On the Response of Hemispheric Mean Temperature to Stratospheric Dust: An Empirical Approach

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ABSTRACT

The cooling effects of stratospheric dust and the corresponding response times for hemispheric mean temperature changes are studied using an empirical approach and a simple time-dependent formulation. The approach couples estimates of stratospheric dust injections by volcanic eruptions to an available record of mean temperature anomalies. The time period examined is 1883–1968. Cooling coefficients (per unit mass of dust) are developed for continued loadings from the transient data; the results are compared to theoretical values developed elsewhere. Response times are also compared. The many uncertainties are noted, as are difficulties introduced by apparent underlying temperature trends. The effort is exploratory in nature, but the overall results are consistent with past arguments that stratospheric dust (as from large volcanic eruptions) is of climatic significance. It is suggested, since climatic changes involve integrals over time, that long-term records of stratospheric dust, as well as other recognized climate-determining parameters, should be established and maintained.

1. Introduction

In this paper a simple, empirically calibrated, time-dependent formulation is used to estimate the effects of stratospheric dust on hemispheric mean temperatures. The formulation utilizes a cooling coefficient per unit mass of stratospheric dust that applies to a steady-state change—in a time frame of years to decades[1]—but which is developed from the dynamic changes that occur following large volcanic eruptions. The temperature changes are obtained from an available record of mean temperature anomalies over the period 1883–1968. The cooling effect of an eruption depends on the response time of the ocean-land surface-atmosphere system; thus, a lumped response time is also developed empirically. The results so obtained are compared to theoretical results reported or developed elsewhere (Hidalgo, 1974; Budyko, 1974). Because of the uncertainties involved the paper is exploratory in nature; however, it is believed that the method should be sufficiently sensitive to show up any serious inconsistencies between theory and available data.

Many previous investigators have considered the question of cooling effects of volcanic dust, as evidenced by the climatic record, generally considering each eruption independently, but apparently without reaching an established consensus (see, e.g., Arakawa *et al.*, 1955; Mitchell, 1961, 1971; Humphreys, 1964; Reitan, 1971; Budyko[2]; Landsberg and Albert, 1974; Bauer and Oliver, 1975). Budyko (1974) has also considered briefly the averaged effects of frequent volcanic activity, arguing that the volcanically active period 1883–1912 was cooler than the subsequent similar period because of greater stratospheric dust. In addition, very recent data from deep-sea drilling sections (Kennett and Thunnel, 1975) suggest a correlation between strong volcanic activity and ice ages, tending to confirm speculations often made in the past.

The approach used here differs from that used in prior studies; it is described in detail following a discussion of the climatic and volcanic records. In essence, however, it is a time-integrated approach in which it is assumed that volcanic eruptions are a major factor in short-term climate change, and that such eruptions provide valuable, if unfortunately not completely satisfactory, indications of the dynamics of climatic temperature changes. Other possible causes of climatic change are discussed briefly, but are not treated quantitatively.

2. The climatic and volcanic records

A Northern Hemisphere yearly mean annual temperature anomaly record over the period 1880–1968 is shown in Fig. 1, along with a record of major volcanic

[1] Possible very long term effects, such as the postulated ice albedo feedback effect [which may require thousands of years to be significant (Budyko, 1974a)] are not included. Climate undoubtedly responds on a variety of time scales.

[2] Private communication to A. J. Grobecker, U. S. Dept. of Transportation.

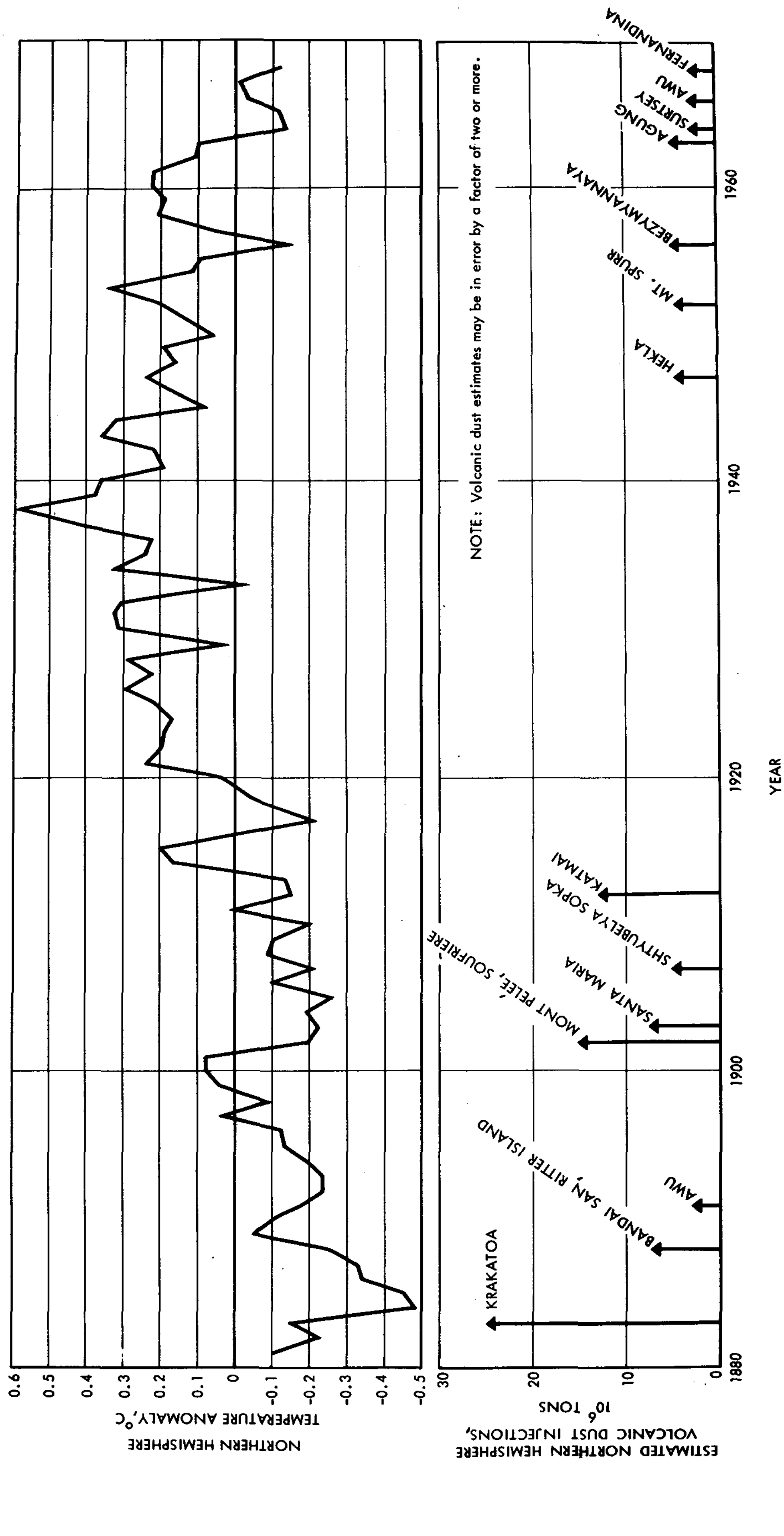

FIG. 1. The climatic record (see text for data source) and principal volcanic dust injections (various sources; cf. Table 1 and Appendix).

TABLE 1. Stratospheric dust injection estimates and related data.

Year	Eruption	Latitude	Mitchell class (10^6 tons)	Deirmendjian[a] (10^6 tons)	Sapper's Class[b]	Lamb[c] total dust (km^3)	Lamb D.V.I.[d] Northern Hemisphere	Megatons assigned in Northern Hemisphere: Event	Megatons assigned in Northern Hemisphere: Year's total
1883	Krakatoa	6°S	100	30	1	6.0	1000	25	25
1885	Falcon Island	20°S	1	—	2 (?)	—	100	0.1	0.1
1886	Tarawera	38.5°S	10	—	1	1.5	(400)[e]	0.5	
1886	Niafu	16°S	1	—	2	—	100	0.2	0.7
1888	Bandai San	38°N	10	—	1	3.0	(250)	5	5.0
1888	Ritter Island	5.5°S	10	—	1	1.7	(125)	2.5	2.5
1890	Bogoslov	54°N	1	—	—	—	(50)	0.5	0.5
1892	Awu	3½°N	10	—	—	—	(100)[e]	2.5	2.5
1898	Una Una	9°S	1	—	3	—	70	0.3	0.3
1902	Mont Pelée	15°N	10	—	2	—	(100)[e]	3.8	
1902	Soufrière	13½°N	10	—	1	1.0	(300)[e]	3.8	
1902	Santa Maria	14½°N	20[f]	—	1	5.4	(600)[e]	7.5	15.1
1903	Santa Maria	14½°N	20[f]	—				7.5	7.5
1904	Minami Iwoshima	24°N	—	—	2 (?)	—	30	0.3	0.3
1907	Shtyubelya Sopka	52°N	10	—	1	3	150	5	5.0
1911	Taal	14°N	1	—	2	—	15	0.4	0.4
1912	Katmai	58°N	10	13.4	1	(0.006–21)	150	13	13
1914	Sakurashima	31.5°N	1	—	2	0.5	20	0.5	0.5
1929	Asama	36.5°N	0.1	—	—	0.1	4	0.1	0.1
1931	Kluchev	56°N	0.1	—	—	0.1	2	0.1	0.1
1947	Hekla	64°N	10	—	—	0.18	(20)	5	5
1951	Mt. Lamington	9°S	—	—	—	—	10	0.5	0.5
1953	Mt. Spurr	61°N	10	—	—	—	2	5	5
1956	Bezymyannaya	56°N	10	—	—	1.0	(10)	5	5
1963–65	Surtsey	63°N	1	—	—	0.7	15	0.5	0.5
1963	Agung	8.5°S	30	9	—	—	400	5	5
1966	Awu	3.5°N	10	—	—	~3	150–200	2.5	2.5
1968	Fernandina	0.5°S	10	—	—	1–2 (?)	50–100	2.5	2.5

[a] Assumes dust in stratosphere at density of 1.0 gm cm^{-3}.
[b] Sapper's scale for explosive eruptions: Magnitude $1 = >10^9$ m^3 ejected, magnitude $2 = 10^8 - 10^9$, etc. Only explosive eruptions are considered here. Source: Lamb, 1970.
[c] Note that Lamb's estimates are for total, rather than stratospheric, dust.
[d] The dust veil index (D.V.I.) is a time- and area-weighted radiation loss parameter, deduced by Lamb in various ways; the values in parentheses are estimates.
[e] World value. No Northern Hemisphere value given.
[f] Timing split according to Mitchell (1974, private communication).

events; a more detailed listing of volcanic events is given in Table 1. The temperature anomaly curve is from a tabulation by Budyko (1969), updated by T. Asakura[3] in 1974, and provided in digital form by J. M. Mitchell, Jr. (private communication, 1974). The validity of this temperature anomaly curve has been questioned (H. E. Landsberg, private communication, 1975), but with minor exception, it is accepted here without further examination.

The volcanic data are from several sources, but are based largely on the tabulation published by Mitchell (1970); however, Mitchell's dust mass estimates have generally been adjusted downward by a factor of 2, based on independent estimates for several of the events (Deirmendjian, 1972; Junge, 1974), and have been further adjusted for latitude of the event as discussed below. The volcanic dust data are discussed briefly in the Appendix. In general, the well-known dust veil index (D.V.I.) concept of Lamb was not utilized in the tabulation in view of its incorporation fo time, dust quantity and area factors (as well as resultant climate change in certain circumstances); for purposes here, dust quantities were of primary interest. The uncertainties in the volcanic dust numbers are large (i.e., a factor of 2 or more). Furthermore, the listing may well be incomplete. The comprehensive work by Macdonald (1972) suggests additional, possibly important events, but no estimates of stratospheric dust injections are provided.

The latitude, strength and probable season of an eruption can all be expected to play a role in determining the effect of the eruption on the mean annual hemispheric temperature anomaly. Dust in the polar night, for example, would be expected to have a quite different effect on the radiation balance than dust in the tropics. The physical phenomena involved are far

[3] Japan Meteorological Agency (unpublished results).

TABLE 2. Assumed latitudinal effects.

Eruption latitude	Northern Hemisphere dust fraction	Year affecting Northern Hemisphere mean temperature
90° N–20° N	1.0	Same
20° N–10° N	0.75	Same
10° N–0°	0.5	Same
0°–10° S	0.5	1 year later
10° S–20° S	0.33	1 year later
20° S–40° S	0.1	1 year later
40° S–90° S	Ignore	

beyond the scope of this paper. Nevertheless, two corrections for latitudinal effects were felt to be necessary and were utilized. These related to the fraction of total dust reaching the Northern Hemisphere and the time delay between the event and its impact on the Northern Hemisphere mean annual temperature anomaly. The assumptions made are shown in Table 2.

These assumptions should be examined and the use of a finer time resolution considered in more detailed work.

Several points should be noted from Fig. 1 and Table 1:

1) The quantity of the dust injected from major volcanic eruptions is indeed very great. The 1883 Krakatoa eruption, for example, apparently put 30–100 million tons of dust into the global stratosphere; for purposes here an initial Northern Hemisphere value of 25 million tons is assigned (see Appendix). Effects were noted essentially worldwide and lasted for several years. Losses in direct solar beam of 20–30% were noted, although about 85% of this (Budyko, 1974; Herman, 1974) is regained due to an increase in scattered light. This event was extensively studied at the time and is described in detail by various authors (Lamb, 1970; Deirmendjian, 1972).

2) A short cooling wave apparently followed the Krakatoa eruption and several of the later eruptions. Other fluctuations, however, showed no evident relation to volcanic events.

3) The overall climatic trend from the 1880's to the 1930's was one of warming. Furthermore, this warming trend apparently took place even in the early part of this period, a time during which a number of major eruptions occurred.

4) A period of several decades existed (~1915–1945) in which volcanic activity was unusually light and, as mentioned earlier, the temperatures were higher than the preceding or, in fact, the subsequent (current) period.

Volcanic eruptions and dust in the stratosphere could affect climate in a variety of ways, including alteration of planetary albedo, heating of the stratosphere by absorption of solar radiation, with alteration of the radiation balance, changes in cloud nuclei (Wexler, 1951), as well as changes in snow albedo due to deposition of tropospheric dust (Landsberg, 1970). Volcanoes put a variety of materials into the stratosphere (e.g., mineral dust, water, CO_2, HCl and SO_2) any of which could contribute to climatic change. The SO_2, however, on conversion to sulfuric acid is believed to be the most important component in a climatic sense.[4] These are all complex effects; the goal here is to estimate empirically the *net* effect. In doing so it is necessary to at least mention other possible causes of climatic effects.

Numerous possible causes of climate change have been discussed in the literature, including both anthropogenic and natural factors. Two principal anthropogenic sources are often considered: changes in atmospheric carbon dioxide and changes in tropospheric dust. These and other factors, such as changes in stratospheric water vapor, ozone, solar constant, sunspots and orbital parameters have been reviewed by various authors (e.g., Mitchell, 1961, 1970, 1975; Reitan, 1971; Dyer, 1974; Lamb, 1972; Bryson, 1974). The possible effects due to changes in CO_2 are perhaps most readily subject to analysis, for good data do exist on atmospheric CO_2 and its increase over recent decades. Thus, according to Reitan (1971), based on calculations by Manabe and Wetherald (1967), the increase in CO_2 between the 1880's and the 1960's could have caused a mean temperature increase of 0.3°C. Unfortunately, however, such computations are based on an assumption of constant cloudiness, and possible changes in cloud cover are exceedingly important. Manabe and Wetherald (1967) show, for example, that a 1% increase in low cloudiness would cause an 0.8°C decrease in mean temperature; thus, a 0.3°C warming could be compensated by a change of about 0.4% in low cloudiness. A change of only 0.4% in low cloudiness would obviously be exceedingly difficult to detect. The influence or even the sign of other possibly significant factors, such as tropospheric dust, is less easy to establish than CO_2, and might well compensate for CO_2 changes. In view of these many uncertainties, no attempt is made here to correct for these other factors. In any event, Mitchell (1975) concluded that neither tropospheric particulates nor atmospheric CO_2, in concert or separately, could have accounted for the major part of the observed temperature changes in the past century.

[4] Friend (1972) gives an "average" analysis for volcanic gases of 95% H_2O, 4% CO_2 and 1% SO_2 by volume for a mass ratio of $H_2O:SO_2$ of 53:1. In CIAP studies (Grobecker *et al.*, 1974) it was reported that the effect of water vapor is about one-half as great as the effect of SO_2 (and of opposite sign) at a mass ratio of 1250:1. Thus, the SO_2 and other dust quantities should generate the dominant effect.

3. The approach

a. Development

The approach was developed following a suggestion by Budyko (1974) that the earth's surface temperature may respond to perturbations in a simple fashion, according to

$$\frac{dT}{dt} = -\lambda(T - T_R). \tag{1}$$

As used here, λ is the climatic response constant (per year) and t is time (years); T and T_R, however, need discussion. In an idealized sense T represents an instantaneous hemispheric, seasonally adjusted mean temperature, and T_R represents the corresponding instantaneous reference temperature toward which T is "moving" with time. The data in Fig. 1, however, represent both a time (annual) and space (Northern Hemisphere) average, and instantaneous data are not extractable therefrom. To proceed, an assumption is necessary: The assumption made is that the mean annual anomaly data (Fig. 1) provide a satisfactory *approximation* to the desired instantaneous anomaly curve. Some justification for this assumption is provided in a later section.

Returning to the procedure development, for conditions following a volcanic eruption T_R clearly varies with time. To account for this it is assumed that

$$T_R = T_0 - \alpha M(t), \tag{2}$$

where T_0 is an (unknown) equilibrium mean temperature in the long-term absence of stratospheric dust, α a cooling coefficient (degrees Celsius per million tons of dust), and $M(t)$ the mass of dust as a function of time.[5]

As Eq. (2) is written, T_0 is implied to be a constant. In fact, T_0 will certainly vary with time since CO_2 and other factors unrelated to volcanic eruption change. The manner in which T_0 has varied with time is unknown, but two limiting assumptions can be made as follows:

- T_0 is constant over a 10- or 20-year period and is estimable from the mean temperature existing after 5 or 10 years of light volcanic activity.
- T_0 is constant over an 80- to 100-year period and is estimable from the mean temperatures prevailing after several decades of very light activity.

As will be seen, the first assumption gives small values for the cooling coefficient and the response time, and the second assumption gives higher values for both. Presumably, "true" values would be between these limiting values, but this cannot be proven to be the case because of the many variables that could affect climate.

The first assumption is used in exploring two periods near the turn of the century; the second is used for the full period (1883–1968). In effect, the second assumption implies either that the lower temperatures prevailing prior to 1883 were due to previous volcanic activity[6] or that an underlying, fairly rapid warming trend began about 1883 with the increase in measured temperature delayed by volcanic activity.

Under both of the above assumptions, T_0 is treated as a constant so that the same mathematical formulation applies. For convenience the quantities θ and β are now defined, where θ is the anomaly relative to the reference line used by Budyko, and β represents a constant difference between this reference line and T_0; i.e.,

$$T(t) - T_0 = \theta(t) - \beta, \tag{3}$$

so that Eq. (1) becomes

$$\frac{dT}{dt} = -\lambda[T - T_0 + \alpha M(t)] \tag{4a}$$

or

$$\frac{d(\theta-\beta)}{dt} = -\lambda[\theta - \beta + \alpha M(t)]. \tag{4b}$$

Eq. (4b) leads to

$$\theta = (\theta_0 - \beta)e^{-\lambda(t-t_0)} + \beta + \int_{t_0}^{t} e^{-\lambda(t-\tau)}[-\lambda\alpha M(\tau)]d\tau, \tag{5}$$

where θ_0 is the value of the temperature anomaly at the beginning of the integration period being considered. To integrate Eq. (5) $M(t)$ is needed; for simplicity, volcanic dust loadings were assumed to follow a relationship of the form

$$M(t) = M_0 e^{-k(t-t_0)}, \tag{6}$$

where M_0 represents the mass of dust in the stratosphere at the beginning (t_0) of the time period in question, and k^{-1} represents a dust-residence time. Substituting Eq. (6) into (5) and integrating gives

$$\theta = (\theta_0 - \beta)e^{-\lambda(t-t_0)} + \beta - \frac{\lambda\alpha M_0}{\lambda - k}[e^{-k(t-t_0)} - e^{-\lambda(t-t_0)}]. \tag{7}$$

Note that either by Eq. (2) or Eq. (7) the steady-state temperature reductions achieved by M_0 tons

[5] In principle, of course, if data on CO_2, H_2O, tropospheric dust, cloudiness, etc., over the course of time were available, and cooling or warming factors were known for each, their effects in the treatment could be included.

[6] For example, Lamb (1972) lists an eruption of Ghaie (4° S, 152° E) in 1878 with a possible D.V.I. 25% greater than that of Krakatoa. It is doubtful, however, that volcanic eruptions were of sufficient frequency or magnitude to depress mean temperatures, prior to the Krakatoa event, to values measured, relative to temperatures in the 1930's.

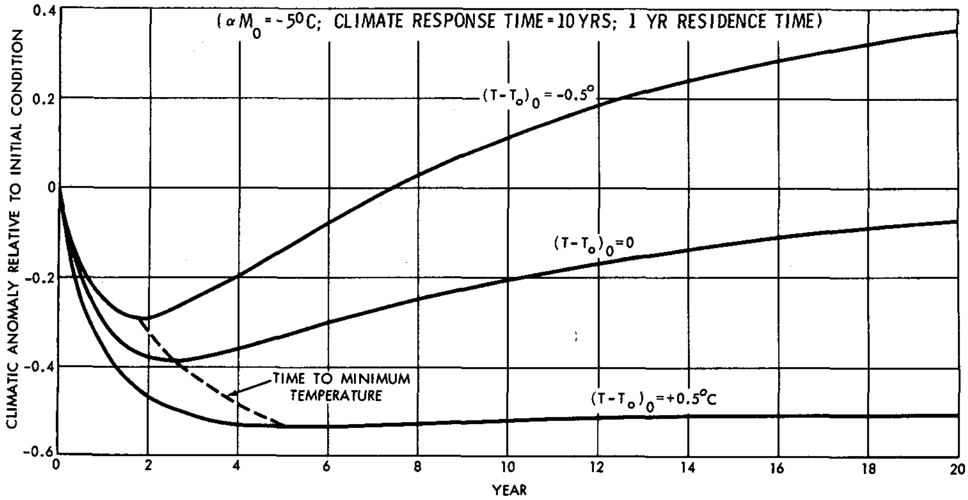

FIG. 2. Apparent climatic response for the same volcanic eruption under three assumptions as to initial temperature anomaly with the same arbitrary value for climatic response time.

maintained continuously would be αM_0. Note also that M_0 is usually estimated from radiation measurements, and that the product αM_0 can be estimated from these measurements and theoretical radiative equilibrium (heat balance) arguments. In one sense, therefore, this work can be considered to be a test (although obviously not a fully satisfactory one) of these theoretical arguments.

Again, for simplicity, in order to use Eq. (7) for sequential events, it was assumed that all dust injections could be characterized by the same value of k, i.e., by the same residence time (k^{-1}). With this assumption, Eq. (7) was applied on a yearly step basis, taking M_0 at the beginning of each time step as the sum of residual old dust and new dust added during the year. Treating residence time values as reasonably known quantities, empirical values of λ and α were sought to obtain a fit to the temperature record (Fig. 1). In most cases the computation was begun at the 1883 point and the analytical curve was forced, by selection of α for an assumed λ, to pass from the $-0.14°C$ point in 1883 through a minimum value of $-0.48°C$, which the data show was reached a year or so later. Dust estimates were used from Table 1. The only exception to this general procedure was made in a study of the period 1901–20 (described later).

For convenience in the following discussion, the terms "climatic response time" C (equal to λ^{-1}) and "dust residence time" R (equal to k^{-1}), both in years, are utilized.

b. *Some characteristics of the formulation*

Before proceeding, it is of interest to note the manner in which the initial anomaly ($\theta_0-\beta$) in Eq. (7) affects the relative shape of the climatic response to a given volcanic event. The point is illustrated in Fig. 2 for an arbitrary set of assumptions. In Fig. 2 all curves are normalized to start at the same point. Note that where the initial anomaly is negative (i.e., the climate without volcanic eruptions is in a warming trend[7]) the downward displacement is much less than if no trend exists or if a cooling trend is under way. Note also that the time to minimum temperature is reduced in the presence of a warming trend. With a small eruption and a strong warming trend it is conceivable that no cooling below the initial value might be noted; similarly, with a strong cooling trend no recovery might result, and the mean temperature might simply drop to the new equilibrium value, accelerated in the process by the volcanic eruption.

Another point of interest with regard to Eq. (7) is that the time-integrated cooling effect of a given eruption (from t_0 to ∞) is simply $\alpha M_0/k$, and is independent of λ. It is evident then that the residence time (k^{-1} or R) is fully as important as M_0 in determining the effects of a given eruption. [The Lamb (1970) D.V.I. incorporates this concept, but not in a manner directly usable here.] The uncertainties introduced by the assumption of a constant residence time for all eruptions are discussed in a later section; clearly, however, the assumption will overstate the effects of some eruptions and understate the effects of others. The effects of variation in λ are illustrated below.

c. *Preliminary utilization of the formulation—the need for an empirical evaluation of constants specific to this formulation*

The formulation shown in Eq. (7) involves several constants, for all of which it might appear that estimates are available from the literature or by reasonable

[7] Note that these "trends" take place with a constant value of T_0; the temperature at the beginning of the period is simply assumed to be displaced from its equilibrium value.

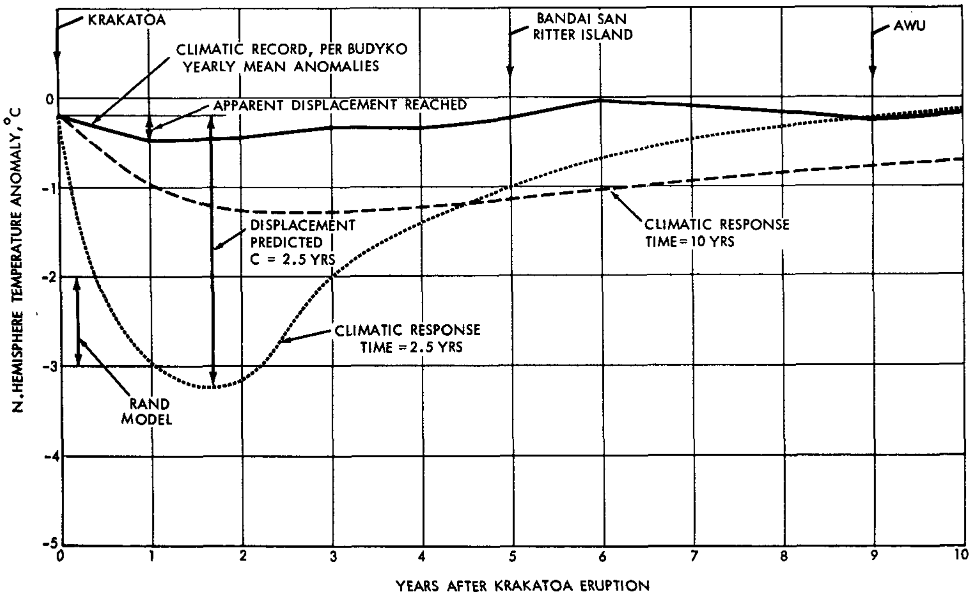

FIG. 3. Climatic record and certain analytical results, treating the Krakatoa eruption as an isolated event and ignoring any underlying trends. The Bandai San and Awu events are not included.

assumptions. As a preliminary exercise, it is of interest to see how well these values might apply to the period following the Krakatoa eruption in 1883. Thus, based on the seasonal march of temperatures, Budyko (1974) developed a value of λ of 0.4 year^{-1}, or a 2.5-year response time. Mitchell (1970) assumed a residence time for stratospheric dust of 14 months, but considerable variability is involved (Mitchell, 1961); a figure of 1 year would seem to be reasonable here. The value of β over a short period might reasonably be taken as zero.

As discussed later, a value of α of 0.53°C per megaton (Mt) can be derived from results of the CIAP effort, as reported by Grobecker *et al.* (1974); this figure is for a hypothesized 10-year, steady-state loading of SO_2-derived stratospheric dust. [Note, however, that CIAP studies also recommend that about one-third of this value should be used for 1-year changes (see Hidalgo, 1974, p. F-122). This smaller figure, however, includes the effects of non-equilibrium response.]

The above constants and an M_0 value of 25 million tons were used with Eq. (7) to develop the dotted curve in Fig. 3; the dashed curve was developed for the same values of α and M_0, but it assumes a 10-year response time. Data (arrow) from a Rand (Mintz-Arakawa) model run (Batten, 1974) that simulates a Krakatoa cloud, which did not include oceanic thermal inertia, are also indicated as a matter of interest. Note that the computational results show far larger displacements than do the actual data. With a climatic response time of 10 years, the displacement is greatly reduced but the duration of the effect is greatly increased. However, the total time-integrated effect as noted above would be unchanged by changes in λ.

A number of factors could enter into the apparent poor agreement in Fig. 3 between computed and measured changes in mean temperature. It appears, however, that an alternative empirical approach may be useful in finding what values of these constants would be required to fit the historical data. Given these data, the empirical and theoretical results will then be reexamined.

At this point it is worth noting that the theoretical curve in Fig. 3, even for a climatic response time of 2.5 years, passes through a broad minimum of the order of a year or more in length. The computed curve, of course, represents idealized instantaneous temperature data; however, a yearly *mean* anomaly for the second year would be close to the theoretical minimum instantaneous value. For longer response times the minimum is broader. In the empirical work which follows, constants are chosen to force the theoretical instantaneous curve through certain measured mean *annual* anomalies; the above observations provide some justification for this approach. Obviously, the exact timing of the averaging period is significant in this regard, but the time resolution used in the following is too coarse for this point to be pursued.

4. Results and discussion

a. Results from short-period studies—possible long-term trends ignored

Two short-period (~20 years) cases were examined to determine empirical constants. In these cases, possible underlying long-term trends were ignored.

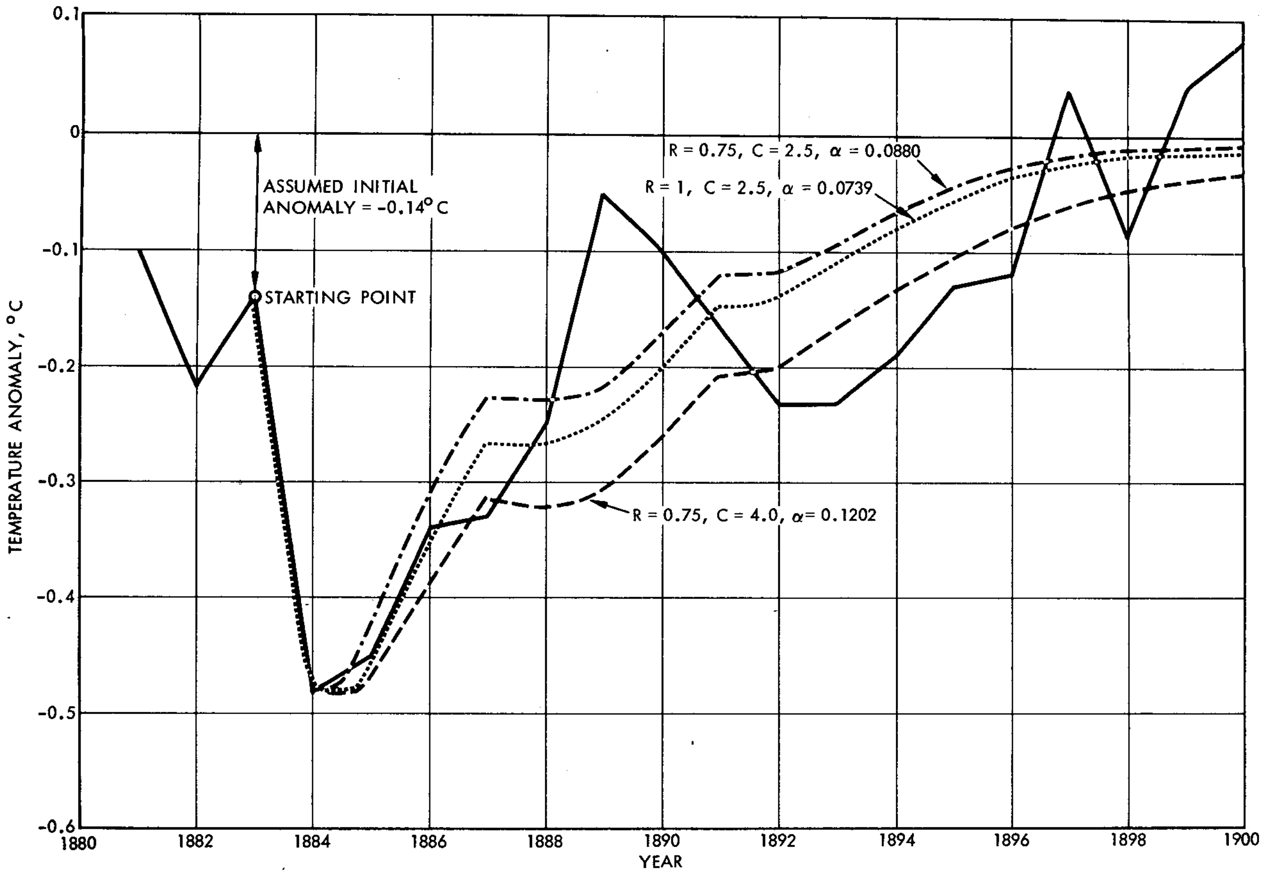

FIG. 4. Climatic record and model results using constants shown. Initial anomaly set at −0.14°C. No other trends. Period 1883–1900.

1) THE 1883–1900 PERIOD

The first set of results is given in Fig. 4 in which analytical curves with several sets of constants are shown. In these cases the starting point was as shown, and the zero anomaly reference value was taken as the appropriate value for T_0 (i.e., β was set equal to zero). As noted earlier, values of α were selected to force the analytical curves through a minimum at −0.48°C; the precise time at which this occurred was allowed to "float". Note that a reasonable fit is given by a climatic response time of 2.5 years, a residence time of 1 year and an α of 0.0739 (an α considerably smaller than that used in Fig. 3).

Note that the assumption of shorter residence times leads to larger values for α. By inspection of the three curves, it is evident that a climatic response time of about 3 years and an α of about 0.1 would, with a residence time of 0.75 year, about duplicate the middle curve. It is clear, therefore, that one of the variables in Eq. (7) must be taken as a known value to avoid multiple apparent solutions. Here residence time is taken as a known value, normally 1 year.

2) THE 1901–1920 PERIOD

For the 1901–20 short-term case, β was taken as +0.08°C so that the initial anomaly $(\theta_0-\beta)$ was zero. (A step change in T_0 of 0.08°C from the previous plot is implied.) In this case the curves were forced through −0.24°C by selection of α. Here a climatic response time of 2.5 years does not give nearly as good a fit as one of 5 years, at least in the period before the Katmai eruption in 1912. The large swing in temperature following the Katmai eruption makes any comparison in this latter period rather suspect.

Note that doubling the climatic response time from 2.5 to 5 years leads, in this case, to a 50% increase in the value of α, an effect that is opposite to that of increasing the residence time.

As discussed in the approach, these figures should provide lower bound estimates of the cooling effects, in that warming trends are essentially ignored. Note, for example, in Fig. 5 that the addition of a warming trend to the analytical result would raise the analytical curves in the 1906–10 period above the values shown; a larger cooling coefficient and a longer response time would then be required to force the curves through the

TABLE 3. Constants used in Figs. 6–8.

Figure no.	β (°C)	$(T-T_0)_{1883}$ (°C)	R (years)	C (years)	α (°C Mt^{-1})
6	0.36	−0.5	1.0	10	0.227
7	0.36	−0.5	1.0	9	0.211
8	0.26	−0.4	1.0	6.5	0.162

recorded data. These facts will become evident in the next section.

b. *Full period results*

1) PRINCIPAL RESULTS

In this case T_0 in 1883 was based on temperatures being approached in the 1930's. Values of β of 0.26 and 0.36 ($\theta_0-\beta$ or $T-T_0$ of −0.4 or −0.5 in 1883) were used.

Several results are shown in Figs. 6, 7 and 8. Again, the curves are the result of stepwise year-by-year application of Eq. (7), using dust masses as in Table 1 with events adjusted in a time sense according to Table 2. The constants used in Figs. 6–8 are given in Table 3.

Note that the analytical results all seem to follow the full curve in a general way, although deviating somewhat in recent years in a quantitative sense. Note also that the C and α values are about twice those found previously.

2) EXPLORATORY OPTIMIZATION STUDIES

The question of preferred constants, even under the assumptions made in this long-term case, is a difficult one and merits some discussion. Note first in Fig. 6 the large deviation between the analytical curve and the record in the 1885–90 period, unmatched by any offsetting deviation later. This effect is less pronounced with a shorter climatic response time in Fig. 7, and is even better balanced out with the set of constants shown in Fig. 8. The post-1945 period is not matched well with any of the assumptions: either a new trend was under way (perhaps the most likely explanation) or the masses of dust used were too small (a point that is discussed shortly).

To reduce subjectivism in the selection of preferred curves, but without attempting a full multidimensional optimization, the root-mean-square (rms) deviation was determined for a number of cases, as shown in Table 4.

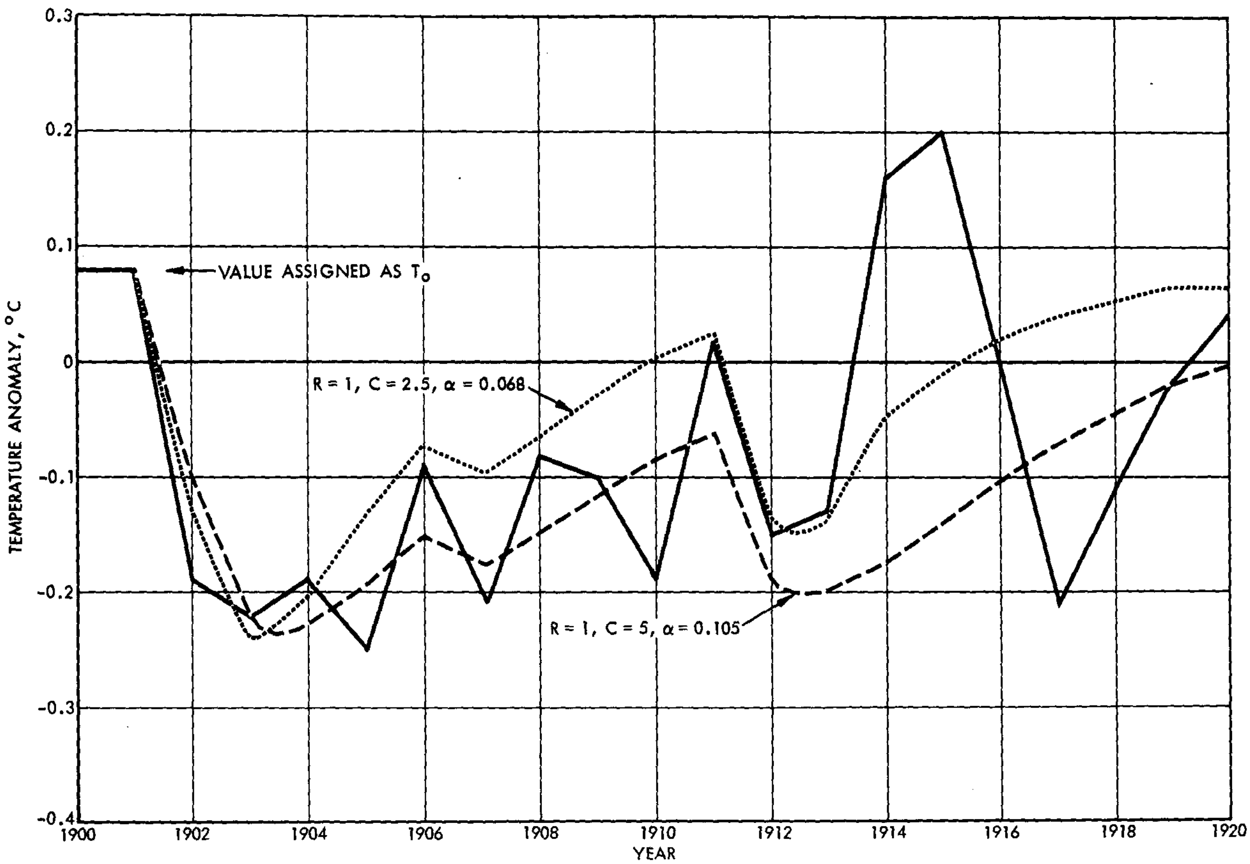

FIG. 5. As in Fig. 4. Underlying trends not considered. Period 1900–1920.

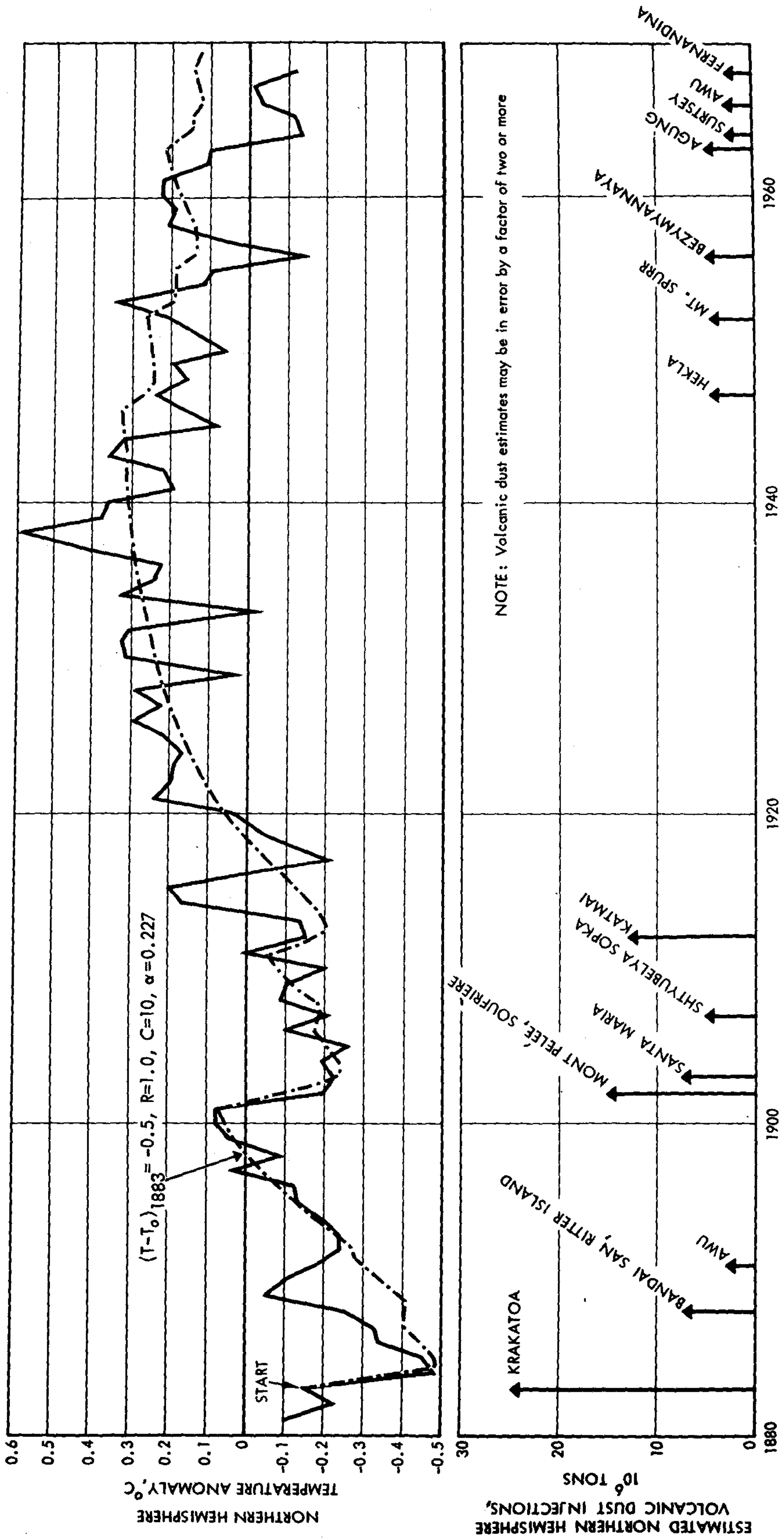

FIG. 6. Analytical result, using $(T-T_0)_{1883}=-0.5$, $R=1.0$, $C=10$ and $\alpha=0.227$, superimposed on the climatic record.

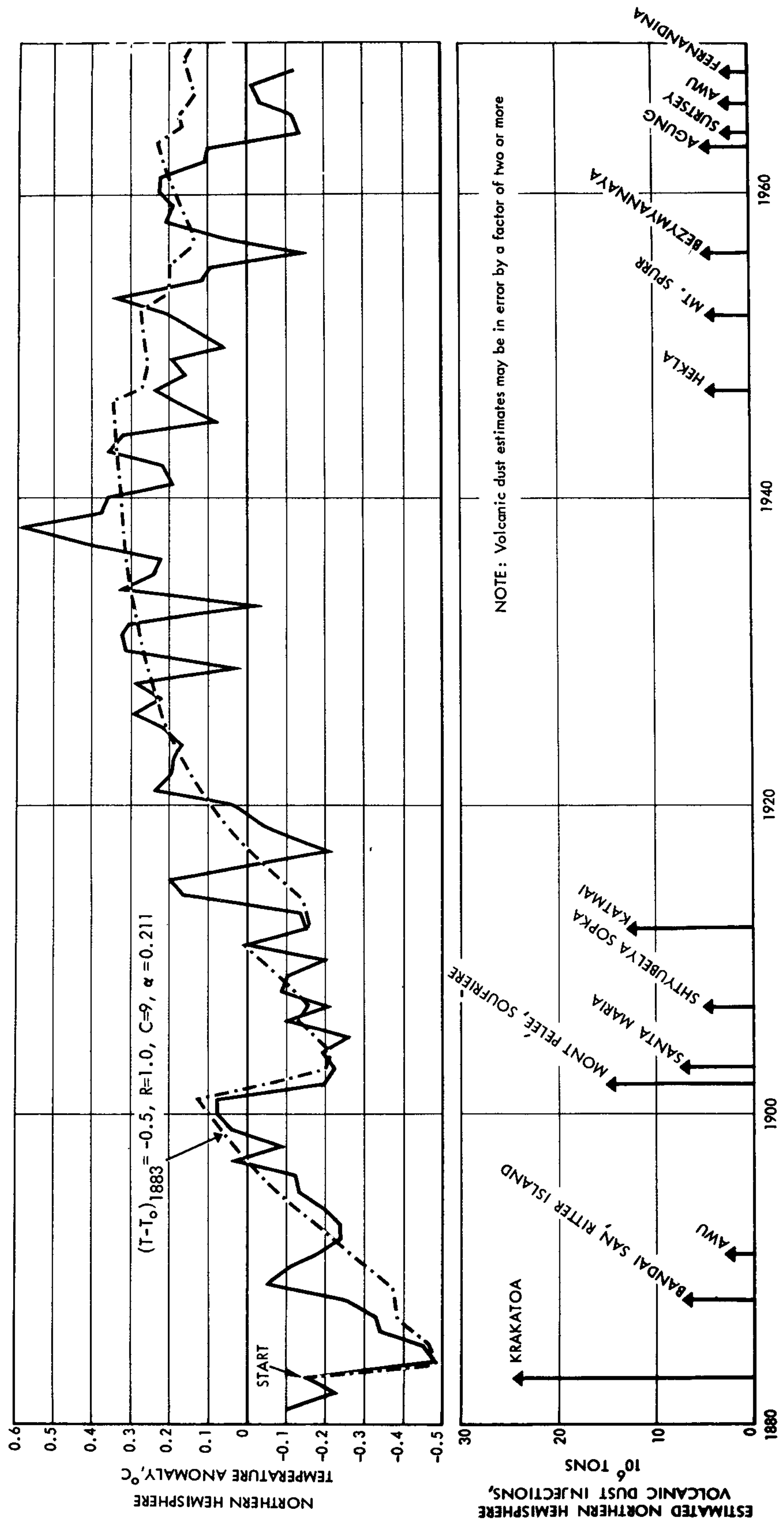

FIG. 7. As in Fig. 6 except for $(T-T_0)_{1883}=-0.5$, $R=1.0$, $C=9$ and $\alpha=0.211$. The effect of a shorter climatic response time can be seen by comparing this figure to Fig. 6.

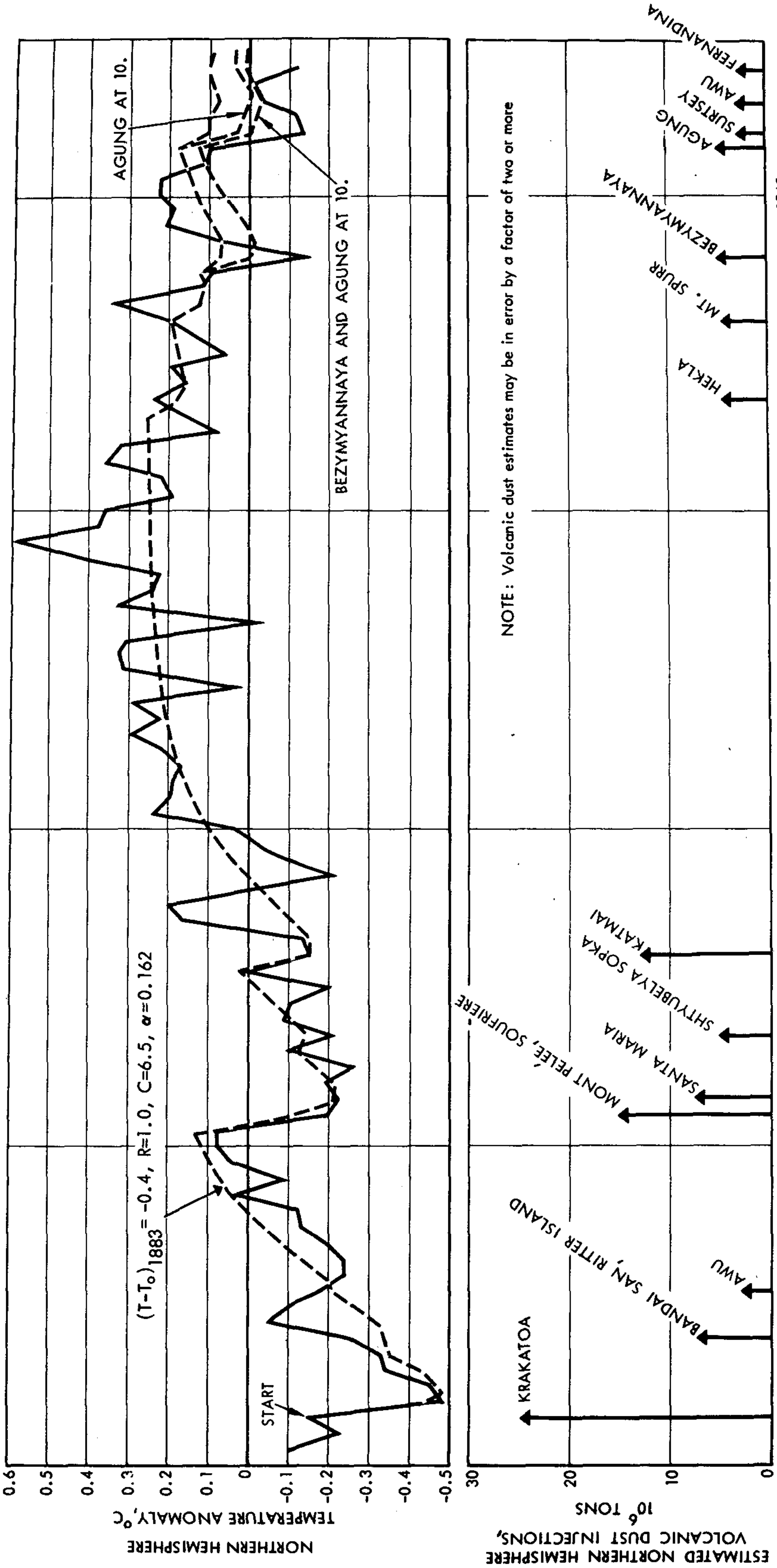

FIG. 8. As in Fig. 6 except $(T-T_0)_{1888}=-0.4$, $R=1.0$, $C=6.5$ and $\alpha=0.162$. The assumed initial temperature anomaly is smaller ($-0.4°C$) than in Fig. 7; a shorter climatic response time and smaller cooling coefficient result.

TABLE 4. Some exploratory optimization results.

$(T-T_0)_{1883}$ (°C)	Constants R	C	α	rms deviation (°C) 1883–1968	1883–1942	Plotted as
−0.5	1.0	8	0.1949	0.1363	0.1232	
−0.5	1.0	9	0.2109	0.1325	0.1202	Fig. 7
−0.5	1.0	10	0.2267	0.1332	0.1242	Fig. 6
−0.5	0.9	10	0.2441	0.1319	0.1208	
−0.4	1.0	6	0.1537	0.1233	0.1233	
−0.4	1.0	6.5	0.1619	0.1220	0.1220	Fig. 8
−0.4	1.0	7	0.1700	0.1222	0.1228	
−0.4	1.25	5.5	0.1269	0.1242	0.1250	
−0.4	1.25	6	0.1337	0.1233	0.1243	
−0.4	1.25	6.5	0.1405	0.1240	0.1257	

Several points should be noted:

- As is visually apparent on comparing Figs. 6 and 8, if the entire 1883–1968 period is to be fitted, a value of $(T-T_0)_{1883}$ smaller than −0.5°C (i.e., about −0.4°C) gives a smaller standard deviation than does the curve shown in Fig. 6, since the deviations in the latter half of the period are reduced considerably. Also, the constants in Fig. 8 give a better fit than those in Fig. 7.
- These data show that the best fit is a function of the time period selected. Thus, the minimum in the tabulated values of rms deviation for the full period is with $(T-T_0)_{1883}=-0.4$°C and with a 6.5-year climatic response time; however, for the shorter period, the minimum is with $(T-T_0)_{1883}=-0.5$°C and a climatic response time of about 9 years (Fig. 7). The best overall fit among the cases studied appears to be the case plotted in Fig. 8.
- In general it was found that as the assumed absolute magnitude of $(T-T_0)_{1883}$ is increased, the optimum climatic response time increases, as does the proper value of α to force the curve through the −0.48°C minimum selected. This can be seen in the α column in Table 3. Also, as was noted earlier, if climatic response time is held constant and R is varied, α increases with decreasing R.

3) SENSITIVITY AND UNCERTAINTIES

(i) Sensitivity to dust masses (post-1945 period). In view of the deviation of the analytical results for post-1945 from the climatic curve, it was of interest to explore briefly the sensitivity of the analytical results to the dust injection numbers used from this period. Two following arbitrary cases were considered: the Agung value was doubled from 5 to 10 million tons, and the Agung and the Bezymyannaya values were both doubled from 5 to 10 million tons.

Results are shown in Fig. 8. In both cases, considerable change is noted in the analytical curve, as would be expected. It seems clear that, if the formulation is valid and if by chance *all* or most post-1945 eruptions have been underestimated relative to those near the turn of the last century by a factor of about 2, i.e., by factors that are probably within the uncertainty, then the post-1945 analytical result would show a greater displacement than does the measured curve [see, however, subsection (ii) below]. This observation suggests a need for a more detailed examination of the events or of dust masses in the stratosphere during this period. Note that fair precision is needed since a factor of 2 is evidently unacceptable.

(ii) Uncertainties in residence times. It was mentioned in an earlier section that dust residence times are as important as initial dust masses in determining total cooling effects, and that the use of a constant residence time for all eruptions overstates the effects of some eruptions and understates the effects of others. In effect, since smaller eruptions would be expected to have shorter residence times than those of larger eruptions, this work may well overstate the effects of small eruptions. Also, if latitude plays a part in residence times, as may well be the case, with high-latitude eruptions possibly having lesser residence times (Mitchell, 1961) than those of tropical eruptions, then the effects of small northern latitude eruptions may be overstated on two counts. These points need further study.

(iii) Uncertainty and sensitivity to the climate record. A ground rule of this paper, as noted at the outset, was that the temperature record would not be questioned in detail. The absolute accuracy of this curve is unknown. Nevertheless, a few comments may be in order.

First, it may be noted that extreme short-term positive swings seem to be balanced by short-term negative swings. These fluctuations may arise from the use of land-dominated averages, i.e., from too few stations to obtain a valid hemispheric average. For changes of the magnitudes shown to be valid, either strong (and unknown) perturbations must occur or thermal inertia (response time) estimates would seem to be in question.

Second, it was observed earlier that all full-term computations, as well as the short-term computations from 1883–1900, involved starting at −0.14°C with a forced minimum at −0.48°C. If this difference of −0.34°C had been in serious error, however, the subsequent curve fits would have been poor. However, the sensitivity of the results to errors in these two data points was given some study by considering two possibilities:

1. The 1883 initial point was erroneously low by 0.1°C, but all other points were correct.
2. The 1883 point and the minimum temperature reached following the Krakatoa eruption were both erroneously low by 0.1°C.

In case 1, the effect was to increase α, inasmuch as the Krakatoa eruption under this assumption caused a cooling of 0.44°C instead of 0.34°C. Small effects were noted on optimum response time, etc., but the predominant effect was to create a corresponding excess displacement in the 1901–03 period. The larger α, of course, made the post-1945 analytical curve look better relative to the data measured.

In case 2, the displacement created by Krakatoa was still taken as 0.34°C, so α was not strongly affected. Optimum response times were increased by a year or so, but no dramatic effect was evident.

These brief examples suggest that the overall results are not extremely sensitive to absolute uncertainties in the temperature anomaly record near the initiation period.

(iv) Other uncertainties. Uncertainties obviously exist in all facets of this work, ranging from questions as to the basic validity of the formulation used to questions of dust masses used, volcanic eruptions not recognized, residence times and differences between eruptions, etc. Analysis of all these factors has not been attempted; the justification is, of course, that this paper is intentionally exploratory.

Two points should be noted with regard to dust masses and the values found for the cooling coefficient. First, what is actually used in the computations is the product αM_0, which is in units of temperature. The entire work could have been done in some unit of this type with different eruptions rated in terms of each other or in radiation loss, etc. (a point that is discussed later). Obviously, if the dust masses used are, say, twofold high, then the α figures are twofold low. Second, the density question should be noted. Optical effects of a given mass of dust of given particle size and properties are inversely related to particle density. Deirmendjian (1972) gives masses in unit density; other authors do not quote the densities being used. Unit densities may be suitable for porous tephra particles; sulfuric acid particles, however, which would be expected to have the longer lasting effects, have densities near 1.7 gm cm^{-3}.

5. Comparison of results

a. Summarized results

The results from the two approaches, as just described, lead to the following two sets of results.

Period considered	α (°C Mt^{-1})	C (years)
Short term	0.07–0.10	2.5–5
Long term	0.16–0.21	6–9

It is important to note that these results were developed (and should always be taken) in pairs, i.e., a small value of α should be used with a short response time, and a larger value of α with a longer response time.

As pointed out earlier, the two assumptions on which these figures were developed may represent limiting values in that in the short-term case, long-term trends are ignored, and in the long-term case, the effects of long-term trends are probably overstated. The question, of course, arises as to which set may be more realistic; and this question cannot be answered absolutely. Insight may, however, be gained by comparing these sets of numbers to results of other investigators. These arguments will favor the larger numbers.

b. Cooling coefficients

The α results can be compared to theoretical values developed under the CIAP program on the basis of radiation calculations on the effects of added dust, and on resultant theoretical temperature changes. One value of α was used for illustrative purposes in Fig. 3. Other values from various authors' work are reported by Hidalgo (1974). These results indicate the change in equivalent solar constant ($\delta\sigma/\sigma$) assuming a uniformly distributed aerosol over the Northern Hemisphere in a 10 km thick layer; this change in equivalent solar constant is then converted to an estimated temperature change if maintained for certain time periods. The equivalent solar constant change computations are made assuming an average solar zenith angle of 60°, which doubles the normal value of optical depth. Given the change in effective solar constant, the change in steady-state temperature is obtained, using an estimated χ proportionality constant according to

$$\Delta T = \chi \frac{\delta\sigma}{\sigma}, \tag{8}$$

where ΔT is the temperature change (°C) after specified time periods, and $\delta\sigma/\sigma$ the fractional effective change in solar constant.

A value of χ of 150°C per percent change in equivalent solar constant is recommended by Leith *et al.*

TABLE 5. Comparison of results.*

Author	ω	β	$-\delta\sigma/\sigma$	α	Remarks
Grobecker	—	—	$0.009\ \Delta S_d$	0.53	**
Herman†	1.0	0.19	$0.010\ \Delta S_d$	0.59	$\lambda = 0.5\ \mu m$ (nonabsorbing)
Herman†	0.9	0.18	$0.016\ \Delta S_d$	0.94	
Pollack-Toon†	0.99	0.12	$0.0061\ \Delta S_d$	0.36	Wavelength integration
Coakley-Schneider†	1.0	0.10	$0.0046\ \Delta S_d$	0.27	Wavelength integration
Luther†	0.95	0.095	$0.011\ \Delta S_d$	0.65	Wavelength integration
This work				0.07–0.23	

* The variations in theoretical numbers include differences in assumed particle-size distribution and density, as well as in computational technique.
** As used in Grobecker *et al.* (1974, p. 43).
† Taken from Hidalgo (1974).

[as reported by Hidalgo (1974)] and by Budyko[8] for time periods of perhaps a decade; a value of 250 that incorporates ice albedo feedback effects is recommended for a time period of the order of 1000 years. A smaller value (50) is recommended for a time period of one year. The different values are in recognition of thermal inertia and feedback effects. While the empirical and theoretical approaches are not strictly comparable, it would appear that the appropriate value for comparison here is the quasi-steady-state value of 150 (see also subsection 5c).

The various workers' results are summarized in Table 5 using a χ value of 150 and converting units. As before, the term ΔS_d in Table 5 refers to a hypothetical particulate concentration, assuming uniform dispersal of added aerosol over a 10 km thick band; the units are $\mu g\ m^{-3}$. The value of α is obtained from χ, $\delta\sigma/\sigma$, and the stratospheric volume involved (2.55×10^{18} m^3), correcting for the units employed.[9]

Table 5 includes reference to values of ω and β which describe assumed optical properties of the particles. The term $(1-\omega)$ defines the probability of absorption and $\beta\omega$ the probability of backscattering. It is now believed (H. Hidalgo, private communication, 1975) that ω for sulfuric acid particles is very near to 1.0; values of 0.9, as one value used by Herman in studying effects parametrically or 0.95 as used by Luther, are too low and lead to values of $\delta\sigma/\sigma$ and thus α which are too high. Thus, the CIAP most-applicable values can be quoted as being about 0.27 to 0.36 as compared to values found here of about 0.07 to 0.2. The midpoint of the empirical values (if taken as 0.14 ± 0.07) is clearly within a factor of about 2–3 of the theoretical values using a χ value of 150, and likely within the uncertainties in the dust masses. Also, quite obviously, the theoretical results themselves involve a number of uncertainties; longwave effects, which reduce the values by 10–25%, are not included and, of course, the value of χ can be questioned. [Grobecker *et al.* (1974) give a range of values that in terms of α would be expressed as 0.5 ± 0.7.] Also, it should be remembered that the α value developed here includes the net effect of all the material injected by volcanic eruptions, including water vapor and perhaps certain feedback effects (such as changes in cloudiness or tropical tropopause warming with resultant water vapor enhancement) which could cause some countervailing warming effects. Thus, the empirical value might well be somewhat lower than the theoretical values that consider aerosols only, all other factors remaining constant.

This comparison can also be phrased somewhat differently. Thus, it was noted earlier that the term αM_0, by definition, represents a steady-state change in temperature if M_0 were to be maintained indefinitely (decades) at a constant level. Also, as stated earlier, stratospheric dust masses have, in fact, been estimated from the optical properties of aerosol clouds, and with the aid of the same theoretical arguments these radiation loss measurements can be used to estimate the product αM_0. Thus, in the case of the Krakatoa eruption, direct beam losses of 20–30% were measured, even after the dust was well dispersed. If it is assumed, following Budyko (1974), that net losses were about 15% of the direct losses, the fractional net loss received at the ground would have been 0.03–0.045. With a χ value of 150 as above, the Krakatoa dust cloud (25 Mt assumed as the initial Northern Hemispheric loading), if maintained indefinitely, should have by these arguments resulted in a 4.5–6.75°C[10] loss in surface temperature. By comparison, the equivalent steady-state loss with the short-term results for this event would be $(0.07\text{–}0.10)(25) = 1.75\text{–}2.5°C$, and for

[8] See footnote 2.

[9] Thus, in the first case in Table 5,

$$\alpha = \frac{150\times 0.009[°C\ (\mu g\ m^{-3})^{-1}]}{2.55[Mt\ (\mu g\ m^{-3})^{-1}]} = 0.53°C\ Mt^{-1}.$$

[10] In fact, figures twice these values could be developed by CIAP methodology, if the 20–30% loss were assumed to represent losses with a direct overhead sun rather than losses at the average solar zenith angle of 60°. However, as the radiation loss data were taken in northern latitudes [30°–60°N, Lamb (1972)], the approach taken above seems more appropriate.

the long-term case $(0.16\text{–}0.21)(25)=4.0\text{–}5.25°C$. This argument would favor the larger pair of numbers, assuming, of course, that the χ value of 150 is correct.

It should also be noted that possible errors introduced by questions about mean annual anomalies versus instantaneous anomalies would be more significant for presumed short response times (see Fig. 3). On this basis, the smaller figures are more suspect.

As a separate matter, it might be noted that a cooling coefficient, but of a completely different sort, can be developed from a plot given by Mitchell (1970) noting stratospheric volcanic loadings versus time and the equivalent cooling for each event. This coefficient, which can be converted to about 0.02°C per megaton of stratospheric dust, relates to the short-term climatic displacement following an eruption and not the steady-state value. An analogous value based on the Krakatoa data used here would be 0.34°C per 25 megatons or about 0.014°C Mt^{-1}.

c. Climatic response times

Response times in the sense used here are not treated explicitly in the studies on which the previous comparison was made; rather, responses as a function of time were given. The first two χ values (50 at 1 year, 150 at 10 years), however, are fitted by $\chi=152.9(1-e^{-0.396t})$, implying a short-term response time (1/0.396) of about 2.5 years which is in agreement with Budyko who estimated response time based on the seasonal march of temperatures. A figure in this range was found to fit short-term events (Figs. 4 and 5) reasonably well, but this comparison should be treated with caution. Budyko refers to a response time for less-than-yearly changes, whereas here the response time refers to longer term changes.

A response time comparable to that used here was developed by Reitan (1971), intercomparing 5-year periods, in attempting to develop estimates of the relative importance of volcanic dust in the stratosphere, tropospheric dust and carbon dioxide. He found, after some study, that a climatic half-response time [as in Eq. (1), but not carried through as in Eq. (7)] of 5 years was most appropriate. A half-time of 5 years corresponds to an e-folding time of 5/0.693 = 7.2 years, very similar to the 6.5-year value found in this work in the full-period study.

d. Effect of trends and other pitfalls

Values of α, k and λ alone, even though precisely known, would be insufficient to predict the "effect" of a given volcanic eruption or of any other specified perturbation, unless underlying trends are understood. This point was made in Fig. 2 and represented part of the wide deviation of "predicted" from measured values in Fig. 3. Other reasons for the deviation shown there can be postulated, including (if the work here is accepted) too large a value for α (or χ). An important point, however, is that use of a large value of α(0.5) with a short response time (2.5 years) as was done here for illustrative purposes in one of the curves in Fig. 3, would seem never to be appropriate.

6. Concluding comments

First, there is qualitative compatibility between empirically derived steady-state cooling effects of stratospheric aerosols and theoretically derived values. Quantitatively the theoretical calculations imply a greater cooling[11] per unit mass than do the empirical results developed here, but the theoretical estimates involve a range of values, and there are obvious uncertainties in the empirical approach.

Second, the results, taken together, are consistent with past arguments that large volcanic eruptions cause short-term climatic cooling. The effects of a given eruption are moderated by the thermal inertia of the earth-ocean-atmosphere system; with repeated eruptions the effects would be increased. However, much more detailed examination of the phenomena and data involved, using finer temporal and spatial resolution, would be required to establish such effects unequivocally. Underlying temperature trends would need to be recognized in any such evaluation: if the climate happened to be in a warming trend at the time of an eruption, the eruption might merely slow the process, and no cooling effect might be detectable.

Finally, the time-integrated aspects of climate changes should be noted. Based on this work, it is suggested that long-term records of stratospheric dust should be developed and maintained. However, if theory is eventually to be tied to observation, accurate long-term records will be needed of all the known

[11] Three papers published subsequent to the preparation of this paper should be noted. Of particular interest (see Pollack *et al.*, 1976: Estimates of the climatic impact of aerosols produced by space shuttles, SST's and other high flying aircraft. *J. Appl. Meteor.*, **15**, 247–258) is a recent theoretical assessment of cooling effects of stratospheric sulfuric acid aerosols on climate. The authors give figures which, converted to units used here, correspond to an α value of 0.1 °C Mt^{-1}, in good agreement with the short-term values derived here. A second theoretical paper by much the same group (see Pollack *et al.*, 1976: Volcanic explosions and climate change: A theoretical assessment. *J. Geophys. Res.*, **81**, 1071–1083) considered the effects of volcanic eruptions; they concluded that volcanic explosions cause cooling and that continued activity, as in the 1880's, can cause cooling effects of 1 K to be realized. In a third paper (Schneider, S. H., and C. Mass, 1975: Volcanic dust, sunspots and temperature trends. *Science*, **190**, 741–746), a strong influence of solar variability, as correlated with sunspot number, was indicated. A model calculation beginning in 1600, incorporating solar variability and stratospheric dust (Lamb's D.V.I.) resulted in a curve showing similarities to known temperature histories. However, the model results for the period examined here showed larger oscillations than do the Budyko data and some disagreements as to the times at which intermediate minima and maxima occurred. The thermal inertia used by Schneider and Mass was based on a 75 m mixed ocean layer, which corresponds to a 6.9-year relaxation time.

factors, of which stratospheric dust is certainly only one, that may control climate.

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APPENDIX

Comments on Certain Volcanic Eruptions

The large eruption of Krakatoa, on which much of the discussion in this paper is based, took place 26–27 August 1883 at 6° S, 105½° E, although smaller explosive phases started 20 May 1883 (Lamb, 1970). Estimates of the amount of solid matter blown up range from 6 to 18 km³ (Lamb, 1970). Lamb suggests that 6 km³ ($\sim 10^{10}$ tons) and Deirmendjian (1972) that 4 km³ were dispersed as dust in the atmosphere; only a small portion, however, went into the stratosphere [1/100 used by Mitchell (1970); 1/600 according to Deirmendjian (1972); recognized but not given by Lamb]. Losses of 20–30% in the direct solar beam were noted essentially worldwide, occurring as long as 2–3 years after the eruption; much of this loss [about 85% (Budyko, 1974)] was made up by an increase in the diffuse radiation.

Several estimates of the stratospheric dust loading from this event are available (although it is not clear as to what degree they are independent) as follows:

Author	Global stratospheric burden
Mitchell (1970)	10^8 tons initially; however, only half this value was assumed to be effective the first year (50 million tons)
Junge (1974), based on Volz (1972)	32×10^6 tons, global, 1 year after the event; apparently roughly the same in both hemispheres
Deirmendjian (1972)	30×10^6 tons, if unit density, computed from optical criteria (a "few months" after the event)

Portions of the dust reached great heights (>27 km) according to Lamb (1970); however, the mean stratospheric cloud height must have been much less. Mitchell (1970) ascribes a residence time of 14 months to the dust so injected. Junge (1974), quoting Volz (1972), gives a worldwide loading of 32 Mt 1 year after the event; correcting for an assumed exponential decay (14 months' residence time), an initial figure of 75 Mt results. Deirmendjian, however, assuming unit density, estimates 30 Mt. Correcting for say 3 months' time this would imply 37 Mt initially, while if a density of say 1.7 gm cm^{-3} is used, the corrected initial value would be 63 Mt. Presumably somewhat less than half these global figures should be used for Northern Hemisphere loadings; in fact, much less if the Agung experience (see below) is representative. However, based on the fact that Junge (1974) implies substantially equal radiation loss values in both hemispheres a year after the event, the split may have been almost equal. Any assigned value involves some arbitrariness; a Northern Hemispheric value of 25 Mt was selected (half of Mitchell's first-year global value), decaying directly from this value thereafter. This figure was then used in prorating subsequent events, using Mitchell's tabulation of magnitudes (Table 1), except as noted below for Katmai and Agung.

Two small (0.5 Mt each) Northern Hemispheric events also took place in 1883. However, by the assumptions of this work (Table 2), they were assumed to have already affected the 1883 starting point climatic anomaly and were ignored in comparison to Krakatoa for the subsequent period.

No independent studies of the large eruptions of 1902–04 were available. Mitchell's prorated estimates and his split in terms of the years involved were used.

In the case of Katmai (1912), Deirmendjian (1972) gives an estimate of 13.4 Mt (at unit density). This particular eruption involved very great discrepancies in estimates ranging, according to Lamb (1970), from 0.006 to 21 km³ of solid ejecta. Deirmendjian estimated Katmai dust at 44% of Krakatoa in total dust, which would have given Katmai roughly equal magnitude in the Northern Hemisphere to Krakatoa. Junge (1974) quotes a figure of 16 Mt, 0.2 year after the eruption, which at 14-months' residence time, corrects to 19 Mt initially. Jung's (Volz's) quoted value, as noted earlier, for the Krakatoa event on a global basis corresponds (at 14-months' residence time) to 75 Mt initially, thereby putting Katmai at ~25% of Krakatoa, or 50% in the Northern Hemisphere. Normalizing by using 50% of the value used for Krakatoa (25 Mt), a figure of 12.5 Mt is obtained. Lamb's D.V.I. for Katmai in the Northern Hemisphere is only 15% of that for Krakatoa. A figure of 13 Mt was used.

In the case of the Agung (1963) eruption, it has been reported that most of the dust remained in the Southern Hemisphere. Junge (1974) quotes a value for the Northern Hemisphere as 2 million tons, 0.8 year after the eruption (versus 12 million tons in the Southern Hemisphere). If the total of 14 Mt is corrected, assuming 14 months' residence time, to the initial injection, a figure of 28 Mt is found, in essential agreement with the 30 Mt quoted by Mitchell (1970). Deirmendjian, however, gives only 9 Mt at unit density; at a density of 1.7 gm cm^{-3}, the figure would be 15.3 Mt. By Table 2 rules, 15 Mt at the latitude of Agung should have put 7.5 Mt in the Northern Hemisphere; however, the Junge (1974) figure of 2 Mt corrects only to 4 Mt initially. These data could not all be rationalized. A figure of 5 Mt was assigned, with a figure twice this also considered (see Fig. 8).

Considerable question also exists in the Hekla (1947), Mt. Spurr (1953) and Bezymyannaya (1956) eruptions, all in the far north. The values shown for Hekla and Mt. Spurr were again prorated from Mitchell's figures. Much smaller values would be estimated from the small total dust figures quoted by Lamb. The Bezymyannaya figure of 5 Mt was also doubled (Fig. 8) for illustrative purposes.

In view of the large uncertainties in the dust estimates, no attempt was made to adjust these figures as residence time assumptions were varied in the main text.

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