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Konrad A. Hughen, *et al.*

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pose of v. This computation yields a downslope projected velocity of 5.5 times the measured range-change rate.
16. Data obtained from the NOAA National Climatic Data Center (http://cdo.ncdc.noaa.gov/CDO/cdo).
17. Time intervals between each pair of SAR satellite scenes varied from year to year. To avoid bias in seasonal cumulative displacements, we manually inspected the range-change time series and found that displacement of the slides occurred almost exclusively during the wet season, so displacements from the wet season capture the total yearly change.
18. The north slide was closest to a M_L (catalog local magnitude) = 4.1 earthquake on 4 December 1998. Although the temporal resolution of the time series is insufficient to clearly resolve triggered sliding caused by this event, the seasonal displacement along the north slide during this period was anomalously high given the amount of precipitation during this interval (*8*), possibly indicating that the sliding may have been accelerated by seismic shaking.
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27. Head scarps of these types of slides show evidence of fissuring and failure in tension, so *CFF* may not provide direct information about where this fissure may reach the surface. However, opening localizes where displacement gradients along the landslide failure surface are large, and so we expect these features to root into the deeper landslide surface where *CFF* is large.
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Materials and Methods
Fig. S1
References

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Abrupt Tropical Vegetation Response to Rapid Climate Changes

Konrad A. Hughen,[1]* Timothy I. Eglinton,[1] Li Xu,[2] Matthew Makou[2]

Identifying leads and lags between high- and low-latitude abrupt climate shifts is needed to understand where and how such events were triggered. Vascular plant biomarkers preserved in Cariaco basin sediments reveal rapid vegetation changes in northern South America during the last deglaciation, 15,000 to 10,000 years ago. Comparing the biomarker records to climate proxies from the same sediment core provides a precise measure of the relative timing of changes in different regions. Abrupt deglacial climate shifts in tropical and high-latitude North Atlantic regions were synchronous, whereas changes in tropical vegetation consistently lagged climate shifts by several decades.

In order to evaluate the relative roles of high and low latitudes in initiating and propagating abrupt global climate changes, we need precise information regarding the relative timing of abrupt changes in different regions. Dating uncertainties, however, are typically too large to constrain the timing of the briefest decadal events in records from different sites (*1*–*4*). Another approach is to identify high- and low-latitude climate proxies in the same high-resolution record, and determine the relative timing of changes stratigraphically (*3*–*6*). For instance, increased methane concentrations attributed to the expansion of tropical wetlands (*7*) have been measured in air trapped in Greenland ice and used to infer shifts to warmer and/or wetter tropical climate during the abrupt Glacial/Bølling and Younger Dryas/Preboreal transitions (*3*, *4*). Temperature changes over Greenland were also reconstructed from the same samples using nitrogen and argon isotopes, allowing the precise identification of relative timing for rapid changes between the tropics and high latitudes. The tropics were found to lag Greenland by 20 to 30 and 0 to 30 years for the Bølling and Preboreal warmings, respectively (*3*, *4*), favoring a North Atlantic trigger at least for the Bølling event (*4*). In similar studies, radiocarbon measured in planktonic foraminifera from tropical Cariaco basin sediments was shown to have increased steeply during the onset of Younger Dryas cooling (*5*, *6*). Atmospheric concentration of cosmogenic beryllium-10 does not show a similar increase during the Younger Dryas onset (*8*), suggesting that the ^{14}C increase was not caused by changes in production rate, but that it instead reflects an abrupt decrease in high-latitude North Atlantic Deep Water (NADW) formation and export (*5*, *6*, *9*). In addition, relative reflectance (gray scale) and laminae thickness from the same sediments reveal rapid shifts in Cariaco upwelling and trade-wind intensity caused by shifts in the mean position of the Intertropical Convergence Zone (*2*, *5*, *6*). Direct comparison of Cariaco radiocarbon and gray-scale data showed that high- and low-latitude climate shifts during the onset of Younger Dryas cooling were synchronous within 10 years (*6*), and thus allows either a North Altantic or a tropical trigger for this rapid climate cooling. If abrupt deglacial warming and cooling events were manifestations of the same millennial-scale shifts in global climate, this subtle discrepancy in timing between Cariaco and Greenland data must be resolved before we can understand the mechanisms responsible for abrupt climate change.

The delayed increase in atmospheric methane following abrupt warmings may have been caused by the release of gas hydrates rather than expansion of tropical wetland vegetation (*10*). However, studies have shown fluctuations in deglacial tropical moisture balance similar in timing to the Bølling/Allerød and Younger Dryas oscillations in the North Atlantic region (*11*–*17*), supporting tropical wetlands as the source for the atmospheric methane signal. Specifically, detailed pollen records from Central and northern South America (*11*, *12*), including the Cariaco basin watershed (*13*, *14*), show that vegetation shifted between predominantly dry grasslands during the Glacial and Younger Dryas periods, and wet montane forest during the Bølling/Allerød and Preboreal periods. The lag of the methane increase behind Greenland warming may have occurred because the time scale necessary for tropical wetland expansion and development of anoxia following a climate shift may be longer than previously thought (*4*, *18*). However, the response time for changes in vegetation fol-

[1]Department of Marine Chemistry and Geochemistry, [2]Department of Geology and Geophysics, Woods Hole Oceanographic Institution, Woods Hole, MA 02543, USA.

*To whom correspondence should be addressed. E-mail: khughen@whoi.edu

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lowing abrupt shifts in climate is currently unknown, and the precise temporal relation between tropical vegetation shifts and climate changes in tropical and high-latitude North Atlantic regions remains a major question. We address this question by using multiple-proxy records of terrestrial vegetation change from vascular plant biomarkers preserved in Cariaco basin sediments, providing direct comparisons to climate proxies from the same core in order to evaluate precisely the timing of tropical vegetation change relative to local and regional climatic forcing (*18*).

Certain classes of organic compounds that are sequestered in sediments can be traced directly to specific sources, such as aquatic versus terrestrial vegetation, or woody versus herbaceous land plants (*19*). In particular, long-chain (C_{25}-C_{33}) *n*-alkanes exhibiting an odd-carbon-number predominance, and C_{24}-C_{32} *n*-alkanoic acids and *n*-alcohols exhibiting an even-carbon-number predominance [e.g., high Carbon Preference Index (CPI)] (*20*), are synthesized nearly exclusively by vascular plants as components of epicuticular leaf waxes, whereas shorter chain homologs may have a large marine component (*19*, *21*, *22*). The preservation of these lipids in sediments is generally good (*19*), and their analysis in sediment cores can be used to study vegetation change or patterns of atmospheric circulation (*23*–*28*). Although these compounds have been shown to be transported long distances by eolian processes (*24*, *28*), studies of lake and marine sediments show the predominant source of plant waxes to be local vegetation in the immediate watershed (*23*, *25*–*27*). Thus, a near-shore marine basin such as Cariaco is expected to have the greatest concentration of plant waxes derived from local sources and not from long-distance transport.

Stable carbon isotopes in vascular plant biomarkers can be used to distinguish between vegetation using different metabolic pathways, such as C_3 versus C_4 carbon fixation (*29*). The C_4 pathway is used by plants in arid environments to limit water loss and results in less isotopic fractionation (heavier $\delta^{13}C$ values) than the C_3 pathway (*29*). Although leaf waxes are typically depleted in ^{13}C relative to bulk tissue, this isotopic distinction is preserved. Those leaf waxes from C_4 plants typically have $\delta^{13}C$ values of about −23‰, whereas C_3 leaf wax $\delta^{13}C$ values are about −34‰ (*30*, *31*). Such a large isotopic signal can readily be discerned in biomarkers from dry grassland versus more humid forest vegetation sources, even in marine sediments (*26*, *27*). In addition to the information available from carbon isotopes, distributional variations in plant biomarker homologs also record differences between "arid" and "humid" vegetation sources. For example, atmospheric dust samples collected in transects along the West African coast reveal changes in the chain-length distributions of *n*-alkanes responding to aridity in the source regions (*26*, *27*). Similarly, *n*-alkanes from southern Asian paleosols and Bengal Fan sediments show chain-length variability that correlates strongly with *n*-alkane $\delta^{13}C$, reflecting the Miocene expansion of C_4 grasses (*31*). Multiple independent proxies for terrestrial vegetation can thus be used to provide a check against factors other than source vegetation that may bias either record.

Cariaco sediment samples were obtained from high–deposition rate piston core PL07-58PC (*2*, *6*), taken on the saddle between eastern and western subbasins where interferences from turbidites and sediment slumping are minimal (*32*). The sediments are annually laminated, and a floating varve chronology between 15,000 and 10,000 calendar years before present (15 to 10 ka B.P.) was anchored to an absolute age scale by cross-correlating variations in foraminiferal ^{14}C with ^{14}C in the newly revised Preboreal Pine tree-ring chronology (*33*, *34*). Organic carbon concentrations throughout the last deglaciation are ~2 to 5%, indicating good preservation of biomarker compounds. We investigated *n*-alkanoic acids because of their high concentrations and because *n*-alkanes showed evidence of contamination (e.g., low CPI) from migration of fossil hydrocarbons in the region. Concentrations of individual long-chain ($\geq C_{24}$) *n*-alkanoic acids ranged from 0.5 to 5 μg per gram of dry weight. Molecular isotopic measurements were made on C_{14}-C_{32} homologs (*32*).

The $\delta^{13}C$ values for C_{30} and C_{32} *n*-alkanoic acids were enriched by 4 to 5‰ during the Glacial period relative to the Preboreal and Holocene, reflecting a greater proportion of waxes from C_4 grasslands during cold, dry climates (Fig. 1A). This trend is absent from shorter-chain C_{16} *n*-alkanoic acids, likely to derive predominantly from marine sources (*35*) (fig. S1). The $\delta^{13}C$ data show rapid shifts coincident with the Glacial-Bølling and Younger Dryas–Preboreal transitions (Fig. 1A), indicating that C_3 vegetation biomass favoring humid conditions expanded abruptly at those times. Also, $\delta^{13}C$ variability coincides with some centennial-scale variability, including early Bølling and Preboreal cool/dry oscillations at ~14.4 ka B.P. and ~11.3 ka B.P., respectively (Fig. 1A).

An index of average chain length (ACL) (*36*) for C_{24}-C_{28} *n*-alkanoic acid homologs also shows abrupt changes during millennial-scale deglacial oscillations (Fig. 1B). Thus, tropical vegetation biomass in

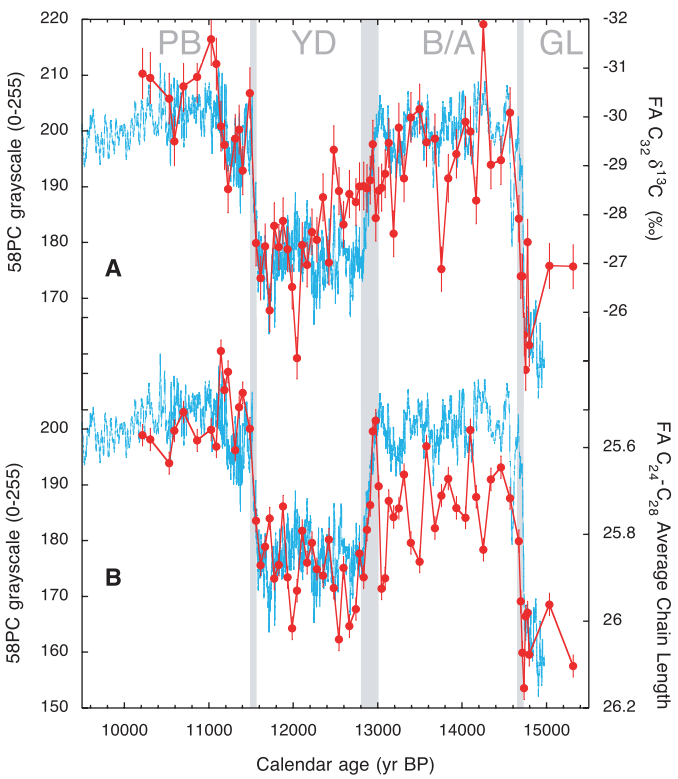


Fig. 1. Analysis of *n*-alkanoic fatty acid (FA) leaf waxes from Cariaco basin sediments. (**A**) $\delta^{13}C$ values of long-chain (C_{32}) *n*-alkanoic acids (red circles) are enriched during Glacial (GL) and Younger Dryas (YD) climate periods, consistent with the increased presence of arid C_4 grasslands. Depleted values consistent with more humid C_3 forest vegetation occur during wetter Bølling/Allerød (B/A) and Preboreal (PB)/Holocene periods. Similar results were observed for C_{30} chain-length FAs (fig. S1). (**B**) An average chain length (ACL) index for C_{24}-C_{28} *n*-alkanoic fatty acids (red circles) (*36*) shows the same deglacial oscillations as $\delta^{13}C$, but includes a rapid transition to drier C_4 vegetation at the beginning of the Younger Dryas. Both biomarker indices are plotted with sediment reflectance (gray scale), a proxy for upwelling and trade-wind intensity (*2*) (blue lines). Abrupt transitions between climatic periods are indicated by gray bars. Error bars are 1σ.

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the Cariaco region shifted rapidly between arid grassland and wet forest during deglaciation, in phase with gray scale and the characteristic trend of climate oscillations in the high-latitude North Atlantic. ACL records an abrupt shift to drier C_4 vegetation during the onset of the Younger Dryas, whereas $\delta^{13}C$ changes more gradually (*37*).

We have chosen to construct high-resolution vegetation records by using a molecular (isotopic) approach because biomarkers provide information independent from, and complementary to, pollen records. For example, the molecular and isotopic composition of leaf waxes are likely indicators of vegetative biomass, rather than reproduction (pollen). In some circumstances, plants can be induced to produce pollen under conditions of stress rather than optimum growth, leading to potential ambiguity in interpreting abrupt pollen shifts (*38*). Similarly, some vegetation living near the limit of its range may be able to survive for years without reproducing, providing a continuous biomarker signal but no pollen. To the extent that plant waxes record biomass of exposed leaf surface area, there is the potential for resolving extremely rapid responses to climate change.

Biomarkers from terrestrial vascular plants must be transported from the continental to the marine environment before deposition in sediments. For plant waxes, there is abundant evidence that this residence time is short, and that shifts recorded in biomarkers from annually laminated Cariaco basin sediments will not substantially lag the corresponding shifts in terrestrial vegetation. Plant wax compounds are transported primarily as aerosols via eolian pathways and are not likely to experience extended periods suspended in the atmosphere (*24*, *26*–*28*). This has been demonstrated with aerosols collected over 3 years in an air-monitoring system in Bermuda that showed a strong seasonal signal in the molecular and isotopic composition of North American land-plant–derived *n*-alkanoic acids, *n*-alcohols, and *n*-alkanes (*28*). The observed seasonality indicates that leaf waxes are introduced into the atmosphere mainly by wind ablation off living vegetation rather than by remobilization of detrital waxes during soil deflation (*28*) and implies a transport time on the order of days to weeks, much shorter than the time span encompassed by individual Cariaco sediment samples (~10 to 15 years).

Previous work has shown that trade-wind strength in the Cariaco region intensified during the Younger Dryas and Glacial periods (*2*, *39*) and that transport of dust from arid African source regions to Cariaco sediments increased during Glacial intervals (*40*). We therefore investigated the possibility that the Cariaco C_4 vegetation signal during Glacial and Younger Dryas climates was not caused solely by changing tropical vegetation, but also resulted from increased long-distance transport and mixing of leaf waxes between arid northwest Africa and humid northern South America. A plot of $\delta^{13}C$ versus ACL for Cariaco basin leaf waxes reveals a significant linear correlation ($r = 0.64$, $P < 0.001$), with "interstadial" and "stadial" end members clearly delineated (Fig. 2). If the Cariaco "stadial" vegetation signals were an artifact of mixing "interstadial" leaf waxes with those transported from arid African regions, the potential African contaminant would lie to the right of the Cariaco Glacial end member along the mixing line. However, Holocene leaf waxes from dust and marine sediments off the coast of northwest Africa (*26*) lie away from the mixing line normal to the Cariaco distribution (Fig. 2). The Cariaco stadial vegetation signals therefore cannot be explained as an artifact of mixing interstadial leaf waxes between tropical sources with different aridities and C_4 plant concentrations, but must rather reflect true changes in northern South American vegetation during cold, dry Glacial and Younger Dryas climates (*41*).

To evaluate the relative timing between Cariaco basin records of climate and vegetation change, we used a stepped cross-correlation approach (*6*, *32*, *42*). We focused initially on the onset of the Younger Dryas event, because it can be identified clearly in Cariaco proxies documenting shifts in North Atlantic ($\Delta^{14}C$) and tropical (gray scale) climate, as well as tropical vegetation (ACL) (*43*) (Fig. 3). Applying this technique to $\Delta^{14}C$ and gray-scale records confirms previous results that high- and low-latitude climate shifts were synchronous (lag of 0 ± 15 years) at that time. Analysis of ACL versus $\Delta^{14}C$ shows that tropical vegetation change lagged North Atlantic climate by 50 ± 25 years (*44*) (Fig. 3). Unfortunately, $\Delta^{14}C$ data across the Younger Dryas/Preboreal and Glacial/Bølling transitions do not show a distinct North Atlantic Deep Water (NADW) signal (*6*), and tropical versus high-latitude comparisons are not possible for these events. However, cross-correlating ACL with gray scale at both of these rapid shifts to warmer and/or wetter conditions reveals that Cariaco vegetation lagged local tropical climate change by at least 25 ± 15 years (*44*) (Fig. 4), and confirms a vegetation lag of 50 ± 25 years at the Younger Dryas onset (Fig. 3). Comparison of leaf wax $\delta^{13}C$ to ACL at the abrupt Bølling and Preboreal transitions shows synchronous changes, a lag of 0 ± 15 years; comparison of leaf wax $\delta^{13}C$ to gray scale shows the same vegetation-climate lag as ACL, 25 ± 15 years (*44*) (Fig. 4). The time scale of vegetation response to rapid climate change thus appears to depend on the rate of change in climate (*45*). This may indicate the ability of some plants to adapt for a longer period, before being replaced by other species, when faced with gradual rather than rapid climate changes.

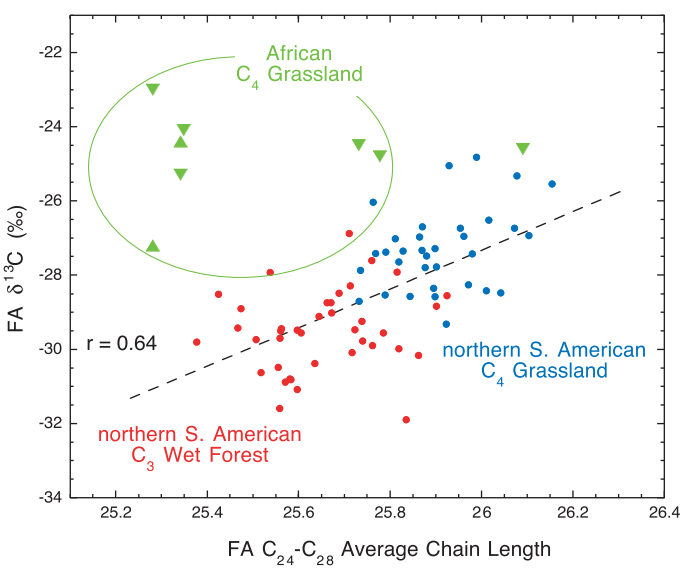


Fig. 2. Comparison of $\delta^{13}C$ for C_{32} and ACL for C_{24}-C_{28} *n*-alkanoic fatty acids (FAs) from Cariaco basin core PL07-58PC. $\delta^{13}C$ and ACL show a significant positive correlation ($r = 0.64$, $P < 0.001$), indicated by the dashed line. A similar correlation and regression line are seen between ACL and C_{30} fatty acid $\delta^{13}C$ ($r = 0.51$, $P < 0.001$) (fig. S2). Data points from warm/wet B/A and PB/Holocene "interstadial," and cool/dry GL and YD "stadial" periods, identified on the basis of gray scale for the same core, are plotted as red and blue circles, respectively. The same indices measured on *n*-alkanoic acid leaf waxes from Holocene African C_4 vegetation are shown for C_{32} FAs (green triangles) and C_{30} FAs (inverted green triangles) (*26*). The majority of African data points lie normal to the South American data distribution and regression line and are inconsistent with potential contamination of Cariaco "stadial" C_4 signals with long-distance transport and mixing of "interstadial" leaf waxes during cold, windy climates (*41*).

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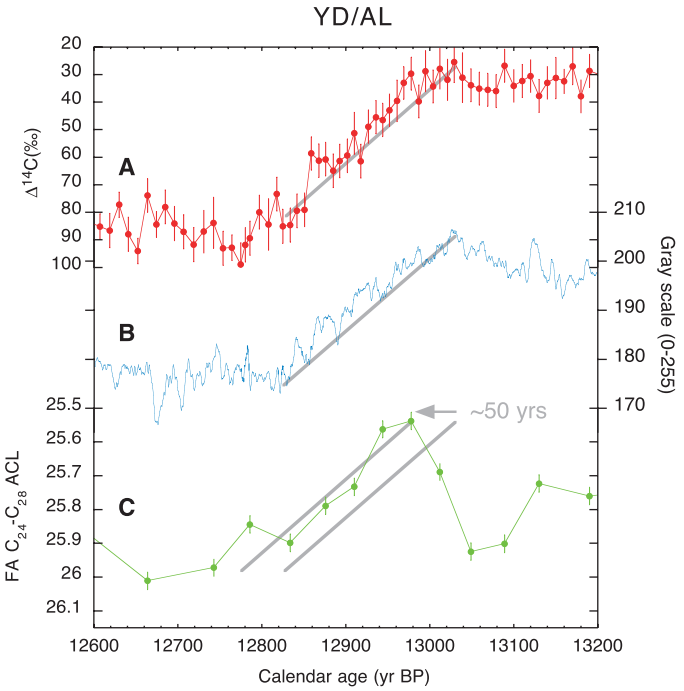


Fig. 3. Details of abrupt change at the beginning of the Younger Dryas interval in multiple proxies from Cariaco basin sediment core PL07-58PC. (**A**) Planktonic foraminiferal $\Delta^{14}C$, indicating a rapid decline in high-latitude NADW formation (*6*, *46*). (**B**) Percent reflectance (gray scale), recording a sharp increase in Cariaco upwelling and tropical trade-wind intensity (*2*). (**C**) Average chain length for C_{24}-C_{28} *n*-alkanoic acid leaf waxes, showing a shift in local Venezuelan vegetation from wetter C_3 forest to arid C_4 grassland. The timing of the rapid shift in each record is indicated by a gray line. Lagged cross-correlation analysis shows that high- and low-latitude climate shifts were synchronous within ± 15 years, but vegetation change in northern South America lagged climate by 50 ± 25 years (*44*). Error bars are 1σ.

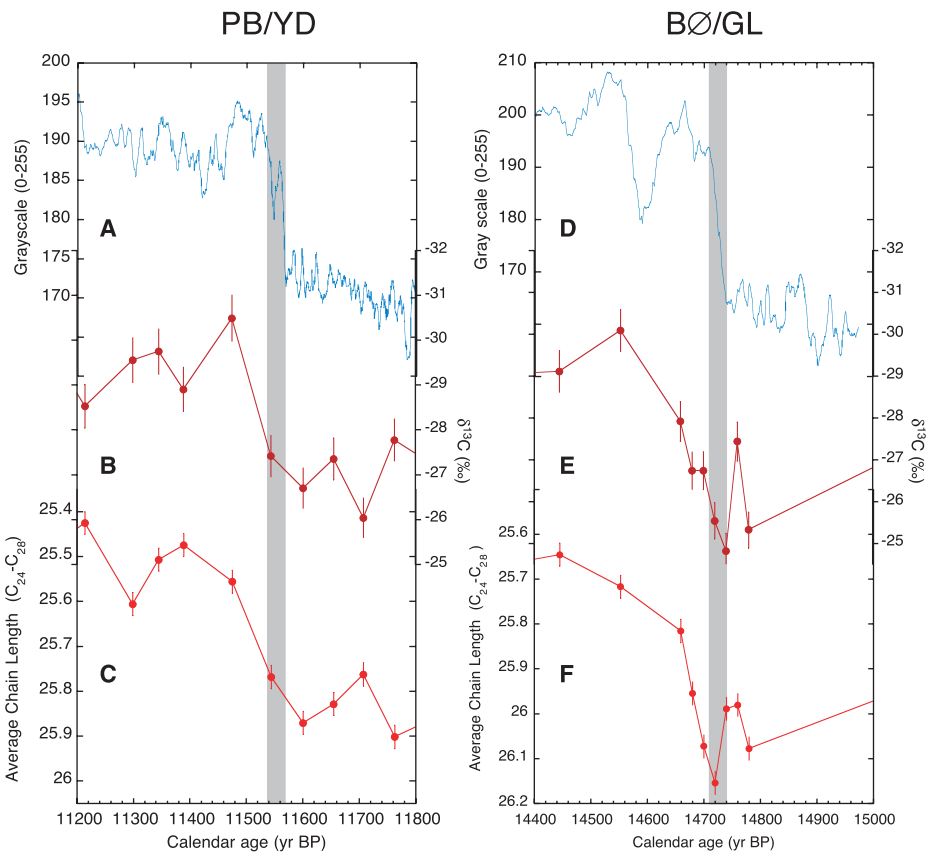


Fig. 4. Details of abrupt changes at the GL/BO (right) and YD/PB (left) transitions recorded in multiple proxies in Cariaco basin core PL07-58PC. (**A** and **D**) Percent reflectance (gray scale), recording abrupt declines in upwelling and trade-wind strength. (**B** and **E**) $\delta^{13}C$ for C_{32} *n*-alkanoic acid leaf waxes. (**C** and **F**) ACL index for C_{24}-C_{28} *n*-alkanoic acids. Both biomarker indices record abrupt shifts from arid C_4 grassland to wetter C_3 forest biomass at the GL and YD terminations. Lagged cross-correlation analysis shows that vegetation lagged local tropical climate changes by 25 ± 15 years for both events (*44*). Vertical gray bars indicate timing of rapid climate shifts based on gray scale. Error bars are 1σ.

Although the Cariaco basin varve chronology is anchored to newly revised tree-ring ages (*33*) by matching ^{14}C variations between ~10 and ~12.5 ka B.P. (*34*), independent (i.e., not ^{14}C) climate proxies can be compared within the interval of overlap. The end of the Younger Dryas event in tree rings is placed at 11,590 ± 0 years B.P. and in Cariaco at 11,580 ± 16 years B.P. (*33*, *34*). Taken together, these data and our new results suggest that climate shifts in both high-latitude and tropical North Atlantic regions were synchronous, and that tropical vegetation change lagged abrupt climatic shifts by several decades. This has profound implications for the interpretation of very rapid climate changes from vegetation records and may help reconcile the lag of atmospheric methane behind Greenland temperature change with Cariaco evidence for synchronous climate shifts at high and low latitudes.

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35. Comparing $\delta^{13}C$ for C_{16} versus C_{32} *n*-alkanoic acids shows no significant correlation (slope = 0.11 and correlation coefficient $r = 0.14$).
36. The ACL index used here is confined to the most abundant homologs and is defined as $\Sigma(iX_i)/\Sigma X_i$, where X = abundance and i = 24, 26, and 28 carbon atoms.
37. It is possible that the range of $\delta^{13}C$ values for arid C_4 versus humid C_3 plant populations lie closer together than the respective ACL distributions. In particular, plants using crassulacean acid metabolism (CAM) metabolism exhibit $\delta^{13}C$ values intermediate between those for C_3 and C_4 plants (*29*, *30*), and may have been present during the Younger Dryas transition. Although $\delta^{13}C$ distributions for modern C_3 and C_4 plants are well known, comparable measurements for ACL indices have not been made.
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41. African *n*-alkanoic acid leaf waxes during Glacial and Younger Dryas climates were almost certainly offset from the Holocene values shown in Fig. 2 for both $\delta^{13}C$ and ACL indices. Isotopic data for *n*-alkane leaf waxes off northwest Africa show a −4‰ shift from the Last Glacial Maximum to the Holocene (*48*). Any similar Glacial shift in *n*-alkanoic acid $\delta^{13}C$ would place the African data in Fig. 2 even farther from the Cariaco data and necessitate an unrealistically large change in ACL to place the African data as an end member on the Cariaco mixing line. Similarly, if African source regions changed, the new region would require a drier ACL, but wetter or unchanged $\delta^{13}C$ signal, inconsistent with the $\delta^{13}C$-ACL relationships seen in African leaf waxes (*27*). The simplest explanation, supported by local pollen data (*13*, *14*), is that Cariaco leaf wax data represent changes in local vegetation during deglacial climate changes.
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43. The biomarker records were used at their original 20- to 40-year sampling resolution for cross-correlation with gray scale and $\Delta^{14}C$. High-resolution gray-scale data were averaged and $\Delta^{14}C$ records were interpolated and resampled to provide even annual resolution. For comparison with gray-scale data, $\Delta^{14}C$ was used at its original 10- to 15-year resolution.
44. The lag calculation is dependent on cross-correlations using all data points within the interval of overlap, rather than the timing of the initial change recorded between data points (*42*). Therefore, although biomarker sampling resolution is ~40 years during the Younger Dryas onset, the transition itself lasted ~200 years, encompassing five or six data points, and the cross-correlation can be evaluated uniquely for each annual time step. Similarly, the biomarker sample resolution during the Glacial/Bølling transition is ~20 years, providing two or three data points within the duration of the shift. However, the biomarker resolution across the Younger Dryas termination is low (~40 years) relative to the climate shift, and the transition appears to occur between data points; thus, the calculated vegetation lag of 25 ± 15 years must be considered a minimum estimate. The robustness of these analyses relies on the assumption that the lower-resolution proxy data are not systematically offset above or below the true trend (*32*).
45. It is possible that some leaf waxes may be transported to the Cariaco basin by rivers, and thus the apparent lag could in part reflect the residence time of leaf waxes during riverine transport. However, leaf waxes are predominantly ablated and transported via eolian processes (*22*, *26*–*28*), and thus if the vegetation change were synchronous with local climate, there should be an immediate response in leaf waxes carried by eolian transport following an abrupt climate shift, with a slower rise due to fluvial input of leaf waxes following the climate change. However, the timing of rapid change in leaf waxes distinctly lags the rapid changes in climate, especially during the Younger Dryas onset (Fig. 3), in agreement with an overall lag of ~50 years.
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Materials and Methods
Figs. S1 and S2
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Antarctic Timing of Surface Water Changes off Chile and Patagonian Ice Sheet Response

Frank Lamy,[1,2]* Jérôme Kaiser,[2] Ulysses Ninnemann,[3] Dierk Hebbeln,[2] Helge W. Arz,[1] Joseph Stoner[4]

Marine sediments from the Chilean continental margin are used to infer millennial-scale changes in southeast Pacific surface ocean water properties and Patagonian ice sheet extent since the last glacial period. Our data show a clear "Antarctic" timing of sea surface temperature changes, which appear systematically linked to meridional displacements in sea ice, westerly winds, and the circumpolar current system. Proxy data for ice sheet changes show a similar pattern as oceanographic variations offshore, but reveal a variable glacier-response time of up to ~1000 years, which may explain some of the current discrepancies among terrestrial records in southern South America.

Resolving the origin and consequences of the millennial-scale climate changes evident during the last glacial period, and particularly during the last deglaciation, is a cardinal challenge of current paleoclimate research. A comprehensive understanding of such climate changes is invaluable for validating and improving prognostic climate models. In the North Atlantic region, large and abrupt temperature changes are clearly documented at intervals of ~5 to 10 and ~1 to 4.5 thousand years (named the Heinrich and Dansgaard-Oeschger events, respectively). The origins of these events are commonly construed to be linked to reorganizations in Atlantic thermohaline circulation (THC) (*1*). More recently, abrupt millennial-scale climatic oscillations have been recognized in many places around the globe. Their source is also postulated to be variations in the North Atlantic ocean-climate system propagating by means of global oceanic and atmospheric feedbacks (*2*).

Recent modeling studies, however, indicate that the tropics and the southern high latitudes may also play an active role in initiating rapid climate change (*3*, *4*). For example, it has been shown that rapid climate oscillations may originate from the tropical Pacific, potentially involving a long-term modulation of interannual-to-decadal climate changes of the eastern trop-

[1]GeoForschungsZentrum-Potsdam, Telegrafenberg, 14473 Potsdam, Germany. [2]Deutsche Forschungsgemeinschaft (DFG) Research Center Ocean Margins, Klagenfurter Strasse, 28359 Bremen, Germany. [3]Department of Earth Sciences and Bjerknes Centre for Climate Research, University of Bergen, Allégaten 41, 5007 Bergen, Norway. [4]Institute of Arctic and Alpine Research, University of Colorado, Boulder CO, 80309–0450, USA.

*To whom correspondence should be addressed. E-mail: flamy@gfz-potsdam.de

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