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Re-evaluating the link between the Laacher See volcanic eruption and the Younger Dryas

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Abstract. The Younger Dryas is the most well-documented millennial-scale cooling event of the Quaternary, but the mechanisms responsible for its initiation remain elusive. Here we use a recently revised chronology for the GISP2 ice core ion dataset to identify a large volcanic sulphur spike coincident with both the sulphur-rich Laacher See volcanic eruption and the onset of Younger Dryas-related cooling (GS-1) in Greenland. Lake sediment and stalagmite records confirm that the eruption's timing was indistinguishable from the onset of cooling across the North Atlantic, but that it preceded westerly wind repositioning over central Europe by ~200 years. We suggest that the initial short-lived volcanic sulphate aerosol cooling was amplified by oceanic circulation shifts or sea ice expansion, gradually cooling the North Atlantic region and incrementally shifting the mid-latitude westerlies to the south. The aerosol-related cooling probably only lasted 2-4 years, and the majority of Younger Dryas-related cooling was instead due to this positive feedback, which was particularly effective during the intermediate ice volume conditions characteristic of ~13 ka BP. We conclude that the large and sulphur-rich Laacher See eruption should be considered a viable trigger for the Younger Dryas.

1 Introduction

The Younger Dryas (YD) represents the archetypal millennial scale climate shift. The YD occurred during the last deglaciation and is often described as a brief return to near-glacial conditions in northern Europe. Research now indicates that the YD was indeed characterised by cold conditions across the North Atlantic and Europe (Carlson et al., 2007; von Grafenstein et al., 1999), but also by a southward shifted westerly wind belt over Europe (Bakke et al., 2009; Brauer et al., 2008; Lane et al., 2013; Baldini et al., 2015b), a southwardly shifted Intertropical Convergence Zone (Shakun et al., 2007), increased moisture across the southwest of North America (Polyak et al., 2004), and potential warming in parts of the Southern Hemisphere (Kaplan et al., 2010).

The most widely accepted explanation for the YD involves meltwater forcing-induced weakening of Atlantic Meridional Overturning Circulation (AMOC) (Berger, 1990; Alley, 2000; Broecker et al., 2010). The freshwater, generally proposed to have originated from North American proglacial lakes, is theorised to have formed a freshwater cap over the North Atlantic, encouraging sea ice expansion, prohibiting the formation of North Atlantic Deep Water, and consequently weakening (or shutting down) AMOC. Initial support for this theory included elevated $\delta^{18}O$ values in Gulf of Mexico sediment dating from

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the early YD (implying that meltwater was rerouted elsewhere) (Broecker et al., 1988; Flower and Kennett, 1990; Teller, 1990). However, the meltwater was originally proposed to have travelled to the North Atlantic via the St Lawrence Valley, but exhaustive searches have revealed only limited geological evidence for a massive flux of freshwater coincident with YD initiation (Broecker, 2006a; Rayburn et al., 2011). The freshwater pulse may have followed another route to the ocean, and more recent research has proposed the Mackenzie Valley as an alternative (Condron and Winsor, 2012; Murton et al., 2010). Another complication is that efforts to model the AMOC response to freshwater inputs under deglacial boundary conditions have yielded equivocal results (Meissner, 2007). Consequently, although meltwater forcing remains the most widely researched cause of the YD, it is not universally accepted. Broad agreement does exist that AMOC weakening was associated with the YD onset, but the driver of this weakening remains unclear.

Other recent research has proposed that a large impact event or events over North America may have triggered the YD (Firestone et al., 2007; Kennett et al., 2009). The Younger Dryas Impact Hypothesis (YDIH) is supported by the discovery of iridium, shocked quartz, and millions of tons of impact spherules at the YD boundary layer (Kennett et al., 2009; Wittke et al., 2013; Wu et al., 2013). The YDIH has proven remarkably controversial, and different researchers suggest either terrestrial origins for the same evidence or that the YD boundary layer was misidentified (Pinter et al., 2011; van Hoesel et al., 2015).

Here we do not argue extensively for or against any of these established hypotheses. Instead, we re-introduce and provide new support for the hypothesis that the YD was triggered by the ~12.9 ka BP eruption of the Laacher See volcano, located in the East Eifel Volcanic Field (Germany). Early research considered the eruption as a possible causative mechanism for the YD (Berger, 1990). However, the concept was dismissed because lacustrine evidence across central Europe appeared to indicate that the YD's clearest expression appeared ~200 years after the Laacher See Tephra within the same sediments (e.g., Brauer et al., 2008; Brauer et al., 1999; Hajdas et al., 1995). However, recently revised stratigraphic frameworks for key climate archives now suggest that the Laacher See eruption was in fact synchronous with cooling associated with the YD onset (i.e., GS-1), but preceded atmospheric circulation shifts over central Europe. Here we use GISP2 ion data (Zielinski et al., 1997) on the GICC05modelext chronology to identify a large sulphur spike whose timing is consistent with the Laacher See eruption, and argue that short-lived volcanogenic aerosol cooling triggered a positive feedback that led to dynamical changes most closely associated with the YD.

2 Background

Laacher See volcano, Germany, is situated in the East Eifel Volcanic Field, which is part of the West European rift system (Baales et al., 2002; de Klerk et al., 2008) (Figure 1). The Laacher See Eruption (LSE) occurred at ~12.880 ± 40 ka BP based on the position of tephra within regional varved lake sequences (e.g., Wulf et al., 2013; Brauer et al., 1999; Lane et al., 2015; Bronk Ramsey et al., 2015), consistent with radiocarbon (12.934 ±165 cal ka BP (Baales et al., 2002)) and $^{40}Ar/^{39}Ar$ (12.9 ± 500 ka BP (van den Bogaard, 1995)) ages for the eruption; the absolute age of the eruption is therefore well constrained. The eruption was one of the largest in Europe during the late Quaternary and dispersed over 20 km^3 of pumice and ash over >230,000 km^2 of central Europe and beyond (Baales et al., 2002; Bogaard and Schmincke, 1985). The eruption consisted of alternating Plinian and phreatomagmatic phases; Plinian columns exceeded 20 km height and injected ash and volcanic gas into the stratosphere (Harms and Schmincke, 2000). Direct effects of the LSE included ash deposition, acid rain, wildfires, and increased precipitation, all of which could have affected the local ecology (de Klerk et al., 2008; Baales et al., 2002; Engels et al., 2016; Engels et al., 2015).

The LSE discharged 6.3 km^3 of sulphate-rich, evolved phonolite and sulphide-rich mafic phonolite magma from a strongly compositionally zoned reservoir (Harms and Schmincke, 2000). Petrologic studies suggest that the LSE released ~2 megatons (Mt) of sulphur into the stratosphere (Harms and Schmincke, 2000), considerably more than petrologic estimates of the amount of sulphur erupted during the 1991 Pinatubo eruption (0.11 Mt) (Schmincke et al., 1999). However, petrologic methods typically severely underestimate the total sulphur released to the atmosphere (often by a factor of 10–100) (Harms and Schmincke, 2000). For example, satellite-derived (TOMS) estimates of the SO_2 content of the 5 km^3 Pinatubo eruption are considerably higher than the 0.11 Mt petrologic estimate, at 15–20 Mt (Sheng et al., 2015), and similar discrepancies between petrographic and satellite-derived estimates of sulphur emissions exist for other modern eruptions. Explanations for this excess sulphur include the presence of a sulphur-rich vapour phase that is released prior to and during the eruption (Harms and Schmincke, 2000). Consequently the true sulphur content of the LSE was probably considerably higher than the petrologic estimates, with Baales et al.(2002) suggesting a maximum sulphur release of 150 Mt, approximately nine times larger than that associated with the 1991 Pinatubo eruption. However, unlike the Pinatubo eruption, sulphate aerosols released during the LSE were largely restricted to the Northern Hemisphere (NH) (Figure 2), leading to strong cooling in the stratosphere that affected the NH disproportionately (Graf and Timmreck, 2001).

3 Results and Discussion

3.1 The timing of the Laacher See Eruption relative to the Younger Dryas

Key research using European lake sediment archives suggested that the LST preceded the YD onset by ~200 years (Hajdas et al., 1995; Brauer et al., 1999). For example, the LST appears very clearly in the Meerfelder Maar sediment core (Germany), and it does indeed appear to predate the YD (e.g., Brauer et al., 1999). However, the recent discovery of the Icelandic Vedde Ash in Meerfelder Maar sediments has revised the stratigraphic framework of climate archives around the North Atlantic (Lane et al., 2013). It is now apparent that the clearest hydroclimatic expression of the YD in central Europe lags Greenland cooling associated with Greenland Stadial 1 (GS-1) by 170 years (Figure 3). Specifically, several lake sediment cores from different latitudes have revealed that westerly wind position gradually shifted to the south following the NGRIP-defined initiation of GS-1(Bakke et al., 2009; Lane et al., 2013; Brauer et al., 2008; Muschitiello and Wohlfarth, 2015). An absolutely-dated, speleothem-based wind strength proxy record from northern Spain has independently confirmed this general southward migration of the westerlies (Baldini et al., 2015b).

Although the LSE apparently preceded the most clearly expressed dynamical climatic change associated with the YD in central Europe, it is indistinguishable from the Greenland temperature decrease associated with the GS-1 onset (Figure 3). In central Europe, hydrogen isotope ratios of land plant-derived lipid biomarkers from Lake Meerfelder Maar indicate that North Atlantic atmospheric cooling began at 12.880 ka BP (Rach et al., 2014), preceding the atmospheric response associated with meridionally displaced westerly winds at the same site but synchronous with the onset of Greenland cooling. This observation is further supported by speleothem $\delta^{18}O$ records from La Garma and El Pindal caves in northern Spain (Baldini et al., 2015b; Moreno et al., 2010), Swiss lake sediment $\delta^{18}O$ records (Lotter et al., 1992), and Swiss lake organic molecule-based (BIT and TEX_{86}) based temperature reconstructions (Blaga et al., 2013), all showing synchronous North Atlantic cooling beginning with the LSE (Figure 3). We note that some lacustrine proxy records suggest that cooling began a few years after the LSE (e.g., the Gerzensee $\delta^{18}O$ stack (van Raden et al., 2013)). The existence of a lag between the temperature signal and the recording of that signal in some lacustrine archives may reflect differences in the type of archive used (i.e., terrestrial versus lacustrine archives, etc.), a decadal-scale residence time of groundwater feeding certain lakes, or differences in moisture source regions for different lakes. We note that the trigger responsible for the cooling must coincide (or predate) the earliest evidence for the cooling, and a lag must logically exist between the forcing and any later response (rather than the effect preceding the cause).

Recent work used volcanic marker horizons to transfer the GICC05modelext to the GISP2 ice core (previously on the Meese/Sowers chronology) (Seierstad et al., 2014), improving comparisons with records of volcanism (Figure 4). This also facilitates a direct comparison between the GISP2 volcanic sulphate record (Zielinski et al., 1997) and both the GISP2 and the NGRIP $\delta^{18}O$ records. We find that a large sulphate spike in the GISP2 record on the GICC05modelext timescale is contemporaneous with the LSE's timing based on i) independent Ar-Ar dates for the eruption and ii) layer counting from the Vedde Ash within NGRIP (Figure 4); we therefore ascribe this sulphate spike to the Laacher See eruption. The coincidence between this sulphate spike, the LSE, and the onset of North Atlantic cooling associated with the YD is striking.

3.2 A complex response of climate to volcanic eruptions

Explosive volcanic eruptions can inject large amounts of sulphur-rich gases into the stratosphere, which gradually become oxidised to form sulphuric acid vapour (Rampino and Self, 1984). Within a few weeks, the vapour can condense with water to form an aerosol haze (Robock, 2000), which is rapidly advected around the globe. Once in the atmosphere, sulphate aerosols induce summer cooling by scattering incoming solar radiation back to space (Timmreck and Graf, 2006; Baldini et al., 2015a). An eruption's sulphur content, rather than its explosivity, determines most of the climate response (Robock and Mao, 1995; Sadler and Grattan, 1999).

Although the radiative effects associated with volcanic aerosols are reasonably well understood, the systematics of how volcanic eruptions affect atmospheric circulation are less well constrained. Research suggests that volcanic eruptions affect a wide variety of atmospheric phenomenon, but the exact nature of these links remains unclear. For example, Pausata et al. (2015) used a climate model to conclude that high latitude NH eruptions trigger an El Niño event within 8-9 months by inducing a hemispheric temperature asymmetry leading to southward ITCZ migration and a restructuring of equatorial winds. The model also suggests that these eruptions could lead to AMOC shifts after several decades, illustrating the potential for climate effects extending well beyond sulphate aerosol atmospheric residence times. Quantifying the long-term influences of single volcanic eruptions is confounded by the effects of subsequent eruptions and other factors (e.g., solar variability), which can overprint more subtle feedbacks. Consequently, despite recent hints that volcanic eruptions have considerable long-term consequences for atmospheric circulation, the full scale of these are currently not well understood over the last two millennia, and are essentially unknown under Glacial boundary conditions.

Despite these uncertainties, indications that volcanism may have had longer-term climate effects during the last Glacial do exist. Bay et al.(2004) found a very strong statistical link between evidence for Southern Hemisphere (SH) eruptions and rapid Greenland warming associated with Dansgaard-Oeschger (DO) events, although no causal mechanism was proposed. More recent research determined that every large, radiometrically-dated SH eruption (five eruptions) across the interval 30-80 ka BP occurred within dating uncertainties of a DO event (Baldini et al., 2015a). The same research also found a strong statistical correlation between large NH volcanic eruptions and the onset of Greenland stadials over the same 30-80 ka BP interval (Baldini et al., 2015a), and proposed that during intermediate ice volume conditions, a positive feedback involving sea ice, glacier extent, and/or AMOC weakening may have amplified the initial aerosol injection. The model suggests that this positive feedback continued to operate until it was superseded by another external forcing promoting warming at high latitudes in the NH, or an equilibrium was reached. This model also appears consistent with observations during the YD: *i)* a large NH volcanic eruption, *ii)* long-term NH high latitude cooling, *iii)* NH mid-latitude westerly wind migration to the south, and *iv)* slow recuperation out of the event following rising 65°N insolation and/or a SH meltwater pulse. The LSE was

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twice the magnitude and injected up to nine times the amount of sulphur into the stratosphere as the 1991 AD Pinatubo eruption (Baales et al., 2002; Schmincke, 2004). The Pinatubo eruption cooled surface temperatures by 0.5°C globally, and up to 4°C over Greenland, Europe, and parts of North America, for two years following the eruption (Schmincke, 2004). Existing models suggest that the LSE created a sulphate aerosol veil wrapping around the high northern latitudes (Graf and Timmreck, 2001) (Figure 2). The model suggests that NH summer temperatures dropped by 0.4°C during the first summer following the eruption, though it assumes that the eruption released substantially less sulphur than current maximum estimates (15 versus 150 megatons) (Graf and Timmreck, 2001), so actual aerosol-induced cooling may far exceed these estimates. It does not seem unreasonable that an eruption of this size, sulphur content, and geographic location could have catalysed a NH cooling event.

3.3 The nature of the positive feedback

Volcanogenic sulphate aerosols typically settle out of the atmosphere within five years; aerosol-induced cooling alone therefore cannot explain the YD's extended duration. We suggest that the LSE initiated North Atlantic cooling, which consequently triggered a positive feedback as proposed for earlier Greenland stadials (Baldini et al., 2015a). Existing evidence strongly suggests that North Atlantic sea ice extent increased (Bakke et al., 2009; Baldini et al., 2015b; Broecker, 2006b) and AMOC weakened (Broecker, 2006b; McManus et al., 2004) immediately after the GS-1 onset, and therefore both could have provided a powerful feedback. The feedback may have resided entirely in the North Atlantic, and involved sea ice expansion, AMOC weakening, and increased albedo, as previously suggested within the context of meltwater forcing (Broecker et al., 2010). Alternatively, hemispherically asymmetrical volcanic sulphate loadings may have induced ITCZ migration away from the hemisphere of the eruption (Ridley et al., 2015; Hwang et al., 2013), and these ITCZ shifts may have forced wholesale shifts in atmospheric circulation cells. Within the context of GS-1, LSE-related NH cooling could have shifted the ITCZ to the south, thereby expanding the NH Polar Cell and shifting the NH Polar Front to the south. Sea ice tracked the southward shifted Polar Front, resulting in more NH cooling, a weakened AMOC, and a further southward shift in global atmospheric circulation cells. Such a scenario is consistent with recent results based on tree ring radiocarbon measurements suggesting that GS-1 was not caused exclusively by long-term AMOC weakening, but instead was forced by NH Polar Cell expansion and southward NH Polar Front migration (Hogg et al., 2016).

Our hypothesis that the YD was triggered by the LSE and amplified by a positive feedback is further supported by modelling results suggesting that a combination of a moderate negative radiative cooling, AMOC weakening, and altered atmospheric circulation best explain the YD (Renssen et al., 2015). Furthermore, AMOC consists of both thermohaline and wind-driven components, and atmospheric circulation changes can therefore dramatically affect oceanic advection of warm water to the North Atlantic. Recent modelling suggests that reduced wind stress can immediately weaken AMOC, encouraging southward sea ice expansion and promoting cooling (Yang et al., 2016), illustrating a potential amplification mechanism following an initial aerosol-induced atmospheric circulation shift. 20th Century instrumental measurements further support this by demonstrating that westerly winds strength over the North Atlantic partially modulates AMOC (Delworth et al., 2016). However, despite increasingly tangible evidence that eruptions trigger atmospheric circulation shifts that can subsequently affect AMOC strength and sea ice extent, the exact nature of any positive feedback is still unclear. Future research should prioritize the identification and characterisation of this elusive, but potentially commonplace, feedback that amplifies otherwise subtle NH temperature shifts.

3.4 Sensitivity to ice volume

Magnitude 6 (M6) eruptions are large but not rare; for example, over the last two thousand years, twelve M6 or larger eruptions occurred, but none produced a cooling event as pronounced as the YD. Abrupt millennial scale climate change is characteristic of glacial intervals, and the forcing responsible only appears to operate during intervals when large ice sheets are present. The lack of large-scale and prolonged climate response to external forcing over the Holocene implies either the absence of the forcing (e.g., lack of large meltwater pulses) or a reduced sensitivity to a temporally persistent forcing (e.g., volcanism). The subdued climate response to volcanic eruptions over the recent past could reflect low ice volume conditions and the absence (or muting) of the requisite positive feedback mechanism. However, millennial-scale climate change was also notably absent during the Last Glacial Maximum (~20-30 ka BP), implying that very large ice sheets also discourage millennial-scale climate shifts. The apparent high sensitivity of the climate system to millennial-scale climate change during times of intermediate ice volume is well documented (e.g., Zhang et al., 2014).

Here, we examine the timing of Greenland Stadials relative to ice volume (estimated using Red Sea sea level (Siddall et al., 2003)). The timings of 55 stadial initiations as compiled in the INTIMATE initiative (Rasmussen et al., 2014) are compared relative to ice volume, and indeed a strong bias towards intermediate ice volume conditions exists, with 73% of the millennial scale cooling events occurring during only 40% of the range of sea level across the interval from 0-120 ka BP (Figure 5). A Gaussian distribution exists, with the most sensitive conditions linked to ice volume associated with a sea level of -68.30 m below modern sea level. This intermediate ice volume was commonplace from 35-60 ka BP, and particularly from 50-60 ka BP. However, over the interval from 0-35 ka BP, these ideal intermediate ice volume conditions only existed during a short interval from 11.8-13.7 ka BP, and optimal conditions were centred around 13.0 ka BP (Figure 5). Additionally, a frequency distribution of the sea level change rate associated with each stadial indicates that whatever mechanism is responsible for triggering a stadial operates irrespective of whether sea level (i.e., ice volume) is increasing or decreasing (Figure 5b). Because active ice sheet growth should not promote meltwater pulses, this observation seemingly argues against meltwater pulses as the sole trigger for initiating stadials,

3.5. Comparison with the climate response to the Toba supereruption

It is useful to compare the LSE to another well-dated Quaternary eruption occurring under different background conditions. The magnitude >8 Toba supereruption occurred approximately ~74 ka BP, and, was the largest eruption of the Quaternary

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(Brown et al., 2014). Despite its size, the climate response to the eruption remains vigorously debated, with some researchers suggesting that the eruption triggered the transition to GS-20 (Polyak et al., 2017; Baldini et al., 2015a; Carolin et al., 2013; Williams et al., 2009) but others arguing for only a small climate response (Haslam and Petraglia, 2010). However, it does not appear that the eruption triggered a sustained global volcanic winter (Lane et al., 2013b), and therefore globally homogenous long-term aerosol induced cooling probably did not occur. The Toba supereruption occurred during an orbitally-modulated cooling trend with intermediate ice volume (-71 m below modern sea level, very near the sea level associated with the most sensitive ice volume conditions (-68.3 m below modern sea level)), and it is possible that the eruption expedited the transition to GS-20 through a combination of an initial short-lived aerosol-induced cooling amplified by a positive feedback lasting hundreds of years. The high latitude NH cooling was accompanied by southward ITCZ migration and Antarctic warming, consistent with sea ice growth across the North Atlantic and a weakened AMOC.

The rates of cooling into GS-1 (the YD) and GS-20 are nearly identical (Figure 6), implying that a similar process was responsible for the cooling in both instances. The cooling rate may reflect the nature of the post-eruptive positive feedback, and was largely independent of the initial radiative cooling effects of the two eruptions, provided that they were large enough to trigger a feedback. The reconstructed climate responses following the LSE and Toba eruptions diverge after the achievement of maximum cooling. In the case of Toba, cold conditions persisted for ~1.7 ka before the very rapid temperature increase characteristic of the onset of GI-19 (Figure 6). GS-1 ended after ~1.2 ka, but unlike GS-20, which had stable cold temperatures throughout, the coldest conditions in Greenland occurred near the beginning of the GS-1 and were followed by gradual warming (Litt et al., 2003). The LSE and Toba eruptions occurred under different orbital configurations, and the contrasting topologies characteristic of GS-1 and GS-20 may reflect differing insolation trends. At ~13 ka BP, summer insolation at 65°N was increasing, rather than decreasing as during the Toba eruption ~74 ka BP. The Toba eruption may have catalysed a shift to the insolation-mediated baseline (cold) state, whereas the LSE may have forced a temporary shift to cold conditions opposed to the insolation-driven warming characteristic of that time interval, resulting in a short-lived cold event followed by gradual warming. Radiative cooling events in the Southern Hemisphere (e.g., a SH eruption, an Antarctic meltwater pulse, etc.) may have abruptly terminated both GS-1 and GS-20; the GS-1 termination coincides with Meltwater Pulse 1B (MWP-1B) (Leventer et al., 2006; Ridgwell et al., 2012), whereas any SH trigger for the termination of GS-20 is unidentified. MWP-1B may have cooled the SH and strengthened AMOC, prompting northward migration of the ITCZ and NH mid-latitude westerlies to achieve equilibrium with high insolation conditions, thereby rapidly reducing sea ice extent and warming Greenland, but this requires further research.

3.6 Response to other late Quaternary eruptions

The strongest argument against the LSE contributing to YD cooling is that several other similar or larger magnitude eruptions must have occurred during the last deglaciation, and that therefore the LSE was not unusual. However, the LSE *i)* was unusually sulphur-rich, *ii)* was high-latitude, and *iii)* coincided with ideal ice volume conditions. It is in fact the only known sulphur-rich high latitude eruption coinciding with the most sensitive ice volume conditions during the last deglaciation. Furthermore, other eruptions may also have contributed to climate change but current chronological uncertainties preclude establishing definitive links. For example, the ~14.2 ka BP Campi Flegrei eruption may have caused cooling but this is currently not possible to confirm with age uncertainties of ±1.19 ka.

Other large sulphate spikes exist in the GISP2 sulphate record, and in fact the two spikes at 12.6 ka BP and 13.03 ka BP are both larger than the potential LSE sulphate spike. A large eruption of Nevado de Toluca, Mexico, is dated at 12.45 ± 0.35 ka BP (Arce et al., 2003), and could correspond to the large sulphate spike at 12.6 ka BP within the GISP2 ice core. The eruption was approximately the same size as the LSE, so the lack of climate cooling may reflect a different climate response due to the eruption's latitude, which caused a more even distribution of aerosols across both hemispheres, or a lower sulphur load. The 12.6 ka BP sulphate spike is associated with a short but dramatic cooling; therefore the lack of long-term cooling may simply reflect the fact that temperatures had already reached the lowest values possible under the insolation and carbon dioxide baseline conditions characteristic of that time. The sulphate spike at 13.03 ka BP is likely related to the small eruption of Hekla (Iceland) (Muschitiello et al., 2017; Mortensen et al., 2005), which deposited sulphate in the nearby Greenland ice but was climatologically insignificant.

4 Conclusions

We suggest that the ~12.9 ka BP Laacher See volcanic eruption may have contributed to GS-1 cooling and the atmospheric reorganisation associated with the YD event. Recent revisions to the chronological framework of key European climate archives now suggest that the onset of GS-1 related North Atlantic cooling occurred simultaneously with the LSE. Notably, we have identified a large volcanic sulphur spike within the GISP2 ion data (Zielinski et al., 1997) on the new GICC05modelext chronology that coincides with both the LSE and the onset of GS-1. Lipid biomarker hydrogen isotope ratios from Lake Meerfelder Maar further corroborate that GS-1 atmospheric cooling began at 12.880 ka BP, coincident with the Laacher See Tephra within the same sediment and preceding the larger dynamical atmospheric response associated with the YD in central Europe (Rach et al., 2014). Aerosol induced cooling following the eruption may have triggered a positive feedback involving sea ice expansion and AMOC weakening, as previously proposed for other Greenland stadials over the interval 30-80 ka BP (Baldini et al., 2015a). Viewed from this perspective, the YD was simply the latest, and last, manifestation of a Last Glacial stadial.

Intermediate ice volume conditions around ~13 ka BP, driven by rising insolation during the Last Glacial termination, may have promoted a positive feedback following the LSE's injection of up to 150 megatons of sulphur into the stratosphere. This is also consistent with observations that YD-like events were not unique to the last deglaciation, but existed during older deglaciations as well (Broecker et al., 2010). At least one high-latitude M6 eruption likely occurred during most deglacial intervals. GS-1 and earlier stadials may therefore reflect the convergence of a large, sulphate-rich high latitude NH eruptions with intermediate ice volume conditions. NH continental ice sheet decay induced continental lithospheric unloading, and may have triggered high latitude NH volcanism (Zielinski et al., 1997; Sternai et al., 2016), introducing the intriguing possibility that eruptions such as the LSE were not randomly distributed geographically and temporally, but instead were intrinsically linked to deglaciation.

More research is necessary to better characterize the sensitivity of Last Glacial climate to volcanic eruption latitude, sulphur content, and magnitude. Accurate dating of other large volcanic eruptions during intermediate ice volume conditions is key to testing the link between volcanism and Greenland stadials; this information could eventually also support, or refute, the LSE's role in triggering the YD. Despite these uncertainties, the currently available evidence strongly supports the concept that the Laacher See eruption played a key role in catalysing the Younger Dryas event.

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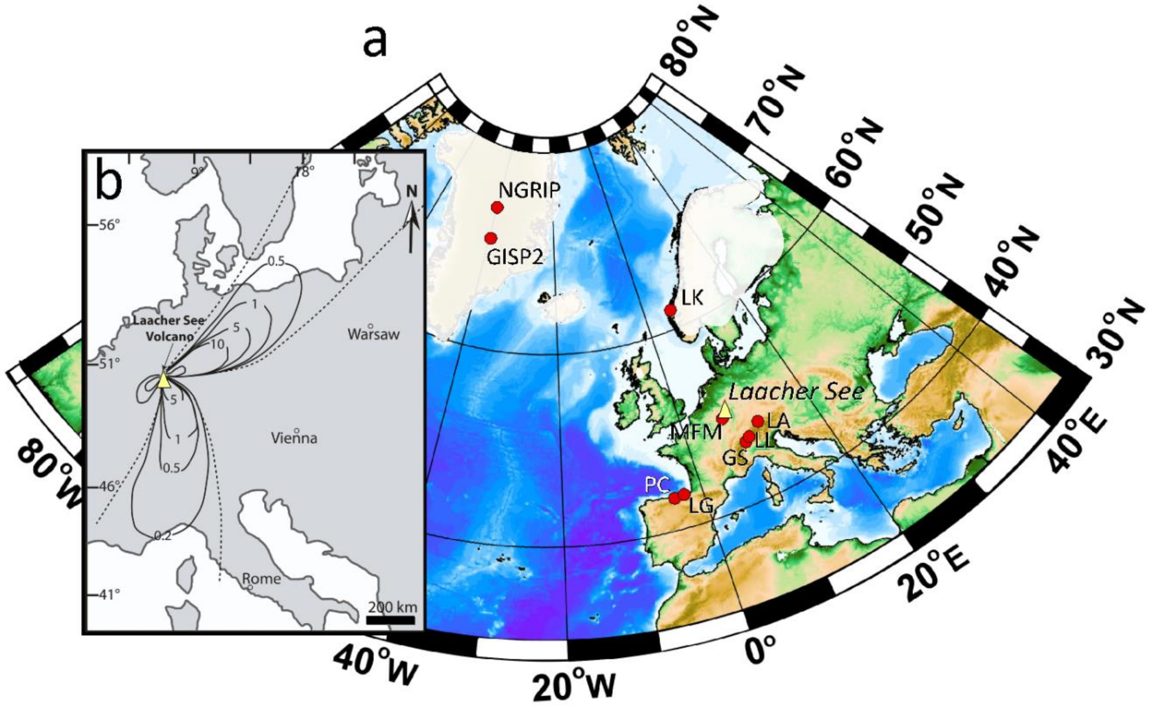

Figure 1. a) Map with the locations of sites discussed and (b) an isopach map of the Laacher See tephra fall deposits across central Europe. Semi-transparent white areas in (a) demarcate continental glaciers, and the reconstruction and map are adapted from Baldini et al. (2015b). The sites shown are: Laacher See volcano (yellow triangle); LG, La Garma Cave; MFM, Meerfelder Maar; LA, Lake Ammersee; PC, El Pindal Cave; LK, Lake Krakenes; LL, Lake Lucerne; GS, Gerzensee; NGRIP, NGRIP ice core. The dashed line in (b) is the outer detection limits of the distal tephra layers [adapted from Bogaard and Schmincke, 1985]. Isopach line labels in (b) are in cm.

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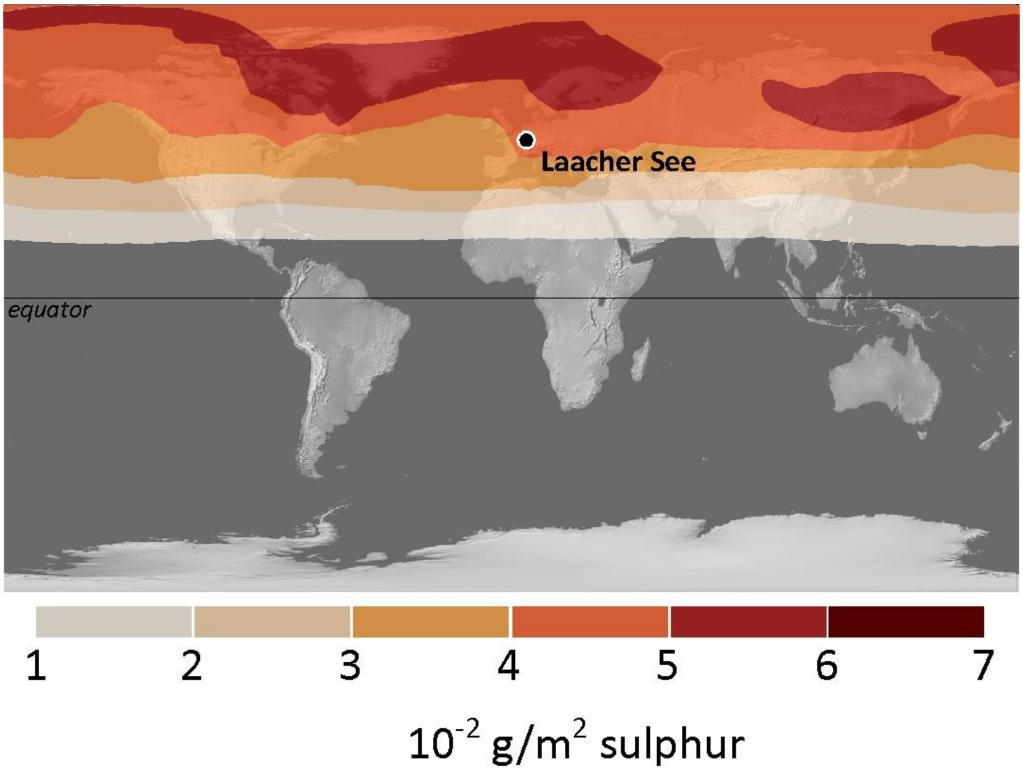

Figure 2. The distribution of the Laacher See volcanic cloud six months after the ~12.9 ka BP eruption based on existing climate model outputs (MAECHAM4 model) (Graf and Timmreck, 2001). The figure is redrawn based on the original in Graf and Timmreck (2001), which contains details regarding the model and simulation. Note that this simulation assumed that the eruption injected 15 megatons of SO_2 into the lower stratosphere, compared with more recent estimates of up to 150 megatons sulphur (Baales et al., 2002; Mortensen et al., 2005).

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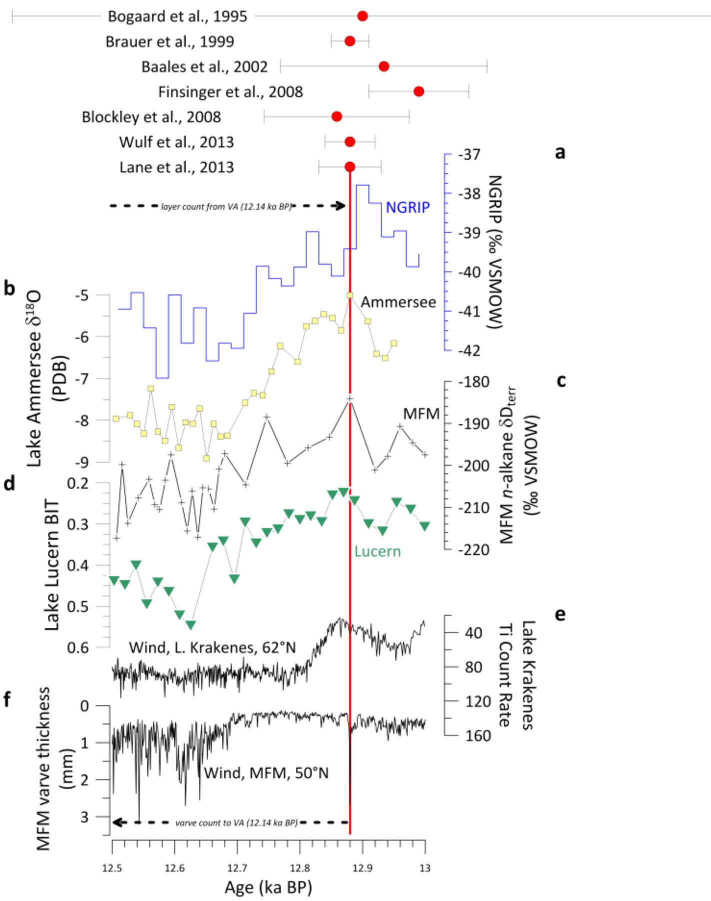

Figure 3. Temperature (a-d) and wind strength (e, f) proxy records from the North Atlantic and Europe. a) NGRIP ice core $\delta^{18}O$ (Greenland, GICC05) (Rasmussen et al., 2006). b) precipitation $\delta^{18}O$ reconstructed using deep lake ostracods from Lake Ammersee, southern Germany. c) Meerfelder Maar (MFM) (Eifel Volcanic Field, Germany) *n*-alkanes δD_{terr} (terrestrial lipid biomarkers) (Rach et al., 2014), d) Lake Lucern, Switzerland, isoprenoid tetraether (BIT) record (Blaga et al., 2013). e) wind strength as reconstructed using Ti count rate in Lake Kråkenes (Norway); three-point moving average shown. f) Wind strength as reconstructed using MFM varve thickness data(Brauer et al., 2008; Mortensen et al., 2005). The records are arranged so that cooling is down for all the records. The LST (vertical red line) is present or inferred within the MFM, Lake Ammersee, and Lake Lucern cores. The LST is not evident in the NGRIP or Lake Kråkenes cores, and the eruption's timing relative to NGRIP $\delta^{18}O$ and Lake Kråkenes Ti is based on layer counting from the Vedde Ash ('VA'), a tephrochronological marker (12.140 ± 0.04 ka BP) also found in MFM(Lane et al., 2013; Brauer et al., 2008). The published chronologies for Lake Ammersee and Lake Lucern were shifted slightly by a uniform amount (0.115 and -0.093 ka, respectively) to ensure the contemporaneity of the LSE in all the records. This adjustment does not affect the LSE's timing relative to the Lake Ammersee or Lake Lucern climate records. Published radiometric dates for the LSE are shown (red circles) with errors, although the absolute age is not as important as its timing relative to the apparent climate shifts.

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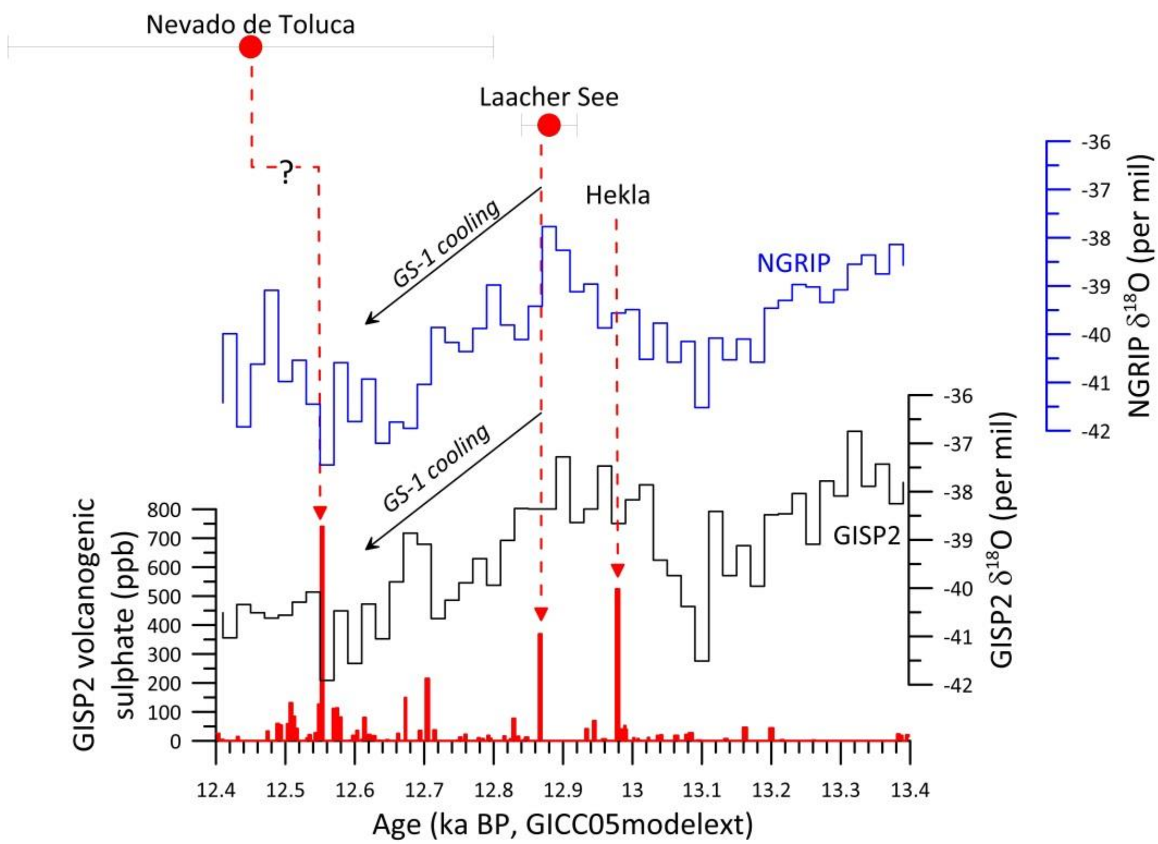

Figure 4. The GISP2 (bidecadal (Stuiver et al., 1995)) (black) and NGRIP (blue) ice core $\delta^{18}O$ records synchronised on the GICC05modelext chronology (Seierstad et al., 2014). The red bars indicate GISP2 volcanological sulfate record (also synchronised on GICC05 timescale). The best age estimate of the LSE is shown (red circle) (Brauer et al., 2008; Lane et al., 2015) and a possible correlation to a synchronous sulfate spike highlighted by the vertical dashed arrow. The large spike at ~13 ka BP represents a smaller but more proximal Icelandic eruption (Hekla) associated with a volcanic ash layer (Mortensen et al., 2005).

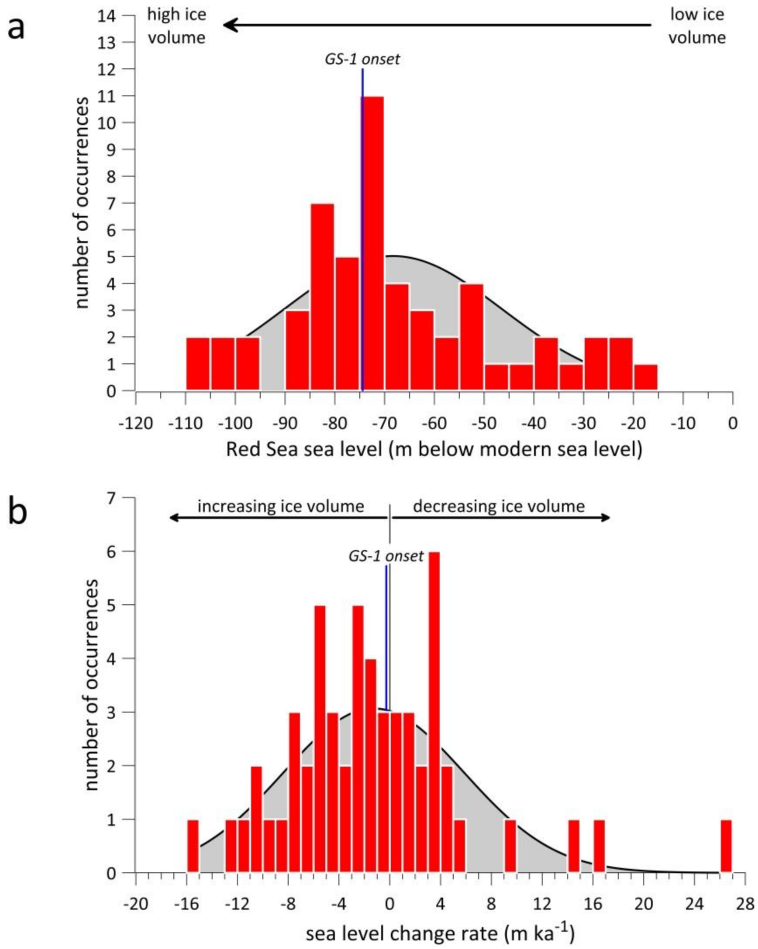

Figure 5: Histogram of the frequency of abrupt Greenland cooling events relative to ice volume and ice volume change. Red Sea sea level (Siddall et al., 2003) is used as a proxy for ice volume and to determine (a) the distribution of ice volume conditions associated with all Greenland cooling events identified by Rasmussen et al.(2014) over the last 120 ka (n = 55). A Gaussian best-fit curve describes the distribution (grey-filled curve) (mean = -68.30 m, σ = 21.85 m), and sea level during the GS-1 onset is marked with a blue dashed line. b) The distribution of sea level change rates (reflecting global ice volume shifts) associated with 55 Greenland cooling events over the last 120 ka. This distribution is also described by a Gaussian best-fit curve (grey-filled curve) (mean = -68.30 m ka^{-1}, σ = 21.85 m ka^{-1}); 21 events are associated with decreasing ice volume, and 34 are associated with increasing ice volume.

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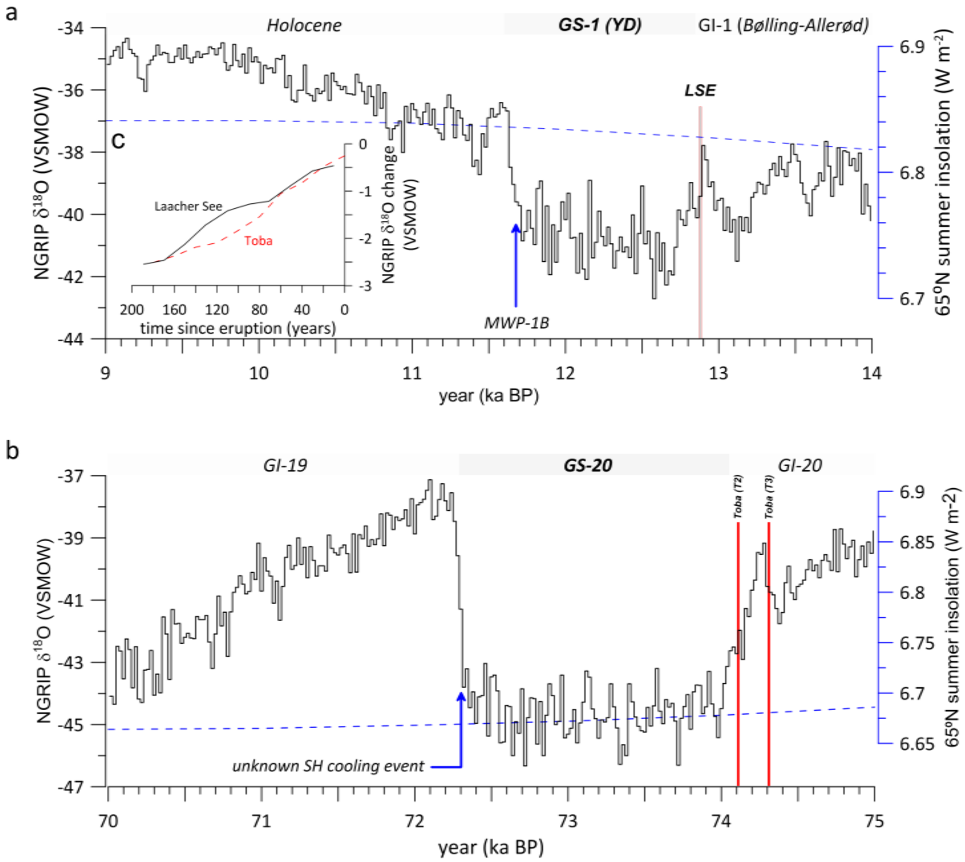

Figure 6: Two five-ka intervals of the NGRIP $\delta^{18}O$ record bracketing (a) the ~12.9 ka BP Laacher See Eruption and (b) the ~74 ka BP Toba supereruption (Rasmussen et al., 2006). The two eruptions are among the most well-dated eruptions of the Quaternary. The LST is not apparent in the NGRIP core (although we identify a candidate sulphate spike in the GISP2 record; see Figure 4), and the timing of the LST relative to the NGRIP $\delta^{18}O$ record is based on the difference in ages between the LST and the Vedde Ash found in European lake sediments (Lane et al., 2013). The timing of the Toba eruption relative to NGRIP is based on the position of the most likely sulphate spikes identified by Svensson et al. (2013). Blue arrows indicate the timing of possible SH cooling events. The inset panel (c) shows NGRIP $\delta^{18}O$ during the 200 years immediately following the two eruptions (the sulphate spike T2 is assumed to represent Toba, as suggested by Svensson et al. (2013)).