The Younger Dryas Event: A Review of Observations and Mechanisms

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Contents:

Abstract
The Younger Dryas event (Y-D event) is a widely studied, if not the most well-studied, example of abrupt climate change. Since its original identification, scientists have made continued efforts in identifying changes in the past climate and attributing it to the sudden global climate reversal that is the Y-D event, which occurred from about 12,900 years to 11,600 years before present (BP). Through various paleoclimatic proxies utilized at various sites around the world, it will be shown that the Y-D event had a clear, global effect. These global observations will be presented on a location-by-location basis, starting with where the Y-D signal appears to be the strongest, in Greenland and Europe. Next, descriptions of the Y-D event in the Americas, Asia, the Pacific Ocean, and Africa will be given, followed by a review of the signal in Antarctica, where it is the most uncertain. Globally, the Y-D event can generally be seen in the form of a cooler and dryer Northern Hemisphere, a warmer Southern Hemisphere, and a displacement of the tropical rain belt which in turn weakens northern summer monsoons. After the summary of global observations, a review on proposed mechanisms will be conducted. First, this paper will examine how ocean circulation could affect such a drastic change. Next, it will discuss the widely accepted theory that a freshwater influx disrupted production of North Atlantic deep water which in turn slowed the overturning circulation, which resulted in the Y-D cold plunge in the Northern Hemisphere. After an examination of other potential mechanisms and triggers, a synthesis of these proposed theories will be done. It ultimately appears that ocean circulation is the main cause behind the Y-D event, though there is still a lot of opportunity for further research in identifying the trigger. Overall, the aim of this paper is to present a comprehensive understanding of what the Y-D climate was like globally, and how this paleoclimatic event was able to occur.

Introduction
The Y-D event is quite likely the most studied climate event on a millennial scale. Having been identified in the early 1900s, this cold snap in the Northern Hemisphere was a first indication that climate could change abruptly, within a matter of decades or years (Weart, 2003). This idea of abrupt climate change was strongly exemplified by the Y-D event, and when it was first noted, it was contrary to the then-current views on climate, which assumed gradual changes over the course of millennia (Weart, 2003). Since then, this idea of rapid climate change has been strongly embedded within paleoclimatic studies, and within the general culture. A quick search on Google Scholar for the "Younger Dryas" term yields over 40,000 results. Furthermore, as anthropogenic global warming becomes more and more visible to the general populace, the popular media has begun joining scientists in looking at the past climate. As a result, the Y-D event has been sensationalized as an ancient "Day After Tomorrow" scenario, conjuring the imagery of drastic disaster movies where a stoppage of ocean circulation instantly plunges the world into cold (Enos, 2020). This disaster movie brought so much attention to the idea of rapid climate change that NASA and the National Snow & Ice Data Center (NSIDC) held a Q&A clarifying the "science" behind the movie, including about the Y-D event (NSIDC, 2004). Evidently, people have noticed that understanding these past climate events could offer potential insights on what is to come in the future. Although the Y-D event was not as extreme as what Hollywood envisions abrupt climate change to be, it still strongly impacted the globe and understanding its driving mechanisms is beneficial for understanding the global climate system in general. Therefore, this paper will present a thorough review on the current knowledge about the history of the Y-D event, its global effects, and its driving mechanisms.

Initial Identification of the Younger Dryas

To start, this paper will give a brief history about how this cold period in the Northern Hemisphere became known as the Younger Dryas. Paleoclimatic indicators of the Y-D event were first identified in Scandinavia around the early 20th century. In Scandinavia, there exists a plant called *Dryas octopetala*. *Dryas octopetala* is an Arctic flower that grows in tundra conditions. Paleobotanists such as Alfred Nathorst noted that one could identify these plants in Scandinavian clay pits (Nathorst, 1870, as cited in Mangerud, 2020). He started identifying their leaves in freshwater clay in southern Sweden, noting that they were not abundant. Since these plants are less populous in milder conditions, paleobotanists can use them as a proxy in understanding past climate by counting their numbers in layers of clay. In 1901, Hartz and Milthers came across a clay bed containing small fossils of many plants that existed in milder conditions, such as birch trees and juniper. This bed, dubbed the "Allerød *gyttja*" (Allerød being the village they found the pit in, and *gyttja* being mud formed from peat decay) was sandwiched between layers of *gyttja* that contained plants that grew in colder conditions, including *Dryas octopetala* (Hartz and Milthers, 1901 as cited in Mangerud, 2020). A photographic example of these alternating gyttja can be seen in Figure 1. This transition was noted to be a climate oscillation, and in 1912, Hartz gave a talk where he described the two inferred cold period layers as Younger Dryas and Older Dryas Clay (Hartz, 1905 as cited in Mangerud, 2020). Thus, the first appearance of the Younger Dryas term was used. This observation would later be confirmed by scientists of many fields. In the 1930s, palynologists (those that studied pollen records) such as Knud Jessen in Scandinavia noted a warm period as well in their records, correlating to the Bølling-Allerød warm period, which is followed by the Y-D event (Jessen, 1935 as cited in Manten, 1966). These palynologists acknowledged the earlier work of Hartz and Milthers and kept with their Younger Dryas/Older Dryas terminology. From the work of these Scandinavian scientists, it can be seen that there is a significant cooling event that occurred in the past in the Nordic regions. This cooling event now known to be the Y-D event will later be shown to have global effects.

Dating Efforts

On the topic of establishing a chronology of the Younger Dryas, there have been many efforts to establish one through various methods. Most recently, the Y-D event in the North Atlantic is thought to have started at 12,870±30 years BP and terminating between 11,700±40 and 11,610±40 years BP (Cheng et al., 2020). In earlier years, when radiocarbon dating first became prevalent, scientists were able to estimate the boundaries for the Y-D event by carbon dating pollen samples. These first estimates determined that the Y-D event occurred from about 11,000 to 10,000 radiocarbon years BP (Mangerud et al., 1974). There have also been efforts to date the Y-D event using tree-rings in Germany. In 1993, Becker and Kromer demonstrated that chronologies dating back as far as 11,400 BP can be pulled from oak and pine tree samples recovered in alluvial deposits (Becker and Kromer, 1993). They concluded that the termination of the Y-D event occurred a couple of centuries after what was determined by the Greenland ice cores and posited that a lag in climate between Greenland and central Europe was what was observed (Becker and Kromer, 1993). On the topic of Greenland ice cores, very strong and visible Y-D event evidence can be observed in them, in the form of $\delta^{18}O$ variations. Since the late 1960s, scientists have noted large fluctuations in $\delta^{18}O$ in ice cores, which indicate large swings in temperature (Dansgaard et al., 1969). Dansgaard, after drilling new cores in 1981, concluded that these violent fluctuations were large climatic events in the North Atlantic

(Dansgaard et al., 1982). Quickly, these fluctuations were observed by various scientists and were dubbed to be Dansgaard-Oeschger oscillations. It was then suggested that a similar abrupt temperature swing is what terminated the Y-D cold period 10,700 years ago (Hammer et al., 1986 as cited in Dansgaard et al., 1989). From the $\delta^{18}O$ profiles, Dansgaard et al. also suggested that during the termination of the Y-D event, a 7ºC increase in temperature occurred over a period of only 50 years (Dansgaard et al., 1989). This indicated that the Y-D event terminated quite abruptly and as a result it is now considered one of the canonical examples of rapid climate change. The Greenland Ice-core Project (GRIP) occurred in the early 1990s and produced a core that was over 3,000 m long coinciding with a 250 kyr record (Dansgaard et al., 1993). This core, along with the NGIRP core drilled in 2003 (North Greenland Ice Core Project members, 2004), was used to construct a time scale for the period of 7,900 to 14,800 years BP, a period of time that includes the Y-D event (Rasmussen et al., 2006). This timeline was constructed by identifying annual layers within the cores using a method called Continuous Flow Analysis. For the Y-D portion of the core, they identified 1232 annual layers, with 78 layers being uncertain (Rasmussen et al., 2006). For NGRIP, the timescale was then set by an 8,200-year BP cold event associated with a volcanic eruption that has a maximum counting error of 47 years (Rasmussen et al., 2006). With this scaling, they estimate the onset of the Y-D event to have occurred 12,896 years BP (with a counting error of 138 years), while the Y-D-Preboreal transition occurred 11,703 years BP (with a counting error of 99 years) (Rasmussen et al., 2006). Other Greenland ice core projects, such as the Greenland Ice Sheet Project II (GISP2), share similar timescales in terms of the transition times (Meese et al., 1997). The paper by Meese et al. also provides a good illustration of the annual layers found in the cores (see Figure 2), where the summers (which have higher dust than the winters) appear as lighter bands due to the increase in dust scattering more light (Meese et al., 1997). A later study done by Cheng et al. examining the $\delta^{18}O$ profiles of speleothems (cave formations) around the world noted that the North Atlantic speleothem record from the Seso Cave in Spain was strongly correlated with that of the NGRIP profile and utilized those to provide the latest set of dates for the Y-D event chronology (presented at the start of this section) (Cheng et al., 2020). The speleothem samples from Cheng et al. are collected from locations scattered around the world, and they generally all carry some form of the Y-D signal (Cheng et al., 2020). This global nature of the Y-D event will be the focus of following section.

Global Observations

Soon, scientists found various trends that can be associated with the Y-D event around the world through various proxies. These next few sections will be dedicated to identifying these observations on a continental/regional basis. For each region, a summary will be given looking into what proxies were utilized, along with a summary of the local implications of the Y-D event. To start, Figures 3a-c shows a comparison between various proxy profiles around the world with the NGRIP $\delta^{18}O$ profile, from the Cheng et al. study in 2020 (Cheng et al., 2020). These include speleothem samples from Asia (Figure 3a), the tropical Pacific and Atlantic (Figure 3b), along with ice cores from Greenland and Antarctica (Figure 3c). These figures can be referred to throughout the location-by-location analysis. Some of these additional proxies in the figure will be discussed among others as an examination of the Y-D event around the globe is conducted. A focus will be made first in Greenland, where the Y-D signal seems strongest. Next, the Y-D signal will be examined in Europe and North America, two areas that were quite clearly impacted during the Y-D event. Afterwards, terrestrial proxy signals will be examined throughout the rest of the world to determine the global extent of the Y-D event.

Greenland

As mentioned earlier, the evidence of the Y-D event is strikingly clear in Greenland ice cores. On top of providing a timeline of events, these ice cores provide insight on snow accumulation rates through their layer thicknesses, temperatures through their $\delta^{18}O$ profiles, and atmospheric composition through other trapped gasses and aerosols (Alley, 2000). Examining the NGRIP $\delta^{18}O$ profile in any one of Figures 3a-c, we can see the highlighted drop and subsequent rise in $\delta^{18}O$ that marks the onset and the termination of the Y-D event. These records show that the Y-D event followed a gradual cooling before a cold plunge that occurred within a century (Broecker et al., 2010). Figure 4 includes the GISP2 core, of which a longer time window is shown so the gradual cooling followed by a cold plunge can be seen more clearly (Shakun and Carlson, 2010). This cooling dropped temperatures by up to 9 ºC (Alley, 2000). Denton et al. noted an even greater drop, with a mean annual temperature drop of up to 15 °C when compared with modern temperatures (Denton et al., 2005). Furthermore, there was a seasonality, where the coldest months dropped by almost 25 ºC when compared to the Bølling-Allerød warm period which occurred prior to the Y-D event (Denton et al., 2005). On the other hand, summer temperatures only dropped about 7 ºC (Denton et al., 2005). This seasonality can also be seen through anomalously warm summers noted by Björck et al in 2002 (Björk et al., 2002). This extreme seasonality is thought to be caused by an increase in sea-ice formation over the North Atlantic, which in turn stems from a shutdown of the ocean conveyor (Denton et al., 2005). This controlling mechanism will be discussed in further detail in a later portion of the paper. From interpreting the ice accumulation in the ice cores, it was also inferred that the amount of precipitation decreased by 0.11-0.07 m year^{-1}, or by about 40% (Alley, 2000). An interesting point to note is that although the Y-D event brought cooler temperatures and less precipitation, it did not have a significant impact on the retreating trend of the southern portion of the Greenland ice sheet, which has been retreating since before 14,100 years BP (Carlson et al., 2008b). Carlson et al. came to this conclusion by measuring the [Ti] and [Fe] marine sediment records. Ti and Fe come from continental sources and represent the sediment brought to the ocean by meltwater streams. They observed heightened [Ti] and [Fe] concentrations corresponding with warming periods, but these elevated concentrations remained constant during the Y-D event (Carlson et al., 2008b). This would imply warm and mild summers in southern Greenland throughout the Y-D event (Carlson et al., 2008b). On the other hand, based off dates associated with moraines on the north coast of Greenland, there was a readvancement of local glaciers in the northern Greenland ice margins (Larsen et al., 2016). Therefore, it can be seen that the Y-D event in Greenland is a return to cold conditions as demonstrated by the temperature drop and some glacial advances, but it does not entirely revert the pre-existing warming trend as shown by the steady retreat of the southern Greenland ice sheet.

Compared to the onset, the ending of the Y-D event was very sharp. Dansgaard et al. noted this with their 7ºC measured increase in temperature occurring over a period of only 50 years (Dansgaard et al., 1989). A later study examining deuterium excess as a proxy for Greenland precipitation moisture sources noted a mode switch occurring over the course of between 1 to 3 years (Steffensen et al., 2008). They point out that this switch initiated a gradual change of increasing temperature that lasted over 50 years, which correlates with the $\delta^{18}O$ fluctuation observed by Dansgaard et al. in 1989. This is still very abrupt when compared to the onset of cooling which lasted over 200 years (Steffensen et al., 2008). Alley took particular interest in noting this abruptness; through a synthesis of all the different markers, he points out many different things occurring with the end of the Y-D event. From the $\delta^{18}O$ profiles, it was concluded that a 5-10ºC warming occurred (in line with the Dansgaard et al. observation). Also noted is a doubling of snow accumulation in central Greenland and a reduced wind speed from the large drop in wind-blown materials found in the layers (Alley, 2000). Finally, Alley also notes that there is a large increase in methane in the cores at the end of the Y-D event, and says it is a result of the expansion of global wetlands (Alley, 2000). These aspects of the end of the Y-D event all occurred within decades, if not a couple of years (Alley, 2000). In short, with the end of the Y-D event, a return to interglacial conditions was observed through increased temperature and precipitation.

Europe

The initial field observations of the Y-D event in Europe were made in Scandinavia. The event was associated with a drastic cold period during the transition between the last glacial period and the present interglacial, as indicated by the various layers of gyttja with different flora observed by Hartz and Milthers (Hartz and Milthers, 1901 as cited in Mangerud, 2020). One example of more recent evidence of the Y-D event can be found in lakebed sediments from Lake Meerfelder Maar in Germany (Brauer et al., 1999). From these lakebed cores, scientists were able to identify pollen samples that were used to date and characterize the past climates. Brauer et al. saw that during the Y-D cold period, non-arboreal pollen counts were higher as compared to the higher arboreal pollen counts from warmer periods (Brauer et al., 1999). Therefore, it can be theorized that in Western Europe during the Y-D event, there was a reduction of pine-birch forests due to the colder temperature (Brauer et al., 1999). This shift from forested woodlands to more open tundra vegetation was recorded by palynological studies throughout Europe, including the Netherlands, Switzerland, and the United Kingdom (Hoek, 2001; Lotter et al., 1992; Jones et al., 2002). Further evidence of a decrease in temperature comes in the form of fossil remains of Chironomidae (non-biting midges) (Heiri et al., 2007). The distribution of various taxa of Chironomidae have been shown to be significantly correlated with summer surface water temperature, so a chironomid-temperature relationship can be quantified and used as a temperature proxy (Walker et al., 1990). Heiri et al., using the temperature proxy nature of chironomids, showed that during the Y-D event, the summer temperatures at Hijkermeer in the northwest Netherlands were approximately 2-3°C cooler compared to the Allerød before it and the Holocene afterwards (Heiri et al., 2007). Around the Mediterranean, subfossil pines in southern France acted as a proxy through their ring-width/counts and their stable carbon and oxygen isotopes (Pauly et al., 2018). Like the other proxies discussed so far, these proxies indicated a cold period during the Y-D event and a southward shift of the polar front (Pauly et al., 2018). However, unlike the record at Lake Meerfelder Maar, the of Pauly et al. hinted at a bi-directional change in humidity and precipitation, meaning that there was an enhancement of extrema (Pauly et al., 2018). They attribute the upswings to local climate anomalies in the Mediterranean, as these shifts are not observed in the rest of Europe or the North Atlantic/Greenland (Pauly et al., 2018). From the fossils of various flora and fauna, it is evident that during the Y-D event, there was a cooling that led to tundra-like conditions in northern and western Europe, and strong climatic effects further south in Europe.

There also exist non-biological proxies that can be used to indicate the existence of the Y-D event in Europe and illustrate its climatic effects. One such example is periglacial evidence, which include fossilized cryogenic structures that can be used to identify various climatic thresholds. A subset of these features includes certain thermal contraction cracks (includes ice, sand, and composite wedges), which is associated with continuous permafrost. Looking at these features around Europe, a permafrost distribution can be determined. Further evaluation of these features can allow one to determine mean temperatures of the year and of the coldest month. After surveying various sites across Europe, it was determined that during the Y-D event, continuous permafrost existed as south as around 54°N, which includes Fennoscandia and the north of the British Isles (Isarin, 1997). Isarin's permafrost distribution is illustrated in Figure 5 for reference. Other thermal contraction cracks, such as soil wedges, are formed when there is seasonal permafrost. Identifying these, studies conclude that discontinuous permafrost existed as south as 50°N, which includes Ireland, England, and the European lowlands (Isarin, 1997). Making temperature inferences, Isarin proposes average winter temperatures during the Y-D event to be at least 15°C cooler than those of present day in continental Europe, and potentially more than 25°C cooler than present day in Ireland and Scotland (Isarin, 1997). On the other hand, summers did not have such a dramatic drop in temperature; temperature reconstructions seem to demonstrate only a drop of 4°C (Renssen and Isarin, 1998). This demonstrates that the extreme seasonality seen in Greenland can also be observed in parts of Europe. These reconstructions also seem to indicate an overall temperature range of 30°C in northwestern Europe, which is a much larger range than the 7°C range today C (Renssen and Isarin, 1998). For southern Europe, it was initially questionable whether or not any cooling occurred during the Y-D event (Rind et al., 1986). For places where permafrost didn't necessarily form, the aforementioned speleothem records can help one identify the Y-D signal. Examining the Seso cave speleothem record closer, it is shown that there was a potential temperature drop of 1.3°C (determined from $\delta^{18}O$) along with a drop in humidity (determined from $\delta^{13}C$) during the Y-D event (Bartolomé et al., 2015). This Seso Cave $\delta^{18}O$ drop can be seen in Figures 3a-c. Therefore, throughout Europe and Greenland, there was an observable drop in temperature and humidity during the Y-D event, though those trends vary in strength locally. On top of these temperature and humidity changes, there is also a potential change in wind patterns. Through the analysis of layers of sediment in Lake Meerfelder Maar, one can determine the wind stresses on the surface of the lake by making inferences about the resultant mixing (Brauer et al., 2008). From this data, Brauer et al. determined that a sudden increase in storminess during the autumn to spring seasons occurred at the onset of the Y-D event and attributed it to an increase in strength in westerly winds (Brauer et al., 2008). These westerlies were shifted south to a more zonal path, and the cause was identified to be a southward expansion of North Atlantic sea ice, and also aligns with the drop in wind measured in Greenland (Brauer et al., 2008). This is illustrated in Figure 6, which compares the modern-day winter wind patterns with the patterns proposed by Brauer et al. based off their Lake Meerfeldar Maar measurements. This wind shift demonstrates how the westerlies are mainly controlled conditions in the North Atlantic, so their increase in strength and shift in path, along with the temperature and humidity drops, demonstrate how a possible change in the North Atlantic could lead to drastic changes across Europe.

Focusing on ice sheets specifically, unlike the ones in southern Greenland, there were more notable responses within the remnants of the Scandinavian and Barents-Kara ice sheets. By studying the moraines (ice sheet deposits) of various glaciers and dating them, Andersen et al. tracked the retreat and advances of the Scandinavian Ice Sheet and identified a belt of moraines spanning across Europe (Andersen et al., 1995). They noted that in most areas, prior to the Y-D event, these ice sheets were in retreat, only to be followed by a readvancement during the Y-D event, with some advancements being up to 40 km (Andersen et al., 1995). More recent and local studies further demonstrate how there is concrete evidence suggesting that these ice sheets had large readvances in places such as western Norway and Iceland (Mangerud et al., 2016; Pétursson et al., 2015). Through the examination of sediments found in the Barents and Kara Seas carried by their ice sheets, it was also determined that remnants of the Barants-Kara Ice Sheet (located at the northern rim of Eurasia) had stopped its retreat or had even advanced during the Y-D event (Svendsen et al., 2004). Like the trends observed in the Scandinavian and Barents-Kara Ice Sheets, smaller local glaciers also saw a stoppage of retreat or an advancement. In the western highlands of Great Britain, a regrowth in glaciers was also observed along with a large drop in annual temperatures of up to 10°C in Scotland due to the Loch Lomond Stadial, which is the local name for the Y-D event in the region (Bickerdike et al., 2018). Further south in the Alps, it has been shown that there was an advancement in glaciers in Italy and Austria through the dating of moraines (Moran et al., 2016). Moran et al. also pointed to a mean summer temperature that is 1.5°C cooler than those of today (Moran et al., 2016). Overall, it can be seen that in Europe, the cooler temperatures and drop in moisture brought by the Y-D event generally halted the retreat of ice throughout the continent, and in some cases, resulted in the advancement of glaciers.

Newer studies of paleoclimatic proxies showed that the conditions in Europe during the Y-D event were not necessarily constant – they note an increase in strength of the Atlantic meridional overturning circulation (AMOC) during the middle of the Y-D event which caused a bipartition in European climate during the period. Pearce et al. noted that the oceans led the termination of the Y-D event, and a northward shift of the polar front started at around 12,300 years BP (Pearce et al., 2013). Speleothem records in northern Iberia show the potential for a 12,150 years BP event that indicated the beginning of a northward migration of westerlies (recall they had migrated south to become zonal jets at the onset of the Y-D event) (Baldini et al., 2015). This time period also indicated the slow northward movement of the polar front (Baldini et al., 2015). This shift is also recorded in Lake Lake Meerfelder Maar in Germany, dated to be at 12,240 years BP, and associated with a sudden increase in snowmelt in the spring (Lane et al., 2013). Bakke et al. also recognized a warming during the late Y-D and attributed it to atmospheric changes (Bakke et al., 2009). These changes caused rapid oscillations between the two positions for the westerlies and occurred right before the termination of the Y-D event (Bakke et al., 2009). Figure 7 shows the measurements done by Bakke et al. from Lake Kråkenes which show these rapid oscillations at the end of the Y-D event. Reflecting upon earlier records, this bipartition was also recognized in the Greenland ice cores through a decrease in dust in the latter portion of the Y-D event, though it is uncertain what the controlling mechanism was (Lane et al., 2013). Reflecting on older evidence, it can be seen that this bipartition was also indicated throughout Europe by plant fossils where the coldest temperatures occurred closer to the beginning of the Y-D event, and later temperatures being generally warmer (Isarin and Bohncke, 1999). Through these studies, it can be seen that although the Y-D event was a cold period in Europe, it still had climatic changes throughout its duration, and resulted in rapid climate oscillations right before its termination.

North America

In North America, we can see a similar cooling trend as compared to Europe, though not as strong in certain areas. Initially, there was doubt about the Y-D event affecting climate in all parts of North America. In 1986, Rind et al. compiled existing studies of paleoclimatic indicators across North America and drew the conclusion that the Y-D event being a worldwide event is a controversial claim (Rind et al., 1986). They had determined that only locations directly adjacent to the North Atlantic had records demonstrating evidence of a climatic oscillation (Rind et al., 1986). In eastern Canada, palynological studies showed that at around ~11,000 years BP, the vegetation in the southern Maritime provinces reverted to fewer trees and more tundra-based shrubs and herbs (Mott et al., 1986). Furthermore, a readvance of glaciers in Newfoundland and southern Quebec was observed (Mott et al., 1986). More recent studies will go on to prove that the Y-D signal can indeed be observed throughout the Americas, not just places near the North Atlantic. Moving to Ontario, various climate proxies provided additional evidence that the Y-D signal existed in central North America. Examining the sedimentary records from two small lakes in Ontario, Yu and Eicher extracted $\delta^{18}O$, $\delta^{13}C$, mollusk shells, and pollen profiles amongst others to indicate that the Y-D event in central North America started off with a temperature drop along with a thinning of forests and accelerated soil erosion (Yu and Eicher, 1998). Their determined records also aligned with other sites in the North Atlantic region, which indicates that the Y-D event carried similar effects throughout the central and eastern North America. Additionally, at some sites such as Twiss Marl Pond and Lake Crawford in Ontario, the $\delta^{18}O$ levels were at their lowest at the onset of the Y-D event and had a slight increase in the latter half of the period, matching with some of the bipartition behaviour observed in Europe (Yu and Wright, 2001). However, these changes within the Y-D event aren't observed in other sites (Yu and Wright Jr., 2001). On top of the temperature drop, sediment records in the Midwestern United States demonstrated an increase in aridity during the Y-D event (Wang et al., 2012). This correlates with an earlier soil analysis in Missouri that had demonstrated that there was a large expansion of grassland in the midcontinent, of which an increase in aridity is thought to be the cause (Dorale et al., 2010). These results further confirm the summarized palynological evidence of Shane and Anderson, which noted a reduction in precipitation and a decline in temperatures during the Y-D event in sites throughout Ohio and Indiana (Shane and Anderson, 1993). Overall, it can be seen that throughout northeast and central North America, the Y-D event was observed through a decrease in forested areas along with a temperature drop, a decrease in humidity, and a re-advancement of glaciers. This similarity in behaviour to that of western Europe hints that the North Atlantic has a strong role in the Y-D event, something that will be discussed further later.

The Y-D event also had a noticeable impact on the Laurentide Ice Sheet, which covered a significant portion of northeastern North America at the time. In Marquette, Michigan, Lowell et al. determined through the analysis of moraines that the Laurentide Ice Sheet had a re-advancement in the Lake Superior region (Lowell et al., 1999). They note that this Marquette readvancement is consistent with observations made by other groups, such as those made by LaSalle and Shilts in 1993, which saw a readvancement of the Laurentide Ice Sheet in the St. Lawrence Lowlands of southern Quebec during the Y-D event (LaSalle and Shilts, 1993). Furthermore, on the northwestern margin of the Laurentide Ice Sheet, there was an increase in iceberg discharge throughout the Y-D event (Andrews and Tedesco, 1992). This lies with the observations of advances of remnant Laurentide Ice Sheet glaciers on the northern portion of the Laurentide Ice Sheet on Baffin Island (Young et al., 2012). Young et al. also estimated a mean annual temperature drop from modern day of approximately 15°C, which correlates with some of the records made in Greenland (Young et al., 2012). They also point out a summer cooling that is at most 4.5 ºC to 5.5 ºC, which reinforces the seasonality of the Y-D event around the North

Atlantic (Young et al., 2012). On the southern edge of the Laurentide Ice Sheet is the glacial lake of Lake Agassiz. Prior to the Y-D event, Lake Agassiz had an outlet to the Gulf of Mexico through the Mississippi River (Broecker et al., 1989). At the onset of the Y-D event, this inflow of freshwater into the Gulf of Mexico stopped, indicated by a changed in oxygen isotope ratios in planktonic foraminifera, and remained that way until around the end of the Y-D event (Brocker et al., 1989). This indicates a potential rerouting of the Lake Agassiz drainage system at the onset of the Y-D event, due to the receding Laurentide Ice Sheet opening up potential new drainage sites (Broecker et al., 1989). The Laurentide Ice Sheet prior to the Y-D event was in recession, as the Northern Hemisphere was experiencing the Bølling-Allerød warm period. As to be discussed later, the route by which Lake Agassiz drained at the onset and during the Y-D event is still a topic of debate. Furthermore, Lake Agassiz can potentially be instrumental in triggering the Y-D event, as to be discussed later as well. During the Y-D event, the glacial advances observed by Lowell et al. likely closed up the rerouted drainage of Lake Agassiz, and the outflow returned to the Mississippi (Broecker et al., 1989). These glacial measurements are able to aid in the understanding of the climate across North America through capturing the behaviour of the Laurentide Ice Sheet, and also provide potential insight into what caused the Y-D event to start.

Turning focus from the east coast of North America to the west, similar cooling trends can be observed along with glacial readvances. Along the Pacific coast spanning from Washington to Alaska, palynological and sedimentary evidence indicate that there is a similar climate oscillation in terms of timing and effects as the Y-D event, with the exception of an increase in moisture (Mathewes, 1993; Kaufman et al., 2010). Kaufman et al. note that this increase in moisture is not documented by many other studies in this region and proposes that its occurrence is due to a strengthening of the Aleutian low-pressure system, which has been shown to occur when the North Atlantic receives a pulse of fresh water (Kaufman et al., 2010; Okumura et al., 2009). This further strengthens the idea that a change in the North Atlantic caused the Y-D event. In the northwest, the same transition from forest to non-arboreal plants in eastern North America is observed, though Mathewes notes that it is a result of both cooler and wetter climates as opposed to cooler and drier in the east (Mathewes, 1993). Focusing in on Alaska, the cooling trend was further emphasized with an advance in alpine glaciers in the Ahklun Mountains (Briner et al., 2002). Further south, the North American Cordillera Ice Sheet, which used to cover Washington and British Columbia and was in retreat prior to the Y-D event, saw periods of re-advance throughout the cold period and is marked by moraines located in the sheet's southwest margin (Kovanen, 2002). Furthermore, mountain glaciers throughout the Cordillera mountain range in North America also saw a readvancement, as evidenced by many dated moraines found throughout the Canadian and American Cordillera ranges (Friele and Clague, 2002; Davis et al., 2009). As we continue even further south to California, speleothem records indicate a drop in $\delta^{18}O$ and $\delta^{13}C$ at around 12,400 years BP, slightly after the accepted onset of the Y-D event, though Oster et al. note that it is within the dating error (Oster et al., 2009). These shifts in isotope ratios along with the increased growth rate of speleothems during this period indicate that there was a drop in temperature, and an increase in moisture. This high-resolution speleothem record is one of many studies that predict this cooler and wetter change in the region (MacDonald et al., 2008; Oviatt et al., 2005). On the other hand, some other studies have correlated cold, dry events in the region with the cooling events found in the north Atlantic (Benson et al., 1998). Nonetheless, a more recent speleothem study in Arizona seems to further bolster the idea that cold periods in the southwestern United States are correlated with an increase in moisture (Wagner et al., 2010). Whether drier or wetter, these paleoclimatic studies agree that in the southwestern United States, a cooling was observed during the Y-D event. Through the examination of fossil packrat pellets to determine their diet, and subsequently associating their plant diet with the temperature limits of Utah agave, Cole and Arundel were able construct a paleotemperature proxy to obtain a value for Y-D temperatures in the region (Cole and Arundel, 2005). They ultimately determined a minimum temperature drop of between 7.5-8.7°C for the Y-D period when compared to present temperatures (Cole and Arundel, 2005). Finally, when focusing on the southeastern United States, it appears that there may have been an increase in temperatures off the coast of Florida shown by $\delta^{18}O$ profiles of sediment cores (Carlson et al., 2008a). Furthermore, sediment cores carrying debris indicate that there is a likely an increase in hurricane activity during the Y-D event (Toomey et al., 2017). Like the wetness in Alaska, these rather unique effects (when compared to the rest of North America) seem to relate to the effects of the AMOC slowing down, a key mechanism driving the Y-D event which has consequences in the Gulf of Mexico (Toomey et al., 2017). This slowdown/shutdown will be discussed further when an investigation on the mechanism behind the Y-D period is conducted.

Central/South America

Continuing south to Central America and South America, it can be seen that there is very strong evidence of a climatic oscillation occurring within the tropics during the Y-D period, though there seems to be varying effects from location to location. At 4 degrees latitude south of the equator in the Atlantic off the coast of Brazil, sediment cores show that there was a 1.3°C increase in temperature during the Y-D event (Weldeab et al., 2006). They attained this value by reconstructing the temperature record through measurements of $\delta^{18}O$ found in planktonic foraminifera remnants in the sediment cores (Weldeab et al., 2006). Like Toomey et al. in 2006, Weldeab et al., believed this increase to be partially due to the weakened AMOC. Additionally, they cited this increase in temperature to be associated with a southward displacement of the Inter Tropical Convergence Zone (ITCZ) (Weldeab et al., 2006). The ITCZ is a narrow belt spanning the globe centred around 6 degrees north of the equator and is where rainfall is typically the most intense (Schneider et al., 2014). It seasonally migrates to whatever hemisphere that warms relative to the other, and during climatic events like the Y-D event, it resides in the warmer hemisphere (Schneider et al., 2014). Therefore, during the Y-D cooling of the Northern Hemisphere, the ITCZ undergoes a southward migration, potentially explaining the increases in temperature just south of the equator. Figure 8 shows how the ITCZ has migrated in the past ~65,000 years and demonstrates a southward movement during the Y-D event along with other periods of cold. This should result in a cooling in the north and a warming in the south along with respective decreases and increases in precipitation. However, this trend is not fully followed, as warming trends are shown to have occurred in the Caribbean Sea and the Gulf of Mexico above the equator (Schmidt et al., 2004; Flower et al., 2004). On the other hand, Lea et al. indeed noted a cooling off the northern coast of Venezuela, around 10 degrees north of the equator (Lea et al., 2003). After reconstructing sea surface temperatures from foraminifera, they concluded that during the Y-D event, the temperatures dropped by 3-4°C, which puts it in synchroneity with the effects of the North Atlantic (Lea et al., 2003). These cooling effects can be observed inland through glacial advances in the Andes mountains. Mahaney et al., dated moraine deposits in the northern Venezuelan Andes, and concluded that the latest glacial advances occurred during the Y-D event (Mahaney et al., 2008). Further cooling in Central America was observed in Costa Rica and Guatemala where temperature drops of 1.5-2.5°C was inferred, though a site in Panama did not observe any cooling (Leyden, 1995). From these different observations, it can be seen that the Y-D event is quite a complex event in tropical Central and South America in terms of temperature. However, both the studies of Lea et al. and Weldeab et al. have consensus that their observations are a result of a southward displacement of the ITCZ, which would have caused an upwelling of cold subsurface water in the Cariaco Basin, where Lea et al. took their measurements (Lea et al., 2003; Weldeab et al., 2006). There is stronger consensus about precipitation changes during the Y-D event, where it appears to have gotten stronger south of the equator, and weaker towards the north due to monsoon pattern changes. This can be observed in speleothems in Brazil and sediment cores off the coast of Brazil that indicate an increase in precipitation during the Y-D event (Wang et al., 2007; Jaeschke et al., 2007). In the same location studied by Lea et al., in the Cariaco Basin of the northern coast of Venezuela, Haug et al. observed a drier Y-D event in the titanium and iron concentration data of the sediment cores (Haug et al., 2001). Like with fluctuations in temperatures, this increase in aridity during the Y-D event is explained by a southward migration of the ITCZ (Haug et al., 2001). The southward migration of the ITCZ also heavily influenced the South American Monsoon. A study done by Novello et al., showed a wetter South American Monsoon south of the equator during the Y-D event (Novello et al., 2017). This trend is consistent with the observations in the region and also buttresses the idea that there was a migration south of the ITCZ (Novella et al., 2017). Overall, it can be seen that the climate changes in Central America and parts of South America are strongly influenced by changes in the ITCZ.

Further south in South America, the Y-D signal remains inconsistent. Although it appears that in the Atlantic just south of the equator off the coast of Brazil, temperatures had a slight increase, measurements throughout the continent give mixed results (Weldeab et al., 2006). In the tropical Andes Mountains, an advance of the mountain glaciers was observed though various moraine deposits dated to the Y-D period, indicating a cooling (Clapperton, 1993; Rodbell et al., 2009; Jomelli et al., 2014). Hajdas et al. also observed a cooling in mid-latitude South America and attributed to be related to the Y-D event even though it preceded the onset of the Y-D event in Europe by about 550 years (Hajdas et al., 2003). From their radiocarbon chronology of sediment cores, they also recognized that there was a lack of warming in this region prior to the Y-D event, differentiating it from trends in the Northern Hemisphere, which was undergoing the Bølling-Allerød warm period (Hajdas et al., 2003). Further south, it is seen that the southernmost Andes do not seem to experience any glacial advances (Rodbell et al., 2009). At the very bottom of South America in southwestern Patagonia, an onset of warm conditions was associated with the timing of the Y-D event (Moreno et al., 2009). These observations in southern Patagonia contrast with other observations that indicate a cooling until 11,300 years BP in northern Patagonia (Moreno et al., 2009). Furthermore, these variations seem to be predominantly influenced by another event called the Antarctic Cold Reversal, which occurred prior to the Y-D event 14,500 to 12,900 years ago (Rodbell et al., 2009; Jomelli et al., 2014). During this event, glacial advances were far stronger throughout the Andes as compared to the Y-D event (Jomelli et al., 2014). This potentially indicates that even though glaciers still advanced in the Andes during the Y-D event, the climate was actually warmer than the period that occurred just before it, the Antarctic Cold Reversal. Moreno et al. also noted that glacial activity was a lot stronger during this time in Patagonia, and that the Y-D event brought warm conditions that caused a retreat of southern Patagonian glaciers (Moreno et al., 2009). Kaplan et al. also note that there is still a lot of uncertainty in dating, so it may still seem ambiguous whether or not colder climates in South America are attributed to the Antarctic Cold Reversal or the Y-D event (Kaplan et al., 2008). The Antarctic Cold Reversal will be elaborated further when an evaluation of the Y-D signal in Antarctica is performed. Nonetheless, as to be shown later, the warming trends in southern South America seem to be generally consistent with the Southern Hemisphere trend of warming generally observed throughout. The more general cooling trends in the north could be considered to be in line with the phenomena observed with tropical Central America, due to the southern migration of the ITCZ.

Asia

From an earlier section, it was seen that the remnants Barents-Kara Ice sheet had halted their retreat in some places and advanced in others during the Y-D event, indicating that at least for the northwesternmost portion of Asia, the Y-D event effects can be seen. For northern Asia, there is still a lack of data for a full understanding of the area, however some pollen studies have shown various temperature drops due to the Y-D event throughout eastern Europe and Siberia (Velichko et al., 2002). From a review of these palynological studies, it is shown that during the Y-D event in this northern region, January temperatures dropped by between 4-6°C while July temperatures dropped by 2-11°C (Velichko et al., 2002). The more severe temperature drops occurred closest to the Scandinavian ice sheets, but overall, these much colder winters along with slightly cooler summers seem to be in trend with the aforementioned European observations. Furthermore, like in Europe, precipitation decreased throughout northern Asia, varying from being 80% of present-day values in central West Siberia all the way to being 50% of present-day values in central Yakutia (northwestern Russia) (Velichko et al., 2002). Furthermore, the permafrost line saw a migration further south during this period (Velichko et al., 2002). Outside of palynological studies, cores from Lake Baikal in Russia also indicate a potential cooling along with a change in moisture (Morley et al., 2005). They also note that Lake Baikal had an increase in river discharge in the north, and a decrease in the south, and attributed it to changes in atmosphere controlled by the oscillation in the North Atlantic (Livingstone, 1999; Morley et al., 2005). Overall, it appears that Siberia as a whole seems to behave very much like Europe during the Y-D event.

For the vast majority of the rest of Asia, the climate is mostly controlled by the Asian Monsoon and the Asian Westerlies. The Asian Monsoon transports heat and moisture all the way from northern Australia to northeastern China and Japan (Cheng et al., 2012). The Asian Westerlies, which are a predominant factor in central Asian climate, also seem to be linked to the Asian Monsoon (Cheng et al., 2016). The westerlies carried cold European temperatures and aridity into central China during the Y-D event, and the resulting drops in temperature can be seen in the oxygen isotope profiles from various caves in Figure 3a (Cheng et al., 2020). Taking cores from the northern East China Sea, Kubota et al. determined the temperatures based off the oxygen isotope ratios of foraminifera and found that they were about 2°C lower than those of present day (Kubota et al., 2010). They also suggest a linkage with the North Atlantic, as they noticed that cold events in the North Atlantic correlated with weaker East Asian Summer Monsoons, meaning a decrease in precipitation as well (Kubota et al., 2010). In the Okinawa Trough, Sun et al. observed a similar drop in sea surface temperatures, further strengthening the idea that changes in the North Atlantic had a strong effect on monsoon climate (Sun et al., 2005). On land in Japan, Nakagawa et al. noted a cold reversal as well, though they dated it to occur a few centuries after the onset of the Y-D event in the North Atlantic (Nakagawa et al., 2003). With their data from sediment cores in Lake Suigetsu, Nakagawa et al. expected this cold reversal to be due to a transmission of effects from the North Atlantic to the Northwest Pacific, though they acknowledge that a mechanism is yet to be defined (Nakagawa et al., 2003). A revisitation of Lake Suigetsu by Schlolaut et al. found that through their proxy data, the bipartition observed in Europe was also observed in eastern Asia, but in reverse (Schlolaut et al., 2017). They suggest that in the second phase of the Y-D event in Europe, the northward shift of the westerlies would increase the transport of moisture towards central Asia. This increase in moisture would then increase snow cover in Eurasia which would then strengthen the East Asian Winter Monsoon and weaken the East Asian Summer Monsoon (Schlolaut et al., 2017). This could potentially help explain the lag in the cold period observations made by Nakagawa et al. Further speleothem samples made in the Chinese Loess Plateau have helped in confirming that it is highly likely that the AMOC is a driver of changes in the East Asian winter and summer monsoons (Sun et al., 2011). The teleconnection carrying this signal is identified to be the northern westerlies (Sun et al., 2011). This further demonstrates how the North Atlantic can strongly influence the climate around the globe through various teleconnections. Thus, it can be seen that in China and Japan, there's a weakening of the summer monsoon, a strengthening of the winter monsoon, and an overall drop in temperatures during the Y-D event.

Continuing on the topic of monsoon changes, there is much evidence supporting a weaker Indian Summer Monsoon during the Y-D event. Through the analysis of glacial outwash sedimentary profiles from the Himalayas, it was shown that between ~12,700 and 11,600 years BP, the Indian Summer Monsoon was observed to generally be weaker than that of today's, with the weakest point being about a 63% drop in rainfall (Ali et al., 2018). Ali et al. also noted an abrupt increase in strength of the Indian Summer Monsoon at around 12,400 years BP that lasted for a century, which could potentially be further evidence of a bipartition within the Y-D event (Ali et al., 2018). Overall, their observed weakening of the summer monsoon is consistent with the aforementioned studies in China and Japan, in addition to many other investigations around India (e.g., Rashid et al., 2007; Kudrass et al., 2001). Overall, the Asian Monsoon changes seem to lie synchronous with the initial onset of the Y-D event, implying a North Atlantic trigger that is quickly propagated through the atmosphere (Cheng et al., 2020). Furthermore, due to the Y-D event onset seemingly lasting longer in these regions, it suggests that the Y-D event onset propagates from northern high latitudes to lower latitudes (Cheng et al., 2020). A likewise longer termination in this region suggests the same directionality for the termination of the Y-D event (Cheng et al., 2020). The Y-D event can also be identified in the Asian tropics, with Indonesian pollen diagrams containing a possible signal as well as sediment cores from the Sulu Sea (Maloney, 1995; Linsley and Thunell, 1990). Overall, these changes observed in tropical Asia seem to be impacted by similar mechanisms as those of the American tropics. Namely, there is a southward migration of the ITCZ, which in combination with the changes in the AMOC and southern shift of westerlies, cause the weakening of the Asian Summer Monsoons (Cheng et al., 2020).

Pacific Ocean
There seems to be a variety of effects of the Y-D event in the Pacific as it covers both the Northern and Southern Hemisphere. In the north, it has been demonstrated that the North Pacific is heavily influenced by sea-ice variability, which is in turn heavily influenced by the North

Atlantic (Max et al., 2012). This implies that the Northern Pacific is closely coupled with the North Atlantic, and such a correlation is demonstrated in various cores from the western Bering Sea, the continental slope of east Kamchatka, and the southern Sea of Okhotsk, which show a temperature decrease of between 1.5-5°C during the Y-D event (Max et al., 2012). This cold trend in the Pacific is similar to the coolings observed in Alaska and Siberia as explored earlier (Briner et al., 2002; Velichko et al., 2002), and further reinforces the general Northern Hemispheric cooling effect of the Y-D event. Moving towards the tropics, it can be seen that the tropical Pacific exhibits similar effects during the Y-D event as other tropics around the world. As mentioned earlier when describing the Asian Monsoons, in the tropical western Pacific (the Sulu Sea), the Y-D signal can be identified (Linsley and Thunell, 1990). One of the earlier studies indicating the global nature of the Y-D event, Linsley and Thunell recorded the reappearance of cool water planktonic foraminifera during the onset of the Y-D event in sediment cores, indicating a drop in temperature in that region (Linsley and Thunell, 1990). Furthermore, from speleothem records in the Philippines, it was further shown that the northern summer monsoons were weakened (Partin et al., 2015). This cooling/drop in moisture trend seems to generally align with trends above the equator in South America on the other side of the Pacific Ocean (Lea et al., 2003; Hajdas et al., 2003). As we move south past the equator in the Pacific, the western Pacific is dominated by the Australian-Indonesian Monsoon. Partin et al. suggested a mild strengthening of the summer monsoon during the Y-D event after evaluating various proxy records from Indonesia, however acknowledging that there is no direct evidence to date (Partin et al., 2015). A summary of the results collected by Partin et al. can be found in Figure 9 and demonstrate the strong trend that is the weakening of the summer monsoon in the north. An analysis of further speleothem records by Cheng et al. (with some isotope profiles being shown in Figure 3b) seem to suggest an increase in temperature and rainfall in this region during the Y-D event, though more gradual than the sudden changes observed in the Northern Hemisphere (Cheng et al., 2020). On the other side of the Pacific, there is likewise an increase in wet conditions corresponding with the Y-D event (Novello et al., 2017). The overall weakening of summer monsoons in the north/strengthening of monsoons in the south are uniform with the southward movement of the ITCZ during the Y-D event (Benway et al., 2006).

The tropical Pacific is also home to the El Niño-Southern Oscillation (ENSO), which strongly affects the tropical eastern Pacific. During the Y-D event, effects on this system can be observed. Using reconstructed sea surface temperatures from sea-floor sediments off the coast of the Galápagos islands and the South China Sea, Koutavas et al. noted a relaxation of temperature gradients during the Y-D event (Koutavas et al., 2002). The gradient typically consists of warmer waters in the western Pacific with cooler waters off the coast of South America and El Niño conditions would imply warm conditions in the east which weaken the temperature gradient; during the Y-D event, the cooling observed in the South China Sea creates a persistent El Niño-like state as it weakens that temperature gradient (Koutavas et al., 2002). Through these records, it can be shown that in general, El Niño conditions correlate with stadials (cold periods) at high latitudes while La Niña conditions correlate with interstadials (Stott et al., 2002). It has also been concluded that the switch from the El Niño state to a La Niña state signals the termination of the Y-D event in the tropical Pacific, and that it seems to end before the termination in the North Atlantic (Cheng et al., 2020). This leads Cheng et al. to claim that the trigger for the termination of the Y-D event lies potentially in the tropics or the Southern Hemisphere in general (Cheng et al., 2020). Although it is still up to debate where the triggers for onset and the termination of the

Y-D event lie, it can be seen with confidence that during the Y-D event, the tropical Pacific saw a weakening of its northern monsoon, along with a consistent El Niño-like state that shifted towards a more modern La Niña state during/after the Y-D termination.

Finally, in the Southern Pacific Ocean, there is further uncertainty in the effects of the Y-D event as studies seem to give conflicting results. It seems initial studies noted an advance of glaciers on land by the southern Pacific and used that to push the idea of a consistent global cooling during the Y-D event. Lowell et al. noted glacial advances in the Southern Alps in New Zealand in addition to glacial advances in southern Chile on the opposite side of the Pacific and used these observations to claim a synchronous cooling of both hemispheres during the Y-D event (Lowell et al., 1995). Speleothem records analyzed by Goede et al. in Australia also pointed towards a cooling in the Southern Hemisphere during this period (Goede et al., 1996). This dating by Geode et al. would later be reanalyzed and refuted by Green et al. in 2013, whose refutation went on to suggest a lack of synchroneity between the Northern and Southern Hemisphere temperatures (Green et al., 2013). Additionally, McGlone conducted a vegetation change and glacial trend analysis in New Zealand and claimed that the poor chronological resolution and no apparent trend in different types of evidence make the Y-D signal in New Zealand quite ambiguous (McGlone, 1995). Analysis of pollen records by Newnham and Lowe, along with additional multiproxy results from Turney et al., suggest that there was a warming in New Zealand during the Y-D event (Newnham and Low, 2000; Turney et al., 2003). Turney et al. further noted an increase in westerly airflow (Turney et al., 2003). Contrary to the observations that suggest a unanimous cooling, these observations of a warmer Southern Hemisphere support the "bipolar seesaw" mechanism that involves deep water formation, as to be explained further on in this paper. In 2005, Williams et al. conducted an analysis of speleothems in New Zealand and concluded that there was a period of warming in New Zealand during the Y-D event (Williams et al., 2005). They also noted that the observed glacier advances by those earlier do not necessarily prove the occurrence of a Y-D cooling due to differences in timing, duration, and structure of the climate reversal in New Zealand when compared with other regions (Williams et al., 2005). Furthermore, newer moraine dating techniques in the Southern Alps glaciers using beryllium-10 by Kaplan et al. in 2010 and Koffman et al. in 2017 further demonstrate a warming and glacier recession during the Y-D event (Kaplan et al., 2010; Koffman et al., 2017). These studies also generally demonstrate a correlation with the Antarctic Cold Reversal, where glacier advances were stopped or reversed at the onset of the Y-D event, just like the southernmost locations in South America (Putnam et al., 2010). Therefore, the Southern Pacific, although with some uncertainty, most likely observed a warming trend during the Y-D event as shown with more recent evidence.

Africa

Africa generally seems to follow similar trends observed by places of similar latitude around the world during the Y-D event. Starting by focusing on Northern Africa, it appears that it keeps in trend with the effects in Europe, where there is a colder and drier climate. Genty et al. analyzed speleothem samples from Southern France and Northern Tunisia and noted a clear cool and arid period during the Y-D event for both regions (Genty et al., 2006). Prior to the Y-D event, subtropical North Africa was undergoing the African Humid Period, which is estimated to have occurred 14,800 to 5,500 years BP based off eolian sediment deposits (sediment being blown into the sea by wind) off the coast of Mauritania (deMenocal et al., 2000). During this period, the

Sahara was covered in vegetation, and even had lakes. As a result, less sediment would get blown away enabling it to act as a proxy for humidity. From the eolian sediment deposits, deMenocal et al. noted an interruption in the African Humid Period where there was a return to aridity, and its dating corresponded with the Y-D event (deMenocal et al., 2000). The African Humid Period is further associated with a strengthening of the African monsoon, so then it makes sense to assume that the arid interruption would consist of a weaker African monsoon (deMenocal et al., 2000). In Lake Masoko in Tanzania, sediment cores reveal a strong transition dated to around the termination of the Y-D event; from pollen records, it shows that in equatorial eastern Africa the end of the Y-D event was marked by a resumption of monsoonal activity and a return from severe dry seasons associated with the Y-D event (Garcin et al., 2007). Garcin et al. also point out a very pronounced migration of the ITCZ over Africa, an observation consistent with those made in the South American and Asian tropics (Garcin et al., 2007). The weakening of the African monsoon during the Y-D event can also be observed in cores in Lake Tanganyika, Lake Magadi in Kenya, and also in speleothem records from Socotra Island in the Indian Ocean (Tierney and Russell, 2007; Williamson et al., 1993; Shakun et al., 2007). Further south of the equator, a Y-D warming was observed off the coast of Madagascar and was implied to be caused by a change in the THC transport from the Atlantic Ocean to the Indian Ocean (Levi et al., 2007). Therefore, throughout Africa, the impacts of the Y-D event are quite similar to places of equivalent latitude around the world, mostly due to the impact of the southern displacement of the ITCZ which in turn weakens the African monsoon.

Antarctica

As implied earlier, a large climatic event that influenced the Southern Hemisphere was the Antarctic Cold Reversal, and its phasing with the Bølling-Allerød warm period in the Northern Hemisphere and the Y-D period is uncertain. Blunier et al. note that the Arctic Cold Reversal onset occurred around 1,800 years before the cooling of the Y-D event (Blunier et al., 1997). They determined this offset through the synchronization of ice cores from Byrd Station and Vostok in Antarctica with the GRIP ice cores from Greenland through methane measurements (Blunier et al, 1997). The Byrd core profile can be seen in Figure 4. Furthermore, throughout the Y-D event, an increasing temperature trend was observed in the Antarctic cores (Blunier et al., 1997). On the other hand, stable isotope measurements done by Steig et al. at the Taylor Dome suggest a synchroneity between the Northern Hemisphere and Southern Hemisphere, where the Y-D event is a cold interval (Steig et al., 1998). Jouzel et al. had a separate core drilled on the East Antarctic Plateau, and their results further pointed at a possible synchroneity between the hemispheres (Jouzel et al., 2001). The results of these cores are significant as they suggest that the "bipolar seesaw" model of ocean circulation (which will be expanded upon further) is inaccurate (Jouzel et al., 2001). Samples from the Law Dome, like the samples made by Steig et al. and Jouzel et al., note that the Antarctic Cold Reversal is not in phase with the Northern Hemispheric warming as assumed by the "bipolar seesaw" model, though unlike those earlier samples, they note a warming during the Y-D event (Morgan et al., 2002). More recently, cores made by the European Project for Ice Coring in Antarctica (EPICA) in Dronning Maud Land and Dome C demonstrate a one-to-one coupling between all warm events in the Antarctic and cold stadials in the Northern Hemisphere, not just the Y-D event (EPICA Community Members, 2006). This interhemispheric coupling of all bipolar climate variations demonstrated by their $\delta^{18}O$ records makes it seem that a warming in the Southern Hemisphere is more likely during the Y-D event than a cooling (EPICA Community Members, 2006). The last 40,000 years of this trend observed by EPICA Members is also in agreement with a core made in the Siple Dome by Ahn et al. in 2004, which tracked climatic changes through carbon dioxide concentration levels throughout the core (Ahn et al., 2004). Stenni et al. point out that the cores from the Taylor Dome and the Law Dome that suggest a synchroneity come from peripheral sites where the stratigraphy is highly deformable, as opposed to the records from EPICA (Stenni et al., 2011). They note that in the time interval for the Antarctic Cold Reversal, the timescale at the Taylor Dome and the time resolution at the Law dome could produce questionable conclusions (Stenni et al., 2010). Stenni et al. go on to analyze a new core drilled in the Talos Dome, which is a peripheral dome as well, and find that its record is clearly asynchronous with those from the Northern Hemisphere (Stenni et al., 2011). This demonstrates a correlation between the Antarctic Cold Period with the Bølling-Allerød warm period in the Northern Hemisphere, and also links the Y-D event with a gradual warming (Stenni et al., 2011). These connections are illustrated in Figure 10, which illustrates the warming trend across various Antarctic domes along with the trends in Greenland. Therefore, although there is not a complete consensus in the behaviour of Antarctic climate during the Y-D event, based off the more recent studies, along with other studies throughout the Southern Hemisphere, it is most likely that a warming trend occurred during that time period.

Global Summary

As demonstrated, through the various paleoclimatic studies analyzing past flora, fauna, and abiotic climate proxies from sources such as glacial deposits, ice cores and marine sediment cores, researchers are able to reconstruct the past environment and observe the global effects of the Y-D event. A quick summary will be given based off latitudinal observations, starting at the north and travelling down south. To start, the Y-D signal is clearest in ice cores made in Greenland, where a cool period was observed (e.g., Denton et al., 2005). The Y-D event appeared in Europe primarily as a cold and dry event during the last deglaciation (e.g., Bartlolomé et al., 2015). It saw a reduction of forests, and a return of tundra flora (e.g., Brauer et al., 1999). There was a halting of retreat in glaciers and ice sheets throughout Europe, and even a readvancement in certain areas (e.g., Anderson et al., 1995). A shift in the westerlies was also observed as they switched to a more zonal path (Brauer et al., 2008). This switch provides a mechanism in aiding the drop in temperatures and moisture (Brauer et al., 2008). Furthermore, although there was a cooling trend throughout Europe, the latter portion of the Y-D event saw a slight warming, and a more unstable climate (e.g., Baldini et al., 2015). This indicates a possible bipartition during the time period. These changes in Europe are strongly linked to a change in the North Atlantic, which is thought to be the epicentre of the event. The European and Greenland records also mark a strong seasonality throughout the Y-D event, with winter temperatures dropping a lot more than summer temperatures (e.g., Renssen and Isarin, 1998). North America, like Europe, also saw a general cooling and drying trend during the Y-D period (e.g., Yu and Eicher, 1998; Young et al., 2012). Eastern North America, being closest to the North Atlantic, saw the strongest drops in temperature and moisture (e.g., Mott et al., 1986). There is also a possibility of a bipartition of the period akin to the one observed in some areas in Europe (Yu and Wright Jr., 2001). In central North America, the Laurentide Ice Sheet, which was in retreat prior to the onset, saw a readvancement (e.g., Lowell et al., 1999). Glaciers in the western Cordillera also saw a stoppage in retreat or a re-advancement (e.g., Friele and Clague, 2002). Further south in North America, the cooling trend can still be observed, though near the Gulf of Mexico, the changes in that region are different than elsewhere possibly due to its proximity to the Gulf which is strongly affected by the changes in the Atlantic that caused the onset of the Y-D event (Carlson et al., 2008a).

Another prominent feature of the Y-D event is a southward displacement of the Inter Tropical Convergence Zone. The shift in the ITCZ was observed throughout the tropics in South America, Asia, Africa, and the Pacific Ocean (e.g., Novello et al., 2017; Cheng et al., 2020; Garcin et al., 2007). In South and Central America, this generally led to an increase in aridity in the Northern Hemisphere, and a wetter tropical climate south of the equator (e.g., Haug et al., 2001; Wang et al., 2007). However, there is no solid consensus on the overall temperature trend in certain regions in South America (e.g., Lea et al., 2003; Weldeab et al., 2006). In the northern subtropics in Asia, a weakening of the summer monsoon, a strengthening of the winter monsoon, and an overall drop in temperature was observed (e.g., Cheng et al., 2020). In the Southern Hemisphere this trend was likely flipped, so the summer monsoons may have gotten stronger (Partin et al., 2015). Along with these changes in the Asian Monsoon, similar changes in the South American, Indian, African, and Australian-Indonesian Monsoon patterns were observed (e.g., Cheng et al., 2020). Due to the weakening of the African Monsoon resulting from the southward migration of the ITCZ, Africa saw an interruption to the African Humid Period, where a return to aridity occurred in the Sahara during the Y-D event (e.g., deMenocal et al., 2000). In the Pacific, the shift in the ITCZ strongly influenced the ENSO patterns, and resulted in a consistent El Niño-like state throughout the duration of the Y-D event (e.g., Koutavas et al., 2002). Further south, there is an apparent warming trend throughout the Southern Hemisphere. In South America, there was a general increase in temperatures which lead to glacial retreats (e.g., Moreno et al., 2009). However, there are conflicting studies that also suggest a drop in temperatures. Some confusion arises with timing, as in the Southern Hemisphere, the Antarctic Cold Reversal occurred before the Y-D event, and it is uncertain whether or not the Y-D event is a continuation of this cold reversal (hence resulting the advancement of in certain glaciers in the Andes, New Zealand, and Australia) or an interruption to this reversal (causing warmer temperatures and retreats in those same locations) (e.g., Jomelli et al., 2014; Lowell et al., 1995; Turney et al., 2003). Further insight can be gleamed from Antarctic ice cores, which generally support the Antarctic Cold Reversal being synchronous with the Bølling-Allerød warm period, which means the Y-D event was a warm interruption in the Southern Hemisphere (e.g., Stenni et al., 2011).

One major synchronizing factor that was briefly touched upon earlier was the usage of atmospheric methane to match trends between different locations. Methane records from ice cores are usually utilized to reflect terrestrial climates, as they depend on a variety of factors, including temperature, hydrologic balance, and wetland production; warmer, wetter, and more productive conditions result in higher methane concentrations (Brook et al., 2000). Notably, there was a drop in methane during the Y-D event, and it can be identified in ice cores in both Greenland and Antarctica, even though their $\delta^{18}O$ profiles differ (Grootes et al., 1993; Steig et al., 2000; Brook et al., 2000). This also demonstrates that it was likely a global atmospheric change rather than local changes, as this trend is observed at both poles (Brook et al., 2000). This global dip in methane is what allowed scientists to synchronize the time scales in the various cores made at the Greenland and Antarctic ice sheets. Figure 10 shows how the methane records in Greenland closely resemble the ones in Antarctica. Furthermore, Brook et al. go on to claim that this dip in atmospheric methane during the Y-D event represents a terrestrial ecosystem response – both tropical and boreal environments were affected by the Y-D event, and their changes (e.g., a drop in wetland coverage) could reduce methane output (Brook et al., 2000). Another atmospheric gas of interest throughout the Y-D event is carbon dioxide. Throughout the Y-D event, there was a steady increase in atmospheric CO_2 as recorded in Antarctic ice cores (Monnin et al., 2001). Figure 10 has a plot of Antarctic carbon dioxide levels from the past 25,000 years, in which a steady rise is observed during the Y-D event. Likewise, a rise in CO_2 can be found in Greenland ice cores (Liu et al., 2012). This rise in atmospheric CO_2 has been associated with a release of carbon from the oceans, which is in turn related to a switch in ocean circulation (Marchitto et al., 2007). This has been shown to occur in the Pacific and also the Southern Ocean (Marchitto et al., 2007; Anderson et al., 2009). Overall, the changes in atmospheric gasses help reflect global changes during the Y-D event, including a reduction of wetland productivity illustrated by a drop in methane, and a change in ocean circulation demonstrated by an increase in carbon dioxide.

Finally, there have been various studies that have coalesced paleoclimatic proxy measurements into one study to observe global trends more closely. Shakun and Carlson compiled 90 temperature records from the Y-D event and utilized empirical orthogonal function (EOF) analysis to determine what modes drove climate variability during that time (Shakun and Carlson, 2010). Shown in Figure 11, it can be seen that EOF1 has a large southern hemisphere positive loading and EOF2 has a large northern hemisphere positive loading, which demonstrates the "southern" and "northern" modes of variability, demonstrating the evidence of the "bipolar" seesaw that has been gleamed upon throughout this paper (Shakun and Carlson, 2010). These two EOFs explain over half of the variability between 15,000-11,000 years BP (Shakun and Carlson, 2010). Furthermore, their principal components (shown in Figure 12) demonstrate the Antarctic Cold Reversal followed by increasing temperatures during the Y-D event in the Southern Hemisphere along with the warm Bølling-Allerød period and the cold Y-D event in the Northern Hemisphere (Shakun and Carlson, 2010). From these different proxies, a temperature estimate can also be made (Figure 13); Shakun and Carlson note an average temperature anomaly of -5°C for northern high latitudes, -2°C for northern mid latitudes, -1°C to 1°C for the tropics, and about 1°C for the Southern Hemisphere (Shakun and Carlson, 2010). These changes average out to an average global cooling of about 0.6°C throughout the Y-D event (Shakun and Carlson, 2010). A similar EOF study done by Clark et al. added an additional 97 records to determine the main modes of variability on a regional and global basis from 20,000 years to 11,000 years BP (Clark et al., 2012). In addition to similar findings from Shakun and Carlson, they further emphasized that the global EOF1 has a strong correlation with greenhouse gasses and the overall global warming of the last glaciation, and EOF2 reflects changes in ocean circulation (Clark et al., 2012). From these EOF studies, a good global perspective is given through average temperature anomalies and through identifying the driving mechanisms behind these changes. As a result, we come to understand that the Y-D cooling in the Northern Hemisphere resulted in an overall global cooling even though the Southern Hemisphere warmed up, and that changes in ocean circulation were likely a driving mechanism.

Although an in-depth review is out of the scope of this paper, a brief discussion of Y-D event impacts on humans and wildlife will be presented, to give a quick insight on how rapid climate change has affected ecology and humanity in the past. Firstly, the return to cold conditions had an impact on humans throughout North America, Europe, and Asia. In North America, it's been estimated that the onset of the Y-D event resulted in a significant decline or reorganization in human population (Anderson et al., 2011). A resurgence in population and settlements was observed at the end of the Y-D event, these population proxies being determined by increases in projectile points and quarry usage (Anderson et al., 2011). In Europe, archaeologists identified various cultural re-organizations dating to have occurred around the onset of the Y-D event, some examples being the re-establishment of reindeer hunting in central Europe, and the end of certain art/burial traditions in the Iberian Peninsula (Weber et al., 2011; Aura et al., 2011). In Asia, Moor and Hillman assert that the stresses due to the climate change during the Y-D event helped force a transition from hunting and gathering to farming (Moore and Hillman, 1992). This period happened to coincide also with a large extinction of ice-age megafauna – in North America, between 13,800 years to 11,400 years BP, up to 35 genera of mammals went extinct (Faith and Surovell, 2009). However, this was also a period when human hunters existed in North America, and other studies suggest that the decline of megafauna predates the onset of the Y-D event (Faith and Surovell, 2009; Gill et al, 2009). Regardless, the die-offs of all this megafauna could have potentially resulted in a notable impact on the atmosphere, namely a large reduction in methane (Smith et al., 2010). Although this methane decrease was not responsible for the whole decrease in methane recorded during the Y-D event, it could have played a contribution (Smith et al., 2010; Brook and Severinghaus, 2011). These megafaunal extinctions and changes in human activity are particularly significant in the study of the Y-D event as they are part of the evidence presented by some who believe that an extraterrestrial impact caused the onset of the Y-D event (Firestone et al., 2007). This claim that a bolide impact caused the extinction of megafauna, the disappearance of the Clovis (an early North American people) culture, and the Y-D event has been the topic of much debate and popular media attention (Kunzig, 2013). The flurry of discussion behind identifying the cause of this very significant climate event goes to demonstrate the importance in understanding the underlying mechanisms behind the Y-D event. As a result, the remainder of this paper will be focused on identifying the mechanisms behind the Y-D event.

Mechanisms: What caused the Younger Dryas?

Generally, there are three primary causes of forcing behind the onset of the Y-D event that have been discussed by scientists: ocean circulation, atmospheric circulation, and solar radiation (Fiedel, 2011). After a deeper discussion of the predominantly followed ocean-based theory, an investigation of atmospheric and solar radiation forcing factors will be conducted. Afterwards, some more novel theories will be addressed, demonstrating that identifying the trigger of this dramatic and abrupt climatic event is still a topic of much discussion.

The Younger Dryas and Ocean Circulation

Presently, it is generally accepted that ocean reorganizations along with sea-ice formation is a major factor in the onset of the Y-D event. More specifically, it involves the slowing of the Atlantic Meridional Overturning Circulation (AMOC). This connection between the switching of ocean circulatory modes and the Y-D event was discussed by Broecker and Denton in 1989, where they used the Y-D event as the “smoking gun” in identifying two modes of ocean circulation (Broecker and Denton, 1989). Per Broecker and Denton, the circulation of today is driven by the “Atlantic Conveyor”, which consists of salty deep water in the north Atlantic (NADW) that flows around Africa, then through the southern Indian Ocean, and then northward through the Pacific (illustrated in Figure 14). This circulation is important for local climates; during the winters when NADW forms, the heat released from the sea is comparable to solar forcing as a part of the northern Atlantic heat budget. A local result of this heat is that it goes to northern Europe – if it weren't for it, the winter temperatures in Europe would be lower (Broecker and Denton, 1989). In fact, models demonstrate that a cold North Atlantic sea surface temperature would result in a cooling of western and central Europe, which correlates with much of the Y-D evidence described earlier (Rind et al., 1986). Manabe and Stouffer had used a coupled sea-air model in 1988 to show that the Atlantic Ocean could have two different modes: one that is similar to today's conveyor belt, and the other with no conveyor circulation (Manabe and Stouffer, 1988). Manabe and Stouffer, like Broecker and Denton, suggest that this second mode where the outflow of NADW is weaker is characteristic in glacial times, including during the Y-D event (Manabe and Stouffer, 1988). From these studies, the onset of the Y-D event seems to be connected with a slowdown or shutdown of the NADW formation, which is attributed a reduction of surface water salinity in the North Atlantic. This is also further supported by all the various cooling trends detailed earlier. Additionally, per Broecker, this shutdown of NADW production plays a part in the "bipolar seesaw", where the Southern Hemisphere warms while the Northern Hemisphere cools (Broecker, 1998). Although there is less certainty when compared with the north, it seems like Southern Hemisphere temperatures were indeed generally warmer during the Y-D event, as discussed earlier. Furthermore, this slowdown/shutdown of the AMOC followed by a cooling of the North Atlantic explains other phenomena observed, such as the southward migration of the ITCZ as mentioned earlier when describing the effects of the Y-D event in the tropics (Schneider et al., 2014). On the other side of the Y-D timeline, it would be reasonable to assume that the Y-D termination resulted in a restart of NADW formation, and studies quickly confirmed that this was likely the case. Fawcett et al. is an example of this: using two general circulation models and then comparing those results with Greenland ice core data, they demonstrated that when the deep water production is turned back on, the warming associated with the end of the Y-D event is observed (Fawcett et al., 1997). Therefore, through the comparison between model results and paleoclimatic data, it can be seen that the Y-D event was probably the result of a shutdown and subsequent restart of NADW production. This idea has been around since the early 1980s, and was particularly championed by Broecker (Alley, 2007). Initially there was some questioning of this theory, as some cited a shutdown of NADW production during warm periods as evidence that it does not lead to cooling (Berger, 1990). Alternate mechanisms involving deep ocean mixing were also proposed to challenge Broecker's AMOC weakening theory (Wunsch, 2005). Regardless, more recent reviews seem to solidify the consensus around a slowdown/stoppage of NADW production through measurable evidence and modelling results (Alley, 2007).

Outside of model results and correlations with Greenland ice core data, there is also oceanic evidence that further strengthens the validity of a change in the AMOC occurring at the onset of the Y-D event. As early as 1987, sediment cores extracted from the Northern Atlantic containing foraminifera shells have been shown to carry proxies that indicate a reduced NADW flux during the Y-D event in which surface temperatures cooled (Boyle and Keigwin, 1987). This is demonstrated by higher Cd/Ca and lower $^{13}C/^{12}C$ ratios, which indicate a depletion of nutrients in intermediate waters which are typically influenced by NADW formation (Boyle and Keigwin, 1987). In 2004, McManus et al. conducted a study of a ^{231}Pa and ^{230}Th ratios in a sediment core from the subtropical North Atlantic Ocean (can be seen in Figure 10) and determined that around the onset of the Y-D event, the AMOC declined sharply in strength (McManus et al., 2004). The two isotopes are created when dissolved ^{235}U and ^{234}U undergo radioactive decay. Because of different residence times, which in turn results in different removal rates by from the tropics by the overturning circulation, the ratio between the isotopes acts as a proxy for AMOC strength (McManus et al., 2004). They notice a large ratio increase at 12,700 years BP, correlating it with the onset of the Y-D event, also noting how this increase did not last long, which hints at a gradual re-invigoration of the AMOC which potentially corresponds with the bipartition observed by some studies (McManus et al., 2004). Böhm et al. also used Pa and Th isotope ratios, and determined that NADW production was weakened, but not fully shut off as compared to other stadial events (Böhm et al., 2015). This results in a "cold" mode in the AMOC that is not quite fully off, as illustrated with Figure 15 (Böhm et al., 2015). These changes in ocean currents have also been observed through Nd isotopic measurements, including a study by Pahnke et al. which demonstrated an increased presence of Antarctic Intermediate Water in the North Atlantic, which is associated with a weakening in NADW production (Roberts et al., 2010; Pahnke et al., 2008). This weakening of NADW production also explains the sudden increase in ^{14}C observed globally, for when the AMOC stops/slows, less CO_2 would have moved from the ocean surface into deeper water, resulting in the buildup observed in the atmosphere (Fiedel, 2011). This release of CO_2 from deep water can be observed in sediment cores off the coast of California, where a drop in carbon uptake was observed right at the onset of the Y-D event (Marchitto et al., 2007). Similar depletion of CO_2 can be observed in cores made in the Southern Ocean (Anderson et al., 2009). In all, many oceanic effects that would have resulted from a slowdown of the AMOC can be observed and dated to the onset of the Y-D event, further cementing the AMOC as the main driver behind it.

Although it generally is of consensus that the Y-D event occurred due to a slowing of the AMOC, there is much debate as to what triggered this ocean current reorganization. When Broecker and Denton suggested the two modes in circulation caused the Y-D event, they said it was potentially due to a substantial input of freshwater stemming from a rerouting of the drainage basin of a meltwater lake called Lake Agassiz (Broecker and Denton, 1989). This mechanism for a trigger was proposed by Rooth in 1982, when he suggested that a diversion of drainage from the Mississippi River to the St. Lawrence River would seriously alter the circulation within the North Atlantic and subsequently cause the Y-D event (Rooth, 1982). This diversion (illustrated amongst other proposed routes in Figure 16) would disrupt the ocean circulation by allowing a serious influx of fresh water to the North Atlantic (Broecker, 2006). Before the Y-D period, Lake Agassiz flowed southwards through the Mississippi River into the Gulf of Mexico. As the southern Laurentide Ice Sheet margin retreated northwards, the northern and eastern shorelines of Lake Agassiz had a sudden drop in elevation, which allowed outflow of the lake to go east through Lake Superior and subsequently into the North Atlantic through the St. Lawrence Lowlands, per Broecker and Denton's initial proposal (Broecker, 2006). This drop was evidenced by sediment cores from the southern outlet of Lake Agassiz, which indicated a disruption in meltwater delivery at around the onset of the Y-D event (Fisher, 2006). Upon reaching the North Atlantic, the surge of fresh glacial meltwater would cause the surface water salinity of that region, and thus the density, to drop. This drop is characteristic of the shutdown/slowdown of NADW formation as described earlier. From this logic, it seems that if there is evidence of a shift in drainage flow from Lake Agassiz right before the Y-D event was observed, this mechanism could be potentially be viable. Broecker et al. first presented evidence of such a shift in 1989, when they presented cores from the Gulf of Mexico containing planktonic foraminifera which indicated that there was a significant drop in meltwater flow from the Mississippi River to the Gulf of Mexico (Broecker et al., 1989). Prior to this study, it had already been established that a drainage of Lake Agassiz had occurred, as the lake level was observed to have dropped by over 40 meters during the Y-D event (Teller and Thorleifson, 1983 as cited in Broecker et al., 1989). Broecker et al. identified the reduction of meltwater flow into the Gulf of Mexico in the form of a drop in ^{18}O-depleted meltwater that was recorded in the planktonic foraminifera (i.e., there was a rise in ^{18}O recorded in the foraminifera) (Broecker et al., 1989). Later evidence shows that there was potentially an outflow of Lake Agassiz through the Champlain Sea, which leads to the North Atlantic (Brand and McCarthy, 2005). They observed a drop in ^{18}O in shells taken from south of Ottawa and attributed it to freshening from additional glacial meltwater (Brand and McCarthy, 2005). However, per Broecker, their timing is poorly constrained, so their observed drop may have occurred after the onset of the Y-D event (Broecker, 2006). Furthermore, Broecker et al. later acknowledge that the drop in meltwater observed in the Gulf of Mexico may not have arose from a diversion meltwater flow, but rather a reduction of ice sheet melting during the Y-D event, as it did observe cooler climates in the Northern Hemisphere (Broecker et al., 2010). Therefore, it seems although there is strong evidence that meltwater flow to the Gulf of Mexico dropped around the onset of the Y-D event, this flow may not have been diverted directly to the North Atlantic through the St. Lawrence.

There are further studies that demonstrate how a sudden flood through the St. Lawrence River was unlikely right before the onset of the Y-D event. An earlier study of sediment cores done by de Vernal et al. found that during the Y-D event, there was an increase in salinity at the mouth of the St. Lawrence pathway, which hints at a reduction of meltwater from the Laurentide Ice Sheet (de Vernal et al., 1996). Therefore, from this study, it seems like there is no eastward flow of meltwater during or just prior to the Y-D event, or at least not enough to cause its onset (de Vernal et al., 1996). A study involving determining the paleontography of the region using various isostatic rebound models further suggested that an overflow to the east did occur, but not at the right time (Teller et al., 2005). They cited a failure to find coarse flood deposits associated with large overflows as their main reason as to why the Lake Agassiz overflow was unlikely to have been routed through the Great Lakes during the Y-D event (Teller et al., 2005). Likewise, Lowell et al. also concluded from surveys that there is no evidence of a massive discharge prior to the end of the Y-D event through either the eastern or northwestern outlets (Lowell et al., 2005). This topic is still a subject of debate as in 2007, Carlson et al. suggested that there was indeed a freshwater influx into the St Lawrence River that caused a slowdown of the overturning circulation (Carlson et al., 2007). They used new geochemical proxies (ΔMg/Ca, U/Ca, and $^{87}Sr/^{86}Sr$) taken from sediment cores at the mouth of the St. Lawrence to determine the changes in freshwater influx (Carlson et al., 2007). However, their results did not explain the lack of an outlet in the Great Lakes, nor do they agree with the earlier results of de Vernal et al., and as a result some disagreements with their method of acquiring a proxy arose (Peltier et al., 2008). More recent evidence has made this argument for an eastern rerouting more compelling, and it cannot be completely written off. In 2015, Levac et al. took new marine sediment cores from the Laurentian Channel and through a strong evidence of lowered salinity at the time of the Y-D onset, re-established the possibility of the St. Lawrence being a possible outlet (Levac et al., 2015). In 2018, Leydet et al. used ^{10}Be to date surface exposure ages on the eastern edge of Lake Aggasiz to determine if there was an eastern outlet available (Leydet et al., 2018). They found that the main eastern channel opened up between 13,000 and 12,700 years BP, which makes it possible for an eastern drainage through the St. Lawrence (Leydet et al., 2018). However, as

Fisher notes in his 2020 review on possible outlets, there is still no evidence to explain a 90 m drop in lake level from a southern or eastern outlet perspective, which occurred at the onset of the Y-D event (Fisher, 2020). Levac et al. also acknowledge that while some drainage may occur through the St. Lawrence, there is still possibility of drainage through other outlets (Levac et al., 2015). Therefore, although there is a possibility of drainage through the eastern end of Lake Agassiz, it is uncertain if this drainage consists of the initial pulse that triggered the Y-D event.

Outside of the St. Lawrence re-route, there are other possibilities of a Lake Agassiz drainage triggering the Y-D event. From a coupled ocean-atmosphere model, Manabe and Stouffer suggested that meltwater pulse 1a (MWP1A) entered the Gulf of Mexico and was advected into the North Atlantic by the Gulf Stream (Manabe and Stouffer, 1997). MWP1A is dated to have occurred around 14,000 years ago (Fairbanks, 1989). Manabe and Stouffer found that this influx of freshness into the Gulf of Mexico would affect the AMOC and noted how it resembles the Y-D event (Manabe and Stouffer, 1997). This implies that rather than a rerouting, a freshwater pulse went directly into the Gulf of Mexico down the same drainage path that Lake Agassiz had utilized at the time. They however note that the magnitude of the thermohaline response is much smaller than if the freshwater was directly discharged into the North Atlantic (Manabe and Stouffer, 1997). Furthermore, the proxy results of MWP1A from McManus et al. indicate no significant effect on the MOC, and they also pointed out that it occurred too early to directly cause the Y-D event (McManus et al., 2004). Additionally, there was no similar pulse during the onset of the Y-D event, as the earlier listed observations demonstrated a drop in meltwater flow (Broecker et al., 1989). Tarasov and Peltier also point out that due to the Mississippi being a sediment-laden river and the Gulf Stream being baroclinically unstable, it is unlikely that the meltwater discharged into the Gulf of Mexico would get advected to the North Atlantic, so a more direct North Atlantic forcing would be required to cause the onset of the Y-D event (Tarasov and Peltier, 2005). Like with the St. Lawrence, Fisher notes that the southern outlet of Lake Agassiz into the Mississippi was unable to accommodate the necessary drop in lake level (Fisher, 2020). Therefore, MWP1A or a similar event was unlikely to have caused the Y-D event.

Contrary to Lake Agassiz emptying through the St. Lawrence, there is also a possibility that a drainage into the Artic Ocean is what triggered the onset of the Y-D event. This mechanism is proposed by Tarasov and Peltier to be the most likely cause for the Y-D event (Tarasov and Peltier, 2005). From their glacial systems model results, they determined that the largest freshwater discharge was directed into the Arctic Ocean through the Fram Strait into the Greenland-Iceland-Norwegian Seas (shown in Figure 16) (Tarasov and Peltier, 2005). They proposed that the Keewatin ice dome (also shown in Figure 16) had a sudden melting and acted as a source of the discharge Tarasov and Peltier, 2005). Using a modern coupled atmosphere-ocean sea ice-land surface processes model (the NCAR CSM1.4 Model), Peltier et al. then showed that the drainage of freshwater into the Arctic Ocean would cause a response that is essentially identical to that of a freshening of the North Atlantic (Peltier et al., 2006). They also point out that once the Arctic freshwater forcing ended, the oceanic circulation began its recovery, signalling the termination of the Y-D event (Peltier et al., 2006). There does not seem to be a detailed chronology yet of the Keewatin ice dome (Tarasov and Peltier, 2005), but there exists evidence of a freshwater pulse coming from Lake Agassiz and emptying to the north. This proposed mechanism is supported by paleoclimatic proxy records from the Arctic regions. From the Mendeleyev Ridge in the west Arctic Ocean, a rise in Ba/Ca ratios in planktonic foraminifera were dated to have occurred at the onset of the Y-D event, indicating an influx of freshness (Hall and Chan, 2004). To further strengthen this proposed mechanism of a northern outlet for Lake Agassiz, a study done by Murton et al. seems to identify a flooding event along the Mackenzie River entering the Artic Ocean at around the time of the onset of the Y-D event (Murton et al., 2010). They identify this flood path through gravels and regionally eroded surfaces throughout the Mackenzie River system and dated them to be attributed to two floods, one about 13,000 years ago and the other 9,300 years ago (Murton et al., 2010). This first event is likely to be associated with the onset of the Y-D event which occurred shortly afterwards. Further sediment evidence in the Chukchi margin in the Arctic Ocean shows a minimum in $\delta^{18}O$, which is dated to have occurred right before the onset of the Y-D event and approximates the modeled peak of Laurentide meltwater discharge suggested by Tarasov and Peltier (Polyak et al., 2007). More recently, using an ocean sea-ice model, Condron and Winsor showed that meltwater discharge from the Artic would reduce the AMOC strength by over 30%, whereas a discharge through the St. Lawrence would only reduce it by less than 15%, thereby demonstrating the viability of Tarasov and Peltier's mechanism further (Condron and Winsor, 2012). In his review of possible outlets, Fisher also agrees that only the northwest outlet draining into the Artic Ocean supports the drop in lake level in Lake Agassiz at the onset of the Y-D event (Fisher, 2020). In all, the modelled results, the paleoclimatic proxy evidence of a freshwater discharge throughout the Mackenzie River delta, and the evidence of a meltwater pulse at various spots in the Artic Ocean, when combined with the droppage of meltwater flow to the Gulf of Mexico, all point towards a northern redirection of Lake Agassiz drainage to be a trigger for the onset of the Y-D event.

Continuing the thread on a northern flux of freshwater, yet another trigger that can shut down the AMOC a large ice-rafting event. A discharge of icebergs from the north into the Arctic ocean was suggested to be part of the meltwater discharge mechanism proposed by Tarasov and Peltier (Tarasov and Peltier, 2005). It has been proposed that large iceberg rafting events, called Heinrich events, have triggered similar rapid changes in climate (Timmerman et al., 2003). Tarasov and Peltier refer to an older climate oscillation, the Preboreal Oscillation, as an example of an event being triggered by a Heinrich-event like introduction of pack ice through the Fram Strait (Tarasov and Peltier, 2005). Sediment cores done by Bauch et al. and Norgaard-Pedersen et al. seem to support such an event near the onset of the Y-D event (Bauch et al., 2001; Nørgaard-Pedersen et al., 2003). Bradley and England go farther with this idea about ice being the main forcing agent, proposing a "sea of ancient ice", or paleocrystic ice that was driven from the Arctic Ocean into the Greenland Sea due to rising sea levels during the Bølling-Allerød (Bradley and England, 2008). Although not quite "paleocrystic" sea ice, the freshwater outflow suggested by Tarasov and Peltier allows for the freshwater to be delivered to the Greenland Sea through the form of pack ice in a manner similar to that proposed by Bradley and England (Tarasov and Peltier, 2005). Further freshwater forcing from ice could have stemmed from the Fennoscandian Ice Sheet; Muschitiello et al. suggest that the melting ice sheet could explain the hydroclimate observed at the onset of the Y-D event in Northern Europe and Greenland, and also suggest that it can cause an entrance into stadial conditions in general (Muschitiello et al., 2015). Like Peltier et al., Muschitiello et al. demonstrate that a freshening of the Nordic Seas can weaken the AMOC (Muschitiello et al., 2015). However, they acknowledge that further studies are required to demonstrate that a rapid export of freshness stemming from the Fennoscandian Ice Sheet occurred at the onset of the Y-D event (Muschitiello et al., 2015). Therefore, given the current evidence and modelling results, it can be seen that ice can certainly affect the AMOC, though it is most likely just a medium for which freshwater forcing due to a rerouting of Lake Agassiz drainage can be applied to the Greenland-Iceland-Norwegian Seas.

The Younger Dryas and Atmospheric Circulation

In their 2007 review, Seager and Battisti claimed that although the Y-D event almost certainly involved changes in the North Atlantic Ocean circulation, those changes do not account for all the global changes associated with events like the Y-D event (Seager and Battisti, 2007). One striking example is how models do not seem to be able to cause observed cooling in South America in the Andes, nor cooling in the southeast Atlantic (Seager and Battisti, 2007). As a result, Seager and Battisti suggest additional changes that needed to occur within atmospheric circulation regimes. In their proposal, they illustrate two modes in atmospheric circulation (see Figure 17) that correspond with the two modes of ocean circulation described in the previous section and ultimately imply that atmospheric changes in the tropics would cause the subsequent oceanic circulation changes in the North Atlantic (Seager and Battisti, 2007). They suggest that a switch to zonal flow across the Atlantic would cause the temperatures in Europe to drop, along with causing the change in North Atlantic circulation that has been observed by others (Seager and Battisti, 2007). This theory can be accompanied with some of the evidence seen in Europe; as discussed earlier, Brauer et al. observed an increase in an increase in windiness in Europe which is resultant of such a switch to a zonal jet (Brauer et al., 2008). Brauer et al. also note that AMOC changes cannot be wholly responsible for these observed changes, so they also suggest that the wind shift is part of the mechanism that caused the Y-D event (Brauer et al., 2008). For the termination of the Y-D event, this wind effect is also suggested as a mechanism. Bakke et al. noted that the westerlies had rapid alternations between drifting more north or south in a feedback cycle with the North Atlantic, which resulted in oscillations between warming or cooling in Northern Europe (Bakke et al., 2009). These rapid oscillations were noted to occur right before the termination of the Y-D event (Bakke et al., 2009). Steffensen et al. noted that prior to dramatic termination of the Y-D event, a sudden atmospheric shift also occurred in Greenland (Steffensen et al., 2008). Therefore, from these studies, it may seem like large wind pattern changes caused the shutoff of circulation in the North Atlantic.

It appears from these studies that although atmospheric effects may very well influence the onset and termination of the Y-D event, it still heavily depends on circulation within the ocean. For example, in Steffensen et al., although it seems like atmospheric changes led the terminating warming in Greenland, these atmospheric reorganizations occurred from Southern Hemisphere/tropical warming (Steffensen et al., 2008). Cheng et al. also suggest that the termination for the Y-D event was triggered by changes in the tropics or the Southern Hemisphere (Cheng et al., 2020). As described earlier, the warming in the Southern hemisphere is a result of the AMOC slowdown/shutdown due to the bipolar seesaw nature of the thermohaline circulation. Therefore, these atmospheric reorganizes seem to be dependent on the initial ocean circulation shift. This creates a “chicken and egg” problem as the two triggers seem dependent on one another based off the conclusions of various studies (Fiedel, 2011). One factor behind this ambiguity is the inaccuracies associated with radiocarbon dating, as it is difficult to tell which changes occurred before the other as uncertainties can range as high as centuries (Pearce et al., 2013). Pearce et al. however refer to a single sedimentary archive when reconstructing their multiproxy paleoclimatic record, and as a result can confidently present a sequence of events without worrying about potential age offsets between events (Pearce et al., 2013). They suggest that the termination starts with a gradual decrease in the Labrador Current influence (stemming from a slow warming of the North Atlantic), which is then followed by a northern shift of the Gulf Stream, which then sets the decline of the sea-ice cover into action, in all emphasizing the fact that ocean circulation changes occur first (Pearce et al., 2013). Broecker had also argued that for the onset of the Y-D event, the global atmospheric changes must have stemmed from perturbations in tropical dynamics which in turn were the result of oceanic changes (Broecker, 2003). A tropical atmosphere-ocean interaction trigger has been put forward as a potential mechanism for abrupt events similar to the Y-D event, namely Heinrich events and Dansgaard-Oeschger oscillations (Broecker, 2003). Broecker's key point is that the Y-D event onset was quite synchronous globally, and that the tropical trigger needs to either attribute this synchroneity to either be of coincidence or to another climate change trigger (Broecker, 2003). In all, considering the evidence that seem to indicate that the atmospheric changes depend on oceanic ones, it appears that although atmospheric changes are instrumental in the onset and termination of the Y-D event, they are not the primary trigger.

The Younger Dryas and Solar Forcing

In 1999, van Geel et al. proposed that solar variability played a large role in the climate shifts that occurred in the Late Pleistocene, including the Y-D event (van Geel et al., 1999). Two potential solar mechanisms for causing climate change were proposed by van Geel and Renssen in 1998 (van Geel and Renssen, 1998). The first possibility is that reduced solar activity (reductions in UV radiation, which causes increases in atmospheric ^{14}C) can reduce stratospheric ozone content, which can in turn relocate the mid latitude storm tracks in both hemispheres and decrease the latitudinal extent of Hadley cells (van Geel and Renssen, 1998). This atmospheric change would in turn cause cooling in both hemispheres and a shift in the precipitation belts (the latter effect indeed being observed) in addition to perturbing the thermohaline circulation (van Geel and Renssen, 1998). A second potential mechanism, which is not exclusive of the first, is that an increase in cosmic ray flux (from solar wind variations) could increase cloud formation, which in turn reflects more incoming radiation, which cools the Earth as a whole (van Geel and Renssen, 1998). These mechanisms are illustrated in Figure 18. These solar forcing hypotheses are supported by the idea that variations of ^{14}C and ^{10}Be observed in ice cores and tree rings could only be explained by changes in solar variation (van Geel et al., 1999). They cite more recent climatic events such as the Maunder Minimum, which occurred from 1645 to 1715 and was recorded through ^{14}C and ^{10}Be variations, as examples of changes in solar variability causing abrupt climate change (van Geel et al., 1999). At the start of the Y-D event, there was a large increase in atmospheric ^{14}C that was assumed to be due to a reduction in ocean circulation causing a decrease in CO_2 exchange between the atmosphere and ocean (Björk et al., 1996). This change in atmospheric ^{14}C was one of the tools utilized to align climate records but has been proposed to also represent a solar forcing at the onset of the Y-D event (Renssen et al., 2000). Furthermore, Renssen et al. suggest that the Y-D event may have been a part of a ~2500-year quasi-cycle that is of solar origin (Renssen et al., 2000). Broecker et al. also acknowledge that the Y-D event could have been part of a natural cycle of sorts (though not necessarily due to solar forcing); they claim that the Y-D event was just a natural part of the last glacial termination, due to the past four terminations having similar cold reversals (Broecker et al., 2010). This would imply that a catastrophic trigger such as a flood was not necessarily needed (Broecker et al., 2010). Finally, as a third piece of evidence from Renssen et al., they propose that global cooling would not be explained by the ocean circulation mechanism (which generally indicates a "bipolar seesaw" as explained earlier), but a change in solar forcing would explain it, and cited potential evidence of cooling in the Southern Hemisphere to support their proposal (Renssen et al., 2000). However, as evidenced by later studies, it seems the Southern Hemisphere did indeed undergo a warming during the Y-D event, thus supporting the ocean circulation mechanism (e.g., Stenni et al., 2010). This uncertainty with Southern Hemisphere temperatures was acknowledged by Renssen et al., who also proposed that this solar forcing could be applied in parallel with oceanic mechanisms, and that more modelling needed to be done to test it (Renssen et al., 2000). The proposed solar minimum of Renssen et al. as a result does not seem to be the primary cause of the Y-D event, though it brings the interesting idea that the onset of the Y-D event could partially be the result of changes in radiative forcing.

Other potential triggers of the Younger Dryas

Continuing with the idea of a change in radiative forcing, recently a topic of much conversation, especially in the media, is the possibility that a cosmic impact caused the onset of the Y-D event. In 2007, Firestone et al. suggested that a potential extra-terrestrial impact caused the climatic changes that are observed during the Y-D event (Firestone et al., 2007). Their primary evidence comes in the form of a layer rich with carbon that has been dated to be 12,900 years BP, or just before the onset of observed Y-D effects. This layer, which is observed across many sites in North America, along with many potential impact sites, is used to suggest how a potential impact destabilized the Laurentide Ice Sheet to cause the flooding that subsequently led to the Y-D event (Firestone et al., 2007). They also lay claims that this impact caused the extinctions of North American megafauna and resulted in large changes within the movements of Paleoamericans (Firestone et al., 2007). This was through the shock wave and biomass burning associated with such a proposed event. The potential observations of impact material such as nanodiamonds dated to occur at the same time as Firestone et al.'s proposed impact helped attract attention to this proposal (Kennett et al., 2009). However, very quick rebuttals followed, with multiple studies failing to identify such markers of an extra-terrestrial impact (Surovell et al., 2009; Daulton et al., 2010). This has been followed by a constant back and forth between various groups, with certain groups uncovering new evidence and other groups refuting it (van Hoesel et al., 2014). Even presently, there are still new studies pushing discoveries that claim to prove the impact hypothesis (Moore et al., 2020). Moore et al. found yet another site with nanodiamonds, and as a result proposed that at the same time as the impact noted by Firestone et al., an impact also occurred in Syria, possibly due to the impactors being from a comet's debris stream (Moore et al., 2020). This would result in an airburst of multiple impactors, which explains how multiple sites have been dubbed a location for which an extra-terrestrial impact could have occurred to trigger the Y-D event (Firestone et al., 2007; Kjaer et al., 2018; Moore et al., 2020). Like the new studies giving proof to the extraterrestrial impact hypothesis, there are still new studies refuting this hypothesis as well (Jorgeson et al., 2020). Jorgeson et al. compiled ^{14}C measurements across all the sites that claimed evidence of a Y-D impact, and then analyzed the variability of the data set with a Monte Carlo simulation to determine if this proposed impact was a synchronous event (Jorgeson et al., 2020). When compared with ^{14}C measurements associated with a known synchronous event, the Laacher See volcanic eruption, Jorgeson et al. determined that the Y-D impact data is not likely to be synchronous, which undermines a central requirement of the impact hypothesis which needs the deposition of material to be a synchronous global event (Jorgeson et al., 2020). Nonetheless, Moore et al. note that there are still many studies that claim to refute and confirm this impact hypothesis, so it can be seen that this topic is still hotly debated (Moore et al., 2020). Furthermore, as this impact theory already utilizes the same freshwater disrupting NADW formation mechanism as the earlier flood proposal by Broecker et al., it already supports the oceanic mechanism for the Y-D event. Therefore, this impact, if it actually happened, would only serve as a pre-trigger of sorts – the main driver would still be a freshwater influx disrupting the AMOC. Until more extensive evidence can be found to address the holes in this hypothesis that are still being pointed out, it seems unlikely that an extraterrestrial impact was the trigger for the Y-D event.

On the topic of the Laacher See volcano mentioned earlier, some have proposed that its eruption could have actually triggered the Y-D event. This volcano, located in Germany, erupted around 12,900 years BP and was briefly mentioned as a possible trigger for the Y-D event early in the 1990s, but was brushed aside due to the existing and more popular meltwater pulse hypothesis that had already been proposed (Berger, 1990; Baldini et al., 2018). Using the latest data from the Greenland Ice Sheet Project ice core data set, Baldini et al. proposed that the Laacher See volcano erupted at the onset of the Y-D event (Baldini et al., 2018). Previously, this theory had been dismissed due to the chronology of the eruption and the Y-D event not aligning (Brauer, 2008). The volcano's signal is shown through a large spike in volcanic sulfur identified within the GISP2 ice core (Baldini et al., 2018). This signal is thought to be observed outside of Europe as well; in Hall's Cave in Texas, Sun et al. examined $^{187}Os/^{188}Os$ ratios in sediments, and determined them to be consistent with an extraterrestrial or volcanic event at the onset of the Y-D event (Sun et al., 2020). However, their records of concentrations of highly siderophile elements (which consist of Os, Ir, Ru, Pt, Pd, and Re abundances) demonstrate that the layer marking the event contains volcanic aerosols, and not extraterrestrial materials, thereby adding more evidence to the volcanic event hypothesis while weakening the validity of the extraterrestrial impact hypothesis (Sun et al., 2020). Baldini et al. hypothesize that this eruption would have brought many sulfates into the atmosphere, and lead to the initial cooling in the Y-D event (Baldini et al., 2018). However, they acknowledge that this aerosol related cooling probably lasted only 1-3 years and that most of the remaining cooling in during the Y-D event was due to sea-ice-ocean circulation pattern changes (Baldini et al., 2018). Baldini et al. also further stress the importance of modelling such an event to test their hypothesis (Baldini et al., 2018). In all, although the Laacher See volcano likely erupted sometime before or during the onset of the Y-D event, the main driver of the cooling in the Northern Hemisphere still seems to be changes in ocean circulation.

Summary of Mechanisms

In summary, due to model evidence and correlations with many paleoclimatic proxies, a slowdown of the AMOC seems to be the most viable mechanism driving the Y-D climate reversal. The AMOC slowdown has been shown through models to cause the resulting Northern Hemisphere cooldown, Southern Hemisphere warm up, and ITCZ displacement, which have all been observed around the globe. Furthermore, there is increasing evidence pointing to a re-routing of Lake Agassiz drainage being the trigger for this slowdown, as it allows for a large influx of freshwater into the North Atlantic to slow NADW production and thus the AMOC. This rerouting has been generally identified to be through the Mackenzie River to the north of Lake Agassiz, with pack ice being a method by which the freshwater is delivered to the Greenland-Iceland-Norwegian Seas, though there are still studies pointing at a possible routing through the originally proposed St. Lawrence. Although this seems to be a definitive mechanism, there are still inconsistencies with certain models (Renssen et al., 2015). Additionally, there seem to be many points of evidence that both confirm or reject potential atmospheric mechanisms and solar forcing mechanisms (e.g., Seager and Battisti, 2007; Renssen et al., 2000). The catastrophic event triggers proposed by Firestone et al. and Baldini et al., while exciting, still need more evidence to be viable triggers. There is also the possibility that a combination of these mechanisms was involved in causing the Y-D event; Renssen et al. found that a combination of a weakened AMOC, moderate negative radiative forcing, and an altered atmospheric circulation best simulated existing proxy evidence (Renssen et al., 2015). They used an intermediate complexity climate model, the LOVECLIM1.2 global model, which has been accurate in reconstructing other paleoclimatic events, such as the last glacial maximum and an event 8,200 years ago (Goosse et al., 2010). This intermediate model has some setbacks, in that it has limited detail in representing atmospheric circulation, but Renssen et al. note that the model response in the extratropics to radiative and freshwater forcing is similar to that of other models, including general circulation models of higher complexity (Renssen et al., 2015). Figure 19 and Table 1 show the results of their modelling experiments, of which the COMBINED model matches with their reconstructed proxy data the best (Renssen et al., 2015). The experiment done by Renssen et al. however does not seem to address results in the Southern Hemisphere. Regardless, the idea that a combination of triggers caused this climate reversal is an interesting one and provides an opportunity for further experimentation with perhaps higher resolution models.

Conclusions

From this review, it can be seen that the Y-D event from about 12,900 years to 11,600 years BP had strong effects on the global climate. It cooled the Northern Hemisphere by a significant amount and resulted in a return to tundra-like conditions throughout Europe and North America. This can be seen through various paleoclimatic proxies that indicate a change in vegetation and a general advance in glaciers. The Y-D event weakened monsoons throughout the northern tropics through the southward migration of the ITCZ, which caused a cooling and aridity throughout areas in Central/South America, Asia, Africa, and the Pacific Ocean. This had various local effects, including an interruption to the African Humid period and a persistent El Niño. In the Southern Hemisphere, a general warming trend can be seen, supporting a “bipolar seesaw” mechanism driving the Y-D event. Studying the global effects of the Y-D event allows for a good exercise in determining the abilities of dating techniques and paleoclimate proxies, as demonstrated by the plethora of different efforts used to map and understand this global event. Furthermore, the Y-D event presents a great natural example in testing the rigour of climate models in trying to determine its underlying mechanisms/triggers. From the latter portion of this review, it can be seen that the most likely trigger/mechanism for the Y-D event was a freshwater influx from Lake Agassiz into the Greenland-Iceland-Norwegian Seas in the form of pack ice through the Fram Strait. This causes a subsequent weakening of the AMOC, which through various teleconnections, such as the southward migration of the ITCZ and a switch in westerlies to form a more zonal jet, carry the Y-D signal around the globe. Also discussed is the potential for other mechanisms to be a factor in driving the Younger Dryas, such as atmospheric shifts and changes in solar radiation. In all, these remaining uncertainties in understanding the mechanisms and effects of the Y-D event demonstrate the complexity of the event and of climate in general. This uncertainty presents further opportunities to learn more about our past climate (and thus our present) through combinations of paleoclimatic proxy, dating, and climate modelling techniques.

Figures

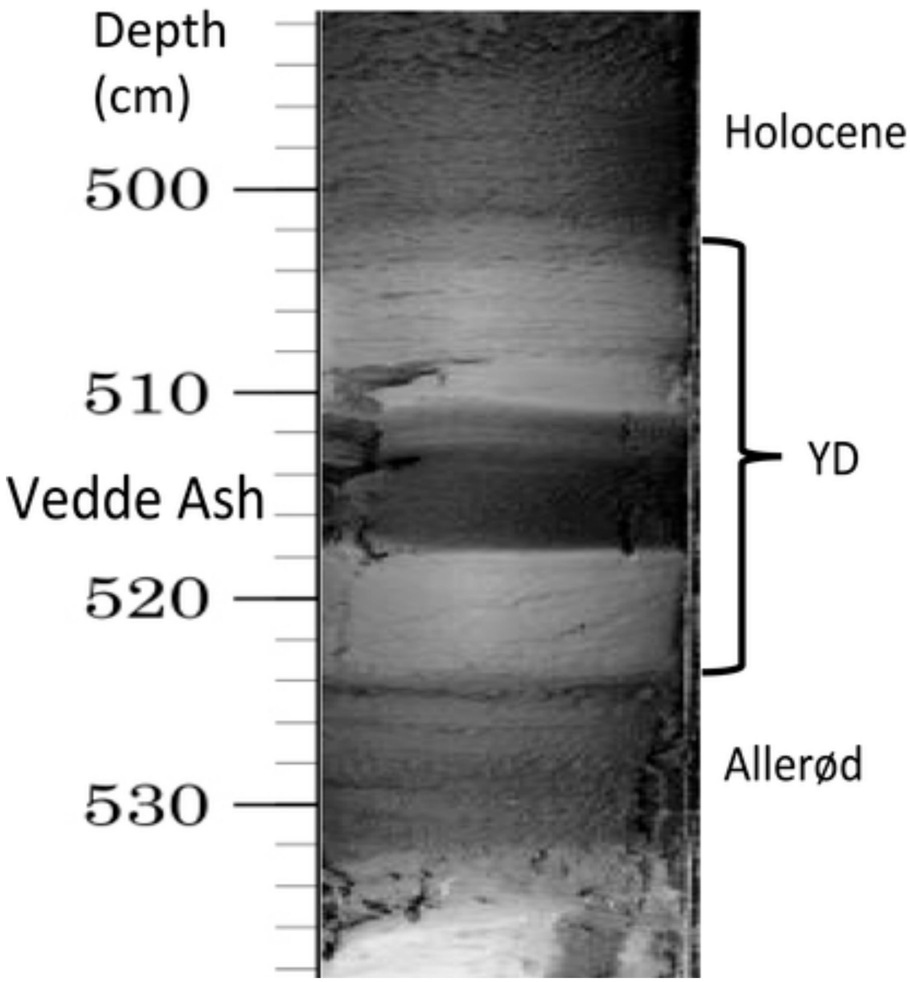

Figure 1: Lake core from western Norway (Photograph by Carl Regnéll, as cited in Mangerud, 2020). This photograph shows the Younger Dryas silt (which is more minerogenic) sandwiched between two *gyttja* with higher organic sediments. The higher amount of organic sediment in these layers (the Allerød and Holocene) correspond to the more temperate conditions that allow for more flora (Mangerud, 2020).

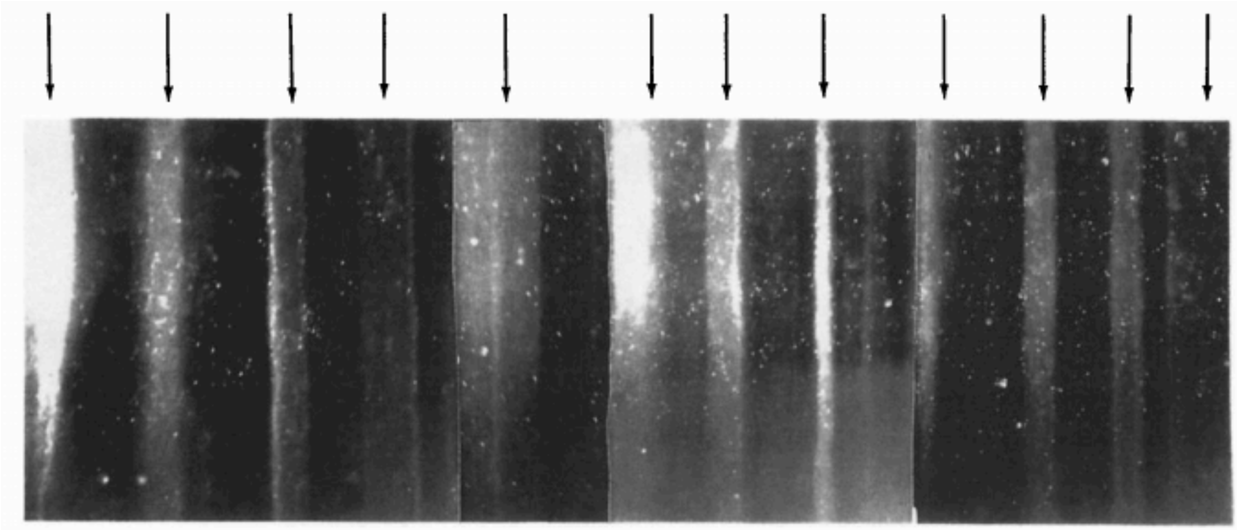

Figure 6. Photograph of a 19-cm-long section of ice from 1855 m showing annual stratigraphic layers in the Wisconsinan. This section contains 12 summer layers (arrowed) sandwiched between darker winter layers. The lighter layers are a result of increased scattering of light due to higher concentrations of dust.

Figure 2: An example of an ice core from Greenland containing layers representing summers and winters (Meese et al., 1997).

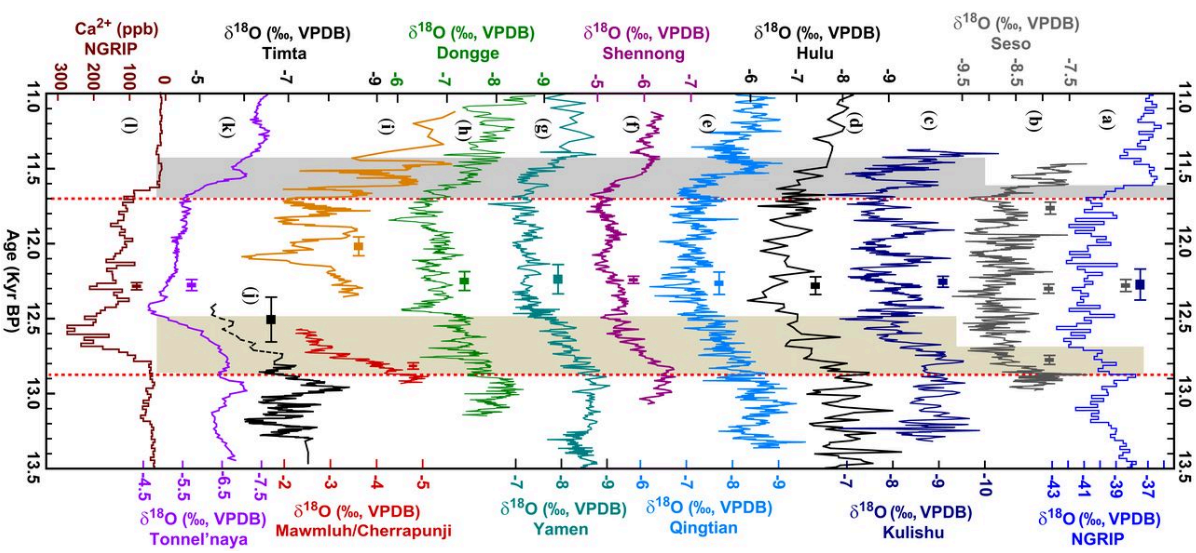

Figure 3a: Speleothem samples throughout Asia showing similar Younger Dryas oxygen isotope patterns, as compared to the NGRIP ice core and the Seso Cave sample (Cheng et al., 2020). Here it can be seen that there is a general trend of cooling during the Younger Dryas.

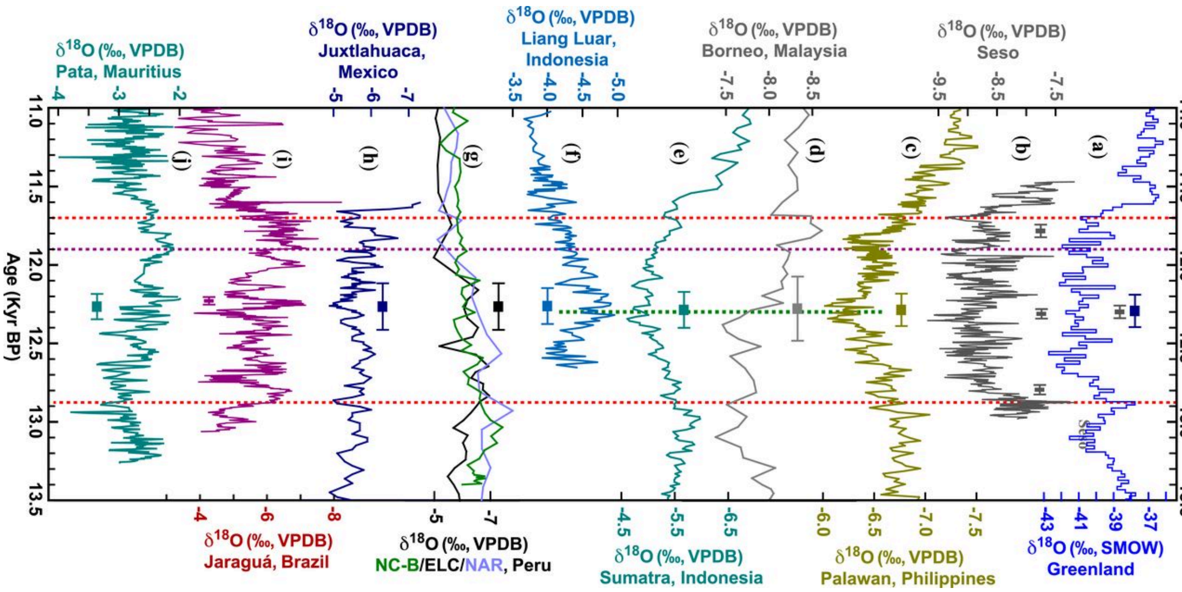

Figure 3b: Speleothem samples throughout the tropics showing similar Younger Dryas oxygen isotope patterns, as compared to the NGRIP ice core and the Seso Cave sample (Cheng et al., 2020). Here, the temperature trends are less certain, with some showing cooling, others showing warming, and others showing no discernable temperature trend during the Younger Dryas.

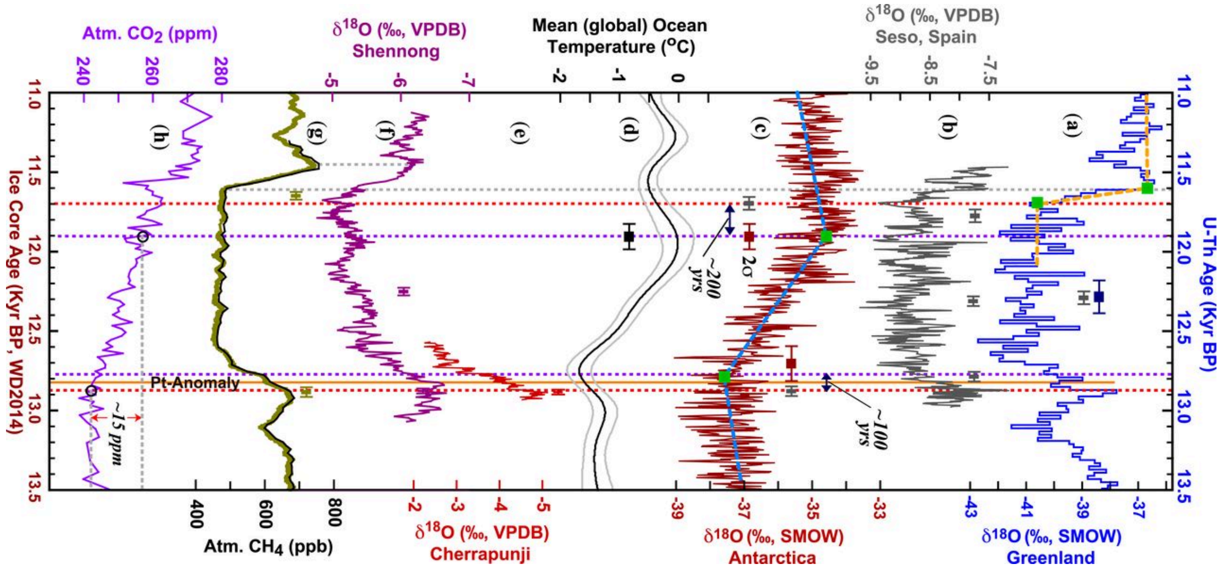

Figure 3c: Proxy data indicating differences in trends at the poles, along with atmospheric trends (Cheng et al., 2020). Here, it can be seen that Antarctica generally warmed while Greenland cooled. Additionally, there is a steady increase in atmospheric carbon dioxide and a large drop in atmospheric methane during the Younger Dryas.

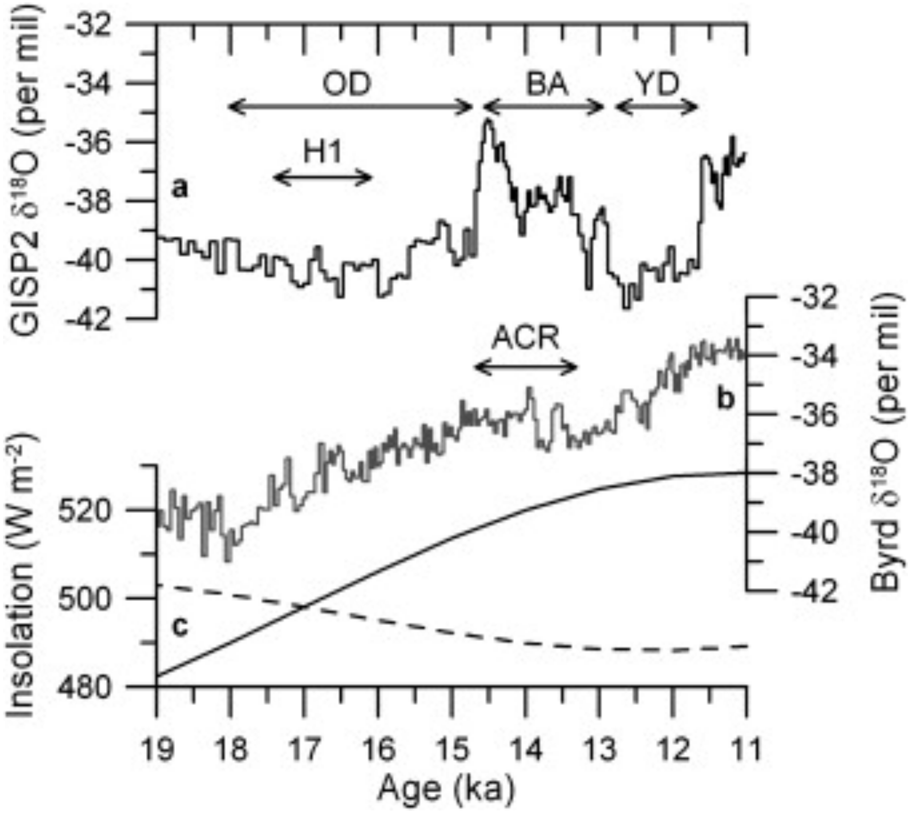

Figure 4: Greenland (GISP2) and Antarctic (Byrd) ice cores from 19,000-11,000 years BP (Shakun and Carlson, 2010). From the GISP2 core, one can identify the gradual cooling prior to the Younger Dryas, followed by an abrupt plunge into the cold period, which is in turn followed by an even more abrupt rise into modern climate conditions. The Antarctic core demonstrates a general warming trend in the Southern Hemisphere throughout the Younger Dryas

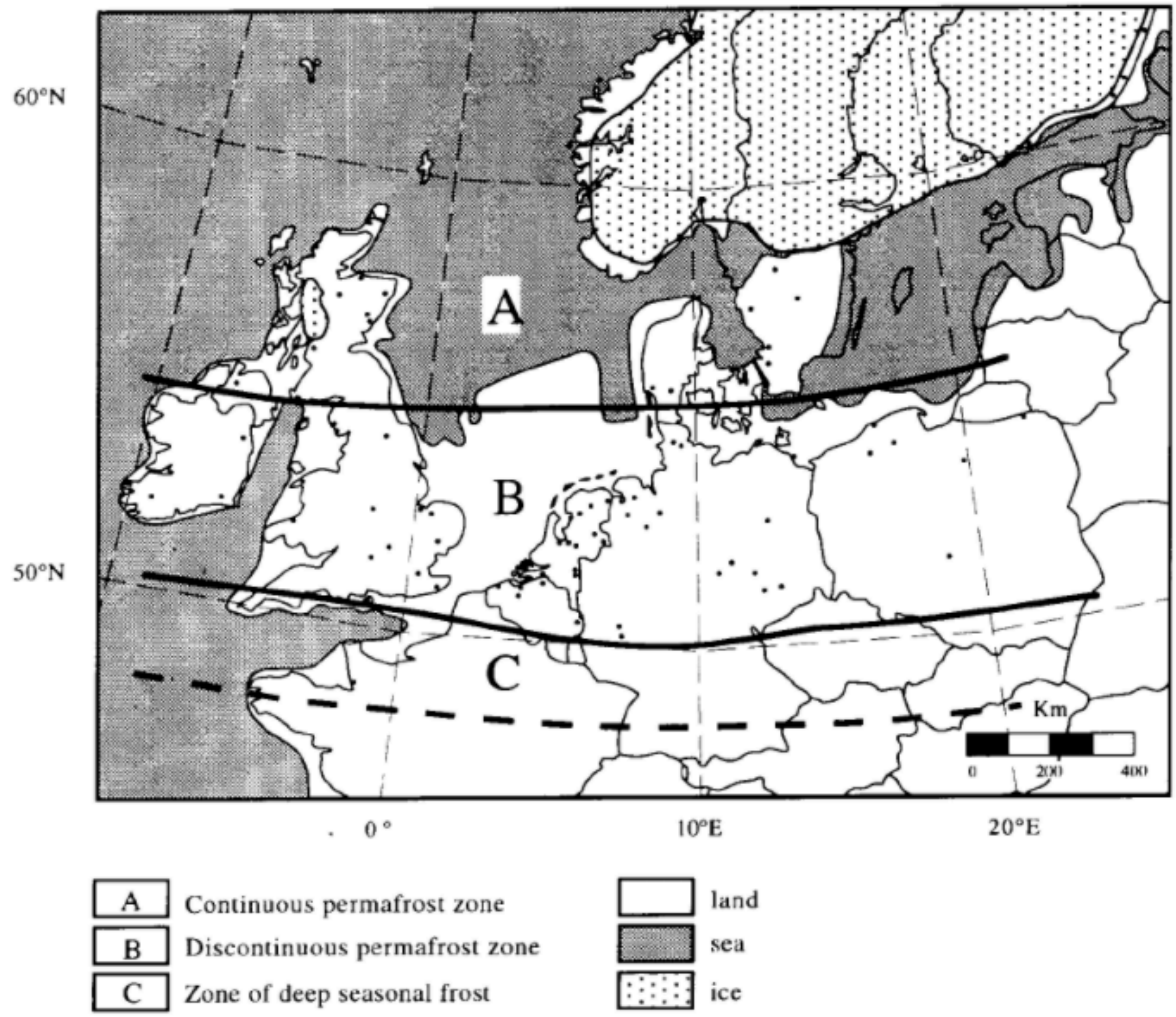

Figure 5: Distribution of permafrost in western Europe during the Younger Dryas based off periglacial phenomena (sites marked by black dots) (Isarin, 1997). Here it can be seen that the permafrost extent is a lot broader than today's climate.

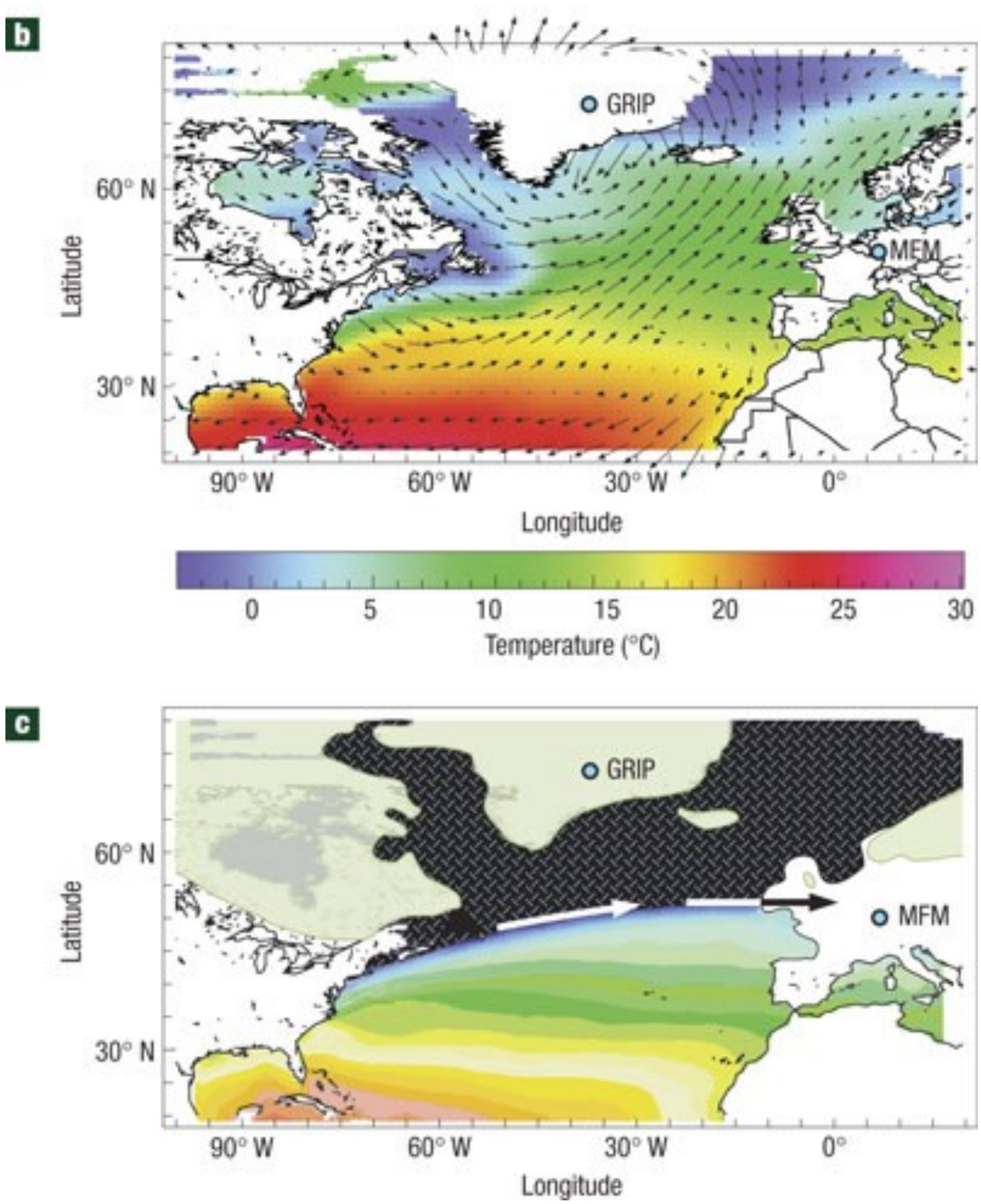

Figure 6: Modern day winter wind patterns (b) and Younger Dryas wind patterns (c) (Brauer et al., 2008). This figure illustrates how the extension of sea ice coverage forces the westerlies to form a zonal jet that goes directly through Europe.

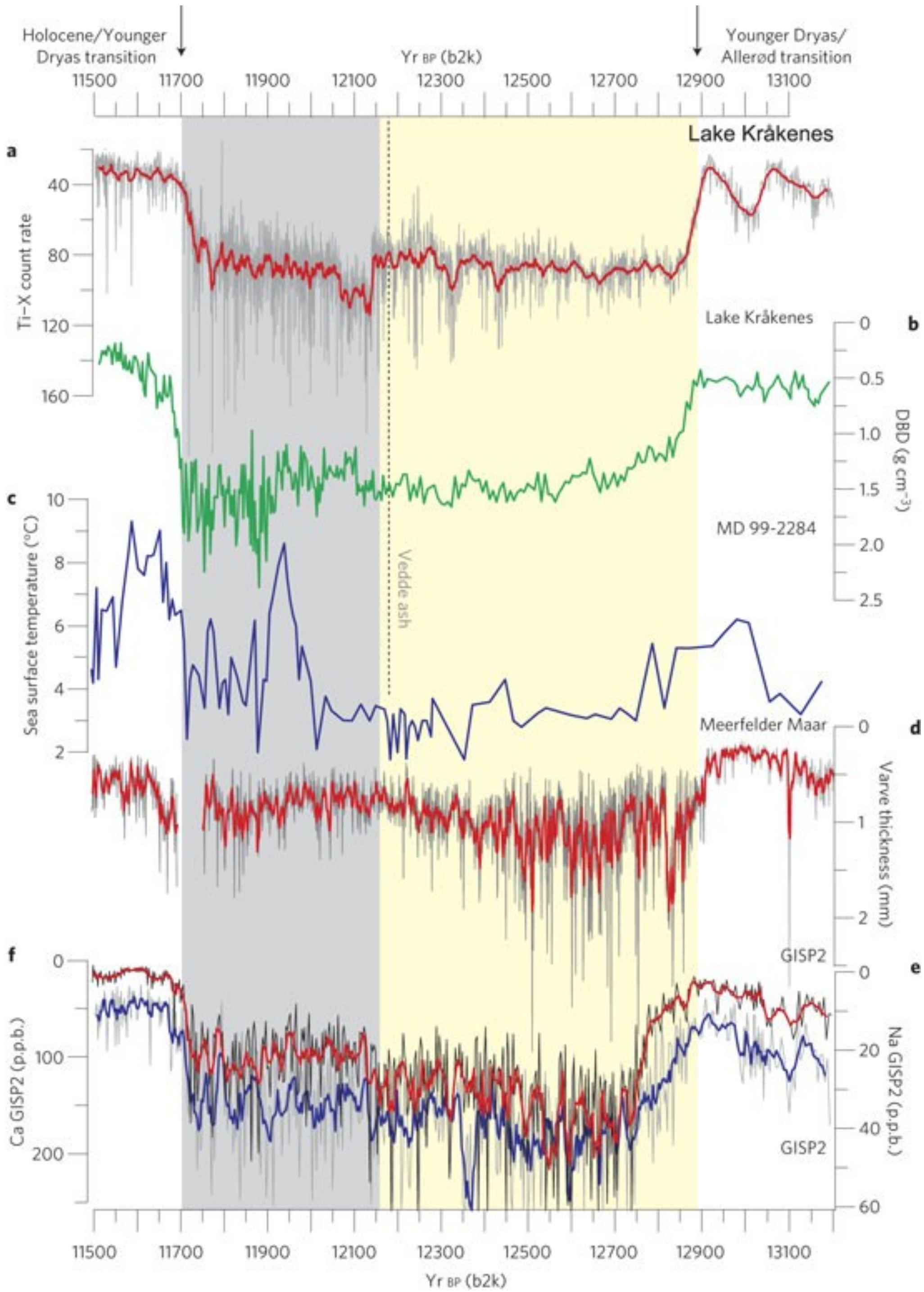

Figure 7: Four different proxies showing the extent of the Younger Dryas (Bakke et al., 2009). Especially noticeable in the Lake Kråkenes dry bulk density (DBP) in lake sediments and temperatures inferred from Lake Meerfelder Maar is the bipartition in the Younger Dryas where larger oscillations are observed before the exit into the Holocene.

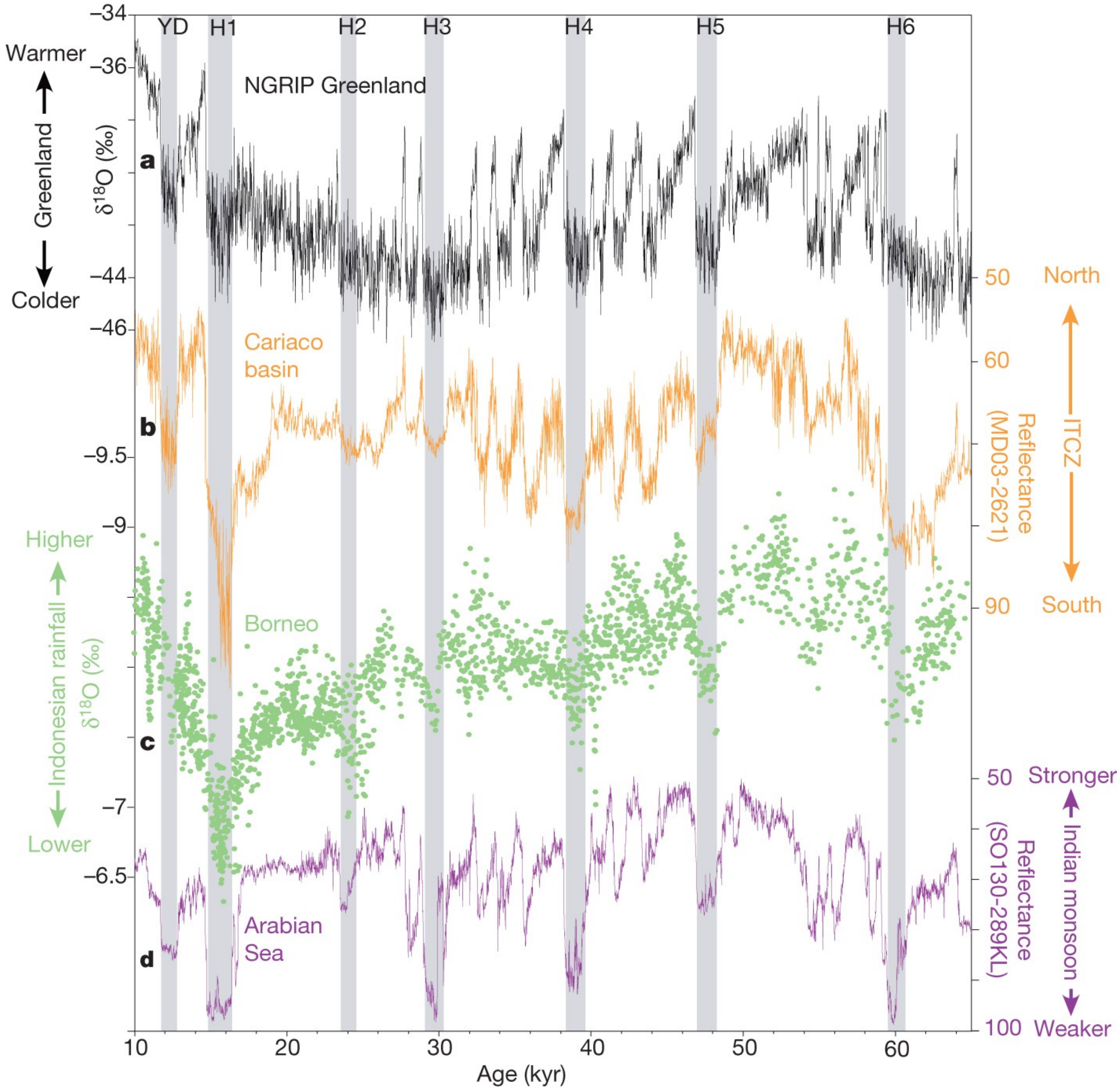

Figure 8: Greenland temperatures (A), Migrations of the ITCZ (B), and tropical monsoon strengths (C and D) from the last ~65,000 years (Schneider et al., 2014). Here it can be seen that during the Younger Dryas, temperatures in Greenland were colder, the ITCZ moved south, and the tropical monsoons generally weakened.

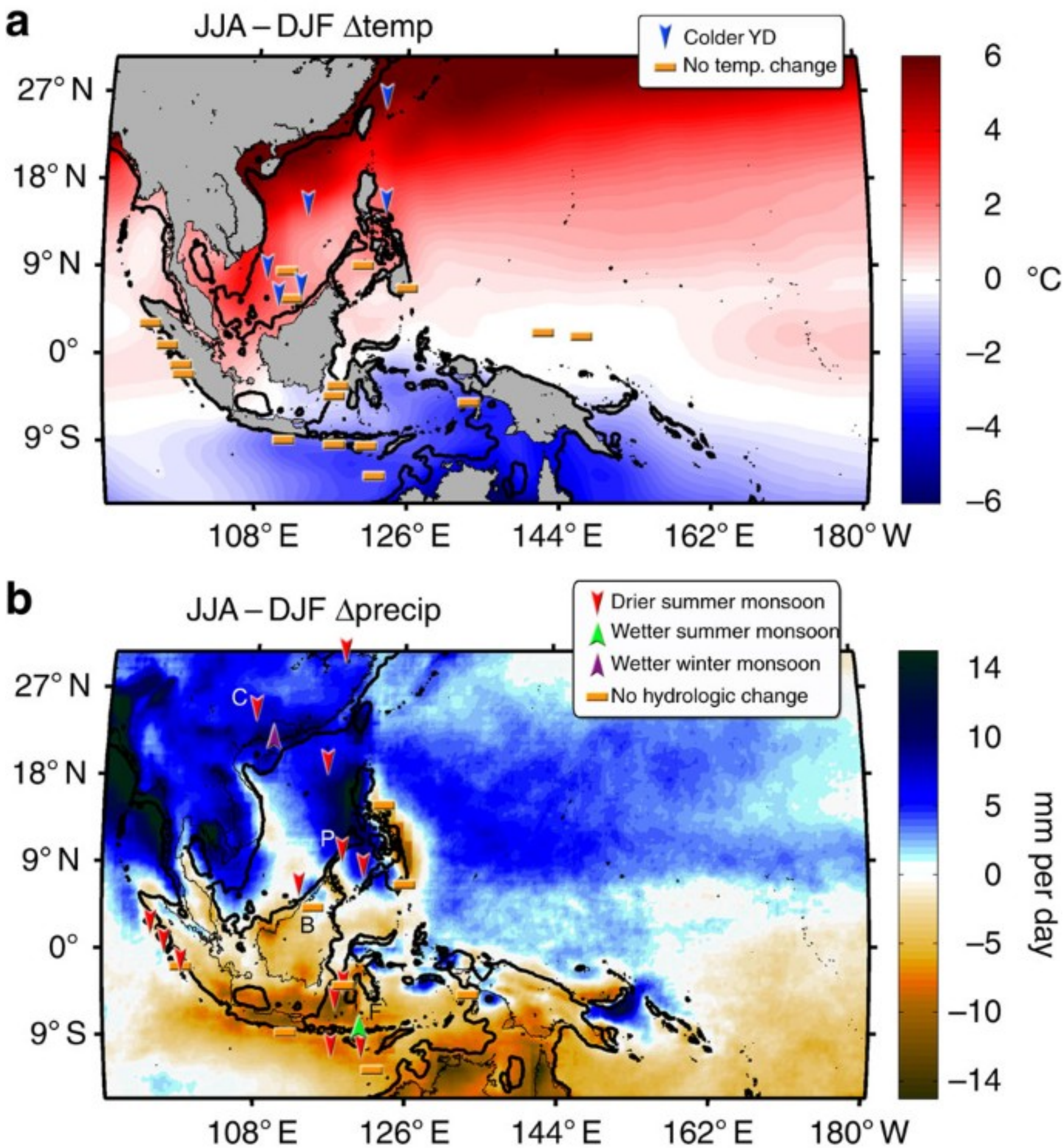

Figure 9: Seasonal differences in temperature (a) and precipitation (b) in the western tropical Pacific (Partin et al., 2015). As can be seen in both maps, it is clear that the summer monsoon is weakened in the north, whereas the data from the southern hemisphere is less certain.

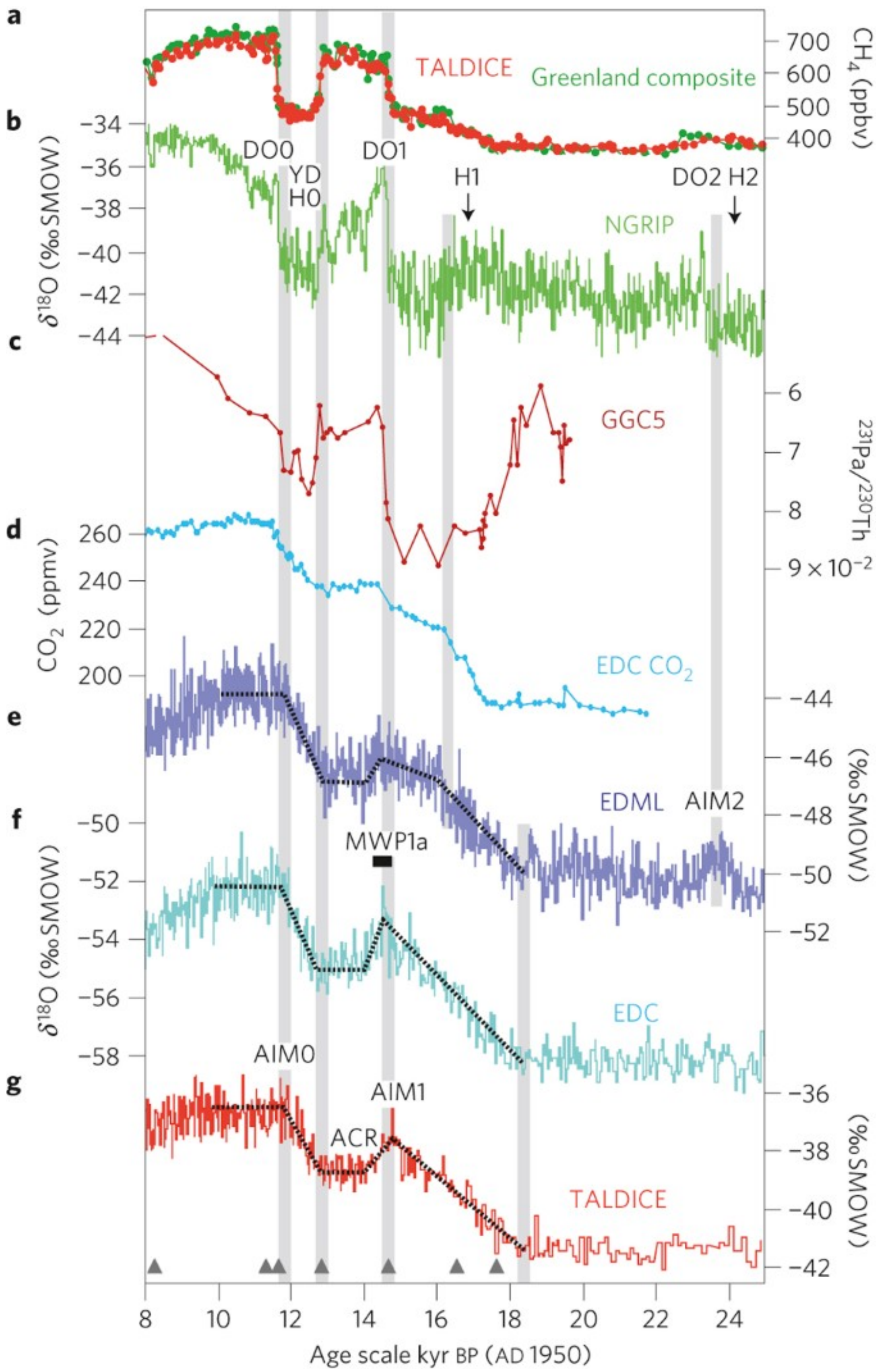

Figure 10: The combined methane records from Greenland and Talos Dome ice core used to synchronize records (TALDICE) (a), the δ^{18}O profile of NGRIP (b), the ^{231}Pa/^{230}Th record from the Bermuda Rise (a proxy for AMOC strength) (c), along with 4 Antarctic core profiles that demonstrate an increase in temperature during the Younger Dryas (d-g) (Stenni et al., 2011).

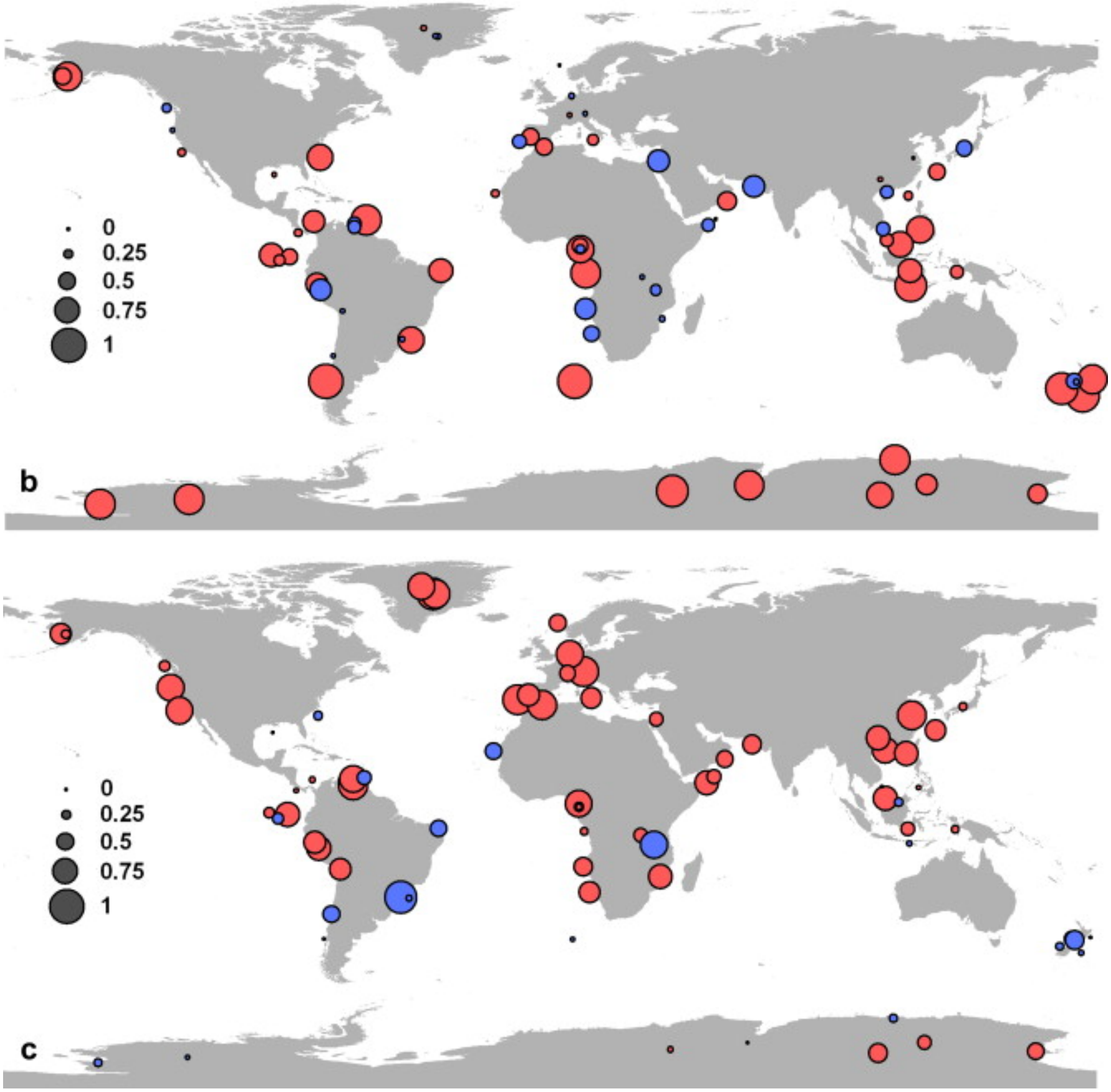

Figure 11: EOF1 (b) and EOF2 (c) for an analysis 15,000-11,000 years BP (Shakun and Carlson, 2010). Positive loadings are shown in red and negative loadings are blue. From here it can be seen that EOF1 has a strong positive loading in the south, whereas EOF2 has a strong positive loading in the north.

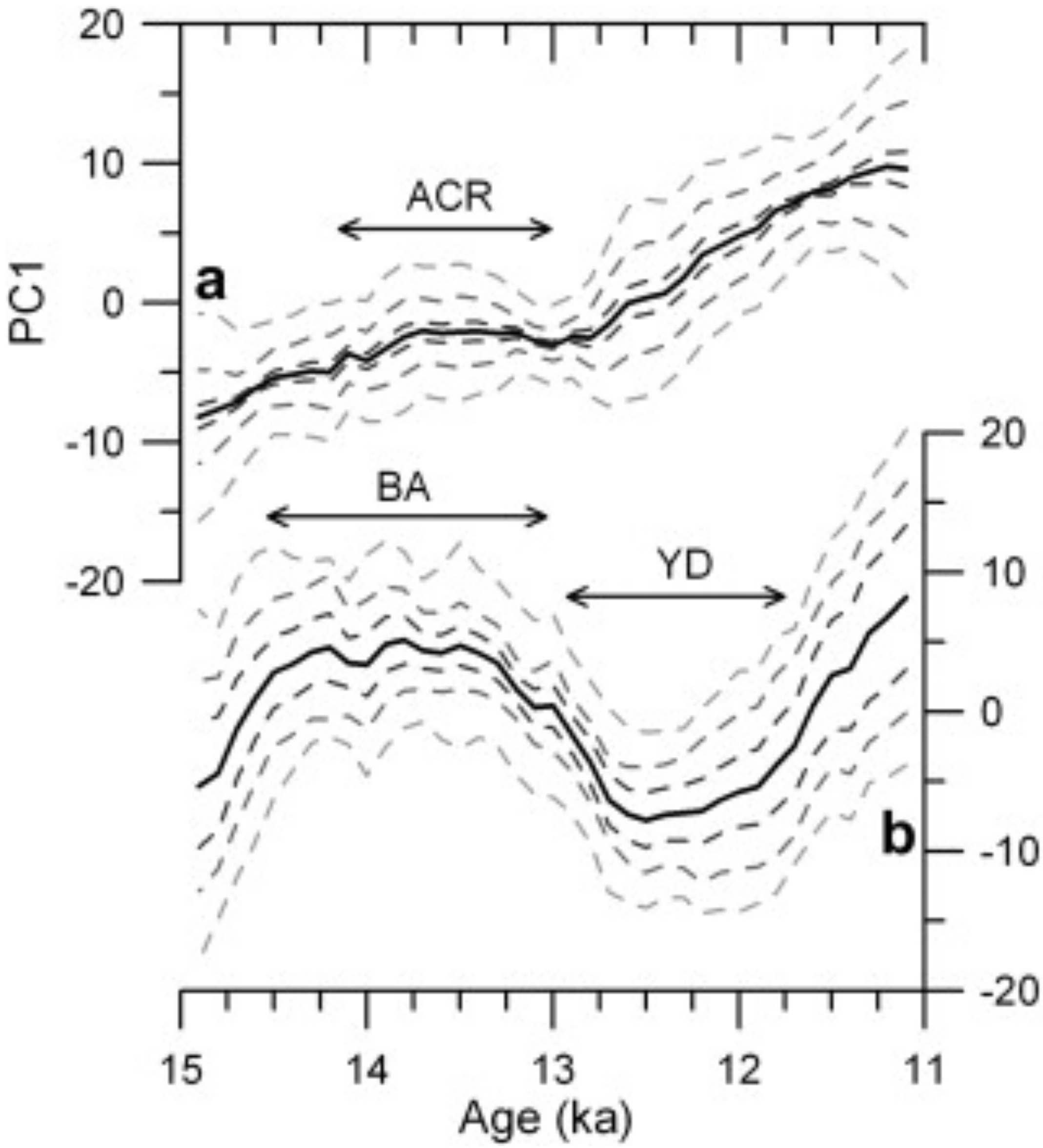

Figure 12: Principal components of EOF1 (a) and EOF (b) (Shakun and Carlson, 2010). Here, clearly defined warming temperature trends in the south and cooling temperature trends in the north can be seen for the duration of the Younger Dryas.

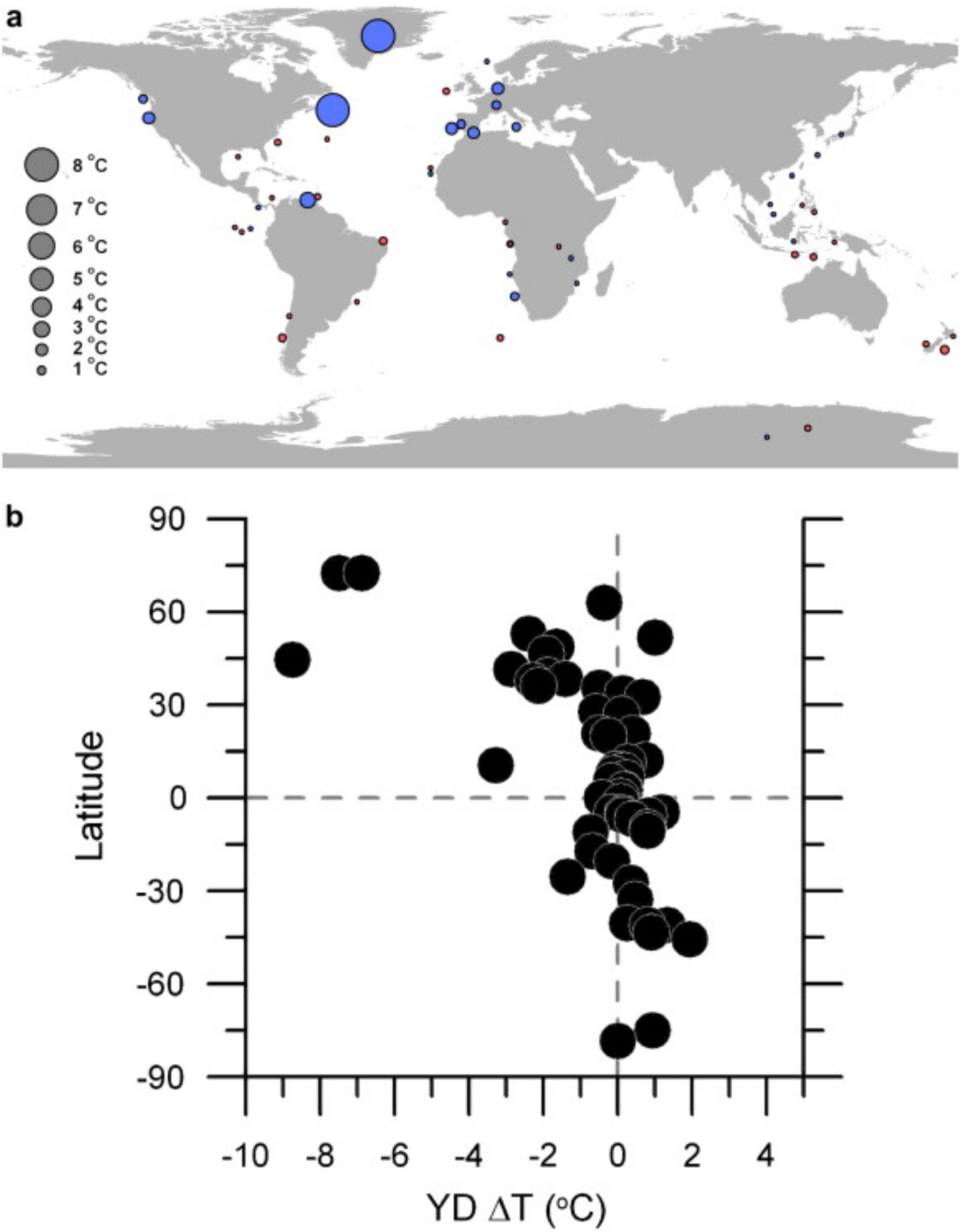

Figure 13: Map of Younger Dryas temperature anomalies (a) and a plot of anomalies per latitude (b) (Shakun and Carlson, 2010). Warmings are shown in red and coolings are shown in blue. From this, the general northern cooling trend and the southern warming trend can be seen.

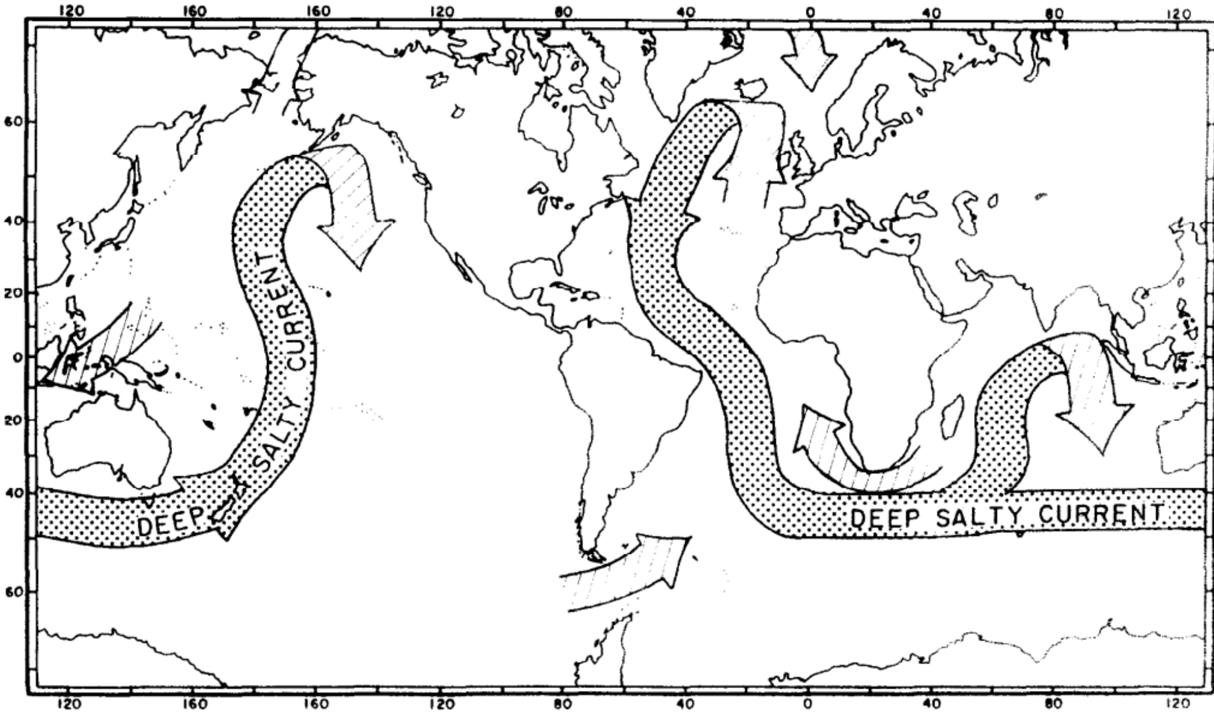

Figure 14: A map illustrating the global thermohaline circulation (Broecker and Denton, 1989). Here we see that a starting point in this circulation pattern lies in the North Atlantic, where NADW forms and sinks.

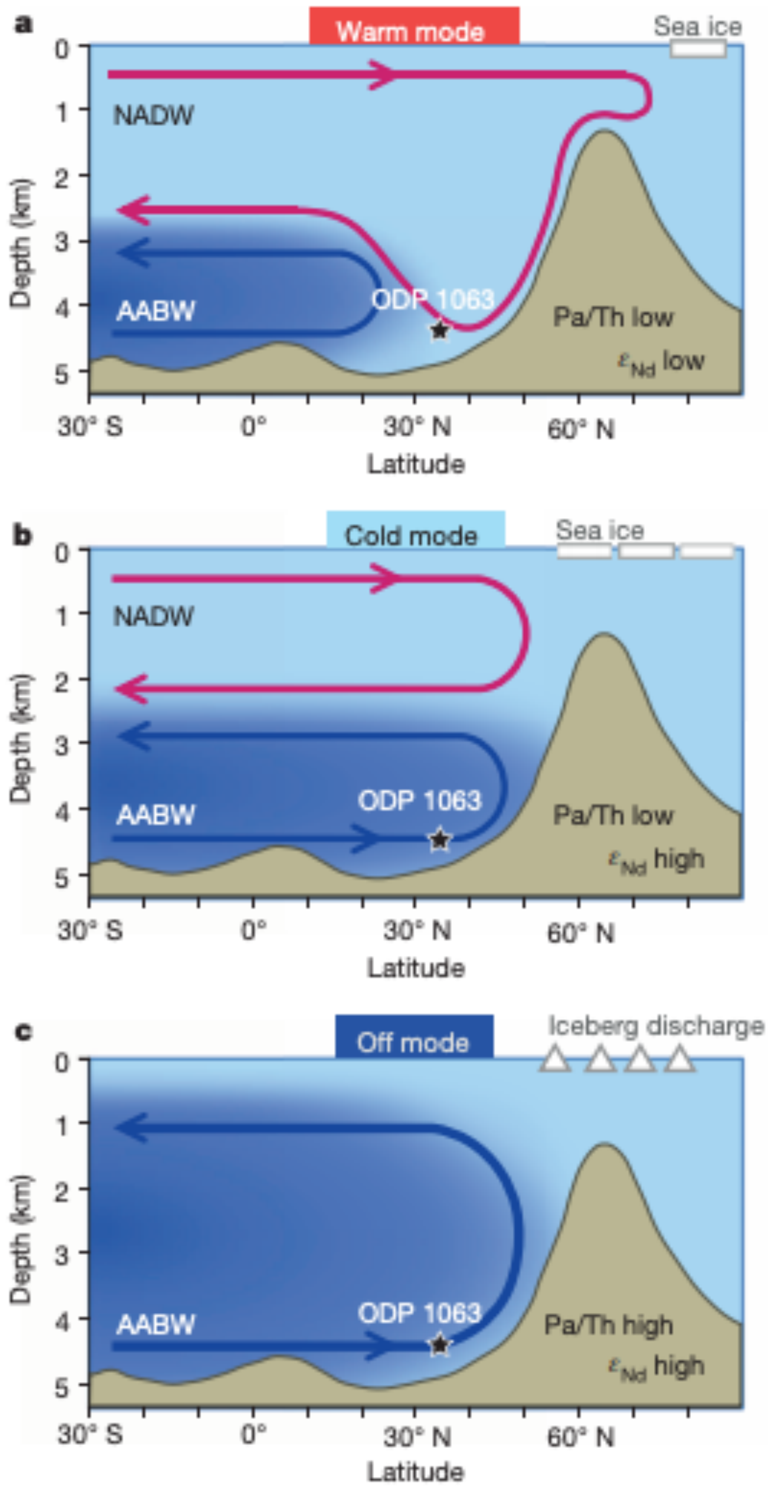

Figure 15: An illustration of various modes of the AMOC with deep NADW ventilation (a), weakened ventilation (b), and little to no ventilation (c) (Böhm et al., 2015). The Younger Dryas is an example of a cold mode, where the slowdown of NADW production prevents the North Atlantic from acting as a heat reservoir.

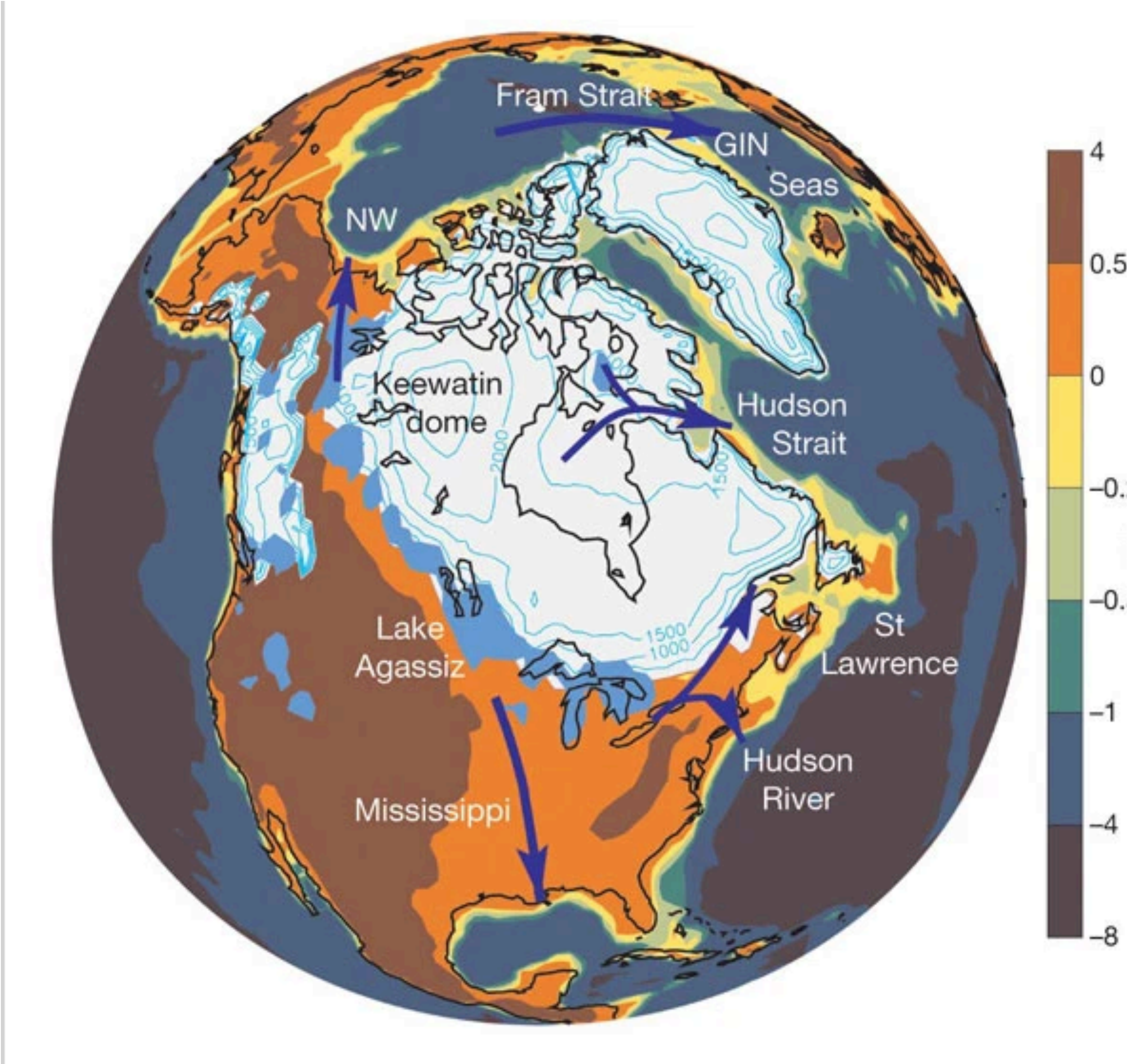

Figure 16: A map of the Laurentide Ice Sheet and Lake Agassiz along with possible drainage routes through the south (Mississippi), the east (Hudson/St. Lawrence Rivers), and the north (northwest through the Mackenzie River into the Fram Strait, or ice discharge through the Hudson Strait) (Tarasov and Pelter, 2005). It appears the discharge through the Mackenzie River is the most likely based off the current evidence.

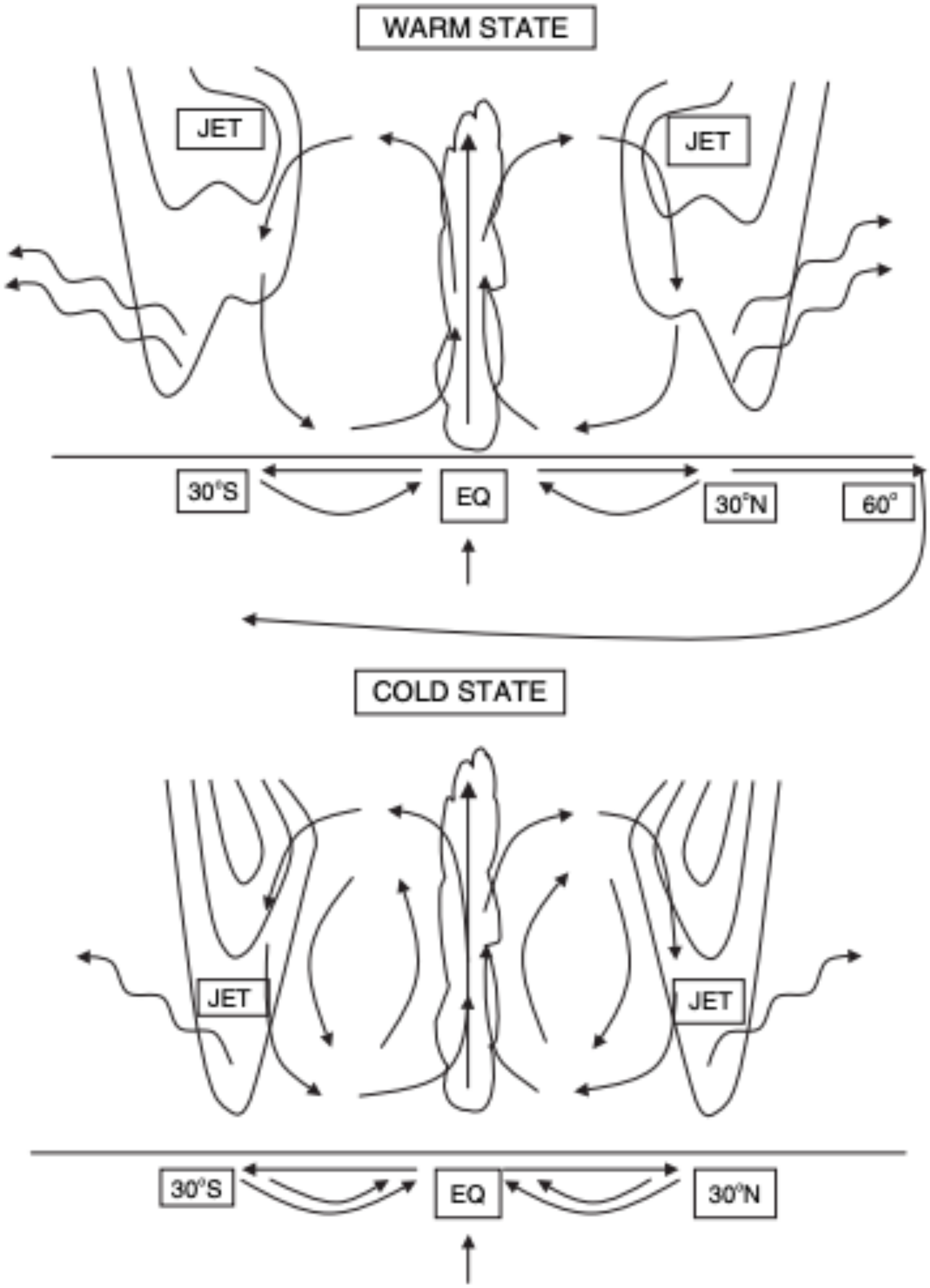

Figure 17: Two modes of atmospheric circulation that are paired with the two modes of oceanic circulation (Seager and Battisti, 2007). In the cold state, there is stronger tropical overturning, and a stronger subtropical jet (this is the zonal jet we see observed in Europe). Furthermore, there is a weaker polar front jet along with weaker eddy heat transport, which are characteristic in stadial periods.

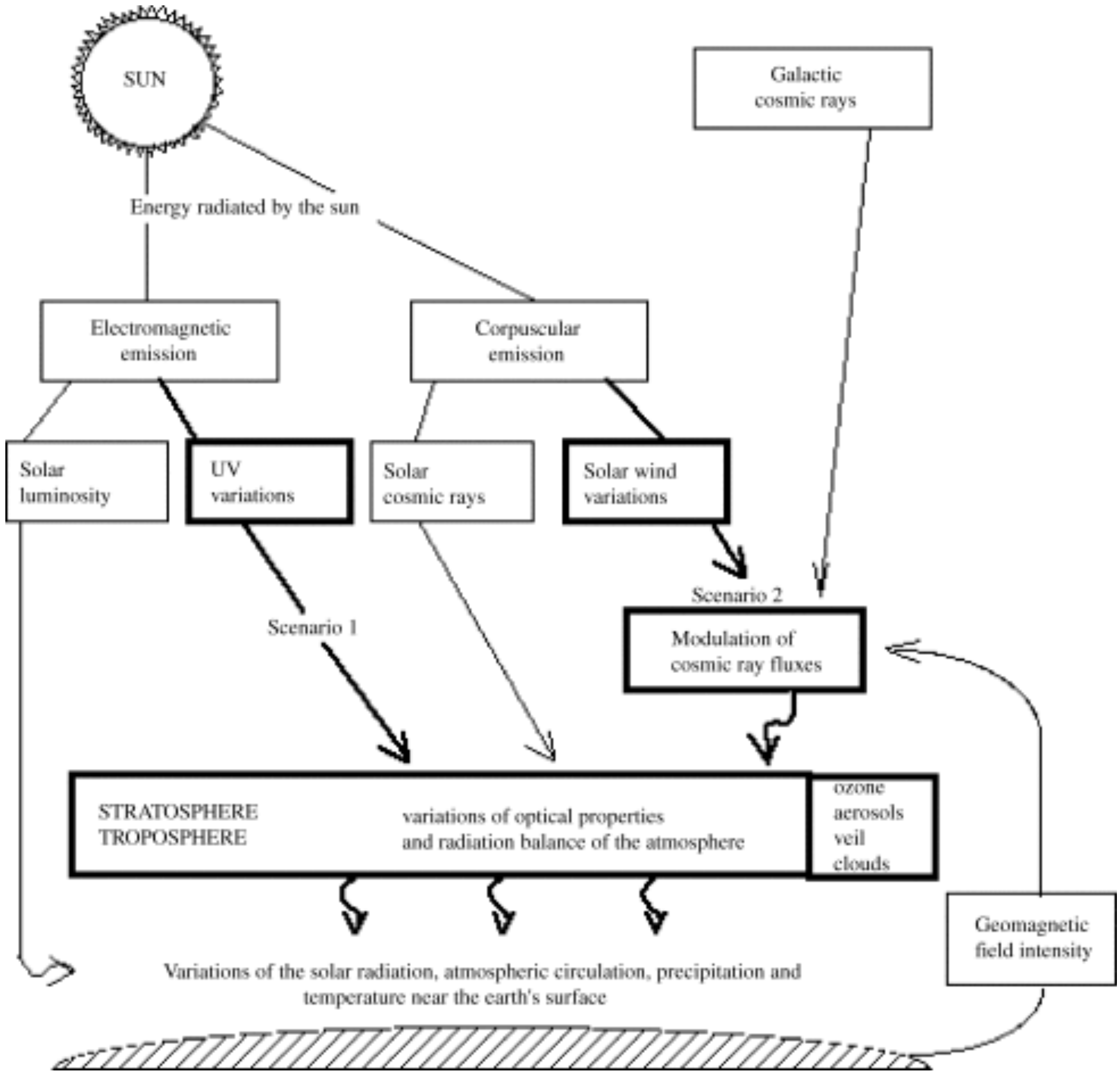

Figure 18: Two solar forcing scenarios (in bold) that lead to climate change (Renssen et al., 2000). The first scenario is based off UV radiation influencing atmospheric ozone, whereas the second is changes in cosmic ray flux through solar wind variations that leads to an increase in cloud cover.

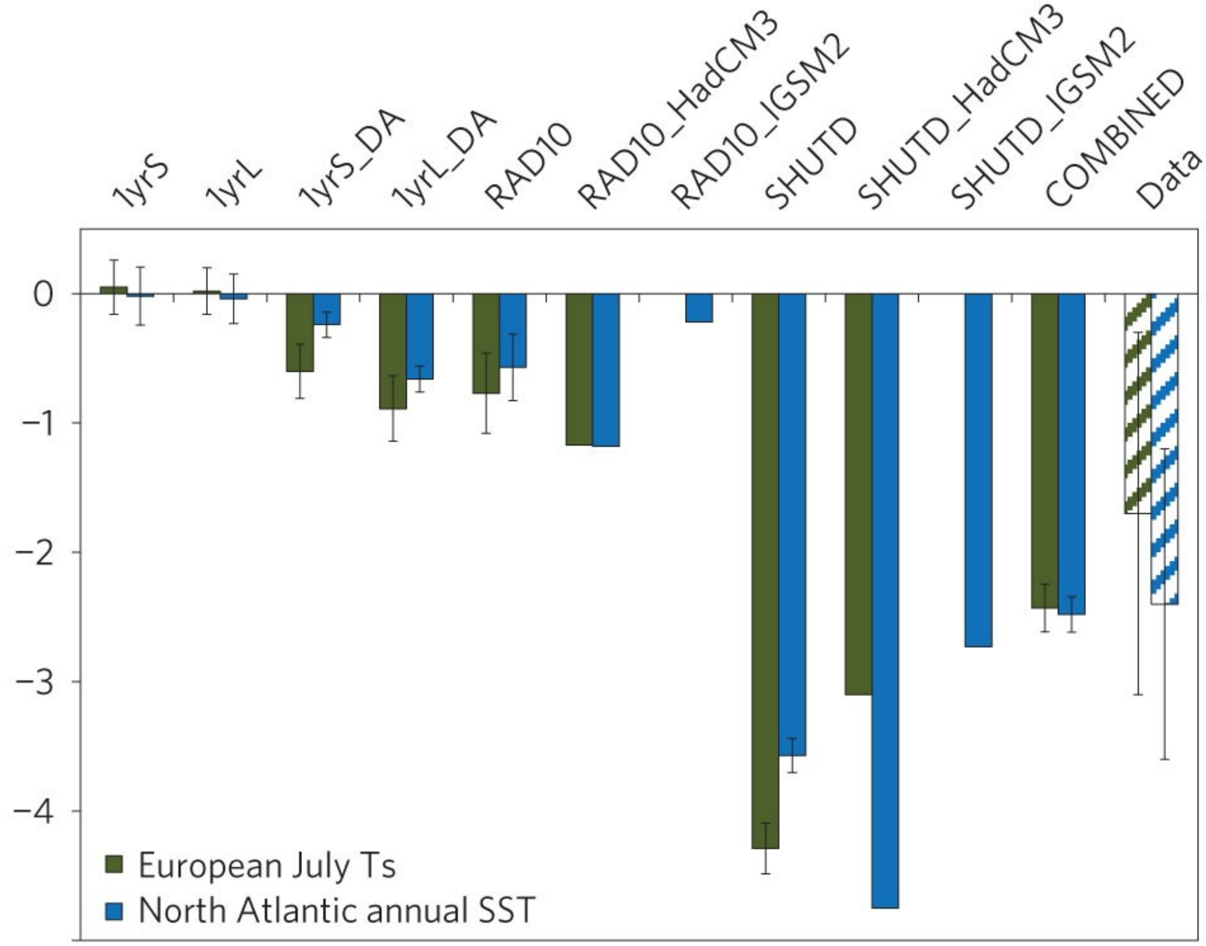

Figure 19: European and North Atlantic temperatures from various model experiments compared with reconstructed Younger Dryas temperatures (Renssen et al., 2015). The different model parameters are shown in Table 1. It can be observed that the model with both freshwater and radiative forcing resulted in the most accurate replication of the proxy-based temperature reconstruction.

Table 1: Experimental model parameters for Renssen et al., 2015

Experiments	Duration (yr)	Additional FW forcing (Sv)	Radiative forcing (W m^{-2})	Ensemble members	Data assimilation
noFW	500	0	0	10	No
1yrS	500	0.5 (1 yr)	0	10	No
1yrL	500	5 (1 yr)	0	10	No
noFW_DA	100	0	0	32	Every 1 yr
1yrS_DA	100	0.5 (1 yr)	0	32	Every 1 yr
1yrL_DA	100	5 (1 yr)	0	96	Every 1 yr
SHUTD	500	4 × Backgr FW	0	10	No
3yrL	100	5 (3 yr)	0	10	No
RAD10	100	0	−10	10	No
3yrLRAD2	100	5 (3 yr)	−2	10	No
COMBINED	1,500	5 (3 yr)	−2	96	Every 5 yr

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