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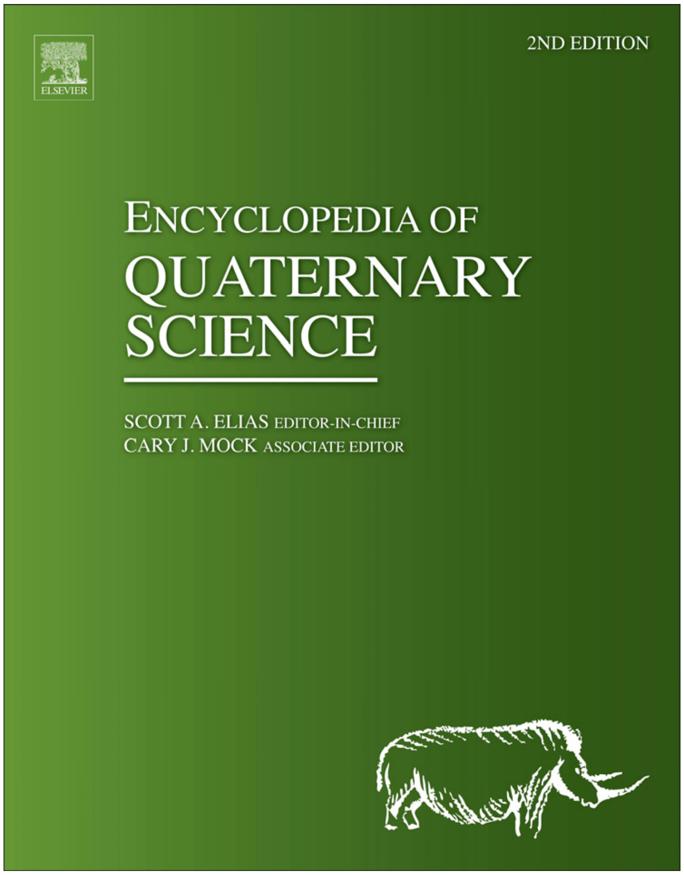

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The Younger Dryas Climate Event

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Introduction

The Younger Dryas has long been viewed as the canonical abrupt climate change event. The name is derived from the late 1800s Swedish and Danish pollen records that indicated a return of the cold-tolerant plant *Dryas octopetala* following a warm interval (the Ållerød warm period) (Andersson, 1896; Hartz and Milthers, 1901). The advent of radiocarbon dating techniques constrained this European cooling event to ~11–10 ka (ka: 1000 years ago) in radiocarbon years (Mangerud et al., 1974) or ~12.9–11.7 ka in calendar years (Rasmussen et al., 2006). At this time, boreal summer insolation (the driving force behind deglaciation) neared its peak (**Figure 1(a)**) and the sea level was approximately halfway through its deglacial rise of 120–130 m (**Figure 1(e)**). Subsequent research identified climate changes correlative to the European Younger Dryas in the North Atlantic and North Pacific Oceans, across Asia and North America, and in the tropics (e.g., Clark et al., 2002; Shakun and Carlson, 2010). The footprint of Younger Dryas climate change suggests that it was caused by a reduction in Atlantic meridional overturning circulation (AMOC) (Stouffer et al., 2006), for which there has been growing evidence since the mid-1980s (Boyle and Keigwin, 1987). This article reviews the Younger Dryas climate event, starting with the geographic pattern of climate change, followed by a discussion of ocean circulation changes, and ending with its ultimate forcing.

Extent of the Younger Dryas

The Northern Hemisphere Younger Dryas

Greenland ice cores

In the Northern Hemisphere, Greenland ice cores provide the best expression of the Younger Dryas as a cold event. Ice core δ^{18}O (δ signifies a change in isotopic ratio relative to a standard multiplied by 1000) consistently decreases by ~3‰ with an ~3.5‰ increase at the end of the event, reflecting up to 9 °C of cooling and 11 °C of warming (**Figure 1(b)**) (Alley, 2000). Isotopic ratios of nitrogen and argon gases from the GISP2 (Greenland Ice Sheet Project 2) ice core suggest 10 ± 4 °C of warming at the end of the Younger Dryas (Grachev and Severinghaus, 2005). The δ^{18}O records could also have an imprint of precipitation source change because Greenland Ice Sheet accumulation decreased by ~40% (0.11–0.07 m year^{-1}) (**Figure 1(c)**) and dust records suggest atmospheric reorganization in only several decades (Alley, 2000). Greenland deuterium excess, a proxy of precipitation source, shifted in 1–3 years at the start of the Younger Dryas, suggesting that 2–4 °C of source cooling was imprinted on the isotopic records (Steffensen et al., 2008). The trapped nitrogen isotopes also indicate that high-altitude Greenland was ~15 °C colder than present, but this estimate should not be extended to the ice margins as it may reflect a strengthening of the near-surface atmospheric inversion that is common over ice sheets (Severinghaus et al., 1998). Indeed, east Greenland valley glacier records suggest that summers were only 6–7 °C cooler than present (Kelly et al., 2008). Greenland runoff to the ocean was also enhanced (**Figure 1(d)**) (Carlson et al., 2008b) with mild terrestrial and marine summer conditions (Björck et al., 2002; Williams, 1993) during the Younger Dryas.

Europe

European pollen, chironomid, lake δ^{18}O, and speleothem records indicate that a cooler (2–6 °C) and/or drier climate characterized the Younger Dryas, which potentially changed in decades (**Figure 2(a)** and **2(b)**); e.g., Brauer et al., 1999; Genty et al., 2006; Heiri et al., 2007; von Grafenstein et al., 1999). Mountain glaciers readvanced in the Swiss Alps implying 3–4 °C of cooling (e.g., Ivy-Ochs et al., 2009). The Scandinavian Ice Sheet also readvanced, depositing an extensive moraine system across southern Norway, Sweden, and Finland (Andersen et al., 1995). Records from glacially influenced lakes in western Norway show an increase in glacier activity (Nesje, 2009), and the Iceland Ice Cap halted retreat during the Younger Dryas (Licciardi et al., 2007).

North Atlantic records

In the eastern North Atlantic, Iberian Margin and Mediterranean sea surface temperature (SST) records show 1–3 °C of cooling (**Figure 2(d)**) (e.g., Bard et al., 2000; Cacho et al., 2001). Northward up to the Norwegian Sea, SST records indicate 1–7 °C of cooling (**Figure 2(c)**) (e.g., Benway et al., 2010; Chapman and Maslin, 1999; Dolven et al., 2002; Ebbesen and Hald, 2004; Karpuz and Jansen, 1992). Westward, SSTs cooled by 1–2 °C in the subtropical gyre (Carlson et al., 2008a; Chapman and Maslin, 1999) and by 5–10 °C near maritime Canada (de Vernal et al., 1996; Keigwin and Jones, 1995). In contrast, SST increased by ~1.5 °C near southeastern United States (**Figure 2(e)**), which was related to the slowing of AMOC and decreased northward heat transport (Carlson et al., 2008a). The Gulf of Mexico also cooled slightly near the end of the Younger Dryas (**Figure 2(f)**) (Flower et al., 2004).

North America

Pollen, lake δ^{18}O, and soil δ^{13}C records from maritime Canada to north-central United States suggest a cooler and drier Younger Dryas (e.g., Dorale et al., 2010; Shuman et al., 2005; Yu and Eicher, 1998). In the southeastern United States, however, the Younger Dryas is characterized by a warmer and/or wetter climate, reflecting the trapping of heat in the western subtropical gyre due to reduced AMOC (Grimm et al., 2006). The Pacific Coast of the United States and Canada experienced 2–3 °C of Younger Dryas cooling (**Figure 3(a)** and **3(b)**) and a wetter climate (e.g., Barron et al., 2003; Kaufman et al., 2010; Kienast and McKay, 2001; MacDonald et al., 2008; Mathewes

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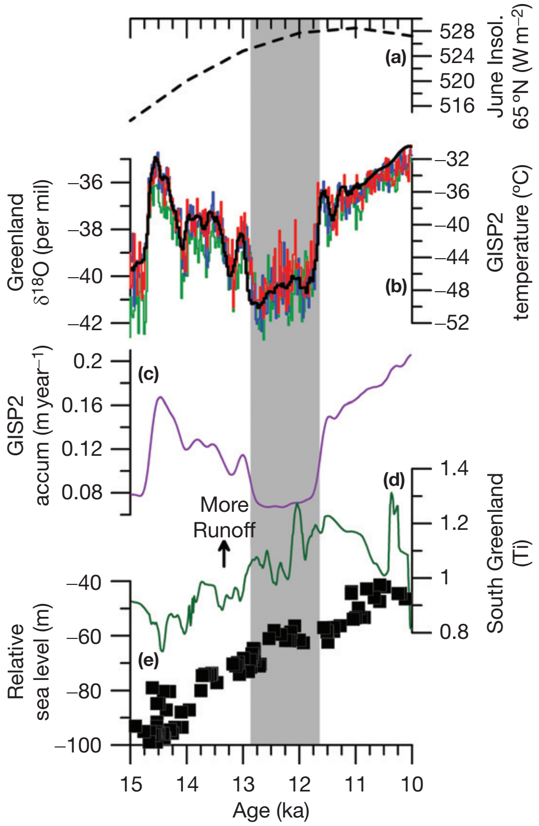

Figure 1 Greenland. (a) June insolation at 65°N (Berger and Loutre, 1991). (b) Greenland ice core $\delta^{18}O$ (red = GISP2 (Greenland Ice Sheet Project 2), blue = GRIP (Greenland Ice Core Project), green = NGRIP (North Greenland Ice Core Project); left) (Rasmussen et al., 2006) and GISP2 temperature (black; right) (Alley, 2000). (c) GISP2 ice core accumulation rate (Alley, 2000). (d) Ti concentration record off southern Greenland (Carlson et al., 2008b). (e) Relative sea-level data (Clark et al., 2009). Gray bar on this and other figures indicates timing of the Younger Dryas.

et al., 1993; Sea and Whitlock, 1995; Vacco et al., 2005). In the North American southwest, packrat midden $\delta^{13}C$ temperature reconstructions imply >3 °C of cooling (Cole and Arundel, 2005), and lake, speleothem, and water table records suggest an increase in net precipitation during the Younger Dryas (**Figure 3(c)** and **3(d)**) (Asmerom et al., 2010; Benson et al., 1997; Pigati et al., 2009; Polyack et al., 2004; Wagner et al., 2010). Valley glaciers readvanced or reoccupied cirques in many of the mountain ranges of the North American Cordillera (e.g., Davis et al., 2009). The southern margin of the Laurentide Ice Sheet also readvanced, depositing a moraine from western Lake Superior to southeastern Quebec (Lowell et al., 1999), with its northwest margin in Hudson Strait increasing iceberg discharge during the Younger Dryas (Andrews and Tedesco, 1992).

North Africa and the Middle East

In North Africa, climate cooled and net precipitation decreased during the Younger Dryas. A record of terrigenous input off the coast of northwest Africa indicates an increase in dust during the Younger Dryas, implying drying of the Sahara Desert (**Figure 4(a)**) (deMenocal et al., 2000), which is consistent

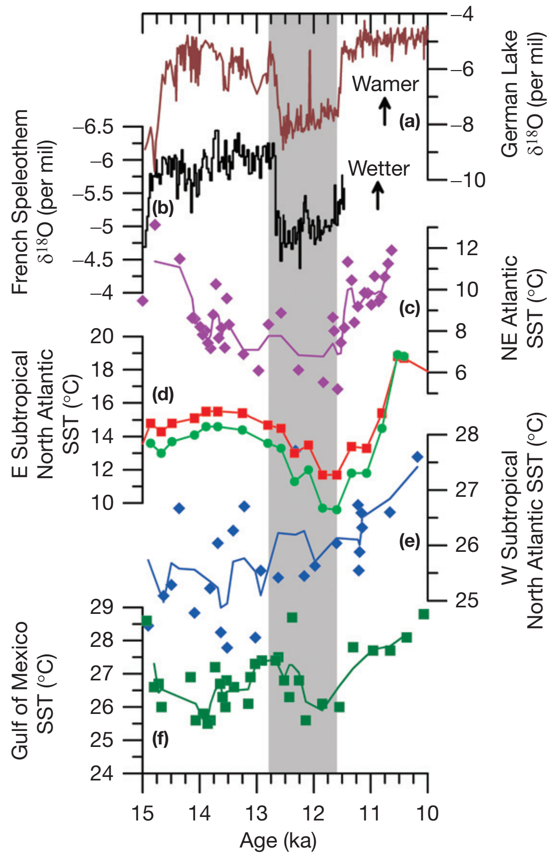

Figure 2 Europe and North Atlantic. (a) German lake $\delta^{18}O$ (von Grafenstein et al., 1999). (b) French speleothem $\delta^{18}O$ (Genty et al., 2006). (c) Feni Drift Mg/Ca-SST (Benway et al., 2010). (d) Iberian Margin alkenone-SST (two different alkenone–SST relationships shown) (Bard et al., 2000). (e) Blake Outer Ridge Mg/Ca-SST (Carlson et al., 2008a). (f) Gulf of Mexico Mg/Ca-SST (Flower et al., 2004). Symbols are individual measurements; line is 3-point running average.

with speleothems from North Africa and the Middle East (**Figure 4(b)**) (e.g., Genty et al., 2006; Shakun et al., 2007). This precipitation decrease is attributable to the southward shift in the Intertropical Convergence Zone (ITCZ) in response to reduced AMOC and North Atlantic cooling (Stouffer et al., 2006). Reduced upwelling in the Arabian Sea because of decreased winds is also consistent with southward ITCZ migration (Altabet et al., 2002; Schulz et al., 1998). North African SSTs cooled 0.5–1 °C (deMenocal et al., 2000; Zhao et al., 1995), but Arabian SSTs were relatively constant during the Younger Dryas (Saher et al., 2007).

Asia

Large portions of central to northern Asia currently lack paleoclimate records spanning the Younger Dryas interval. Nevertheless, a decrease in $\delta^{18}O$ from Lake Baikal in southern Russia implies shifts in moisture sources to the lake and potential cooling during the Younger Dryas (Morley et al., 2005). Further east, a pollen record from central Japan suggests ≤3 °C of cooling during the Younger Dryas (Nakagawa et al., 2003), whereas SST records from the China Sea indicate 0.5–1 °C of cooling (Kubota et al., 2010; Sun et al., 2005). Multiple

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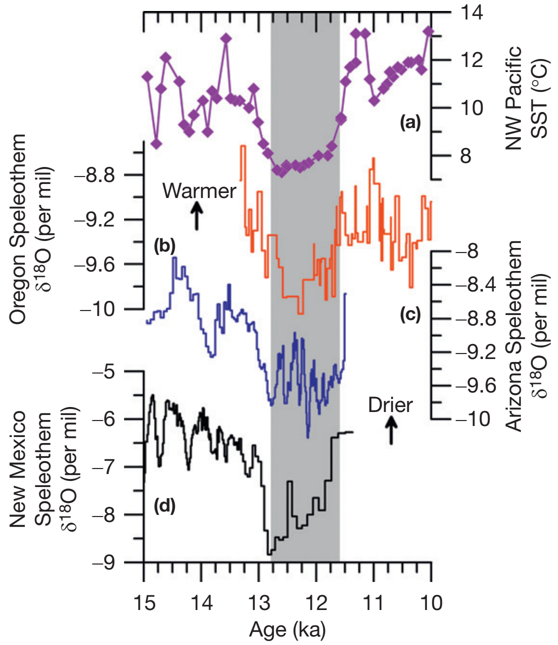

Figure 3 Western North America. (a) Northwest Pacific alkenone-SST (Barron et al., 2003). (b) Oregon speleothem $\delta^{18}O$ (Vacco et al., 2005). (c) Arizona speleothem $\delta^{18}O$ (Wagner et al., 2010). (d) New Mexico speleothem $\delta^{18}O$ (Asmerom et al., 2010).

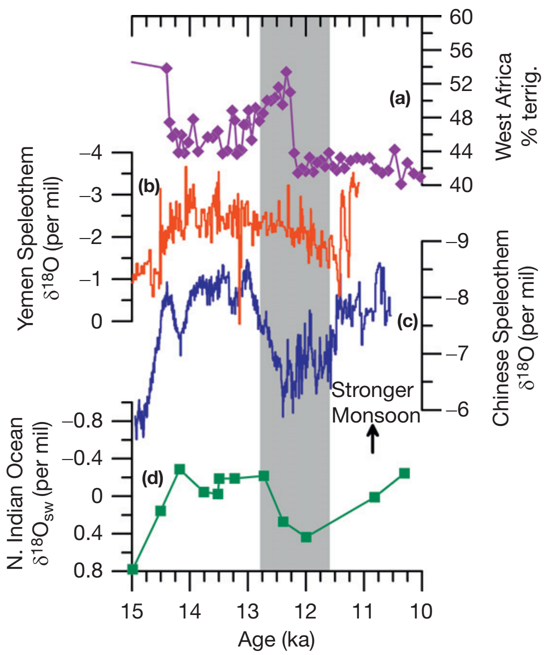

Figure 4 North Africa and Asia. (a) West Africa dustiness (deMenocal et al., 2000). (b) Yemen speleothem $\delta^{18}O$ (Shakun et al., 2007). (c) Chinese speleothem $\delta^{18}O$ (Wang et al., 2001). (d) North Indian Ocean $\delta^{18}O$ of seawater (sw) (Rashid et al., 2007).

speleothem $\delta^{18}O$ records from China and India and Indian Ocean marine salinity proxies imply a weakening of the summer Indo-Asian monsoon (**Figure 4(c)** and **4(d)**) (Dykoski et al., 2005; Rashid et al., 2007; Sinha et al., 2005; Wang et al., 2001), with a south China lake record suggesting an increase in winter Asian monsoon strength during the Younger Dryas (Yancheva et al., 2007). Overall, these paleomonsoon proxies are consistent with a southward shift in the ITCZ.

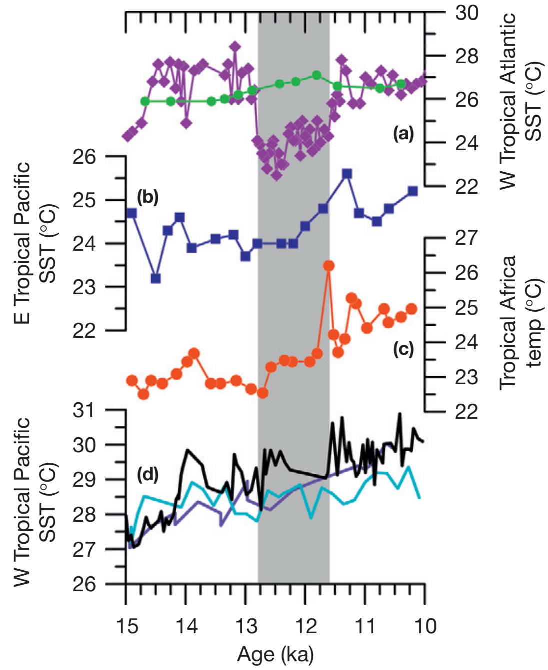

Figure 5 Tropics. (a) Mg/Ca-SST from Cariaco Basin (purple) (Lea et al., 2003) and alkenone-SST from Caribbean Sea (green) (Rühleman et al., 1999). (b) Mg/Ca-SST from eastern tropical Pacific (Lea et al., 2006). (c) Lake Tanganyika, Africa temperature (Tierney et al., 2008). (d) Mg/Ca-SST from western tropical Pacific (three records) (Stott et al., 2007).

Tropical Climate and the Younger Dryas

East Atlantic and Central America

The temperature pattern observed in the tropics during the Younger Dryas is complex (Shakun and Carlson, 2010). SSTs show a warming of 0.25–1.2 °C in the Caribbean Sea and off northeast Brazil (Jaeschke et al., 2007; Rühleman et al., 1999; Schmidt et al., 2004; Weldeab et al., 2006), but a cooling of ~3.3 °C north of Venezuela (**Figure 5(a)**) (Lea et al., 2003), possibly from a southward shift in the ITCZ that increased trade wind strength and upwelling (Hughen et al., 1996). Indeed, precipitation proxies suggest that precipitation decreased north of the equator and increased south of the equator (Haug et al., 2001; Jaeschke et al., 2007; Wang et al., 2007). Ice cores from Peru and Bolivia show an ~2‰ decrease in $\delta^{18}O$ (Thompson et al., 1998), concurrent with higher lake levels in nearby Lake Titicaca and consistent with southward ITCZ migration (Baker et al., 2001). Despite wetter conditions, a warmer climate would explain glacier recession in Peru during the Younger Dryas (Rodbell and Seltzer, 2000).

West Atlantic and Africa

Off tropical West Africa, SST records indicate ~0.2 °C of warming during the Younger Dryas (Shefuβ et al., 2005; Weldeab

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et al., 2007) in agreement with air temperature proxies that suggest ~0.8 °C of warming (Weijers et al., 2007). Another Younger Dryas SST record from southwestern tropical Africa, however, suggests a slight cooling from increased winds and an enhanced local upwelling of colder waters (Kim et al., 2002), which may extend to ~25°S (Farmer et al., 2005). Eastern tropical Africa temperature records indicate a complex pattern with warming at Lake Tanganyika (**Figure 5(c)**) and potential cooling at Lake Malawi and surrounding regions (Gasse et al., 2008; Tierney et al., 2008; Weldeab et al., 2007). Off the southeast coast of Africa near Madagascar, SSTs warmed during the Younger Dryas (Levi et al., 2007). Interestingly, despite evidence for a southward shift in the ITCZ in tropical Africa (e.g., Gasse et al., 2008; Johnson et al., 2002; Tierney and Russell, 2007), leaf wax $\delta^{13}C$ from Lakes Malawi and Tanganyika indicate arid conditions, highlighting a complex tropical African climate response to reduced AMOC (Castaneda et al., 2007; Tierney et al., 2008).

Tropical Pacific

In the eastern tropical Pacific, most SST records indicate little temperature change or slight warming during the Younger Dryas (**Figure 5(b)**) (e.g., Benway et al., 2006; Lea et al., 2006), with one record showing ~0.4 °C of cooling (Kienast et al., 2006). Paleosalinity reconstructions suggest a net decrease in moisture transport from the Atlantic to the Pacific because of a southward displacement of the ITCZ (Benway et al., 2006). Further west, the Mauna Kea Ice Cap of the Hawaiian Islands may have readvanced during the Younger Dryas from a southward shift in the ITCZ, with enhanced extratropical storms and northward local winds that transported more moisture to Hawaii (Anslow et al., 2010). In the tropical west Pacific, SSTs generally warmed but with regions of cooling potentially reflecting proxy biases (**Figure 5(d)**) (Kienast et al., 2001; Levi et al., 2007; Linsley et al., 2010; Rosenthal et al., 2003; Stott et al., 2007).

Southern Hemisphere Climate During the Younger Dryas

During the Younger Dryas, the extratropical Southern Hemisphere generally warmed (Shakun and Carlson, 2010). Antarctic ice cores record increasing $\delta^{18}O$ and δD (**Figure 6(e)** and **6(f)**) (e.g., Blunier and Brook, 2001; EPICA Community Members, 2006; Jouzel et al., 2007). SST records show a warming of 0.3–1.9 °C from the southeast Atlantic to New Zealand (**Figure 6(a)–6(c)**) (Barker et al., 2009; Barrows et al., 2007; Carlson et al., 2008b; Lamy et al., 2004; Pahnke and Sachs, 2006). Speleothem and pollen records from New Zealand and pollen records from South America confirm that the Younger Dryas was generally a period of warming (e.g., Hajdas et al., 2003; Moreno et al., 2009; Newnham and Lowe, 2000; Turney et al., 2003; Williams et al., 2005), consistent with New Zealand and Patagonian glacier retreat (Kaplan et al., 2008; 2010; Moreno et al., 2009; Putnam et al., 2010).

Geographic Response Summary

Globally, the Younger Dryas was a period of climate change. Plotting Younger Dryas temperature and climate anomalies against latitude shows that climate anomalies increased in magnitude toward the poles with opposite signs in the

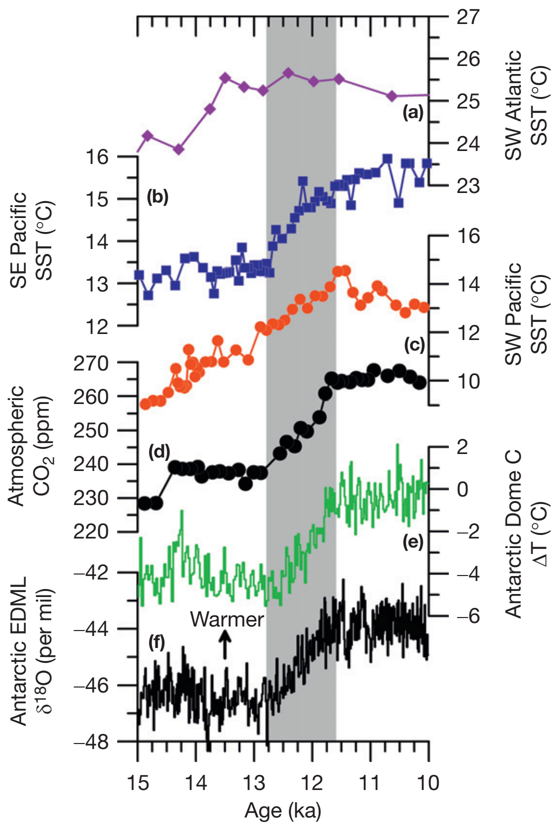

Figure 6 Southern Hemisphere. (a) Mg/Ca-SST near southeast Brazil (Carlson et al., 2008a). (b) Alkenone-SST near southern Chile (Lamy et al., 2004). (c) Alkenone-SST near New Zealand (Pahnke and Sachs, 2006). (d) Atmospheric CO_2 (Monnin et al., 2001). (e) Antarctic Dome C change in temperature (Jouzel et al., 2007). (f) Antarctic EDML ice core $\delta^{18}O$ (EPICA Community Members, 2006).

Northern and Southern Hemispheres (**Figure 7**) (Shakun and Carlson, 2010), reflecting the bipolar seesaw response (Blunier and Brook, 2001). Nevertheless, greater cooling at high northern latitudes than warming at high southern latitudes results in a net global cooling of ~0.6 °C likely caused by more extensive snow and sea-ice cover increasing Northern Hemisphere albedo during the Younger Dryas (Shakun and Carlson, 2010).

Ocean Circulation During the Younger Dryas

The first evidence for a change in AMOC during the Younger Dryas came from North Atlantic proxies of water mass nutrients (benthic $\delta^{13}C$ and Cd/Ca) that showed that Southern Ocean Deep-water volume increased in the North Atlantic at the expense of North Atlantic Deep-water (**Figure 8(b)** and **8(c)**) (Boyle and Keigwin, 1987). More recently, a proxy of water export ($^{231}Pa/^{230}Th$ in marine sediments) from the North Atlantic suggests a ~30% reduction in deep AMOC strength during the Younger Dryas (**Figure 8(a)**) (McManus et al., 2004), consistent with a reduction in the North Atlantic bottom water currents (**Figure 8(d)**) (Praetorius et al., 2008). Other tracers of water mass source and age confirm increased Southern Ocean Deep-water and Antarctic Intermediate-water and reduced North Atlantic Deep-water volume

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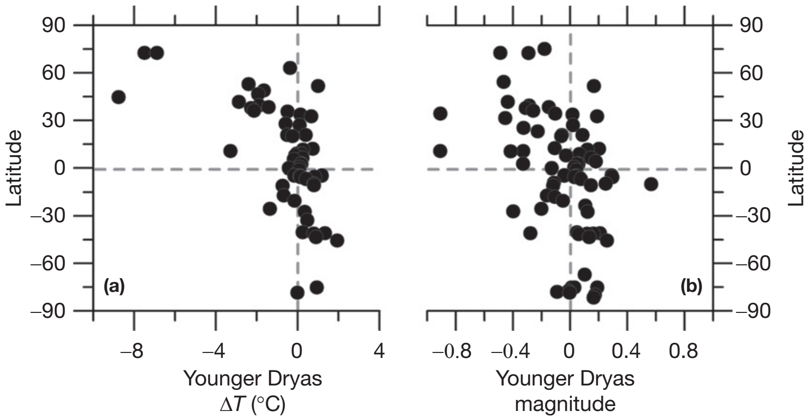

Figure 7 Latitudinal plots of the change in temperature between the Younger Dryas and Allerød (a) and the magnitude of Younger Dryas climate change relative to the glacial–interglacial change in a proxy record (b). Data from Shakun and Carlson (2010).

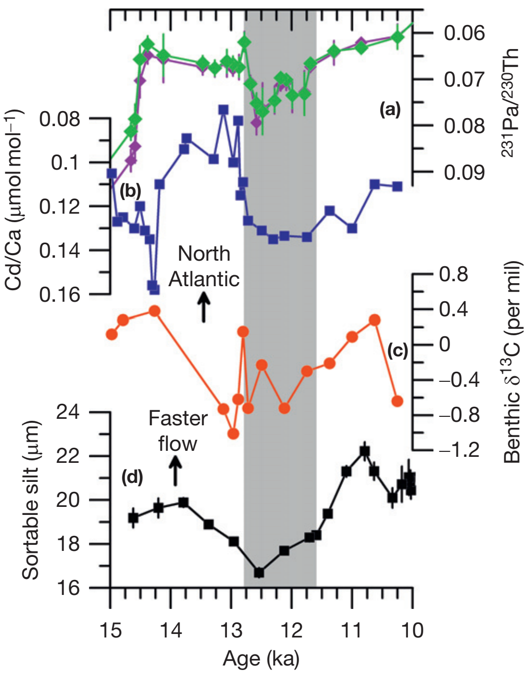

Figure 8 North Atlantic overturning circulation. (a) $^{231}Pa/^{230}Th$ for the North Atlantic with up indicating more water export past Bermuda Rise (McManus et al., 2004). (b) Benthic Cd/Ca from Bermuda Rise (Boyle and Keigwin, 1987). (c) Benthic $\delta^{13}C$ from Bermuda Rise (Boyle and Keigwin, 1987). (d) Sortable silt from the Iceland-Scotland Overflow (Praetorius et al., 2008). Gray bar indicates timing of Younger Dryas.

(Pahnke et al., 2008; Roberts et al., 2010; Robinson et al., 2005). The rise in atmospheric CO_2 during the Younger Dryas (Figure 6(d)) implies increased upwelling and degassing of carbon-rich water, evidence for which exists in the Southern Ocean and along Baja California (Anderson et al., 2009; Marchitto et al., 2007). The reduction in AMOC strength during the Younger Dryas provides the causal mechanism for Northern Hemisphere cooling and Southern Hemisphere warming, as well as the southward displacement of the ITCZ. Coupled atmosphere-ocean climate models consistently simulate a reduction in AMOC and these attendant climate impacts in response to increased freshwater discharge to the North Atlantic (Liu et al., 2009; Meissner and Clark, 2006; Otto-Bliesner and Brady, 2009; Peltier et al., 2006; Stouffer et al., 2006).

The Cause of the Younger Dryas

The forcing of the Younger Dryas was originally inferred to be northward retreat of the southern Laurentide Ice Sheet margin out of the Great Lakes, thereby routing western Canadian Plains freshwater from the Mississippi River to the St. Lawrence River, with the attendant increase in freshwater discharge to the North Atlantic slowing AMOC (Johnson and McClure, 1976; Rooth, 1982). This hypothesis has more recently been questioned based on marine, terrestrial, and ice-sheet model results.

In the St. Lawrence Estuary, planktic $\delta^{18}O$ records show a 0.5–0.8‰ decrease at the start of the Younger Dryas (de Vernal et al., 1996; Keigwin and Jones, 1995; Keigwin et al., 2005) when $\delta^{18}O$ increases in the Gulf of Mexico (Figure 9(b)) (e.g., Flower et al., 2004). Concurrent Younger Dryas cooling of 5–10 °C in the St. Lawrence Estuary (de Vernal et al., 1996; Keigwin and Jones, 1995) may have masked a much larger $\delta^{18}O$ signal associated with eastward freshwater routing (1.8–2.7‰) (Carlson et al., 2007). Four independent geochemical freshwater source and amount proxies and mollusk $\delta^{18}O$ from the St. Lawrence River and Estuary support the eastward routing of western Canadian Plains freshwater at the start of the Younger Dryas (Figure 9(c) and 9(d)) (Brand and McCarthy, 2005; Carlson et al., 2007; Cronin et al., 2008). Modeling of the geochemical records indicates a base-flow freshwater discharge increase of ~0.12 Sv (Sv = Sverdrup, 10^6 m^3 s^{-1}) (Carlson et al., 2007), which is sufficient to slow AMOC (Liu et al., 2009; Meissner and Clark, 2006; Otto-Bliesner and Brady, 2009; Peltier et al., 2006; Stouffer et al., 2006). Although dynoflagellate cyst reconstructed sea surface salinity does not decrease, which led de Vernal et al. (1996) to argue that St. Lawrence River freshwater discharge did not

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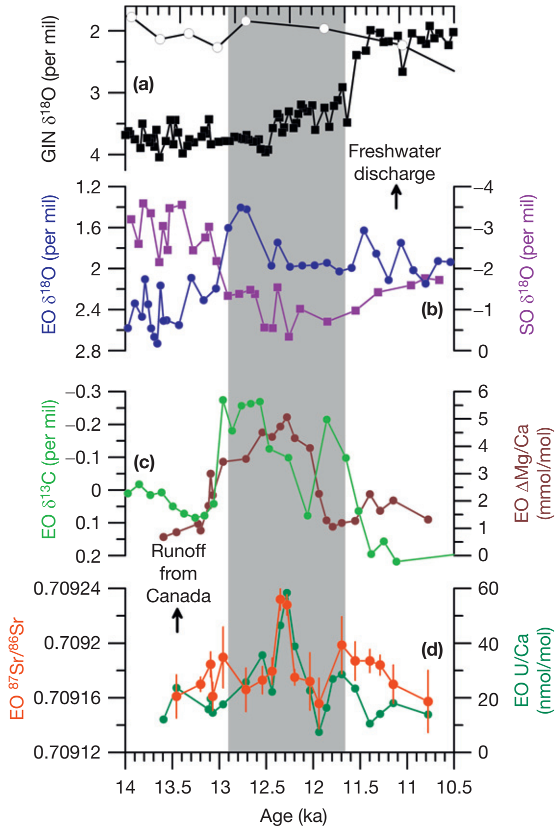

Figure 9 North American runoff records. (a) Two planktic $\delta^{18}O$ records from the Greenland-Iceland-Norwegian (GIN) Sea (Fram Strait, open symbols; Norwegian Sea, diamonds) (Dokken and Jansen, 1999; Nørgaard-Pedersen et al., 2003). (b) Planktic $\delta^{18}O$ records from the Gulf of Mexico (Southern Outlet: SO) (purple; Flower et al., 2004) and St. Lawrence Estuary (Eastern Outlet: EO) (blue; Keigwin et al., 2005). (c) Planktic $\delta^{13}C$ (light green; de Vernal et al., 1996) and $\Delta Mg/Ca$ (brown; Carlson et al., 2007). (d) Planktic $^{87}Sr/^{86}Sr$ (red) and U/Ca (dark green) from the St. Lawrence Estuary (Carlson et al., 2007).

increase during the Younger Dryas, subsequent research has shown that dynoflagellate cysts are insensitive to salinity changes above 12 practical salinity units (Telford, 2006). Similarly, ^{14}C dates from the eastern outlet of the western Canadian Plains have been interpreted as implying that the eastern outlet was not open until after the start of the Younger Dryas (Lowell et al., 2009). The dates, however, are only minimum limiting and indicate that the eastern outlet was open prior to ~12.6 ka, in agreement with the Mississippi and St. Lawrence runoff records (Carlson et al., 2009).

Arguments for a northern routing of western Canadian Plains freshwater and an abrupt discharge of Laurentide meltwater to the north have also been proposed as the forcing of the Younger Dryas. Murton et al. (2010) used optically stimulated luminescence (OSL) ages from sand layers bracketing an erosional surface on the Mackenzie River Delta to argue that the erosional event (e.g., flood) occurred shortly after ~13 ka based on the average of the stratigraphically lower dates. However, the OSL ages overlying the erosional surface of ~11.9 ka would be a closer age constraint on the erosion event, because erosion removed some unknown amount of sand and the overlying ages are conformable with the end of the erosion event as flow waned, depositing the sand. Therefore, the erosion event is not related to the onset of the Younger Dryas. Tarasov and Peltier (2005) suggested that a 0.09 Sv discharge of Laurentide meltwater into the Arctic Ocean for 0.3 ka could have forced the Younger Dryas. Climate models do not, however, produce a >1 ka long reduction in AMOC from a 0.3 ka discharge of freshwater (Liu et al., 2009; Meissner and Clark, 2006; Otto-Bliesner and Brady, 2009; Peltier et al., 2006; Stouffer et al., 2006). Although such a melting event could be longer, sea-level records constrain the global meltwater discharge to <5 m of sea-level rise (**Figure 1(f)**) (Bard et al., 2010) or ~0.05 Sv over 1.2 ka. Furthermore, there is no existing paleoceanographic evidence for an increase in freshwater discharge to the Arctic Ocean at the start of the Younger Dryas (**Figure 9(a)**) (Carlson and Clark, 2008).

Another recent hypothesis for the source of the freshwater that forced the Younger Dryas is a comet that impacted on or near the Laurentide Ice Sheet (Firestone et al., 2007), but evidence for Younger Dryas-like events during earlier deglaciations precludes a unique comet forcing (Carlson, 2008). These earlier events occur at approximately the mid-point of the deglacial sea-level rise, similar to the sea level during the Younger Dryas, suggesting that the AMOC reduction is related to ice-sheet volume/extent (Carlson, 2008). Only continental routing from the Laurentide Ice Sheet retreat pertains to ice-sheet extent/size and the existing records all suggest that this freshwater routing was from the southern outlet to the eastern outlet, consistent with the originally hypothesized forcing of the Younger Dryas.

See also: **Glacial Climates:** Thermohaline Circulation. **Ice Core Records:** Correlations Between Greenland and Antarctica. **Paleoceanography:** Paleoceanography An Overview. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events; Younger Dryas Oscillation, Global Evidence.

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