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Evidence for a volcanic cooling signal in a 335-year coral record from New Caledonia

Thomas J. Crowley,¹ Terrence M. Quinn,² Frederick W. Taylor,³ Christian Henin,⁴ and Pascale Joannot,⁵

Abstract. Although volcanic cooling events have been detected in tree ring records, their occurrence in marine records has received much less attention. Herein we report results from a 335-year oxygen isotope record (1657-1992) from a New Caledonia coral indicating that as many as 16 interannual-scale cooling events occur within 1 year of a volcanic eruption as determined by ice core records. There are also pentadal/decadal-scale cooling events beginning in 1675, 1813, and 1903 that immediately postdate volcanic eruptions. However, the interannual correspondences are complicated by the fact that some of the cooling events also coincide with El Niños, which cause cooling in this part of the western South Pacific. If our conclusions are substantiated by further work, occurrence of distinct volcanic cooling signals may enable refinement of coral chronologies by use of the "event stratigraphic" approach, with the most promising correlation horizons being associated with the following eruptions: 1808 (Unknown), 1813-1821 (several eruptions), 1835 (Coseguina), 1883 (Krakatau), and possibly 1963 (Agung).

Introduction

The role of volcanism as an agent of climate change has been a topic of interest for some time because stratospheric injections of sulphate aerosols should increase backscatter of solar radiation and cause cooling. After many years of debate, studies of large volcanic eruptions indicate a significant effect on interannual climate change [e.g., Bradley, 1988; Angell, 1990]. Volcanic signals have also been detected in tree ring records [e.g., LaMarche and Hirschboeck, 1984; Jones *et al.*, 1995]. The role of volcanism with respect to decadal-scale variability is more uncertain. Although earlier studies suggested such an influence [Robock, 1979; Hammer *et al.*, 1980; Porter, 1986], more recent work has called into question the strength of this correlation [Wigley, 1991; Crowley *et al.*, 1993].

Coral records have recently emerged as an important source of information on interannual-decadal climate change [e.g., Cole *et al.*, 1993; Quinn *et al.*, 1993; Dunbar *et al.*, 1994; T. M. Quinn *et al.*, A multicentury coral stable isotope record from New Caledonia, submitted to *Paleoceanography*, 1997 (hereinafter referred to as T. M. Quinn *et al.*, submitted manuscript, 1997)]. The primary features of interest in these studies have been variations in the tropical Pacific circulation on El Niño and decadal timescales. Recently, however, Genin *et al.* [1995] reported on an anomalous algal bloom in a coral record from the Gulf of Eilat following the eruption of Mount Pinatubo in June 1991. The increase in water column overturn is consistent with observations of a strong winter cooling in this region after volcanic eruptions [cf. Graf *et al.*, 1993; Robock and Mao, 1995].

Although Genin *et al.* [1995] found evidence for significant overturn in their coral record, they speculated that isotopic variations may be more difficult to detect in coral records. However, Gagan and Chives [1995] found evidence for Pinatubo cooling in a northwest Australia record. In this paper we provide further evidence for imprint of large volcanic eruptions on a 335-year coral record from New Caledonia. There is also some evidence linking smaller isotope excursions with smaller volcanic eruptions.

Methods

We cored a living coral near Amedee Island, New Caledonia (22°S, 166°E), in September 1992. This region was chosen because of the very large data gap in the South Pacific; for example, many sea surface temperature (SST) records are relatively incomplete before about 1960. The selection of New Caledonia was based on availability of an excellent calibration data set (daily SST and salinity measurements) developed by the French oceanographic group at Institut Français de Recherche Scientifique pour le Développement en Coopération (ORSTOM) (cf. T. M. Quinn *et al.*, submitted manuscript, 1997). Details of the site and a full description of the long record are given by Quinn *et al.* [1996a; submitted manuscript, 1997]. Correlation of the monthly oxygen isotope record with a nearby high-quality sea surface temperature (SST) record from Amedee Island is 0.86, indicating that the coral is primarily recording information on SSTs on the seasonal time scale. Although the mean annual $\delta^{18}\text{O}/\text{SST}$ calibration is different, the coral still can be used to estimate SSTs on time scales longer than seasonal (c.f. Quinn *et al.*, submitted manuscript, 1997).

The chronology in the coral was determined by band counting (T. M. Quinn *et al.*, submitted manuscript, 1997) density bands in this coral are quite distinct. The dark bands are laid down in austral winter when growth rates slow; this marks the beginning of the "coral year" in our record. Higher resolution sampling (12/yr) over a calibration interval indicates that 4/yr sampling yields estimates of mean annual conditions similar to 6/yr and 12/yr [Quinn *et al.*, 1996a].

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The data set will be available in the National Geophysical Data Center following publication of the full description of the coral record (T. M. Quinn et al., submitted manuscript, 1997).

In order to compare our coral results with the volcanic record for the southern hemisphere, we compiled a composite estimate of middle- to low-latitude eruptions based on records of conductivity and sulphur in ice core records from Antarctica (Table 1). The details involved in the construction of this index are discussed in the appendix. The volcanic peaks have been compared to catalogs of volcanism, primarily from *Simkin and Siebert* [1994] but augmented by *Lamb* [1970]. Although many of the volcanic peaks are based on occurrence in more than one ice core, a number of peaks in the eighteenth century are from only one core [*Moore et al.*, 1991]. Thus the identification of peaks and attributions in this section require more verification. As the most compelling evidence for a volcano/coral cooling occurs in the post-1800 record (see below), we do not consider the uncertainty with respect to volcano identifications in the eighteenth century as being critical to the main conclusion of our paper.

Results

The principal results of the coral-volcano comparison are shown in Figure 1. In addition to a long-term trend and multidecadal variability (which are discussed by T. M. Quinn et al. (submitted manuscript, 1997)), there are striking excursions in the mean annual isotope record coincident within the accuracy of the coral record with the eruptions of Krakatau (1883), Coseguina (1835), a large unknown eruption in 1808 [cf. *Dai et al.*, 1991], and possibly a paired set of eruptions in 1752 (Little Sunda) and 1754 (Taal). In order to investigate this phenomenon more closely, we identified cooling events of less than 3 years duration and compared them to the subset of the volcanism composite that is not associated with cooling events of greater than 3 years duration (the rationale being that it is less easy to distinguish between cause and effect for the longer cooling events). The magnitude of the cooling event was determined as the decline from a preexisting value representing either a peak or a "ledge" in the $\delta^{18}\text{O}$ record.

Comparison (Figure 2) of the interannual oxygen isotope excursions with the composite volcanic record indicates that 16 of the 20 volcanic eruptions coincide with or are followed within 1 year by an interannual isotope excursion. The correlations are most striking after 1800, where there is a sequence of nine "direct hits" in a row between 1803 and 1886. The noncorrespondence of the 1917 and 1920 ice core peaks with a cooling in the coral record may reflect the unidentified sources of the 1917 and 1920 peaks (see footnotes d and e in Table 1). Although cooling occurs after the powerful 1963 Agung eruption, maximum cooling postdates the eruption by 2 years and blends with the strong El Niño event documented by *Quinn* [1992] and others. This complication will be further discussed below. However, the results indicate that if the 1917 and 1920 volcanic events are ignored, then all 10 post-1800 peaks identified in ice core records from the southern hemisphere are associated with cooling events in the coral record. An eleventh cooling event (1857) could be associated with the 1855 eruption of Cotopaxi and has been listed in the

volcanic compilation of *Robock and Free* [1995] but not in our record (again, see discussion in appendix).

With respect to the pre-1800 record, where the origin and chronology of the volcanic events are less well known, six of the nine volcanic events (1693, 1720, 1744, 1752, 1754, and 1788) coincide within a year of an eruption. There are two other possible volcanic events (1668 (Pacaye?), 1799 (Fuego)) occurring in the *Moore et al.* [1991] record that are below the threshold level used in this study (see appendix) but which also coincide with cooling events in the New Caledonia record.

Even if further analysis modifies the suggested low-latitude origin of some of the volcanic peaks (Table 1), there is still a large amount of evidence indicating that many of the coral cooling events occur at the same time as low-latitude eruptions. If there were no other complicating factors associated with these cooling events (see below), the likelihood of 16 volcanic "hits" occurring in such a record is exceedingly small ($p < 0.00001$). This probability was determined in the following manner. First, we determined the number of interannual cooling events (47) and multiplied this number by 2 to account for the fact that we are determining whether the volcano peak is coincident or predates by a year a cooling event in the coral record. Then we divided the entire length of the record (335 years) by this 2-year criteria to determine the number of independent opportunities for a volcano to record a hit. Cooling events of greater than 3 years duration (a total of 70 years) were first removed from the original time series before determining that there are 132 two-year "samples" that are not associated with multiyear cooling events. Of this number, approximately one third (47) actually have cooling events (Figure 2). Then we determined the chance that 16 of the 20 volcanic eruptions would coincide with the smaller category of 47 cooling events. A way to envision this exercise is to imagine a table with 132 pockets in it, 47 of which are blue. Then 20 red balls (volcanic eruptions) are randomly cast on the table. By chance alone one would expect about one third of the red balls to fall within the blue pockets. However the actual number is about 3 times that amount; thus the very low probability of the coincidence occurring by chance.

Although the previous exercise is instructive, a significant complication results from the fact that El Niño/Southern Oscillation (ENSO) events are associated with cooling in the western South Pacific. For example, the strong 1940-1941 El Niño is clearly recorded as a cooling event in the New Caledonia record (Figures 1 and 2). Likewise, the weaker 1976-1977 El Niño is associated with a significant isotopic excursion (the stronger 1982-1983 ENSO is more muted and blends into a 6-year "cool" interval). In fact, a record of ENSO events for the past few centuries [*Quinn*, 1992] indicates that 12 of the 16 volcano/coral hits also occur within a year of cooling events linked with volcanism. The only four volcano/coral excursions that do not seem to be associated with ENSOs are the 1746 cooling following the November 1744 eruption of Cotopaxi, an unknown 1788 eruption that has also been found in a reanalysis of northern hemisphere ice core [*Crowley et al.*, 1993; cf. appendix note c], the strong 1808 eruption of unknown origin [*Dai et al.*, 1991], and a weak cooling following the 1822 eruption of Galunggung.

Table 1. Conductivity/Sulphate Peaks in Antarctic Ice Core Records Used to Construct a Composite Index of Volcanism That Could Have a Hemispheric-Scale Impact on Southern Hemisphere Climate

Year	Volcano	Latitude	Longitude	Volcano Reference	Scaled Amplitude	Ice Core Reference
1600	Huaynaputina, Peru	16.6°S	70.9°W	S-5	1.21	CH, Z, DM, M
1605	Momotombo, Nicaragua	12.4°N	86.5°W	S-4	0.45	Z, M
1622	Colima?, Mexico	19.5°N	103.6°W	S-4	1.02	CH, DM, M
1641	Kelut, Indonesia	7.9°S	112.3°E	S-4	0.98	CH, Z, DM, M
	Awu, Indonesia	3.7°N	125.5°E	S-5		
1675 ^a	Long Island, N.G.?	5.4°S	147.1°E	S-6	0.91	CH, M
1680	Tongkoko, Indonesia	1.5°N	125.2°E	S-5	0.89	CH, M
	Krakatau, Indonesia	6.0°S	105.4°E	LMB		
1693	Serua, Indonesia	6.3°S	130.0°E	S-4	0.67	Z, M
	Gunung Api (1694)	4.5°S	130.0°E	LMB		
1711 ^b	??	—	—	—	0.91	M
1720	Bravo Cerro, Columbia	5.1°N	75.3°W	S-4	0.45	M
1744	Cotopaxi, Ecuador	0.7°S	78.4°W	S-4	0.45	M
1752	Little Sunda, Indonesia	8.0°S	118.0°E	LMB	0.50	M
1754	Taal, Philippines	14.0°N	121.0°E	S-4	0.45	M
1768	Cotopaxi, Ecuador	0.7°S	78.4°W	S-4	0.45	M
1772	Gunung Papandajan, Indonesia	7.5°S	108.0°E	LMB	0.56	M
1788 ^c	??	—	—	—	0.64	CH, M
1803	Cotopaxi, Ecuador	0.7°S	78.4°W	LMB	0.50	M
1808	??	—	—	—	1.93	LW, Z, DM, D, LG, M
1812	Soufriere, Caribbean	13.3°N	61.2°W	S-4	0.50	M, D?
	Awu, Indonesia	3.7°N	125.5°E	S-4		
1815	Tambora, Indonesia	8.0°S	118.0°E	S-7	2.83	CH, Z, LW, DM, D, LG, M
1817	Roung, Indonesia	8.1°S	114.0°E	S-4	0.98	D, M
1822	Galunggung, Indonesia	7.2°S	108.0°E	S-5	1.28	D, M
1826	Kelut, Indonesia	7.9°S	112.3°E	S-4	0.67	M
1831	Babuyan, Philippines	19.5°N	121.9°E	S-4	0.68	Z, DM, LG
1835	Coseguina, Nicaragua	13.0°N	87.6°W	S-5	0.90	Z, LW, DM, LG, M
1877	Cotopaxi, Ecuador	0.7°S	78.4°W	S-4	0.49	DM, M
1883	Krakatau, Indonesia	6.1°S	105.4°E	S-6	1.00	CH, Z, LW, DM, LG, M
1886	Tarawera, N.Z.	38.2°S	176.5°E	S-5	0.66	DM, M
1902	Santa Maria, Guatemala	14.8°N	91.6°W	S-6	0.57	Z, LG, M?
	Soufriere, Caribbean	13.3°N	61.2°W	S-4		
	Pelee, Caribbean	14.8°N	61.2°W	S-4		
1917 ^d	Agrigan, Marianas?	18.8°N	145.7°E	S-4	0.54	Z?, M
1920 ^e	??	40.6°S	72.1°W	S-4	0.57	M
1963	Agung, Indonesia	8.3°S	150.5°E	S-4	1.23	DM, M
1991 ^f	Pinatubo, Philippines	15.1°N	120.4°E	S-6	0.78	Mc

The eruption chronology was derived from published analyses of Antarctic records (see below), with peaks defined as increases above local background [cf. Crowley *et al.*, 1993] and amplitudes scaled to the 1883 Krakatau peak in all records except for the Dai *et al.* [1991] core, which was scaled to Tambora and then rescaled to Krakatau using the Krakatau/Tambora relationship in other cores. "Year" refers to estimated year of the eruption. "Volcano reference" refers to the suggested volcano(s) responsible for the ice core peak, as determined by a search in the volcano chronologies of Lamb [1970] and Simkin and Siebert [1994]. The number after "S" refers to the relative magnitude of eruption as estimated by Simkin and Siebert [1994]. These latter references were used to eliminate some ice core peaks that appear to correlate with high-latitude eruptions. "Ice core reference" refers to the ice core reference used to identify or corroborate peak, with northern hemisphere sites used as an aid in assigning a low-latitude origin (CH, Crowley *et al.*'s [1993] analysis of Hammer *et al.*'s [1980] results; D, Dai *et al.* [1991]; DM, Delmas *et al.* [1992]; LG, Legrand and Delmas [1987]; LW, Langway *et al.* [1995]; M, Moore *et al.* [1991]; Mc, McCormick *et al.* [1995]; and Z, Zielinski [1995]). Note that low-latitude northern hemisphere eruptions can influence the southern hemisphere. See appendix and footnotes for further discussion.

^a The attribution of this peak to the Long Island eruption is very uncertain. The ¹⁴C age determination of 1660 ± 20 years [Simkin and Siebert, 1994] has also resulted in this eruption being identified with other events in this time interval [Jones *et al.*, 1995; Zielinski, 1995].

^b This peak could be a local eruption; it is included in the list because there is no reason to exclude it.

^c Even though no volcano compilations list an eruption at this time, occurrence of a prominent peak in the Crete core at the same time [Crowley *et al.*, 1993] suggests it could be a low-latitude eruption. This peak also occurs in an ice core from Mount Logan, Yukon Territories, Canada [Mayewski *et al.*, 1993]. However, further verification of this peak is needed in other cores.

^d The attribution of this peak to Agrigan is very uncertain; it could be an unknown eruption in the high latitudes of the southern hemisphere. The only reason it is listed as possibly having a low-latitude influence is because a well-dated 1917 peak occurs in the Greenland Ice Sheet Project 2 record 1 year earlier than the local Icelandic (Katla) eruption it has been attributed to [Zielinski, 1995].

^e The origin of this peak is unknown; it could be noise or some undocumented eruption in the high latitudes of the southern hemisphere.

^f As discussed in the appendix, the Pinatubo estimate is based on satellite measurements. Although the global Pinatubo aerosol loading was approximately 20% greater than Agung [McCormick *et al.*, 1995], the Pinatubo aerosol veil was approximately evenly distributed between the northern and southern hemispheres, while the Agung forcing was primarily in the southern hemisphere. Thus the estimated magnitude of the Pinatubo versus Agung eruptions in this table reflects only the estimated relative magnitude of their southern hemisphere aerosol loading.

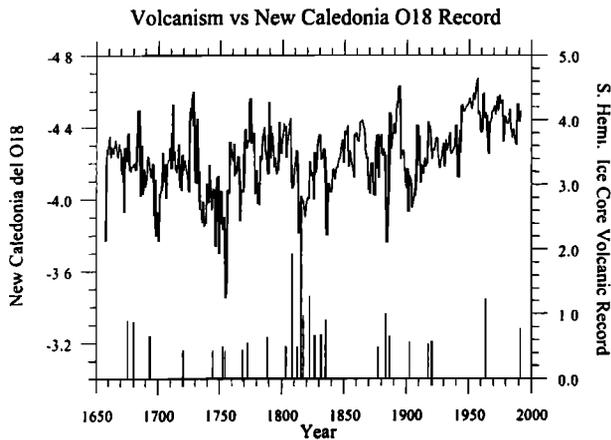


Figure 1. Comparison of mean annual $\delta^{18}\text{O}$ isotope (per mil) record (top) from a *Porites* coral head near Amedee Island ($22^{\circ}29'S$, $166^{\circ}27'E$), New Caledonia, with a composite southern hemisphere ice core record (bottom) of volcanic sulphur; note that low-latitude northern hemisphere eruptions can also affect the southern hemisphere (Table 1).

Because of the coincidence of some of the coral cooling events with ENSOs, it is therefore difficult to unambiguously ascribe all the volcano-cooling coincidences to cause and effect, especially given the uncertainty in some of the volcano identifications. It is also beyond the scope of this paper to address the more complicated relationship from the probabilistic viewpoint. However, given the well-established cooling following significant volcanic eruptions [Bradley, 1988; Angell, 1990], a volcanic cause for some of the New Caledonia cooling events seems quite plausible. Furthermore, some of the cooling events (1746, 1754, and 1836) are quite large compared with the magnitude of the El Niños estimated by Quinn [1992]. With respect to the largest interannual isotope fluctuations discussed above (1752/1754, 1836, and 1884), the correspondence is quite compelling for a strong volcano imprint. Evaluation of the smaller peaks would require more verification from future studies.

One interesting feature of the interannual isotope excursions involves comparison of the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ records from these events. Results (not shown) indicate a general tendency for $\delta^{13}\text{C}$ to become more positive when $\delta^{18}\text{O}$ increases (indicating cooler temperature). This response is opposite to what would be expected from a cooling large enough to cause water column overturn, as a negative $\delta^{13}\text{C}$ signal would be expected from such a response. As $\delta^{13}\text{C}$ variations in the symbiotic algae in corals [McConnaughey, 1989; Winter et al., 1991] have also been linked to changing photosynthetic activity due to light availability (cloudiness), one possible explanation for the observed $\delta^{13}\text{C}$ response at New Caledonia involves changes in cloudiness due to air mass changes associated with cooler temperatures. For example, increased outbreaks of high-pressure systems from higher latitudes could cause both cooling and decreased cloudiness. Because the entire subject of interpretation of $\delta^{13}\text{C}$ variations in corals is open to uncertainty and often contentious, this problem would have to be investigated further before any firmer conclusions could be made as to the explanation. We discuss it here for the record only.

In addition to the interannual cooling association, there are also several cases where volcanic events (1675, 1812, and 1902) occur just prior to pentadal-scale cooling events in the coral record. The most prominent cooling in the coral record from approximately 1813 to 1821 correlates with the most intense cluster of eruptions found in the composite ice core index. In fact, the interval 1803-1835 contains 9 of the 27 volcanic peaks identified in the composite ice core index. Lamb [1970] documents a further 1814 eruption of Mayon (Philippines) that cannot be distinguished clearly in the ice core records. This time interval was probably the coolest period in the last 200 years [Groverman and Landsberg, 1979; Bradley and Jones, 1993]. Because there is also some evidence for a solar-induced cooling in the early nineteenth century [Lean et al., 1995; Crowley and Kim, 1996], the total cooling in the 1813-1822 interval may reflect a combined effect of volcanism and solar irradiance decreases. However, the sharp onset of cooling seems most likely due to volcanism.

The prominent cooling from 1813 to 1821 is part of a longer pattern that occurs in the above hemispheric indices. There is a warm-cold-warm-cold set of pentadal/decadal-scale oscillations between about 1803 and 1837 occurring in these records, with the boundaries between the oscillations being 1813, 1821, and (approximately) 1835. Because of the similarity of the $\delta^{18}\text{O}$ changes in New Caledonia and northern hemisphere climate records over this longer interval, and the large geographic scale of the cooling to be expected from such an extensive amount of volcanic aerosol (and possibly solar) forcing, we tentatively suggest that the warm-cold-warm-cold oscillations may be used to refine chronologies in other coral records. To test this idea we illustrate (Figure 3) two other Pacific coral records [Dunbar et al., 1994; Quinn et al., 1996a], one from "nearby" Vanuatu (15°S , 167°E) and the other from the eastern equatorial Pacific (Galapagos, 0°S , 91°W). We also identify tentative time lines that could be used from our volcanic forcing index, assuming that the response from large-scale forcing will occur within a year.

Comparison of the different coral records suggests that all illustrated corals appear to record the unidentified volcanic

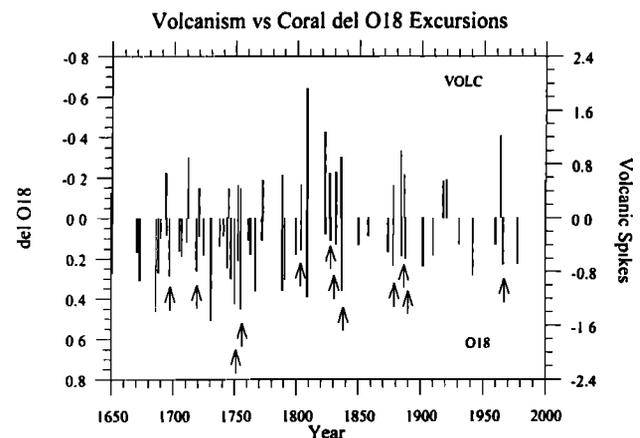


Figure 2. Comparison of interannual coral isotope excursions (bottom) with the subset of Table 1 conductivity/sulphate peaks (top) that do not coincide with pentadal decadal-scale cooling events in the coral record. Arrows denote El Niño-Southern Oscillations (ENSOs) that correlate with volcanic spikes. See text for further discussion.

eruption in 1808-1809 [Dai *et al.*, 1991; Zielinski, 1995]. Then there is the longer cooling subsequent to the start of the main pulse of eruptions in 1812. A prominent cooling in 1836 seems to mark the very large eruption on Coseguina (1835) and the end of the main burst of volcanic activity. Although earlier studies [Palais and Sigurdsson, 1989; Crowley *et al.*, 1993] suggested that the magnitude of the Coseguina eruption had been overestimated by Lamb [1970], there are clear sulphate/conductivity peaks in Antarctic ice core records [Legrand and Delmas, 1987; Moore *et al.*, 1991; Delmas *et al.*, 1992] indicating that this eruption resulted in a significant stratospheric perturbation. Zielinski [1995] and Langway *et al.* [1995] have also found evidence for the Coseguina eruption in Greenland ice cores (Greenland Ice Sheet Project 2 (GISP2) and Dye 3).

Results of this trial exercise (Figure 3) suggest that the chronology of the Vanuatu record may be 2-3 years too young and the Galapagos record 2-3 years too old in the sections analyzed. We emphasize again that the results from this exercise need to be treated with caution until they can be further tested with other records. Nevertheless, if this "event stratigraphic" approach is substantiated by other studies, it could be of considerable value in further refinement of coral chronologies for testing, for example, model predictions of phase relationships of decadal-scale oscillations in different parts of the Pacific Basin [cf. Latif and Barnett, 1994; Gu and Philander, 1997]. For example, a 1.7% lengthening in the chronology of the Vanuatu record could change the computed decadal oscillation from 14.5 to 14.8 years [cf. Quinn *et al.*, 1996b], bringing it closer in line to the 15.4 yr. periodicity determined by T. M. Quinn *et al.* (submitted manuscript, 1997) for the New Caledonia record.

Discussion and Conclusions

To summarize, evidence is quite good for the effect of large volcanic eruptions on the New Caledonia oxygen isotope record. Smaller eruptions may also have an impact, although it is not always easy to separate the volcanic signal from El Niño-induced cooling. There is also good evidence that the abrupt cooling in 1813 may have been triggered by volcanic eruptions, but solar irradiance changes may contribute to the full magnitude of the decadal-scale cooling. Finally, some of the large volcanic eruptions, and the decadal-scale warm-cold-warm-cold oscillations from about 1800-1840, may be useful for correlative purposes in coral studies.

The reason volcanic cooling events in corals have rarely been identified or discussed before is difficult to determine. For oxygen isotope time series dominated by precipitation variations [Cole *et al.*, 1993], the apparent lack of a volcanic cooling contribution to the records may reflect the fact that the precipitation response to volcanism is difficult to predict. Some investigators may simply have overlooked the relationship; it is difficult to detect unless mean annual averaging is used, and not all authors choose this option. Another consideration involves the observation that although some of Quinn's [1992] ENSO events occur in the New Caledonia record, almost two thirds do not. This lack of response may reflect the fact that New Caledonia is just west of the "hinge line" in the Southern Oscillation Index [Trenberth and Shea, 1987] separating regions of positive and negative

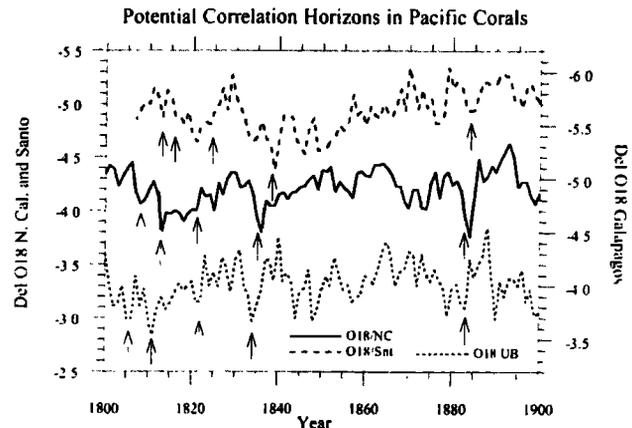


Figure 3. Comparison of volcanic "events" in New Caledonia $\delta^{18}\text{O}$ record (middle) with mean annual $\delta^{18}\text{O}$ variations at Espiritu Santo, Vanuatu [Quinn *et al.*, 1996b] (top) and Urvina Bay (UB, Galapagos) (bottom) [Dunbar *et al.*, 1994]. Arrows indicate proposed time stratigraphic horizons related to volcanic eruptions. These involve cooling associated with the 1808 eruption, the beginning (1813) and end (1821) of a decadal-scale cold event associated with several volcanic eruptions, and the coolings associated with the 1835 Coseguina and 1883 Krakatau eruptions. Note that the proposed correlation with the 1883 Krakatau eruption in the Galapagos record is in our opinion more uncertain than some of the other events.

response to Southern Oscillation changes. The decreased imprint of ENSO events in this region could therefore facilitate clearer detection of other sources of interannual variability. An additional factor to consider is that the response from volcanic eruptions may differ by latitude. For example, although radiative forcing from the 1991 Pinatubo eruption was greatest in low latitudes [Stenchikov *et al.*, 1997], the cooling resulting from the eruption persisted longer in higher southern latitudes [Dutton and Christy, 1992]. This cooling extended approximately to the latitude of New Caledonia. As the latter analysis utilized zonally averaged data, further work would have to be done to determine if there are preferred sectors of the southern subtropics that are influenced by cold air outbreaks from higher latitudes. Whatever the explanation(s), future studies of volcanic cooling events in coral records may be of interest not only to the paleoceanographic and climate communities but also to those interested in the effect of environmental change on coral populations.

Appendix

The composite record of volcanism from 1600 to the present [Table 1] was determined in the following manner. Prior work [Robock and Free, 1995, 1996] suggests that compositing of volcanic sulphur/conductivity spikes in ice cores represents a promising approach to development of an objective index of hemispheric volcanism. This approach was adopted in the present study. However, the methodology was somewhat different than that used by Robock and Free [1995, 1996], who used the standard deviation of short-interval peaks above local background [cf. Crowley *et al.*, 1993] and then averaged the values for different sites. In this study we took the published estimates of some peaks from two studies [Delmas *et al.*, 1992; Langway *et al.*, 1995] and then identified

additional peaks from three other cores by determining the magnitude of increase above local background. However because of chronology uncertainties [Legrand and Delmas, 1987] and different levels of noisiness in different cores, we defined peaks as only those peaks larger than a small peak that could be relatively unambiguously tied to a volcanic eruption from published volcano reference sources. For example, the 1803 Cotopaxi eruption was set as the lower cutoff in the Moore *et al.* [1991] time series. By contrast, only the largest peaks in the Legrand and Delmas [1987] time series could be identified with any confidence, especially since the timescale was more difficult to interpret in this record.

Relative magnitudes of the peaks were then scaled to the 1883 Krakatau eruption, except for the peaks in the Dai *et al.* [1991] core, which only included variations from the first half of the nineteenth century. Here the peaks were first scaled to Tambora and then rescaled to Krakatau based on the average Tambora/Krakatau ratios in other southern hemisphere ice cores. Compositing involved averaging values from only those cores that recorded a volcanic eruption. The rationale here is that wind scouring can remove the sulphate record from snow but can less easily magnify them, so only cores that contained peaks should be considered. Low latitude northern hemisphere eruptions are included because they can affect southern hemisphere climate. Finally, data from the 1991 Pinatubo eruption were added for completeness, based on relating estimated global stratospheric loading from McCormick *et al.* [1995] to the Agung peak, the latter which appears prominently in southern hemisphere ice core records (see also footnote f of Table 1).

To test the validity of our method, we first compared the results of our composite with those of Robock and Free [1995]

from 1850 to the present. Results indicated general agreement in relative magnitude for the largest peaks but some differences in the smaller peaks, perhaps because we only averaged sites for which peaks occurred, and not all cores record all the volcanic events. Another difference between our composite and that of Robock and Free [1995] is that their record has more volcanic events than ours. This subset of volcanic peaks in their composite record, which do not occur in our index, is consistently smaller in amplitude than our small peaks. This relationship suggests that the Robock and Free [1995] criterion for inclusion of volcanic "events" included a lower cutoff level than ours. The relative merits of the two approaches are a subject for further discussion. The approach of Robock and Free [1995] may include more eruptions but it also includes a number of peaks we could not relate to known eruptions. Our composite can be confidently tied to most eruptions since 1800, but it has a higher cutoff level that might exclude some legitimate volcanic peaks. We nevertheless believe our approach is a viable alternative. One could also argue that the smaller peaks included in the Robock and Free [1995] record, even if they all reflected volcanic eruptions, are likely to have a proportionately smaller climate impact. Since the latter is the objective of the present study, we believe our composite is of some merit, all the while recognizing it is subject to further modification.

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