</tr>
<tr>
<td>NOG</td>
<td>325-57A-5R-1 (1-3)</td>
<td>11.746</td>
<td>0.054</td>
<td></td>
<td>0 - 5</td>
<td>-0.53</td>
<td>0.15</td>
</tr>
<tr>
<td>NOG</td>
<td>325-57A-5R-1 (41-50)</td>
<td>11.875</td>
<td>0.055</td>
<td></td>
<td>0 - 5</td>
<td>-0.27</td>
<td>0.14</td>
</tr>
<tr>
<td>NOG</td>
<td>325-57A-5R-1 (126-133)</td>
<td>12.317</td>
<td>0.073</td>
<td></td>
<td>0 - 5</td>
<td>0.40</td>
<td>0.23</td>
</tr>
<tr>
<td>NOG</td>
<td>325-57A-6R-1 (103-114)</td>
<td>12.674</td>
<td>0.074</td>
<td></td>
<td>0 - 5</td>
<td>-0.63</td>
<td>0.07</td>
</tr>
<tr>
<td>NOG</td>
<td>325-57A-6R-2 (68-80)</td>
<td>12.767</td>
<td>0.064</td>
<td></td>
<td>0 - 5</td>
<td>0.21</td>
<td>0.11</td>
</tr>
<tr>
<td>HYD</td>
<td>325-31A-2R-1 (27-32)</td>
<td>13.197</td>
<td>0.003</td>
<td></td>
<td>0 - 20</td>
<td>0.40</td>
<td>0.13</td>
</tr>
<tr>
<td>HYD</td>
<td>325-31A-2R-CC (5-10)</td>
<td>13.238</td>
<td>0.002</td>
<td></td>
<td>0 - 20</td>
<td>0.46</td>
<td>0.29</td>
</tr>
<tr>
<td>HYD</td>
<td>325-33A-9R-1 (7-16)</td>
<td>15.167</td>
<td>0.097</td>
<td></td>
<td>0 - 5</td>
<td>0.82</td>
<td>0.13</td>
</tr>
<tr>
<td>HYD</td>
<td>325-33A-10R-1 (42-48)</td>
<td>15.331</td>
<td>0.070</td>
<td></td>
<td>not available</td>
<td>0.65</td>
<td>0.07</td>
</tr>
<tr>
<td>NOG</td>
<td>325-53A-6R-1 (7-9)</td>
<td>17.146</td>
<td>0.011</td>
<td></td>
<td>0 - 10</td>
<td>1.00</td>
<td>0.03</td>
</tr>
<tr>
<td>NOG</td>
<td>325-53A-15R-1 (33-37)</td>
<td>20.899</td>
<td>0.005</td>
<td></td>
<td>0 - 10</td>
<td>1.28</td>
<td>0.28</td>
</tr>
<tr>
<td>NOG</td>
<td>325-53A-15R-1 (42-48)</td>
<td>20.930</td>
<td>0.004</td>
<td></td>
<td>0 - 10</td>
<td>1.54</td>
<td>0.07</td>
</tr>
<tr>
<td>HYD</td>
<td>325-33A-16R-1 (54-60)</td>
<td>25.038</td>
<td>0.156</td>
<td></td>
<td>0 - 30</td>
<td>1.20</td>
<td>0.06</td>
</tr>
</tbody>
</table>

* Sample ID Palaeo can be slightly different to the corresponding Sample ID Dating, as the dating sample in some cases is a subsample of a larger coral section that was sampled for palaeoclimate. See Felis et al. (2014) for details.

** U-Th ages are reported as kyr before the present relative to AD 1950. See Felis et al. (2014) for details.

*** U-Th age represents the mean of 2 datings, the error given here is the range of the two datings.

The JCp-1 reference composition used in this study is $-1.58\%$ for $\delta^{13}C$. See Felis et al. (2014) for details.

The corals are massive and robust/branching Isopora palifera/cuneata colonies.

The grey shading indicates the anomalous 12.8 to 11.7 ka interval of the Great Barrier Reef coral record that reveals the pronounced decrease in skeletal $\delta^{13}C$ values. For this time interval, a larger water depth of coral growth relative to the preceding period cannot be inferred. Coral paleowater depth estimates are from Webster et al. (2018) and Humblet et al. (2019). A few relatively large paleowater depth ranges (0-20 m, 0-30 m, 0-40 m) are samples that may lack the necessary multiproxy paleoenvironmental indicators or are indicative for possible transport from more shallower habitats (Webster et al., 2018; Humblet et al., 2019).

Table S2. Coral $\delta^{13}C$ data, U-Th ages, and paleowater depth estimates.
<table>
<thead>
<tr>
<th>Term</th>
<th>Units</th>
<th>Values</th>
</tr>
</thead>
<tbody>
<tr>
<td>$C_b = \text{DIC}_b$</td>
<td>mol C m$^{-3}$</td>
<td>2215</td>
</tr>
<tr>
<td>$C_a = \text{CO}_{2\text{atm}}$</td>
<td>ppm (or $\mu$atm)</td>
<td>240</td>
</tr>
<tr>
<td>$C_r = \text{DIC}_r$</td>
<td>mol C m$^{-3}$</td>
<td>2100</td>
</tr>
<tr>
<td>$\delta^{13}C_b$</td>
<td>$%_0$</td>
<td>1.5</td>
</tr>
<tr>
<td>$\delta^{13}C_a$</td>
<td>$%_0$</td>
<td>−6.5</td>
</tr>
<tr>
<td>$\delta^{13}C_r$</td>
<td>$%_0$</td>
<td>1.5</td>
</tr>
<tr>
<td>ALK$_r$</td>
<td>mol m$^{-3}$</td>
<td>2600</td>
</tr>
<tr>
<td>ALK$_b$</td>
<td>mol m$^{-3}$</td>
<td>2550</td>
</tr>
</tbody>
</table>

**Table S3.** Initial values, largely taken from Köhler et al. (2005a).
<table>
<thead>
<tr>
<th>Term</th>
<th>Units</th>
<th>Content</th>
</tr>
</thead>
<tbody>
<tr>
<td>$r_{2b}$</td>
<td>mol C m$^{-3}$ yr$^{-1}$</td>
<td>$\text{DIC}<em>r \cdot f</em>{\text{exchange}}$</td>
</tr>
<tr>
<td>$b_2r$</td>
<td>mol C m$^{-3}$ yr$^{-1}$</td>
<td>$C_{b_2}/\tau$</td>
</tr>
<tr>
<td>$a_{2b}$</td>
<td>mol C m$^{-3}$ yr$^{-1}$</td>
<td>0.031 \cdot area_{a_0}(i) \cdot CO_{2\text{atm}} [\text{Köhler et al., 2005a}]</td>
</tr>
<tr>
<td>$b_{2a}$</td>
<td>mol C m$^{-3}$ yr$^{-1}$</td>
<td>0.031 \cdot area_{a_0}(i) \cdot pCO_{2\text{box}}</td>
</tr>
<tr>
<td>pCO$_{2\text{box}}$</td>
<td>$\mu$atm</td>
<td>$f(\text{DIC}_b, \text{ALK}_b, T = 20^\circ\text{C}, S = 36.2 \text{ psu}, p = 0 \text{ bar})$ [Zeebe and Wolf-Gladrow, 2001]</td>
</tr>
<tr>
<td>$a_{2r}$</td>
<td>mol C m$^{-3}$ yr$^{-1}$</td>
<td>0.051 \cdot (area_{\text{global}} - area_{a_0}(i)) \cdot CO_{2\text{atm}} [\text{Köhler et al., 2005a}]</td>
</tr>
<tr>
<td>$r_{2a}$</td>
<td>mol C m$^{-3}$ yr$^{-1}$</td>
<td>0.051 \cdot (area_{\text{global}} - area_{a_0}(i)) \cdot pCO_{2\text{box}}</td>
</tr>
<tr>
<td>pCO$_{2\text{ref}}$</td>
<td>$\mu$atm</td>
<td>$f(\text{DIC}_r, \text{ALK}_r, T = 15^\circ\text{C}, S = 35.9 \text{ psu}, p = 0 \text{ bar})$ [Zeebe and Wolf-Gladrow, 2001]</td>
</tr>
<tr>
<td>volume$_{a_0}(i)$</td>
<td>m$^3$</td>
<td>area$_{a_0}(i) \cdot \text{depth}_b$</td>
</tr>
<tr>
<td>area$_{\text{global}}$</td>
<td>m$^2$</td>
<td>350 \cdot 10^{12}</td>
</tr>
<tr>
<td>area$_r$</td>
<td>m$^2$</td>
<td>area$<em>{\text{global}} - area</em>{a_0}(i)$</td>
</tr>
<tr>
<td>depth$_b$</td>
<td>m</td>
<td>6</td>
</tr>
<tr>
<td>$f_{\text{exchange}}$</td>
<td>m$^3$ yr$^{-1}$</td>
<td>$\text{volume}_{a_0}(i) / \tau = \text{lifetime wrt b-r exchange}$</td>
</tr>
<tr>
<td>$\tau$</td>
<td>yr</td>
<td>0.25, increased by $\times 2$ @ 13 ka in scenarios 101</td>
</tr>
<tr>
<td>C$_{\text{ref}}$</td>
<td>mol C m$^{-3}$ yr$^{-1}$</td>
<td>$\rho_{\text{ref}} \cdot 1000/12 \cdot area_{a_0}(i)/\text{duration}(i)$</td>
</tr>
<tr>
<td>$\rho_{\text{ref}}$</td>
<td>kgC m$^{-2}$</td>
<td>30 (scenarios 110, 111, before 13 ka in best-guess), 0 (else), 60 (after 13 ka in best-guess scenario) [Köhler et al., 2005b]</td>
</tr>
<tr>
<td>$\epsilon_{a_0}$</td>
<td>$%$</td>
<td>$-$2, isotopic fractionation during air2sea flux [Mook, 1986]</td>
</tr>
<tr>
<td>$\epsilon_{2b}$</td>
<td>$%$</td>
<td>$-$10, isotopic fractionation during sea2air flux [Mook, 1986]</td>
</tr>
<tr>
<td>$\delta^{13}C_{\text{corals}}$</td>
<td>$%$</td>
<td>$\delta^{13}C_b + \epsilon_{\text{corals}}$</td>
</tr>
<tr>
<td>$\epsilon_{\text{corals}}$</td>
<td>$%$</td>
<td>$-$2, isotopic fractionation during CaCO$_3$ growth in corals [Linsley et al., 2019]</td>
</tr>
<tr>
<td>$\delta^{13}C_{\text{ref}}$</td>
<td>$%$</td>
<td>$-$21, mean of glacial/interglacial range of global mean terrestrial values in BICYCLE</td>
</tr>
</tbody>
</table>

**Table S4.** A description of the individual terms contained in the differential equations and the chosen parameter values.
### Table S5. Overview of simulation scenarios performed with the 3-box model.

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>000</td>
<td>CTRL, only a2b, r2b, a2r, r2a fluxes, no corals, no respired land carbon, no change in $\tau$</td>
</tr>
</tbody>
</table>

**scenarios**

| 100     | Coral-Baseline: 000 + corals, $\epsilon_{\text{coral}} = -2.0\%$ |
| 110     | 100 + release of C from respired land carbon, 30 kgC m$^{-2}$ of flooded shelf area with $\delta^{13}C = -21\%$ |
| 101     | 100 + reduced $f_{\text{exchange}}$ after 13 ka, or $\tau$ rises by a factor of 2 |
| 111     | 100 + all 3 effects together (corals, land carbon, change in $\tau$) |

**sensitivity runs — basics**

| 100-eps | fractionation corals $\epsilon_{\text{coral}}$: $-2.0, -3.3, -0.5\%$ |
| 100-depth | GBR box depth: 6, 4.5, 7.5 m |
| 100-a2s | gas exchange fluxes: ctrl, 0.75-ctrl, 1.25-ctrl |
| 100-b2r | exchange flux box-ocean: ctrl, 0.75-ctrl, 1.25-ctrl |

**sensitivity runs — scenario sensitives**

| 100-corals | carbonate fluxes varied around ctrl, 0.75-ctrl, 1.25-ctrl |
| 110-resp | C release from land carbon: 30, 22.5, 37.5 kgC m$^{-2}$ |
| 101-tau | change in b2r fluxes @ 13ka: $\times = 2, 1.75, 2.25$ |

**Best guess scenarios**

| 111-best-internal | 111 + C released from flooding rises at 13 ka from 30 to 60 kgC m$^{-2}$ (implemented in BICYCLE) |
| 111-best-prescribed | 111-best-internal + atmospheric $\delta^{13}CO_2$ prescribed (6.8% at 13.0–12.0 ka; -6.6% else) |