Major Volcanic Eruptions and Climate: A Critical Evaluation

CLIFFORD F. MASS AND DAVID A. PORTMAN*

Department of Atmospheric Sciences, University of Washington, Seattle, Washington

(Manuscript received 22 September 1987, in final form 18 November 1988)

ABSTRACT

This paper examines whether major volcanic eruptions of the past century have had a significant impact on surface land and ocean temperatures, surface pressure and precipitation. Both multieruption composites and individual eruption time series are constructed and analyzed. Included in this work is an attempt to remove one source of interannual variability: the El Niño/Southern Oscillation (ENSO). These exercises indicate that only the largest eruptions (in terms of producing a stratospheric dust cloud) are suggested in the climatic record. Removing the ENSO signal in the composite and individual eruption series enhances the apparent volcanic effect of the largest eruptions. No volcanic signal is obvious in pressure and precipitation records.

1. Introduction

During the past few decades much has been written about the effects of volcanic eruptions on climate. Although the majority of the literature suggests a significant volcanic signal, most often in surface air temperature (e.g., Lamb 1970; Mass and Schneider 1977; Taylor et al. 1980; Self et al. 1981; Kelly and Sear 1984), several papers have expressed doubts about the existence of a volcanic influence on surface climate. For example, Landsberg and Albert (1974) provided evidence that the great eruption of Tambora may not have caused "the year without a summer" of 1816, Gentilli (1948) noted the small effects of volcanic eruptions on total radiation, and Ellsaesser (1977) questioned whether Mount Agung produced a global temperature anomaly in 1963–64. After examining the temperature evolution accompanying the largest eruptions since the late eighteenth century, Angell and Korshover (1985) concluded that "while volcanic eruptions certainly do not cause warming, the evidence that they cause cooling is not overly impressive." A careful reading of the literature proposing a volcanic impact on climate provokes questions about the selection of eruptions, the statistical techniques used for evaluating volcanic signals, and the adequacy of the databases applied. These problems, coupled with the apparent lack of a significant cooling following the recent El Chichón eruption (which created a dense stratospheric dust cloud in the Northern Hemisphere), suggests that a careful reevaluation of the meteorological effects of volcanic eruptions is in order.

This paper attempts to appraise the reality of volcanic influence on surface climate by examining the response of surface air temperature, sea surface temperature, surface pressure and rainfall after major volcanic events of the past century. Both multieruption composites and individual eruption time series are constructed and analyzed. Included in this work is an attempt to remove one source of interannual variability: the El Niño/Southern Oscillation (ENSO). These exercises suggest that only the largest eruptions (in terms of producing a stratospheric aerosol cloud) are apparent in the climatic record.

2. Recognition of a volcanic signal

a. Expected characteristics

How does one identify the effects of a volcanic eruption on temperature, precipitation and rainfall? A review of the literature on the subject, coupled with knowledge of the earth's climate system, suggests some guidelines for identifying the presence of a volcanic signal in multieruption composites or single-eruption time series. Specifically, a volcanic signal:

1) should first appear *after* an eruption. An anomaly that is a continuation of a trend established before an eruption can not be assumed to be of volcanic origin.

2) should be observed within 2–3 yr after an eruption. Several studies (e.g., Wexler 1951; Lamb 1970; Roosen et al. 1973) have shown that little volcanic aerosol remains in the stratosphere after this period.

* Present affiliation: Atmospheric and Environmental Research, Inc., Cambridge, Massachusetts.

Corresponding author address: Dr. Clifford Mass, Dept. of Atmospheric Sciences, University of Washington, Seattle, WA 98195.

3) a multieruption composite should be similar to the individual component eruptions in magnitude and sign. For example, a posteruption temperature drop in a multieruption composite could result from a large (and perhaps random) variation associated with only one or a few events.

4) should maintain the same sign and approximate magnitude for several months since volcanic aerosols spread relatively uniformly within the affected latitude zones. Thus, short-period excursions lasting only 1 or 2 months are probably not of volcanic origin.

Intercomparisons between several climate parameters can also help appraise the reality of a potential volcanic signal. For example, because of the large heat capacity of the ocean, the response of sea surface temperature to volcanic events might be expected to lag and be of lesser magnitude than variations in air temperature (McCracken and Luther 1984; Harvey and Schneider 1985).

b. *The importance of ENSO-related variability*

In order to determine whether volcanic eruptions produce climatic change, it is helpful to first consider other sources of interannual variability that might mask or mimic a volcanic signal. One important source of such variability is the El Niño/Southern Oscillation (ENSO). During a canonical ENSO event, which typically begins during a Northern Hemisphere spring, an area of anomalously warm surface water is observed in the eastern Pacific and subsequently expands westward during the next 6 months (Rasmusson and Wallace 1983; Philander 1983). Associated with the warm water is above-normal sea level pressure in the western Pacific, greatly enhanced precipitation at equatorial stations to the east of 160°E, weakening or reversal of the easterly winds in the equatorial eastern Pacific, and teleconnections with weather anomalies in the extratropical latitudes. The increased sea surface temperatures that accompany ENSO events are reflected in global tropospheric temperature anomalies (e.g., Wright 1977; Angell and Korshover 1978; Newell 1979; Angell 1981; Horel and Wallace 1981; Quiroz 1983; Pan and Oort 1983). ENSO events occur at irregular intervals and typically last 1 to 2 yr. A comprehensive chronology of ENSO events over the past century is presented in Quinn et al. (1978) and is summarized in Table 1.

Many of the single-eruption time series and multieruption composites presented in the next section reflect the influence of ENSO events. For example, for the eruption of Mt. Agung (Fig. 18) there are large air temperature rises in the latitude band of the eruption only in 1957 (yr −6), 1965 (+2), 1969 (+6), and 1972 (+9); all are ENSO years and, with the exception of 1969, all are strong or moderate events (Table 1).

TABLE 1. Chronology of ENSO events.*

1873	1875	1877–78**	1880**	1884–85**
1887–89**	1891**	1896**	1899–00**	1902*
1905**	1911–12**	1914**	1917	1918–19**
1923	1925–26**	1929–30**	1932	1939–40
1941**	1943–44	1946	1948	1951
1953**	1957–58**	1963	1965**	1969
1972–73**	1975	1976**	1977–78**	1982–83**

* From Quinn et al. (1978).
** Strong or moderate events.

Sea surface temperatures in this band undergo increases of 0.2°C or more in 1957 (−6), 1959 (−4), 1963 (0), and 1972 (+9), all but one an ENSO year. For sea surface temperature the largest eruption band increases were in 1895, 1896, 1899, 1900, 1905, with only one (1895) being a non-ENSO year. Many additional examples could be provided to show that ENSO signals are apparent in both the air and sea surface temperatures and that warm years are generally associated with ENSO events.

c. *Estimating the climatic impact of volcanic aerosols*

Deirmendjian (1973) and others have noted that although stratospheric aerosols from volcanic events can dramatically attenuate the direct solar beam (by up to 20% or 30%), a concomitant increase in downward-scattered radiation results in a relatively small decrease in global or total radiation. This occurs because volcanic stratospheric aerosols scatter most of the intercepted solar radiation in the forward direction (Pollack et al. 1976; Turco et al. 1982). Actinometric measurements indicate that volcanic stratospheric aerosols decrease total radiation by up to 5%–7% for a limited period of time in the polar latitudes of the eruption hemisphere. For example, Wendler (1984) noted a 5.6% decrease in total radiation at Fairbanks, Alaska for 5 months (November 1982–March 1983) after the eruption of El Chichón; this attenuation decreased to ~3.5% by April–May 1983. After the eruption of Mt. Agung in 1963, Viebrock and Flowers (1968) observed a ~7% reduction in total radiation at the South Pole during the first year, which decreased to ~2% the following year. In contrast to these polar observations, the perturbation of total radiation at Aspendale, Australia (38°S) after the Mt. Agung eruption barely rose above the noise level (Dyer and Hicks 1965). (It should be noted that the El Chichón and Mt. Agung eruptions produced denser and more extensive stratospheric aerosol clouds than most of the other eruptions considered in this study.)

The potential climatic impact of volcanic eruptions is also reduced by the inhomogeneous spread of volcanic dust. The largest attenuation of solar radiation does not occur globally, but rather is confined predominantly to the eruption hemisphere. Even in the eruption hemisphere the stratospheric aerosols do not spread uniformly. For example, following the recent El Chichón eruption (17°N) the dust veil spread quickly between the equator and 30°N, but then slowed down dramatically; 6 months after the eruption the dust veil was apparent only from 10°S to 35°N.

It is clear from a variety of studies (e.g., Turco et al. 1982) that most of the conversion from sulfur dioxide to sulfuric acid aerosol occurs quite rapidly, apparently within the first few months after an eruption. By the first anniversary of the volcanic event the aerosol cloud is already in considerable decline, and after 2 or 3 yr aerosol levels approach those of the preeruption stratosphere (Volz 1970; Cadle et al. 1976). A range of theoretical studies (e.g., Schneider and Dickinson 1974; Chou et al. 1984; Harvey and Schneider 1985) have shown that 4 to 5 yr is required for the ocean mixed layer to reach a steady-state temperature after a change in solar forcing. Therefore, with a response time considerably longer than the time-scale of the radiation perturbation, the earth's climate system cannot realize the maximum cooling implicit in simple radiative calculations.

The cooling accompanying volcanic events is also mitigated by the volcanic aerosol's infrared opacity and the associated "greenhouse" warming. Detailed radiative calculations of Pollack et al. (1976) and Chou et al. (1984) indicate that such infrared warming reduces volcanic cooling at the surface by about 40%. Similarly, Harshvardhan (1979) found that infrared warming by stratospheric aerosols reduces the net radiative loss of an atmospheric column by approximately 30%.

Many references (e.g., Lamb 1970; Pollack et al. 1976) have discussed the influence of volcanic eruptions on the optical depth and radiative properties of the atmosphere. There is, in fact, general agreement that large volcanic events can enhance optical depths by as much as 0.1 for a period of a year over the eruption hemisphere, and that the largest events (e.g., El Chichón) can enhance optical depth by as much as 0.3 within limited areas of the eruption hemisphere during the first year. To accurately model the radiative effects of such optical depth perturbations it is essential to consider latitudinal, seasonal and variable albedo (surface and cloud) effects. Such detailed calculations, as presented in Harshvardhan (1979), show that the change in albedo due to a hemispherically uniform stratospheric aerosol varies substantially by latitude and time of year. For example, an optical depth enhancement of 0.1 over an entire hemisphere results in an albedo increase of less than 1% in the midlatitudes during the warm months of the year and in the tropics during most of the year; in the polar regions such an aerosol cloud increases albedo by as much as 10%. Averaged for an entire year and a complete hemisphere the albedo enhancement for such a case would be approximately 1% to 1.5%. These theoretical results are consistent with actinometric observations in which attenuation of total radiation is larger in the polar than lower latitudes.

Considering previous observations of volcanic dust veils and current theoretical knowledge of their radiative effects, what is the best estimate one can make of the magnitude of surface temperature anomalies that should accompany the largest volcanic eruptions? For a large volcanic eruption with a dense stratospheric dust veil let us assume a hemispheric aerosol cloud with a spatially uniform optical depth of 0.1. Using the results of Harshvardhan (1979), such a cloud will increase hemispheric albedo by 1%–1.5%, which we will reduce to 0.7%–1.05% (average of ~0.9%), to take account of the mitigating infrared effects noted above. In Harvey and Schneider (1985) a seasonal, land–ocean, energy-balance climate model is used to explore the temperature effects of a 2% reduction in solar radiation produced by a volcanic event. Their simulations begin with the maximum solar radiation perturbation, which then falls off with time. Scaling their results to a 0.9% reduction in net total radiation and assuming an aerosol half-life of 1 yr produces maximum negative anomalies in land–air temperature of 0.34°C (Northern Hemisphere) and 0.27° (Southern Hemisphere) after about 6 months into the first year. Over the ocean the surface air temperature anomalies reach only 0.14°C (Northern Hemisphere) and 0.16°C (Southern Hemisphere) approximately a year later. These volcanic temperature anomalies may be too large since volcanic stratospheric aerosols do not immediately "turn on" throughout the entire eruption hemisphere, but develop and spread during the first 6 months and are in considerable decline by the first anniversary of the eruption (e.g., Volz 1970; Cadle et al. 1976).

The implication of this analysis is that the surface temperature signal of even the largest eruptions should be weak, with maximum negative surface temperature perturbations over an entire hemisphere not expected to exceed ~0.2°C. Considering the limited number of large eruptions, the substantial noise and errors resident in current datasets, and the existence of other sources of short-term variability (such as ENSO events), it is not surprising that finding a clear volcanic signal in surface temperature is a challenging exercise.

3. A compositing study of volcanic influence

a. Methodology

In this study the method of superposed epoch analysis (also known as compositing) is used to search for a volcanic signal in surface air temperature, sea surface temperature, precipitation and surface pressure. As noted in Mass and Schneider (1977) and Taylor et al. (1980), this technique is based on the temporal averaging of data with respect to specified "key dates" and

TABLE 2. Major volcanic events selected for this study.

Event	Latitude	Date	DVI	VEI	Greenland ice[1]	Antarctic sulfate[2]	Ice Pb/Zn[3]	Turbidity[4]	Lunar eclipse[5]
1. Krakatau	6°S	8/1883	1000	6	yes	yes	yes	yes	n.a.[6]
2. Tarawera	38°S	6/1886	800	5	yes	yes	no	yes	n.a.
3. Mount Pelee	15°N	5/1902	100	4	no	no	no	no	n.a.
Soufriere	13°N	5/1902	300	4	no	no	yes	yes	n.a.
Santa Maria	15°N	10/1902	600	6	no	no	yes	yes	n.a.
4. Ksudach	52°N	3/1907	500	5	no	no	no	yes	n.a.
5. Katmai	58°N	6/1912	500	6	yes	no	no	yes	n.a.
6. Agung	8°S	3/1963	800	4	yes	yes	yes	yes	yes
7. Awu	4°N	8/1966	200	4	no	yes	no	yes	no
8. Fuego	15°N	10/1974	250	4	no	yes	yes	n.a.	yes
9. El Chichón	17°N	4/1982	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	yes

[1] Hammer et al. (1980).
[2] Delmas and Boutron (1980).
[3] Boutron (1980).
[4] Dyer (1974).
[5] Keen (1983).
[6] n.a. indicates not available.

can be used to identify weak, nonperiodic signals in a noisy climatic record.

Nine volcanic events since 1883 were selected for this study. The main selection criterion was whether an eruption created a significant stratospheric dust veil. Our evaluation of stratospheric dust veils made use of a variety of data sources including the explosive magnitude of the eruption (the Volcanic Explosivity Index—VEI—of Newhall and Self 1982), sulfate and trace-metal concentrations in Arctic and Antarctic ice samples (e.g., Hammer et al. 1980; Delmas and Boutron 1980; Boutron 1980), actinometric measurements (e.g., Dyer 1974), lunar brightness during eclipses (Keen 1983), and quasi-objective indices of volcanic dust veils (e.g., the Dust Veil Index—DVI—of Lamb 1970). Table 2 lists the volcanic events used in this study as well as some of the evaluation parameters mentioned above. Appendix A reviews the selection criteria and appendix B gives further details about each of these eruptions. Although our weighting of a variety of volcanic parameters is unavoidably subjective, we feel that the most significant stratospheric aerosol events since 1883 (i.e., Krakatau, Tarawera, Pelee/Soufriere/Santa Maria, Agung, El Chichón) are contained within our list and that the lesser events (i.e., Ksudach, Katmai, Awu and Fuego) established at least minimal stratospheric dust veils.

Data for this analysis came from the World Monthly Surface Station Climatology (WMSSC) and the Comprehensive Ocean-Atmosphere Data Set (COADS), both available at the National Center for Atmospheric Research (NCAR). WMSSC is a global collection of historical surface data reports from over 4400 sites that were taken from the Smithsonian Institution's *World Weather Records*, the National Climatic Data Center's (NCDC) *Monthly Climatic Data for the World* and from a variety of additional data sources. COADS, a collection of global oceanic surface data gathered from ships-of-opportunity, ocean weather ships, buoys, and bathythermographs, summarizes marine observations in 2° lat by 2° long boxes. Monthly mean surface air temperature, station pressure and precipitation were obtained for the years 1873 through 1983 from WMSSC and for January 1984 through January 1985 from *Monthly Climatic Data for the World*. Monthly mean sea surface temperature for the years 1873 through 1979 was taken from COADS.

In order to composite the above monthly datasets it is necessary to define a key month for each volcanic eruption. If an eruption occurred on or before the seventeenth of a calendar month, that month was designated the key month; if the eruption occurred after that day the following month was selected.[1] As noted in a variety of studies (e.g., Lamb 1970), little evidence of stratospheric aerosol is observed after 3 to 5 yr following even the largest eruptions. Only following the 1963 Agung eruption did stratospheric aerosol concentrations remain high enough to produce a measurable attenuation of the direct solar beam for a longer period (nearly 7 yr)—a phenomenon attributed by Pueschel et al. (1972) to a "recharging" of stratospheric aerosols by smaller eruptions in the years following Agung. With such studies in mind, we selected a 21-yr compositing period of 10 yr before until 11 yr after the key month.

The use of a homogeneous dataset for all of the selected eruptions would have limited this study to the few stations reporting during the earliest eruption (i.e.,

[1] The key month is defined as the first calendar month following the eruption during which the volcanic cloud completed one entire circuit around the globe. Assuming a 2-week period between the cloud's initial injection and its first global circuit leaves the seventeenth as the dividing day.

Krakatau). Therefore, to take advantage of improving data availability with time, we allowed the number of stations to vary between eruption periods. All land stations reporting less than 90% (228 months) of a given 21 yr (253 months) eruption period were omitted. To prevent unrepresentative weighting of data-rich regions, one station was omitted whenever two stations were within 330 km of each other. COADS boxes with less than two ship observations or a mean day-of-the-month of the contributing ship reports outside the middle third of the month (days 10 to 20) were omitted. For the eruptions prior to 1950 a COADS box was omitted if it did not encompass at least 65% (164 months) of a 21 yr eruption period. After 1950 this criterion was increased to 80%. Finally, to insure data quality excessively large monthly anomalies were omitted.[2]

Figure 1 presents the number of fixed surface stations reporting temperature, station pressure and precipitation in each hemisphere, and the number of $2° \times 2°$ boxes reporting sea surface temperature (after the above quality control procedures) for the eruption periods used in this study. For all parameters and eruption periods the majority of reports came from the Northern Hemisphere. The amount of data increases dramatically after the late 1800s, with the best data availability during the 1960s (except for pressure which had the best coverage during the recent El Chichón eruption).

For each 21 yr eruption period, the monthly data at every selected WMSSC station or $2° \times 2°$ COADS box was expressed as an anomaly from the 10-yr mean of the corresponding months prior to the eruption. The use of such a 10-yr average has many advantages: it is long enough to remove much of the interannual variability, but short enough to reflect long-term climate trends. Monthly mean anomalies were then calculated for 12 zonal bands of 15° lat. Zonal bands with less than three reports were omitted from further analysis. Hemispheric mean anomalies were then computed weighting each latitude band by its fractional area. The eruption year (yr 0) was defined as the first 12 month period, beginning with the eruption key month, following an eruption, with all other years defined relative to it. Superposed epoch analyses were constructed on monthly, annual and seasonal bases by averaging the monthly anomalies for several eruptions. In addition, eruption hemisphere composites were created.

As noted in section 2b, sea surface temperature variations in the tropical Pacific are reflected in surface and tropospheric air temperature over a large portion of the globe. Much of this SST variability appears to be connected with quasi-periodic ENSO cycles. Such ENSO-related temperature change might potentially mask the temperature effects of a volcanic eruption (Angell and Korshover 1984, 1985), or could leave the spurious impression of a volcano influence on climate. Therefore, in some of the figures presented below we have attempted to remove the ENSO signal from the surface temperature time series in order to clarify the volcanic effect.

Three indices of monthly mean tropical Pacific SST were constructed using the COADS 2° by 2° data boxes and are presented in Fig. 2. The first (SST1) is a time series of SST data for the eastern Pacific (6°N–6°S, 80°W–92°W) from 1872–1944 (Fig. 2a). The second

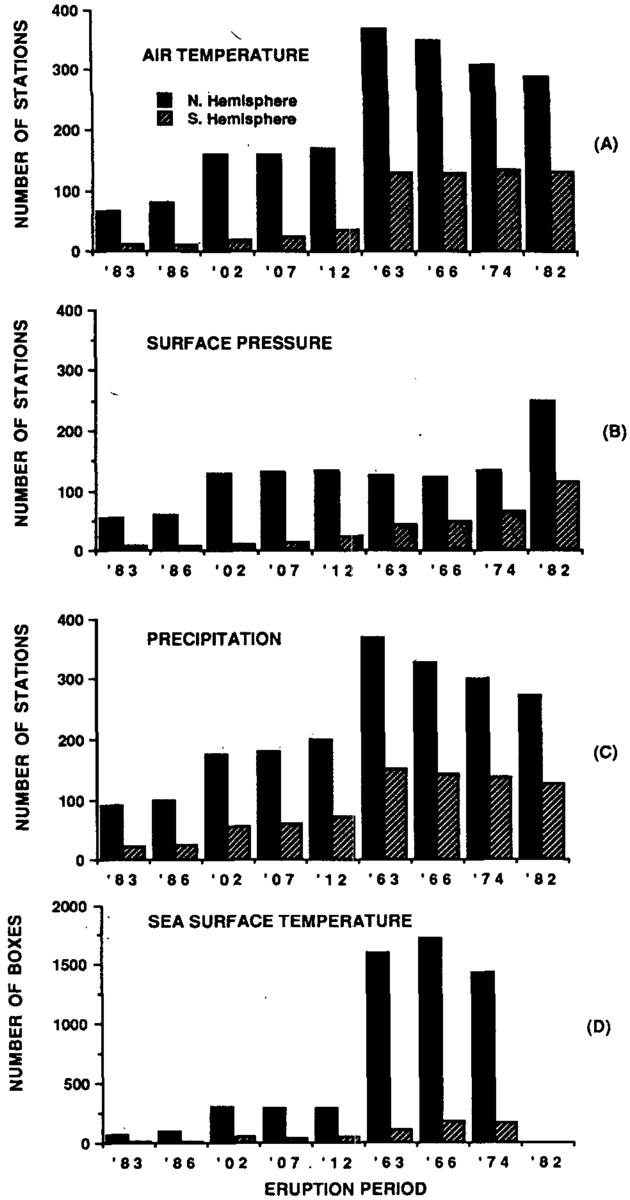

FIG. 1. Number of stationary surface stations and $2° \times 2°$ ocean boxes reporting in the Northern and Southern hemispheres for (a) air temperature, (b) surface pressure, (c) precipitation, and (d) sea surface temperature for the eruption periods considered in this study.

[2] Rejection criteria are as follows: monthly mean surface temperature anomalies exceeding 17.5°C, monthly mean pressure anomalies exceeding 30 mb, monthly sea surface temperature anomalies greater than 4.8°C, and monthly total precipitation values exceeding 2 m.

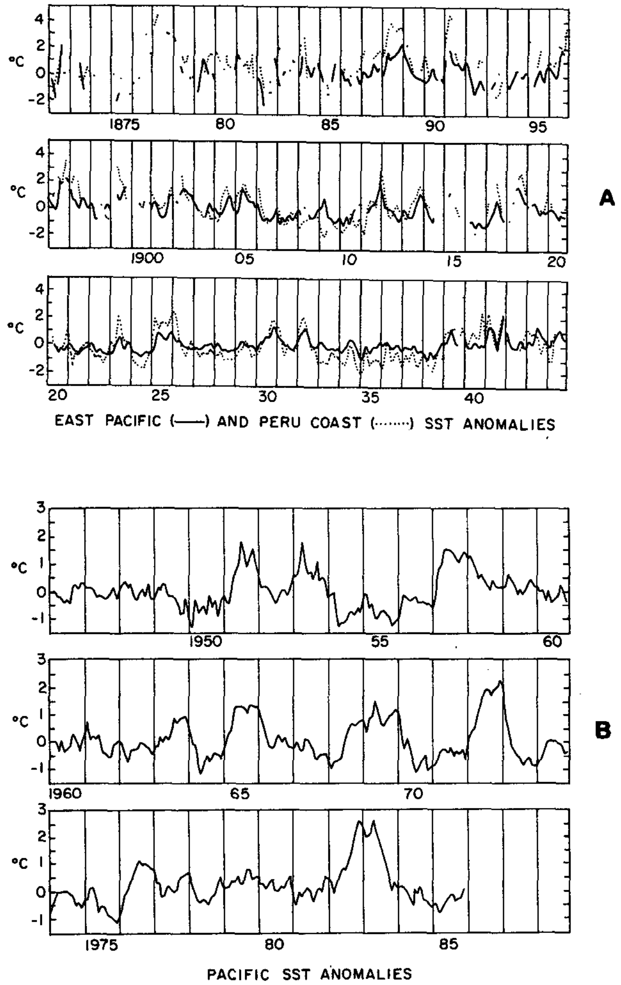

FIG. 2. Tropical sea surface temperature indices. (a) presents the SST anomalies for the East Pacific (6°N–6°S, 80°W–92°W) and the Peru Coast (4°S–10°S, 82°W–coast) region for 1872 through 1944. (b) presents Pacific SST anomalies for 1946 through 1985 for 6°N–6°S, 178°–80°W.

(SST2) is a SST series for the Peru Coast (4°S–10°S, 82°W–coast) for the same period (Fig. 2a). The third SST series (SST3), shown in Fig. 2b, is for a more extensive region (6°N–6°S, 80°W–178°W) from 1946–1985.

There is little in the published literature on the quantitative relationship between tropical SST and surface air temperature other than the observation that tropical tropospheric temperatures lag behind SST by ~6 months (e.g., Angell 1981). To help provide additional quantitative information, a calculation of the linear regression coefficient between tropical Pacific SST and surface air temperature has been made by regressing the tropical SST indices SST1 and SST3 against the Northern and Southern hemisphere surface temperature series of Jones et al. (1986). The correlation and linear regression coefficients between SST3 and the Jones et al. hemispheric datasets are shown in Table 3 for a variety of lags. For the Northern Hemisphere the maximum correlation coefficient (0.33) occurs when tropical Pacific SST leads surface temperature by 8 months; the corresponding linear regression coefficient is 0.13. The correlation of Southern Hemisphere surface air temperature with tropical SST is largest when the latter leads the former by 4 months (the corresponding linear regression coefficient is 0.15). It is important to note that for both hemispheres the variation of the correlation and linear regression coefficients is not great for adjacent lags. The same procedure was applied for the longer SST1 series, with very similar results. Using the optimal lag (L) and regression coefficient (R) between tropical SST and hemispheric temperature, an "ENSO-corrected" hemispheric temperature (T_c) was calculated for each eruption event:

$$T_c(t) = T_s(t) - T_{sst}(t - L) \times R, \qquad (1)$$

where t is time, T_s is the uncorrected hemisphere surface air temperature, and T_{sst} is either SST1 or SST3. The second term on the right side of (1) will be termed the "ENSO-correction" in further discussions.

TABLE 3. Correlation and linear regression coefficients between hemispheric surface temperature and tropical pacific SST.

Lag[1] (months)	Correlation coefficient	Regression coefficient
Northern Hemisphere temperature		
−12	.24	.10
−11	.27	.11
−10	.29	.12
−9	.32	.13
−8	.33	.13
−7	.31	.13
−6	.32	.13
−5	.31	.13
−4	.29	.12
−3	.31	.13
−2	.28	.11
−1	.22	.09
0	.19	.08
Southern Hemisphere temperature		
−12	.20	.06
−11	.27	.08
−10	.32	.09
−9	.39	.11
−8	.46	.13
−7	.51	.15
−6	.52	.15
−5	.53	.15
−4	.54	.15
−3	.52	.15
−2	.50	.14
−1	.46	.13
0	.41	.12

[1] A negative lag indicates that air temperature lags tropical Pacific SST.

Several investigators (e.g., Angell and Korshover 1985; Lough and Fritts 1987) have suggested that volcanic eruptions might produce longitudinal variations in climate. Except for some sea surface temperature composites, this paper does not examine climatic variability along latitude circles. An examination of such regional variations is made difficult by the inhomogeneous distribution of observing sites, with large data voids over the oceans and even over many land areas. The fact that volcanic aerosols spread quite uniformly in the longitudinal direction suggests that a careful examination of possible hemispheric and zonal responses is a valuable first step. Furthermore, it is hard to imagine a scenario in which a large volcanic eruption has no influence on hemispheric and zonal climate while at the same time producing significant regional anomalies.

The reader will find no mention of statistical significance tests in this article. Instead we have attempted to use physical reasoning to determine whether an "apparent" volcanic signal is the result of volcanic aerosol or is derived from some other source. Our skepticism about the use of statistical tests (e.g., the "Student t" and Monte Carlo methods) is derived from the examination of several papers that purport to find statistically significant volcanic signals in air temperature records; applying physical insight and additional information often reveals other sources of the "volcano-induced" variability.

b. *Annual and monthly composites of air and sea surface temperatures for the eruption hemispheres*

All nine events: Figure 3 presents the annual eruption hemisphere composites for all nine events. The temperature composite (Fig. 3a) indicates a modest drop in the eruption year, a recovery by yr 3, and a relatively small (~0.2°C) temperature range. The size of the negative anomalies in the two posteruption years are not unique since similarly sized anomalies exist in other years. The SST composite (Fig. 3b) also indicates modest cooling during the first 2 yr after the eruption key date; again, the posteruption negative anomalies are no larger than others observed during the composite period. The difference between hemispheric temperature and SST (Fig. 3c) is rather featureless, while the "ENSO-corrected" temperature (Fig. 3d) is slightly more suggestive of a volcanic signal (i.e., of an isolated posteruption cooling) than the uncorrected temperature series.

The monthly eruption hemisphere composites for all nine events are displayed in Fig. 4. The temperature composite (Fig. 4a) indicates a ~0.25°C drop in temperature during the first few months after the eruptions, followed by warming during the subsequent year. Temperature declines again between months 10 and 18, and then slowly returns to normal during the following year. In contrast, SST (Fig. 4b) remains nearly constant for nearly 10 months after the eruption date, then falls until month 20, and recovers during the following year. The difference in phasing of temperature and SST creates a substantial negative excursion in T-SST during the eruption year (Fig. 4c). The composite ENSO-related temperature correction (Fig. 4d) is rather small; therefore, the corrected temperature (Fig. 4e) is quite similar to the original temperature series.

Five large events: In an attempt to strengthen the volcanic signal, the five largest eruptions (based on the

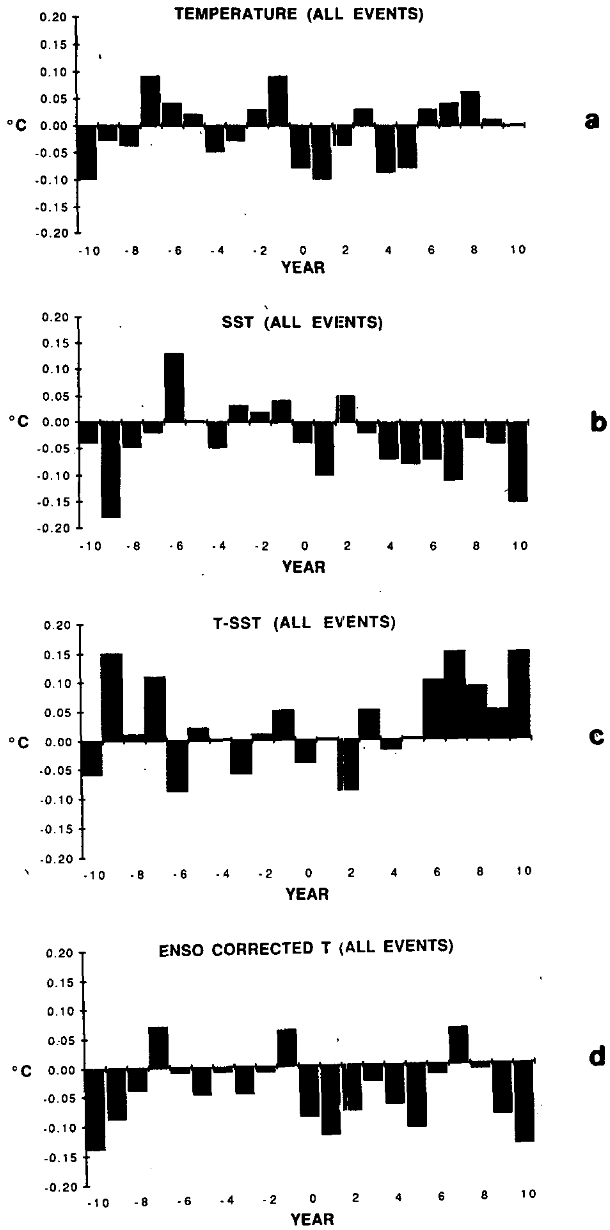

FIG. 3. Annual eruption hemisphere composites for all nine eruptions for air temperature (a), sea surface temperature (b), air temperature minus sea surface temperature (c), and ENSO-corrected temperature (d).

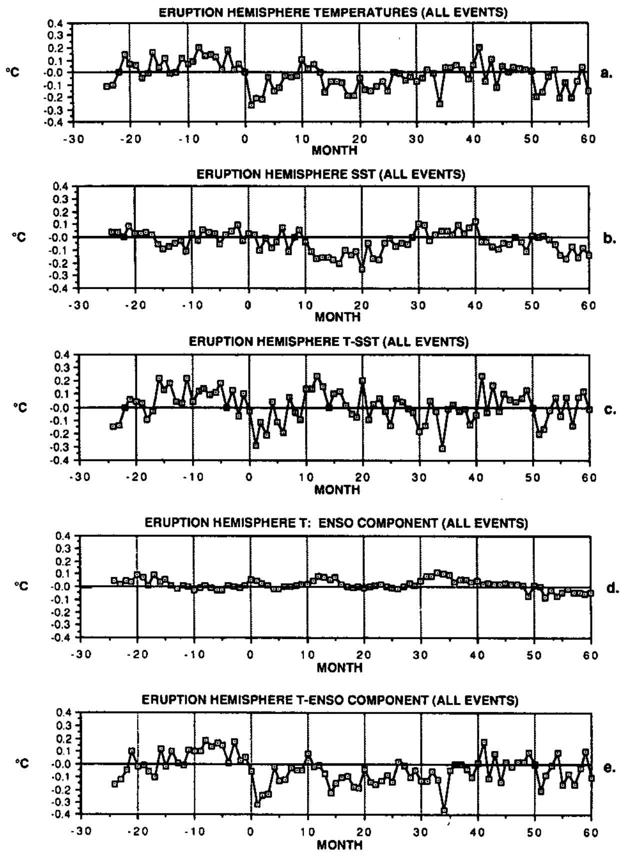

FIG. 4. Monthly eruption hemisphere composites for all nine eruptions for air temperature (a), sea surface temperature (b), temperature minus sea surface temperature (c), ENSO temperature correction (d), ENSO-corrected temperature (e).

information presented in appendix B regarding stratospheric injection)—Krakatau, Tarawera, Pelee/Soufriere/Santa Maria, Agung and El Chichón—were composited together. Figure 5 presents the annual eruption hemisphere temperatures for these events. The temperature composite (Fig. 5a) shows preeruption warming and a substantial posteruption cooling (over 0.3°C between yr −1 and +1), with the negative anomaly in yr 1 being the largest of any year of the composite. The SST composite (Fig. 5b) also suggests a warming trend during the 10 yr prior to yr 0, followed by 2-yr posteruption cooling. The difference between temperature and SST for these events (Fig. 5c) shows alternating positive and negative anomalies prior to the eruption date, and negative anomalies during the three posteruption years. Finally, application of the ENSO-correction to the temperature time series (Fig. 5d) decreases the preeruption warming and enhances the posteruption cooling.

The monthly time series for these five large events are presented in Fig. 6. The temperature series (Fig. 6a) shows a sharp drop during the 2 months following the eruption key date, and generally below-normal temperatures until approximately month 40. In contrast, the sea surface temperatures (Fig. 6b) do not drop until month 10, and subsequently remain below normal until around month 26. The lag is SST response creates a 1 yr negative excursion in the difference between temperature and SST (Fig. 6c). The ENSO-corrected temperature series (Fig. 6e) is similar to the un-

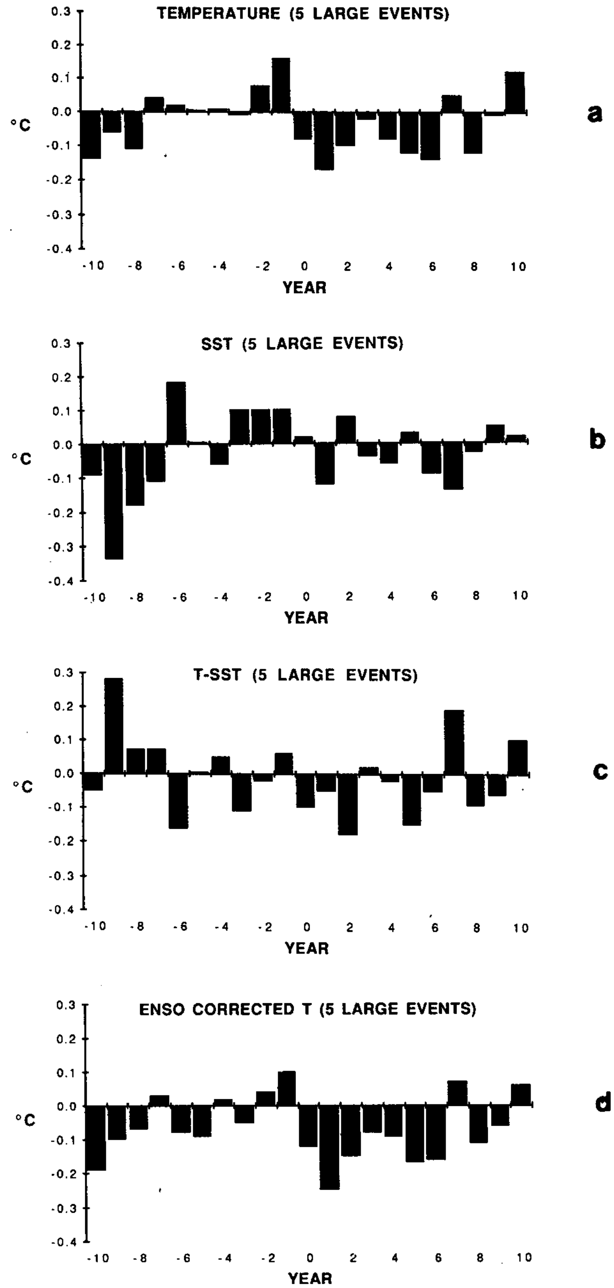

FIG. 5. Annual eruption hemisphere composites for five large eruptions for air temperature (a), sea surface temperature (b), air temperature minus sea surface temperature (c), and ENSO-corrected temperature (d).

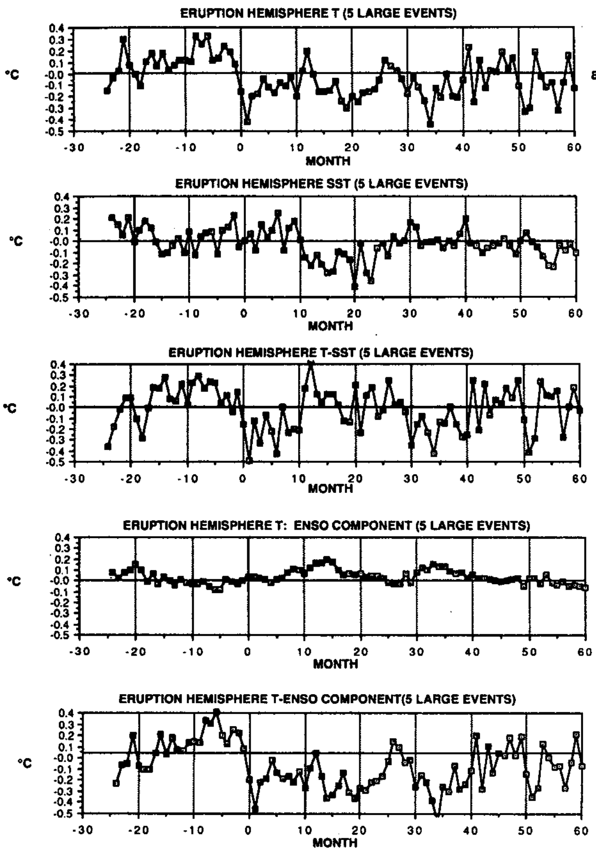

FIG. 6. Monthly eruption hemisphere composites for five large eruptions for air temperature (a), sea surface temperature (b), temperature minus sea surface temperature (c), ENSO temperature correction (d), ENSO-corrected temperature (e).

corrected temperatures, except for enhanced cooling during the first 3 posteruption yr.

Three large twentieth-century events: Since air and sea surface temperature data during the 1880s is relatively sparse but increases rapidly in availability by 1900, we created a composite of the three most significant stratospheric events during the present century: Pelee/Soufriere/Santa Maria, Agung, and El Chichón. The resulting annual composite (Fig. 7a) indicates a general warming trend during the 10 yr prior to the eruption key date, followed by substantial cooling (0.33°C) during the 3 yr after the composite event. Temperatures make an unimpressive recovery during subsequent years. The SST composite for two of the events (Pelee/Soufriere/Santa Maria and Agung[3]), presented in Fig. 7b, indicates virtually no cooling during the eruption year, followed by a substantial temperature fall during yr 1. The difference between temperature and SST (Fig. 7c) exhibits a large drop during the eruption year. The ENSO-corrected temperature series (Fig. 7d) shows substantially less preeruption warming and larger posteruption cooling than the uncorrected temperature series. Furthermore, the corrected temperature series evinces a smooth, monotonic temperature recovery during yr 1 through 5. Clearly, this temperature series is suggestive of a volcanic signal.

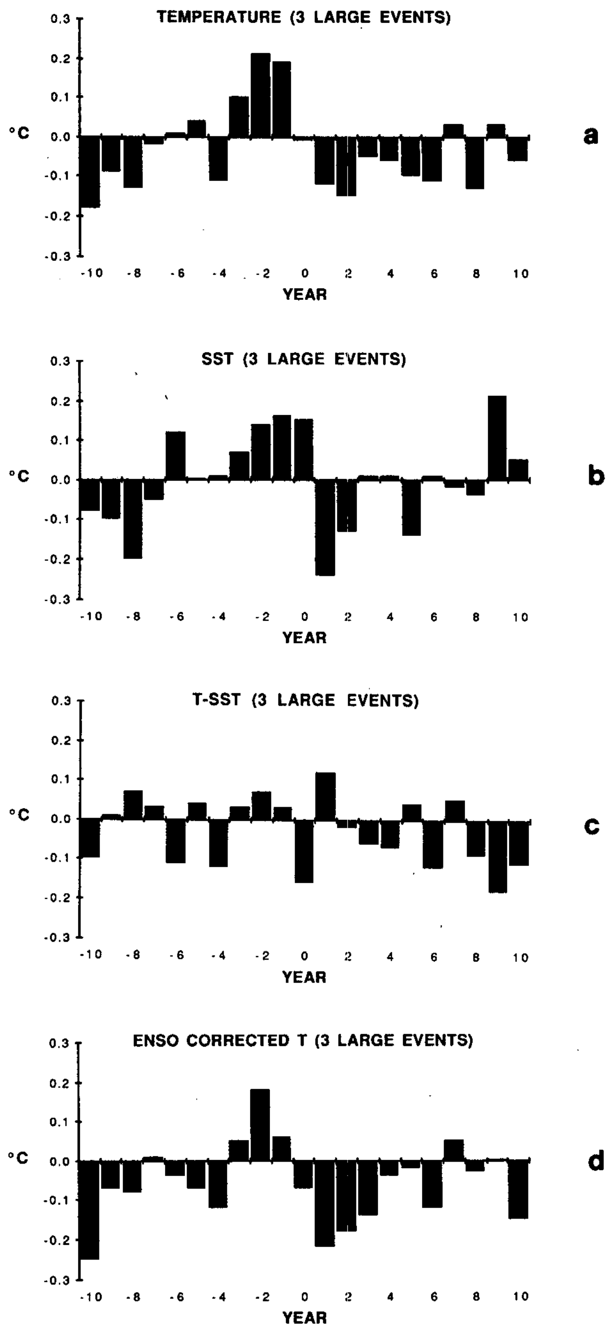

FIG. 7. Annual eruption hemisphere composites for three large eruptions for air temperature (a), sea surface temperature (b), air temperature minus sea surface temperature (c), and ENSO-corrected temperature (d).

[3] The El Chichón SST values were not available at the time the calculations were made.

The monthly eruption hemisphere temperature series for the three large events is shown in Fig. 8. A drop in temperature is apparent around the eruption key date, with high amplitude oscillations during the subsequent year (Fig. 8a). Predominantly negative anomalies are observed during the following 2 yr, with an episode of particularly cool values around month 35. Sea surface temperatures (Fig. 8b) are fairly steady and above-normal prior to month 10, followed by a steady decrease to negative anomalies of ~0.3°C during the next 6 months. SST recovers during the third and fourth year following the eruption key date. The difference between temperature and SST reaches ~0.55°C during the first year after the eruption key date (Fig. 8c). The ENSO-corrections (Fig. 8d) are fairly large for this composite (reaching ~0.2°C) and result in a corrected temperature composite (Fig. 8e) with larger posteruption negative anomalies than displayed in Fig. 8a.

Four weaker events: Annual eruption hemisphere temperature series for the four weaker eruptions—Ksudach, Katmai, Awu and Fuego—are presented in Fig. 9. The uncorrected temperatures for this composite

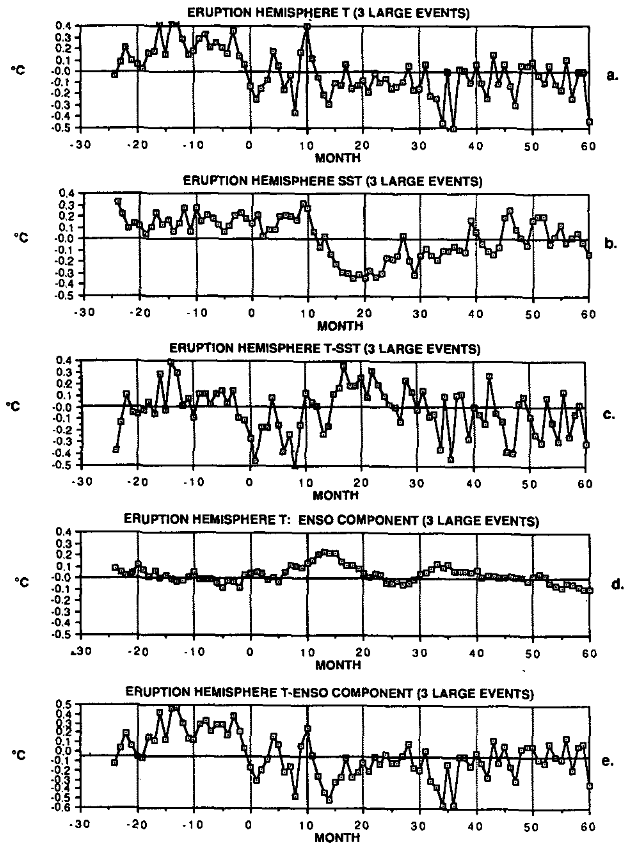

FIG. 8. Monthly eruption hemisphere composites for three large eruptions for air temperature (a), sea surface temperature (b), temperature minus sea surface temperature (c), ENSO temperature correction (d), ENSO-corrected temperature (e).

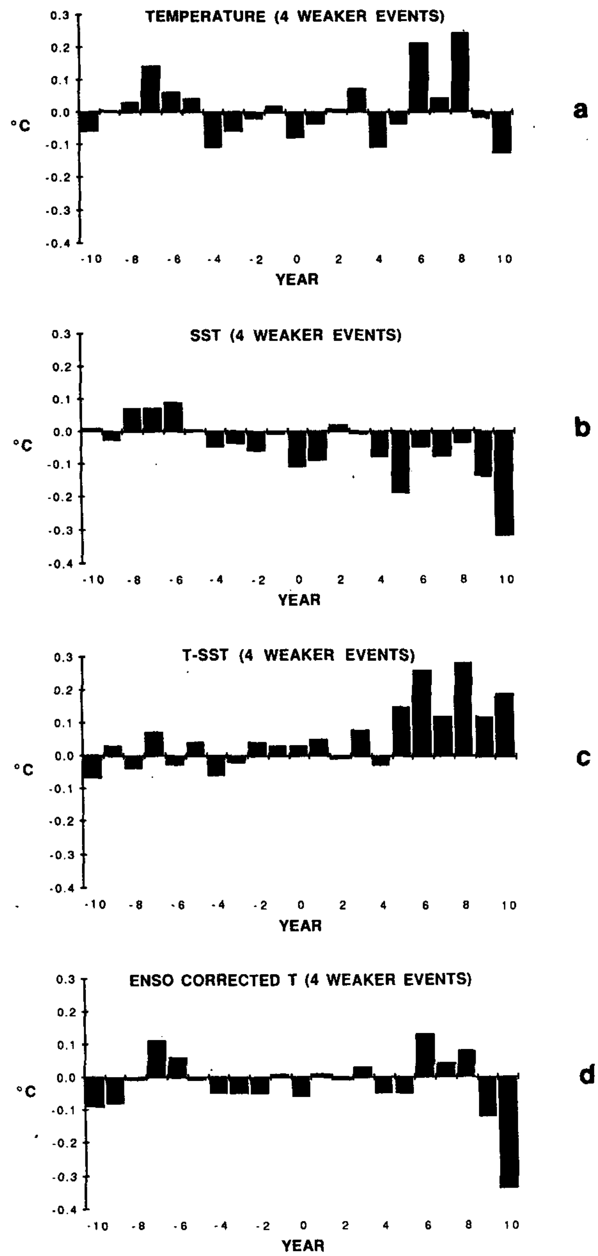

FIG. 9. Annual eruption hemisphere composites for four smaller eruptions for air temperature (a), sea surface temperature (b), air temperature minus sea surface temperature (c), and ENSO-corrected temperature (d).

(Fig. 9a) are far less suggestive of a volcanic signal than those of the previous figures: temperature decreases by only ~0.1°C in the eruption year and other years possess both larger temperature falls and greater negative anomalies. SST (Fig. 9b) decreases in the 2 yr following the eruption key date, but is not exceptional with respect to the amplitude of the negative anomaly or the magnitude of the temperature decrease. Both the differences between temperature and SST (Fig. 9c) and the corrected temperature series (Fig. 9d) give little suggestion of a volcanic signal.

The monthly eruption hemisphere composites for the four weaker eruptions are shown in Fig. 10. Uncorrected temperature (Fig. 10a) shows only minor excursions during the two years before and after the eruption key date, with small negative anomalies during the first 6 months and at approximately 26 months after the events. SST (Fig. 10b) indicates a cooling trend during the first 5 months, with steady warming during the subsequent 3 yr. The difference between air and sea surface temperature (Fig. 10c) is rather featureless, except for minor dips at approximately 25 and 53 months. With only a small ENSO-correction (Fig. 10d), the corrected temperatures (Fig. 10e) are very similar to those shown in Fig. 10a.

c. *Volcanic effects on surface pressure*

Volcanic aerosols might affect surface pressure directly by changing the temperature of the column of air above an observing site or indirectly by altering the atmospheric circulation. With regard to the direct hydrostatic effects, stratospheric volcanic aerosols can in-

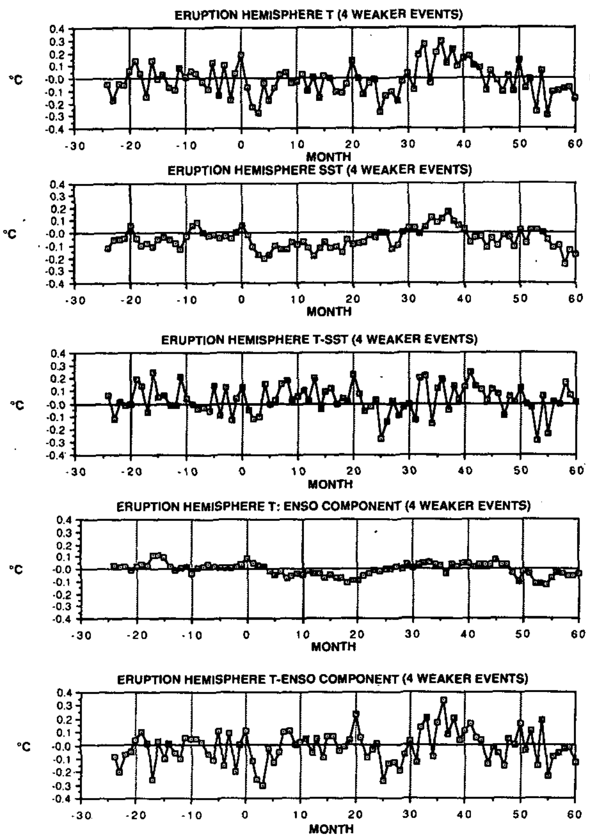

FIG. 10. Monthly eruption hemisphere composites for four smaller eruptions for air temperature (a), sea surface temperature (b), temperature minus sea surface temperature (c), ENSO temperature correction (d), ENSO-corrected temperature (e).

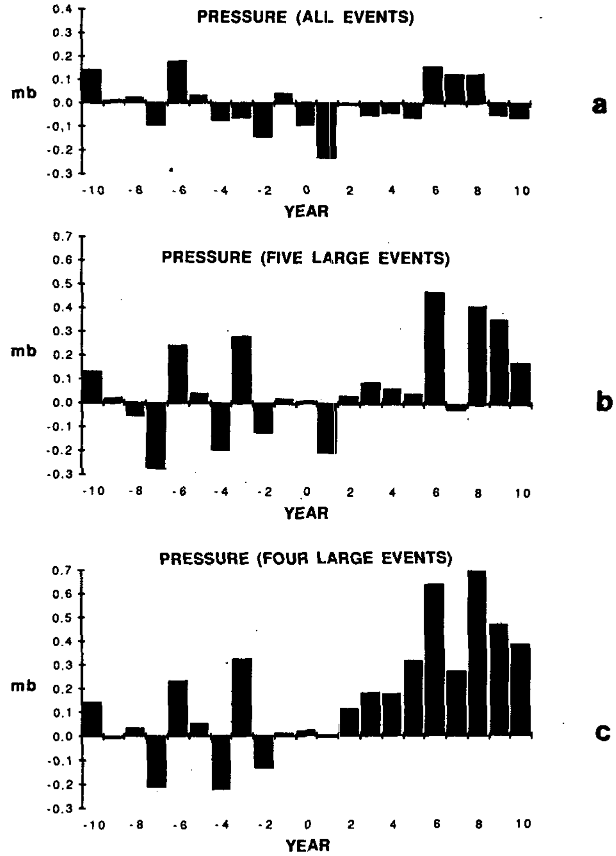

FIG. 11. Eruption hemisphere pressure composites for all nine events (a), five large events (b), and four large events (c).

crease albedo and thus induce tropospheric cooling, which would contribute to higher pressure; on the other hand, aerosol-induced stratospheric warming (e.g., Newell 1970) might result in pressure falls. The sign or magnitude of the surface pressure variations would depend on the net effect of these two mechanisms which can vary according to a variety of factors including the height, optical properties and latitudinal distribution of the volcanic dust cloud.

Figure 11 presents annual eruption hemisphere composites of surface pressure. The composite of all nine events (Fig. 11a) shows a small decrease in pressure following the eruption key date, with the largest negative anomaly (−0.24 mb) observed during yr 1. Compositing the five largest events (i.e., Krakatau, Tarawera, Pelee/Soufriere/Santa Maria, Agung, El Chichón) produces a noisy pattern with an inconsequential drop in the eruption year and a 0.24 mb decrease in yr 1. However, even the latter pressure drop is suspect since it is derived from a very large (1.02 mb) decrease in only one eruption (Agung). A composite of the remaining four strong events (i.e., without Agung), shown in Fig. 11c, indicates little pressure change in the 2 yr following the eruption key date. Clearly these composites do not suggest an appreciable volcanic signal in (eruption hemisphere) surface pressure.

d. Volcanic effects on precipitation

Several investigators have suggested that volcanic aerosols might have an influence on precipitation. Some have theorized that as volcanic aerosols settle out of the stratosphere they might act as cloud condensation nuclei (e.g., Humphreys 1940; Wexler 1951, 1956). Such volcanic cloud seeding could either increase precipitation in regions deficient in such nuclei or reduce precipitation by overseeding. Other papers have suggested that the radiative effects of volcanic aerosols might alter atmospheric circulation and thus indirectly modify the amount and distribution of precipitation (e.g., McCracken and Luther 1984; Handler 1984, 1986). Using a zonally averaged, nine-layer model with parameterized eddy transport, McCracken and Luther (1984) simulated the climatic effects of the El Chichón eruption of 1982. They found that the simulated precipitation rate following this event was nearly unchanged in the Southern Hemisphere and that most of the Northern Hemisphere experienced only minor (<5%) changes. The only major impact was in the vicinity of the southward-displaced Intertropical Convergence Zone (ITCZ); north of the zone precipitation dropped by as much as 6%, while to its south enhancements of up to ~10% were found.

Figure 12a presents the annual precipitation composite for the eruption hemispheres of three large events (Pelee/Soufriere/Santa Maria, Agung, El Chichón).[4] This figure does not suggest a volcanic signal in precipitation, and supports the results of McCracken and Luther (1984) of minimal hemispheric effects. To test McCracken and Luther's suggestion of latitudinal variations in precipitation accompanying a volcano-induced shift in the ITCZ, precipitation series (for the three-eruption composite) were prepared for the bands 0°–15°S, 0°–15°N and 15°–30°N (Figs. 21b–d). In the McCracken and Luther model, precipitation increases in the band from 0°–15°S during the first year; in Fig. 12d this band possesses a preeruption negative anomaly that decreases in amplitude during the eruption year and turns positive during the subsequent 2 yr period. In their model results precipitation decreases in the 15°–30°N band during the first year, while in Fig. 12b precipitation increases slightly in that zone. A large eruption-year precipitation decrease is observed from 0°–15°N (Fig. 12c), a zone in which McCracken and Luther found little change. Considering the simplified nature of the McCracken and Luther model and the limited number of eruptions considered in this study, our results can not be considered a conclusive evaluation of McCracken and Luther's hypothesis of

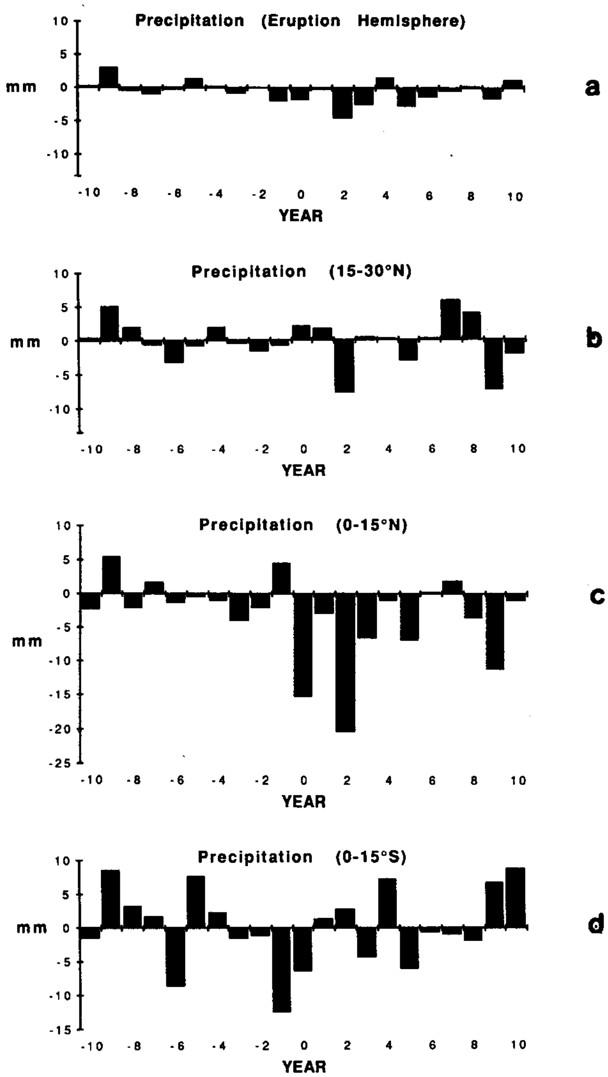

FIG. 12. Precipitation composites for three large events for the eruption hemisphere (a), 15°–30°N (b), 0°–15°N (c), 0°–15°S (d).

a volcano–precipitation connection. However, our results suggest that a significant volcanic signal in precipitation does not appear to exist.

4. Evaluation of the individual time series

In order to better appraise the reality of a possible volcanic signal, we shall now consider the individual eruptions in detail, keeping in mind the important influence of ENSO events on air and sea surface temperatures. Figures 13–21 present air and sea surface temperature time series for each of the nine individual eruptions.

Krakatau (6°S, 8/1883): In the year following this eruption, air temperature (Fig. 13) decreased by ~0.5°C in the only latitude zone (30°–45°S) in the eruption hemisphere with sufficient data. Although this

[4] Insufficient precipitation data was available from the remaining large eruptions (Krakatoa, Tarawera) for reliable results.

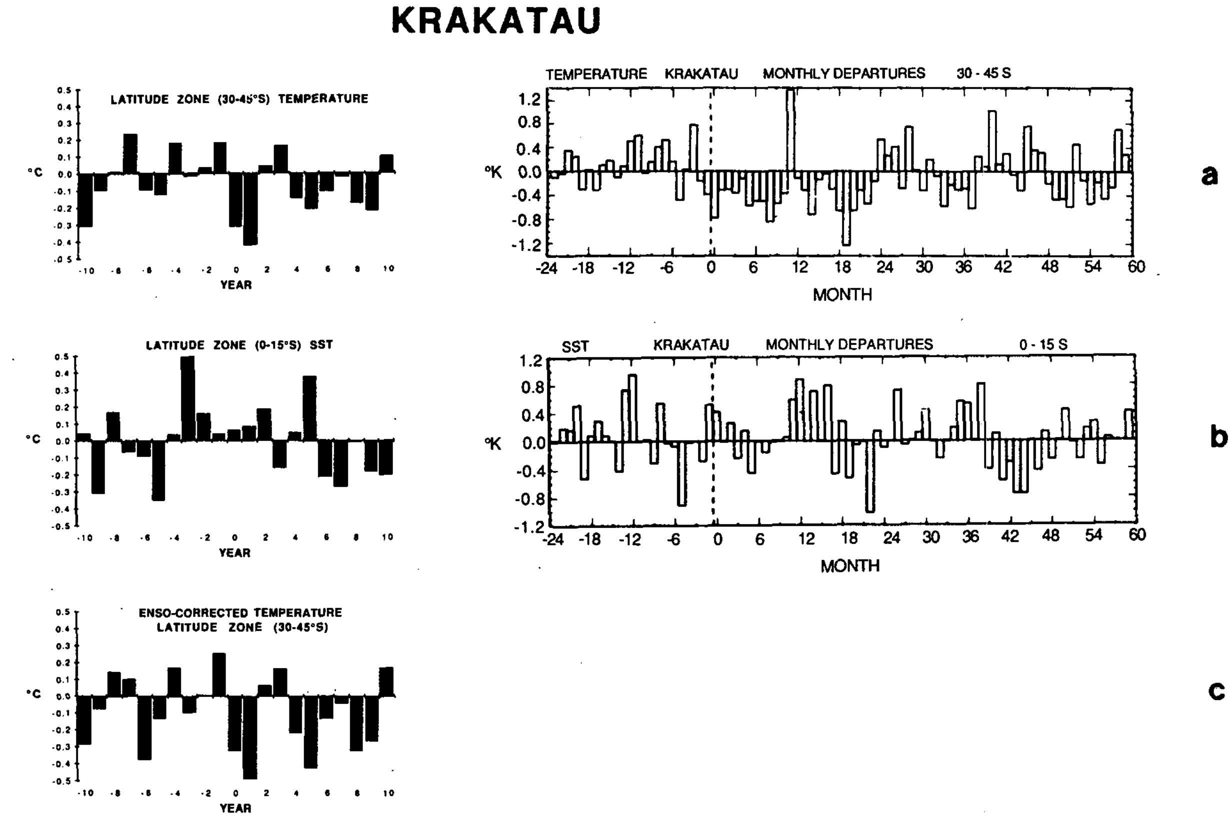

FIG. 13. Annual and monthly air and sea surface temperatures for Krakatau.

result is suggestive of a volcanic effect, one should note that monthly air temperatures in this zone began to decrease prior to the eruption key month. For sea surface temperature (Fig. 13b) the only band with enough data to be included (0°–15°S) experienced warming during the 2 posteruption yr. Figure 2 and the chronology of Quinn et al. (1978) indicate that 1884 was an ENSO year with SST warming. The fact that air temperatures still cooled in the eruption hemisphere even though an ENSO event was occurring suggests a sizable volcanic cooling. As shown in Fig. 13c, applying the ENSO-correction enhances the posteruption cooling.

Tarawera (38°S, 6/1886): The air temperature in the band 30°–45°S (again, the only eruption hemisphere band with sufficient data) shows minor cooling during the first 2 yr after this eruption, with the resulting negative anomaly (~0.05°C) being maintained for 5 yr (Fig. 14a). As shown in Figs. 2 and 14b, tropical Pacific SST drops substantially in 1886 but climbs rapidly during the ENSO event of 1887. Thus, it is not the minor temperature decrease in yr 0 that is suggestive of a volcanic influence, but rather the fact that temperatures continue to decline in year +2 when SST was rapidly increasing. The ENSO-corrected temperature for the 30°–45°S latitude band (Fig. 14c) illustrates this hypothesis with enhanced posteruption cooling.

Pelee/Soufriere/Santa Maria (5–19/1902, 15°N): Air temperature fell sharply during yr 0 and +1 in the eruption hemisphere and during yr +1 in the eruption band (Fig. 15a, c). SST declined in the eruption hemisphere in yr 0 and 1, but rose in yr 0 in the eruption band (Fig. 15b, d). The fact that air temperature decreased in yr +1, and subsequently rose in yr 3, is not as suggestive of a volcanic signal as one might think, since the tropical-Pacific SST showed a nearly simultaneous variation (Fig. 2). What does suggest a volcanic effect is that Northern Hemisphere air temperature fell in yr 0, even though a moderate ENSO event was occurring. In fact, yr 0 (1902) is the *only* year shown in this figure in which an ENSO event is *not* associated with hemispheric warming. The eruption band (0°–15°N) temperature (Fig. 15c) shows only a minor air temperature decrease during the eruption year. Could it be that this tropical eruption band is so strongly influenced by the ENSO signal that air temperature hardly falls and SST rises slightly? Is tropical air temperature so strongly coupled to tropical SST that it cannot fall substantially until the ocean surface begins to cool? ENSO-correction does enhance the apparent posteruption cooling (Fig. 15e).

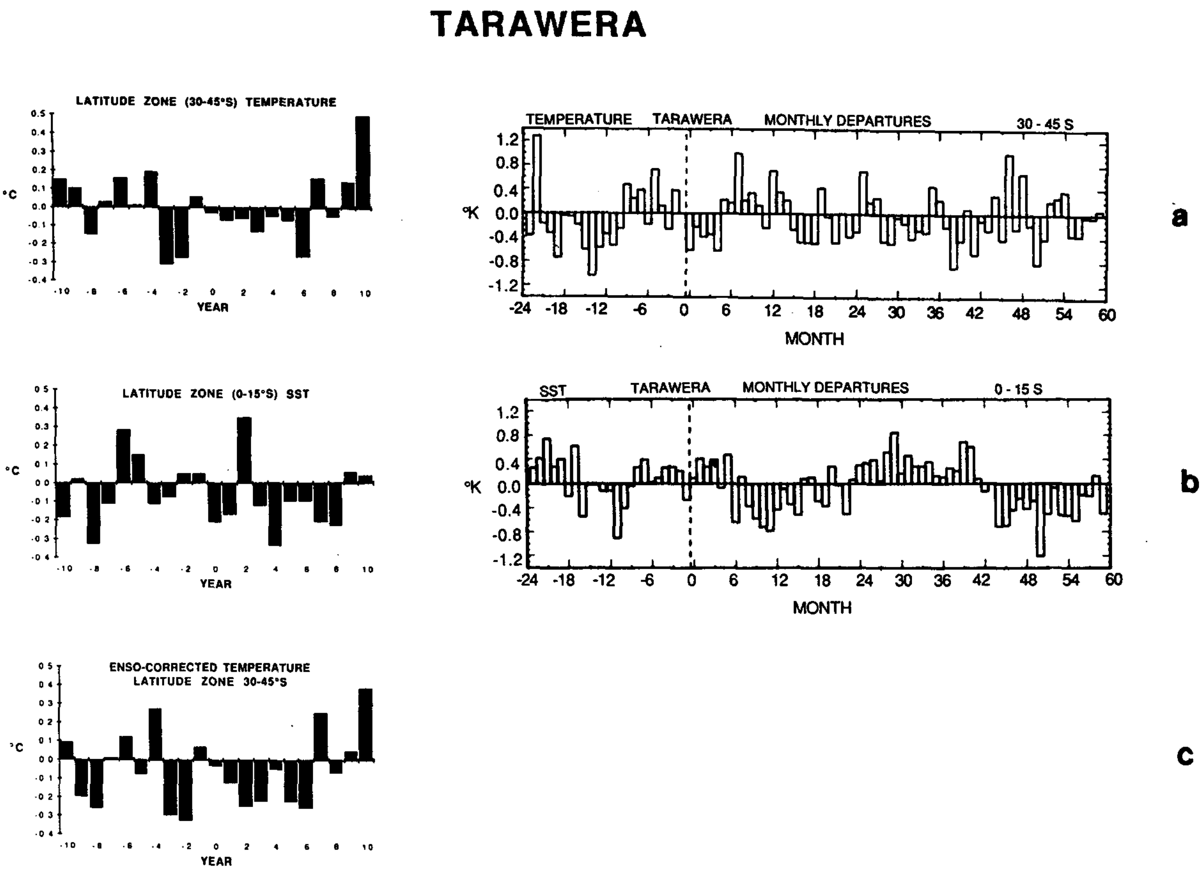

FIG. 14. As in Fig. 13, for Tarawera.

Ksudach (52°N, 3/1907): In both the eruption hemisphere and band Ksudach's annual air temperature fell during the eruption year and recovered over a period of ~3 yr (Fig. 16a, c). SST also dropped during the eruption year in the hemispheric and eruption band series (Fig. 16b, d). However, there are several reasons to suspect that a volcanic aerosol cloud was not the predominant cause of the posteruption temperature depressions. First, the eruption occurred in the high latitudes of the Northern Hemisphere and little aerosol would be expected to reach the Southern Hemisphere; yet, the air temperature and SST depressions in the Southern Hemisphere (calculated but not shown) were only slightly smaller than those in the Northern Hemisphere. Second, the Pacific SST dropped rapidly during the year prior to the eruption and reached a minimum approximately 6 months before the eruption—assuming a 6 month lag in air temperature response, this produces a evolution not unlike that observed in the monthly series shown in Fig. 16. ENSO-correction does not enhance any apparent posteruption cooling (Fig. 16e). In the eruption band air temperatures decreased prior to the eruption. For these reasons it appears that Ksudach may not have had a significant effect on surface temperature.

Katmai (58°N, 6/1912): In both the hemisphere and band of the eruption, air and sea surface temperatures fell (slightly) during yr 0, and then rose in yr 1 (Fig. 17a, c). However, there are reasons to question Katmai's role in this minor posteruption cooling. First, the evolution of air temperature and SST (Fig. 17c, d) in the eruption hemisphere and band are very similar to the tropical Pacific SST lagged by ~6 months. Thus, temperature declines during the first 6 months after the eruption were simply following the lagged-SST. Eruption zone and hemisphere cooling in air and sea surface temperature occur predominantly during the first 5 months after the eruption, a period too short for SST response to volcanic forcing. (The initial 5 month drop is even suggested in the Southern Hemisphere SST.) Finally, ENSO-correction (Fig. 17e) does not enhance the apparent posteruption cooling. For these reasons it appears unlikely that the yr 0 temperature drop after Katmai is caused by volcanic aerosol.

Agung (8°S, 3/1963): In the eruption hemisphere air temperature fell during the 2 yr following the event, and then slowly recovered over the next five years (Fig. 18a). In the eruption zone air temperature actually increased in yr 0, and then fell by ~0.3°C the following year. Southern Hemisphere and eruption band SST

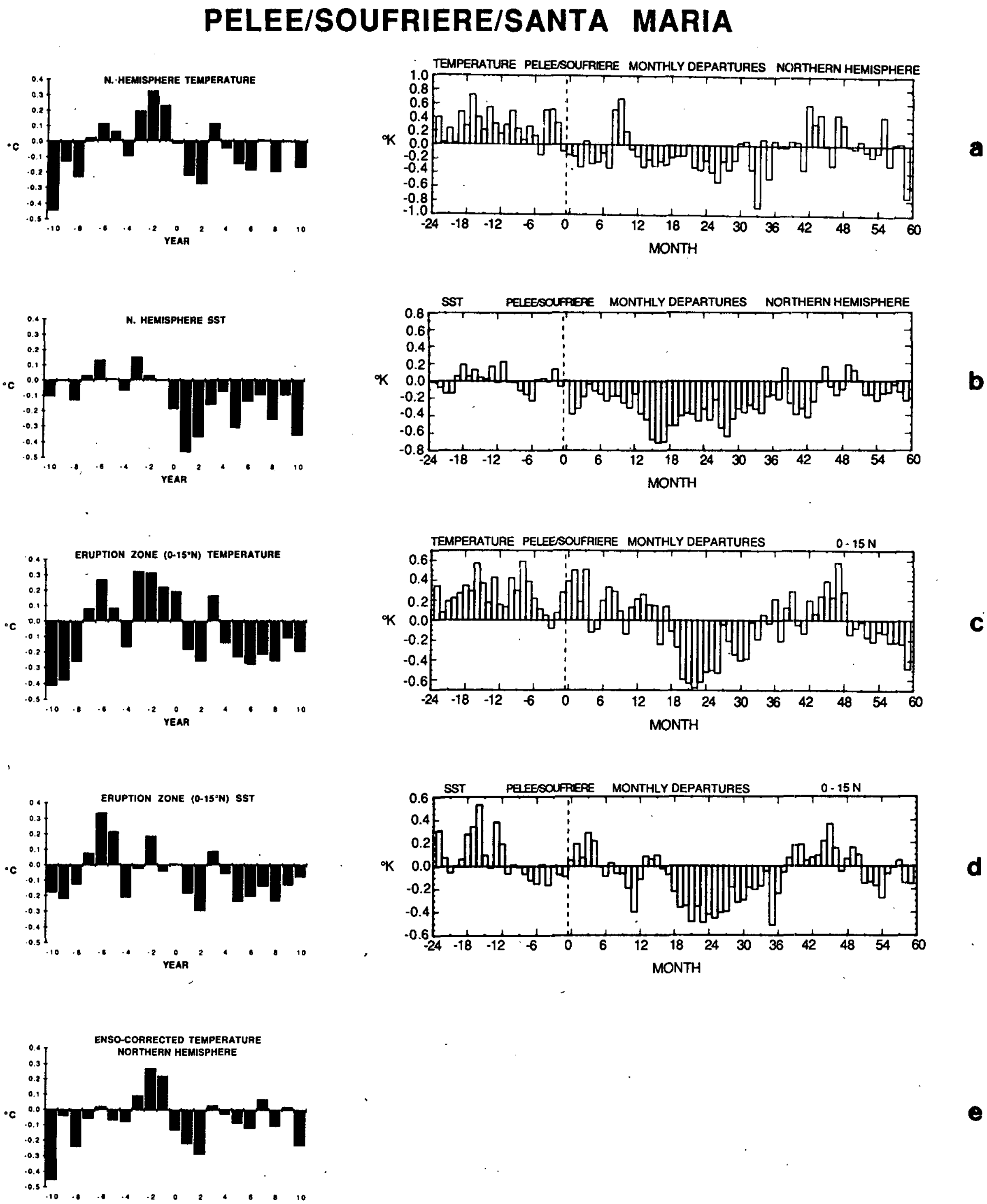

FIG. 15. As in Fig. 13, for Pelee/Soufriere/Santa Maria.

(Fig. 18b, d) anomalies are nearly identical, with rising SST from yr −2 through 0, followed by a large decrease in yr 1. Could it be that the relatively weak ENSO warming of 1963 was overwhelmed by the cooling effect of the volcano in the eruption hemisphere, while in the tropical eruption band the tropical Pacific SST increase (see Fig. 2) overpowered the volcanic cooling? In yr 1 the ENSO event entered its cool phase; thus, with ENSO-related cooling and volcanic effects superposed, a large negative anomaly was established in yr

KSUDACH

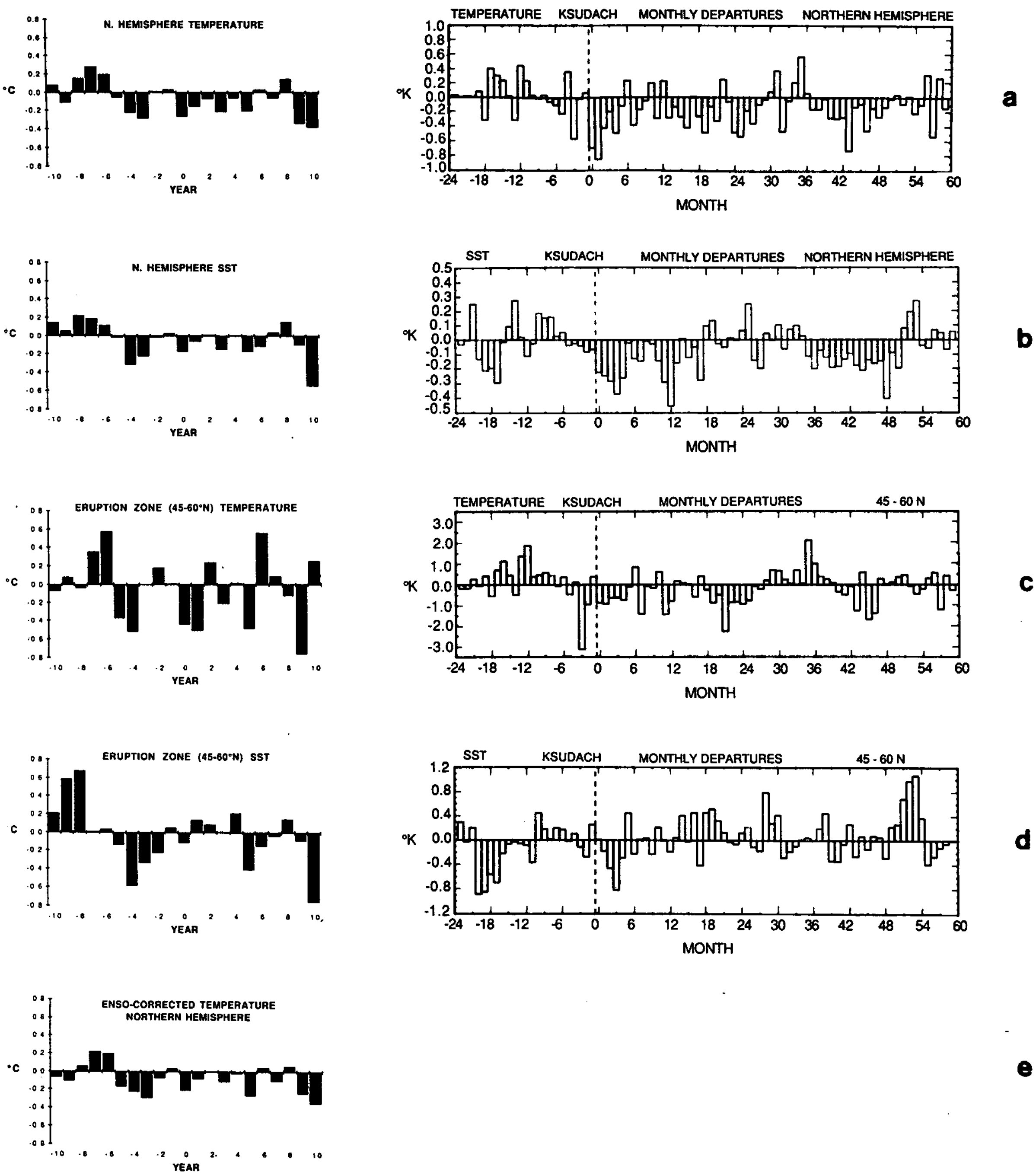

FIG. 16. As in Fig. 13, for Ksudach.

1. Increasing tropical SST in late 1964 and 1965 contributed towards returning the air temperature to normal.

Awu (4°N, 8/1966): The annual and monthly air and sea surface temperature anomalies in the eruption band indicate cooling during the first 2 yr after this event (Fig. 19c, d). However, there are several reasons to suspect that this temperature decrease has little to do with the volcano. First, in the eruption (Northern) hemisphere (Fig. 19a) there is no indication of a temperature fall during the first 2 yr, in fact, there is actually a slight increase. Second, the monthly series of air and

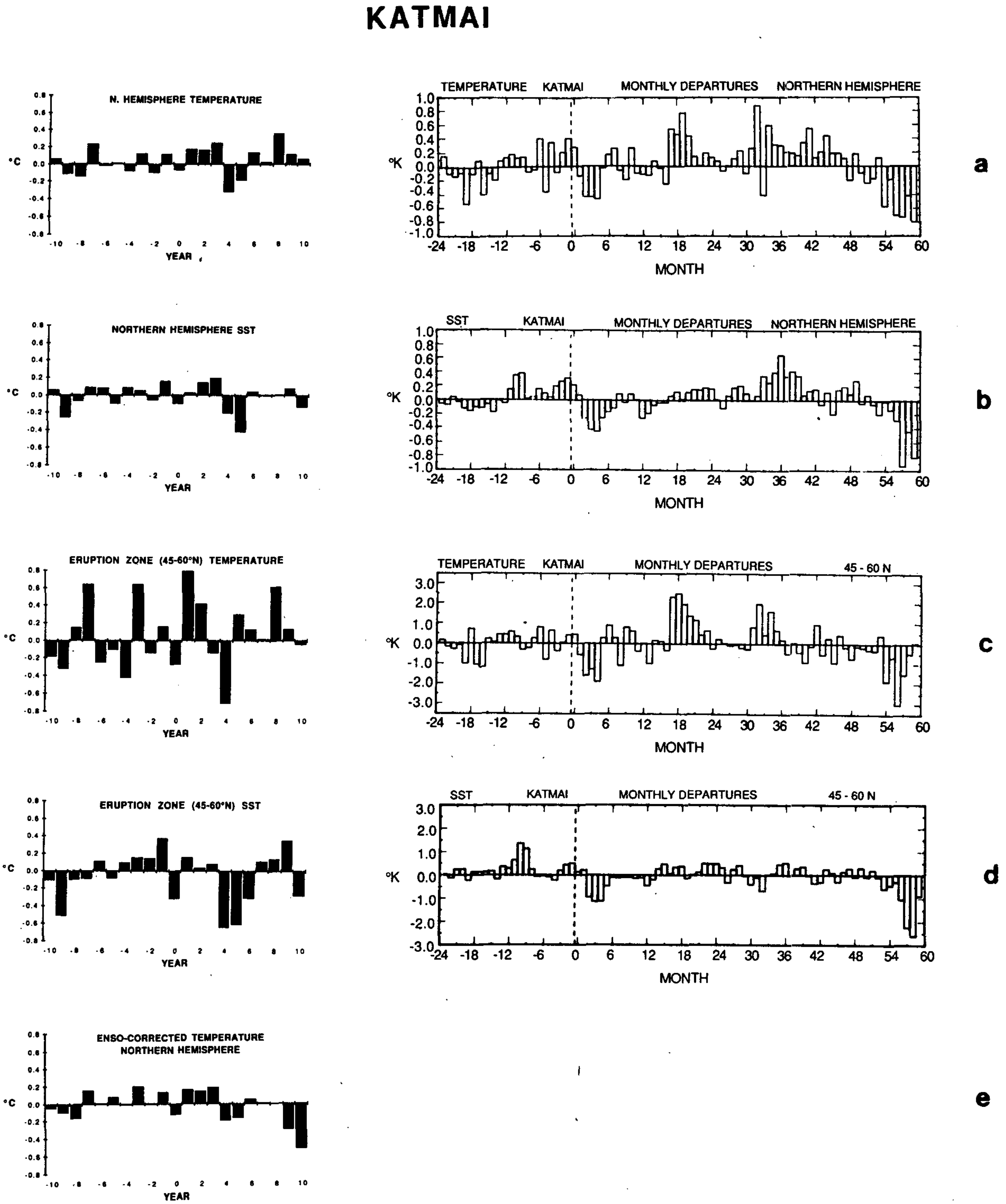

FIG. 17. As in Fig. 13, for Katmai.

sea surface temperature are nearly identical, with declining temperatures during approximately the first 18 months. These temperature variations appear to follow the tropical Pacific SST with a lag of ~5 to 6 months (Fig. 2b). Thus, the apparent volcanic signal after Awu is simply the result of a change in Pacific SST, and explains why the Southern Hemisphere (calculated but not shown) experiences a larger drop in temperature than either the eruption hemisphere or band.

Fuego (15°N, 10/1974): There is no evidence of an air temperature drop during the eruption year in either the eruption zone or hemisphere, while during yr +1

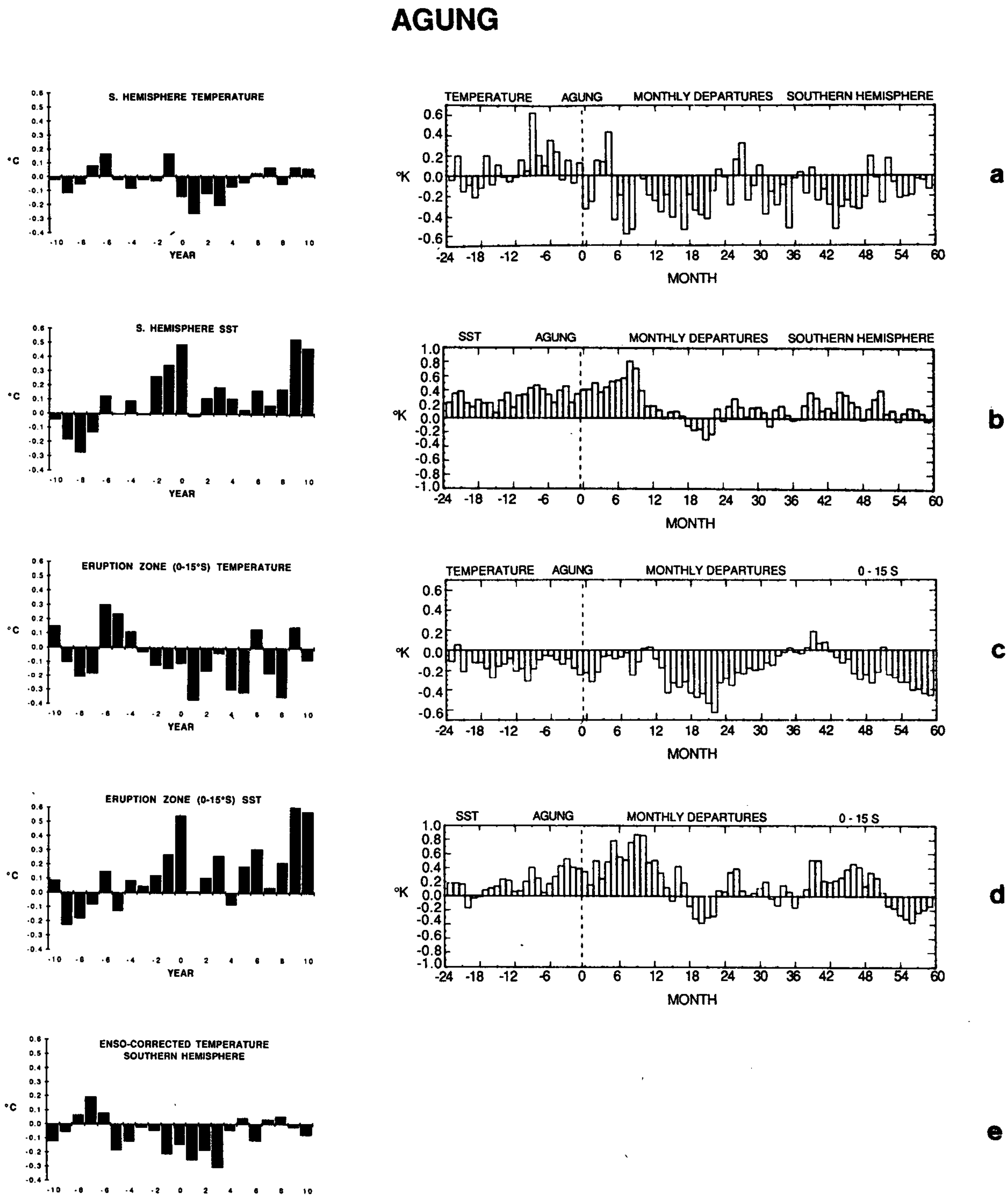

FIG. 18. As in Fig. 13, for Agung.

there is a small (~0.1–0.2°C) decrease in the eruption zone and hemisphere anomalies (Fig. 20a, c). However, it is unlikely that the relatively minor eruption of Fuego was the cause of the posteruption cooling. The signal is strongest in the noneruption hemisphere (not shown) and there is no evidence that Fuego's aerosol cloud ever made it that far. Second, the air temperature evolution is clearly responding to variations in sea surface temperature. As shown in Fig. 2b, temperature dropped to a relative minimum in the equatorial Pacific during late 1975, and the warm air and sea surface temperatures in yr −2 are simply the response to the 1972–73 El Niño.

El Chichón (17°N, 4/1982): Although the El Chi-

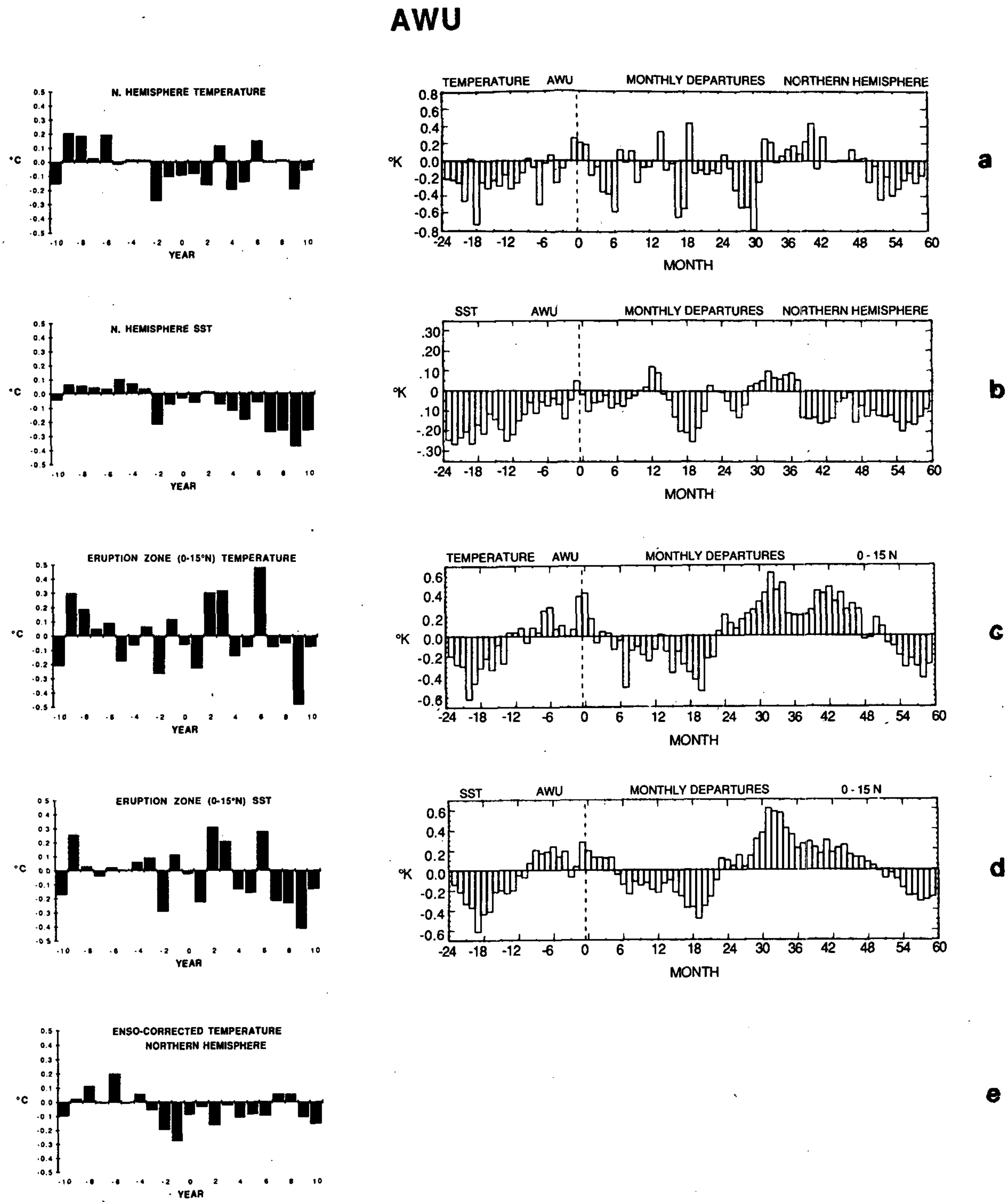

FIG. 19. As in Fig. 13, for Awu.

chón eruption of 1982 may have produced the densest Northern Hemisphere stratospheric aerosol cloud of the past 100 yr, the annual and monthly hemispheric anomalies (Fig. 21a) show only a slight temperature decrease during the 2 yr following the eruption. However, in the eruption band (15°–30°N) the temperature deviation fell to ∼−0.20°C in the eruption year and maintained this anomaly the following year even though one of the strongest ENSO events of recent record was occurring (Fig. 21b). Could it be that global warming due to the ENSO event (see Fig. 2 for tropical SST) was masking the volcanic cooling with the exception of the latitudes adjacent to the eruption, where the aerosol veil was densest and longest lived? (N.B.:

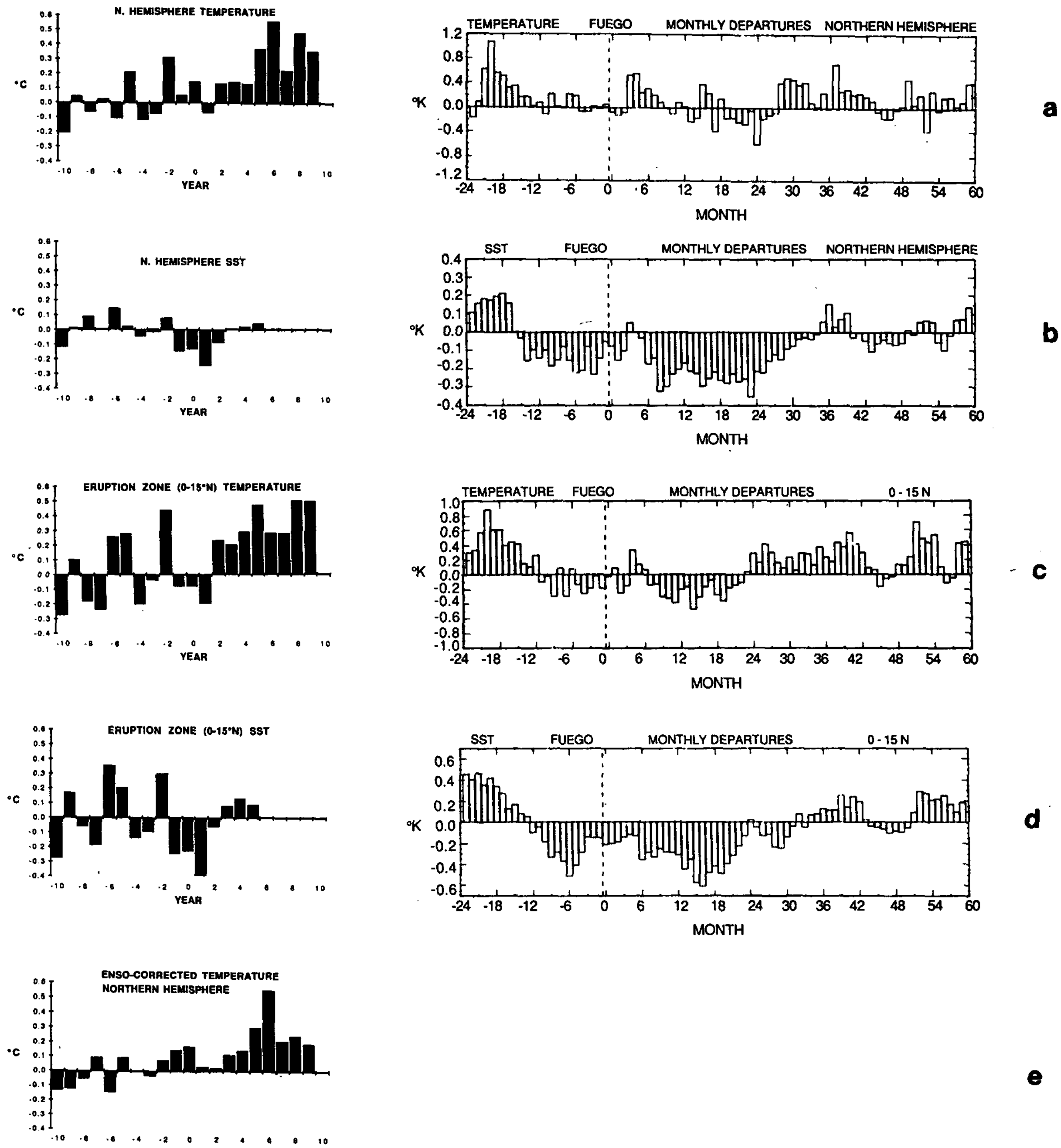

FIG. 20. As in Fig. 13, for Fuego.

the El Chichón stratospheric veil was limited to 5°S to 35°N during the first 6 months.) The ENSO-corrected eruption-hemisphere temperatures (Fig. 21c) do indicate substantial posteruption cooling.

5. Discussion

a. General conclusions

The multieruption composites, shown in section 3, suggest posteruption cooling at the surface. For the complete set of eruptions some volcanic signal seems to exist, and this signal is enhanced by considering only the larger eruptions and by subtracting out another source of interannual variability, the El Niño/Southern Oscillation. The composite air temperature decrease following the larger eruptions (~0.3°C) is similar to what would be expected from theoretical arguments (as described in section 2c). Although the posteruption cooling is suggestive, there is also a disturbing preeruption warming in several of the composites. This

EL CHICHON

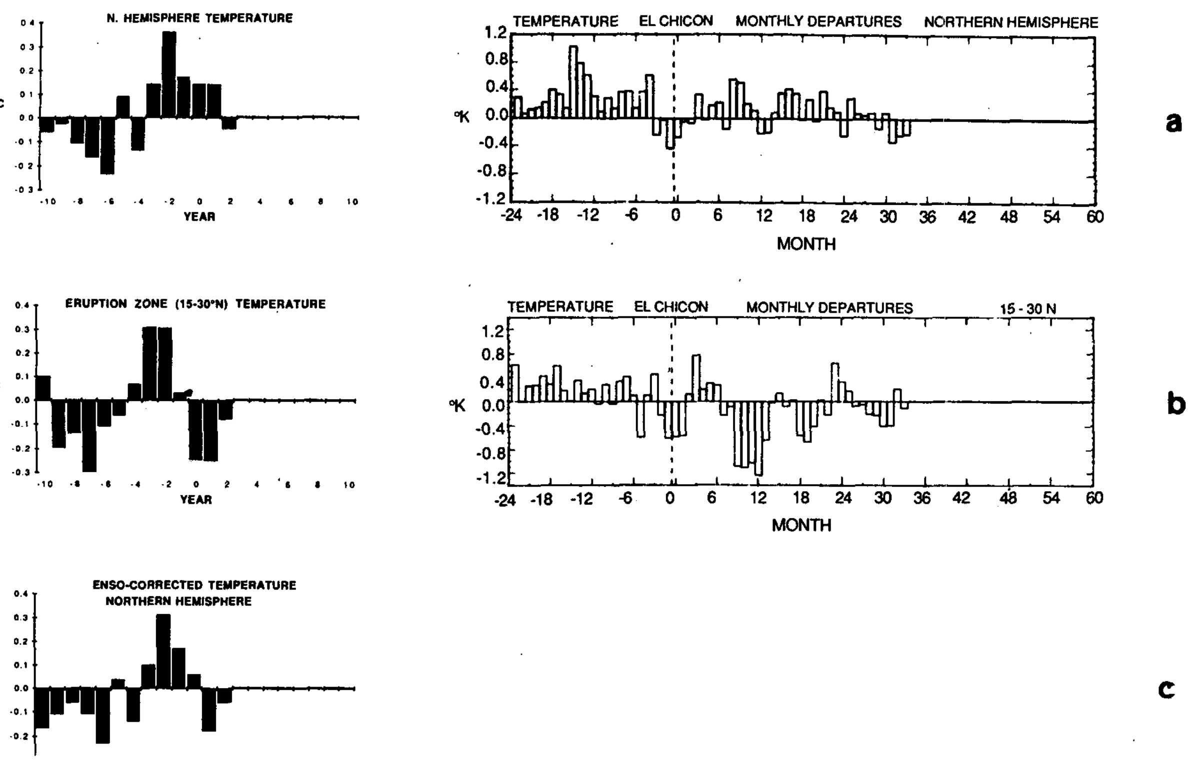

FIG. 21. As in Fig. 13, for El Chichón.

warming mainly is the result of temperature increases prior to two eruptions: Pelee/Sourfriere/Santa Maria and El Chichón. Since it is unlikely that atmospheric and ocean warming initiate volcanic eruptions, this feature provides a warning about conclusions made with such a small collection of volcanic events. On the other hand, some confidence in the reality of a volcanic impact on temperature comes from the fact that weaker eruptions produced little or no composite temperature perturbation.

Six (five) of the nine events discussed in this paper were associated with two (one) year[5] temperature falls in the eruption hemisphere of at least 0.1°C, while in the eruption zone six (seven) of the nine events had a similar temperature decrease. For the five larger eruptions, four (four) of the five experienced at least a 0.1°C temperature decline in the eruption hemisphere over 1 (2) yr (in both cases El Chichón was the exception). In the eruption zone, three (four) out of the five large eruptions experienced similar temperature falls over 1 (2) yr. These results indicate a greater frequency of cooling in the posteruption years than would be expected by chance, especially for the larger events. However, it is important to note that the eruptions that produced the densest and most widespread stratospheric aerosol clouds (e.g., El Chichón) did not necessarily create the greatest cooling or the largest negative temperature anomalies. Both the phasing of the temperature drops and maximum posteruption negative anomalies vary between volcanic events, and cooling seems to begin *prior* to several of the eruptions. For some of the events, posteruption cooling is maintained for more years than would be expected from a volcanic dust cloud, and the apparent signals in the eruption latitudes are not always stronger than those of the eruption hemisphere. As noted in section 4, many of these discrepancies can be explained by ENSO-related variability or the lack of a significant stratospheric aerosol cloud for the smaller eruptions.

The composite and individual eruption chronologies shown above suggest two periods of posteruption decline in air temperature: one occurring during the first few months after the eruption key date and another peaking around a year and a half later. Sea surface temperature holds relatively steady during the first 10 posteruption months and then experiences a 1-yr decline. Thus, the primary air temperature perturbation during the second year occurs at a time when the ocean has had a chance to respond to the change in insolation and appreciable aerosol is still resident in the stratosphere. The amplitude (0.1° to 0.5°C) and phasing of

[5] Two year: yr +1 minus yr −1; one year: yr 0 minus yr −1.

the primary posteruption cooling in the second year is similar to those established by other investigators (e.g., Mass and Schneider 1977; Taylor et al. 1980; Self et al. 1981; Kelly and Sear 1984) and from the theoretical estimates described in section 2 of this paper.

A difference in phasing of the air and sea surface temperature response to volcanic aerosols is consistent with a variety of modeling studies. Harvey and Schneider (1985), using a global energy balance model, found that air temperature decreased over water with about half the amplitude found over land, and that the ocean–air temperature response was delayed by approximately 3 months. The posteruption decline of temperature in the ocean mixed layer was about one-quarter of the land air temperature decrease and was delayed nearly a year (again, compared to air temperature over land). McCracken and Luther (1984), using a statistical–dynamical climate model, found a similar result, with ocean temperatures changing more slowly and with less amplitude than land air temperatures.

b. *Is there an immediate response to volcanic forcing?*

Perhaps the greatest surprise in the composite results is the substantial decline in air temperature during the months immediately following the volcanic events. Similarly, Kelly and Sear (1984) found substantial cooling of Northern Hemisphere temperature during the first 2 or 3 months following several major volcanic eruptions in the Northern Hemisphere. It is difficult to understand how a volcanic eruption could induce such an immediate hemispheric temperature response. Months are required for a sulfuric aerosol cloud to become established and several months more are needed for the volcanic aerosols to spread beyond the latitude band of ejection. Furthermore, ocean mixed layer temperatures, which significantly control and modulate atmospheric air temperatures, cannot respond to a change in external forcing over a period of 1 to 2 months.

Is the short-term response in the above composites real or is it a reflection of random (or unrelated) events occurring during a limited number of volcanic eruptions? Let us attempt to answer this question by first examining the composite of three large eruptions (Fig. 8), which shows a relative minimum in eruption hemisphere surface temperature during month +1. First, it is obvious that this negative excursion began prior to the eruption key date. An examination of the individual component eruptions reveals that eruption hemisphere temperatures in one (Pelee/Soufriere/Santa Maria) fell to nearly the posteruption level prior to the eruption key month. Furthermore, rapid cooling is not observed in the eruption band for that event. Eruption hemisphere air temperature in the period around El Chichón went through its largest negative excursion in the month *before* the eruption and was *recovering* during the few months after the event. Negative temperature anomalies ($\sim$0.25°C) observed for a few months after the Agung eruption were gone by month +3, and were greatly exceeded by large negative anomalies during the next 4 yr. In the Agung eruption band the posteruption negative deviations are clearly a continuation of a preeruption trend. A similar posteruption cooling spike is observed in the composite of five large events (Fig. 6), which includes the substantial events of Krakatau and Tarawera, neither of which possess a significant and immediate posteruption cooling. Thus, one is forced to conclude that the posteruption cooling spike is not of volcanic origin.

The large and immediate posteruption cooling found by Kelly and Sear (1984) also does not appear to be volcanic in origin. First, they used several of the same eruptions as in the present study (e.g., Pelee/Soufriere/Santa Maria); therefore, as explained above there would be a tendency to get posteruption cooling that had nothing to do with the eruptions themselves. They also used some additional (and lesser) eruptions, such as Ksudach and Katmai (which they call Novarupta), which are associated with rapid posteruption cooling. As noted in the previous section the sharp decline in air temperature associated with Ksudach (Fig. 16) followed pre-eruption cooling in the equatorial Pacific, and in the eruption band the negative anomalies began *prior* to the event. Even more suspicious is the strong cooling in the noneruption (Southern) hemisphere (not shown), which occurs as rapidly and with as great an amplitude as that of the eruption (Northern) hemisphere. The Katmai eruption (Fig. 17) also evinced a large air temperature drop in the first 6 months following the eruption. As noted above, there are many reasons to doubt that this negative excursion was of a volcanic origin. In the eruption hemisphere and band cooling occurred simultaneously in air and sea surface temperature; certainly, one would not expect the sea surface temperature to react that rapidly to a change in solar insolation. Furthermore, the posteruption temperature decline appears to be simply following previous changes in the equatorial Pacific SST.

Other problems with the conclusions of Kelly and Sear (1984) are associated with their analysis techniques. In their study they normalized the composite temperature anomalies by the monthly standard deviations, which vary from 0.64 (January) to 0.23 (July). Since the majority of the eruptions occurred in the spring (seven of the nine events occurred in March through June), the division by the mean monthly standard deviations amplified any "signal" during the first few months (i.e., spring and summer) and subsequent late-spring/summer periods. Ksudach and Katmai, the two eruptions with the largest posteruption cooling, were preceded by moderate ENSO events (1905 and 1911–12) that would be associated with above normal sea surface and air temperatures (Quinn et al. 1978; Rasmusson and Wallace 1983). The method used by Kelly and Sear to calculate the anomalies (simply subtracting out the mean of the 12 months preceding the eruption) would tend to create a negative anomaly in the years following such ENSO-precursor eruptions. In summary, because of the eruptions they selected and the compositing techniques used to analyze them, Kelly and Sear (1984) found a rapid posteruption cooling that could not have been produced by the volcanic events considered in their study.

c. *Are volcanic eruptions associated with ENSO events?*

Handler (1984) has suggested that large volcanic eruptions can initiate ENSO events. In a seasonal composite of 11 low-latitude (20°S–20°N) eruptions he found increased sea surface temperatures over the eastern tropical Pacific after these eruptions; in contrast, 20 extratropical (>20° from the equator) eruptions produced cooler sea surface temperatures in the same area. From this evidence Handler hypothesizes that low latitude eruptions, such as the recent El Chichón eruption, can induce ENSO events.

The results of this study do not support Handler's hypothesis. First, it is clear that the ENSO corrections displayed in the monthly composite series do not indicate any consistent relationship between ENSO warming and the volcanic events considered in the composites. Second, as a check on the volcano/ENSO connection we composited monthly sea surface temperatures in the eastern equatorial Pacific (10°N–10°S, 80°W–120°W) using three low-latitude eruptions (Agung, Awu, and Fuego). This composite (Fig. 22) indicates cooling, not warming, during the first 2 yr after the eruption key month. The ENSO volcano relationship is cast into further doubt by Fig. 23, which presents the frequency distribution of the number of moderate and strong ENSO events (as defined by Quinn et al. 1978) relative to the nine eruptions discussed in this study. A relationship between ENSO events and volcanic eruptions is not evident.

There are several additional reasons to question Handler's theory. First, he uses the Volcanic Explosivity Index (VEI) as the sole measure of the amount of aerosol injected into the stratosphere by a volcanic event. However, explosivity is not the only or even the best measure of the yield of stratospheric aerosol after an eruption and other parameters such as sulfur emission are as or more important. As a result, Handler's composites include many eruptions that had little or

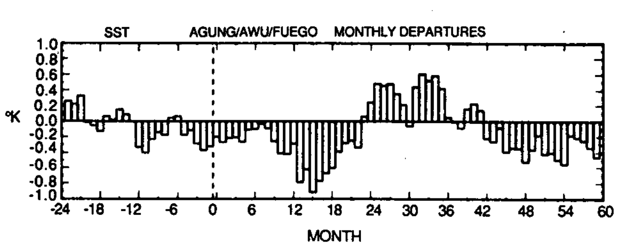

FIG. 22. Monthly sea surface temperature anomalies for three low-latitude eruptions (Agung, Awu and Fuego) for the region (10°N–10°S, 80°W–120°W).

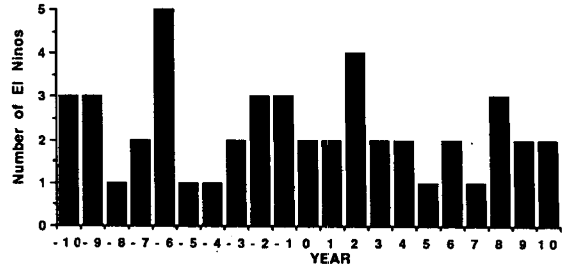

FIG. 23. Frequency distribution for the number of moderate and strong El Niño events (from Quinn et al. 1978) versus volcanic eruption key dates.

no impact on stratospheric aerosol loading. Handler's results indicate that anomalous sea surface temperature warming occurs one season (3 months) after volcanic events. Such a rapid response to external forcing is inconsistent with the large thermal inertia of the oceans and is not supported by modeling studies (e.g., Harvey and Schneider 1985). Handler does not give a credible theory for the very different response following low latitude and extratropical eruptions and shows only two seasons prior to the eruption key dates, making it impossible to judge the significance of the purported volcanic signal. Nicholls (1988), in a comparison of the Darwin pressure anomaly with the dates of 10 eruptions used in Handler (1986), found a significant relationship between ENSO events and some volcanic eruptions, but with the ENSO signal *preceding* the volcanic events. He concluded that this relationship is most likely due to chance and small sample size.

d. *The historical perspective: The question of the "year without a summer"*

Probably the most cited eruption in the volcano–climate debate is the cataclysmic explosion of Tambora in April 1815. Milham (1924), Hoyt (1958), Stommel and Stommel (1979) and many others have suggested that the dust veil from Tambora produced unusually cold temperatures and crop failures in eastern North America and northern Europe during the summer of 1816.

Landsberg and Albert (1974) examined surface temperatures for a collection of stations in western Europe and North America in order to appraise the climatic significance of this eruption. They found that although the summer of 1816 was cold in eastern North America and western Europe, such temperatures were "neither unprecedented in either of these areas at the time, nor statistically particularly unique, with similar events occurring also in later years that were not especially distinguished by major volcanic eruptions." Temperatures in eastern Europe and central North America were considerably milder than those of the above regions. Angell and Korshover (1985), examining both regional and hemispheric temperature variations after Tambora, found little evidence for posteruption cooling for the Northern Hemisphere or Europe. Only at New Haven, Connecticut was a suggestive temperature decrease observed.

Landsberg and Albert suggested another mechanism for the cold anomalies in parts of Europe and North America: a persistent long-wave pattern with cold meridional flows one wavelength apart over the stricken regions and milder temperatures at other locations. Corroboration of this hypothesis is found in Catchpole and Faurer (1983), which shows that the sea ice patterns of 1816 were consistent with a highly meridional atmospheric circulation over eastern North America that allowed southward excursions of Arctic air. Lamb and Johnson (1966) reconstructed the mean pressure pattern for July 1816 over the North Atlantic Ocean and came to the same conclusion. The question is whether major volcanic eruptions can induce or trigger such long-wave amplification.

We know of no general circulation model simulation that suggests that a small reduction in net insolation, as one might expect from a volcanic dust veil, would dramatically amplify the meridional motions of the atmosphere. In fact, the general circulation experiments of Hunt (1977) indicate that the "impact of the volcanic debris on the dynamical behavior of the general circulation was quite small, at most slightly reducing the zonal wind." However, it should be noted that the Hunt model was relatively primitive, and did not include land/ocean contrasts or topography. A suggestion that volcanic eruptions can produce large longitudinal variability is proffered in the study of Lough and Fritts (1987), which, using tree-ring data, concludes that major volcanic eruptions (VEI > 3) are associated with above-normal temperatures over the western U.S. and below-normal temperatures in the central part of the nation. Lough and Fritts' work indicates that low latitude eruptions produce the largest climatic effects; their finding of contrasting and oppositely phased signals in adjacent geographical regions suggests an amplification in the hemispheric wave pattern. In contrast, LaMarche and Hirchboeck (1984) used frost-damage zones in the annual rings of trees to conclude that major volcanic eruptions (as determined from Lamb's Dust Veil Index) are associated with below-normal temperatures in the western U.S. Clearly, the question of volcano-induced long-wave amplification has not been resolved.

6. Conclusions and summary

This paper appraises the reality of volcanic influence on surface climate by examining the response of surface air temperature, sea surface temperature, surface pressure and rainfall after major volcanic events of the past century. Both individual eruption time series and multieruption composites are constructed and analyzed. Included in this work is an attempt to remove one source of interannual variability: the El Niño/Southern Oscillation (ENSO). These exercises indicate that only the largest eruptions (in terms of producing a stratospheric dust cloud) are suggested in the climatic record, and that modest cooling (~0.1°–0.2°C) is observed for 1 to 2 yr after these large events. Previous suggestions of large cooling during the first few months after volcanic events appear to be unwarranted.

The interannual variability produced by tropical sea surface temperature variations (i.e., ENSO cycles) is very apparent in individual eruption air temperature series. For some eruptions this ENSO signal masks the volcanic signal, while during others it gives a spurious impression of a volcanic effect on temperature. An attempt to remove the ENSO signal from the temperature composites for the major eruptions strengthened the apparent volcanic signal. In contrast, for the composite of weaker eruptions attempts to eliminate the ENSO signal did not enhance or create an apparent volcanic signal.

There is little hint of a volcanic influence on surface pressure or precipitation.

Why is the volcanic effect on surface climate so small? An examination of the actinometric observations made during the past century indicates that major volcanic eruptions significantly influence the earth's radiation balance. Volcanic stratospheric aerosols (composed mainly of sulfuric acid) can drastically reduce (by up to tens of percent) the direct solar beam; however, as a result of a nearly commensurate augmentation of diffuse radiation from the strongly forward-scattering volcanic aerosols, decreases in total radiation are relatively minor (only a few percent). This small decrease in net insolation, coupled with warming due to the infrared opacity of the stratospheric aerosols, leaves only a minor deficit in radiative heating as a result of even a major volcanic eruption. Furthermore, volcanic aerosol clouds usually cover only a limited portion of the globe, often extend slowly away from the eruption latitude, and exist for only a short period (1–3 yr) compared to the response time of the ocean/atmosphere system (4–5 yr). As a result, the volcanic temperature signal of even a major eruption is expected to be small, probably no more than a few tenths °C. Such a signal is difficult to find in a limited and noisy climatic record where interannual variability (such as produced by ENSO events), seemingly random variations of the climatic system, instrument error, and a host of other factors can easily obscure a weak signal. A further complication is the very limited number (~5) of truly large volcanic stratospheric aerosol events during the past century, when a relatively representative global dataset has been available.

There is much that still needs to be done in the investigation of volcanic effects on climate. First, further exploration and verification of proxy records of climate and volcanic magnitude should be given high priority; only in this way can we acquire enough cases to enable a more conclusive statement on the influence of volcanos on climate. Additional research is required on the modulation of longitudinal and temporal variability by volcanic events. Do major volcanic eruptions enhance longitudinal temperature differences as suggested by Stommel and Stommel (1979) and Lough and Fritts (1987)? An acute need also exists for atmospheric and coupled atmosphere/ocean general circulation model simulations to test many of the present speculations about atmosphere/ocean response to a small decrease in solar forcing. Such simulations would also help direct the examination of the observed climatic record to the regions and modes of maximum response. Finally, better analysis and statistical tools must be forged for the evaluation of weak, nonperiodic signals in a noisy climatic record.

Acknowledgments. This work has been funded by the Climate Dynamics Program of the National Science Foundation (ATM-8313647). Special thanks goes to Clara Deser, who provided Pacific sea surface temperature data, regression statistics, and invaluable advice and assistance. We also appreciate the constructive comments and suggestions of Alan Robock and an anonymous reviewer. The bulk of the computations were completed at the National Center for Atmospheric Research, which is supported by the National Science Foundation.

APPENDIX A

Review of Selection Criteria

In order to appraise the climatic impact of volcanic eruptions a variety of eruptive indices have been considered. Each index represents a different approach to determining the quantity of radiatively active stratospheric aerosol injected by volcanic events. This appendix reviews several indices and describes their strengths and weaknesses.

1. Major volcanic indices

a. The Volcanic Explosivity Index (VEI)

This index, described in Newhall and Self (1982) and Simkin et al. (1981), is an attempt to semiquantitatively evaluate the explosive magnitude of over 8000 eruptions. A value of 0 to 8 is assigned (ranging from 0 for nonexplosive eruptions to 8 for the most energetic events) based on a variety of nonatmospheric factors including volume of ejecta, column height, duration, qualitative description and other parameters.

Although the explosive magnitude of an eruption does significantly influence the amount of material injected into the stratosphere, it is not the only important parameter. For example, Pollack et al. (1976), Hamill et al. (1977), Self and Rampino (1988) and others have noted that the most important constituent of long-lived stratospheric dust veils is sulfate aerosol, and that the amount of this aerosol is crucially dependent on the quantity of sulfur gas released by the volcanic event. Furthermore, large VEIs can be associated with powerful eruptions in which most of the energy is radiated laterally rather than in the vertical. For these and other reasons, the explosive magnitude of an eruption, expressed by the VEI, is not a sufficient measure of the resulting stratospheric dust veil. The VEI may be a useful measure of how effectively a volcano could potentially loft sulfur compounds *if* they are present.

b. The Dust Veil Index (DVI)

The DVI was devised by Lamb (1970) to provide a numerical assessment of the amount of volcanic dust injected into the stratosphere. Its value is based on the depletion of the direct solar beam, temperature-lowering in midlatitudes, and the quantity of solid materials dispersed as dust; furthermore, the DVI gives a measure of not only the opacity of dust veils, but their extent and duration as well. To mitigate circular reasoning in studies of the climatic impact of volcanic eruptions, Lamb provides a version of the DVI that is not based on temperature effects. A consideration of the qualitative nature of much of the data used, the variations in data type and quality, and the necessarily subjective nature of many of the evaluations, suggests that this index alone may not be a sufficient basis for determining the magnitude of aerosol injection into the stratosphere.

c. Actinometric observations

Many papers discuss the effects of volcanic eruptions on atmospheric turbidity, with particular emphasis given to the transmission of the direct solar beam. Wexler (1951) noted that a variety of sites around the world experienced a significant reduction in normal solar radiation following the Krakatau, Santa Maria and Katmai eruptions. After Krakatau solar radiation flux took nearly 3 yr to recover, while for the other two eruptions the recovery period was about half as long. Dyer (1974) presented solar transmission data for a variety of stations from 1883 to 1972. He found that although some eruptions are clearly evident in these records (e.g., Mt. Agung, Ksudach, Katmai, Santa Maria/Soufriere, Krakatau) others are not because of their proximity to earlier eruptions, lack of data, or their own inherent weak signal. Mendonca et al. (1978) examined the monthly transmission of normal-incidence solar radiation at Mauna Loa Observatory in Hawaii. They found very strong evidence of reduced transmission for 2 to 3 yr after the eruption of Mt. Agung in 1963, with some suggestion of weaker, shorter-duration signals after the 1966 Awu and 1974 Fuego events. Hoyt et al. (1980), in an examination of normal-incidence pyrheliometric measurements at three sites in the United States, found evidence of the eruptions of Mt. Agung, Awu and Fuego. Keen (1983) studied the illumination of the moon during 21 lunar eclipses (since 1960) as a measure of stratospheric aerosol loading produced by volcanic eruptions. His results indicate that Agung and El Chichón had significant, and approximately equal effects, with little evidence of enhanced stratospheric aerosol after the eruptions of Fuego and Fernandina. Wendler (1984) found that at Fairbanks, Alaska the direct solar beam was reduced by 24.8%, and the global (or total) radiation was reduced by ~5% during the year after El Chichón.

d. Chemical analyses of polar ice cores

Several studies have been made of sulfate and trace metal concentrations in polar ice samples (Delmas and Boutron 1978, 1980; Boutron 1980; Hammer et al. 1980, 1981). It has been suggested that stratospheric volcanic aerosols precipitate onto the polar ice sheets, and thus ice cores can be used to date and, to some degree, quantify the magnitude of volcanic events. A careful examination of the above references suggests that sulfate and trace metals concentrations in ice cores are useful, but imperfect, measures of the extent and density of volcanic dust veils. Although the most significant stratospheric aerosol events (e.g., Krakatau, Mt. Agung) are quite evident in the polar ice cores of the eruption hemisphere, there are substantial excursions in the concentration histories that are not associated with any major eruption. It has been suggested (e.g., Self and Rampino 1988) that lesser eruptions in the high latitudes might produce a substantial signal in ice core chronologies without producing significant stratospheric aerosols. Thus, polar ice core chronologies cannot be used alone to appraise volcanic aerosol concentrations but are valuable supplements to the indices described above.

APPENDIX B

Review of the Selected Volcanic Events

1. Krakatau

Located on the island of Krakatau in Indonesia (6.1°S, 105.4°E), this volcano began erupting on 20 May 1883. On 26–27 August 1883 a cataclysmic eruption destroyed much of the mountain and created a column of gas and dust that reached nearly 27 km ASL. A stratospheric dust cloud spread throughout the world producing brilliant sunsets and attenuated the direct solar beam by up to 20%–30% for a period of 2 to 3 yr. This eruption was given a very high DVI of 1000 and the large VEI of 6. Ice cores in both the Arctic and Antarctic indicated enhanced acidity and trace metal content following this event.

2. Tarawera

This energetic eruption (DVI of 800 and VEI of 5) occurred on New Zealand's North Island (38.2°S, 175.5°E) on 10 June 1886. Polar ice cores in both hemispheres indicated enhanced acidity after this eruption. A significant increase of turbidity, as indicated by a reduction in the direct solar beam, has been noted.

3. Pelee/Soufriere/Santa Maria

Due to their temporal and geographic proximity, these three eruptions are considered as one event in this study. Clearly, more stratospheric aerosols were produced by Soufriere (13.3°N, 61.2°W, 17 May 1902) than Mont Pelee (14.8°N, 61.2°W, 8 May 1902); however, the latter is infamous for claiming over 30 000 lives as a cloud of avalanching hot gases and ash descended on nearby towns. Soufriere was a moderately explosive eruption (DVI of 300, VEI of 4); Mount Pelee has a similar VEI but a lesser DVI of 300. On 24 October 1902 Santa Maria (14.8°N, 91.5°W) erupted violently and has been assigned a large VEI of 6 and a DVI of 600. The combined Pelee/Soufriere/Santa Maria dust veil was evident in both the attenuation of the direct solar beam and in trace metals in Antarctic ice cores. Wexler (1951) noted that the direct solar beam at European sites was attenuated by up to 20% for approximately a year and a half and did not return to preeruption values until 1905. Twilight phenomena were reported until 1904.

4. Ksudach

Also known as Shtyubelya Sopka and located in Kamchatka, U.S.S.R., Ksudach (51.8°N, 157.5°E) erupted vigorously (DVI of 500, VEI of 5) on 28 March 1907. Although ice core evidence is lacking, twilight effects were observed until 1909. Dyer (1974) reported that the direct solar beam was attenuated by this event.

5. Katmai

Also referred to as Novarupta, this volcano is located on the Alaskan Peninsula (58.3°N, 155.0°W) and erupted on 6 June 1912 (DVI of 500, VEI of 6). The direct solar beam was weakened by up to 25% in the Northern Hemisphere during late 1912, and did not recover until the end of 1914 (Lamb 1970). Kimball (1918) noted diminished atmospheric transparency and unusually brilliant sunsets throughout the U.S., Europe and North Africa during the second half of 1912.

6. Agung

This volcano, located on the Indonesian island of Bali (8.3°S, 155.5°E) began erupting on 19 February 1963, with the greatest explosions occurring on 17 March. The dust veil spread globally during 1963 and was detectable in anomalous twilight effects until early 1966. The dust veil was far larger in the Southern than the Northern Hemisphere, with Southern Hemisphere direct solar beam deficits ranging as high as 30%. This moderately explosive volcano (VEI of 4) was given a very large DVI of 800 by Lamb (1970). Evidence of this eruption was also reported in polar ice cores, and in darkening of the moon during lunar eclipses (Keen 1983).

7. Awu

Awu erupted on Great Sangihe Island in Indonesia (3.7°N, 125.5°E) on 12 August 1966. A modest eruption compared to the others considered in this study, it was assigned a DVI of 200 and VEI of 4. Dyer (1974) reported an attenuation in the direct solar beam and Delmas and Boutron (1980) indicated that there was some enhanced acidity in Antarctic ice following this eruption.

8. Fuego

This Guatemalan volcano (14.5°N, 90.9°W) erupted violently on 14 October 1974, producing a column of gas and dust that rose to 20 km. The resulting stratospheric dust veil spread over the Northern Hemisphere and was observed in twilight effects for over a year. Keen (1983) reported a noticeable darkening of the moon after this eruption. It is curious that Delmas and Boutron (1980) and Boutron (1980) report a signal from this volcano in the Antarctic ice, yet Hammer et al. (1980) found nothing in the Greenland ice cores where the evidence should have been much stronger. Lidar observations at Hampton, Virginia (McCormick et al. 1978) indicate that the aerosol cloud decayed to background levels within 1.5 yr (March 1976).

9. El Chichón

Located in Chiapas, Mexico (17.3°N, 93.2°W), El Chichón began erupting on 28 March 1982, with the largest eruption occurring on 3–4 April 1982. The eruption column was observed to reach nearly 30 km and spread into both hemispheres (Mitchell 1982). The emissions of this volcano were particularly rich in sulfur and produced the densest stratospheric dust veil in the Northern Hemisphere since the 1883 eruption of Krakatau (Rampino and Self 1984; Robock 1983).

10. Volcanos that were not selected

There are several well known volcanoes of the past century that were not included in the above list (e.g., Mt. St. Helens 1980; Bezymiannaya 1956). This exclusion is based on our assessment that these eruptions had a lesser contribution to stratospheric aerosol loading. For example, it is generally agreed (see Robock 1981; Self and Rampino 1988) that Mt. St. Helens did not have a significant climatic effect. Although this eruption was relatively energetic (VEI of 5), the combination of low-sulfur content of the volcanic effluent and the predominantly lateral nature of the blasts, resulted in only a minor increase in the stratospheric aerosol burden.

REFERENCES

Angell, J. K., 1981: Comparison of variations in atmospheric quantities with sea surface temperature variations in the equatorial eastern Pacific. *Mon. Wea. Rev.*, **109**, 230–243.

——, and J. Korshover, 1978: Estimate of global temperature variation in the 1000–30 mb layer between 1958 and 1977. *Mon. Wea. Rev.*, **106**, 1422–1432.

——, and ——, 1984: Comparison of tropospheric temperatures following Agung and El Chichón volcanic eruptions. *Mon. Wea. Rev.*, **112**, 1457–1463.

——, and ——, 1985: Surface temperature changes following six major volcanic episodes between 1780 and 1980. *J. Climate Appl. Meteor.*, **24**, 937–951.

Boutron, C., 1980: Respective influence of global pollution and volcanic eruptions on the past variations of the trace metals content of Antarctic snow since the 1880's. *J. Geophys. Res.*, **6**, 7426–7432.

Cadle, R. D., C. S. Kiang and J.-F. Louis, 1976: The global scale dispersion of the eruption clouds from major volcanic eruptions. *J. Geophys. Res.*, **81**, 3125–3132.

Catchpole, A. J. W., and M.-A. Faurer, 1983: Summer sea ice severity in Hudson Strait, 1751–1870. *Climatic Change*, **5**, 115–139.

Chou, M.-D., L. Peng and A. Arking, 1984: Climate studies with a multilayer energy balance model. Part III: Climatic impact of stratospheric volcanic aerosols. *J. Atmos. Sci.*, **41**, 759–767.

Deirmendjian, D., 1973: On atmospheric turbidity anomalies. *Adv. Geophys.*, **16**, 267–297.

Delmas, R., and C. Boutron, 1978: Sulfate in Antarctic snow: spatio-temporal distribution. *Atmos. Environ.*, **12**, 723–728.

——, and ——, 1980: Are the past variations of the stratospheric sulfate burden recorded in central Antarctic snow and ice layers? *J. Geophys. Res.*, **85**, 5645–5649.

Dyer, A. J., 1974: The effect of volcanic eruptions on global turbidity and an attempt to detect long-term trends due to man. *Quart. J. Roy. Meteor. Soc.*, **100**, 563–571.

——, and B. B. Hicks, 1965: Stratospheric transport of volcanic dust inferred from solar radiation measurements. *Nature*, **208**, 131–133.

Ellsaesser, H. W., 1977: Comments on "Estimate of the global temperature, surface to 100 mb, between 1958 and 1975." *Mon. Wea. Rev.*, **105**, 1200–1201.

Gentilli, J., 1948: Present-day volcanicity and climatic change. *Geol. Mag.*, **85**, 172–175.

Hamill, O., O. Toon and C. Kiang, 1977: On the formation of stratospheric aerosols. *J. Atmos. Sci.*, **34**, 1104–1119.

Hammer, C. U., H. B. Claussen and W. Dansgaard, 1980: Greenland ice sheet evidence of post glacial volcanism and its climatic impact. *Nature*, **288**, 230–235.

——, —— and ——, 1981: Past volcanism and climate revealed by Greenland ice cores. *J. Volcanol. Geotherm. Res.*, **11**, 3–10.

Handler, P., 1984: Possible association of stratospheric aerosols and El Niño type events. *Geophys. Res. Lett.*, **11**, 1121–1124.

——, 1986: Possible association between the climatic effects of stratospheric aerosols and sea surface temperatures in the eastern tropical Pacific Ocean. *J. Climate*, **6**, 31–41.

Harshvardhan, 1979: Perturbation of the zonal radiation balance by a stratospheric aerosol layer. *J. Atmos. Sci.*, **36**, 1274–1285.

Harvey, L. D. D., and S. H. Schneider, 1985: Transient response to external forcing on 10^0–10^4 year time scales, 2, sensitivity experiments with a seasonal, hemispherically averaged, coupled atmosphere, land, ocean energy balance model. *J. Geophys. Res.*, **90**, 2207–2222.

Horel, J. D., and J. M. Wallace, 1981: Planetary-scale atmospheric phenomena associated with the southern oscillation. *Mon. Wea. Rev.*, **109**, 813–829.

Hoyt, D. V., C. P. Turner and R. D. Evans, 1980: Trends in atmospheric transmission at three locations in the United States from 1940–1977. *Mon. Wea. Rev.*, **108**, 1430–1439.

Hoyt, J. B., 1958: The cold summer of 1816. *Ann. Assoc. Amer. Geogr.*, **48,** 118–131.

Humphreys, W. J., 1940: *Physics of the Air,* Third ed. McGraw Hill, 676 p.

Hunt, B. G., 1977: A simulation of the possible consequences of a volcanic eruption on the general circulation of the atmosphere. *Mon. Wea. Rev.*, **105,** 247–260.

Jones, P. D., S. C. B. Raper, R. S. Bradley, H. F. Diaz, P. M. Kelly and T. M. L. Wigley, 1986: Northern Hemisphere surface air temperature variations, 1851–1984. *J. Climate Appl. Meteor.*, **25,** 161–179.

Keen, R. A., 1983: Volcanic aerosols and lunar eclipses. *Science,* **222,** 1011–1013.

Kelly, P. M., and C. B. Sear, 1984: Climatic impact of explosive volcanic eruptions. *Nature,* **311,** 740–743.

Kimball, H. H., 1918: Volcanic eruptions and solar radiation intensities. *Mon. Wea. Rev.*, **46,** 355–356.

LaMarche, V. C., and K. K. Hirschboeck, 1984: Frost rings in trees as records of major volcanic eruptions. *Nature,* **307,** 121–126.

Lamb, H. H., 1970: Volcanic dust in the atmosphere; with a chronology and assessment of its meteorological significance. *Phil. Trans. Roy. Soc. London,* **266,** 425–533.

——, and A. I. Johnson, 1966: Secular variations of the atmospheric circulation since 1750. *Geophysical Memoirs,* No. 110, Meteorological Office, London, 125 pp.

Landsberg, H. E., and J. M. Albert, 1974: The summer of 1816 and volcanism. *Weatherwise,* **27,** 63–66.

Lough, J. M., and H. C. Fritts, 1987: An assessment of the possible effects of volcanic eruptions on North American climate using tree-ring data, 1602 to 1900 A.D. *Climatic Change,* **10,** 219–239.

Mass, C., and S. Schneider, 1977: Statistical evidence on the influence of sunspots and volcanic dust on long-term temperature records. *J. Atmos. Sci.*, **34,** 1995–2004.

McCormick, M. P., T. J. Swissler, W. P. Chu and W. H. Fuller, Jr., 1978: Post-volcanic stratospheric aerosol decay as measured by lidar. *J. Atmos. Sci.*, **35,** 1296–1303.

McCracken, M. C., and F. M. Luther, 1984: Preliminary estimate of the radiative and climatic effects of the El Chichón eruption. *Geofis. Int.*, **23-3,** 385–401.

Mendonca, B., K. Hanson and J. DeLuisi, 1978: Volcanically related secular trends in atmospheric transmission at Mauna Loa Observatory, Hawaii. *Science,* **202,** 513–515.

Milham, W. I., 1924: The year 1816—the cause of abnormalities. *Mon. Wea. Rev.*, **52,** 563–579.

Mitchell, J. M., Jr., 1982: El Chichón, weather-maker of the century. *Weatherwise,* **35,** 252–259.

Newell, R. E., 1970: Stratospheric temperature change from the Mt. Agung volcanic eruption of 1963. *J. Atmos. Sci.*, **27,** 977–989.

——, 1979: Climate and the ocean. *Amer. Sci.*, **67,** 405–416.

Newhall, C. G., and S. Self, 1982: The volcanic explosivity index (VEI): an estimate of explosive magnitude for historical volcanism. *J. Geophys. Res.*, **87,** 1231–1238.

Nicholls, N., 1988: Low latitude volcanic eruptions and the El Niño-Southern Oscillation. *J. Climate.*, **8,** 91–95.

Pan, Y. H., and A. H. Oort, 1983: Global climate variations connected with sea surface temperature anomalies in the eastern equatorial Pacific Ocean for the 1958–73 period. *Mon. Wea. Rev.*, **111,** 1244–1258.

Philander, S. G. H., 1983: El Niño Southern Oscillation phenomena. *Nature,* **302,** 295–301.

Pollack, J. B., O. Toon, C. Sagan, A. Summers, B. Baldwin and W. Van Camp, 1976: Volcanic explosion and climatic change: A theoretical assessment. *J. Geophys. Res.*, **81,** 1071–1083.

Pueschel, R. F., L. Machta, G. F. Cotton, E. C. Flowers and J. T. Peterson, 1972: Normal incidence radiation trends on Mauna Loa, Hawaii. *Nature,* **240,** 545–547.

Quinn, W. H., D. O. Zopf, K. S. Short and R. T. W. Kuo Yang, 1978: Historical trends and statistics of the Southern Oscillation, El Niño, and Indonesian droughts. *Fishery Bulletin,* **76,** 663–678.

Quiroz, R. S., 1983: The climate of the "El Niño" winter of 1982–1983—a season of extraordinary climatic anomalies. *Mon. Wea. Rev.*, **111,** 1685–1706.

Rampino, M. R., and S. Self, 1984: The atmospheric effects of El Chichón. *Sci. Amer.*, **250,** 48–57.

Rasmusson, E. M., and J. M. Wallace, 1983: Meteorological aspects of the El Niño/Southern Oscillation. *Science,* **222,** 1195–1202.

Robock, A., 1981: The Mount St. Helens volcanic eruption of 18 May 1980: minimal climatic effect. *Science,* **206,** 1402–1404.

——, 1983: The dust cloud of the century. *Nature,* **301,** 373–374.

Roosen, R. G., R. J. Angione and C. H. Klemcke, 1973: Worldwide variations in atmospheric transmission. Part 1: baseline results from Smithsonian observations. *Bull. Amer. Meteor. Soc.*, **54,** 307–316.

Schneider, S. H., and R. E. Dickinson, 1974: Climate-modeling. *Rev. Geophys. Space Phys.*, **12,** 447–493.

Self, S., and M. R. Rampino, 1988: The relationship between volcanic eruptions and climate change: Still a conundrum? *Eos,* **6,** 74–86.

——, —— and J. Barbera, 1981: The possible effects of large 19th and 20th century eruptions on zonal and hemispheric surface temperatures. *J. Volcanol. Geothermal Res.*, **11,** 41–60.

Simkin, T., L. Seibert, L. McClelland, W. G. Melson, D. Bridge, C. Newhall and J. Latter, 1981: *Volcanoes of the World.* Hutchinson Ross, 232 pp.

Stommel, H., and E. Stommel, 1979: *Volcano Weather: the Story of the Year Without a Summer, 1816.* Seven Seas Press, 177 pp.

Taylor, B., T. Gal-Chen and S. Schneider, 1980: Volcanic eruptions and long-term temperature records: an empirical search for cause and effect. *Quart. J. Roy. Meteor. Soc.*, **106,** 175–199.

Turco, R. P., R. C. Whitten and O. B. Toon, 1982: Stratospheric aerosols: observation and theory. *Rev. Geophys.*, **20,** 233–279.

Viebrock, H. J., and E. C. Flowers, 1968: Comments on the recent decrease in solar radiation at the South Pole. *Tellus,* **20,** 400–411.

Volz, F. E., 1970: Atmospheric turbidity after the Agung eruption of 1963 and size distribution of the volcanic aerosol. *J. Geophys. Res.*, **75,** 5185–5193.

Wendler, G., 1984: Effects of the El Chichón volcanic cloud on solar radiation received at Fairbanks, Alaska. *Bull. Amer. Meteor. Soc.*, **65,** 216–218.

Wexler, H., 1951: On the effects of volcanic dust on insolation and weather. *Bull. Amer. Meteor. Soc.*, **32,** 10–15.

——, 1956: Variations in insolation, general circulation and climate. *Tellus,* **8,** 480–494.

Wright, P. B., 1977: The southern oscillation-patterns and mechanisms of the teleconnections and their persistence. Hawaii Institute of Geophysics, Rep. HIG-77-13, 107 pp.