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Do volcanic eruptions enhance or diminish net primary production? Evidence from tree rings

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[1] Low growth rates of atmospheric CO_2 were observed following the 1991 Pinatubo (Luzon) volcanic eruption. One hypothesis for this CO_2 anomaly is that since diffuse light is more efficiently used by forests than direct light, the increase in the diffuse fraction of sunlight due to scattering by volcanic sulfur aerosol in the years following the eruption substantially increased forest net primary production (NPP). However, other observations suggest a decrease in northern forest NPP because of the cooler conditions following the eruption. Here we used a global database of dated tree ring widths (which correlate with forest NPP) to test this hypothesis. Ice core records of sulfur deposition allowed us to identify the timing and magnitude of 23 Pinatubo-scale eruptions since 1000 CE. We found a significant decrease in ring width for trees in middle to high northern latitudes (north of 45°N) following eruption sulfur peaks. Decreases in tree ring widths were in the range of 2–8% and persisted for ~8 years following sulfur peaks, with minima at around 4–6 years. Ring width changes at lower latitudes in the Northern Hemisphere (30°N to 45°N) and in the Southern Hemisphere (30°S to 56°S) were not significant. In the tropics (30°N to 30°S) the paucity of tree ring records did not permit the evaluation of NPP changes. Given that elevated aerosol levels and summer cooling last only ~2–3 years after an eruption, the persistence of declines in northern tree growth for up to 8 years after eruptions implies some additional mechanism that links these shorter-lived global eruption effects to sustained changes in tree physiology, biogeochemistry, or microclimate. At least for this sample of trees, the beneficial effect of aerosol light scattering appears to be entirely offset by the deleterious effect of eruption-induced climate change. *INDEX TERMS:* 0315 Atmospheric Composition and Structure: Biosphere/atmosphere interactions; 1615 Global Change: Biogeochemical processes (4805); 0305 Atmospheric Composition and Structure: Aerosols and particles (0345, 4801); 8409 Volcanology: Atmospheric effects (0370); *KEYWORDS:* NPP, Pinatubo, tree rings, carbon sink, diffuse light, boreal forest

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1. Introduction

[2] The June 1991 Pinatubo eruption was the century's largest in terms of stratospheric sulfur emissions and effects on global climate [*Hansen et al.*, 1996; *McCormick et al.*, 1995]. It was followed by ~3 years of reduced atmospheric CO_2 accumulation [*Prentice et al.*, 2001]. Concurrent measurements of atmospheric $\delta^{13}C$ provide evidence that the carbon sink was terrestrial [*Battle et al.*, 2000; *Francey et al.*, 2001] while inverse modeling based on remote CO_2 flask measurements implicates the northern middle and high latitudes [*Bousquet et al.*, 2000; *Fan et al.*, 1998; *Rayner et al.*, 1999].

[3] Following the Pinatubo eruption, scattering by sulfur aerosols globally increased the diffuse fraction of incident light with only modest reductions in total light levels [*Molineaux and Ineichen*, 1996]. An increase in diffuse light fraction at constant total light levels is thought to enhance photosynthesis, particularly in forests, by distributing light more evenly among leaves, decreasing the shade volume within the canopy, and thus increasing canopy light use efficiency. Modeling of light use by forest and crop leaves [*Cohan et al.*, 2002; *Sinclair et al.*, 1992] along with eddy covariance based measurements of carbon uptake under cloudy as compared with clear conditions [*Gu et al.*, 2003, 1999] clearly show that instantaneous, canopy-level light use efficiencies are higher for diffuse as compared to direct radiation.

[4] *Roderick et al.* [2001] have hypothesized that the carbon sink following the Pinatubo eruption resulted from a volcano-induced increase in the diffuse fraction of incident sunlight boosting terrestrial photosynthesis. Building

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on the canopy-level studies showing diffuse-light growth enhancement, *Roderick et al.* [2001] estimate a 7% increase in global plant net primary production (NPP), and wood production specifically, as a result of the increase in diffuse light fraction in the year following the Pinatubo eruption; this would roughly match the observed carbon uptake. *Gu et al.* [2003] have explicitly tested this hypothesis using tower eddy covariance measurements of CO_2 uptake in the Harvard forest, finding ~15% increases in noontime photosynthesis rates under clear skies in 1992–1993.

[5] In contrast, satellite measurements of Normalized Difference Vegetation Index (NDVI) suggest that leaf area and NPP in Northern Hemisphere extratropical forests actually decreased following the Pinatubo eruption as a result of cooler summer temperatures [*Zhou et al.*, 2001] and a reduction in the length of the growing season [*Lucht et al.*, 2002], although interference of volcanic aerosols with the satellite observations of vegetation reflectance complicates interpretation [*Myneni et al.*, 1998; *Shabanov et al.*, 2002]. Modeling studies suggest that an enhanced carbon sink in northern biomes after Pinatubo can be consistent with reduced NPP assuming that heterotrophic respiration decreased by an even greater extent than NPP in response to summer cooling [*Jones and Cox*, 2001; *Lucht et al.*, 2002].

[6] Previous studies have used the correlation with climate of tree ring parameters such as width and maximum latewood density to reconstruct climate responses to past volcanic eruptions [*Briffa et al.*, 1998; *D'Arrigo and Jacoby*, 1999; *Lough and Fritts*, 1987]. The existing body of tree ring width chronologies can also be more directly used to test the hypothesis that the increase in diffuse light associated with volcanic events enhances NPP. Tree radial increment is proportional to annual NPP in a variety of forest types [*Gower et al.*, 1992; *Graumlich et al.*, 1989; *Grier and Logan*, 1977; *LeBlanc*, 1996; *Rathgeber et al.*, 2000]. This link with NPP makes the spatially extensive tree ring chronologies from the last millennium a potentially useful means for evaluating the response of NPP to a variety of environmental factors, including volcanic events, at both regional and global scales. Since many tree ring chronologies extend over a period of several centuries, we can obtain additional confidence in any patterns found by assessing growth across multiple volcanic eruptions.

2. Methods

[7] We compiled all dated tree ring width data files from the International Tree Ring Data Bank (ITRDB) that included site longitude and latitude. This yielded 1498 sites, with a median of 25 cores per site, and a total of 43,447 cores in our analysis (Figure 1a). (In the studies contributing to the ITRDB, typically two cores were sampled from each tree [*Schweingruber*, 1988].) The ITRDB [*World Data Center for Paleoclimatology*, 2003] represents data gathered by over 100 different research groups. The largest single block, 31% of the sites we used, came from the Northern Hemisphere temperate and boreal tree ring network developed by *Schweingruber et al.* [1991] for reconstructing regional summer temperatures [*Briffa et al.*, 1998, 2001, 2002]; no other single research group contributed more than 5% of the sites used. Conifers (gymnosperms) accounted for 86% of the sites in our analysis, with *Pinus* (pine), *Picea* (spruce), and *Larix* (larch) respectively representing 30%, 21%, and 7% of the total. Broadleaf trees (angiosperms) accounted for the remaining 14% of the sites, with *Quercus* (oak) the most prevalent genus (7% of total sites).

[8] Width series were first divided by a 41-year moving average to remove tree-age as well as low-frequency climatic effects. All cores from a given site were averaged to produce a non-dimensional width index time series with a mean of 1 [*Fritts*, 1976]. Then all available site indices were averaged to produce ring width indices for site subgroups of interest that extended back as far as 1000 CE, which were again normalized to a unit mean.

[9] Even for the last few centuries, climatically important volcanic eruptions are incompletely known, and the different published compilations differ somewhat; a volcanic aerosol "dust veil" often cannot be assigned to a historically known eruption [*Bradley and Jones*, 1995]. However, sulfate levels in ice cores, which show pronounced peaks as aerosol from a large eruption is deposited, can provide the basis for consistent estimates of past volcanic aerosol levels [*Zielinski*, 2000]. Points within an ice core can be dated to the year by counting annual layers downward from the surface, either visually or using the annual cycle in such ice properties as $\delta^{18}O$. (For a comprehensive discussion of stratigraphic dating of ice cores, see *Alley et al.* [1997].)

[10] We used the time series of annual mean Northern Hemisphere 550-nm optical depth since 1000 CE of *Crowley* [2000] (http://www.ngdc.noaa.gov/paleo/pubs/crowley.html) to identify eruption years. This time series was derived primarily from high-resolution ice core sulfate measurements calibrated against atmospheric observations after modern eruptions. Eruption years were defined as those that showed a peak in volcanic aerosol forcing; this was often the year after the actual eruption implicated, for example, 1992 for Pinatubo (Table 1). Many of these eruption years, and all of the eruption years before 1500, do not correspond to well-dated known large eruptions (Table 1), presumably because of the incompleteness of the historical record of volcanism. Using an optical depth estimate of 0.1 as a threshold (Pinatubo peak depth in this series was 0.123) we obtained 23 eruption years during the period 1000–1970, or 2.4 per century (Table 1). Note that width changes after Pinatubo itself (and the 1982 El Chichón eruption) were not included in our averaging of ring width index departures across eruption years because ring widths were not available for a long enough period after these eruptions to allow the same filtering scheme to be used.

[11] The possibility of missing or double-counting annual layers introduces the potential for error in ice core chronologies, so that the uncertainty in dating increases going back in time from the known surface date. The *Crowley* [2000] time series of Northern Hemisphere sulfate aerosol levels since 1000 is based primarily on measurements from two extensively studied Greenland ice cores, Crete and GISP2 [*Hammer et al.*, 1980; *Zielinski*, 1995]. An idea of the accuracy of the dating of these cores over this period

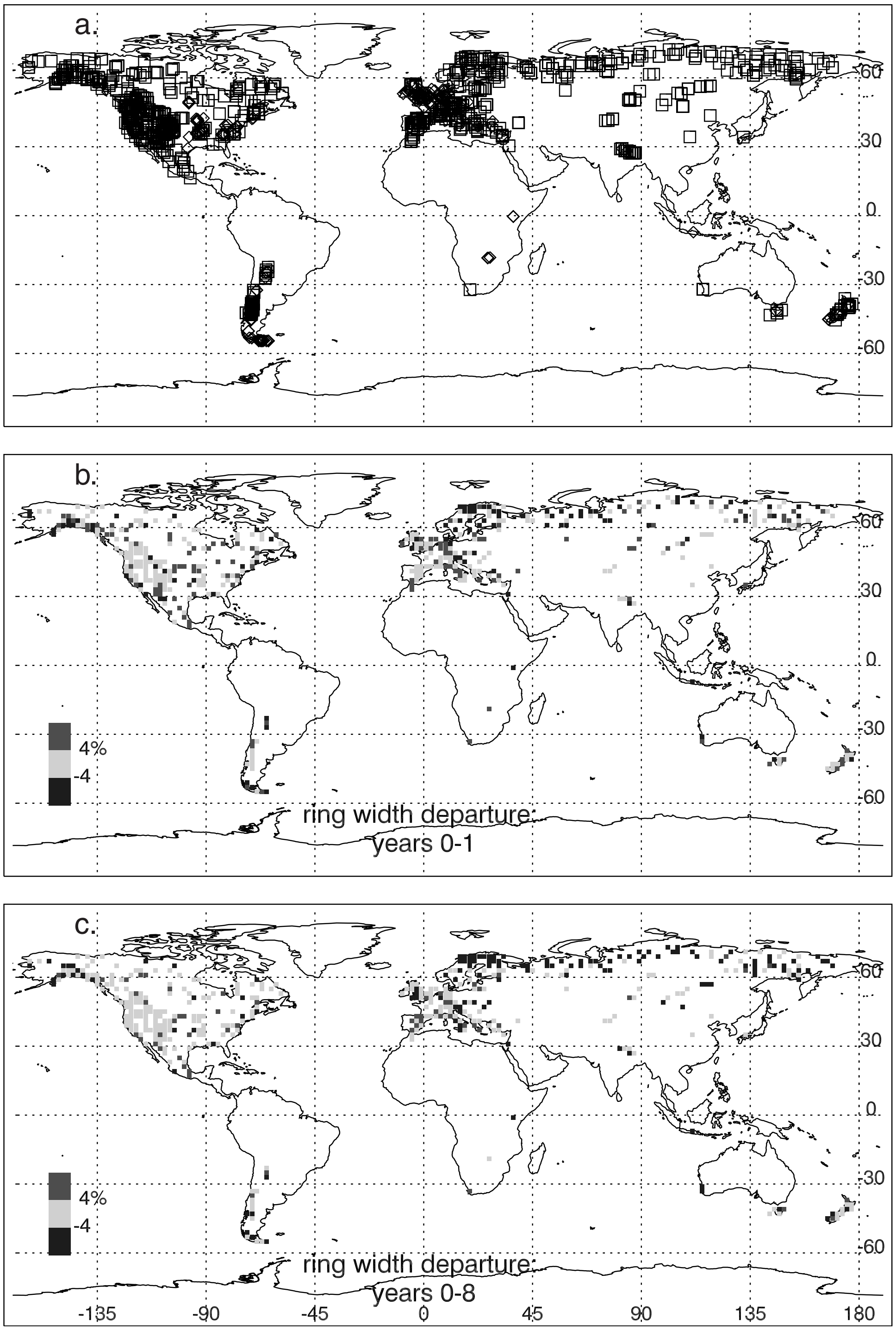

Figure 1. (a) Location of sites used for this analysis (n = 1498). Symbols indicate tree type: needleleaf (squares) or broadleaf (diamonds). (b) Ring widths for years 0 through 1 following eruption years compared with the long-term average. Site responses were averaged over 2° × 2° cells. (c) Ring widths for years 0 through 8 following eruption years compared with the long-term average. Note the preponderance of negative width departures at middle and high northern latitudes. See color version of this figure at back of this issue.

Table 1. Eruption Years Used in This Study (Defined as Years With High Atmospheric Levels of Volcanic Aerosol) and the Possible Responsible Historical Eruptions

	Optical Depth[b]		Eruption Description[c]					
Eruption Year[a]	NH	SH	Volcano	Locality	Beginning Month/Year	Latitude	Longitude	VEI[d]
1026	0.11		Sheveluch	Kamchatka	1000 ± 50	56.7N	161.4E	5
			Billy Mitchell	Bougainville	1030 ± 25	6.1S	155.2E	5
1058	0.15		Baitoushan	China	1050 ± 10	42.0N	128.1E	7
1175	0.23		Okataina	New Zealand	1180 ± 20	38.1S	176.5S	5
1229	0.17							
1259	0.39							
1285	0.13							
1295	0.12							
1329	0.10							
1453	0.15		Kuwae	Vanuatu	1452 ± 10	16.8S	168.5E	6
1460	0.15							
1586		0.19	Kelut	Java	1586	7.9S	112.3E	5
1587	0.15							
1594		0.09	Raung	Java	1593	8.1S	114.0E	5
1600		0.16	Huaynaputina	Perú	2/1600	16.6S	70.9W	6
1601	0.18							
1622	0.10							
1641	0.18		Komaga-take	Japan	7/1640	42.1N	140.7E	5
			Parker	Phillipines	12/1640	6.1N	124.9E	5
1674	0.11		Gamnokara	Halmahera	5/1673	1.4N	127.5E	5
1693		0.10	Hekla	Iceland	2/1693	64.0N	19.7W	4
			Serua	Indonesia	6/1693	6.3S	130.0E	4
1695	0.12	0.14						
1729	0.11							
1783	0.11	0.09	Grímsvötn	Iceland	5/1783	64.4N	17.3W	4
			Asama	Japan	5/1783	36.4N	138.5E	4
1809	0.18							
1810		0.28						
1815	0.20	0.67	Tambora	Sundas	4/1815	8.3S	118.0E	7
1831	0.16		Babuyan Claro	Phillipines	1831	19.5N	121.9E	4
1836		0.19	Cosiguina	Nicaragua	1/1835	13.0N	87.6W	5
1883	0.12		Krakatau	Indonesia	8/1883	6.1S	105.4E	6
1884		0.09						
1902	0.12		Santa Maria	Guatemala	10/1902	14.8N	91.6W	6
1951		0.09	Ambrym	Vanuatu	12/1950	16.2S	168.1E	4
			Lamington	New Guinea	1/1951	9.0S	148.2E	4
1963		0.10	Agung	Bali	3/1963	8.3S	115.5E	4
1983	0.10		El Chichón	México	3/1982	17.4N	93.2W	5
1992	0.12	0.12	Pinatubo	Phillipines	6/1991	15.1N	120.4E	6

[a]Italics denote Northern Hemisphere volcanic aerosol peak, boldface denotes Southern Hemisphere.
[b]NH: Northern Hemisphere, SH: Southern Hemisphere. Optical depths are from *Crowley* [2000] and *Robertson et al.* [2001], respectively, except for the 1992 Southern Hemisphere value which is from *Sato et al.* [1993], and were used to define the eruption years in column 1.
[c]Likely contributing eruptions compiled from *Simkin and Siebert* [1994] as updated at http://www.volcano.si.edu/gvp/world/. Note that in many cases the most important contributing eruption is uncertain or unknown; the location and date of a contributing eruption were not directly used in compiling the list of eruption years in column 1, which is based on ice core rather than historical eruption records.
[d]VEI = Volcanic Explosivity Index [*Simkin and Siebert*, 1994]; 4 is "large"; 5 and up "very large."

can be obtained from considering the very large mid-thirteenth century sulfur peak from an unidentified tropical eruption, which is accompanied by distinctive volcanic tephra in ice cores from both hemispheres [*Palais et al.*, 1992]. The highest sulfate concentrations from this event are in the 1259 layer for both cores, as well as for at least two other arctic cores independently dated by counting layers [*Langway et al.*, 1988; *Zielinski*, 1995]. This suggests that drift between the two cores and between each core and calendar years is ≤ 1 year at least as far back as the thirteenth century.

[12] For Southern Hemisphere sites, we used Southern Hemisphere volcanic-aerosol optical depths from the time series of *Robertson et al.* [2001] (http://www.ngdc.noaa.gov/paleo/pubs/robertson2001/robertson2001.html). This time series extends back to 1500 and was also derived primarily from ice core records. For this time series, we used a threshold optical depth of 0.085 because this time series' value for Northern Hemisphere post-Krakatau forcing is 15% lower than that from *Crowley* [2000]. This threshold yielded 12 Southern Hemisphere eruption years (Table 1).

[13] Standard errors for width departures following eruptions were calculated from the distribution of width departures across eruption years. Since the distribution of ring widths tends to be skew rather than normal [*Schweingruber*, 1988], departures were tested for significance using a Monte Carlo approach. For the same group of sites, ring width departures were separately calculated (using the same filtering and averaging approach described above) for 1000 different scenarios in which random groups of years were substituted for the actual eruption years. This

Table 2. Ring Width Departures Following Eruption Years

Site Grouping	Number of Sites	Number of Cores	Ring Widths, % Difference From Average (±SE): Equal Weighting[a], Years 0–1	Equal Weighting[a], Years 0–8	Weighting by NPP[b], Years 0–1	Weighting by NPP[b], Years 0–8
All N of 60°N	259	7948	−3.4 ± 2.6	−5.0[c] ± 1.4	−4.0[d] ± 2.7	−5.5[c] ± 1.5
All 45°–60°N	465	12182	0.0 ± 0.9	−1.9[d] ± 0.8	0.0 ± 0.9	−2.1[d] ± 1.0
All 30°–45°N	527	15893	0.5 ± 1.2	0.4 ± 0.9	1.1 ± 0.9	0.4 ± 0.8
All 30°–45°S	144	4242	2.1 ± 1.5	−0.9 ± 1.3	3.0 ± 1.4	−0.6 ± 1.5
All 45°–56°S	31	890	−0.4 ± 3.1	−3.5 ± 1.3	0.6 ± 3.1	−3.3 ± 1.8
			North of 45°N			
Spruce	267	7226	−1.3 ± 2.5	−2.9[c] ± 1.7	−1.5 ± 2.0	−2.7[c] ± 1.4
Pine	148	4612	−1.5 ± 1.5	−3.1[c] ± 0.7	−0.7 ± 1.5	−3.1[c] ± 0.9
Larch	104	3090	−6.3[d] ± 2.8	−7.2[c] ± 1.8	−7.0[c] ± 2.7	−7.2[c] ± 1.8
Oak	72	1826	2.6 ± 2.2	−0.7 ± 2.1	2.0 ± 1.9	−0.4 ± 2.2
All trees	723	20101	−1.6 ± 1.3	−3.2[c] ± 0.8	−1.3 ± 1.3	−3.1[c] ± 0.9

[a]Average of sites within a grouping, weighted equally.
[b]Sites averaged by 1° × 1° cell and weighted by contemporary NPP per unit area [*Randerson et al.*, 1997] for that cell.
[c]With p ≤ .01 (probabilities derived from Monte Carlo resampling).
[d]With p ≤ .05.

approach provides an estimate of the probability that a given width anomaly following eruptions would arise by chance. We used p = 0.05 (two-tailed) as the significance level.

[14] We combined all sites within 15° wide latitude zones to produce zonal ring width time series before and after eruption events. Since sites were concentrated in the northern extratropics (Figure 1a), we excluded the tropics (30°N–30°S). Ring width departures in percent are reported in Table 2 for two combinations of years: years 0 to 1 relative to the eruption year, when atmospheric aerosol levels and thus enhancement of diffuse radiation should have been greatest [*Roderick et al.*, 2001], and years 0 to 8, to examine longer-term effects on tree growth.

3. Results

[15] For years 0–1, trees north of 60°N showed a trend toward narrower widths that was significant when sites were weighted according to regional levels of NPP but not when all sites within the latitude zone were weighted equally (Table 2; Figure 2a). No other zone examined showed significant year 0–1 width departures (Table 2; Figure 1b).

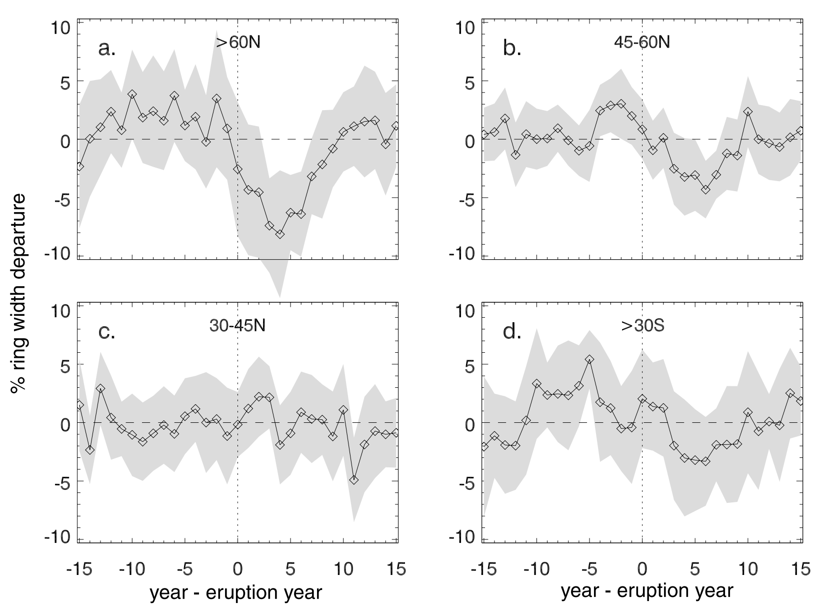

Figure 2. (a–d) Mean ring width departures around eruption years, by latitude zone (with the 30°S–45°S and 45°S–56°S zones combined). Shading shows 2-standard error confidence limits based on the variation across eruption years. Because width departures are expressed relative to a centered moving average, zones that have substantial negative departures following eruption years tend to show positive departures preceding eruption years. See Table 2 for number of sites and significance levels of departures for each zone. Table 1 lists the eruption years used.

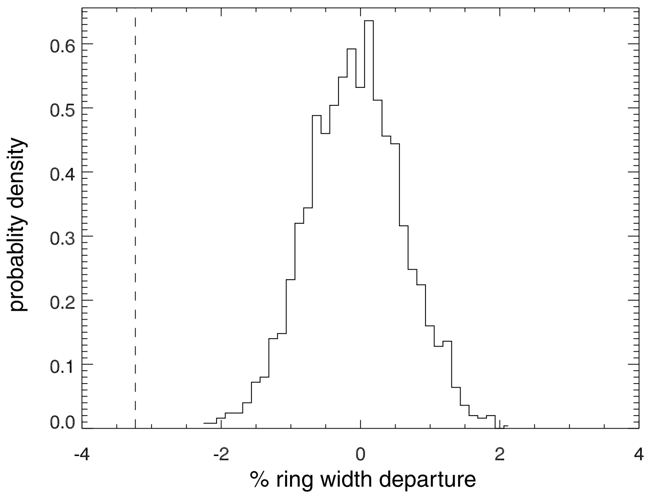

Figure 3. An example of the Monte Carlo analysis used to test regional width departures for significance: width departure for sites north of 45° N for years 0–8 following eruption years (dashed line) as compared to the distribution of mean widths from the same sites following 1000 sets of random "eruption years" (histogram). Here the negative departure is significant, $p < 0.001$.

[16] Significantly narrowed tree rings (implying decreased levels of NPP) were found in regions north of 45°N for the period 0–8 years after eruption years (Figure 3; Table 2). Trees north of 60°N showed a multiyear decline in mean ring width following eruptions, with the maximum reduction in ring width (8.1 ± 2.7%) occurring in year 4 and an average decrease in years 0–8 of 5.0 ± 1.4% (Figure 2a). Sites 45°N–60°N showed a similar pattern but with smaller amplitude: The maximum reduction in ring width (4.3 ± 1.2%) occurred in year 6, and the average decrease in years 0–8 was 1.9 ± 0.8% (Figure 2b). In the two remaining zones, 30°N–45°N and south of 30°S, the ring width departures for years 0–8 were not significant (Table 2), nor do the individual years show any clear anomalies (Figures 2c and 2d).

[17] Among trees north of 45°N, we found significant multiyear width reductions for a number of individual conifer genera (Figures 4a–4c, Table 2). However, *Quercus*, the primary broadleaf genus in this zone in our data set, did not show any significant ring width departures (Figure 4d). Width reductions were also widely distributed longitudinally, though they appear to be greater over Eurasia than over North America (Figures 1b and 1c).

[18] Most of the individual Northern Hemisphere eruption years we considered were associated with negative ring width departures for trees north of 45°N (20 out of 23;

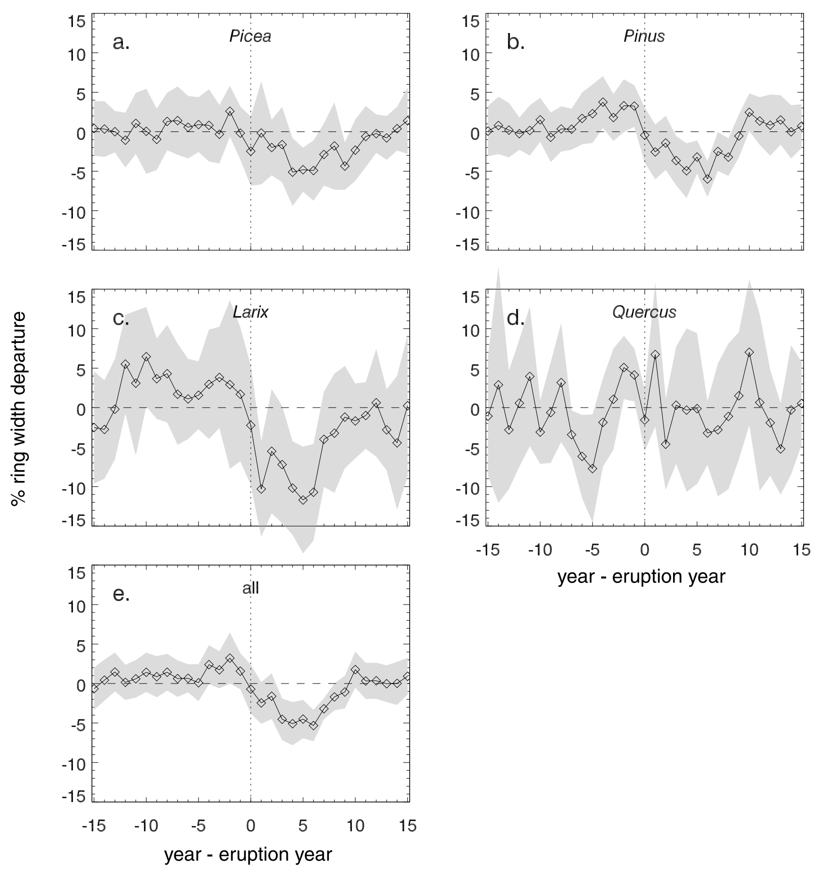

Figure 4. Mean ring width departures around eruption years among sites north of 45° N, for (a–d) the most common tree genera in the database and (e) the mean of all sites north of 45° N. Shading shows 2-standard error confidence limits based on the variation across eruption years. See Table 2 for number of sites and significance levels of departures for each group. Table 1 lists the eruption years used.

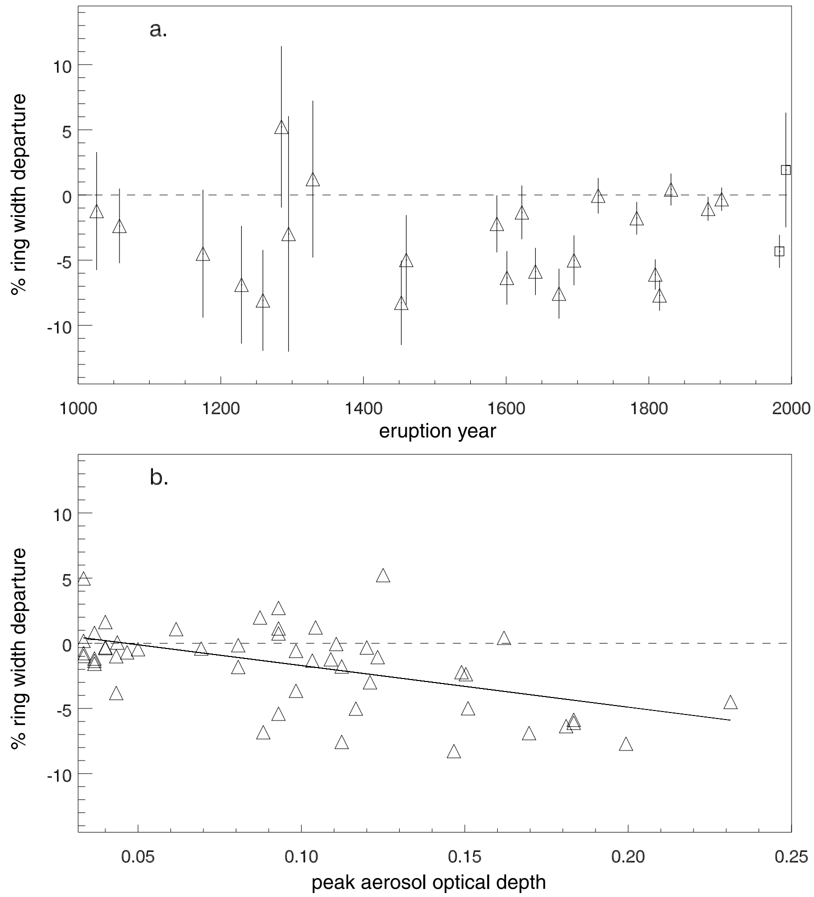

Figure 5. Year 0–8 ring width departure, mean of sites north of 45°N, by individual Northern Hemisphere eruption year. (a) Triangles: years considered in our averaging of width departures across eruptions (optical depth > 0.1; n = 23) plotted by date. Error bars show standard error across sites. These are larger for the earliest eruptions considered because fewer site chronologies exist for the early part of the period. Squares: the Chichón and Pinatubo eruptions, with ring widths normalized against the last 41 years of data for each tree rather than the centered 41-year moving average used for earlier eruptions. Note that the pattern of a width decrease after eruptions is consistent across centuries. (b) For all aerosol optical depth peaks with magnitude 0.033–0.250 (n = 51) as a function of the optical depth. High peak optical depth correlates with negative width departures. A linear fit of the relationship is drawn: $y = 1.5 - 31.9x$, weighted $R^2 = 0.36$ ($p < 0.01$ (Student's t test)).

Figure 5a). Over a range of peak optical depths [from *Crowley*, 2000], there was a positive correlation between the degree of width reduction north of 45°N and the peak optical depth (Figure 5b). Although we did not include the recent Chichón and Pinatubo eruptions in our analysis of response patterns across eruption years, ring widths around these eruptions appear to be broadly consistent with the response patterns observed after earlier eruptions (Figure 5a).

[19] In the earlier part of our study period there are fewer site chronologies available (75% of sites in our data set have widths for 1800, 17% for 1500, 5% for 1200, 3% for 1000) and eruption magnitudes and timings are less well known; however, restricting our analysis to only the eruption years since 1500 that correspond to historically known large eruptions in Table 1 (10 Northern Hemisphere, 11 Southern Hemisphere eruption years) results in the same general pattern of width reductions at mid- to high-northern latitudes (graphs in supplementary material[1] figures; compare Figure 5a).

4. Discussion

4.1. Does Diffuse Light Enhance Forest NPP?

[20] We found no increase in NPP in our data set immediately following eruptions over the past millennium. Our findings suggest that for extratropical trees, any diffuse light growth enhancement is offset by other, deleterious consequences of eruptions, such as summer cooling and a decrease in the length of the growing season.

[21] Tree rings provide a method of assessing NPP changes after eruptions that complements field-level eddy covariance measurements of net carbon uptake such as

[1]Supporting materials are available at ftp://ftp.agu.org/apend/gb/2003GB002076.

those which provided evidence for an increase in photosynthesis rates following Pinatubo [*Gu et al.*, 2003]. While eddy covariance measurements provide near-instantaneous rates of net ecosystem carbon exchange, a tree ring provides a measure of plant growth integrated over weeks to months. There are a number of possible reasons that tree ring analysis and eddy covariance might lead to different conclusions about the effect of eruptions on forest carbon uptake. Modeling of forest diffuse light response suggests that increased aerosol scattering may enhance photosynthesis under clear conditions but impede photosynthesis on cloudy days [*Cohan et al.*, 2002]. Even assuming substantial overall diffuse light enhancement of photosynthesis, aerosol-induced climate change such as nocturnal warming [*Roderick and Farquhar*, 2002] raising nighttime plant respiration and overall cooling resulting in delayed onset and early end of the growing season may lead to a smaller increase, or a decrease, in annual NPP. As proxies for NPP, tree rings have the additional advantages of straightforward replication and extension, permitting assessment of NPP changes over large spatial and temporal distances, and of providing a means to evaluate the effect of eruptions on individual tree species and habits within a canopy.

[22] Using tree rings as a proxy for NPP following eruptions also has limitations. Diffuse light growth enhancement would be expected to be most pronounced for closed-canopy forests with high leaf area indices [*Roderick et al.*, 2001], and presumably especially significant for understory plants with more leaves shaded from direct light. The format of the ITRDB did not allow us to group trees by canopy leaf density, and tree ring chronologies measured for climate reconstructions generally use dominant rather than understory trees [*Fritts*, 1976]. Analyses of ring width series carefully chosen to represent different canopy leaf area indices and tree positions within the canopy may be able to build on our findings and more sensitively assess the magnitude of diffuse-light growth enhancement on a forest-wide basis. Also, few ring chronologies are available for tropical forests. The development of more tropical ring chronologies [*Worbes*, 1999] may in the future allow more comprehensive assessment. Finally, the magnitude of growth enhancement that an analysis like ours can detect is limited by interannual variability in ring width due to other factors. While our uncertainties are small enough that we would likely have detected the 7% increase in global wood production lasting 1–2 years after Pinatubo-scale eruptions that *Roderick et al.* [2001] hypothesized, our analysis did not have the power to detect a smaller growth enhancement or a merely regional one.

[23] An additional source of uncertainty in our analysis comes from the volcanic aerosol time series used to construct our list of eruption years. Even assuming no error in the ice core datings, sulfate levels from any one core reflect imperfectly eruption sulfur emissions. Sulfur deposition over the Antarctic after Pinatubo varied ~20% even over a few kilometers [*Jihong and Mosley-Thompson*, 1999], and sulfate loading for well-documented eruptions such as Novarupta 1912 is undetectable in some Greenland cores while substantial in others [*Zielinski*, 1995; *Zielinski et al.*, 1997]. An index of past aerosol optical depth that includes information from more ice cores measured at high resolution along with atmospheric observations (e.g., of eclipses [*Stothers*, 2002]) is desirable and would likely improve our precision in detecting volcanic aerosol effects on tree growth.

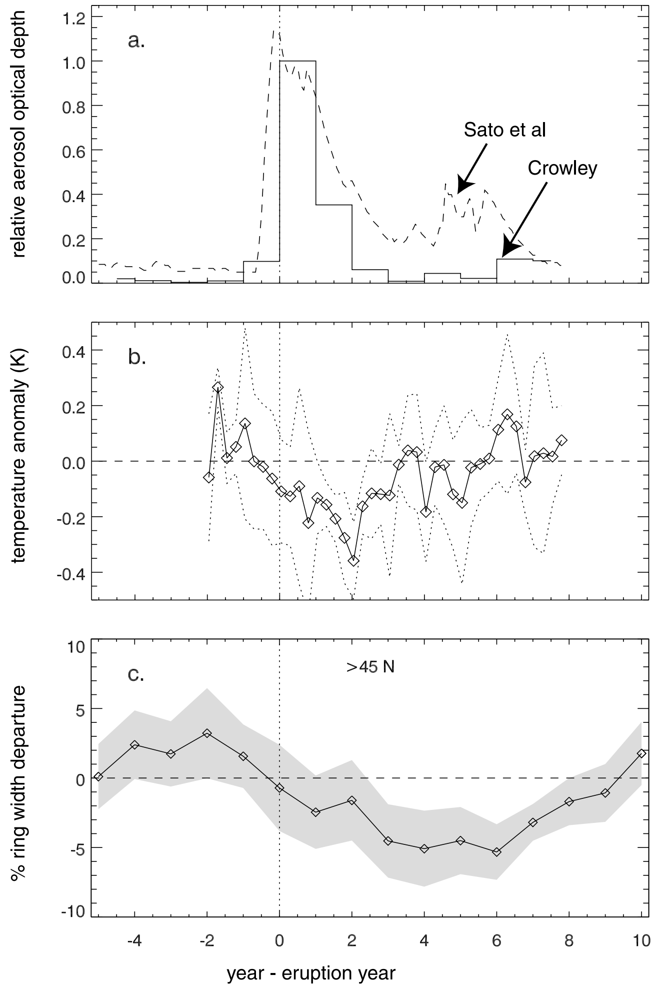

Figure 6. Timescale of responses after large eruptions. (a) Aerosol optical depth as a fraction of that for the eruption year, mean of 25 eruptions since 1000 from *Crowley* [2000] (Northern Hemisphere) and globally by month median of five large eruptions since 1880 from *Sato et al.* [1993] (http://www.giss.nasa.gov/data/strataer/). The peak shows an initial approximately 1-year decay time, although a fraction of the aerosols appears to last several years. (b) Solid line: mean seasonal Northern Hemisphere surface temperature anomalies around the five eruptions since 1880, calculated from the temperature compilation of *Hansen et al.* [1999] (http://www.giss.nasa.gov/data/update/gistemp/). Dotted lines: mean ±1 standard deviation. (c) Ring widths among trees north of 45°N for eruptions since 1000 (this study): same data as Figure 4e. Comparing Figures 8b and 8c, the lag between the peak post-volcanic temperature and ring width departures can be seen.

[24] Eddy covariance and tree rings each have advantages and drawbacks in measuring forest parameters relevant to uptake of atmospheric carbon. Ultimately, combining tech-

niques such as eddy covariance that measure instantaneous carbon fluxes with techniques such as tree ring analysis that retrospectively measure biomass accumulation offers the best prospects for understanding the full impacts of volcanic eruptions and changes in diffuse light fraction on photosynthesis rates, NPP, and carbon uptake.

4.2. Why is There a Decadal-Scale Decrease in Tree Growth Following Eruptions?

[25] Ring narrowing for $\geq$10 years following eruptions has been reported at individual high-latitude and alpine sites [*Gervais and MacDonald*, 2001; *Scuderi*, 1990]. This study shows that narrower rings occur after eruptions across a broad array of mid- and high-latitude northern trees with the period of decreased growth lasting considerably longer than would be expected from the initial aerosol radiative forcing.

[26] Trees that grow in cool-summer climates are often temperature-limited, so that their ring widths correlate well with annual mean and particularly summer temperatures. Invoking the cooling observed in the summers following eruptions [*Groisman*, 1992; *Hansen et al.*, 1996; *Robock and Mao*, 1995] as the major cause for ring width departures following eruptions explains the latitudinal distribution of ring narrowing well. A reduction in temperature would not reduce growth in more temperate sites that are not temperature-limited, such as most of those in the 30°N–45°N band. In fact, for trees in this band that grow in more xeric climates, cooling might result in growth enhancement due to a reduction in water stress, although we do not find significant growth enhancement for the band as a whole after eruptions.

[27] Unlike trees in the 45°N–60°N band, trees at the equivalent latitudes in the Southern Hemisphere did not show a significant growth reduction (Table 2). This could be due to reduced severity of post-eruption cooling in the southern as compared to the northern midlatitudes due to greater marine influence on land climate. However, since the departures for the two bands are the same within error, an interhemispheric difference remains to be demonstrated.

[28] Volcano-derived stratospheric aerosols have an initial e-folding time of $\sim$1 year (Figure 6a) [*Robock*, 2000], and large eruptions result in reductions in measured global temperatures lasting 2–3 years (Figure 6b) [*Sear et al.*, 1987] and changes in $\delta^{18}O$ of Greenland snow indicating cooling for $\sim$2 years following the deposition of volcanic sulfur [*Stuiver et al.*, 1995]. The effect found here on ring width is longer lasting, peaking at $\sim$4–6 years and lasting on average for 8 years (Figure 6c). This raises the question of whether post-eruption cooling can explain this effect. To do so would require either cooling lingering more than $\sim$3 years after eruptions or a lag of several years between cooling and tree growth reduction.

[29] The climate impact of eruptions is not uniform but has distinctive seasonal and regional patterns [*D'Arrigo and Jacoby*, 1999; *Groisman*, 1992; *Lough and Fritts*, 1987; *Robock and Mao*, 1995], and it is possible that some mid- and high-latitude cooling can last considerably longer than 2–3 years. Thus, an examination of glacier movements has suggested that eruptions trigger decade-long episodes of glacial advance [*Porter*, 1986]. A spectral analysis of global temperature fields since 1950 raises the possibility that eruptions cause long-term climate perturbations through persistent El Niño-like ocean temperature patterns [*Lee and Fang*, 2000], and a long-lasting ocean effect is consistent with models of the land-ocean response to radiative forcing [*Lindzen and Giannitsis*, 1998]. Up to decadal-scale ocean cooling tentatively linked to volcanic eruptions has also been found in a study of South Pacific coral [*Crowley et al.*, 1997]. Nevertheless, it appears puzzling that the maximum growth reduction seen in this study lags by several years the period of maximum eruption cooling.

[30] Tree wood increment is often largely based on previous-year productivity, so that a period of distinctive climate conditions will affect ring width for several years afterward [*Fritts*, 1976]. For evergreen needleleaf trees, the multiyear needle lifetime means that the effect of previous-year climate is particularly pronounced [*Jacoby et al.*, 1996]. However, the large width reduction found for larch (Figure 4c), which are deciduous [cf. *Colenutt and Luckman*, 1996], implies that needle retention cannot be the primary explanation for the response lag.

[31] For trees growing on permafrost, permafrost encroachment on tree roots following even one or two abnormally cool summers [*Romanovsky and Osterkamp*, 1997] may persist in subsequent years due to the long response time of soil temperature to surface temperature changes. Shallower thaws may cause water and nutrient stress because more of the trees' roots remain frozen during the growing season. This is an attractive explanation for the particularly large width reduction observed in larch in our data set, many of which grew within the Siberian permafrost zone [cf. *Kobak et al.*, 1996]. In very moist environments, cool summers may similarly lead to nutrient limitation following root damage from waterlogging, as has been suggested for Irish oaks [*Baillie and Munro*, 1988].

[32] Carbon in wood deposited from trees showing drought stress due to permafrost encroachment following eruptions would be predicted to show reduced discrimination against ^{13}C [e.g., *Hubick and Farquhar*, 1989] in rings following eruption years. Similarly, nutrient stress due to permafrost encroachment or flooding might be expected to result in reduced discrimination against ^{15}N [e.g., *McKee et al.*, 2002]. Variations in tree ring carbon and nitrogen isotope ratios have been used to deduce changes in tree water and nutrient status in response to such factors as arctic warming, increased ambient CO_2, and acid rain [*Barber et al.*, 2000; *Bert et al.*, 1997; *Leavitt and Long*, 1991; *Penuelas and Estiarte*, 1997; *Tang et al.*, 1999]. Measurements of the composition of rings formed around volcanic events can thus potentially test whether drought stress, nutrient stress, or other factors are significantly affecting tree physiology following eruptions.

[33] The work of *Braswell et al.* [1997], which shows a 2-year lag between changes in mean surface temperature and NDVI-derived vegetation extent in a number of biomes, and a similar lag between changes in mean surface temperature and the reduction of CO_2 growth rates at Mauna Loa and the South Pole, suggests a positive correlation between temperature and lagged NPP. A 2-year or perhaps somewhat longer lag between temperature and NPP changes for

northern forests is consistent with our findings as well. *Braswell et al.* [1997] suggest temperature-sensitive heterotrophic nutrient cycling as a mechanism by which a temperature change leads to a delayed change in NPP. Analyses of tree ring isotopic composition may be able to test whether cooling indeed affects tree nutrient status.

5. Conclusions

[34] In our analysis of tree ring widths, including over 40,000 cores from more than 1000 sites, we find evidence for a decrease in northern forest NPP following eruptions. This decrease appears at least initially to be a consequence of the surface cooling effects of volcanic aerosols. Some additional feedback mechanism is required to explain the observed persistence of this reduced growth for several years beyond the cooling peak.

[35] Our finding of no significant increase in wood accumulation around eruption years favors lower soil respiration rather than increased plant growth as the main factor in northern forest post-Pinatubo carbon uptake. Diffuse light enhancement of growth of temperate understory trees and herbaceous plants, or of tropical vegetation, is still possible; the importance to plants of light relative to temperature would be expected to increase moving into deep shade and toward the tropics, and may have been greater after Pinatubo than after earlier eruptions because of warming in recent decades. Eruption-induced ocean fertilization [*Watson*, 1997] and tropical fire suppression [cf. *Langenfelds et al.*, 2002] provide additional possible avenues for carbon uptake.

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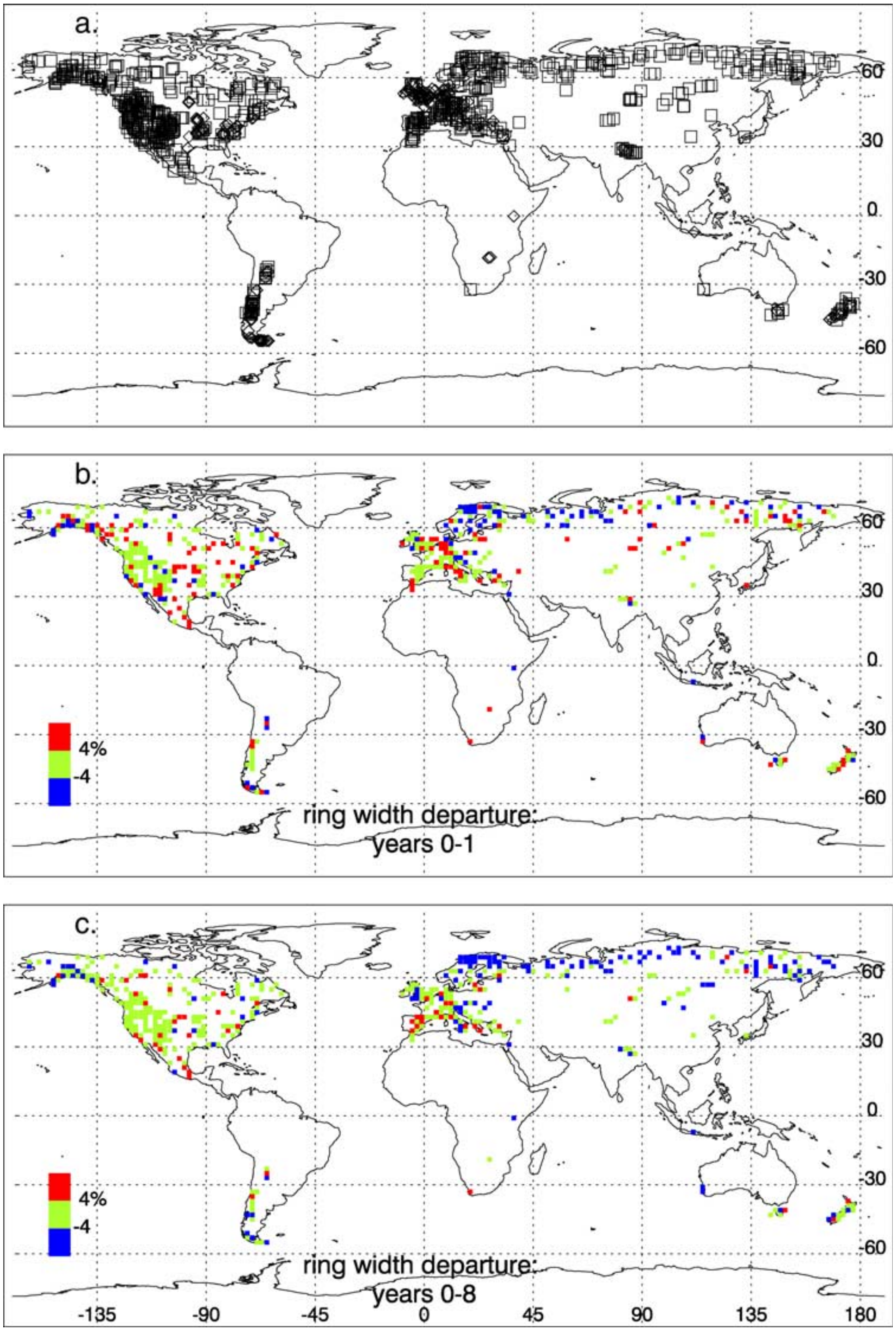

Figure 1. (a) Location of sites used for this analysis (n = 1498). Symbols indicate tree type: needleleaf (squares) or broadleaf (diamonds). (b) Ring widths for years 0 through 1 following eruption years compared with the long-term average. Site responses were averaged over 2° × 2° cells. (c) Ring widths for years 0 through 8 following eruption years compared with the long-term average. Note the preponderance of negative width departures at middle and high northern latitudes.