Plateaus and jumps in the atmospheric radiocarbon record – potential origin and value as global age markers for glacial-to-deglacial paleoceanography, a synthesis

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Abstract. Changes in the geometry of ocean meridional overturning circulation (MOC) are crucial in controlling past changes of climate and the carbon inventory of the atmosphere. However, the accurate timing and global correlation of short-term glacial-to-deglacial changes of MOC in different ocean basins still present a major challenge. The fine structure of jumps and plateaus in atmospheric and planktic radiocarbon (14C) concentration reflects changes in atmospheric 14C production, ocean–atmosphere 14C exchange, and ocean mixing. Plateau boundaries in the atmospheric 14C record of Lake Suigetsu, now tied to Hulu Cave U/Th model ages instead of optical varve counts, provide a stratigraphic “rung ladder” of up to 30 age tie points from 29 to 10 cal ka for accurate dating of planktic oceanic 14C records. The age differences between contemporary planktic and atmospheric 14C plateaus record the global distribution of 14C reservoir ages for surface waters of the Last Glacial Maximum (LGM) and deglacial Heinrich Stadial 1 (HS-1), as documented in 19 and 20 planktic 14C records, respectively. Elevated and variable reservoir ages mark both upwelling regions and high-latitude sites covered by sea ice and/or meltwater. 14C ventilation ages of LGM deep waters reveal opposed geometries of Atlantic and Pacific MOC. Like today, Atlantic deep-water formation went along with an estuarine inflow of old abyssal waters from the Southern Ocean up to the northern North Pacific and an outflow of upper deep waters. During early HS-1, 14C ventilation ages suggest a reversed MOC and ~1500-year flushing of the deep North Pacific up to the South China Sea, when estuarine circulation geometry marked the North Atlantic, gradually starting near 19 ka. High 14C ventilation ages of LGM deep waters reflect a major drawdown of carbon from the atmosphere. The subsequent major deglacial age drop reflects changes in MOC accompanied by massive carbon releases to the atmosphere as recorded in Antarctic ice cores. These new features of MOC and the carbon cycle provide detailed evidence in space and time to test and refine ocean models that, in part because of insufficient spatial model resolution and reference data, still poorly reproduce our data sets.
1 Introduction

1.1 A variety of terms linked to the notion “$^{14}$C age”

The $^{14}$C concentration in the troposphere is mainly determined by $^{14}$C production, atmospheric mixing, air–sea gas exchange, and ocean circulation, which vary over time (e.g., Alves et al., 2018; Alveson, 2018). The $^{14}$C content of living terrestrial plants is in equilibrium with the atmosphere via processes of photosynthesis and respiration. Accordingly, the $^{14}$C of terrestrial plant remains in a sediment section directly reflects the amount of radioactive decay, and thus the time passed since the plant’s death and the $^{14}$C composition of the atmosphere during the time of plant growth.

In contrast, $^{14}$C values of marine and inland waters are cut off from cosmoenic $^{14}$C production in the atmosphere, and hence depend on the carbon transfer at the air–water interface and the result of local transport and mixing of carbon in the water. For surface waters, the air–sea transfer involves a time span of 10 years or less (e.g., Nydal et al., 1980). However, vertical and horizontal water mixing results in surface ocean $^{14}$C concentrations that are on average 5% lower than those in the contemporaneous atmosphere, a difference expressed as “Marine Reservoir Age” (or “reservoir effect”, sensu Alves et al., 2018). These “ages” reflect the local oceanography and are highly variable through time (~200–2500 years; e.g., Stuiver and Braziunas, 1993; Grootes and Sarnthein, 2006; Sarnthein et al., 2015). Apart from U/Th dated corals (many papers have been published on their reservoir age since Adkins and Boyle, 1997), the $^{14}$C age of planktic foraminifers is the most common tracer in marine sediments, providing a rough estimate of the time passed since sediment deposition. Soon after, however, marine geologists were confronted with age inconsistencies that implied a series of unknowns, in particular the surface ocean $^{14}$C “reservoir age” that finally became a most valuable tracer for oceanography.

The $^{14}$C records of benthic foraminifers in deep-sea sediments reflect the time of radioactive decay since their deposition with the apparent “ventilation age” of the deep waters in which they lived. Ventilation age is primarily the time span from the moment when carbon dissolved in the local surface waters with a somewhat reduced $^{14}$C level lost contact with the atmosphere until the precipitation of benthic carbonate from the down-welled deep waters. Details on the derivation of ventilation ages are provided in Cook and Keigwin (2015) and Balmer and Sarnthein (2018). In addition, however, ventilation ages include difficult to quantify lateral admixtures of older and/or younger water masses, as well as $^{14}$C-enriched organic carbon supplied by the biological pump, and are thus referred to as “apparent”. Today, the apparent transit times of carbon dissolved in the deep ocean range from a few hundred up to the ~1800 $^{14}$C years found in upper deep waters of the northeastern North Pacific (Matsumoto, 2007).

The reservoir ages of surface waters and the ventilation ages of deep waters present robust and high-resolution tracers essential for drawing quantitative conclusions on past ocean circulation geometries, marine climate change, and the processes that drive both past ocean dynamics and carbon budgets, given that the ages rely on a number of robust age tie points. Obtaining such tie points presents a problem, since any attempt to date a deep-sea sediment record by means of $^{14}$C encounters a number of intricacies of how to disentangle the effects of global atmospheric $^{14}$C variations due to past changes in cosmoenic $^{14}$C production and carbon cycle from (i) local depositional effects such as sediment hiatuses and winnowing, differential bioturbational mixing depths, and sediment transport by deep burrows; (ii) the effects of local atmosphere–ocean exchange and ocean mixing, resulting in reservoir and ventilation ages that change through time and space (e.g., Alves et al., 2018; Grootes and Sarnthein, 2006); and (iii) quantitatively “pure” $^{14}$C ages due to radioactive decay from the final target. These problems are exacerbated by the need for a generally accepted high-precision atmospheric reference record for the period 14–50 cal ka, which is beyond tree ring calibration.

Current $^{14}$C-based chronologies of deep-sea sediment records, used to constrain and correlate the age of glacial-to-deglacial changes in ocean dynamics and climate on a global scale, are often of insufficient quality when they are (i) based on age tie points that are spaced far too widely (e.g., using Dansgaard Oeschger (DO) events 1, 2, and 3 only and/or sporadic tephra layers for the time span 30–14 cal ka), (ii) disregarding atmospheric $^{14}$C plateaus, (iii) making the risky assumption of ± constant planktic $^{14}$C reservoir ages and other speculative stratigraphic correlations or compilations, and (iv) ignoring small-scale major differences in low-latitude reservoir age. Likewise, clear conclusions are precluded by an uncertainty range of 3–4 kyr that is sometimes accepted for tie points during the glacial-to-deglacial period (Stern and Lisiecki, 2013; Lisiecki and Stern, 2016), where significant global climate oscillations occurred on decadal-to-centennial timescales, as widely shown on the basis of speleothem and ice-core-based records (Steffensen et al., 2008; Svensson et al., 2008; Wang et al., 2001). Thus, marine paleoclimatic and paleoceanographic studies today focus on the continuing quest for a high-resolution and global (and thus necessarily atmospheric) $^{14}$C reference record.

1.2 Review of tie points used to fix calibrated and reservoir ages in marine $^{14}$C records

The tree-ring-based calibration of $^{14}$C ages provides a master record of decadal changes in atmospheric $^{14}$C concentrations back to ~14 cal ka (Reimer et al., 2013, 2020) with floating sections beyond (from ~12.5–14.5, around 29–31.5, and 43 cal ka; Turney et al., 2010, 2017; Reimer et al., 2020). The evolution of Holocene and late deglacial $^{14}$C ages with time is not linear but reveals variations with numerous dis-
Distinct jumps (i.e., rapid change) and (short) plateau-shaped (slow or no change or even inversion) structures indicative of fluctuations in atmospheric $^{14}$C concentration. Prior to 8500 BP, various plateaus extend over 400–600 cal year (yr) and beyond (Fig. 2). Given the quality of the tree ring calibration data, these fluctuations can be considered real and suitable for global correlation (Sarnthein et al., 2007, 2015; Umland and Thunell, 2017; Sarnthein and Werner, 2018). Air–sea gas exchange transfers the atmospheric $^{14}$C fluctuations into the surface ocean, where they can provide high-resolution tie points to calibrate the marine $^{14}$C record and marine reservoir ages back to $\sim 14$ ka (via $^{14}$C wiggle matching”). In the near future, however, it is unlikely that a continuous tree-ring-based record will become available to trace such atmospheric $^{14}$C variations further back over the period 14–29 cal ka crucial for the understanding of last-glacial-to-interglacial changes in climate. Hence, various other carbonate-based $^{14}$C archives have been employed for this period to reconstruct past changes in atmospheric $^{14}$C concentration and age and tie them to an “absolute” or “calibrated” (e.g., incremental and/or based on speleothem carbonate) age scale.

Suites of $^{14}$C ages of paired marine and terrestrial plant-borne samples, e.g., paired planktic foraminifers and wood chunks, provide the most effective but rarely realizable absolute-age markers and reservoir ages of local ocean surface waters (Zhao and Keigwin, 2018; Rafter et al., 2018; Schroeder et al., 2016; Broecker et al., 2004). Likewise, the alignment of $^{14}$C-dated variations in downcore sea-surface temperatures (SST) with changes in hydroclimate as recorded in age-calibrated sedimentary leaf-wax hydrocarbon isotope (δD) records from ancient lakes (Muschitiello et al., 2019) appears to be successful and is assumed to be coeval. Further tie points are derived from volcanic ash layers (Waelbroeck et al., 2001; Siani et al., 2013; Davies et al., 2014; Sikes and Guilderson, 2016), paired U/Th- and $^{14}$C-based coral ages (Adkins and Boyle, 1997; Robinson et al., 2005; Burke and Robinson, 2012; Chen et al., 2015), and the (fairly fragmentary) alignment of major tipping points in $^{14}$C dated records of marine SST and planktic δ$^{18}$O to the incremental age scale of climate events dated in polar ice core records (Waelbroeck et al., 2011). Such well-defined tie points, however, are spaced widely in peak glacial to early deglacial ice core records and too widely to properly resolve a clear picture of the spatiotemporal pattern of marine paleoclimate events. Finally, various data compilations tentatively rely on the use of multiple age correlations amongst likewise poorly dated marine sediment records, an effort that is necessarily problematic. Skinner et al. (2019) recently combined new and existing reservoir age estimates from North Atlantic and Southern Ocean to show coherent but distinct regional reservoir age trends in subpolar ocean regions, trends that indeed envelop the range of actual major small-scale and short-term oscillations in reservoir age revealed by our technique of $^{14}$C plateau tuning for the subpolar South Pacific (Küssner et al., 2020a).

Lacking robust age tie points, several authors resort to $^{14}$C reservoir age simulations for various sea regions by ocean General Circulation Models (GCMs) (e.g. Butzin et al., 2017; Muglia et al., 2018) to quantify the potential difference between marine and atmospheric $^{14}$C dates for glacial-to-interglacial periods. In view of the complexity of ocean Meridional Overturning Circulation (MOC) and the global carbon cycle, it is not surprising that the results of a comparison of a selection of robust empiric vs. simulated $^{14}$C reservoir ages are not that encouraging yet (as discussed further below).

Beyond accepting a generally close link between $^{14}$C concentrations in the troposphere and in the surface ocean, the fine structure of planktic $^{14}$C records with centennial-scale resolution can provide a far superior (though costly) link from the marine sediment records to the reference suite of narrow-standing jumps and boundaries of the plateaus robustly identified in the atmospheric $^{14}$C record of Lake Suigetsu, the only long and continuous record based on terrestrial plant remains (Bronk Ramsey et al., 2012, 2020). Beyond the reach of the tree-ring-based age scale ~ 14 cal ka, the absolute age of the Suigetsu atmospheric $^{14}$C structures can be either calibrated by incremental (microscopy- or XRF-based) varve counts (Schlalaut et al., 2018; Marshall et al., 2012) or by a series of paired U/Th- and $^{14}$C-based model ages correlated from the Hulu Cave speleothem record (Bronk Ramsey, 2012, 2020; Southon et al., 2012; Cheng et al., 2018). The difference in absolute age between these calibrations (Fig. 3) is of little importance for the tuning of planktic plateaus to corresponding atmospheric $^{14}$C plateaus and the derivation of planktic reservoir ages that present the highly variable offset of the $^{14}$C age of a planktic plateau from that of the correlated atmospheric plateau. The offset is deduced by subtracting the average $^{14}$C age of an atmospheric $^{14}$C plateau from that of the correlated planktic $^{14}$C plateau, independent of any absolute age value assigned.

The uncertainty of the Suigetsu atmospheric $^{14}$C record is significantly larger than that of the tree ring-based calibration record because of lower $^{14}$C concentrations, limited sampling density, and uncertainties in the independent age determination. Thus the $^{14}$C fluctuations could be real or represent mere statistical scatter (null hypothesis) in which case the record of atmospheric $^{14}$C ages against time would show a simple continuous rise resulting from radioactive decay and the advance of time, such as is suggested by a fairly straight progression of the highly resolved deglacial Hulu Cave $^{14}$C record plotted vs. U/Th ages (Southon et al., 2012; Cheng et al., 2018).

The unequivocal fluctuations in the tree-ring-based master record of atmospheric $^{14}$C concentration (Fig. 2; Reimer et al., 2013, 2020) are on the order of 2 %–3 % over the last 10 kyr (Stuiver and Brazianus, 1993) and even larger back to ~ 14 ka. Under glacial and deglacial low-CO$_2$ conditions be-
yond 14 ka, when climate and ocean dynamics were less constant than during the Holocene, real atmospheric $^{14}$C fluctuations were, most likely, even stronger and $^{14}$C plateaus and jumps were accordingly larger. Plateau–jump structures are also becoming increasingly evident in the evolving atmospheric calibration record (Reimer et al., 2020). The age-defined plateaus and jumps in the Suigetsu atmospheric $^{14}$C calibration curve may thus be regarded as a suite of “real” structures, extending the calibration provided by the tree ring record for the Holocene and Bølling–Allerød-to-Early Holocene periods (Fig. 2) into the early deglacial and Last Glacial Maximum (LGM) periods.

The plateau–jump structures may partly be linked to changes in cosmogenic $^{14}$C production, as possibly shown in the $^{10}$Be record (Fig. 4; based on data of Adolphi et al., 2018), and are presumably more dominant than short-term changes in ocean mixing and the carbon exchange between the ocean and the atmosphere. The exchange is crucial, since the carbon reservoir of the ocean contains up to 60 (preindustrial) atmospheric carbon units (Berger and Keir, 1984). The apparent contradiction with the smooth Hulu Cave $^{14}$C record (Southon et al., 2012; Cheng et al., 2018) may possibly be explained by the Hulu Cave speleothem precipitation system acting as a low-pass filter for fluctuating atmospheric $^{14}$C concentrations (following statistical tests made by Bronk Ramsey et al., 2020) and, to a very limited degree, by the obvious scatter in the Suigetsu data. The filter for Hulu data possibly led to a loss, especially of short-lived structures in the preserved atmospheric $^{14}$C record, though some remainders were preserved in the $^{14}$C records of Hulu Cave (Fig. 1). So we would rather trust the amplitude of Suigetsu $^{14}$C structures than the timing of Hulu Cave data.

Like a “rung ladder”, the age-calibrated suite of $^{14}$C plateau boundaries and jumps is suited for tracing the calibrated age of numerous plateau boundaries in glacial-to-deglacial marine $^{14}$C records that are likewise densely sampled, even when some rungs have been destroyed by local influences on gas exchange or ocean mixing. In addition, one may record the average offset of planktic $^{14}$C ages from paired atmospheric $^{14}$C ages, i.e. the planktic reservoir age, for each single $^{14}$C plateau (Sarnthein et al., 2007, 2015). We prefer the Suigetsu record to IntCal20, since it is based on original primary atmospheric data and results in small-scale spatiotemporal changes of reservoir age, whereas IntCal20 mixes and smooths a broad array of different data sources with comparatively coarse age resolution, including carbonate-based speleothem and marine records.

For the first time, this suite of tie points may facilitate a precise temporal correlation of all sorts of changes in surface and deep-water composition on a global scale, crucial for a better understanding of past changes in ocean and climate dynamics.

1.3 Items discussed in this synthesis

Section 2 summarizes (1) the means of separating noise and global atmospheric and local oceanic forcings that together control the structure of a planktic $^{14}$C plateau, (2) the choice of a U/Th-based reference timescale (Bronk Ramsey et al., 2012; Cheng et al., 2018) instead of the earlier varve-counted version (Schlautet al., 2018) to date the structures in the global atmospheric $^{14}$C record of Lake Suigetsu (Sarnthein et al., 2015), (3) the extension of the suite of age tie points back from 23 to 29 cal ka, values crucial for an accurate global correlation of ocean events over the Last Glacial Maximum, and (4) potential linkages of atmospheric $^{14}$C plateaus and jumps to cosmogenic $^{14}$C production and/or ocean dynamics.

Section 3 includes the following discussions and implications.

1. A global summary of published marine $^{14}$C reservoir age records (Sarnthein et al., 2015), now enlarged by nine plateau-tuned records from the Southern Hemisphere (Balmer et al., 2016; Balmer and Sarnthein, 2018; Kuüssner et al., 2018, 2020) and the northeastern Atlantic (Ausin et al., 2020a). In total, 18 (LGM) and 19 Heinrich Stadial 1 (HS-1), plus three wood chunk-based records (Broecker et al., 2004; Zhao et al., 2018), now depict the spatiotemporal variability of past reservoir ages of surface waters in different ocean regions.

2. A comparison of our plateau-based reservoir ages with LGM estimates of surface water $^{14}$C reservoir ages simulated by the GCM of Muglia et al. (2018).

3. More detailed insights into the origin of past changes in the global carbon cycle from glacial to interglacial times are provided by the enlarged set of $^{14}$C reservoir and ventilation ages that form a robust tracer of global circulation geometries and the inorganic carbon (DIC) dissolved in different basins of the ocean (Sarnthein et al., 2013).

The discussion highlights $^{14}$C plateau tuning and its revised calibrated timescale for global data–model intercomparison and a new understanding of ocean MOC during the LGM and its reversal during HS-1.

2 Results – age tie points based on $^{14}$C plateau boundaries

2.1 Suite of planktic $^{14}$C plateaus: means to separate global atmospheric from local oceanographic forcings

The basic assumption of the $^{14}$C plateau tuning technique is that the fine structure of fluctuations of the global atmospheric $^{14}$C concentration record can also be found in the surface ocean. In a plot of $^{14}$C age vs. calendar age such
fluctuations lead to a pattern of plateaus and jumps that correspond to decreases and increases in $^{14}$C concentration. Here we refer to the derivation and interpretation of planktic $^{14}$C plateaus, assuming a predominantly global atmospheric origin with occasional local oceanographic forcings. The series of planktic $^{14}$C plateaus and jumps are derived in cores with average hemipelagic sedimentation rates of $>10$ cm kyr$^{-1}$ and a dating resolution of $<100–150$ years. The plateau-specific structures in a sediment age–depth record form a well-defined suite for which absolute age and reservoir age are derived by means of a strict alignment to the reference suite of global atmospheric $^{14}$C plateaus as a whole. Initially, age tie points of planktic foraminiferal $^{18}$O records showing (orbital) isotope stages 1–3 serve as stratigraphic guideline for the alignment under the simplifying assumption that stratigraphic gaps are absent, which is not always true (Fig. S2). Planktic reservoir ages and their short-term changes are derived from the difference in average $^{14}$C age between atmosphere and surface waters in subsequent plateaus. To stick as close as possible to the modern range of reservoir ages (Stuiver and Braziunas, 1993), tuned reservoir ages are kept at a minimum unless stringent evidence requires otherwise.

A close correspondence between $^{14}$C concentrations in atmosphere and surface ocean is expected based on rapid gas exchange. In several cases, however, the specific struc-
Table 1. Summary of varve- and U/Th model-based age estimates (in cal ka) (Schlolaut et al., 2018; Bronk Ramsey et al., 2012) for ~30 plateau (pl.) boundaries in the atmospheric $^{14}$C record identified in Lake Suigetsu Core SG062012 (composite depth, c.d.) by means of visual inspection over the interval 10.5–27 cal ka (modified from the Supplement to Sarnthein et al., 2015). On the right-hand side, three columns give the average ($\bar{\Omega}$) and uncertainty range of $^{14}$C ages for each $^{14}$C plateau. YD stands for the Younger Dryas period. Bold and bold-italic fonts mark ages henceforth preferred in this paper; see Sect. 2.2. for more details.

| Suigetsu | Plateau top | Plateau base | $\bar{\Omega}^{14}$C age | ± Uncertainty $^{14}$C Plateau (1σ range) |
|----------|-------------|--------------|--------------------------|-----------------------------------------|
| SG06_2012 | Varve-based cal. age estimates | U/Th-based cal. age estimates | Varve-based cal. age estimates | U/Th-based cal. age estimates | Depth (cm c.d.) | Depth (cm c.d.) |
| “Preboreal” | 10 525 | 10 560 | 1325 | 11 100 | 11 108 | 1383 | 9525 | −170/+110 | 9356/9635 |
| “Top YD” | 11 290 | 11 281 | 1402 | 11 760 | 11 755 | 1453 | 10 060 | −100/+35 | 9963/10095 |
| “YD” | 11 950 | 12 490 | 1555 | 13 160 | 13 080 | 1582 | 11 000 | −85/114 | 10 915/11 114 |
| “No name” | 13 580 | 13 656 | 1626 | 13 980 | 13 970 | 1657 | 12 000 | −100/125 | 11 857/12 050 |
| 1 | 14 095 | 14 160 | 1666 | 15 095 | 15 100 | 1740 | 12 471 | −100/125 | 12 315/12 683 |
| 2a | 15 310 | 15 420 | 1754 | 16 140 | 16 150 | 1820 | 13 850 | −100/125 | 13 174/13 665 |
| 2b | 16 075 | 16 520 | 1802 | 16 400 | 16 410 | 1820 | 13 850 | −100/125 | 13 808/13 885 |
| 3 | 16 835 | 17 500 | 1847 | 17 500 | 17 510 | 1888 | 14 671 | −100/125 | 14 582/14 792 |
| 4 | 17 880 | 18 650 | 1913 | 18 830 | 18 850 | 1971 | 15 851 | −100/125 | 15 661/16 044 |
| 5a | 18 960 | 19 720 | 1978 | 19 305 | 19 320 | 2032 | 16 670 | −100/125 | 16 570/16 750 |
| 5b | 19 305 | 20 240 | 2003 | 20 000 | 20 010 | 2032 | 17 007 | −100/125 | 16 830/17 247 |
| 6a | 20 190 | 21 000 | 2050 | 20 920 | 20 930 | 2105 | 17 667 | −100/125 | 17 435/17 960 |
| 6b | 20 920 | 21 890 | 2105 | 21 275 | 21 280 | 2132 | 18 075 | −100/125 | 17 960/18 240 |
| 7 | 21 375 | 22 400 | 2140 | 21 790 | 21 800 | 2171 | 18 843 | −100/125 | 18 741/18 975 |
| 8 | 22 835 | 23 940 | 2175 | 22 730 | 22 740 | 2227 | 19 715 | −100/125 | 19 425/20 041 |
| 9 | 22 730 | 24 250 | 2257 | 23 395 | 23 400 | 2312 | 20 465 | −100/125 | 20 238/20 728 |
| 10a | 23 935 | 25 880 | 2358 | 25 085 | 25 090 | 2400 | 22 328 | −100/125 | 21 946/22 600 |
| 10b | 25 080 | 27 000 | 2400 | 25 800 | 25 800 | 2426 | 22 708 | −100/125 | 22 233/23 147 |
| 11 | 26 110 | 27 770 | 2443 | 27 265 | 27 830 | 2525 | 24 088 | −100/125 | 23 727/24 595 |

Figure 2. High-resolution record of atmospheric $^{14}$C jumps and plateaus (i.e., suite of labeled horizontal boxes that envelop scatter bands of largely constant $^{14}$C ages extending over > 300 cal yr) in a sediment section of Lake Suigetsu vs. tree-ring-based $^{14}$C jumps and plateaus 10–14.5 cal ka (Reimer et al., 2013). The blue line averages paired double and triple $^{14}$C ages of Suigetsu plant macrofossils. Age control points (cal ka) follow varve counts (Schlolaut et al., 2018) and U/Th-model-based ages of Bronk Ramsey et al. (2012). YD stands for Younger Dryas, and B/A stands for Bølling-Allerød.
ture and relative length of a planktic $^{14}$C plateau may de-
viate from those of the pertinent plateau observed within
the suite of atmospheric plateaus, and thus indicate local
intra-plateau changes of reservoir age. Though less frequent,
these changes may indeed amputate and/or deform a plateau,
as result of variations in local ocean atmosphere exchange
and oceanic mixing. Two aspects help to sort out short-term
climate-driven intra- and inter-plateau changes in $^{14}$C reser-
voir age. (i) The evaluation of the structure and reservoir age
of an individual plateau strictly includes the age esti-
mates deduced for the complete suite of plateaus. (ii) Our
experience shows that deglacial climate regimes in control
of changes in surface ocean dynamics generally occurred on
(multi-)millennial timescales (e.g., YD, B/A, HS-1), whereas
atmospheric $^{14}$C plateaus hardly lasted longer than a few
hundred up to 1100 years (Figs. 1 and S1). Abrupt changes
in gas exchange or ocean mixing usually affect one or only
a few plateaus of the suite. Absolute age estimates within a
plateau are derived by linear interpolation between the age of
the base and top of an undisturbed plateau assuming constant
sedimentation rates. The potential impact of short-term sed-
imentation pulses on $^{14}$C plateau formation has largely been
discarded by Balmer and Sarnthein (2016).

### 2.2 Suigetsu atmospheric $^{14}$C record: shift to a
chronology based on U/Th model ages

Originally, we based the chronology of $^{14}$C plateau bound-
aries in the Suigetsu record (Sarnthein et al., 2015) on a
scheme of varve counts by means of light microscopy of thin
sections (Bronk Ramsey et al., 2012; Schlolaut et al., 2018).
Over the crucial sediment sections of the Last Glacial Max-
imum (LGM) and deglacial Heinrich Stadial 1 (HS-1), how-
ever, varve quality and perceptibility in the Suigetsu profile
is highly variable (Fig. 5). In parallel, varve-based age esti-
mates were derived from counting various elemental peaks
in μXRF data and interpreted as seasonal signals (Marshall
et al., 2012). The results obtained from the two indepen-
dent counting methods and their interpolations widely sup-
port each other but diverge for older ages. The varve counts
ultimately formed the backbone of a high-resolution chronol-
gy obtained by tying the Suigetsu $^{14}$C record to the U/Th
based timescale of the Hulu cave $^{14}$C record (Bronk Ramsey
et al., 2012). Recently, Schlolaut et al. (2018) amended the
scheme of varve counts. Accordingly, Suigetsu varve preser-
vation (i.e., the number of siderite layers per 20 cm thick
sediment section) is fairly high prior to ~32 ka and over
late glacial Termination I but fairly poor over large parts of
the LGM and HS-1, from ~15–32 cal ka (17.3–28.5 m c.d.
in Fig. 5). Here only few than 20%–40% of the annual
layers expected from interpolation between clearly varred
sections are distinguished by microscope. Varve counts that
use μXRF data (Marshall et al., 2012) can distinguish sub-
tle changes in seasonal element variations that are not dis-
tinguishable in thin-section microscopy, and hence result in
higher varve numbers, especially during early deglacial-to-
peak glacial times. However, some subtle variations are dif-
ficult to distinguish from noise, which adds uncertainty to the
μXRF-based counts. Thus, the results from either counting
method are subject to uncertainties that rise with increased
varve age (Fig. 5).

Bronk Ramsey et al. (2012) established a third timescale
based on $^{14}$C wiggle matching to U/Th dated $^{14}$C records of the
Hulu Cave and Bahamian speleothems. In part, this cal-
ibrated (cal.) age scale was based on Suigetsu varve counts,
in part on the prerequisite of the best-possible fit of a pat-
tern of low-frequency changes in $^{14}$C concentration obtained
from Suigetsu and Hulu Cave. The two $^{14}$C records were fitt-
ed within the uncertainty envelope of the Hulu “Old and
Dead Carbon Fraction” (OCF and DCF) of $^{14}$C concentra-
tion. The uncertainty of this model is still not completely un-
derstood. The U/Th-based age model of Suigetsu may suf-
sfer from the wiggle matching of atmospheric $^{14}$C ages of
Lake Suigetsu with $^{14}$C ages of the Hulu Cave (Southon et
al., 2012) in case of major short-term changes in atmospheric
$^{14}$C concentration due to a memory effect of soil organic car-
bon in carbonate-free regions of the cave overburden. The
speleothem-carbonate-based Hulu ages may have been in-
fluenced far more strongly by short-term changes in the lo-
cal DCF than assumed, as suggested by major variations in
d in the Suigetsu chronology. We compared the results of the two timescales, which
were independently deduced from varve counts, with those of the U/Th-based model age scale using the base of $^{14}$C
Plateau 2b as a test case, which is the oldest tie point con-
strained by μXRF-based counts. In contrast to 16.4 cal ka,
proposed by optical varve counts, μXRF-based counts sug-
gest an age of ~16.9 cal ka (Marshall et al., 2012; Schlolaut
et al., 2018), which closely matches the U/Th-based esti-
mate of 16.93 ka. This is a robust argument for the use of the
U/Th-based Suigetsu timescale as it is the “best possi-
ble” age scale to calibrate the age of 30 $^{14}$C plateau bound-
aries (Fig. 1). In its older part, the U/Th model timescale
is further corroborated by a decent match of short-term in-
creases in $^{14}$C concentration with the low geomagnetic in-
tensity of the Mono Lake and Laschamp events at ~34 and
41.1 ± 0.35 ka (Lascu et al., 2016), independently dated by
other methods. The new U/Th-based model ages of $^{14}$C
plateau boundaries are significantly higher than our earlier
microscopy-based varve ages over HS-1 and LGM, a differ-
ence increasing from ~200 years near 15.3 cal ka to ~530
near 17 ka and 2000 years near ~29 ka (Fig. 3).

Note, any readjustment of the calendar age of a $^{14}$C plateau
boundary does not entail any change in $^{14}$C reservoir ages
before deduced for surface waters by means of the plateau

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respectively) by using a running kernel window (Sarnthein et al., 2018; Bronk Ramsey et al., 2020), we now extended our plateau tuning and also defined the boundaries and age ranges of \(^{14}\text{C}\) plateaus and jumps for the interval \(\sim 23–29\text{ cal ka}\), which results in a total of \(\sim 30\) atmospheric age tie points for the time span \(10.5–29\text{ cal ka}\) (Fig. 1; summary in Table 1; following the rules of Sarnthein et al., 2007, 2015). Prior to \(25\text{ cal ka}\), the definition of \(^{14}\text{C}\) plateaus somewhat suffered from an enhanced scatter of raw \(^{14}\text{C}\) values of Suigetsu. In addition to visual inspection, the \(^{14}\text{C}\) jumps and plateaus were also defined with higher statistical objectivity by means of the first derivative of all trends in the \(^{14}\text{C}\) age-to-calendar age relationship (or \(^{14}\text{C}\) age-to-core relationship, respectively) by using a running kernel window (Sarnthein et al., 2015).

2.3 Linkages of short-term structures in the atmospheric \(^{14}\text{C}\) record to changes in cosmogenic \(^{14}\text{C}\) production vs. changes in ocean dynamics

Potential sources of variability in the atmospheric \(^{14}\text{C}\) record were first discussed by Suviert and coworkers in the context of Holocene fluctuations deduced from tree ring data (e.g., Stuiver and Braziunas, 1993), which have more recently also been simulated (e.g., Hain et al., 2014). Similar to changes in \(^{14}\text{C}\), variations in \(^{10}\text{Be}\) deposition in ice cores reflect past changes in \(^{10}\text{Be}\) production as a result of changes in solar activity and the strength of the Earth’s magnetic field (Adolphi et al., 2018). If we omit assumptions on the modulation of past \(^{14}\text{C}\) concentrations by changes in the global carbon cycle, we can calculate the atmospheric \(^{14}\text{C}\) changes over last glacial-to-deglacial periods with \(^{10}\text{Be}\) and a carbon cycle model and convert them into \(^{14}\text{C}\) ages (Fig. 4). Changes in climate and carbon cycle over this period, however, necessarily modified the \(^{10}\text{Be}\)-based \(^{14}\text{C}\) record if included correctly into the modeling. Between 10 and \(13.5\text{ cal ka}\), the \(^{10}\text{Be}\)-modeled \(^{14}\text{C}\) record displays a number of plateau structures that appear to match the Suigetsu-based atmospheric \(^{14}\text{C}\) plateaus. Between 15 and \(29\text{ cal ka}\), however, \(^{10}\text{Be}\)-based \(^{14}\text{C}\) plateaus are more rare and/or less pronounced than those in the Suigetsu record. Most modeled plateaus are far shorter than those displayed in the suite of atmospheric \(^{14}\text{C}\) plateaus of Lake Suigetsu (e.g., plateaus near the top of 2a, 2b, the top of 5a, and 9), except for a distinct equivalent of plateau no. 6a. On the whole, the modeled and observed structures show little coherence. This may indicate that any direct relationship between variations in cosmogenic \(^{14}\text{C}\) production and the Suigetsu plateau record is largely obscured by the carbon cycle, uncorrected climate effects on the \(^{10}\text{Be}\) deposition, and/or noise in the \(^{14}\text{C}\) data. In addition, a relatively high uncertainty in the measured \(^{10}\text{Be}\) concentrations in the ice (in many cases \(\sim 7\%\); Raisbeck et al., 2017) and a lower sample resolution on the order of 50 to 200 years may contribute to the smoothed character of the \(^{10}\text{Be}\) record in Fig. 4.

On the other hand, the “new” \(^{14}\text{C}\) plateaus of plateau boundaries may suggest some reasonable stratigraphic correlations between peak glacial and deglacial change in atmospheric \(^{14}\text{C}\) and \(^{10}\text{Be}\) plateaus with millennial-scale events in paleoceanography (Fig. 6, Table 2): the suite of deglacial \(^{14}\text{C}\) plateaus no. 2a, 1, and Top YD indeed displays a temporal match with three brief but major deglacial jumps in ocean degassing of \(^{14}\text{CO}_2\) documented in the West Antarctic Ice Sheet Divide (WDC) ice core (Marcott et al., 2014). The two records have been independently dated by means of annual-layer counts in ice cores and \(^{14}\text{C}\) ages of stalagmites. The match suggests that these atmospheric \(^{14}\text{C}\) plateaus may largely result from changes in air–sea gas exchange and, in turn, from changes in ocean dynamics.

In particular, these events may have been linked to a variety of fast changes, such as in sea ice cover in the Southern Ocean and/or in the salinity and buoyancy of high-latitude surface waters (Skinner et al., 2010; Burke and Robinson, 2012). These factors control upwelling and meridional overturning of deep waters, in particular found in the Southern Ocean (Chen et al., 2015) and/or North Pacific (Rae et al., 2014; Gebhardt et al., 2008). Such events of changes in MOC geometry and intensity may be responsible for ocean degassing and the \(^{14}\text{C}\) plateaus. The enhanced mixing of the Southern Ocean and a similar, slightly later mixing event in the North Pacific (MD02-2489; Fig. S2d) may have triggered – with phase lag – two trends in parallel: (1) a rise in at-
mospheric CO$_2$, which was in part abrupt (sensu Chen et al., 2015; Menviel et al., 2018), and (2) a gradual enrichment in $^{14}$C depleted atmospheric carbon, reflected as a $^{14}$C plateau.

Plateau 6a matches a $^{14}$C plateau deduced from atmospheric $^{10}$Be concentrations, and thus suggests changes in $^{14}$C production. Other changes in atmospheric $^{14}$C (plateaus 4 and 8) match short-term North Atlantic warmings during peak glacial and earliest deglacial times, similar to that at the end of HS-1 and during plateau “YD”, and hence may reflect minor changes in ocean circulation and ocean–atmosphere exchange without major degassing of old $^{14}$C depleted deep waters in the North Atlantic (Table 2, Fig. S2a). There is still little information, however, on the origin of several other peak glacial $^{14}$C plateaus during 17.5–29 cal ka. The actual linkages of these plateaus to events in ocean MOC still remain to be uncovered.

### 3 Discussion and implications

#### 3.1 $^{14}$C plateau boundaries – a suite of narrow-spaced age tie points to rate short-term changes in marine sediment budgets, chemical inventories, and climate during 29–10 cal ka

In continuation of previous efforts (Sarnthein et al., 2007, 2015) the tuning of high-resolution planktic $^{14}$C records of ocean sediment cores to the new age-calibrated atmospheric $^{14}$C plateau boundaries now makes it possible to establish a “rung ladder” of ~ 30 age tie points covering the time span 29–10 cal ka. These global tie points have a time resolution of several hundred to thousands of years and are used to constrain the chronology and potential leads and lags of events that occurred during peak glacial and deglacial times (Fig. 1). The locations of 18 (20; depending on the age range covered) $^{14}$C records are shown in Fig. 7. Figures 8 and S2 give the time histories of the planktic and benthic reservoir ages; the information they provide is discussed below.

Six prominent examples showing the power and value of additional information obtained by means of the $^{14}$C plateau-tuning method are as follows.

- **i.** The timing of ocean signals of the onset of deglaciation (sudden depletion of planktic $\delta ^{18}$O and rise in SST) in the North Atlantic and North Pacific can now

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### Table 2. Temporal match of various $^{14}$C plateaus with deglacial periods of major atmospheric CO$_2$ rise and ocean warmings (AA stands for Antarctic, and GIS stands for Greenland Interstadial).

| $p$CO$_2$ rise (~ 12 ppm) | Plateau no. | Plateau boundaries |
|---------------------------|-------------|--------------------|
| AGE based on annual layers in AA ice core (Marcott et al. 2014) | “Top YD” | 11.83–11.3 |
| 11.7–11.5 | 1 | 15.1–14.2 |
| 14.8–14.53 | 2a | 16.52–15.5 |
| 16.4–16.15 | (data gap) | 17.3–17.1 |
| 17.4–~ 17.1 | | |

Further potential correlatives:
- Progressive N. Atlantic warming during the YD at 12.39–12.03 ka$^a$
- Onset of Antarctic$^b$ warming at 18.3–17.6 ka (ice-based timescale) | “YD” | 12.46–11.98 |
- Onset of North Atlantic$^c$ warming at 19.3–18.6 ka (U/Th-based timescale) | 3 | 18.22–17.5 |
- Top H2: GIS 2 N. Atlantic warming at 23.4–23.3 ka$^d$ | 4 | 19.6–18.65 |
- | 8 | 24.25–22.95 |

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Age control based on $^a$Naughton et al. (2019). Age control based on $^b$Kawamura et al. (2007). Age control based on $^c$Balmer and Sarnthein (2018). Age control based on $^d$Grootes and Stuiver (1997).
Figure 4. (a, b) Atmospheric $^{14}$C ages and plateaus (horizontal boxes) deduced from $^{10}$Be production rates vs. GICC05 age scale (Adolphi et al., 2018) compared to the Suigetsu record of atmospheric $^{14}$C plateaus vs. Hulu Cave U/Th-based model ages (Southon et al., 2012; Cheng et al., 2018) for the intervals (a) 10–20 and (b) 19–29 cal ka BP.

be distinguished in detail from those in the Southern Hemisphere, where warming began at 17.6 cal ka, when the cooling of Heinrich 1 started in the North Atlantic (Fig. S2) (Küssner et al., 2020a, in harmony with Schmittner and Lund, 2015), a finding important to further constrain global “bipolar see-saw” (Stocker and Johnsen, 2003).

ii. Likewise, the end of the cooling equated with the Antarctic Cold Reversal (ACR; WAIS Divide Project Members, 2013) in Pacific surface waters off Central Chile was found to be precisely coeval with the onset of the Younger Dryas cold spell in the Northern Hemisphere (Küssner et al., 2020a).

iii. Signals of local deep-water formation in the subpolar North Pacific can now be separated from signals originating in the North Atlantic (Rae et al., 2014; Sarnthein et al., 2013). In this way, we can now specify and tie major short-lasting reversals in Atlantic and Pacific MOC on a global scale.

iv. Signals of deglacial meltwater advection can now be distinguished from short-term interstadial warmings in the northern subtropical Atlantic, which helps to locate meltwater outbreaks far beyond the well-known Heinrich belt of ice-rafted debris (Balmer and Sarnthein, 2018).

v. As outlined above, the timing of marine $^{14}$C plateaus can now be compared in detail with that of deglacial events of climate and atmospheric CO$_2$ rise that are independently dated by means of ice core-based stratigraphy (Table 2; Fig. 6). These linkages offer a tool to explore details of deglacial changes in deep-ocean MOC once the suite of $^{14}$C plateaus has been properly tuned at any particular ocean site.

vi. The refined scale of age tie points also reveals unexpected details for changes in the sea ice cover of high latitudes, as reflected by anomalously high $^{14}$C reservoir
Figure 6. (a) Four sudden steps (pink bars) in the deglacial atmospheric CO$_2$ rise at West Antarctic Ice Sheet Divide ice core (WDC) reflect events of fast ocean degassing, that may have contributed to the origin of deglacial $^{14}$C plateaus. Age control based on ice cores (Marcott et al., 2014). (b) The steps are compared to suite of atmospheric $^{14}$C plateaus dated by Hulu U/Th-based model ages (Bronk Ramsey et al., 2012). Hol stands for Holocene, YD stands for Younger Dryas, B/A stands for Bølling-Allerød, HS stands for Heinrich stadials 1 and 2, LGM stands for Last Glacial Maximum, and GIS-2 stands for Greenland interstadial 2.

Finally, the plateau-based high-resolution chronology has led to the detection of numerous millennial-scale hiatuses (e.g., Sarnthein et al., 2015; Balmer et al., 2016; Küssner et al., 2020a) overlooked by conventional, e.g., AnalySerie-based methods (Paillard et al., 1996), of stratigraphic correlation (Fig. S2). In turn, the hiatuses give intriguing new insights into past changes of bottom current dynamics linked to different millennial-scale geometries of overturning circulation and climate change such as in the South China Sea (Sarnthein et al., 2013, 2015), in the South Atlantic (Balmer et al., 2016), and southern South Pacific (Ronge et al., 2019).

Clearly, the new atmospheric $^{14}$C “rung ladder” of closely-spaced chronostratigraphic tie points has evolved into a valuable tool to uncover functional chains in paleoceanography that actually have controlled events of climate change over glacial-to-deglacial times. The extension of the age range back to 29 ka allows for constraining potential changes in the ocean dynamics expected for Dansgaard-Oeschger (DO) events 2, 3, and 4 as compared to those found for DO-1, though pertinent core records are still missing.

3.2 Observed vs. model-based $^{14}$C reservoir ages that act as a tracer of past changes in surface ocean dynamics provide incentive for model refinements

Radioisotopic plateau tuning of marine sediment sections to the Suigetsu $^{14}$C atmospheric master record allows us to establish the difference between the average $^{14}$C age of coeval atmospheric and planktic $^{14}$C plateaus at semi-millennial-scale resolution. The suite of changing $^{14}$C reservoir ages over time forms a prime tracer of past ocean dynamics influencing local surface waters and a data set crucial to deducing past apparent deep-water ventilation ages (e.g., Muglia et al., 2018; Cook and Keigwin, 2015; Balmer and Sarnthein, 2018).

To better constrain the water depth of past reservoir ages, we dated monospecific planktic foraminifera (Sarnthein et al., 2007): at low latitudes to midlatitudes we used G. bul-
loides, G. ruber, or G. sacculifer with habitat depths of 0–80/120 m (Jonkers and Kucera, 2017), and at high latitudes we mostly used N. pachyderma (s) living at 0–200 m depth (Sistricht et al., 2003). Averaging of $^{14}$C ages within a $^{14}$C plateau helps to remove analytical noise and minor real $^{14}$C fluctuations. Nine plateaus are located in the LGM, 18–27 cal ka (Fig. 1). Here, planktic foraminifera-based reservoir ages show analytical uncertainties of $> 200$ to $> 300$ years each for standard Accelerator Mass Spectrometry (AMS) dating. By comparison, short-term temporal variations in reservoir age reach 200–400 years, and occasionally up to
Figure 8. Global distribution of $^{14}$C reservoir ages of Late LGM surface waters estimated (a) by means of $^{14}$C plateau tuning of planktic $^{14}$C records. (b) Model-based estimates (GCM of Muglia et al., 2018, assuming an Atlantic Meridional Overturning Circulation, AMOC, strength of 13 Sv) for sites with planktic foraminifera-based age values. The x–y graph (d) and map (c) show (rounded) differences between observed and modeled values and their intra-LGM trends. Minor differences are displayed in magenta, and larger differences of $>400$ years are shown in red. Planktic habitat depths and model estimates are largely confined to 0–100 m water depth. Arrows of surface currents delineate different sea regions important to assess potential limits of spatial extrapolation of reservoir ages. Distribution of core numbers and references for $^{14}$C records are given in Table 3 and Fig. 7a.

600 years, particularly when close to the end of the LGM (Table 3).

To better decode the informative value of our $^{14}$C reservoir ages for late LGM we compared average ages of $^{14}$C Plateaus 4–5 (18.6–20.9 cal ka) with estimates generated by various global ocean models, an approach similar to that of Toggweiler et al. (2019) applied to modern reservoir ages of the global ocean. In an earlier paper (Balmer et al., 2016), we compared our empiric reservoir ages for the LGM with GCM-based estimates of Franke et al. (2008) and Butzin et al. (2012). Franke et al. (2008) underestimated our midlatitude values by up to $\sim 2000$ $^{14}$C years, while LGM reservoir age estimates of Butzin et al. (2012) were more consistent with ours. Their GCM considered more realistic boundary conditions, such as the LGM freshwater balance in the Southern Ocean and, in particular, LGM SST and wind fields plus the gas transfer velocity for the exchange of $^{14}$C of CO$_2$ (Sweeney et al., 2007). Further improvements are expected from a model configuration that properly resolves the topographic details of the continental margins and adjacent seas, which frequently form the origin of our sediment-based data sets (Butzin et al., 2020). For the time being, we compared our empirical estimates with estimates from a coarse-resolution GCM, using the results by Muglia et al. (2018; 0–50 m water depth (w.d.); Fig. 8c, d; Table 3) as an example. Their model includes ocean surface reservoir age and ocean radiocarbon fields that have been validated through a comparison to LGM $^{14}$C data compilation made by Skinner et al. (2017). It conforms two plausible, recent model estimates of surface reservoir ages that can be compared to our results (Table 3).

Low LGM values (300–750 years) supposedly document an intensive exchange of surface waters with atmospheric CO$_2$, most common in model- and foraminifera-based estimates of the low-latitude and midlatitude Atlantic. Low empiric values also mark LGM waters in midlatitudes to
high latitudes off Norway and off central Chile, i.e., close to sites of potential deep and/or intermediate water formation. Off Norway and in the northeastern Atlantic, model-based reservoir ages of Muglia et al. (2018) largely match the empiric range. However, the uncertainty envelopes for data shown in Fig. 8c (±560 years; r = 0.59) generally by far exceed the spatial differences calculated for the empiric data. Conversely, model-based reservoir ages only poorly reproduce the low planktic foraminifera-based estimates off central Chile and values in the western Pacific and Southern Ocean.

In part, the differences may be linked to problems like insufficient spatial resolution along continental margins, ignoring east–west differences within ocean basins, and/or the estimates of a correct location and extent of seasonal sea ice cover used as LGM boundary condition, such as east of Greenland, in the subpolar northwest Pacific, and off southern Chile, where sea ice hindered the exchange of atmospheric carbon (per analogy to that of temperature exchange, e.g., Sessford et al. 2019). In addition, model estimates of the annual average are compared to 14C signals of planktic foraminifera that mostly formed during summer only, e.g., when large parts of the Nordic Seas were found to be ice-free (Sarnthein et al., 2003). Hence, models may need to better constrain local and seasonal sealing effects of LGM sea ice cover.

In general, the foraminifera-based reservoir age estimates for our sites that represent various hydrographic key regions in the high-latitude ocean appear to be much higher than model-derived values. These deviations reach up to 1400 years, particularly in the Southern Ocean. In part, they may result from the fact that present models may not yet be suited to capturing small-scale ocean structures such as the interference of ocean currents with local bathymetry and local upwelling cells. Here, model-based reservoir ages appear far too low in LGM regions influenced (i) by regional upwelling such as the South China Sea and thus governed by an estuarine overturning system (Wang et al., 2005; Fig. 9), (ii) by coastal upwelling off northwestern Australia (Xu et al., 2010; Sarnthein et al., 2011), or (iii) when stratified by a meltwater lid, such as off eastern New Zealand (Bostock et al., 2013; Küssner et al., 2020a). Local oceanic features are
likely to be missed in current resolution models. Our more narrow-spaced empiric data could help to refine the skill of models to capture past 14C reservoir ages.

Various differences amongst plankton- and model-based reservoir ages may also result from differential seasonal habitats of the different planktic species analyzed that, in turn, may trace different surface and subsurface water currents. Distinct interspecies differences were found in Baja California that record differential, upwelling-controlled habitats conditions (Lindsay et al., 2015). In the northern Norwegian Sea interspecies differences amount up to 600 years (Werner, 2018). Here 14C records of Arctic *Tuborotalita quinqueloba*, dominantly grown close to the sea surface during peak summer, differ from the paired record of *Neogloboquadrina pachyderma*, formed in subsurface waters, and that of subpolar species *N. incompta*, mainly advected from the south by Norwegian Current waters that are well mixed with the atmosphere during peak winter. This makes closer specification of model results as a product of different seasonal extremes a further target.

### 3.3 Planktic foraminifera-based 14C reservoir ages – a prime database to estimate past changes in the 14C ventilation age of deep waters and past oceanic MOC and DIC

“Raw” apparent benthic ventilation ages (in 14C years; “raw” sensu Balmer and Sarnthein, 2018) express the difference between the (coeval) atmospheric and benthic 14C levels measured at any site and time of foraminifer deposition. These ages are the sum of (1) the planktic reservoir age of the 14C age and the average 14C level has been subject

### Table 3. Continued.

| Sediment core | HS-1 pla. res. age | B/A pla. res. age | LGM be. vent age | LGM b.w. model age |
|---------------|-------------------|------------------|-----------------|-------------------|
| U/Th-based model age | 18–16.5 ka | 16.5–15.5 ka | 14.7–13.6 ka | (year) | strong AMOC weak AMOC |
| 14C Plateau (Pl.) no. | Pt. 3–2b (year) | Error (year) | Pt. 2a (year) | Error (year) | Pl. 1–1a (year) | Error (year) | early late (year) | (year) |
| Atlantic Ocean | PS2644 | 1775–1660 ±105 to 160 | 1900 ±355 – | – | 345 2400 | 948 918 |
| | GIK 23074 | 1730–2000 ±125 to 160 | 670 ±310 140–310 ±250 to ±100 | 375 375 | 960 931 |
| | MD08-3180 | 1420–1610 ±310 to ±160 | 1460 ±390 630–360 ±310 | 600 600 | 1031 1004 |
| | SHAK06-9K | 330–410 | 535 780–925 | – |
| | (= MD99-2334) | | | | |
| | ODP 1002 | –100 to 20 ±140 | 90 ±345 355 ±200 | – | 1247 1175 |
| | GeoB 3910-1 | 630–560 ±160 to ±180 | 175 ±475 210–230 ±220 to ±110 | 2150 2150 | – |
| | GeoB 1711-4 | 660–690 ±195 to ±45 | 420 ±320 880 ±255 | 1500 1500 | 1387 1714 |
| | KNR 159-3-6GGC | 460–340 ±380 to ±300 | 170 ±700 180–230 ±370 to ±310 | 1470 1470 | 1354 1563 |
| | MD07-3076 | 1650 ±180 – | 920 ±230 3640 3640 | 1653 2060 |
| Indian Ocean/Timor Sea | MD01-2378 | 740 ±125 – | 200–185 ±345 to ±135 | 2720 – | 1679 1881 |
| Pacific Ocean | MD02-2489 | 800–550 ±155 to ±120 | 550 ±305 440 ±285 | 2625 | 2332 2595 |
| | MD01-2416 | 1480–1140 ±135 to ±195 – | 720–570 ±285 to ±140 | 3700/5100 | 2400 2683 |
| | ODP 893A | 1065–1490 ±280 to ±125 | 1400 ±370 520 ±185 | 1430 | 1677 1705 |
| | MD02-2503 | 965–1365 ±160 to ±165 | 1215 ±325 395–535 ±240 to ±130 | – – | – |
| | GIK 17940 | 1210–1370 ±200 to ±470 | 1045 ±320 870–970 ±325 to ±100 | 3300–1800 | 1807 1897 |
| | (= SO50-37) | | | | 3225 3225 | 2373 2667 |
| | PS75/104-1 | 1050 ±265 1180 ±350 | 800 ±280 | – – – – |
| | (= SO213-84) | | | | | |
| | MD07-3088 | 800–1090 ±85 to ±125 | 1060 ±275 1310–730 ±125 to ±190 | 13607 | 1600 1808 1701 |
| | SO213-76-2 | 840 ±310 – | – | 3460 | 1712 2001 |
| | PS97/137-1 | 1500–670 ±90 to ±180 | 455 ±270 – | 1400–2400 | 2400/2900 | 1631 1871 |

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Table 3. Continued.

| Sediment core | Data source |
|---------------|-------------|
| Atlantic Ocean | Sarnthein et al. (2015), Benthic data supplemented |
| GIK 23074 | Sarnthein et al. (2015), Benthic data supplemented |
| MD08-3180 | Balmer and Sarnthein (2018) |
| SHAK06-5K | Ausin et al. (2020a) |
| (= MD99-2334) | Skinner et al. (2014) |
| ODP 1002 | Sarnthein et al. (2015) |
| GeoB 3910-1 | Balmer et al. (2016) |
| GeoB 1711-4 | Balmer et al. (2016) |
| KNR 159-5-36GGC | Balmer et al. (2016), data supplemented |
| MD07-3076 | Balmer et al. (2016) |

| Indian Ocean/Timor Sea | MD01-2378 | Sarnthein et al. (2015) |
|------------------------|-----------|-------------------------|
| Pacific Ocean | MD02-2489 | Sarnthein et al. (2015) |
| MD01-2416 | Sarnthein et al. (2015), modified |
| ODP 893A | Sarnthein et al. (2015), data supplemented |
| MD02-2503 | Sarnthein et al. (2015) |
| GIK 17940 | Sarnthein et al. (2015) |
| (= SOSO-37) | Sarnthein et al. (2015) |
| PS75/104-1 | Küssner et al. (2018, 2020a) |
| (= SO213-84) | Ronge et al. (2016) |
| MD07-3088 | Küssner et al. (2020a), Siani et al. (2013) |
| SO213-76-2 | Küssner et al. (2020a), Ronge et al. (2016) |
| PS97/137-1 | Küssner et al. (2020a), data supplemented |

Spheric $^{14}$C that occurred over the (short) time span between deep-water formation and benthic sediment deposition (e.g., Balmer and Sarnthein, 2018; Cook and Keigwin, 2015). In most cases, however, this second step is omitted since its application usually does not imply any major modification of the ventilation age estimates (Fig. S2a; Skinner et al., 2017; Sarnthein et al., 2013).

On the basis of $^{14}$C plateau tuning we now can rely on 18 accurately dated records of apparent benthic $^{14}$C ventilation ages (Fig. S2a–d) to reconstruct the global geometry of LGM and HS-1 deep and intermediate water circulation, as summarized in ocean transects and maps (Figs. 9–11) and discussed below. The individual matching of our 20 planktic $^{14}$C plateau sequences with that of the Suigetsu atmospheric $^{14}$C record is displayed in Sarnthein et al. (2015), Balmer et al. (2016), Küssner et al. (2020a), and Ausin et al. (2020a). In addition, robust estimates of past reservoir ages are obtained for four planktic and benthic $^{14}$C records from paired atmospheric $^{14}$C ages of wood chunks (Rafter et al., 2018; Zhao and Keigwin, 2018; Broecker et al., 2004).

3.3.1 Major features of ocean meridional overturning circulation during LGM (Fig. 10)

Off Norway and near the Azores Islands very low benthic $^{14}$C ventilation ages of $<100$–750 years suggest ongoing deep-water formation in the LGM northern North Atlantic reaching down to more than 3000–3500 m water depth, with a flow strength possibly similar to today (and a coeval deep countercurrent of old waters from the Southern Ocean flowing along the East Atlantic continental margin off Portugal). This pattern clearly corroborates the assembled benthic $\delta^{13}$C record showing plenty of elevated $\delta^{13}$C values for the northwestern, eastern, and central North Atlantic (Sarnthein et al., 1994; Millo et al., 2006; Keigwin and Swift, 2017). Irrespective of unspecified potential zonal variations in deep-water ventilation age at midlatitudes and different from a number of published models (e.g., Ferrari et al., 2014; Butzin et al., 2017), this “anti-estuarine” pattern has been confirmed by a global tracer transport model of Gebbie (2014), MIROC model simulations (Sherriff-Tadano et al., 2017; Yamamoto et al., 2019), and independently by $\varepsilon_{Nd}$ records (Howe et al., 2016; Lippold et al., 2016). The latter suggests an overturning of AMOC that is possibly even stronger than today, in particular due to a “thermal threshold” (Abé-Ouchi, 2018) overlooked in other model simulations.

In contrast to the northern North Atlantic, deep waters in the southern North Atlantic and circumpolar (CP) deep waters in the subpolar South Atlantic show an LGM $^{14}$C ventilation age of $\sim3640$ years, finally rising up to 3800 years (Figs. 10, 11, S2b). These waters were upwelled and admixed from below to surface waters near the sub-Antarctic Front.
during the terminal LGM (Fig. S2b; Skinner et al., 2010; Balmer and Sarnthein, 2016; model of Butzin et al., 2012). In the southwestern South Pacific abyssal, in part possibly Antarctic-sourced waters (Rae and Broecker, 2018) likewise show high apparent $^{14}C$ ventilation ages of 3500 years that drop to 2750 years near the end of the LGM (Figs. 10 and S2c) ($^{14}C$ ages of Ronge et al., 2016, modified by planktic $^{14}C$ reservoir ages of Küssner et al., 2020a). A vertical transect of benthic $\delta^{13}C$ (McCave et al., 2008) suggests that the abyssal waters were overlain by CP waters, separated by pronounced stratification near $\sim$3500–4000 m water depth. In part, the CP waters stemmed from North Atlantic Deep Water. Their apparent ventilation age 3500 years probably came close to the values found in the southern South Atlantic. East of New Zealand the CP waters entered the deep western Pacific and spread up to the subpolar North Pacific, where LGM $^{14}C$ ventilation ages reached $\sim$3700 years, possibly occasionally reaching 5000 years (Fig. S2d).

Similar to today, the MOC of the LGM Pacific was shaped by estuarine geometry, probably more weakened than today (Du et al., 2018) and more distinct in the far northwest than in the far northeast. This geometry resulted in an upwelling of old deep waters in the subarctic northwestern Pacific, here leading to a $^{14}C$ reservoir age of $\sim$1700 years for surface waters at terminal LGM. On top of the Lower Pacific Deep Waters, we may surmise Upper Pacific Deep Waters that moved toward south (Figs. 10top and 11).

The Pacific deep waters were overlain by Antarctic and Pacific Intermediate Waters (IW) with LGM $^{14}C$ ventilation ages as low as 1400–1800 years, except for a ice-covered shelf site at the southern tip of Chile with IW ages of 2400–2900 years, possibly a result of local upwelling of CP waters. In general, however, the low values of Pacific IW are similar to those estimated for South Atlantic IW and likewise reflect a vivid exchange with atmospheric CO$_2$ in their source regions in the Southern Ocean (Skinner et al., 2015).

When entering and crossing the entrance sill to the marginal South China Sea the “young” IW were mixed with “old” CP waters entrained from below, here leading to $^{14}C$ ventilation ages of 2600–3450 years (Figs. 9 and S2d). The LGM South China Sea was shaped by an estuarine-style overturning system marked by major upwelling near its distal end in the far southwest (Wang et al., 1999). This upwelling led to planktic $^{14}C$ reservoir ages as high as 1200–1800 years, values rarely found elsewhere in surface waters of low latitudes.

Our widely spaced distribution pattern of 18 open-ocean $^{14}C$ ventilation ages (plus four values based on paired wood chunks) in Figs. 10 and 11 agrees only in part with the circulation patterns suggested by the much larger data sets of $^{14}C$ ventilation ages compiled by Skinner et al. (2017) and Zhao et al. (2018). Several features in Figs. 10 and 11 directly deviate, e.g., the ages we derive for the North Atlantic and mid-depth Pacific. These deviations may be linked to both the different derivation of our $^{14}C$ ventilation age estimates and the details of our calendar-year chronology now based on the narrow-standing suite of $^{14}C$ plateau boundary ages. The quality of our $^{14}C$ reservoir ages of surface waters also controls the “apparent” ventilation age of deep waters, as it results from direct addition of the short-term average $^{14}C$ age of a planktic $^{14}C$ plateau to a paired, i.e., coeval benthic $^{14}C$ age (formed during the time of benthic foraminiferal growth, somewhat after the actual time of deep-water formation).

### 3.3.2 Major features of meridional overturning circulation during early HS-1 (Fig. 10)

Near the onset of deglacial Heinrich Stadial 1 (HS-1; $\sim$18–14.7 cal ka) major shifts in $^{14}C$ ventilation age suggest some short-lasting but fundamental changes in the circulation geometry of the deep ocean, a central theme of marine paleoclimate research (lower panel of Figs. 10, 11, and S2a, b). Deep waters in the eastern Nordic Seas, west of the Azores Islands,
Figure 10. The 2D transects of the geometries of global ocean MOC. Arrows (blue is high ventilation, and yellow is poor ventilation) suggest average deep and intermediate water currents that follow the gradient from low to high benthic ventilation ages based on paired planktic $^{14}$C reservoir ages derived by means of a $^{14}$C plateau tuning technique (Sarnthein et al., 2013; Balmer et al., 2018; Küssner et al., 2020a). At some Pacific sites reservoir ages are based on paired $^{14}$C ages of planktic foraminifera and wood chunks (marked by a green “w”; Sarnthein et al., 2015; Zhao and Keigwin, 2018; Rafter et al., 2018). Red arrows suggest poleward warm surface water currents. Zigzagging lines indicate major frontal systems separating counter rotating ocean currents (e.g., west of Portugal and north of MD07-307; following Skinner et al., 2014). (a, b) Late LGM circulation geometry (21–18.7 cal ka) that is largely similar to today. Note the major east–west gradient of ventilation ages in the central North Atlantic between Portugal (PORT) and the Mid-Atlantic Ridge west of the Azores (MAR). (c, d) HS-1 benthic ventilation ages reveal a short-lasting MOC reversal leading to Atlantic-style overturning in the subpolar North Pacific and coeval Pacific-style stratification in the northern North Atlantic, with seesaw-style reversals of global MOC at the onset and end of early HS-1 (first proposed by Broecker et al., 1985, albeit for LGM times). Increased ventilation ages reflect enhanced uptake of dissolved carbon in the LGM deep ocean (Sarnthein et al., 2013), major drops suggest major degassing of CO$_2$ from both the deep Southern Ocean and North Pacific during early HS-1. SCS is the South China Sea. AABW is Antarctic Bottom Water. AAIW is Antarctic Intermediate Water. NADW is North Atlantic Deep Water. Small arrows within age numbers reflect temporal trends. Many arrows are speculative, using circumstantial evidence of benthic $\delta^{13}$C records and local Coriolis forcing at high-latitude sites per analogy to modern scenarios. Location of sediment cores are given in Fig. 7, and short-term variations in planktic and benthic $^{14}$C reservoir and ventilation age are given in Fig. S2 and Table 3.

and off northern Brazil show a rapid rise to high $^{14}$C ventilation ages of ~2000–2500 years and up to 4000 years off Brazil, values that give the first proof of a brief switch from “anti-estuarine” to “estuarine” circulation that governed the central North Atlantic and Norwegian Sea during early HS-1. This geometry continued – except for a brief but marked and widespread event of recurring NADW formation near 15.2 ka – until the very end of HS-1 near 14.5 ka (Fig. S2a; Muschitiello et al., 2019). The MOC switch from LGM to HS-1 is in line with changes depicted in paired benthic $\delta^{13}$C data (Sarnthein et al., 1994), but not confirmed by the coeval $\varepsilon_{Nd}$ record that suggests a constant source of “mid-depth waters”, with the $\delta^{13}$C drop being simply linked to a higher age (Howe et al., 2018).

Conversely, benthic $^{14}$C ventilation ages in the northeastern North Pacific (Site MD02-2489) show a coeval and distinct but brief minimum of 1050–1450 years near 3640 m w.d. during early HS-1 (~18.1–16.8 ka; Figs. 10, 11, and S2d). This minimum was produced by extremely small benthic–planktic $^{14}$C age differences of 350–650 years and provides robust evidence for a millennial-scale event of deep-water formation, which has flushed the northeastern North Pacific down to more than 3640 m w.d. (Gebhardt et al., 2008; Sarnthein et al., 2013; Rae et al., 2014). Similar circulation geometries were reported for the Pliocene (Burls et al., 2017). “Young” Upper North Pacific Deep Waters (North Pacific Intermediate Waters sensu Gong et al., 2019) then penetrated as a “western boundary current” far to the south, up to the northern continental margin of the South China Sea (Figs. 9b, 11, and S2d). The short-lasting North Pacific regime of anti-estuarine overturning was similar to that we find in the modern and LGM Atlantic and, most interesting, simultaneous with the Atlantic’s estuarine episode.

Recent data on benthic–planktic $^{14}$C age differences (Du et al., 2018) precisely recover our results in a core at ~680 m w.d. off southern Alaska. However, they do not de-
Figure 11. Global distribution of $^{14}$C reservoir ages obtained (a) for late LGM intermediate waters (100–1800 m w.d.) and (b) for LGM deep waters (> 1800 m w.d., including Site GIK 23074 at 1157 m in the Norwegian Sea).

...describe the “young” deep waters at their Site U1418 at ~3680 m w.d., as corroborated by a paired authigenic $\epsilon_{Nd}$ maximum suggesting a high local bottom water age nearby. We assume that the amazing difference in local deep-water ventilation ages is due to small-scale differences in the effect of Coriolis forcing at high latitudes between a site located directly at the base of the Alaskan continental margin (U1418; Fig. 10b) and that on the distal Murray Sea Mount in the “open” Pacific (MD02-2489; Figs. 7 and 11), which probably has been washed by a plume of newly formed North Pacific deep waters probably stemming from the Bering and/or Okhotsk Seas. In contrast, the incursion of almost 3000 year old deep waters from the Southern Ocean has continued along the continental margin all over HS-1. In summary we may conclude that the geometry of ocean MOC was briefly reversed in the “open” North Pacific over almost 1500 years during HS-1, far deeper than suggested by previous authors (e.g., Okazaki et al., 2012; Gong et al., 2019) but similar to changes in geometry first proposed by Broecker et al. (1985) for an LGM ocean.

3.3.3 Deep-ocean DIC inventory

Apart from the changing geometries in ocean MOC during LGM and HS-1, the global set of $^{14}$C plateau-based (and hence refined) estimates of apparent $^{14}$C ventilation...
ages (Fig. 10) has ultimately also revealed new insights into glacial-to-deglacial changes in deep-ocean DIC inventories (Sarnthein et al., 2013; Skinner et al., 2019). On the basis of GLODAP data (Key et al., 2004), any drop in \(^{14}\)C concentration (i.e., any rise in average \(^{14}\)C ventilation age) of modern deep waters is tied linearly to a rise of carbon (DIC) dissolved in deep ocean waters below the base of the modern ocean. Any rise in average \(^{14}\)C concentration (i.e., any rise in average \(^{14}\)C ventilation age) of modern deep waters is tied linearly to a rise of carbon (DIC) dissolved in deep ocean waters below 2000 m, making for 1.22 mmol C/ \(\pm 1\%e\) \(^{14}\)C. By and large, GCM and box model simulations of Chikamoto et al. (2012) and Wallmann et al. (2016) suggest that this ratio may also apply to LGM deep-water circulation, when apparent \(^{14}\)C ventilation ages in the Southern Ocean increased significantly (from 2400 up to \(\sim 3800\) years), thermohaline circulation was accordingly more sluggish, and transit times of deep waters extended. Accordingly, a “back-of-the-envelope” calculation of LGM ventilation age averages in the global deep ocean suggests an additional carbon absorption of 730–980 Gt (Sarnthein et al., 2013). This estimate can easily accommodate the glacial transfer of \(\sim 200\) Gt C from the atmosphere and biosphere and may also explain 200–450 Gt C that was most probably removed from glacial Atlantic and Pacific intermediate waters. These estimates offer an independent evaluation of ice core-based data, other proxies, and model-based data on past changes in the global carbon cycle (e.g., Menviel et al., 2018).

### 4 Some conclusions and perspectives

- Despite some analytical scatter, \(^{14}\)C ages for the top and base of Lake Suigetsu-based atmospheric \(^{14}\)C plateaus and coeval planktic \(^{14}\)C plateaus do not present statistical “outliers” but instead show real age estimates that are reproduced by tree-ring-based \(^{14}\)C ages over the interval 10–13 cal ka and further back.
- Hulu Cave U/Th model-based ages of \(^{14}\)C plateau boundaries of the Suigetsu atmospheric \(^{14}\)C record appear to be superior to those derived from microscopy-based varve counts only, since U/Th model-based ages match far more closely the age when now deduced from XRF-based varve counts for the tie point of lower plateau boundary 2b, a test case in the early deglacial, and for the age assigned to the Laschamp event prior to the LGM.
- During deglacial times, we show that several atmospheric \(^{14}\)C plateaus paralleled a rise in air–sea gas exchange and in turn distinct changes in ocean MOC. Changes in cosmogenic \(^{14}\)C production rarely provide a complete explanation for the plateaus identified in the Suigetsu \(^{14}\)C data under discussion.
- In total, \(^{14}\)C plateau boundaries in the range now provide a suite of \(\sim 30\) age tie points to establish – like chronological ladder rungs – a robust global age control for deep-sea sediment sections and global stratigraphic correlations of last glacial to deglacial climate events, 29–10 cal ka. U/Th model ages confine the calibrated age uncertainty of Suigetsu plateau boundaries assigned halfway between two \(^{14}\)C ages nearby inside and outside a plateau’s scatter band to less than \(\pm 50\) to \(\pm 70\) years. Nevertheless, stratigraphic gaps may hamper the accurate tuning of planktic \(^{14}\)C plateaus to their atmospheric equivalents and result in major discrepancies.

- The difference in \(^{14}\)C age between coeval atmospheric and planktic \(^{14}\)C plateaus presents a robust tracer of planktic \(^{14}\)C reservoir ages and shows their high temporal and spatial variability for the LGM and HS-1 that is now established for 18 and 20 sediment sites, respectively.
- Paired reservoir ages obtained from different planktic species document the local distribution patterns of different surface water masses and prevailing foraminiferal habitats at different seasons are still insufficiently considered in model simulations.
- New, more robust deep-water \(^{14}\)C ventilation ages, derived on the basis of our robust planktic \(^{14}\)C reservoir ages, reveal geometries of LGM overturning circulation similar to those of today. In contrast, \(^{14}\)C ventilation ages of early HS-1 suggest an almost 1500-year event of widely reversed circulation patterns marked by deep-water formation and brief flushing of the northern North Pacific and estuarine circulation geometry in the northern North Atlantic.
- Increased glacial \(^{14}\)C ventilation ages and carbon (DIC) inventories of ocean deep waters suggest an LGM drawdown of about 850 Gt C into the deep ocean. Starting with HS-1 a drop of ventilation age suggests carbon released to the atmosphere (Sarnthein et al., 2013).
- Site-specific comparisons of planktic and model-based reservoir age estimates highlight the need for further model refinements to make them better reflect the real complex patterns of ocean circulation, including seasonality.

**Data availability.** Published primary radiocarbon data of all sites are available at PANGAEA. The \(^{14}\)C data of five marine sediment cores still under publication by Küssner et al. (2020a) and Ausin et al. (2020a; see the caption of Fig. S2) are deposited at PANGAEA under https://doi.org/10.1594/PANGAEA.922671 (Küssner et al. 2020b) and https://doi.pangaea.de/10.1594/PANGAEA.921812 (Ausín et al., 2020b).

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Author contributions. All authors contributed data and valuable suggestions to write up this synthesis. MS and PG designed the outline of the manuscript. KK, BA, TE, and MS provided new marine \(^{14}\)C records in addition to records previously published. GS displayed the details of Suigetsu varve counts. RM provided a \(^{10}\)Be-based \(^{14}\)C record and plots of raw \(^{14}\)C data sets of Suigetsu und Hulu Cave. Discussions amongst PG, RM, GS, and MS served to select U/Th-based model ages at the best possible timescale. JM streamlined the sections on data–model intercomparison.

Competing interests. The authors declare that they have no conflict of interest.

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