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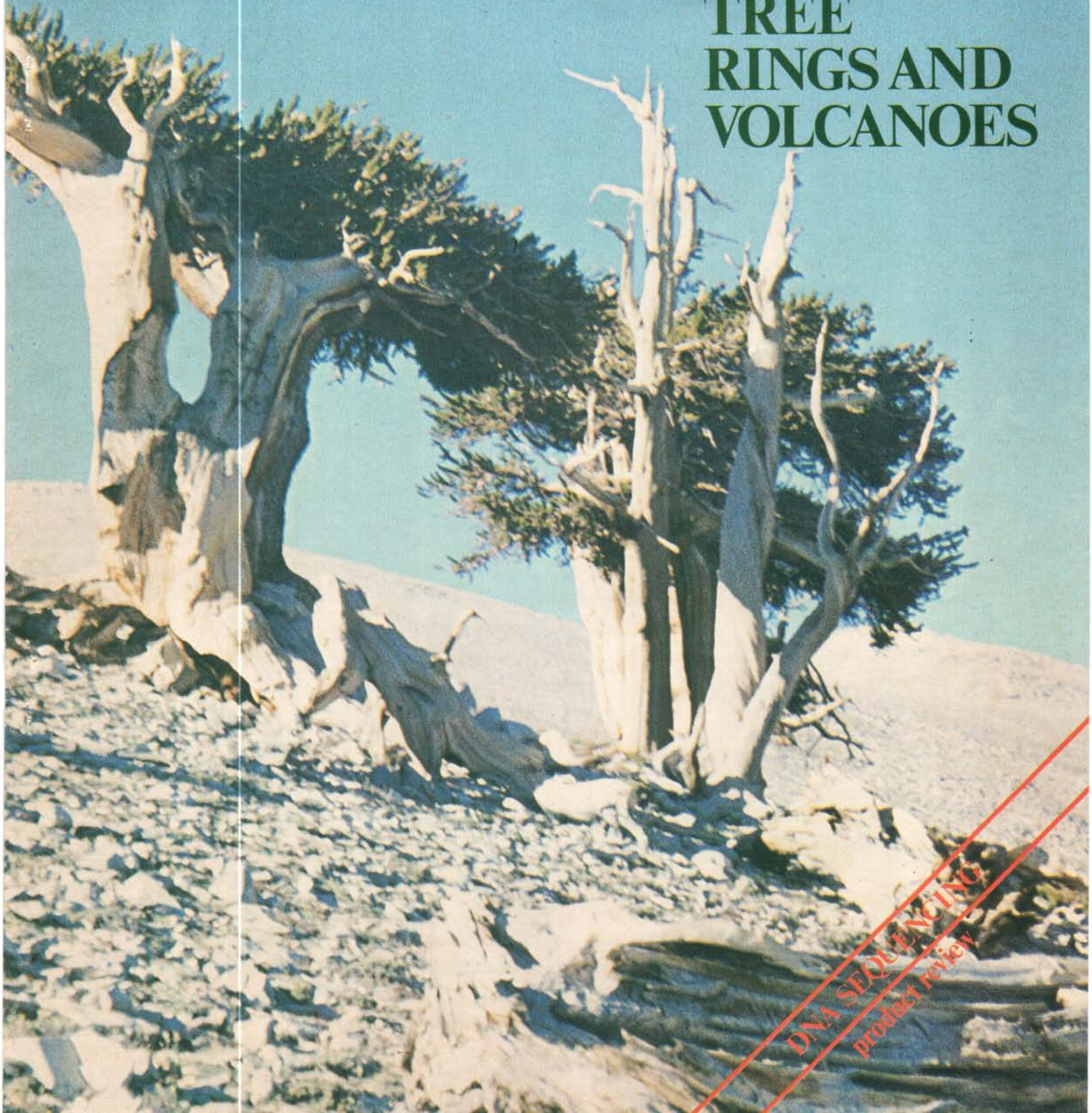
TREE RINGS AND VOLCANOES



DNA SEQUENCING
product review

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Note: the 1627 B.C. date (in red) has been corrected from the originally published date of 1626 B.C. which was incorrect

Frost rings in trees as records of major volcanic eruptions

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New data about climatically-effective volcanic eruptions during the past several thousand years may be contained in frost-damage zones in the annual rings of trees. There is good agreement in the timing of frost events and recent eruptions, and the damage can be plausibly linked to climatic effects of stratospheric aerosol veils on hemispheric and global scales. The cataclysmic proto-historic eruption of Santorini (Thera), in the Aegean, is tentatively dated to **1629-27** BC *from frost-ring evidence.*

VERY occasionally, a major explosive eruption injects large amounts of ash and sulphur aerosols into the stratosphere and upper troposphere. In addition to causing rare atmospheric optical effects—a bluish tint to the Sun and Moon, a milky cast in the daylight sky, and lurid sunsets[1,2]—such events can have pronounced effects on weather and climate that may persist for several years. Current interest[3,4] in the climatic effects of volcanic dust and aerosols has been heightened by a series of eruptions of the Mexican volcano El Chichón in 1982 and the eruption of Mt St Helens. A new dimension has been added by the recent development of proxy records of past eruptions from measurements of acidity, chemical impurities and microparticle content of ice cores from the polar regions[5-7]. We propose here that the occurrence of frost-damage zones in accurately dated tree-ring sequences from subalpine bristlecone pines in the western USA is closely linked to climatological effects of major volcanogenic atmospheric veils, and that these frost rings represent new, independent proxy records of climatically effective eruptions during the past several thousand years.

Frost rings and meteorological events

Frost damage to the wood of mature trees is a rare phenomenon caused by temperatures well below freezing at some time during the growing season, when secondary wall thickening and lignification of immature xylem cells in the annual ring is not yet complete. Freezing promotes extracellular ice formation and dehydration which result in crushing of the outermost zone of weaker cells, leaving a permanent, anatomically distinctive record in the wood[8]. Two successive nights with temperatures reaching −5 °C and an intervening day at about the freezing level are sufficient to cause frost damage[9]. Two types of frost rings can be distinguished: earlywood frost damage in the inner part of the ring, which occurs near the beginning of the growing season and latewood frost damage, which typically involves only the dark zone of small, thick-walled cells formed near the end of the growing season (Fig. 1). If the period of cambial activity is known, observing the position of the damage in a ring often permits dating of the event to within a week or two. If daily meteorological data are available, it is usually possible to identify the specific date on which the damage took place and thus, to characterize the synoptic meteorological situation as well as the antecedent climatic conditions that may have contributed to formation of a frost ring.

Frost-damage zones have been produced in the annual rings of subalpine bristlecone pines (*Pinus longaeva* D. K. Bailey and *P. aristata* Engel.) at intervals of a few decades to a few hundred years for at least the past 4,000 yr. They are observed at localities ranging from California to Colorado, a distance of some 1,300 km. In the course of tree-ring chronology development, the presence and type of frost damage in dated annual rings from living trees[10-12] and sub-fossil wood[13,14] was routinely noted. It soon became clear that frost damage had not occurred randomly, but that characteristic frost-ring dates were common to many trees at the same site or general locality, and even to localities several hundred kilometres apart. For the period of meteorological record it was possible to identify the dates of damaging related meteorological events because the growing season of bristlecone pine is known fairly well. For example[15,16], in the White Mountains, California and the Snake Range, Nevada the frost event that produced latewood damage in AD 1884 (Fig. 1) probably took place on 9–10 September, and similar latewood damage in 1965 probably occurred on 17–19 September. In each case temperatures reached new record lows for the month at nearby meteorological stations. In 1902 an earlywood event recorded in the Snake range almost certainly occurred on 3–4 July when the minimum temperature at upper treeline is estimated to have been −10 °C or below. There is clearly a relationship between such unusual meteorological events and the formation of frost rings in subalpine bristlecone pines.

Antecedent climatic conditions may also be important. Study of two years (1884 and 1965) in which latewood frost damage took place in bristlecone pines in California and Nevada shows that these were notably cool summers in the western Great Basin[17,18]. One effect of such cool conditions is to delay both the onset and the completion of cambial activity. Bristlecone pines near upper treeline normally begin radial growth about late June, and cell maturation is complete by late August, some 3–4 weeks later than in trees near the lower forest border in the same area[19]. During a cool summer, maturation may be delayed to late September, when severely cold weather would be more likely to occur even during a normal year, and this would widen the tree's 'window of vulnerability' to damaging frost.

Climatic effects of eruptions

One important consequence of a large explosive eruption is the spread of a stratospheric veil of fine silicate ash and sulphur aerosols, with resultant surface cooling[20] that may be accentuated at high latitudes by the veil's long residence time in the Arctic stratosphere[21,22]. The degree of concentration of such a cooling effect, and the time lag involved depends in part on the location of the volcano and the composition of the veil[23], on the prevailing circulation, and in part on the season of the year in which the eruption occurs[1].

The effects of an aerosol veil on the large-scale atmospheric circulation have been studied by empirical methods and by modelling. Early work by Wexler[24] on the 700 mbar pattern to be expected following a major eruption suggested that a probable circulation response would be expansion of the circumpolar vortex and intensification of an upper level trough over western North America in January. He proposed a summer pattern of westerlies displaced south of normal, and unseasonable synoptic conditions, such that the July weather chart would resemble that of a normal mid-May. Wexler's basic premise seems to be supported by Lamb's observation[21] of southward displacement of the sub-polar low-pressure zone in the North Atlantic sector in the first July following a great eruption, and continuing in some cases for 3–4 yr. Flohn[25] has speculated that one effect of the intensified polar cold vortices expected to result from high-latitude cooling associated with a stratospheric veil is increased meridionalization of the mid-latitude westerlies. Such quasi-stationary, meridional flow patterns are "... characterized by large-scale standing eddies, extending with diagonal troughs near 200 mbar far into the tropics, and leading to an increased frequency of extreme and unusual weather situations..." Modelling results seem to support the general kind of circulation changes proposed by Wexler, Flohn, and others as consequences of a volcanic aerosol veil. Hunt[26], in an early attempt, incorporated volcanic debris in an experiment using a three-dimensional, multilevel general circulation model with a dynamic dust veil of Krakatoa's magnitude. In the short term, he found an overall

Table 1 Notable frost-ring events in the western USA and dates of major volcanic events

Frost event	Volcanic event			
Date	Date(s)	Volcano(es)	Estimated DVI/E_{max}	Combined DVI/E_{max}
AD 595	–	–	–	–
628		(see Table 3)		
687	–	–	–	–
1200	–	–	–	–
<u>1227</u>	–	–	–	–
1299	–	–	–	–
<u>1453</u>		(see Table 3)		
<u>1485</u>	–	–	–	–
		(Beginning of Lamb chronology AD 1500)		
No (see Table 3)	1500	Unknown, Java	1,000	1,000
No	1586	Kelud, Java	1,000	1,000
No	1593	Ringgit, Java	1,000+	1,000+
1601	1601	Unknown	1,000+	1,000+
No	1614	Little Sunda Islands	1,000	1,000
1640	1638	Roung, Java	500	1,000
	1640	Komagatake, Japan	500	
<u>1660</u>	1660	Katla, Iceland	800	2,400
	1660	Omate, Peru	1,000	
	1660	Pichincha, Ecuador	300	
	1660	Teon, Banda Sea	300	
No	1673	Gamma Kunnora, Moluccas	1,000	1,000
<u>1680</u>	1680	Tongkoko, Celebes	1,000	1,400
	1680	Krakatoa, Sunda Strait	400	
No	1693	Hekla, Iceland	300	1,300+
	1693	Serua, Moluccas	500	
	1694	Amboina, Moluccas	250+	
	1694	Celebes	250+	
<u>1732</u>	No			
No	1752	Little Sunda Islands	1,000	2,500
	1754	Taal, Philippines	300	
	1755	Katla, Iceland	1,200	
<u>1761</u>	No			
No	1783	Eldeyjar and Jökull, Iceland	2,300	5,400
	1783	Asama Yama, Japan	600	
	1785	Vesuvius, Italy	2,500	
<u>1805</u>	No			
<u>1817</u>	1814	Mayon, Philippines	300	3,300
	1815	Tambora, Sunda Strait	3,000	
No	1821	Eyjafjallajökull	300	1,100
	1822	Vesuvius	300	
	1822	Galunggung	500	
<u>1828</u>	No			
1831	No			
<u>1837</u>	1835	Coseguina, Nicaragua	4,000	4,000+
	1837	Avachinskaya Sopka, Kamchatka	–	
1866	No			
1884	1883	Krakatoa, Sunda Strait	1,000	1,000
1902*	1902	Pelée, Martinique	100	1,000
	1902	Soufrière, St Vincent	300	
	1902	Santa Maria, Guatemala	600	
<u>1912</u>	1912	Katmai, Alaska	500†	500+
1941	No			
<u>1965</u>	1963	Agung, Bali	800†	800+

Notable frost rings are those occurring at two or more localities or in 50% or more of sampled trees in any one locality (years of occurrence in both Great Basin and Rocky Mountains are underlined). Period analysed is AD 560–1969. Major volcanic events are individual eruptions or closely spaced eruption sequences (bracketed) having a combined Dust Veil Index (DVI/E_{max}) estimated by Lamb[1] as 1,000 or more. Period analysed for volcanic events is AD 1500–1968.

* Earlywood.

† Estimated aerosol injection has been revised upward[23,27,28].

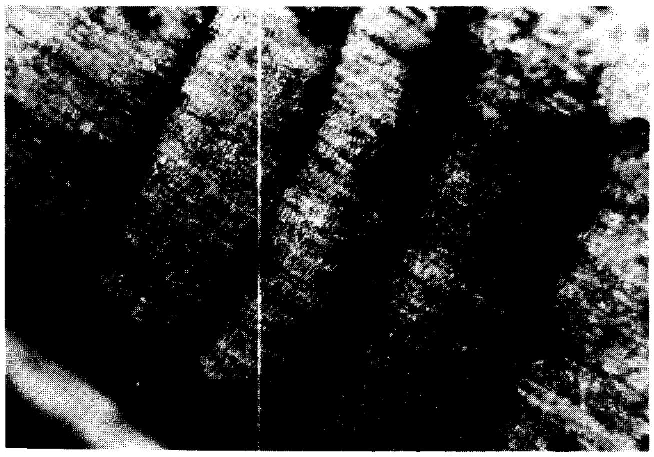

Fig. 1 Latewood frost-damage zone in 1884 annual ring, bristlecone pine, White Mountains, California. Severe frost event occurred 9–10 September 1884.

decrease in zonal wind intensity and changes in the intensity of the mean meridional cells, coupled with a decrease in mean hemispheric surface temperature. Cool summers that may contribute to occurrence of frost rings in the western USA seem to be characterized by such increased meridionality, with frequent development of a deep trough at the 700-mbar level in the middle troposphere[10]. Individual spring, summer, and autumn months can be 2–4 °C cooler than normal under this synoptic regime. Bearing in mind that the climatic impact of any individual veil will be tempered by such factors as the preeruption state of the Earth–atmosphere system and the location of the volcano, we suggest that the anomalous circulation over western North America in years following great eruptions could consist of southerly displacement of the general westerly flow and/or more frequent development of an upper level trough, accompanied by occasional outbreaks of unseasonably cold air from higher latitudes. Synoptic situations more typical of winter may be expected to occur in late spring and in early autumn. Such a scenario seems to have been followed in the frost-ring year of 1884, where an examination of daily surface-pressure maps for 9 and 10 September shows a very large high-pressure area extending from northern Saskatchewan and Manitoba down through California, Nevada and Utah, probably representing an outbreak of cold Arctic air that took place unusually early, near the end of an already cool and delayed growing season. The mid-May snowstorm and late September cold-wave in 1983 in the Rocky Mountains and High Plains of the western USA could be more recent examples.

Frost-ring records

Data from bristlecone pines at seven localities in the western USA[12] were studied in this work (Fig. 2). These include three sites in New Mexico and southern Colorado, two sites in the Colorado Front Range, one site on the Nevada-Utah border, and four sites in the White Mountains of eastern California. The tree-ring records from the Rocky Mountains begin between AD 560 and 1535. The chronology from the central Great Basin begins in AD 737. Although all of the sites have potential for chronology extension based on records from dead trees and remnants, this approach has been most successful at one site (Campito Mountain) in the White Mountains, where the continuous upper treeline chronology begins in 3435 BC [11,13,16]. The chronological control provided by these cross-dated tree-ring records representing large numbers of trees ensures accurate placement in time of both the frost-damage event and any associated volcanic eruptions.

Frost rings vary considerably from one event to another in the severity of cell damage, in their frequency of occurrence at a particular site, and in their range of distribution. Branches, seedlings, and very young trees also show a high incidence of frost damage[15], but the results discussed here are based on fairly homogeneous samples of old, well-established trees. With few exceptions, such as 1902, earlywood damage in a given year is restricted to a single locality, perhaps reflecting a local meteorological event or indicating that only at this locality had the trees begun cambial activity, rendering them susceptible to frost damage. However, latewood frost damage is frequently found to have occurred at several localities in the same year, even where these are separated by hundreds of kilometres. For example, several frost events are common to the Rocky Mountains and to the Snake Range, others to the Snake Range and the White Mountains. These variations in severity and geographical extent of latewood damage provided us with a basis for the stratification of frost-ring years that we have used to identify the more important occurrences, termed 'notable frost-ring events'. The results for all of the western USA for the period represented by chronologies from two or more localities are given in Table 1. There are 25 of these notable events in a total of some 116 individual years during which frost damage occurred in at least one sampled tree somewhere in the region. A similar stratification based on less stringent criteria was used for the White Mountains record alone, covering the much longer period 3435 BC to AD 1971.

Krakatoa effect

The postulated linkage between atmospheric veil effects caused by major volcanic eruptions and the climatological and meteorological setting for severe and widespread frost damage was originally suggested by the remarkable coincidence of frost-ring dates which fell no more than 2 yr after each of the four climatically effective Northern Hemisphere or equatorial eruptions and eruption sequences of the past 100 yr. These dates are 1884 (Krakatoa, 1883), 1902 (Pelée, Soufrière, early 1902), 1912 (Katmai (Novarupta), early 1912), and 1965 (Agung, 1963). In addition to their measured effects on the intensity of the direct solar beam[21], the aerosol veils associated with most of these eruptions seem to have caused widespread surface cooling[20,23]. To provide a much longer, if less accurate data set for further evaluation, we referred to Lamb's volcanic eruption chronology[1] and to his dust-veil estimates. A better eruption catalogue is now available[28], which is longer, more complete, and probably more objective than Lamb's in its assessment of relative magnitudes, because it incorporates the Volcanic Explosivity Index (VEI) of Newhall and Self[29], which emphasizes the explosive eruptions that are most effective in injecting gas and

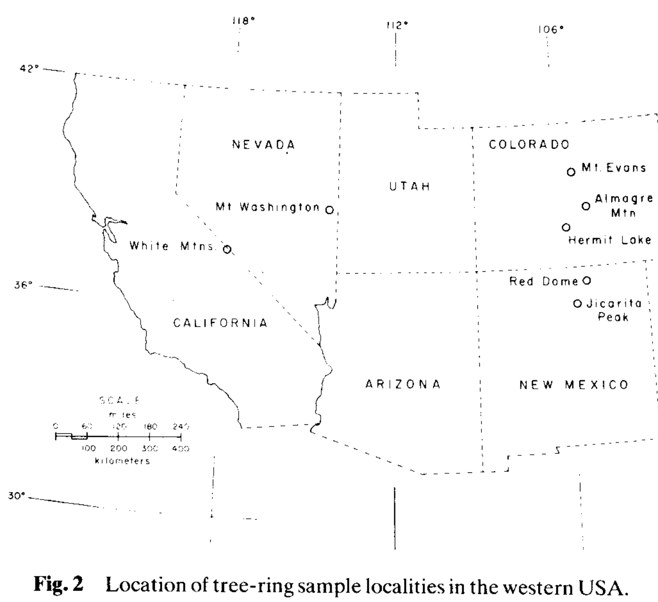

Fig. 2 Location of tree-ring sample localities in the western USA.

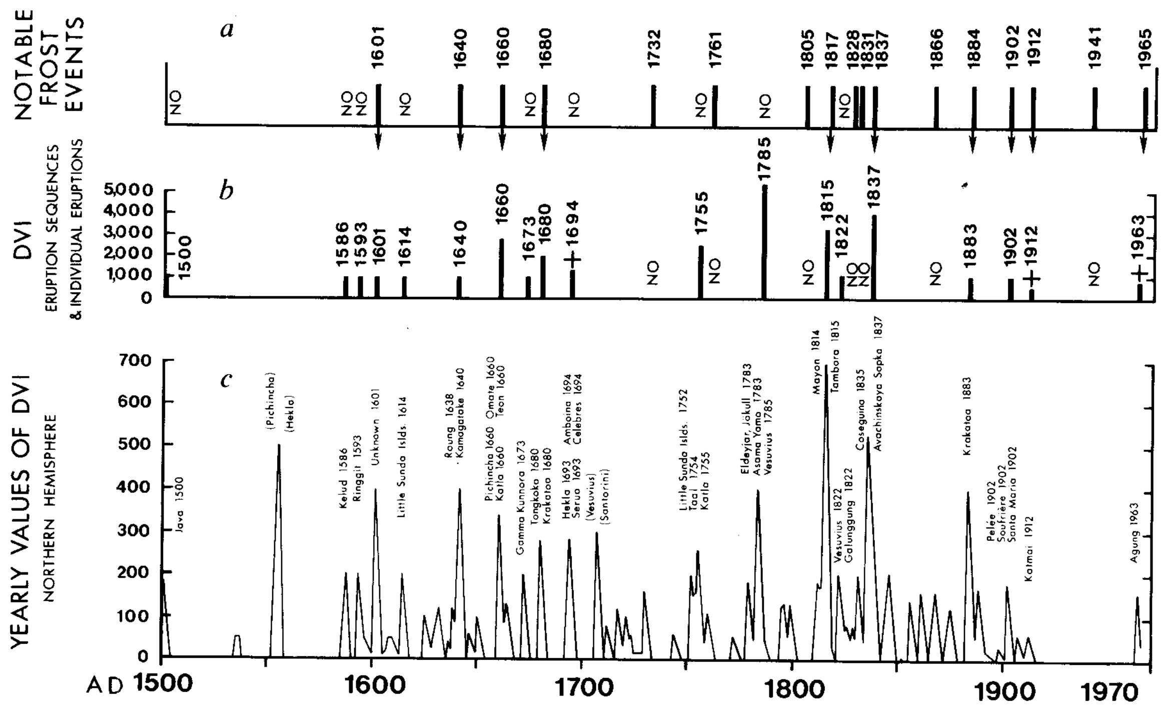

Fig. 3 Frost rings in bristlecone pines in the western USA in relation to major volcanic eruptions on global scale. *a*, Dates of notable frost-ring events. Arrow indicates associated eruption and NO indicates absence of notable frost event at time of major eruption. *b*, Dust Veil Index (DVI/E_{max}) and dates of eruptions and eruption sequences of 1,000 or greater[1]. NO indicates apparent absence of major eruption corresponding to frost-ring date shown in *a*. *c*, Integrated yearly DVI[1], with names of volcanoes and dates of major eruptions. Some other large eruptions are indicated by names of volcanoes in parentheses.

ash into the stratosphere. However, this index does not weight eruptions by geographical location to provide a scaling of probable climatic impact, so we chose to use Lamb's geographically weighted Dust Veil Index (DVI/E_{max}) in this initial comparison. We identified as 'major volcanic events' those eruptions and/or closely spaced eruption sequences, for which Lamb had assigned DVIs of 1,000 or more. For reference purposes, note that Lamb had scaled his indices to give Krakatoa, 1883, an index of 1,000. There are 19 such events in the total period, AD 1500–1968, covered by his chronology (Table 1 and Fig. 3). In 10 cases, a notable frost event as previously defined occurred in the western USA in the same year or within 1 or 2 yr following a major eruption or eruption sequence. It is also true that in nine cases a frost event occurred without an apparent antecedent volcanic event, and seven volcanic events have no known associated frost event.

To test the statistical significance of this apparent relationship, we separated the total period of concurrent frost and volcanic events into two parts. That period from 1882 to 1968 includes the four well-documented volcanic events cited above and, moreover, is a period for which meteorological data are adequate to characterize climatically-effective eruptions through their depression of hemispheric temperatures as well as to study associated regional climatic and meteorological anomalies. Our approach was to divide the 87-yr period into 29 discrete 3-yr intervals (2–3 yr is a typical decay period for stratospheric aerosol veils[20]), or triads, and then to count the number of cases in which a notable frost event occurred in the same triad as a major volcanic event. The starting date of 1882 was chosen so that 1883–84 and 1963–65 would be respectively grouped in the same 3-yr intervals, and so that the last year of Lamb's chronology would be included. From elementary probability theory[30], the expected frequency of joint occurrences of events in two random, completely independent series is equal to the product of their individual probabilities. Thus, if we assume that these probabilities are equal to the observed average frequencies, we would expect such joint occurrences to take place with a frequency of only 0.024, less than one time in 29 triads (Table 2). The observed number is nearly six times that expected by chance (note here that neither in this nor in the following analysis does the frost-ring date precede that of the associated volcanic eruption within a triad). A similar analysis was performed for AD 1500–1880 that included 127 triads and six joint occurrences. The observed number is over four times the expected chance value. The final step in our evaluation was a contingency analysis using the χ^2 statistic calculated from the observed and expected frequencies of joint occurrence[31]. Despite the small sample size, the results suggest that for each subperiod, the observed number of joint occurrences of volcanic and frost events is very unlikely to have arisen by chance. Therefore, we can reject the null hypothesis of random association, and accept the alternative hypothesis—that major eruptions are likely to be closely followed by notable frost events—at better than the 99.9% confidence level.

The analytical results show that we can tentatively associate many notable frost events in the western USA over the past several hundred years with antecedent volcanic eruptions catalogued by Lamb. As shown in Table 3 there are other

Table 2 Basic data and results of contingency analysis of frost-ring occurrence following major eruptions, AD 1500–1968

Period	1882–1968	1500–1880
No. of triads	29	127
No. of volcanic events	4	15
Observed frequency per triad	0.14	0.12
No. of frost events	5	12
Observed frequency per triad	0.17	0.09
Expected frequency of joint occurrences	0.024	0.011
Expected no. of joint occurrences	0.7	1.4
Observed no. of joint occurrences	4	6
χ^2	16.0	14.7
p	<0.0005	<0.0005

Volcanic and frost events are those listed in Table 1.
* Includes Yates correction for continuity due to small sample size[31].

Table 3 Notable frost-ring events in the White Mountains, California, and dates of possible associated eruptions

Frost events	Volcano	Volcanic events Date Lamb[1]	Stothers & Rampino[2]	Hammer et al.[5]	Simkin et al.[28]	VEI
2035 BC	St Helens			—	1900?BC	5
1627	Santorini			1390 ± 50 BC	1470 ± 20	6
42	Etna		44 BC	50 ± 20	44	3+
AD 119	Ilopango		—	gap	AD 260 ± 100	6?
601	White River		—	—	525 ± 100	6
628	Rabaul?		AD 626	AD 622–623	540 ± 100	6
687	—			—	—	—
1003	—			—	—	—
1029	—			—	—	—
1077*	—			—	—	—
1099	—			—	—	—
1171	—			—	—	—
1200	—			—	—	—
1453	Kelud?	—		1453–1454	1451	3+
1500	St Helens	—		—	1500	5
	Unknown (Java?)	1500		—	—	—
1601	Unknown	—		1600–1601	—	—
1884	Krakatoa	1883		1883	1883	6

Notable frost rings occur at two or more sites, or in 20% or more of trees (years of occurrence in 50% or more of trees on any site are underlined). Period analysed is 3435 BC to AD 1971. Northern Hemisphere and equatorial eruptions were listed if they have a Volcanic Explosivity Index (VEI[29]) of 5 or greater[28], or were represented by ice core acidity peaks[5] or well documented atmospheric veils[1,2]. Source 1 covers the period from AD 1500; 2 from 700 BC to AD 630.

* Earlywood.

records that can be used to extend this comparison qualitatively much farther into the past, using the long tree-ring record of frost events from the White Mountains, California. However, because it is based largely on data from a single tree-ring site, this frost-event chronology offers a less complete basis for comparison with volcanic events than does the broadly based regional data set given in Table 1. There are also greater uncertainties in the volcanological records in the earlier time periods. We have used the eruption catalogue of Simkin *et al.*[28], which is much longer than Lamb's. The initial Greenland ice core data[5] also provide a long, but locally biased and perhaps less well dated and possibly less complete eruption record spanning the past several thousand years, based on acidity peaks attributed to volcanogenic aerosols. Finally, we have referred to well-dated eruptions in the late BC-early AD period that are based on careful evaluation[2] of historical and archaeological records of volcanism and atmospheric veils in the Mediterranean region.

The apparent correspondence of frost events and eruptions is not as good as in the more recent past, but some coincidences are worth noting. Severe frost damage occurred in AD 1601, and may be linked to the acidity peak described[5] as one of the strongest of the earlier signals in the Greenland ice record. The ice-core data do not reflect a peak at the time of the 1500 frost ring, but an eruption occurred that produced atmospheric optical effects in Europe and is reported by Lamb to have taken place in Java[1]. A large eruption of Mt St Helens also occurred about this time[28]. The severe frost event of 1453 could be linked to a low acidity peak about 1453 [5], possibly related to a large eruption of Kelud in 1451 [28]. The high acidity peak[5] at about 1258 is not represented in the frost record and, there seem to have been no great eruptions that might account for the numerous frost events from 687 to 1200. However, another strong signal in the Greenland ice is dated at about 623 and may reflect the eruption whose atmospheric veil was recorded in Europe in 626 [2] and which might be related to the 628 frost event. The volcano is unknown, but could be Rabaul, in New Britain[28]. There is no documentary record of an atmospheric veil at the time of the frost event in 601, and no corresponding acidity peak. However, a major eruption took place in the White River field, Alaska, at about that time[28].

Referring to the earliest part of our record, only four notable frost-damage events are recognized in the White Mountains in the 4036-yr period before AD 601. One of these events, in AD 119, falls in a gap between the Crête and Camp Century ice records, but could be related to a major eruption of Ilopango, El Salvador, in the Second or Third Centuries[28]. Another, dated to 42 BC, might correspond to one of three peaks in the Camp Century core, approximately dated to 260, 210, and 50 BC, respectively[5]. It follows closely the great eruption of Etna, well dated to 44 BC from direct observations[2]. A third frost event, in 2035 BC, is the most severe in the entire record, as it occurs in all the trees in the sample and caused severe anatomical damage[16]. There does not appear to be a corresponding acidity peak in Greenland, but the frost-ring date coincides approximately with a large radiocarbon-dated eruption of Mt St Helens[28].

Santorini connection

The remaining frost event in the BC–early AD time period, dated to 1627 BC, is especially notable, not only for its severity, but because it offers the intriguing possibility of dating precisely the cataclysmic eruption on Santorini (Thera) in the Aegean Sea in the second millennium BC. The frost-ring date had seemed to place the eruption too early in time to match the conventionally accepted date of 1500–1450 BC based on the archaeological evidence available in the early 1970s[32], but this discrepancy in dates may now be resolved.

Because the volume of ejecta has been estimated at several times that of Krakatoa, the Santorini eruption seems likely to have produced a major, climatically effective aerosol veil, which is also likely, on present evidence, to have been linked with a notable frost-ring event. Betancourt and Weinstein[33] have carefully reviewed both the archaeological and radiocarbon evidence for the dating of the beginning of the late Bronze Age in the Aegean. In particular, they focus on radiocarbon dates from material obtained during the excavation of a late Minoan settlement near the village of Akrotiri. It lies on the south side of the largest remnant of the original volcanic island, most of which was either ejected or collapsed into the central caldera following the eruption. The excavations were begun by Marinatos in 1967, and the evidence indicates that the buildings were first toppled by earthquake, and soon thereafter buried under many metres of volcanic ash from the eruption. Betancourt and Weinstein evaluate two sets of dates associated with the destruction of ancient Akrotiri, after first calibrating or correcting the raw dates for atmospheric radiocarbon variability, using the 1973 MASCA tables. [For a fuller documentation and a discussion of these samples see Michael[34], who supported the seventeenth century BC eruption date.] The first set of dates is from 'long-lived' material, mostly wood from olive, pine, and cedar that would be expected to predate the eruption considerably; the dates are older than 1700 BC. More relevant here are the dates on six short-lived samples, including charred seeds of legumes from a storage jar, and charred fragments of shrubs. The average date given for these samples is 1688 BC ± 57 yr. More recently, four additional dates have been reported on grain from storage jars from the destruction level[35]. They have a large range, but yield an average date of 1675 BC. Given the uncertainties in the individual radiocarbon determinations coupled with the uncertainty in the calibrations themselves, our proposed eruption date of 1627 BC or perhaps 1 or 2 yr earlier, derived from frost-ring evidence, falls well within the probable range for the true date of the eruption based on radiocarbon.

Conclusions

The potential importance of frost rings as proxy indicators of past eruptions warrants careful weighing of the physical evidence and of the strength of the postulated linkage between aerosol veils, the response of the circulation, and their relation to the climate and weather of the study region. The tree-ring data base is adequate for the past several hundred years, but frost events may be greatly under-represented before about AD 600, because of the decreasing size and geographical extent of the sample. Our method of stratifying 'notable' frost events seems valid, but the summing of magnitudes of closely spaced eruptions (Table 1) may not be justifiable physically in all cases[20]. The comparison of dates of eruptions and frost rings shows that there is not a one-to-one correspondence. Frost events that are unrelated to aerosol veils have occurred, for example, in 1941. Conversely, there were notable eruptions with no associated frost ring in the western USA. That is, some eruptions may result in a frost record at a certain locality while others may not, and frost damage can be due to atmospheric variability from other causes. We also emphasize that comparisons of the kind that we have attempted become increasingly unreliable in the earlier part of the record. This is due partly to uneven reportage of eruptions and partly to the uncertainties in dating of geological, archaeological, glaciological and other sources of evidence.

Further testing of our hypothesis might include the analysis of an improved frost-event chronology for the western USA using an expanded tree-ring data base. Because frost rings are found in trees in the sub-Arctic and in the high mountains of lower latitudes in many parts of the world[36–40], other frost-event records from suitable regions could be compared independently with volcanic chronologies. Our preliminary findings also suggest the need for empirical studies of regional climatic variability and analysis of modelling results to see how important volcanogenic aerosol veils may be as a causal factor in frost-ring formation.

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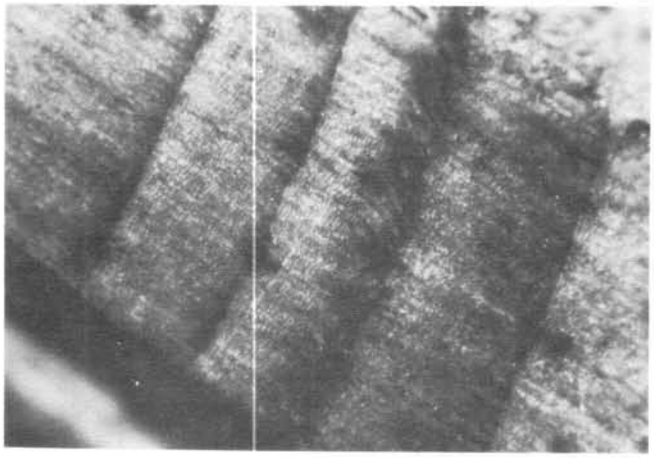

Fig. 1 Latewood frost-damage zone in 1884 annual ring, bristlecone pine, White Mountains, California. Severe frost event occurred 9–10 September 1884.

decrease in zonal wind intensity and changes in the intensity of the mean meridional cells, coupled with a decrease in mean hemispheric surface temperature. Cool summers that may contribute to occurrence of frost rings in the western USA seem to be characterized by such increased meridionality, with frequent development of a deep trough at the 700-mbar level in the middle troposphere[10]. Individual spring, summer, and autumn months can be 2–4 °C cooler than normal under this synoptic regime. Bearing in mind that the climatic impact of any individual veil will be tempered by such factors as the preeruption state of the Earth–atmosphere system and the location of the volcano, we suggest that the anomalous circulation over western North America in years following great eruptions could consist of southerly displacement of the general westerly flow and/or more frequent development of an upper level trough, accompanied by occasional outbreaks of unseasonably cold air from higher latitudes. Synoptic situations more typical of winter may be expected to occur in late spring and in early autumn. Such a scenario seems to have been followed in the frost-ring year of 1884, where an examination of daily surface-pressure maps for 9 and 10 September shows a very large high-pressure area extending from northern Saskatchewan and Manitoba down through California, Nevada and Utah, probably representing an outbreak of cold Arctic air that took place unusually early, near the end of an already cool and delayed growing season. The mid-May snowstorm and late September cold-wave in 1983 in the Rocky Mountains and High Plains of the western USA could be more recent examples.

Frost-ring records

Data from bristlecone pines at seven localities in the western USA[12] were studied in this work (Fig. 2). These include three sites in New Mexico and southern Colorado, two sites in the Colorado Front Range, one site on the Nevada-Utah border, and four sites in the White Mountains of eastern California. The tree-ring records from the Rocky Mountains begin between AD 560 and 1535. The chronology from the central Great Basin begins in AD 737. Although all of the sites have potential for chronology extension based on records from dead trees and remnants, this approach has been most successful at one site (Campito Mountain) in the White Mountains, where the continuous upper treeline chronology begins in 3435 BC [11,13,16]. The chronological control provided by these cross-dated tree-ring records representing large numbers of trees ensures accurate placement in time of both the frost-damage event and any associated volcanic eruptions.

Frost rings vary considerably from one event to another in the severity of cell damage, in their frequency of occurrence at a particular site, and in their range of distribution. Branches, seedlings, and very young trees also show a high incidence of frost damage[15], but the results discussed here are based on fairly homogeneous samples of old, well-established trees. With few exceptions, such as 1902, earlywood damage in a given year is restricted to a single locality, perhaps reflecting a local meteorological event or indicating that only at this locality had the trees begun cambial activity, rendering them susceptible to frost damage. However, latewood frost damage is frequently found to have occurred at several localities in the same year, even where these are separated by hundreds of kilometres. For example, several frost events are common to the Rocky Mountains and to the Snake Range, others to the Snake Range and the White Mountains. These variations in severity and geographical extent of latewood damage provided us with a basis for the stratification of frost-ring years that we have used to identify the more important occurrences, termed 'notable frost-ring events'. The results for all of the western USA for the period represented by chronologies from two or more localities are given in Table 1. There are 25 of these notable events in a total of some 116 individual years during which frost damage occurred in at least one sampled tree somewhere in the region. A similar stratification based on less stringent criteria was used for the White Mountains record alone, covering the much longer period 3435 BC to AD 1971.

Krakatoa effect

The postulated linkage between atmospheric veil effects caused by major volcanic eruptions and the climatological and meteorological setting for severe and widespread frost damage was originally suggested by the remarkable coincidence of frost-ring dates which fell no more than 2 yr after each of the four climatically effective Northern Hemisphere or equatorial eruptions and eruption sequences of the past 100 yr. These dates are 1884 (Krakatoa, 1883), 1902 (Pelée, Soufrière, early 1902), 1912 (Katmai (Novarupta), early 1912), and 1965 (Agung, 1963). In addition to their measured effects on the intensity of the direct solar beam[21], the aerosol veils associated with most of these eruptions seem to have caused widespread surface cooling[20,23]. To provide a much longer, if less accurate data set for further evaluation, we referred to Lamb's volcanic eruption chronology[1] and to his dust-veil estimates. A better eruption catalogue is now available[28], which is longer, more complete, and probably more objective than Lamb's in its assessment of relative magnitudes, because it incorporates the Volcanic Explosivity Index (VEI) of Newhall and Self[29], which emphasizes the explosive eruptions that are most effective in injecting gas and

Fig. 2 Location of tree-ring sample localities in the western USA.