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Tree growth response to the 1913 eruption of Volcán de Fuego de Colima, Mexico

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Abstract

The impact of volcanic eruptions on forest ecosystems can be investigated using dendrochronological records. While long-range effects are usually mediated by decreased air temperatures, resulting in frost rings or reduced maximum latewood density, local effects include abrupt suppression of radial growth, occasionally followed by greater than normal growth rates. Annual rings in Mexican mountain pine (*Pinus hartwegii* Lindl.) on Nevado de Colima, at the western end of the Mexican Neovolcanic Belt, indicate extremely low growth in 1913 and 1914, following the January 1913 Plinian eruption of Volcán de Fuego, 7.7 km to the south. That event, which is listed among the largest explosive eruptions since A.D. 1500, produced ashflow deposits up to 40 m thick and blanketed our study area on Nevado de Colima with a tephra fallout 15–30 cm deep. Radial growth reduction in 1913–14 was ≥30% in 73% of the sampled trees. We geostatistically investigated the ecological impact of the eruption by mapping the decrease in xylem increment and found no evidence of a spatial structure in growth reduction. Little information has been available to date on forest species as biological archives of past environments in the North American tropics, yet this historical case study suggests that treeline tropical sites hold valuable records of prehistoric phenomena, including volcanic eruptions.

Keywords: Dendrochronology; Ecology; Timberline; Treeline; Neotropics; Volcanology

Introduction

Dendrochronological records provide various types of evidence for the impact of volcanic activity on forest ecosystems. Major explosive volcanic eruptions that inject dust and aerosols into the stratosphere are capable of causing large-scale surface cooling (Minnis et al., 1993). Distant, large-scale networks of temperature-sensitive tree-ring chronologies reflect those eruptions either in anatomical xylem features, such as frost damage (LaMarche Jr. and Hirschboeck, 1984), or in measured annual growth parameters, especially maximum latewood density (Briffa et al., 1998). Trees growing close enough to the volcano to be covered with tephra may either be killed or survive depending on tephra layer thickness and coarseness (Yamaguchi, 1985). Surviving trees experience abrupt suppression of radial growth (Druce, 1966; Hinckley et al., 1984), including locally absent rings (Yamaguchi, 1983). Such initial response can generate prolonged periods of reduced radial increment (Segura et al., 1995b; Smiley, 1958) or be followed by greater than normal growth rates (Abrams et al., 1999; Hinckley et al., 1998; Segura et al., 1995a).

Despite the presence of many active volcanoes in the North American tropics, little or no information is available on forest species as biological archives of past eruptions in that heavily populated region. Early work (Eggler, 1967) conducted near the apex of Volcán Paricutín used a total of nine relatively young trees belonging to three different pine species. Abnormal wood growth was associated with the aftermath of volcanic activity, but the absence of crossdating among ring patterns, the limited number of samples, and

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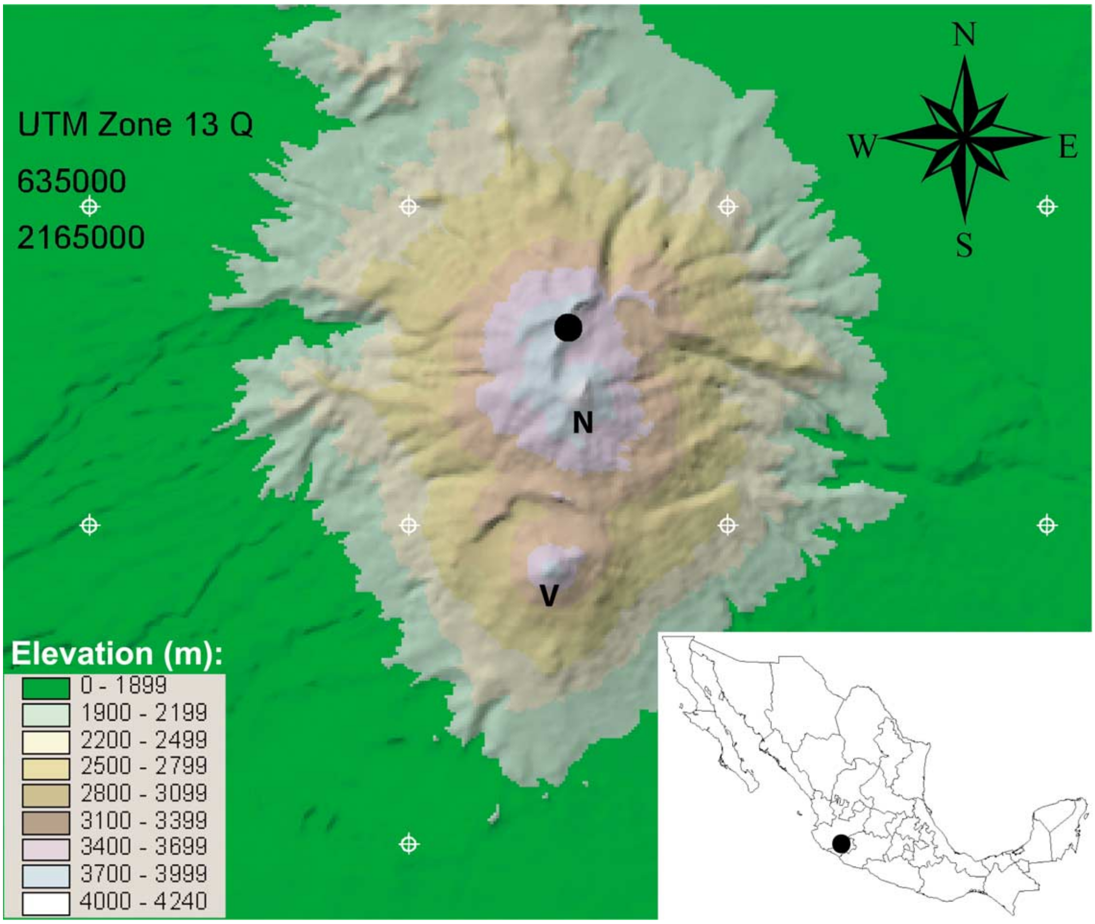

Fig. 1. False-color, shaded relief map of the Colima Volcanic Complex. Grid nodes (white crosshairs) at 10-km intervals are shown for reference, together with Universal Transverse Mercator (UTM) coordinates of the upper left grid node. Volcán de Fuego (V) is about 5.5 km south of Nevado (N), and both peaks rise 2000 m above the surrounding landscape. Our study area (●)was covered by 15 to 30 cm of tephra following the 1913 eruption of Volcán de Fuego.

the below-treeline elevation of the sites cast doubts on the reliability of dates assigned to xylem layers (Biondi and Fessenden, 1999). Based on a recently developed, 400-year tree-ring chronology from tropical North America (Biondi, 2001a), we present here the impact of tephra fallout from the January 1913 Plinian eruption of Volcán de Fuego de Colima on the *Pinus hartwegii* timberline found near the top of the nearby Nevado de Colima. Our objective was to investigate changes in annual growth of forest trees following the deposition of volcanic tephra, as well as to test if tree growth response was organized spatially within that upper elevation forest.

Study area

Volcán de Fuego de Colima is one of the most active volcanoes in the North American continent (Bretón González et al., 2002; Martin del Pozzo and Sheridan, 1993; Medina Martinez, 1983). Located at the western end of the Trans-Mexican Neovolcanic Belt, Volcán de Fuego rises almost 4000 m above sea level and lies about 5.5 km south of the older, higher (ca. 4300 m), and larger Nevado de Colima, an andesitic volcano long extinct and covered by abundant vegetation (Fig. 1). Both peaks are part of the Colima Volcanic Complex, although some confusion has been generated by authors who have used such names as synonyms (e.g., Simkin and Siebert, 1994, p. 267). Volcán de Fuego is a typical composite volcano characterized by periods of reduced activity separated by short, cataclysmic eruptions (Martin del Pozzo et al., 1995). One such event occurred in January 1913 and lasted a few days, causing widespread damage (Saucedo Girón, 1997; Saucedo Girón and Macías Vázquez, 1999). The 1913 event featured a volcanic explosivity index (Newhall and Self, 1982) of 4 and is included among the largest explosive eruptions since A.D. 1500. After the emission of a tall vertical eruptive column, pyroclastic flows descended following major canyons toward the south, southeast, and southwest sides of the cone, reaching as far as 15 km from the crater (Robin et al., 1991). Repeated photographs taken before and after the eruption show that the summit crater lost nearly 100 meters in height, while changing from a nicely rounded shape to a jagged, irregular rim (Luhr, 1981; Waitz, 1914). Ash and pumice fallout extended to the northeast, covering Nevado de Colima with tephra deposits reaching and sometimes exceeding 50 cm in depth (Saucedo Girón, 1997). Wind measurements performed from 1994 to 1997 at 3500 m elevation on the Volcán in a locality called El Volcancito, just a few hundred meters below the summit, indicate that in January predominant wind directions are from W–WNW and S–SSE (Galindo Estrada et al., 1998). By adding wind vectors together, it is therefore expected that ashfall and ash clouds will be most abundant in the resulting direction of NE–ENE. Upper-level winds, which in this region are driven mainly by the strong Hadley circulation system (Waliser et al., 1999), are also expected to point northeastward. In fact, ashfall reached as far as northern Mexico: it was reported in the states of San Luis Potosí (400 km to the northeast) and Coahuila (700 km northward).

While volcanic activity maintains the slopes of Volcán de Fuego free of forests, abundant vegetation is found on Nevado de Colima, especially on its northern side (Madrigal Sánchez, 1970). The southern aspects of Nevado have frequently been disturbed from Volcán de Fuego historic eruptions. For instance, there are reports of fires burning on the southern flanks of Nevado de Colima shortly after the start of the 1913 eruption (Saucedo Girón, 1997). Tree species found on Nevado can be grouped and classified according to elevational zones that change along gradients of temperature and precipitation (McVaugh, 1992). Near the top, up to about 4000 m elevation, the dominant species is *Pinus hartwegii* Lindl., which forms open, uneven-aged stands and endures daily temperature ranges as large as the annual range, approaching 20°C (Galindo Estrada et al., 1998). While *Pinus hartwegii* treelines on Pico de Orizaba (Lauer, 1978) and on the twin peaks Iztaccihuatl and Popocatepetl (Beaman, 1962), all mountains that exceed 5200 m elevation, have already been investigated, the Nevado de Colima treeline has not been studied in detail, not even after the 1936 creation of the Parque Nacional Volcán-Nevado de Colima. Field observations on forest structure and dynamics were recorded during extended trips in April 1998, May 2000, and May 2001. A total of 50 dominant trees were dendrochronologically sampled and used to develop the first multicentury tree-ring chronology in the Americas between 20°N and the equator (Biondi, 2001a). Sampled trees ranged from 0.5 to 1.3 m in diameter at breast height (dbh), reached ages in excess of 500 years, and were located within or near a site called Puerto La Calle, north of Nevado (Fig. 1). Although the site, which covers about 2 km^2, is relatively protected against strong winds by nearby ridges, it received a tephra fallout with a thickness of 15–30 cm from the 1913 eruption (Saucedo Girón, 1997).

Materials and methods

Crossdated tree-ring records were combined by tree to obtain a representation of growth damage caused by the 1913 eruption. Dates were assigned to individual xylem layers by visually matching ring patterns among core samples underneath a stereo-zoom binocular microscope with 7–35× magnification (Fig. 2). Ring width was measured to the nearest micrometer, and numerical verification of dating accuracy was performed using the COFECHA software (Holmes, 1983). A total of 63 samples from 26 trees included years 1913–14 and were used to investigate the impact of the 1913 eruption on the *Pinus hartwegii* forest. Ring-width patterns of 51 core samples that included years from 1910 to 1920 were digitally captured using an image analysis system and combined into a web-accessible animation (Biondi, 2001b). The other 12 samples included in computations of growth reduction were used for X-ray microdensitometry and were not digitally captured. To minimize growth effects related to tree age and size, standardized ring indices were computed as

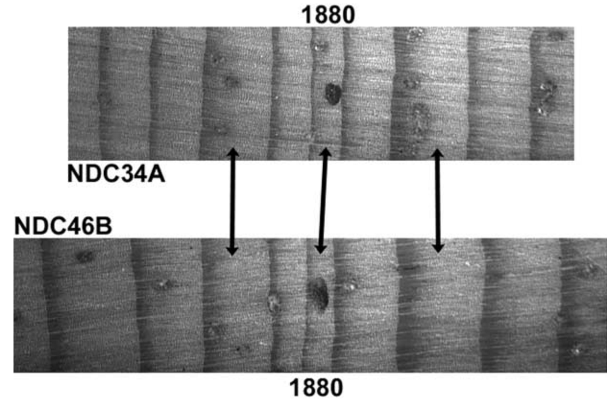

Fig. 2. Example of crossdating two *P. hartwegii* core samples from Nevado de Colima, Mexico. Identification codes are formed by three letters for the site (NDC), two digits for the tree (34 and 36), and a letter for the increment core (A and B). Time is from right to left, and a pencil dot marks year 1880. Interannual growth variability and degree of replication across samples are high enough to allow for matching ring patterns both visually (arrows) and numerically (Biondi, 2001a).

$$i_{it} = \log(w_{it} + k) - y_{it}$$

with i_{it} = ring-index of sample i on year t; w = crossdated ring width (μm); k = constant added to avoid taking the logarithm (log) of zero; y = modified negative exponential or straight line with slope ≤ 0. This computational method is based on an exponential model and is well suited for estimating the most recent portion of historical trends (Biondi, 1993, 1999). Tree summaries were produced by averaging ring indices, and mean ring-index in 1913–14 ($I_{13\text{-}14}$) was compared to the mean of ring indices in 5 previous years (1908–12) and 5 following years (1915–19). Tree growth reduction ($gr_{13\text{-}14}$, %) was then computed for each tree as

$$gr_{13\text{-}14} = \left(1 - \frac{I_{13\text{-}14}}{I_{08\text{-}12,15\text{-}19}}\right) \times 100,$$

with $I_{08\text{-}12,15\text{-}19}$ = arithmetic average of ring indices in 1908–12 and 1915–19.

The main climatic signal recorded by *P. hartwegii* annual wood increment on Nevado de Colima is June rainfall (Biondi, 2001a), which corresponds to the onset of summer monsoon precipitation. To test for potential interactions between volcanic activity and climate as factors influencing tree growth in 1913–14, monthly precipitation data extending back to 1900 were obtained from global datasets developed at the Climatic Research Unit, University of East Anglia, Norwich, UK. We used a historical (1900–1998) monthly precipitation dataset for global land areas gridded at 2.5° latitude by 3.75° longitude resolution, constructed and supplied by M. Hulme (Hulme, 1992, 1999; Hulme et al., 1998). Gridded values were computed using Thiessen polygon weights to average rain gauge data within each gridbox. Weather stations could contribute to the gridded values if they had at least 75% valid monthly measurements during a 40-year reference period. When a monthly station value was missing, an estimate was obtained by calculating the mean anomaly for that location using a maximum of 50 surrounding stations. We used the average of gridboxes 4149 and 4150, which include our study area as they cover 18°45′N–21°15′N latitude and 106°52.5′W–99°22.5′W longitude. The total summer (June–August) rainfall was computed to represent monsoon precipitation.

Coordinates of sampled trees were obtained in the field using handheld global positioning systems (GPSs). Geostatistical techniques (Isaaks and Srivastava, 1989) were used to investigate spatial patterns of growth reduction. In particular, variogram analysis was used to test for spatial dependence, and block kriging was used to produce a map of tree response to the 1913 eruption. The main advantages of kriging versus other interpolation techniques are as follows: (i) it is the best linear unbiased estimator of the regionalized variable at unsampled locations, because it is constrained to produce residuals with zero mean and minimal variance; (ii) it provides an estimate of the error of the contoured surface by producing a map of the deviations from the data values; (iii) block kriging allows both declustering (= weighing different locations so that each weight is inversely proportional to the density of sample locations) and averaging of point estimates for the block area in one single step; (iv) it may be expanded to incorporate additional variables using the technique of cokriging; and (v) its parameters often have biological meaning (Biondi et al., 1994).

It should be noted that the Nevado de Colima tree-ring chronology shows extreme reduction of tree growth not only in 1913–14 but also in 1816–17 and in 1655 (Biondi, 2001a). We decided to focus on the 20th century events because in 1655 the chronology is based on a very small sample (6 trees only), and in 1816–17 tree-ring records could not be compared with climatic records. Furthermore, in the early 1800s the tree-ring evidence is not clear enough to argue in favor of a volcanic effect, from either a local or a remote eruption. Indeed, the 1816–17 negative peak of *P. hartwegii* tree growth was preceded, in 1813, by widespread frost damage (Biondi, 2001a). Although the 1816–17 suppression follows the 1815 Tambora eruption (Harrington, 1992), Volcán de Fuego de Colima also erupted in February 1818 with a volcanic explosivity index of 4 (Newhall and Self, 1982). It would then be tempting to postulate that tree-ring dates should be changed by 2 years, so that 1813 could become 1815 and the growth suppression could coincide with the years 1818–19. In reality, tree rings differ from any other kind of paleoenvironmental record because their dating is *independent* and *internally consistent.* More than 50 samples were visually and numerically crossdated back to 1800,[1] and it is unlikely that all samples included two false rings, although that possibility cannot be excluded a priori.

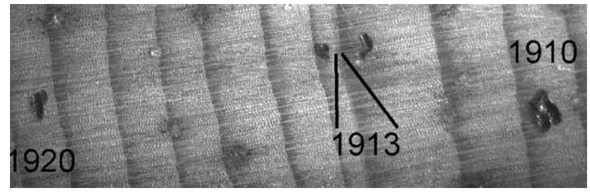

Fig. 3. A representative sample of tree growth response to the 1913 eruption. In this core, the 1913 ring is barely visible (0.092 mm wide). The amount of radial growth suppression in 1913–14 ($gr_{13–14}$) was 52%.

Results and discussion

Comparison among different radii and trees shows that the 1913 eruption resulted in a sudden reduction of stem growth (Fig. 3). Of 63 measured ring-width series, all decreased 20% or more in 1913 compared to 1912. In addition, 22 samples had no visible ring in 1913, and 8 were missing the 1914 ring as well. Growth reduction ($gr_{13–14}$) of individual pines reached a maximum of 83%, with a mean of 43% and a standard deviation of 23%. Only three individuals suffered less than 10% reduction; 73% of the trees showed a reduction ≥30%. Based on measured dbh in 1998 and on collected tree-ring specimens, at coring height (1–1.3 m above ground) all trees were older than 50 years in 1913, and 85% exceeded 100 years. In 1913, access to our study area was possible only by foot, horseback, or burro, the latter being the most common form of transportation in those areas until the late 1900s. Although stumps and other indicators of selective logging are scattered within the study area, together with evidence of cattle grazing and low-intensity fires, the overall synchronicity of wood growth among samples indicates that large-scale, mostly climatic factors have controlled tree ring formation over the past few centuries. As the 1913–14 growth reduction is widespread among sampled trees, but unique over the 1900s, it cannot be related to intermittent and repeated disturbance events.

[1] Dates reported by Biondi (2001a) were independently verified at the Laboratory of Tree-Ring Research, University of Arizona, Tucson.

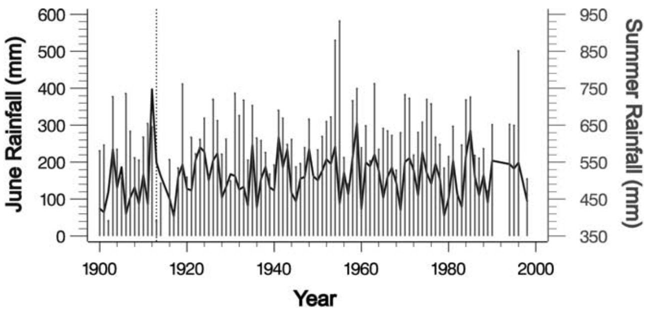

Fig. 4. Time series plot of June rainfall (solid line) and summer (June–August) total precipitation (needles) from 1900 to 1998 (see text for details). The summer of 1913 (dotted line) was very dry. The wettest June on record occurred in 1912. Values were missing in 1915 and in a few years after 1990.

Furthermore, the absence of frost damage in every tree-ring sequence rules out a potential influence of distant volcanic activity (LaMarche Jr. and Hirschboeck, 1984), such as the 1912 Novarupta eruptions in what is now Katmai National Park on the Alaska Peninsula (Fierstein and Hildreth, 1992).

June and summer (June–August) precipitation from 1900 to 1998 is shown in Fig. 4. In 1913, even though summer rainfall was very low (about 390 mm vs the long-term average of 616 mm), June rainfall was above normal (198 mm vs the long-term average of 161 mm). In addition, the June of the year before the eruption, 1912, was the wettest June on record (397 mm, or 2.5 times the long-term average). June rainfall in 1914 was slightly more than normal (163 mm vs 161 mm). Hence, it is unlikely that the extreme growth reduction experienced by *P. hartwegii* trees was caused by climatic factors alone. Nevertheless, rainfall variability could have enhanced tree response to the 1913 eruption, either by increasing the size of the 1912 radial increment (see Fig. 3 for an example) or by further reducing wood formation during the 1913–1914 growing seasons. Even more plausibly, a lowering of summer rainfall delayed the removal of tephra deposited on pine crowns and on the soil surface, which, as explained in the following paragraph, was the most likely mechanism behind the observed tree growth suppression (see Hinckley et al., 1998).

Volcanic ashfall and tephra fallout reduce photosynthesis in multiple ways (Eggler, 1967; Seymour et al., 1983; Yamaguchi, 1985). First, mechanical damage to foliage from lapilli and ash accumulation curtails the amount of photosynthetic tissue. Second, dust in the atmosphere and deposited on tree crowns lowers photosynthetically active radiation reaching pine needles. Third, dust layers plug stomata, impeding gas exchange between plant epidermis and atmosphere. Fourth, besides hampering the supply of photosynthate, crown damage and metabolic malfunction also hinder auxin production and transport to the lower stem, so that wood growth is delayed, asymmetric, and even absent at coring height. Fifth, the mantle of ash (and mud as soon as rain begins to fall) covering the ground decreases aeration of roots and changes soil chemistry, which in turn may affect absorption of water and nutrients. Because of the processes that drive growth suppression, trees on Nevado de Colima that survived the 1913 eruption could recover after monsoon rainfall had washed away the accumulated tephra. The dry summer of 1913 exacerbated the negative impact of the eruption on pine growth, which did not return to normal until 1915 (Biondi, 2001a).

Other abnormal wood increment patterns have also been associated with the aftermath of tephra deposition, such as increased growth rates following an initial reduction of tree radial increment (Abrams et al., 1999; Eggler, 1967; Hinckley et al., 1998). That rebound effect is typical of areas where tree growth is limited by lack of soil nutrients and/or by competition with surrounding individuals. Ash-driven fertilization, mulching, and partial stand mortality favor wood increment of surviving trees after an initial growth decline (Mahler and Fosberg, 1983; Segura et al., 1995a). The openness of *P. hartwegii* stands, and the deep volcanic soils they can exploit on Nevado de Colima, are consistent with the lack of a rebound effect, which was also absent in *P. ponderosa* from northern Arizona (Smiley, 1958).

Variogram analysis of tree response to the 1913 eruption revealed no spatial dependence. The isotropic semivariogram of 1913–1914 growth reduction could best be modeled as a pure nugget effect, indicating a lack of spatial autocorrelation. This derived from the presence of highly variable amounts of growth reduction at either short or long distances (Fig. 5). Because no spatial structure existed in the data, block kriging values provided no improvement over inverse distance weighting interpolation (Isaaks and Srivastava, 1989). The lack of distance-dependent growth reduction within the forest provides useful information. In particular, it shows that tree response to the 1913 eruption did not follow a spatially coherent pattern, presumably because of the stochastic nature of ashfall and tephra deposition on irregular terrain and crown forms. Although the study area is relatively small, with maximum intertree distances of about 1.3 km, the number of pairs in 100-m distance intervals ranged between 11 and 50; hence, it should not have

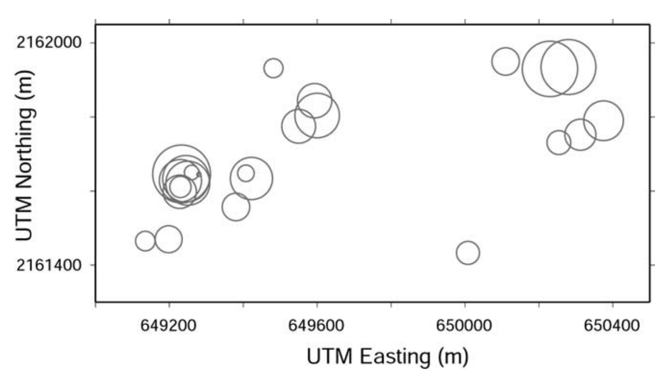

Fig. 5. Map of 1913–14 radial growth reduction in 26 *P. hartwegii* trees sampled on Nevado de Colima. Circle diameter is directly proportional to growth reduction (gr_{13-14}), and the center of the circle is drawn at the tree location. For scale, the largest circle corresponds to an 83% reduction.

biased our results. In a previous study, post-1980 tree-ring reductions in trees and saplings of four coniferous species growing northeast of Mt. St. Helens were found to be largely a function of distance from the volcano (Hinckley et al., 1984). However, subsequent research using old-growth *Abies amabilis* (Segura et al., 1994) found no spatial structure in tree growth reductions following tephra deposition from the 1980 eruption of Mount St. Helens, mostly because of the interaction with other factors (elevation, competition, etc.) affecting radial growth.

As mentioned under Materials and Methods, we focused on the 1913 eruption because we could not find clear evidence for a tree growth response to the eruption of February 1818. Considering that both volcanic events were of similar magnitude, occurred prior to the growing season, and had similar geographical axes of tephra fallout, it is perhaps necessary to further discuss the hypothesis of two false rings being included in the more than 50 samples used to develop the chronology back to 1800. False rings, or intraannual growth bands (Fritts, 1976; Kuo and McGinnes Jr., 1973), are associated with a temporary cessation of apical growth followed by growth resumption during the same year (Larson, 1962; Panshin and de Zeeuw, 1980). Storm and frost damage, drought, insect defoliation, fire, etc., can induce the formation of false rings (Glock and Agerter, 1963; Kramer and Kozlowski, 1979). Anatomical features of tree-ring series used in this study demonstrate that transition from one ring to another is sharp and well defined (see for example Figs. 3 and 4). Stem increment of *P. hartwegii* on Nevado de Colima during the 2001–2002 growing seasons has been measured at half-hour intervals (Biondi, 2002) and shows no evidence of intraseasonal cessation and resumption of growth (P. Hartsough and F. Biondi, unpublished data). Therefore, one would have to assume that the two double rings were formed because of the volcanic eruptions themselves, in 1913 and 1818. In other words, rings considered missing would not be, and what was considered a micro-ring (such as 1913 in Fig. 3) would instead be an intraannual band. While the development of additional dendrochronological records from *P. hartwegii* sites in other parts of Mexico and Central America should provide a test of this hypothesis, currently available evidence does not seem to warrant a revision of Nevado de Colima tree-ring dates.

In conclusion, we were able to demonstrate that *P. hartwegii* trees growing near treeline on Nevado de Colima were negatively affected by the impact of the January 1913 eruption from the nearby Volcán de Fuego. Exactly dated, annually resolved tree-ring records uncovered a sudden growth reduction in 1913, often extending into 1914 as well. Such 2-year response was strengthened by summer drought. Given that little information has been available to date on forest species as biological archives of past environments in the North American tropics, it appears that treeline tropical sites hold valuable records of environmental phenomena, including volcanic eruptions.

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