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Technical details concerning development of a 1200-yr proxy index for global volcanism

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Abstract

This technical report describes details of developing a volcano forcing reconstruction (Crowley et al., 2008) for climate models that is based primarily on sulphate records in Antarctic and Greenland ice cores. The chronology of eruptions is considered accurate to within 1 yr for the interval AD 1104–2000 and 2 yr for AD 800–1103. The reconstruction involves: (1) calibration against satellite aerosol optical depth (AOD) estimates of the 1991 Pinatubo/Hudson eruptions; (2) partial validation against independent lunar estimates of AOD and global sulphate emissions; (3) partial assessment of uncertainties in AOD estimates; (4) assessment of possible tropical "false positives" in ice core reconstructions due to simultaneous occurrence of mid/high-latitude eruptions in each hemisphere; (5) identification of a new category of eruptions, termed "unipolar" tropical eruptions, in which the eruption plume penetrates mainly to polar regions in only the hemisphere of its eruption; (6) use of different growth curves for high- and low-latitude eruptions; (7) specification of 2/3 power shortwave scaling for eruptions larger than the 1991 Pinatubo eruption; and (8) compensatory introduction of an estimate of effective particle size that affects lifetime and scattering properties of stratospheric aerosols.

1 Development of time series of volcanic eruptions

1.1 Determination of volcanic peaks

Because each ice core site (Table 1) has unique features, a "one size fits all" scheme for automatically determining magnitude of sulphate peaks was abandoned in favor of point-by-point inspection of each record. The magnitude of a specific peak was determined by comparing it to the previous years of background variability, choosing the peak background variability as the baseline point for estimating excess flux for a particular event. Sometimes the baseline period can extend for several decades, in which case a confident assessment of peak amplitude can be obtained. Sometimes there are low-frequency trends in background variability, or pulses of volcanic eruptions, that make assessments more difficult. In this case an iterative approach was adopted until a reasonably reproducible estimate could be obtained for a particular peak.

Although all Antarctic cores have been measured for sulphate flux, there is a disappointing lack of published multiple site annual-scale estimates of sulphate flux for Greenland. Only the GISP2 record has a continuous sulphate record (Zielinski, 1995), and that is "only" at approximate biannual resolution. Two other cores (Crete and GRIP) have near-annual scale electrical conductivity measurements (ECM) and spot sulphate measurements. The spot sulphate measurements (Langway et al., 1995; Clausen et al., 1997) were used to scale the higher resolution ECM record in these cores to sulphate (n.b., annual sulphate levels for the North GRIP ice core have not been published). Three additional cores were used in the upper part of the record to better constrain estimates over the last two centuries and develop a bias estimate for the fewer number of records over the entire length of the series.

1.2 Chronology

Dating of Greenland cores can be extended with considerable confidence back to AD 1104 because of cross-checking with historical eruptions on Iceland (e.g. Hammer, 1977). For the interval 800–1103, uncertainty in Icelandic eruptions is greater – probably 1–2 yr for the "Settlement Event" around 871 and the large Eldgja fissure flow around 933. These are the best-guess ages determined by detailed intercomparisons of three high resolution Greenland ice cores (Vintner et al., 2006) and are also the ages adopted here. As far as can be determined, uncertainty in NH volcanic peaks therefore seems to be 1 yr.

Antarctic dating is more difficult, although a number of cores exist with annual-scale resolution. Still, false bands and missing bands can potentially occur in these records. Records can be cross-checked with Greenland for a number of large tropical eruptions – Krakatau (1883), Tambora (1815), Huaynaputina (1600), and the unknown but very large 1257/1258 eruption. Prior to that time Antarctic dating is more uncertain, for it

is difficult to find any confident matchups with the Greenland records, especially since "false positives" – simultaneous mid/high latitude eruptions in both hemispheres – are a distinct possibility. Cross-checking of Antarctic records (cf. Langway et al., 1995) indicates that the records seem to be consistent, but the fits are not airtight. Although it is not possible to make a precise uncertainty estimate of pre-1258 Antarctic peaks, an educated guess would be that the chronology of Antarctic eruptions prior to 1258 is correct to within 2–3 yr.

1.3 Development of composite records for Greenland and Antarctica

Sulphate peaks for the individual ice cores were first composited separately for Greenland and Antarctica. Sometimes, ice cores do not record events identified in other ice cores. This is usually the case for smaller volcanic eruptions, although a notable exception is Krakatau (1883), which cannot be found in the GRIP ice core. The average value for an event across all ice cores on an ice sheet includes "0" for cores that do not record the event. This approach mutes the impact of some volcanic events that can be large in one ice core but small or absent in other ice cores (e.g. large 1588 and 1829 GISP2 sulphate peaks).

Because Greenland is a much smaller ice mass than Antarctica, mean values were just compiled from available records. Some records are only from the upper part of a core, and temporal biasing can occur due to "dropout" of information in the older part of the time series. Subsets from the older interval were compared to the fuller data set over the last two centuries, and a bias correction was added to the initial value for the earlier intervals.

The Antarctic compilation was considerably more complicated. Peninsular and West Antarctic ice cores have substantially higher accumulation rates of both snow and sulphate, with mean sulphate accumulation rates for known eruptions about a factor of two higher than for East Antarctic sites. Yet these regions are only about 10 % of the area of Antarctica. These values was used to area-weight the contribution values from different regions and also estimate a West Antarctic component for eruptions older than the oldest annual-resolution ice core (about AD 1450, cf. Cole-Dai et al., 1997). Temporal bias estimates for Antarctica were calculated in the same manner as for Greenland.

1.4 Correction for latitude of eruptions

Local eruptions can have a disproportionate effect on volcano composites unless adjustments are made to the measured values in ice cores. Icelandic eruptions are particularly troublesome because much of the sulphate may be delivered through the troposphere. Because the index developed for this study is a record of stratospheric AOD, Icelandic eruptions (taken from www.volcano.si.edu/) were generally minimized or eliminated from the ice core composite unless some evidence could be found for larger-scale influence. For example, some Icelandic peaks can be found in ice cores from western Canada (Mayewski et al., 1990), although the values are much-muted from Greenland. In this case this far-field site was used to estimate the potential stratospheric input from some Plinian interludes of Icelandic eruptions; for example, the inferred stratospheric component of the 1783 Laki eruption is only about 15 % of the average sulphate loading for this event on Greenland. The procedure is not foolproof; some of the western Canada sulphur could have been transported in the troposphere across Asia and the North Pacific. As in all geological records, there may also be "site effects" that affects local deposition rates in northwest Canada. Nevertheless, the correction is almost certainly in the right direction and could even be considered a maximum stratospheric estimate due to the above caveats.

A somewhat similar approach can in theory be applied to Antarctic records, although the identification of local eruptions is more problematic. Kurbatov et al. (2006) have identified local ashes in some intervals of the Siple Dome ice core. Sulphate variablity in the Taylor Dome record (Steig et al., 2000) also seems strongly influenced by local (Antarctic) volcanism. In some cases (e.g. 843, 1170, 1710) the size of sulphate peaks in Taylor Dome are very much larger than found in any other Antarctic ice cores, and a local source for the eruptions can be fairly confidently established. If so, such eruptions

have been deleted from the Antarctic composite because the hemispheric effects are likely nil.

A final set of corrections involves large mid/high latitude eruptions (e.g. Japan, Siberia, Alaska) that are significant but still bias regional estimates due to proximity to the deposition sites. Clausen and Hammer (1988) pioneered the use of radioactive nuclides from bomb testing to estimate this local bias effect. Gao et al. (2007) updated the approach with more recent estimates of stratospheric injection of radioactive nuclides. They determined that, for a given size injection, there was about a 20 % "surplus" provided to Greenland cores due to proximity of high-latitude injections. This estimate is similar to the 20 % reduction in peak high-latitude NH AOD for the Pinatubo eruption. The consistency in both these approaches indicates that the critical low-latitude inputs of climatically-important tropical eruptions can be estimated with some confidence.

2 Calibration and validation against AOD

2.1 Calibration of mean sulphate values against stratospheric AOD

Mean sulphate values were scaled against AOD by matching against the maximum twelve-month value for 30–90° S AOD from satellite measurements (Sato et al., 1993; http://data.giss.nasa.gov/modelforce/strataer/). The comparison is only made for Antarctica, because there are not enough Greenland ice core measurements for the 1990's and because of greater noise from background tropospheric aerosol emissions and, we suspect, interannual deposition variability due to changes in the North Atlantic Oscillation.

Inspection of the 30–90° S Sato AOD estimate (Fig. 1) indicates an irregular shape that is due primarily to the additional impact of the August 1991 eruption of Hudson in Chile (VEI = 5; VEI refers to Volcanic Eruption Index – Newhall and Self, 1982). A further kink in the |AOD can be attributed to the April 1993 eruption of Lascar (VEI = 4), also in Chile. An iterative best-fit of likely growth and decay curves for the eruptions indicates that for the first 36 months after the June 1991 eruption of Pinatubo, 84 % of the AOD contribution can be attributed to Pinatubo, 12 % to Hudson, and 4 % to Lascar. The very modest contribution of Lascar to AOD – even in the latitude zone of its origin – suggests that even moderately powerful volcanic eruptions such as VEI = 4 have only very modest effects on the stratospheric AOD record of volcanic forcing.

The area-weighted (see above) mean sulphate flux for Antarctica is $10.1 \pm 0.5\,\mathrm{kg\,km^{-2}}$. The estimate was determined by comparing nine different estimates over a four-year period of study; the interval of study included a different number of ice cores, remeasurements of fluxes for different ice cores to test for consistency (see above), and also assessment of the averaging effect for inclusion of higher accumulation rate cores from West Antarctica (also see above). Since the maximum twelve-month (November 1991 to October 1992) estimated 30–90° S AOD from satellite estimates is 0.120, and the background level is estimated as 0.065, then the Pinatubo/HudsonAOD contribution is 0.1135 (ratio of 18 % of which is from Hudson). Rounding off slightly, the conversion rate for sulphate is therefore $0.011 \pm 0.0055\,\mathrm{AOD\,kg^{-1}\,km^{-2}}$. This is the value used as initial scaling of sulphate to AOD in our reconstruction.

2.2 Partial validation of mass estimates from ice cores

A mean flux of global sulphate can be determined by taking the mean Antarctic values, multiplying by 1.057 for the difference between mean twelve-month maximum Greenland AOD (0.120) and the 1.2 low-latitude multiplier discussed by Gao et al. (2007). This value is consistent with AOD estimates for low- and high-latitudes independently determined by Sato et al. (1993). The estimated mean global sulphate flux is 6.1 ± 0.3 MT sulphate for the interval November 1991 to October 1992 – the interval over which the Antarctic flux levels have been scaled to AOD. Note that the uncertainty is only that due to uncertainty is estimating mean values for Antarctic. Total uncertainties are almost certainly greater than that number but are difficult to constrain.

Two different approaches to initial injection levels of sulphur from the Pinatubo eruption yield values $27–28 \pm 6$ MT sulphate (Guo et al., 2004; the Hudson injection is

almost an order of magnitude lower, 3 MT). The Pinatubo value is four times the estimate from the ice core composite. However, Guo et al. point out that one-half of the initial injection had been scavenged within twenty days by volcanic ash and water vapor injected (or entrained) with the eruption plume. Any curve fits out to 1.5 yr based on the initial twenty days of operation yielded unrealistic values – either zero or levels well above that suggested by satellite observations. Additional processes, including dispersion of the dense low-latitude cloud to higher latitudes, may have significantly delayed the exponential decay based on just the first twenty days. What we can say that the mean 6.1 ± 0.3 MT value for that we obtain does not seem significantly out of line with the 14 ± 3 MT values remaining only three weeks after the 15 June 1991 eruption.

2.3 Life cycle of Pinatubo eruption

Satellite observations of Pinatubo (Fig. 2) were inspected to develop a latitude-dependent canonical growth and decay model for time evolution of the aerosol perturbation. Satellite AOD indicates that for the Pinatubo eruption peak low-latitude (0–30° N) AOD occurred about five months after the eruption, persisted for about three months, and then decayed with an e-folding time of about one year. For the purpose of the reconstruction, AOD values increase linearly for five months, maintain a plateau for the next three months, and then decrease with an e-folding time of one year for tropical eruptions.

The Pinatubo reconstruction for 30–90° N indicates that it took approximately nine months to reach maximum values, which then persisted for approximately three months before decaying with an e-folding time of approximately one year. The 60–90° S time series for Pinatubo indicated that it took nearly a year after the eruption to reach maximum values over Antarctica. These observations were incorporated into the algorithm transferring sulphate fluxes into AOD.

2.4 Comparison with global AOD index

Using observations and methodologies discussed above, a global reconstruction (Fig. 3) was developed from the Antarctic sulphate data and compared to the Sato AOD global composite. Lunar eclipse estimates of AOD for the Pinatubo eruption (courtesy R. Keen, based on method of Keen et al., 1983) were further compared as a check on the reconstruction. The good agreement of the ice-core reconstruction with the satellite reconstruction merely demonstrates that the scaling was done properly. Although there is a slight underestimate of peak forcing (due to the non-standard evolution of the 30–90° S signal), the four-year average AOD, and therefore average radiative forcing, for the ice core reconstruction is identical to Sato et al. (1993). This agreement is very important because the total forcing of a reconstruction is very important for correct computation of ocean heat storage changes. The agreement with the lunar eclipse data is very good for two of the eclipses but quite poor for the first eclipse. We suspect that the size of the aerosol particles at this time may significantly bias lunar AOD estimates (see further discussion below).

3 Development of millennial-scale reconstruction

3.1 Problem of "unipolar" tropical eruptions

Despite the large-size of the 1982 El Chichon eruption, its aerosol is hardly detected over Antarctica, either in the satellite or ice core data. A similar situation applies to the 1963 Agung eruption and the 1902 Santa Maria (Caribbean) eruption. These "unipolar" tropical eruptions are difficult to determine in earlier time series and, because of the climatic importance of tropical eruptions, are a sources of significant uncertainty in ice core reconstructions. Based on the experience with the 20th century unipolar eruptions, this category appears to be occur for medium-large sized volcanic (VEI = 5) eruptions. Their identification in older records is ambiguous. If small sulphate peaks

occur in the opposite hemisphere to a large sulphate peak, it could indicate an eruption of this type. It could also be coincidental. Large cooling persisting for more than a year in proxy records might be some way of inferring a possible tropical source, but then the volcano and climate reconstructions would not be independent.

The relative magnitude of tropical forcing also differs from true bipolar eruptions. As indicated by data from Agung and El Chichon, low latitude forcing is much more restricted to the hemisphere of the eruption, whereas for Pinatubo the southern tropical strip has virtually the same value as the northern tropical strip. In our reconstructions we therefore stipulate that the opposite hemisphere to a unipolar eruption has a tropical value midway between the tropical hemisphic maximum and the very low values observed in polar regions.

In the present study unipolar eruptions were inferred for only a few events, in which the mean flux for one polar region is about 10 kg km^{-2}, at least three "hits" of about 2–3 kg km^{-2} can be found in the opposite hemisphere record, and there is some indication from Northern Hemisphere tree ring records that cooling persisted for more than one year. Because of the very significant impact of volcanic eruptions, and the precarious nature of their identification in ice core records, only a small number of events were identified as candidates for unipolar eruptions. Further work will hopefully reveal a fuller list of criteria to enable more confident assessment of this phenomenon.

3.2 Duration of mid/high latitude aerosol plumes

The time evolution of high-latitude eruptions is different than the low-latitude extension of Pinatubo and is was based on Stothers's (1996) reanalysis of data for the 1912 Novarupta eruption in Alaska and cross-checked against the August 1991 eruption of Hudson in Chile (Fig. 1). Stother's analysis indicates that peak values occur within two months of the eruption, persists for three months, and then decays more rapidly than for an equatorial eruption – the e-folding time for the first step in only three months, but it is nine months for the second step. The longer second step may reflect input from the 20 January 1913 eruption of Colima (19° N) in Mexico.

The Stothers' result is supported by a comparison of biweekly-scale ECM measurements of the 1912 interval in the GISP2 ice core (Fig. 4). The difference between the ECM and AOD estimates for the interval around 1913.0 could reflect either residual higher values of aerosols in the highest latitudes (the core is at 73° N – farther north than any measurement station in Stothers' analysis), superposition of the annual cycle of higher accumulation in the ice core winter (note magnitude of annual cycle in later intervals of the ECM record), or some combination of the two. The Stothers' result is also consistent with the deconvolved inference of the Hudson aerosol plume (Fig. 1). Even allowing for uncertainties due to mixing of Pinatubo and Hudson plumes, the deconvolution of the Hudson plume suggests peak AOD three months after the eruption and a three-month e-folding time for the decay phase. One possible explanation for the more rapid evolution of high-latitude plumes may involve a smaller volume of air containing debris from the eruption (thereby leading to more frequent collisions of particles and faster growth rates and dropout of aerosols). The rule for mid-high latitude eruptions was therefore specified as having a 3.5 month ramp, a peak "plateau".5 months long, and plateau peak, with peak values 1.3 times the initial value (rather than 1.2 for Pinatubo), and an e-folding time of 0.67 yr (we would need to examine more decay times for mid-high latitude peaks before adopting the very short determined for Novarupta).

Stothers' analysis of the 1912 eruption also suggests that, despite its clear power (VEI = 6), the aerosol plume did not reach equatorial regions. Ground-based British AOD observations in Cairo (30° N) indicate the perturbation was barely detectable at that latitude. This result has been generalized to apply to all high-latitude eruptions, plus mid-latitude eruptions (a clear mid-latitude Northern Hemisphere eruption (Japanese eruptions of 1667 and 1739) has not been identified in Antarctic sites).

3.3 Additional scaling for large eruptions

Because collisions between aerosol particles result in size increases, shortwave radiative forcing is reduced by the 2/3 power. This scaling is implicitly taken into account for

eruptions up to the size of Pinatubo (peak AOD about 0.180). Above that level some additional scaling is required. In this study, 2/3 power scaling was applied to all values greater than or equal to 0.200 AOD, as first determined by a linear estimate of AOD based on the mass/AOD relationship discussed above. We tested this approach for the large 1912 eruption (Fig. 5); the results are consistent within the range of uncertainty, but more studies would have to be made to test this assumption further.

3.4 Aerosol size

Aerosol size affects lifetime and scattering properties (Pinto et al., 1989), significantly affecting in particular the climate impact of very large eruptions (e.g. Timmreck et al., 2009). Based on the relation between AOD and Reff (effective radius) in the GISS reconstruction for Pinatubo (Fig. 6) – notably the lag between peak Reff and AOD, and estimates of maximum particle size from Pinto et al. (1989), a simple calculation captures many of the observed properties of Reff in Pinatubo. Because aerosol coagulation results in AOD scaling to the 2/3 power of mass, Reff was determined based on the 1/3 power of AOD. During non-volcano years, Reff was set at 0.200, consistent with the methods of Sato et al. (1993; http://data.giss.nasa.gov/modelforce/strataer/).

3.5 Time of year for eruptions

The Smithsonian volcanism website (www.volcano.si.edu/) provides a wealth of information on eruption dates for more recent eruptions – these were stipulated in the reconstruction. For most unknown eruptions, a start date of 1 January was used as a default. However, if an ice core record of a postulated tropical eruption had maximum sulphate values in year two of the signal, then a mid-year (1 July) start date was inferred (based on observations for length of time it took the Pinatubo signal to reach polar regions). Subsequent to finalization of the present volcanic index, it was discovered that very high-resolution (bimonthly-scale) ECM time series might be used to also estimate the month/season of an eruption, because in Greenland there is a small annual winter peak in ECM associated with higher snowfall rates. For example, the onset of the great “1258” eruption at the beginning of 1258 in the ice core ECM record would, by analogy with the length of time required for the first-traces of the Pinatubo cloud to spread to Greenland and Antarctic, suggest an eruption around September 1257, with an approximate uncertainty of 2–3 months. This result was tested on the AD 79 Vesuvius eruption, with estimated time of eruption consistent within 2–3 months with historical observations. Due to the lateness of the discovery, and the need to release the data for climate simulations (Jungclaus et al., 2010), the methodology has been applied only to the 1257 eruption, but it could with some effort be applied to other eruptions.

3.6 Completed AOD reconstruction

Following the original example of Sato et al. (1993), all of the above information was used to develop a four-equal-area band of AOD and Reff for the interval AD 800–2000 (see uploaded software program for code). We used our forcing estimates from the ice core composite up to 1960, Stothers’ (1956) reanalysis from 1960–1980, and the NASA/GISS satellite reconstruction afterwards. The Stothers’ analysis raises questions about the reliability of the inferred 1968 Fernandes (Galapagos) eruption (VEI = 4) inferred by Keen (1983) based on lunar AOD estimates. We agree. We also would not expect a high-temperature magma characteristic of ocean spreading centers to produce eruptions powerful enough to penetrate the equatorial tropopause (about 20 km). Based on our experience, we also do not think that a VEI = 4 eruption has as much of a radiative impact as calculated by Hansen. We have, however, added in our reconstruction a small eruption identified by Stothers in the 1960s that could be of tropical origin, possibly from Indonesia (Awu?), where Plinian eruptions are common.

Since climate models often update radiative forcing on a time step of approximately 10 days, the reconstruction was partitioned into time steps of 1/36 yr and is on the Paleoclimate Model Intercompariosn Program (PMIP) link (https://pmip3.lsce.ipsl.fr/wiki/lib/exe/fetch.php/pmip3:design:lm:ici5_aod_reff_c.zip).

3.7 Comparison with Gao et al. reconstruction

While it is beyond the scope of this paper to discuss in detail the features of the final reconstruction (Fig. 7), we note that the reconstruction show more or less than the standard pulses of volcanism that have been documented before, with the differences being the hopefully improved calibration of the AOD and addition of effective radius size. Schmidt et al. (2011) illustrate a comparison with the Gao et al. (2007) reconstruction, which also uses cores from sources not available for the present study (the sources were co-authors on that paper). Although there is general agreement in timing of major eruptions over the last 800 yr (with the notorious lack of unquestionble equatorial peaks creating a problem before then – see Sect. 1.2), there are a few significant differences.

There are a number of specific differences for eruptions in the two reconstructions. The Gao et al. reconstruction generally has higher values of AOD for the largest historical eruptions – a feature we suspect is due to the 2/3 power scaling used in our study. The alleged global reach of the 1831 Babuyan eruption has also probably been overestimated; improvements in chronology and identification of regional eruptions see above) suggest that an Antarctic peak about that time actually was deposited in 1833 and appears to be of local origin. We also do not have a large AOD for the 1783 Laki eruption, because we think most of it is tropospheric in origin (see Sect. 1.4). Anciki et al. (2011) recently report sulphur isotope evidence that virtually none of the Laki sulphur deposited in Greenland was in contact with the stratosphere.

Our estimate of the powerful Krakatau eruption again suggests values not much greater than Pinatubo, in agreement with Gao et al. An additional test of this conclusion is illustrated in Fig. 8, which compares lunar AOD estimates from R. Keen (personal communication) to the enlarged growth and decay curve for that interval. The lunar eclipse 14 months after the eruption agrees almost perfectly with our canonical growth and decay curve, while the eclipse about seven months after the eruption shows slightly lower values than we estimated. We suspect that this is in part due to the fact that use a standard nine month buildup for the Krakatau plume to reach northern high latitudes, whereas lessons from the Pinatubo eruption suggest that it probably took 11–12 months to reach the polar hemisphere farthest away from the volcano (see Fig. 2).

The 13th century pulse of eruptions originally identified by Langway et al. (1995) also shows some differences; at least part of this is (we suspect) due to Taylor Dome indications (again, see Sect. 1.4) of at least one coincidental local eruption at the time of one of the Greenland eruptions previously inferred by all investigators (including ourselves – Crowley, 2000) as being tropical in origin.

There is also a large difference in estimate of the 1453 "Kuwae" peak (Gao et al., 2006). In our reconstruction we do not find a large signal in Greenland records. There is a large signal in 1458/1459 of our reconstruction, and the GRIP ECM reconstruction (Vinter et al., 2006). We interpret this as a high-latitude Northern Hemisphere eruption and base the chronology justification on proximity of a very well dated 1477 Icelandic event clearly evident in ice core records – given that the AD 79 Vesuvius signal occurs within a year of band the band-counted signal in the GISP2 and GRIP ice cores, we cannot conceive of a 5 yr error occurring only fifteen years prior to a clearly identified reference level. Nevertheless, we concede that the subject warrants further discussion. Overall, we consider that the comparison between the two reconstructions gives a "reasonable" level of agreement.

Supplementary material related to this article is available online at: http://www.earth-syst-sci-data-discuss.net/5/1/2012/essdd-5-1-2012-supplement.zip.

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Table 1. Location and duration in years AD of sites used in this study.

NGT_B20	79° N	36° W	830–1993	Bigler et al. (2002)
GISP2	73° N	38° W	1–1984	Zielinski (1995)
Site T	72° N	38° W	1731–1989	Mosley-Thompson et al. (1993)
Crete	71° N	37° W	553–1778	Hammer (1977)
Site A	70° N	36° W	1715–1985	Mosley-Thompson et al. (1993)
Mt. Logan	60° N	141° W	1689–1979	Mayewski et al. (1990)
Law Dome	66° S	112° E	1301–1995	Palmer et al. (2002)
Dyer	70° S	65° W	1505–1989	Cole-Dai et al. (1997)
Dome C	74° S	124° E	1763–1973	Legrand and Delmas (1987)
DML B32	75° S	0° W	159–1997	Traufetter et al. (2004)
DML B31	75° S	4° W	463–1994	Sommer et al. (2000)
Siple Station	76° S	84° W	1417–1983	Cole-Dai et al. (1997)
ITASE 01-5	77° S	89° W	1781–2002	Dixon et al. (2004)
ITASE 00-5	77° S	124° W	1708–2001	Dixon et al. (2004)
ITASE 00-4	78° S	120° W	1799–2001	Dixon et al. (2004)
ITASE 01-3	78° S	95° W	1859–2002	Dixon et al. (2004)
ITASE 00-1	79° S	111° W	1651–2001	Dixon et al. (2004)
ITASE 99-1	80° S	122° W	1713–2000	Dixon et al. (2004)
Plat. Remote	84° S	43° E	1–1986	Cole-Dai et al. (2000)
SP2001	90° S		905–1999	Budner and Cole-Dai (2003)
Taylor Dome	78° S	159° E	1–1957	Steig et al. (2000)

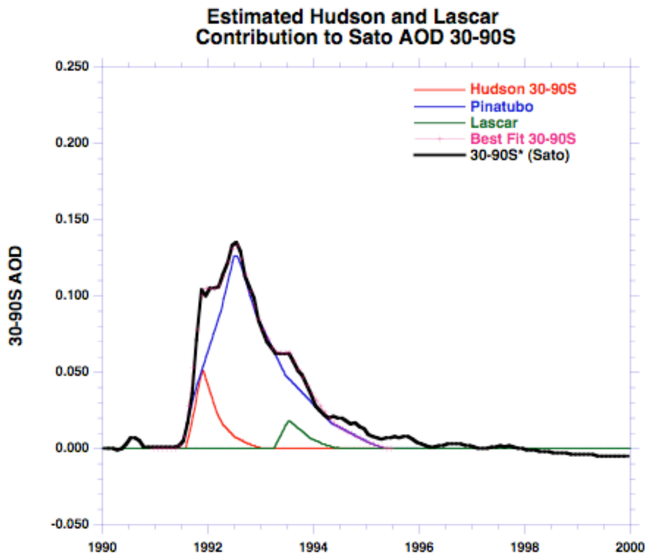

Fig. 1. Best fit iteration of contribution of smaller Southern Hemisphere eruptions to the zonal mean 30–90° S values of Sato et al. (1993).

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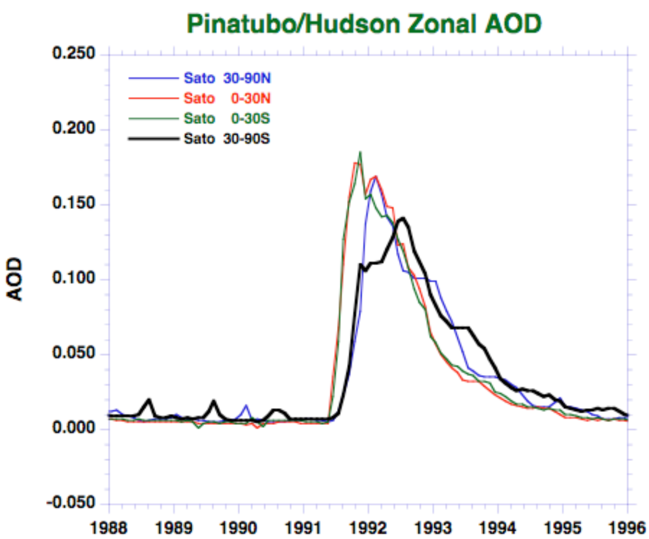

Fig. 2. Observed zonal variability of the Pinatubo plume from Sato et al. (1993).

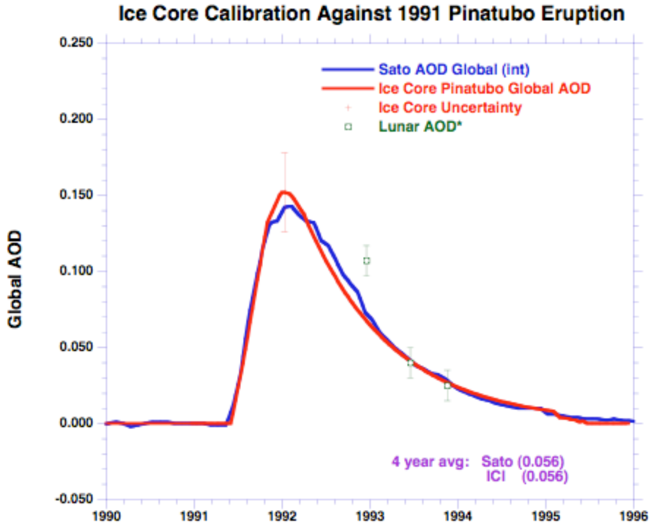

Fig. 3. Calibration of ice core "Pinatubo" plume against Sato et al. (1993) global AOD, using rules discussed in the text. Also shown are lunar AOD estimates from R. Keen (personal communication).

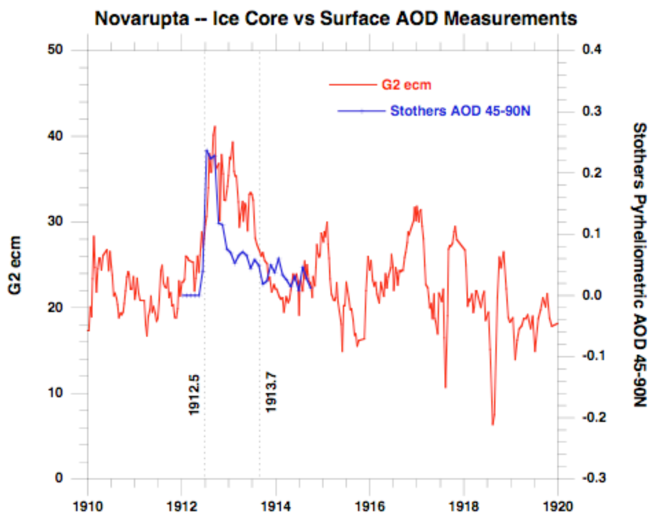

Fig. 4. Supporting evidence of a shorter life cycle for high latitude eruptions based on Stother's (1996) reanalysis of surface observations for 1912 Novarupta eruption and bimonthly-scale electrical conductivity measurements (ECM) from the GISP2 ice core.

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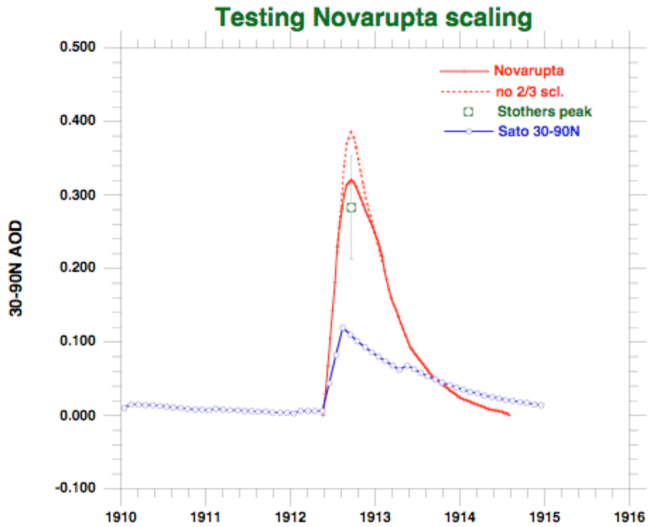

Fig. 5. Independent test of ice core AOD calibration, and 2/3 power scaling rule, against Stother's (1996) reanalysis of 1912 Novarupta eruption. For reference, the Sato et al. (1993) values are also shown.

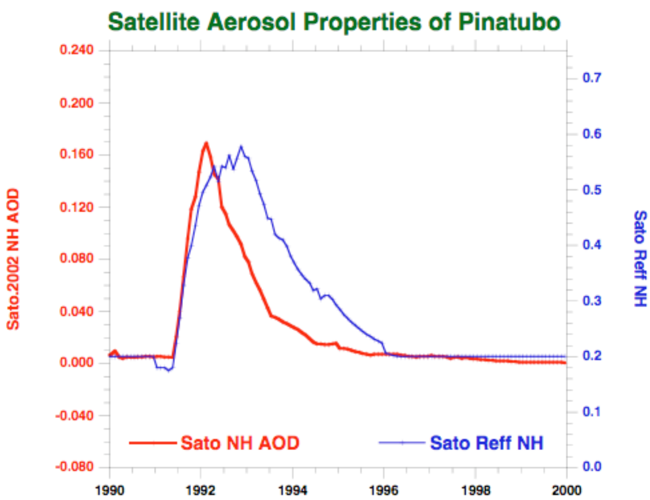

Fig. 6. Comparison of observed AOD and Reff for Pinatubo, indicating the lag in Reff that probably reflects dispersion and then agglomeration of aerosol particles.

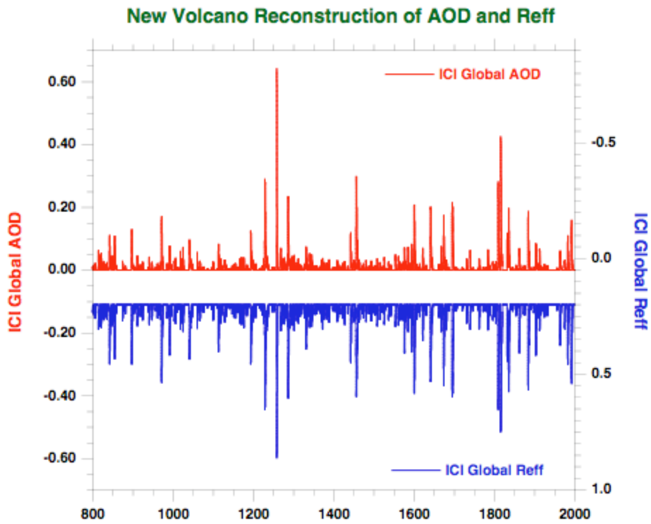

Fig. 7. Comparison of reconstructed AOD and Reff for new ice core reconstruction (ICI stands for "ice core index"); note that Reff is plotted on reverse scale to enable easier comparison.

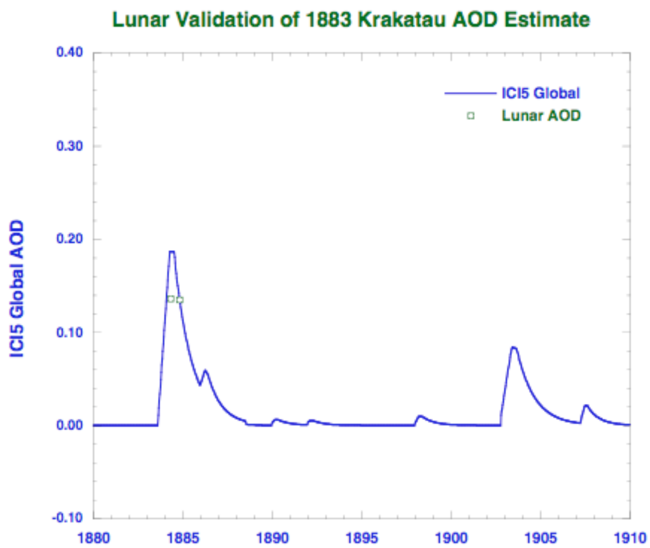

Fig. 8. Further independent test of AOD calibration for 1883 Krakatau eruption, using lunar AOD data from Richard Keen (personal communication). Note that the second "bump" in the reconstruction reflects the addition of new aerosol from the the 1886 Okataina eruption on New Zealand.

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