Climatic and volcanic forcing of tropical belt northern boundary over the past 800 years

Alfaro-Sánchez R^{1,3*}, Nguyen H², Klesse S³, Hudson A³, Belmecheri S³, Köse N⁴, Diaz HF⁵, Monson RK^{3,6}, Villalba R⁷, Trouet V³

Corresponding author: *r.alfarosanchez@gmail.com

¹Centre de Recerca Ecològica i Aplicacions Forestals (CREAF), Barcelona, Spain

²Bureau of Meteorology, Melbourne, Victoria, Australia

³Laboratory of Tree-Ring Research, University of Arizona, Tucson, Arizona, USA

⁴Istanbul University-Cerrahpaşa, Forestry Faculty, Forest Botany Department, Bahçeköy-Istanbul, Turkey

⁵Department of Geography and Environment, University of Hawai'i at Mānoa, Honolulu, USA

⁶Department of Ecology and Evolutionary Biology, University of Arizona, Tucson, Arizona, USA

⁷Instituto Argentino de Nivología, Glaciología y Ciencias Ambientales (IANIGLA - CONICET), Mendoza, Argentina

The position of the northern boundary of the tropical belt has a significant impact on the hydroclimate of many Northern Hemisphere arid and semi-arid regions. Widening of the tropical belt since the 1970s has largely been attributed to anthropogenic forcing. However, prior to this, the relative influence of natural drivers of tropical belt expansion and contraction is poorly understood. Here we use tree-ring width data from five Northern Hemisphere mid-latitude regions to reconstruct early spring tropical belt northern boundary movements over the past 800 years (1203-2003 CE). Our reconstruction targets the latitudinal extent of the Hadley Circulation as a metric of tropical belt boundary position, and explains 45% of its interannual variance. We find that the tropical belt contracted (expanded) significantly during positive (negative) phases of the El Niño Southern Oscillation and Pacific North American teleconnection patterns. The tropical belt also contracted significantly following major volcanic stratospheric sulfur injection events. The longest period of persistent tropical belt expansion occurred in the late 16th century, during the crux of the Little Ice Age. Our results warn of potential socio-economic consequences of future variations in tropical belt width driven by natural climate variability or stratospheric aerosol injections, whether volcanic or artificial.

Main

The boundaries of the tropical belt – climatologically identified based on a range of climatological metrics including the extent of the Hadley Circulation - determine the latitude of the subtropical dry zone¹. Latitudinal movements of the tropical belt boundary influence the hydroclimate in many arid and semi-arid regions and are

critical for sustainable urban resource use, regional agriculture, and ecosystem water services².

On seasonal and interannual time-scales, the position of the tropical belt boundary is associated with external forcings such as volcanic activity, as well as with internal climate variability³, including tropical-extratropical teleconnections, such as the Pacific North American (PNA)⁴ pattern and the El Niño Southern Oscillation (ENSO) system⁵. On multi-decadal time-scales, the Pacific Decadal Oscillation (PDO)⁶ and the Northern and Southern Annular Modes⁷ can influence tropical belt variations.

Since the late 1970s, the tropical belt has shown a poleward expansion in both hemispheres⁸, which is associated with changing global precipitation patterns and has encouraged a suite of observational, reanalysis, and modeling studies investigating the earth system mechanisms and forcings that drive shifts in tropical belt width^{1,3}.

The magnitude of the observed late twentieth century poleward shift is dependent on the data and metrics used⁸. Reanalysis products tend to agree on the Northern Hemisphere (NH) tropical expansion⁷ and to amplify it compared to model simulations^{8,9}, even when the magnitude of the expansion is less pronounced in most recent reanalysis products⁹.

Historical general circulation model (GCM) runs indicate that the observed widening over the 21st century may result from combined greenhouse gas (GHG) forcing^{1,3,10}, black carbon aerosols and tropospheric ozone in the NH^{11,12}, and stratospheric ozone depletion in the Southern Hemisphere¹³. In the NH mid latitudes, this recent tropical expansion has been partly offset by increasing industrial aerosol (e.g., sulfate) emissions that cause tropospheric cooling¹².

However, coupled climate model simulations show no clear consensus on the primary forcing mechanism of NH tropical expansion over the last four decades⁸ and model-observation comparisons show that neither anthropogenic forcing¹⁴ or natural variability^{15,16} alone can explain it. As a result, the relative importance of anthropogenic versus natural drivers is incompletely understood^{10, 11}. The drivers of an observed multidecadal 1945-1980 contraction of the NH tropical belt^{8,17} are even less well known.

Such shortcomings in our understanding of the factors influencing tropical belt width can lead to a large inter-model spread in simulated future trends¹⁸, and thus to large uncertainties in future hydroclimate projections. Advanced models and more refined observations are needed to reduce such uncertainties and to improve the ability of GCMs to reliably project future trends in tropical belt width. This information will provide a better understanding of the relative roles of prospective drivers, particularly aerosols^{1,11}. Here, we contribute to leveraging this challenge by using tree-ring width data to reconstruct the northern boundary of the tropical belt over the last 800 years (1203-2003 CE).

The target of our reconstruction is the latitudinal extent of the Hadley Circulation (HCE), a commonly used metric that is robust in its representation of tropical belt boundary position^{7,19}. We computed monthly HCE indices for the period 1930-2014 CE, averaged over all longitudes in the NH, based on meridional winds derived from the Twentieth Century Reanalysis V2 (20CR) dataset²⁰ according to the method described in ref.21 and modified by ref.22 ([Fig. S1; Methods](#)).

Northern or southern HCE excursions, particularly during early spring, can significantly impact precipitation patterns near the northern edge of the subtropical

region (~35-40N) and precipitation-sensitive tree-ring data from these locations can thus reflect HCE_{FMA} movements (Fig. 1). We used mid-latitude tree-ring width data sensitive to precipitation or water deficit (e.g. the self-calibrating Drought Severity Index, scPDSI) during the target season (FMA) to compile regional tree-ring chronologies for five NH regions (Tables S1-S2; Methods). The five regions of the NH are located in western and central North America (hereafter W USA and C USA), southeastern Europe (Turkey; hereafter E EU), and western and central Asia (northern Pakistan; hereafter W ASIA and the Tibetan Plateau, hereafter C ASIA, Fig. 1). Fluctuations in FMA NH Hadley Circulation intensity (hereafter HCI_{FMA} , see Methods) can also impact subtropical precipitation patterns, particularly in W USA and C Asia (Fig. S2). However, the five regional tree-ring chronologies, and 93 out of the 96 site chronologies contributing to them, are independent of HCI_{FMA} variability (Table S3a; Table S1) and we therefore consider our reconstruction to primarily reflect variations in HCE_{FMA} .

We developed two regression models to reconstruct HCE_{FMA} based on these five regional tree-ring chronologies. The first model includes all five regional chronologies and extends back over their common period (1532-2003 CE). The second model includes three chronologies (W USA, C USA and C Asia) that extend back to 1203 CE and was used to reconstruct HCE_{FMA} for the period 1203-1531 CE (Fig. 2; Fig. S3). Both models are calibrated against a reanalysis-based HCE_{FMA} target²⁰ over their period of overlap (1930-2003 CE), explain 51% and 45% of the total variance in that target, respectively, and show robust statistical skills (Table S3). The five-chronology model and the three-chronology model were merged to obtain the early spring tropical belt northern boundary reconstruction ($RHCE_{FMA}$) covering the period 1203-2003 CE.

Internal climate variability

Our reanalysis-based HCE_{FMA} target is significantly negatively correlated with the leading modes of sea surface temperature (SST) variability: ENSO (1930-2014 CE, $r=-0.60$, $p<0.001$, for the eastern Pacific FMA NIÑO3.4 SST, and $r=-0.52$, $p<0.001$, for the central Pacific FMA NIÑO4 SST) and PDO (1930-2014 CE, $r=-0.41$, $p<0.001$). This coupling is reflected by the warmer (cooler) SSTs found in the tropical Pacific during southern-most (northern-most) HCE_{FMA} positions (Fig. 1a-b). Our results thus confirm that over the instrumental period (1930-2003 CE), the early spring NH tropical belt boundary tends to contract (expand) during the warm (cool) phases of ENSO and PDO, triggered by an increase (decrease) in the meridional temperature gradient^{21,5,9,6}.

Pacific SSTs also modulate PNA phases, with positive (negative) PNA phases associated with warm (cool) phases of ENSO and PDO²³. As such, our HCE_{FMA} target also correlates significantly negatively with the FMA PNA pattern (1950-2014 CE, $r=-0.56$, $p<0.001$). February to April sea level pressure (SLP) composite maps for southern-most (northern-most) HCE_{FMA} positions reflect the positive (negative) phase of the PNA teleconnection pattern (Fig. 1c-d).

In line with our findings for the 20th and 21st centuries, the early spring NH tropical belt boundary (1203-2003 CE) – here represented by the reconstructed FMA Hadley Circulation extent ($RHCE_{FMA}$)– typically moved southward during warm ENSO (El Niño) and positive PNA phases and northward during cold ENSO (La Niña) and negative PNA phases over the past 800 years. These $RHCE_{FMA}$ -ENSO and $RHCE_{FMA}$ -PNA relationships are reflected in the significantly negative correlations between $RHCE_{FMA}$ and two ENSO reconstructions and one PNA reconstruction over their full period of overlap ($r=-0.20$, $p<0.001$ with NIÑO3.4²⁴ (1301-2003 CE); $r=-0.12$, $p<0.001$ with

NIÑO4²⁵ (1203-2003 CE); $r=-0.20$, $p<0.001$ with PNA²⁶ (1203-1998 CE)). RHCE_{FMA} movements during past positive (negative) ENSO and PNA extremes, defined as the 95th and 90th (5th and 10th) percentiles of the reconstructions²⁴⁻²⁶, are further demonstrated in a superposed epoch analysis (SEA; see [Methods](#)). We found a significant contraction of the RHCE_{FMA} during extreme positive ENSO (El Niño) and PNA phases and a significant expansion during extreme negative ENSO (La Niña) and PNA phases ([Fig. 3a-c](#)). These results are confirmed by a reverse SEA analysis that composites reconstructed ENSO and PNA indices during southern-most and northern-most RHCE_{FMA} extremes ([Fig. S4](#)).

Spectral analyses revealed significant periodicity in the position of the RHCE_{FMA} within interannual (3-8 years) and decadal (10-26 years) bands over the past eight centuries ([Fig. S5](#); [Fig. 2](#)), highlighting the association between the RHCE_{FMA} movements and past ENSO²⁴, PNA, and PDO variability²⁷.

Post-volcanic tropical belt contractions

Past GCM simulations have shown that the northern boundary of the tropical belt is projected to contract after major volcanic stratospheric sulfur injection events¹⁶. Reanalysis data have also shown that increases in stratospheric volcanic aerosols drive a weakening and equatorward shift of the subtropical jets, and the associated HCE²⁸, but direct observational evidence of past post-volcanic tropical belt contractions is lacking.

Volcanic sulfate aerosol injections into the stratosphere block the Earth's surface from incoming solar radiation driving cooling at regional-to-global scales that can persist for up to ten years²⁹, along with changes in atmospheric circulation and precipitation patterns that can persist up to two years^{30,31}.

The tropical eruption of Mount Tambora in 1815, for instance, resulted in the following year 1816 being one of the coolest years recorded for the NH³² that is often referred to as ‘the year without summer’³³. Our reconstruction shows that the RHCE_{FMA} in 1816 was in its southernmost position over the past 800 years.

To put the 1816 NH tropical belt contraction in a longer-term context, we applied SEA to determine the response of RHCE_{FMA} to past major volcanic forcing events (1203-2003 CE). These major volcanic events were defined as: (a) tropical volcanic events, and (b) events with sulfate injections one (1P) or two (2P) times larger than the 1991 eruption of Pinatubo, derived from 3 independent, ice-core based volcanic reconstructions (see *Methods*, [Table S4](#)). We found evidence of a significant contraction of 0.25 to 0.86 degrees latitude one year after tropical events and stronger contractions of 0.41° to 0.99° and 1.25° to 1.56°, one year after 1P and 2P volcanic events, respectively ([Fig 3d](#); [Fig. S6a-b](#)).

We also used SEA to detect post-volcanic effects in the five individual mid-latitude regional chronologies contributing to RHCE_{FMA}. We found significant positive post-volcanic responses in tree growth for the W USA and E EU chronologies ([Fig. S6c-e](#)). The observed increases in tree growth are related to the wet conditions following volcanic events reported for both regions³¹, similar to the positive precipitation anomalies caused by southern HCE positions ([Fig. 1a](#)).

Observational data and model simulations have linked the onset of El Niño (positive ENSO phases) to large tropical or NH eruptions^{25,34}. This is consistent with our results showing significant contractions of the RHCE_{FMA} in response to El Niño events ([Fig. 3a-b](#)), which in turn occur up to two years following volcanic events ([Fig. 3d](#)). Such an El Niño-amplified response can be explained by the thermostat mechanism in the

equatorial Pacific³⁵: the western Pacific warm pool is more sensitive to changes in radiative forcing than the eastern Pacific and will cool relatively faster following a uniform radiative reduction such as those associated with large volcanic events. As a result, the zonal SST gradient in the Pacific will initially decrease and the trade winds driven by this gradient will weaken. These effects will, in turn, lead to a further reduction of the SST gradient, and promote the emergence of El Niño conditions³⁴.

Late 16th century tropical belt expansion

The Little Ice Age (LIA)³⁶, a period from ca. 1450 to 1850 CE, was generally characterized by low atmospheric CO₂ concentrations, cooling in the NH³⁶, and strong tropical volcanic activity (Table S4). Low-resolution paleoclimate records that accurately capture centennial-scale variability indicate that the LIA was generally characterized by El Niño-like conditions³⁷ and coincided with a contraction of the tropical belt³⁸ and a southward movement of the Intertropical Convergence Zone (ITCZ)^{39,40}.

Our reconstruction is limited in its ability to capture the centennial-scale variability of the LIA. It best represents HCE_{FMA} variability at interannual to decadal scales (Fig. 4a; Fig. S5). On these decadal time-scales, we find a six-decade expansion period that occurred at the crux of the LIA. The longest period of persistent tropical belt expansion in RHCE_{FMA} extended from ca. 1568 to 1634 CE, with a northern-most expansion peak in 1624 (Fig. 2b; Fig. 4a).

Several NH regions where dry conditions prevail during HCE_{FMA} expansions, such as the southwestern United States, Northern Mexico, the southeastern United States, Turkey, and eastern China, are documented to have experienced the most severe droughts of at least the last 500 years at the end of the 16th century^{41–43}. In particular,

two extreme droughts during the period from 1587 to 1612 CE in the Carolinas and Virginia of the United States have been linked to the disappearance of the Lost Colony of Roanoke Island and the near abandonment at Jamestown Colony, the first English settlements in North America⁴⁴.

In Turkey, the famine, rural flight, epidemics, and violence associated with a severe 1590s drought gave rise to the Celâlî Rebellion (1596-1610 CE), the most severe crisis in pre-World War I Ottoman history, during which rebel armies plundered the Ottoman provinces in their revolt against the authority of the Ottoman Empire⁴⁵. In Asia, the exceptional drought event that occurred in 1586-1589 CE over eastern China was followed by wide-spread famines and peasant rebellions that marked the end of the Ming dynasty, which finally collapsed in 1644⁴³.

Most of these severe droughts have been recorded as summer droughts in tree-ring based paleoclimate reconstructions that are independent from $RHCE_{FMA}$. They commonly targeted drought indices, such as scPDSI, that integrate short- to long-term effects on the soil water balance, and thus recorded not only drought conditions in summer, but also in the spring months leading up to it⁴¹. As such, a northward movement of the RHCE in early spring (FMA) during the late 16th century, could move storm tracks away from these regions and create prolonged spring but also summer drought conditions. The synchronicity of these historically documented societal crises over the NH can thus be linked to the late 16th century – early 17th century $RHCE_{FMA}$ expansion peak.

The driving force behind this peak, however, is not fully understood. A recent climate model simulation for E EU, one of the regions impacted by HCE_{FMA} (Fig. 1a-b), suggests that internal climate dynamics, rather than external forcings, drove decadal-

scale precipitation variability in the late 16th and early 17th centuries⁴⁵. Such climate dynamical patterns, including ENSO and PDO, have also been linked to the late 16th century extreme droughts across North America⁴¹.

In the late 1500s and early 1600s, ENSO was more negative than average^{24,25} (Fig. 4b) and these La Niña-like conditions in the tropical Pacific could also have contributed to a negative PNA pattern²⁶ (Fig. 4c), and thus be in line with northward RHCE_{FMA} positions (Fig. 4a; Fig. 3a-c). Then, it is likely that the tropical belt expansion of the late 16th century – early 17th century was driven by internal climatic variability, rather than external forcings, such as volcanic eruptions.

Role of anthropogenic forcing

Our RHCE_{FMA} extends the observational period of interannual NH tropical belt movements by more than 700 years and shows that prior to 21st century increases in GHG and anthropogenic aerosol emissions, natural atmospheric variability and volcanic forcing were the primary drivers of early spring NH tropical belt variations.

Over the most recent four decades, our instrumental HCE_{FMA} target shows an expansion trend of 0.4° per decade (1979-2014; Fig. S1c), which is consistent with other recent reanalysis-based expansion rates^{1,5,9}. The correlation between HCE_{FMA} and FMA ENSO and PDO modes also increases over this period (1979-2014 CE, NIÑO3.4 SST $r=-0.71$; NIÑO4 SST $r=-0.6$; PDO $r=-0.61$, $p<0.001$), compared to an earlier period of similar length (1943-1978 CE, NIÑO3.4 $r=-0.3$ SST; NIÑO4 SST $r=-0.3$, $p<0.1$; PDO $r=-0.05$, $p>0.1$).

This strong relationship in recent decades suggests that anthropogenic forcing of internal climate modes⁶, such as the recent increase in the frequency of La Niña events⁴⁶ and the post-1979 change of phase in the PDO⁶, might amplify or temper

projected 21st century expansion of the NH tropics due to increased GHG emissions and decreased aerosol emissions⁴⁷.

In closing, we show that like anthropogenic aerosols⁴⁷, volcanic effects lead to a contraction of the RHCE_{FMA}. We find that past major volcanic events have resulted in a short-lived (1-2 years) contraction of the RHCE_{FMA} and our results should be taken into account when considering the potential implications of proposed solar radiation management approaches. In view of delayed actions on GHG emission reductions, continuous and deliberate injection of aerosols into the stratosphere, thus mimicking the effect of major volcanic eruptions, are increasingly considered in an effort to counteract anthropogenic planetary warming⁴⁸. Our results indicate that, in addition to modeled impacts on regional hydroclimate⁴⁹ and on extreme climate events⁵⁰, such artificial stratospheric aerosol injections could result in tropical belt contractions. These contractions can potentially offset projected GHG-driven tropical expansion⁴⁷, but can have hemispheric-scale hydroclimate implications that command careful consideration.

References

1. Seidel, D. J., Fu, Q., Randel, W. J. & Reichler, T. J. Widening of the tropical belt in a changing climate. *Nat. Geosci* **1**, 21–24 (2008).
2. Seager, R. & Vecchi, G. A. Greenhouse warming and the 21st century hydroclimate of southwestern North America. *Proc. Natl. Acad. Sci.* **107**, 21277–21282 (2010).
3. Lucas, C., Timbal, B. & Nguyen, H. The expanding tropics: a critical assessment of the observational and modeling studies: The expanding tropics. *Wiley Interdiscip. Rev. Clim. Change* **5**, 89–112 (2014).
4. Lucas, C. & Nguyen, H. Regional characteristics of tropical expansion and the role of climate variability. *J. Geophys. Res. Atmospheres* **120**, 6809–6824 (2015).
5. Lu, J., Chen, G. & Frierson, D. M. W. Response of the Zonal Mean Atmospheric Circulation to El Niño versus Global Warming. *J. Clim.* **21**, 5835–5851 (2008).
6. Allen, R. J., Norris, J. R. & Kovilakam, M. Influence of anthropogenic aerosols and the Pacific Decadal Oscillation on tropical belt width. *Nat. Geosci.* **7**, 270–274 (2014).
7. Nguyen, H., Evans, A., Lucas, C., Smith, I. & Timbal, B. The Hadley Circulation in Reanalyses: Climatology, Variability, and Change. *J. Clim.* **26**, 3357–3376 (2013).
8. D’Agostino, R. & Lionello, P. Evidence of global warming impact on the evolution of the Hadley Circulation in ECMWF centennial reanalyses. *Clim. Dyn.* **48**, 3047–3060 (2017).
9. Adam, O., Schneider, T. & Harnik, N. Role of Changes in Mean Temperatures versus Temperature Gradients in the Recent Widening of the Hadley Circulation. *J. Clim.* **27**, 7450–7461 (2014).

- 292 10. Hu, Y., Tao, L. & Liu, J. Poleward expansion of the hadley circulation in CMIP5
293 simulations. *Adv. Atmospheric Sci.* **30**, 790–795 (2013).
- 294 11. Allen, R. J., Sherwood, S. C., Norris, J. R. & Zender, C. S. Recent Northern
295 Hemisphere tropical expansion primarily driven by black carbon and tropospheric
296 ozone. *Nature* **485**, 350–354 (2012).
- 297 12. Wang, H., Xie, S.-P. & Liu, Q. Comparison of Climate Response to Anthropogenic
298 Aerosol versus Greenhouse Gas Forcing: Distinct Patterns. *J. Clim.* **29**, 5175–5188
299 (2016).
- 300 13. Polvani, L. M., Waugh, D. W., Correa, G. J. P. & Son, S.-W. Stratospheric Ozone
301 Depletion: The Main Driver of Twentieth-Century Atmospheric Circulation Changes
302 in the Southern Hemisphere. *J. Clim.* **24**, 795–812 (2011).
- 303 14. Garfinkel, C. I., Waugh, D. W. & Polvani, L. M. Recent Hadley cell expansion: The
304 role of internal atmospheric variability in reconciling modeled and observed
305 trends. *Geophys. Res. Lett.* **42**, 10,824–10,831 (2015).
- 306 15. Johanson, C. M. & Fu, Q. Hadley Cell Widening: Model Simulations versus
307 Observations. *J. Clim.* **22**, 2713–2725 (2009).
- 308 16. Lu, J., Deser, C. & Reichler, T. Cause of the widening of the tropical belt since 1958.
309 *Geophys. Res. Lett.* **36**, (2009).
- 310 17. Brönnimann, S. *et al.* Southward shift of the northern tropical belt from 1945 to
311 1980. *Nat. Geosci.* **8**, 969–974 (2015).
- 312 18. Davis, N. & Birner, T. On the Discrepancies in Tropical Belt Expansion between
313 Reanalyses and Climate Models and among Tropical Belt Width Metrics. *J. Clim.* **30**,
314 1211–1231 (2017).

19. Davis, N. & Birner, T. Climate Model Biases in the Width of the Tropical Belt. *J. Clim.* **29**, 1935–1954 (2016).
20. Compo, G. P. *et al.* The Twentieth Century Reanalysis Project. *Q. J. R. Meteorol. Soc.* **137**, 1–28 (2011).
21. Oort, A. H. & Yienger, J. J. Observed Interannual Variability in the Hadley Circulation and Its Connection to ENSO. *J. Clim.* **9**, 2751–2767 (1996).
22. Nguyen, H. *et al.* Variability of the extent of the Hadley circulation in the southern hemisphere: a regional perspective. *Clim. Dyn.* **50**, 129–142 (2018).
23. Horel, J. D. & Wallace, J. M. Planetary-Scale Atmospheric Phenomena Associated with the Southern Oscillation. *Mon. Weather Rev.* **109**, 813–829 (1981).
24. Li, J. *et al.* El Niño modulations over the past seven centuries. *Nat. Clim. Change* **3**, 822–826 (2013).
25. Liu, Y. *et al.* Recent enhancement of central Pacific El Niño variability relative to last eight centuries. *Nat. Commun.* **8**, 15386 (2017).
26. Liu, Z. *et al.* Pacific North American circulation pattern links external forcing and North American hydroclimatic change over the past millennium. *Proc. Natl. Acad. Sci.* **114**, 3340 (2017).
27. D’Arrigo, R. & Wilson, R. On the Asian expression of the PDO. *Int. J. Climatol.* **26**, 1607–1617 (2006).
28. Haigh, J. D., Blackburn, M. & Day, R. The Response of Tropospheric Circulation to Perturbations in Lower-Stratospheric Temperature. *J. Clim.* **18**, 3672–3685 (2005).
29. Sigl, M. *et al.* Timing and climate forcing of volcanic eruptions for the past 2,500 years. *Nature* **523**, 543–549 (2015).
30. Robock, A. Volcanic eruptions and climate. *Rev. Geophys.* **38**, 191–219 (2000).

31. Iles, C. E., Hegerl, G. C., Schurer, A. P. & Zhang, X. The effect of volcanic eruptions on global precipitation. *J. Geophys. Res. Atmospheres* **118**, 8770–8786 (2013).
32. Wilson, R. *et al.* Last millennium northern hemisphere summer temperatures from tree rings: Part I: The long term context. *Quat. Sci. Rev.* **134**, 1–18 (2016).
33. Luterbacher, J. & Pfister, C. The year without a summer. *Nat. Geosci.* **8**, 246 (2015).
34. Emile-Geay, J., Seager, R., Cane, M. A., Cook, E. R. & Haug, G. H. Volcanoes and ENSO over the Past Millennium. *J. Clim.* **21**, 3134–3148 (2008).
35. Clement, A. C., Seager, R., Cane, M. A. & Zebiak, S. E. An Ocean Dynamical Thermostat. *J. Clim.* **9**, 2190–2196 (1996).
36. Grove, J. The Little Ice Age, 498 pp. *Methuen Lond.* (1988).
37. Mann, M. E. *et al.* Global Signatures and Dynamical Origins of the Little Ice Age and Medieval Climate Anomaly. *Science* **326**, 1256–1260 (2009).
38. Denniston, R. F. *et al.* Expansion and Contraction of the Indo-Pacific Tropical Rain Belt over the Last Three Millennia. *Sci. Rep.* **6**, (2016).
39. Ridley, H. E. *et al.* Aerosol forcing of the position of the intertropical convergence zone since ad 1550. *Nat. Geosci.* **8**, 195–200 (2015).
40. Lechleitner, F. A. *et al.* Tropical rainfall over the last two millennia: evidence for a low-latitude hydrologic seesaw. *Sci. Rep.* **7**, 45809 (2017).
41. Stahle, D. W. *et al.* Tree-ring data document 16th century megadrought over North America. *Eos Trans. Am. Geophys. Union* **81**, 121–125 (2000).
42. Touchan, R., Akkemik, Ü., Hughes, M. K. & Erkan, N. May–June precipitation reconstruction of southwestern anatolia, Turkey during the last 900 years from tree rings. *Quat. Res.* **68**, 196–202 (2007).

43. Shen, C., Wang, W.-C., Hao, Z. & Gong, W. Exceptional drought events over eastern China during the last five centuries. *Clim. Change* **85**, 453–471 (2007).
44. Stahle, D. W., Cleaveland, M. K., Blanton, D. B., Therrell, M. D. & Gay, D. A. The Lost Colony and Jamestown Droughts. *Science* **280**, 564 (1998).
45. Xoplaki, E. *et al.* Modelling Climate and Societal Resilience in the Eastern Mediterranean in the Last Millennium. *Hum. Ecol.* (2018). doi:10.1007/s10745-018-9995-9
46. Cai, W. *et al.* Increased frequency of extreme La Niña events under greenhouse warming. *Nat. Clim. Change* **5**, 132–137 (2015).
47. Allen, R. J. & Ajoku, O. Future aerosol reductions and widening of the northern tropical belt. *J. Geophys. Res. Atmospheres* **121**, 6765–6786 (2016).
48. Caldeira, K., Bala, G. & Cao, L. The Science of Geoengineering. *Annu. Rev. Earth Planet. Sci.* **41**, 231–256 (2013).
49. Ricke, K. L., Morgan, M. G. & Allen, M. R. Regional climate response to solar-radiation management. *Nat. Geosci.* **3**, 537 (2010).
50. Jones, A. C. *et al.* Impacts of hemispheric solar geoengineering on tropical cyclone frequency. *Nat. Commun.* **8**, 1382 (2017).

Corresponding author

Correspondence and requests for materials should be addressed to Raquel Alfaro-Sánchez (r.alfarosanchez@gmail.com).

Acknowledgments

This study was supported by the US National Science Foundation CAREER grant AGS-1349942 and the NSF grant 1065790 from the Emerging Frontiers Section in the

Division of Environmental Biology. RAS is supported by the Postdoctoral grant Juan de la Cierva -Formación-FJCI-2015-26848, from Spanish Ministry of Economy, Industry and Competitiveness. SK acknowledges the support of the USDA-AFRI grant 2016-67003-24944 and RKM acknowledges support from NSF Ecosystems Program grant 1754430. RV is partially supported by BNP-PARIBAS Foundation.

Author contributions

V.T, R.V, and R.A.S conceived and designed the research. R.A.S led the analysis and figure preparation and wrote the manuscript, with major contributions by V.T and H.N. H.N. provided the monthly Hadley circulation indices data for the northern hemisphere. S.K. and A.H contributed to data analysis. N.K. provided tree-ring chronologies. V.T. and R.K.M. organized financial support. All authors contributed to scientific discussions and to the preparation of the manuscript.

Figures

Figure 1 | February to April (FMA) Hadley Circulation extent (HCE_{FMA}) influence on

NH climate variability. CRU TS4.01 FMA precipitation (*shading* every 4 mm) and

HadISST1 Sea Surface Temperatures (*contours* every 0.3 °C) composite maps for the **(a)**

10th percentile (southern-most) and **(b)** 90th percentile (northern-most) HCE_{FMA} years

(1930-2014 CE, $p < 0.05$), with *blue* (*red*) colors indicating wet and cool (dry and warm)

anomalies, respectively. Twentieth Century Reanalysis V2 FMA sea level pressure

composite maps for the **(c)** 10th percentile (southern-most) and **(d)** 90th percentile

(northern-most) HCE_{FMA} years (*shading* every 40 Pa; 1930-2011 CE, $p < 0.05$). Black

boxes at each panel indicate the location of the five regional chronologies used in the

reconstruction models.

409 **Figure 2| Early spring tropical belt northern boundary reconstruction (RHCE_{FMA}).**
410 Annually resolved instrumental (*red* line, HCE_{FMA}) and reconstructed (*black* line,
411 RHCE_{FMA}) February-April Hadley circulation extent for (a) the 1930-2014 CE
412 instrumental period and for (b) the 1203-2003 CE period. *Blue* line represents the 50-yr
413 low-pass filtered RHCE_{FMA}. Dashed horizontal lines indicate the average, 5th and 95th
414 percentile positions of RHCE_{FMA}. Shading around the reconstruction represents the
415 combined chronology and calibration error estimation. (c) Sample depth over time for
416 each of the 5 master chronologies and the periods 1532-2003 CE and 1203-1531 CE.

Figure 3| RHCE_{FMA} response to ENSO and PNA extremes and to major volcanic events. Superposed epoch analyses (SEA) showing the early spring tropical belt northern boundary reconstruction (RHCE_{FMA}) response during extreme negative (positive) El Niño Southern Oscillation (ENSO) and Pacific/North-American pattern (PNA) years, defined as the 5th and 10th (95th and 90th) percentiles of two ENSO reconstructions, (a) NIÑO3.4 SST²⁴ and (b) NIÑO4 SST²⁵, and (c) one PNA reconstruction²⁶. (d) SEA showing the RHCE_{FMA} response to tropical volcanic events (T) and events one (1P) or two (2P) times larger than Pinatubo, based on a revised volcanic forcing reconstruction²⁹. Significant ($p < 0.05$) departures are indicated by circles. The list of the events is in [Tables S4, S6](#).

428 **Figure 4| Tropical belt northern boundary, El Niño Southern Oscillation (ENSO) and**
429 **PNA variability (1203-2003 CE).** Fifty-year smoothed time series of reconstructed (a)
430 early spring tropical belt northern boundary position ($RHCE_{FMA}$), (b) winter NIÑO3.4²⁴
431 and spring NIÑO4²⁵ SSTs, and (c) winter Pacific/North-American pattern (PNA)²⁶. Grey
432 shaded area indicates a period of persistent tropical belt expansion and horizontal
433 lines inside this area indicate the average position of the reconstructions.

Methods

Hadley Circulation

Our reconstruction targets the HCE as a metric for the tropical belt extent in the Northern Hemisphere (NH). Monthly HCE indices, averaged over all longitudes of the NH, were computed according to ref.21 modified by ref.22. The modification consists of computing the zonal mean meridional mass stream function from the zonal mean divergent component of the meridional wind instead of directly from the zonal mean meridional wind. We first decomposed monthly horizontal winds into the divergent components in spherical coordinates, which is the only part of the wind that contributes to the vertical motion and can be regarded as the meridional overturning circulation. The global (zonal mean) mass stream function was computed from the monthly divergent meridional wind component in the same way as ref.21. The HCE is defined as the average position of the 25%-peak value of the global stream function between 400-700 hPa. The results of our methodology agree with the classical definition described in ref.21, with correlations exceeding 0.9 when comparing zonal mean HC edges^{22,51}.

HCE indices were based on horizontal wind data derived from the Twentieth Century Reanalysis V2 (20CR) dataset²⁰. The 20CR dataset is available from 1871 to present, with a spatial resolution of 2° and a monthly temporal resolution. We limited the monthly target time series to period 1930-2014 CE because 20CR can be strongly affected by inhomogeneity in its earliest period due to decline in station density and measurement quality⁵², particularly before the arrival of radiosonde measurements in the 1930s²⁰.

We used monthly Hadley Circulation intensity (HCI) indices, defined by the maximum values of the overturning stream function expressed in kg s^{-1} and derived from the 20CR dataset²⁰, to study the intra-seasonal variability of HCI and its association to HCE position. The Hadley circulation, which is thermally driven, responds to increased global temperatures by decreasing its intensity and increasing its latitudinal extension^{5,7}. Our analysis confirmed previous reanalysis results⁶ suggesting that HC strengthens and contracts during winter and early spring and weakens and expands during summer in the 20CR dataset, but the intra-annual peak of the northern-most HCE position is slightly out of phase with the weakest HCI (Fig. S1a). During early spring (from February to April, FMA), the NH HCE (hereafter HCE_{FMA}) is relatively close to the equator, while FMA NH HCI (hereafter HCI_{FMA}) is relatively high (Fig. S1a), and both, HCI_{FMA} and HCE_{FMA} time series, are inversely correlated with each other (1930-2008 CE; $r = -0.38$, $p < 0.001$; Fig. S1b).

We computed field correlations between monthly HCE indices and precipitation and temperature for the NH (0-80°N), derived from the monthly resolved CRU TS4.01 dataset of gridded ($0.5 \times 0.5^\circ$) climate indices. From the field correlation analyses we determined the fraction of the map (%) with significant correlations between monthly HCE with gridded 0.5° CRU TS4.01 temperature and precipitation data (Fig. S1a).

These field correlations between monthly HCE indices and precipitation for the NH (0-80°N) show a high percentage of grid cells with significant correlations for February and March, probably due to HC intensification during winter and early spring (Fig. S1a). HCE movements during winter also have a significant impact on climate (Fig. S1a). Field correlations between monthly HCE indices and temperature for the NH (0-80°N) show the highest percentage of grid cells with significant correlations for

December and January. However, in the NH, tree-ring proxies that capture winter variability are scarce, particularly temperature-sensitive ones. We thus chose early spring as the target season for our tropical belt northern boundary reconstruction, i.e., HCE averaged from February to April (FMA), HCE_{FMA} . The HCE_{FMA} index correlates significantly (positively, $p < 0.05$) with the HCE indices of January and summer (July, August and September, Table S5).

To assess the influence of interannual HCE variability on NH climate we developed composite maps for 10th (southern-most) and 90th (northern-most) percentiles of HCE_{FMA} . NH climate data included gridded FMA averages of sea level pressure from the 20CR dataset, sea surface temperatures from the HadISST1 dataset, and CRU TS4.01 precipitation data⁵³. Field correlation analysis and composite maps were computed using the KNMI Climate Explorer (<https://climexp.knmi.nl>; ref.54).

We calculated Pearson correlation coefficients on an annual basis, from February to April, between our HCE_{FMA} index and the Pacific/North-American pattern (PNA) time series from the U.S. National Weather Service Climate Prediction Center (CPC; <http://www.cpc.ncep.noaa.gov/>) over the period 1950-2014 CE, and between HCE_{FMA} index and two El Niño Southern Oscillation (ENSO) indices (the Central Pacific HadISST1 NIÑO3.4 and the eastern Pacific HadISST1 NIÑO4⁵⁵) and the Pacific Decadal Oscillation (PDO) index time series (<http://research.jisao.washington.edu/pdo/PDO.latest.txt>, ref.56,57) over three-time windows: 1930-2014, 1943-1978 and 1979-2014 CE.

The significance level for field correlations and time series was set to $p < 0.05$.

Tree-ring data and climate response of regional chronologies

We selected a network of precipitation or drought-sensitive tree-ring chronologies (Table S1) to reconstruct NH HCE_{FMA} based on three criteria: (a) located in a NH region that is climatically influenced by HCE_{FMA} (Fig. 1), (b) sensitive to the same FMA climatic variable that is influenced by HCE_{FMA}, and (c) statistically significantly correlated with HCE_{FMA} over the period of overlap. Of the 96 chronologies that fit these criteria, 78 were derived from the International Tree-Ring Data Bank (ITRDB; <http://www.ncdc.noaa.gov/data-access/paleoclimatology-data/datasets/tree-ring>) and 18 additional chronologies were contributed by the authors (Table S1). The chronologies are clustered in five regions in western and central North America (hereafter W USA and C USA), southeastern Europe (Turkey; hereafter E EU), and western and central Asia (northern Pakistan; hereafter W ASIA and the Tibetan Plateau, hereafter C ASIA; Fig. 1, Table S1).

For each region, we combined the individual, tree-ring series of the contributing chronologies into a regional chronology. Between 6 and 47 site chronologies contributed to each of the five regional chronologies (Table S3a). This chronology development approach allows for improved sample depth further back in time and emphasizes the common signal within a region⁵⁸. Between 150 and 2908 individual tree-ring series were thus compiled per regional chronology (Fig. 2c; Table S3a). Age-related trends were removed from the raw measurements of individual tree-ring series using cubic smoothing splines with a 50% frequency-response cutoff equal to 2/3rds of the length of each series (Table S3a). Regional chronologies were then calculated by averaging the detrended series using a biweight robust mean. The variance of each regional chronology was stabilized to avoid biases due to changing series replication over time⁵⁹. Regional chronologies were then scaled to match their

mean and variance with the target HCE_{FMA} time series over the period of overlap (1930-2003 CE). The period spanned reliably by each regional chronology was determined based on the expressed population signal (EPS): the chronology period was cut off for reconstruction when EPS values fell below the commonly utilized yet arbitrary threshold of 0.85. Using this threshold, all five regional chronologies covered the period 1532-2003 CE and three of the chronologies (W USA, C USA, and C ASIA) extended back to 1203 CE (Table S3a; Fig. 2c).

We assessed the climate responses for the five regional chronology using Pearson correlation analyses (significance level set to $p < 0.05$) with monthly and seasonal precipitation, temperature, and self-calibrating Drought Severity Index (scPDSI) data for the target months of the HCE reconstruction (February to April and one to two months combinations thereof), over the period 1930-2003 CE. Monthly and seasonal climate data were averaged for all CRU grid points within a region defined by the most distant north-south latitude and east-west longitude coordinates of all the individual chronologies contributing to each regional chronology (Table S2). Each of the five regional chronologies correlated significantly with HCE_{FMA} (Table S3a) as well as with regional early spring climate variables that are influenced by HCE_{FMA} (Table S2). The signs of the tree-ring chronology-HCE_{FMA} correlations correspond to the signs of the precipitation-HCE_{FMA} correlations for all regions (Fig. 1), suggesting that the statistical link between tree growth and HCE_{FMA} in these regions is founded in HCE and tree-growth climatology.

Early spring tropical belt northern boundary reconstruction

The five regional chronologies were included in a stepwise multiple linear regression model that was calibrated against the instrumental HCE_{FMA} index over the period 1930-

2003 CE ([Table S3b](#)). This model was used to reconstruct HCE_{FMA} over the period 1532–2003 CE – the period common to all five chronologies. We used the split calibration-verification method to determine the skill of the reconstruction over two sub-periods 1930–1967 and 1968–2003 CE, where above-zero values of Coefficient of Efficiency (CE) and Reduction of Error (RE) statistics indicate that the regression model has acceptable reliability in reproducing the HCE_{FMA} target signal in both sub-periods⁶⁰ ([Table S3c](#)).

Uncertainty in the reconstruction arises from (1) the decreasing number of tree-ring series back through time (replication error) and (2) unexplained variance in the stepwise multiple linear regression model (calibration error, ref.61). We estimated the replication error for each regional chronology by bootstrapping⁶²: standardized tree-ring measurements for every year were sampled with replacement 1000 times and arithmetic means were calculated. Two tailed 95% confidence intervals (CI) were estimated based on the distribution of the bootstrapped mean and the upper and lower limits of the CI were used as the replication error boundaries. The calibration error was estimated as the standard error (SE) for the full calibration period 1930–2003 CE. The overall error of the reconstruction was then estimated as the square root of the standardized summed and squared replication error of each regional master chronology and calibration error terms. The same protocol (including model development, calibration, and uncertainty estimation) was used to develop a second HCE_{FMA} reconstruction using only the three regional chronologies that extended back to 1203 CE ([Fig. S3](#); [Table S3c](#)). The two reconstructions were merged to obtain the early spring tropical belt northern boundary reconstruction ($RHCE_{FMA}$), with the five-

chronology model covering the period 1532-2003 CE and the three-chronology model the period 1203-1531 CE (Fig. 2).

We performed a multi-taper method spectrum analysis⁶³ to identify the RHCE_{FMA} reconstruction's periodicity. A wavelet analysis was also performed using a Morlet transform⁶⁴ on the full reconstruction to determine whether low-frequency behavior of the RHCE_{FMA} was stationary over time. Spectrum and wavelet analyses were performed with KNMI Climate Explorer⁵⁴. Statistical significance was tested (95% confidence intervals) against a red-noise background.

ENSO, PNA and PDO reconstructions

We calculated Pearson correlation coefficients (significance level set to $p < 0.05$) on an annual basis between the RHCE_{FMA} and two ENSO reconstructions and one PNA reconstruction over the maximum period of overlap (1203-2003 CE). Weak agreement between various PDO reconstructions for the period prior to the 20th century²⁷ prevents us from confirming with confidence past coupling between PDO and RHCE_{FMA}. The Central Pacific November through January NIÑO3.4 SST reconstruction²⁴ is based on tree-ring chronologies from both the tropics and mid-latitudes of both hemispheres and overlaps roughly 10% with chronologies used in the RHCE_{FMA}. The eastern Pacific NIÑO4 SST reconstruction²⁵ is based on tree-ring oxygen isotopic time series from Taiwan, it targets the spring season (March-May), and is independent of the RHCE_{FMA}. The winter (December through March) PNA²⁶ reconstruction is based on 29 proxies from the NH, and it shares 2 (out of 96) site chronologies with the RHCE_{FMA}.

We compared lagged relationships between RHCE_{FMA} and ENSO and PNA reconstructions in a superposed epoch analysis (SEA; Fig. 3a-c) using the R package dplR⁶⁵. For this purpose, we superposed an 11-year window of contemporaneous and

lagged RHCE_{FMA} values centered over reconstructed positive and negative ENSO and PNA extremes (event years). Positive (negative) event years in the ENSO and PNA reconstructions^{24–26} were defined as the 95th and 90th (5th and 10th) percentiles of the reconstructions, after processing with a 10-yr high-pass filter to isolate interannual variability (Table S6). Monte Carlo simulations (n=10,000) were used to develop bootstrapped confidence intervals to estimate whether RHCE_{FMA} values were significantly (p<0.05) higher or lower than average in the 11-year window (t-5 to t+5) centered around the ENSO and PNA extremes (t0).

To corroborate our results, we also calculated the inverse SEA, in which we determined reconstructed ENSO and PNA anomalies in an 11-year window centered on the southern-most (northern-most) RHCE_{FMA} events years, defined as the 5th (95th) and 10th (90th) percentiles of the RHCE_{FMA}, after processing with a 10-yr high-pass filter (Table S7, Fig. S4).

Global volcanic forcing

Newly revised ice-core records of volcanic forcing²⁹ and two additional independent volcanic forcing reconstructions^{66,67} that extend back to at least 1203 CE, were used to analyze the response of the early spring tropical belt northern boundary reconstruction to major volcanic stratospheric sulfur injection events over the last eight centuries.

We consider major volcanic forcing events those that either have a tropical source (T) or an estimated sulfate injection that was once (1P) or twice (2P) the size of the eruption of Mount Pinatubo in 1991, a commonly used metric of global volcanic forcing equivalent to -7.5 W m⁻² of volcanic forcing²⁹, 15 Tg of sulfate aerosols⁶⁷ or 11 kg km⁻² of sulfate flux⁶⁶ (Table S4).

Eruption years, volcano names and regions, and volcanic explosivity index (VEI) of major volcanic forcing events listed by ref.29 were assigned to known volcanic events. The match of ice core sulfate signals with volcanic eruptions was mainly based on ref.68. Nine out of the 30 volcanic events listed with a tropical source and 6 out of the 14 volcanic events classified as 1P remain unknown. Only 1 out of the 5 volcanic events classified as 2P remain unknown, the volcanic event of 1230. The other four 2P events correspond to the eruptions of Rinjani (Samalas) in 1258, Kuwae in 1458 (ref.69), Grímsvötn in 1783 and Tambora in 1815 ([Table S4](#)).

When ice core sulfate signals were not matched to a known eruption, eruption latitudes -tropical, Northern Hemisphere (NH) or Southern Hemisphere (SH) source- were assigned based on the presence or lack of simultaneous signals in both Greenland and Antarctic ice cores. Therefore, synchronous sulfate deposition on both polar ice sheets indicate a tropical eruption and signals in only one hemisphere are assumed to be extratropical in origin and were attributed accordingly to NH or SH eruptions²⁹.

We conducted SEA to determine the averaged response of $RHCE_{FMA}$ to major volcanic forcing events in the 11-year window ($t-5$ to $t+5$) centered around these volcanic events (t_0). We first used as a list of events, the tropical volcanic eruptions (T) and eruptions one (1P) or two times (2P) larger than the Pinatubo from the 3 volcanic reconstructions^{29,66,67} ([Table S4](#)). NH and SH source events were also tested in SEA but not reported due to the lack of significance. To investigate the volcanic response of the five contributing regional chronologies to the early spring tropical belt northern boundary reconstruction, we also ran SEA using as event years the tropical volcanic eruptions (T) and eruptions one (1P) or two times (2P) larger than the Pinatubo from the volcanic reconstruction of ref.29.

Data availability

The RHCE_{FMA} will be housed at NOAA-Paleoclimatology/World Data Service for Paleoclimatology: <https://www.ncdc.noaa.gov/paleo/study/25050>.

The data that support the findings of this study are available from the corresponding author upon request.

References

51. Studholme, J. & Gulev, S. Concurrent Changes to Hadley Circulation and the Meridional Distribution of Tropical Cyclones. *J. Clim.* **31**, 4367–4389 (2018).
52. Donat, M. G. *et al.* Reanalysis suggests long-term upward trends in European storminess since 1871. *Geophys. Res. Lett.* **38**, L14703 (2011).
53. Harris, I., Jones, P. D., Osborn, T. J. & Lister, D. H. Updated high-resolution grids of monthly climatic observations – the CRU TS3.10 Dataset. *Int. J. Climatol.* **34**, 623–642 (2014).
54. Trouet, V. & Van Oldenborgh, G. J. KNMI Climate Explorer: a web-based research tool for high-resolution paleoclimatology. *Tree-Ring Res.* **69**, 3–13 (2013).
55. Rayner, N. A. Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J. Geophys. Res.* **108**, (2003).
56. Zhang, Y., Wallace, J. M. & Battisti, D. S. ENSO-like interdecadal variability: 1900–93. *J. Clim.* **10**, 1004–1020 (1997).
57. Mantua, N. J., Hare, S. R., Zhang, Y., Wallace, J. M. & Francis, R. C. A Pacific interdecadal climate oscillation with impacts on salmon production. *Bull. Am. Meteorol. Soc.* **78**, 1069–1079 (1997).

669 58. Trouet, V. A Tree-Ring Based Late Summer Temperature Reconstruction (AD 1675–
670 1980) for the Northeastern Mediterranean. *Radiocarbon* **56**, S69–S78 (2014).

671 59. Frank, D., Esper, J. & Cook, E. R. On variance adjustments in tree-ring chronology
672 development. *Tree Rings Archaeol. Climatol. Ecol. TRACE* **4**, 56–66 (2006).

673 60. Cook, E. R., Meko, D. M., Stahle, D. W. & Cleaveland, M. K. Drought
674 Reconstructions for the Continental United States. *J. Clim.* **12**, 18 (1999).

675 61. Esper, J. *et al.* Long-term drought severity variations in Morocco. *Geophys. Res.*
676 *Lett.* **34**, (2007).

677 62. Briffa, K. R. *et al.* Fennoscandian summers from ad 500: temperature changes on
678 short and long timescales. *Clim. Dyn.* **7**, 111–119 (1992).

679 63. Jenkins, G. . & Watts, D. . Spectral Analysis and its Applications. *Louvain Econ. Rev.*
680 **36**, 554–554 (1970).

681 64. Torrence, C. & Compo, G. P. A practical guide to wavelet analysis. *Bull. Am.*
682 *Meteorol. Soc.* **79**, 61–78 (1998).

683 65. Bunn, A. G. A dendrochronology program library in R (dplR). *Dendrochronologia* **26**,
684 115–124 (2008).

685 66. Crowley, T. J. & Unterman, M. B. Technical details concerning development of a
686 1200 yr proxy index for global volcanism. *Earth Syst. Sci. Data* **5**, 187–197 (2013).

687 67. Gao, C., Robock, A. & Ammann, C. Volcanic forcing of climate over the past 1500
688 years: An improved ice core-based index for climate models. *J. Geophys. Res.* **113**,
689 (2008).

690 68. Sigl, M. *et al.* A new bipolar ice core record of volcanism from WAIS Divide and
691 NEEM and implications for climate forcing of the last 2000 years. *J. Geophys. Res.*
692 *Atmospheres* **118**, 1151–1169 (2013).

693 69. Plummer, C. T. *et al.* An independently dated 2000-yr volcanic record from Law
694 Dome, East Antarctica, including a new perspective on the dating of the 1450s CE
695 eruption of Kuwae, Vanuatu. *Clim. Past* **8**, 1929–1940 (2012).

696

RHCE_{FMA} departures RHCE_{FMA} departures RHCE_{FMA} departures RHCE_{FMA} departures