VOLCANIC DUST IN THE ATMOSPHERE; WITH A CHRONOLOGY AND ASSESSMENT OF ITS METEOROLOGICAL SIGNIFICANCE

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CONTENTS

After defining the terms commonly used in reporting volcanic eruptions and noting previous approaches to assessment of their magnitudes, this study proceeds to examine aspects of importance, or possible importance, to meteorology—principally the dust veils created in the atmosphere, particle sizes and distribution, heights, fall speeds and atmospheric residence times. Later sections deal with spread of the dust by the atmospheric circulation and the direct effects apparent upon radiation, surface temperature and extent of ice in the polar regions. These effects, as well as various crude measures of the total quantity of solid matter thrown up, are used to arrive at numerical assessments of volcanic eruptions in terms of a dust veil index (d.v.i.). The latitude of origin of the dust (latitude of the volcano) receives some attention, and apparently affects the course of development of the atmospheric circulation over the three or four years following, at least in the case of great eruptions (d.v.i. > 100 over one hemisphere). Effects upon the extent of ice on the polar seas may be of somewhat longer duration, and thereby influence the atmospheric circulation over a longer period of years, since there seems to be some association with the cumulative d.v.i. values when successive great eruptions occur with only few years between. The time distribution of volcanic dust since the last Ice Age, and since A.D. 1500, are indicated in as much detail as the evidence permits. Some associations with changes of climate are suggested, but it is clear that volcanic dust is not the only, and probably not the main, influence in this.

The appendices give a chronology of eruptions (including those which it seems possible to dismiss as regards any effect on world weather or climate) and a chronology of d.v.i. values. A third appendix displays by means of graphs the variation of some circulation parameters in January and July in the region of northwest Europe over the years immediately following forty of the greatest eruptions since 1680.

1. Introduction: definitions, types of eruption, assessments of magnitude

A *volcanic eruption* is the escape of magma—i.e. molten rock—usually accompanied by steam, water vapour and other gases (initially partly dissolved), from pockets or reservoirs of such material beneath the surface of the Earth's crust. The surface rock may be lifted or shattered just at the place of the eruption so that sometimes large quantities of solid matter broken into various sizes are thrown up into the atmosphere, often along with much fresh 'ash' formed by solidification of the erupting magma. Prominent among the gases emitted are water vapour, sulphur compounds and carbon dioxide.

The water vapour constitutes proportionately a minute addition to the world reservoir of water. Any meteorological or climatic effect of it must be limited to the time during which some of this vapour, and its radiation-absorption effects, constitute an element of any persisting veil of volcanic dust and vapours in the stratosphere. The total effect of such veils upon radiation and prevailing temperatures is examined later in this paper.

The sulphur compounds (H_2S, SO_2 etc.) in the gases emitted become oxidized in the atmosphere and are sooner or later brought to Earth as dilute acids in rain. Their chemical activity fixes them in the Earth's surface materials. While in the stratosphere, however, the sulphur compounds do appear to play a part in the dust veil.

Volcanic eruptions are a continuing source of supply of carbon dioxide to the atmosphere, and this is not so quickly absorbed at the Earth's surface as the other substances mentioned. The extra carbon dixoide might therefore be expected to have some climatic effect through its absorption of the outgoing terrestrial radiation; but the amounts seem to be too small to matter in this way. The addition of CO_2 to the atmosphere from volcanoes proceeds by sudden local injections, which, totalled over the first half of the present century, Plass (1956, p. 143) estimated at no more than one sixtieth part of the CO_2 produced by man's activities (burning of fossil fuels, clearance of forests, land cultivation etc.) in that time and one six-hundredth part of that produced by natural decay of organic matter. Though in periods of greater volcanic activity the additional CO_2 spread throughout the atmosphere by the general circulation may accumulate somewhat, it is not only largely absorbed in the sea and fixed as bicarbonate in the ocean bed (50 ka are probably required to restore equilibrium (Schwarzbach 1961)) but it is also very slowly abstracted from the atmosphere direct by chemical action ('weathering') of old exposed volcanic rocks, the silicates in which become partly replaced by carbonates through the chemical action of carbonic acid in cloud droplets and rain. The CO_2 concentration therefore probably varies rather little over periods of hundreds of millions of years, but anomalies may well have characterized somewhat shorter geological ages associated with great differences of volcanic activity or vegetation cover.† In historical times increments of atmospheric CO_2 due to volcanic activity can at no time have been sufficient to affect the radiation balance of the Earth significantly if Plass's estimate referred to above is, even roughly, of the right order of magnitude. No measurable variation of the general CO_2 content of the atmosphere has been found to follow even the greatest eruptions of modern times. Moreover, the great increase of carbon dioxide put into the atmosphere by man's burning of fossil fuels, which is estimated to have increased the concentration by about 10% from the nineteenth century average of 290 parts/10^6 to 320–330 parts/10^6 by 1967 (J. M. Mitchell personal communication), seems to have been insufficient to offset a world-wide cooling tendency evidently due to other causes, as yet unknown, since about 1940 (see, for example, Mitchell 1961; Flohn 1963, p. 20; *La Nature* 1966); and it remains doubtful whether this CO_2 change has had any climatic effect. On these grounds, the radiation effect of volcanic CO_2 in the atmosphere appears negligible in connexion with meteorological effects at the present time or in the periods surveyed in this paper.

Lava is the name given to molten magma after it has emerged and flows over the previous surface and when it solidifies. The composition, viscosity and gas content of lava vary. Those lavas that come from deepest in the Earth's crust are the most 'basic'. Basic lavas, whatever level they come from, have relatively low percentages of silica (SiO_2), and emerge as the hottest and most fluid lavas. They solidify to form dark basaltic rock. They contain relatively little volatile material and steam; and the eruptions which produce them are rarely explosive. Acid lavas are lighter coloured, less fluid (i.e. they are more viscous) and have a relatively high SiO_2 percentage: they are prone to explosive eruption, since the abundant compressed steam and gases have difficulty in escaping. There are corresponding differences in the 'ash' resulting from solidification of shattered bubbles of the lavas ejected into the atmosphere.

Tephra (or *pyroclastic material*) is the collective term for all broken solid matter ejected in volcanic eruptions. The names (*fine*) *ash* or *dust* are used more or less indiscriminately for particles of the smallest sizes–of diameter less than about 0.2 mm. *Coarse ash* or *volcanic sand* are used to describe grains from about 0.2 to 2 (with some authors 4) mm across. The name *lapilli* refers to bigger grains, up to 10 to 30 mm cross-measurement. The size limits are rarely applied very rigorously, partly because nearly all volcanic deposits contain a mixture of grain sizes. The distinction between *cinders* (or *scoriae*) and *pumice* for larger material is one of porosity, the density of pumice being sometimes so low that it floats on the sea. (Parts of the great raft of light coloured, acid, rhyolitic pumice produced on the Antarctic Ocean by a submarine eruption of 5 March 1962 near the South Sandwich Islands at 55.9° S 27.9° W are believed (Gass, Harris & Holdgate 1963) to have been blown 4000 nautical miles eastwards (7500 to 8000 km) over the ocean to arrive as a largely contiguous mass at Heard Island (53° S, $73\frac{1}{2}$° E) and patches on the seas to

† It has been suggested by Plass and others that the luxuriant vegetation growth of the Carboniferous (270 to 350 Ma ago) implies an atmosphere richer than now in carbon dioxide, possibly persisting from some still earlier geological age, but that the plant growth itself ultimately reduced the CO_2 in the atmosphere so much as to lower world temperature and bring on the ice ages of the late Carboniferous and Permian. However, the most careful recent estimates of the effect of an increase of CO_2 upon world temperature are all very much lower than Plass assumed: they range from negligible (Kondratiev & Niylisk 1963) to a 2 °C rise for a doubling of the present CO_2 (Manabe & Wetherald 1967). There are, moreover, other theories of the causation of those ice ages—in terms of drifting continents or, alternatively, of variable sun. The Carboniferous itself is considered to have been an era of little volcanic activity so that the supposed excess of CO_2 can hardly have been produced during it.

Mountain building tends to be accompanied by volcanic activity, but in very varying degree. It is reasonable to suppose in this connexion, however, that there may have been periods of more volcanic activity than is found at the present time (P. E. Baker, personal communication, 27 October 1965).

the north of there as far as 48° S, 74° E, $10\frac{1}{2}$ months later.) Dust formed from acidic lavas is pale, greyish white (*rhyolitic tephra*) and may have an SiO_2 content of 60 to 70 %. Dust from basic lavas is dark, brown to black, and has been found in Iceland to have an SiO_2 content of 53 to 55 % (Einarsson, Kjartansson & Thorarinsson 1954), though there may be as little as 45 % SiO_2 in some volcanic dust (P. E. Baker, personal communication). Within the course of a single eruption the materials ejected are liable to go from large 'blocks' and 'bombs', parts of the former surface, and acidic tephra from near the surface, thrown out in any initial explosion to dark, more basic material later on; though more complex and varied sequences also occur. Explosive phases may also occur late in an eruption through rock from the crater walls falling in and blocking the vent or, as at Krakatau in 1883, through the sea getting into the crater and temporarily solidifying a crust on the magma.

Eruptions are clearly of two chief types: *effusive eruptions* and *explosive eruptions*. Effusive eruptions are those in which lava flow predominates. Explosive eruptions do not necessarily produce any flow of liquid lava: it may all be thrown up vertically to condense as spray, and solidify as ash and cinders, and the partly solid froth to solidify and fall as pumice. Mud is also liable to be rained down as a result of the convection and condensation of atmospheric moisture. There was no flow of lava in the eruptions of Vesuvius in A.D. 79 or of Mt Pelée in 1902. The residence time of solid ejecta in the atmosphere is greater the smaller the particle size and the higher it is initially cast by the explosion. The explosive nature of an eruption is therefore important for any subsequent meteorological effect, though it may be possible to arrive at a more discriminating assessment of the meteorological importance of any eruption from observations of the height, extent and persistence of the dust veil.

There are *two main types of volcano*:

(1) Volcanoes with a *central*, more or less circular, *vent* (*crater*) surrounded by a cone built up of eruptive material. Most volcanoes in and around the Pacific, and in the Mediterranean and Africa, are of this type.

(2) *Fissure volcanoes*, in which lava issues from a linear crack in the surface rock. Later eruptions usually issue from a new fissure line parallel with the previous ones which remain sealed by the old lava. The greatest lava flows in post-glacial, and in historical, times have been from fissures of this type—in both cases in Iceland—respectively about 8000 years ago and in A.D. 1783 (Thorarinsson, Einarsson & Kjartansson 1959, p. 147). The eruptions which created the basalt rocks of northern Ireland and the Hebrides, and the Columbia basalt in North America, as well as the Deccan Traps (formed in the Tertiary geological era) in India, were evidently also of this type (Stubbs 1961).

Temperatures (over 1000 °C) high enough to melt rock in the Earth's crust—i.e. at depths less than 35 km—can be presumed to arise only in areas of local anomalous heating associated with crustal disturbance. The world average increase of temperature with depth is about 1 °C/33 m in the uppermost layers of the crust, diminishing with depth so that the mean temperature at 35 km depth is probably about 600 °C. In the oldest volcanic areas in Iceland which were active in the Tertiary geological era, 1 to 13 Ma ago, and are more or less quiescent today, average temperature gradients five times as great are found; in the immediate vicinity of contemporary activity at hot springs and steam holes in Iceland gradients several hundred to several thousand times as great are indicated, and the jets of hot water and steam emitted may represent a heat flux 10^{14} to 10^{15} times the global average (Thorarinsson *et al.* 1959, pp. 150–151). Reservoirs of molten material in the Earth's crust are thus of limited extent, are built up over very long periods of time and are associated with distinct chapters of tectonic activity in the part of the world affected.

In some parts of the world groups of volcanoes have appeared to indicate that stresses can build up within the Earth's crust and lead to fractures simultaneously over a sizeable region, or along a line, in some cases possibly even several volcanoes being supplied from one and the same magma pool. Indications of this kind, largely taken from Sapper (1917), may be listed as follows:

Iceland. Broadly simultaneous eruptions in different parts of Iceland in 1727–8 and in 1783.

Kamchatka. Alternating activity of two volcanoes, Kluchev and Shiveluch.

Central America–West Indies. Occurrence within 9 to 10 months of each other of severe eruptions of four volcanoes: Soufrière (St Vincent), Mt Pelée (Martinique), Santa Maria (Guatemala) and Colima (Mexico). Alternating activity over many years of Fuego and Pacaya (both in Guatemala).

Canary Islands. Alternating activity of Teneriffe and Lanzarote.

Mediterranean. Periods of activity of Vesuvius and Etna have tended to alternate, while Stromboli continues. Occasionally all erupt together. Methods of detection available since 1950 indicate that Etna has a long and narrow magma pool, about 60 km long; this may be unusually big.

Aegean. Activity of Santorin and Nisyros appears to be connected.

Java. Group eruptions of several volcanoes in 1772, 1822 and 1826.

Philippine Islands. Alternating activity of Taal and Mayon.

Ecuador. Alternating activity of Cotopaxi and Pinchincha.

Chilean Andes. Simultaneous eruptions at many points in 1835 and of smaller groups in 1932 and 1960.

The *magnitude of an eruption* is generally assessed by some estimate of the total quantity of material, including lava, solid blocks and dust, ejected. Sapper (1927) has defined the scale shown in table 1.

TABLE 1. SAPPER'S SCALE OF ERUPTIONS

magnitude	quantity of material ejected	magnitude	quantity of material ejected
1	$> 10^9\ m^3$ ($> 1\ km^3$)	5	10^5 to $10^6\ m^3$
2	10^8 to $10^9\ m^3$	6	10^4 to $10^5\ m^3$
3	10^7 to $10^8\ m^3$	7	10^3 to $10^4\ m^3$
4	10^6 to $10^7\ m^3$	8	$< 10^3\ m^3$

Sapper also classified eruptions as either (*a*) *mainly lava* in the typical *effusive*, eruptions; or (*b*) *mainly solid matter* ejected, typically *explosive*, eruptions.

In his nomenclature a_1 stands for an effusive eruption of the first magnitude, a_2 second magnitude, and so on, b_1 for an explosive eruption of the first magnitude, b_2 second magnitude, and so on.

In the reporting of eruptions (e.g. in the yearly *Bulletins of volcanic eruptions* issued since 1961 by the Volcanological Society of Japan on behalf of the International Association of Volcanology of the I.U.G.G.) most attention has generally been devoted, for obvious reasons, to the magnitude of the disaster and the most noticeable phenomena on the ground—topographic changes in the neighbourhood of the eruption, lava flows, mud flows, occasionally cinder and ash fall and the depths thereof, destruction of arable land, tidal waves, casualties. The same tendency understandably affects the reports of volcanic eruptions so far received by the collecting Center for Short-Lived Phenomena established in 1968 by the Smithsonian Institution in Washington. A quite different assessment of eruptions is needed for studies of possible meteorological effects. In this connexion the quantity of lava is of no direct interest. What matters most here must be those materials which remain suspended longest in the atmosphere, principally the smallest particles, and possibly also the water vapour injected into the stratosphere. Hence, reporting of the dust cloud, dust deposits and radiation effects are chiefly needed.

Before examining the possible meteorological consequences it is necessary to assemble the scattered data with regard to noteworthy volcanic dust veils since 1500 and attempt to give them some quantitative assessment. Though the list may not be complete for the early years, there is less likelihood of any great dust pall that spread around the world having remained unobserved and unreported than that a lava eruption in a remote area, however devastating its effects in the immediate vicinity, should go undiscovered. The list here given is meant chiefly to give a basis for identifying numbers of similar cases for meterological study. This intention will not be brought to nought if a few cases remain unknown. It is hoped that this publication may also encourage the reporting† in connexion with all future eruptions of those parameters, little related to the local disaster, that may affect world weather.

2. Volcanic dust: particle sizes, heights in the atmosphere, fall-speeds and atmospheric residence times

Our knowledge of the prevailing sizes of the particles that have made up some of the most persistent volcanic dust veils is largely derived from the optical effects observed at the ground, though supplemented by direct sampling after the 1963 Bali eruption. The height of a dust veil can be broadly determined from the times of first and last illumination of the dust layer at sunrise and sunset respectively. In some cases, indications of two or more dust layers have been observed.

(a) *Sizes of the dust particles*

Particles small in comparison with the wavelength of the incident light, as is the case with most atmospheric dust other than volcanic dust and with smoke haze, scatter the short wavelengths more than the long, so that the Sun and Moon appear red or reddish. There are also many observations of dimmed and reddish Sun after volcanic eruptions, even at points far removed from the origin of the volcanic dust in place and time.‡ Many of the particles concerned are less than half a micrometre in cross-section (Mossop 1964).

On rare occasions the Sun and Moon, seen through a cloud of minute water droplets or solid particles in which some required range of sizes predominates, appear blue or green. These are bigger particles than those discussed in the previous paragraph. At times the colour is merely bluish grey or white, in which hints of a blue tint vary with time. Blue Sun or Moon was reported at various places and times, usually of rather brief duration and occasionally with a notable tendency to red colour in the interim, in association with volcanic dust clouds after eruptions in 1783, 1821, 1822, 1831, 1855, and 1883 (*Royal Society* 1888). Within the first nine months after

† Probably the reporting of those aspects of any big volcanic eruption which are only of meterological concern will have to be provided for as an addendum to the routine synoptic weather messages, if it is to reach all those meteorological institutes that need the information for long-range weather forecasting or for research.

‡ Darkened lunar eclipses were noted in Europe in 1884 and 1964 after the volcanic eruptions in the East Indies more than a year previously: normally in a lunar eclipse the shadowed part of the moon can be clearly seen by naked eye and telescope observation reveals colours, but during the total eclipse observed in Britain on 18 December 1964 this was not so (*New Scientist*, 31 December 1964).

the great explosive eruption of Krakatau (6° S 105½° E) in August 1883 the Sun, Moon and planet Venus occasionally appeared blue or green when seen from points (mainly) within the tropical zone. Also after the other eruptions mentioned blue Sun or Moon was seen only either near the volcano or within one latitude zone around the world over which the densest dust veil was carried by the upper winds, and where it may reasonably be supposed that there were greater concentrations of rather larger particles than in most of the dust veil. According to Minnaert (1959), from observations of similar phenomena seen through artificial clouds of steam and smoke, the Sun and Moon appear blue when the particle diameters in the cloud are mainly between 1 and 5 μm. Blue Sun and Moon were again observed over Britain and western Europe on 4 or 5 days at the end of September 1950, when shining through a smoke trail in the upper troposphere from exceptionally great forest fires in northern Alberta, Canada (Bull 1951). This may be a diffraction phenomenon; though the expected surrounding bright, reddish ring, the radius of which depends on particle size, would be unlikely to be seen unless there were a strong predominance of some quite small range of particle sizes.

After the eruption of Krakatau in 1883 an unusual corona surrounding the Sun was described for the first time by the Reverend S. E. Bishop in Honolulu and subsequently observed in many places. The sky near the Sun was generally white or bluish white instead of blue and this area of white illumination, sometimes brilliant enough to be described as a white glare, was occasionally seen to be edged with a ring of pink, red, brown or orange-rose colour at an angular distance of about 20° from the Sun. The angle appeared rather less when the Sun's elevation was high and greater when the Sun was low. This phenomenon has ever since been known as 'Bishop's ring'. The coloured ring is apparently never conspicuous and may often have passed unnoticed. It was observed again after the eruptions in the West Indies in 1902 and in Alaska in 1912. And, after not being seen for 50 years, it was observed again following the eruption in Bali in 1963, at least over central and southern Africa (Burdecki 1964; W. Schuepp personal communication 18 May 1965). Schuepp reports that it was frequently visible on clear days over the Congo from December 1963 onwards, the sky near the Sun's disk being much brighter than in other years. In December 1964 this anomaly became weaker. In the Krakatau case, although the ring was originally described by Bishop within 10 days of the explosive eruption in August 1883, and possibly first seen near the volcano after the preliminary eruptions in May to July, it was most frequently and widely observed in many parts of the world in 1884 and 1885, reached its maximum brightness almost a year after the eruption and was last seen in the early summer of 1886.

The radius of Bishop's ring should, according to theory, vary inversely with the size of the dust particles or droplets causing it. Moreover, there must be a great predominance of particles within quite a narrow size range, or else the diffraction pattern due to particles of one size would obliterate that of another, rendering the colour invisible. Probably this is the chief reason why the white glare is often seen all about the Sun but the coloured ring at its edge only rarely. De Bary & Bullrich (1959) imply that particles of diameter 0.8 to 1 μm would produce a Bishop's ring of 20° radius as observed in 1883–6; smaller particles would produce a bigger ring, radius about 38° for particle diameter 0.5 μm. A faint ring at such a wide angle from the Sun would be very liable to go unobserved, but in fact no ring of radius even approaching 38° has been reported. (Earlier accounts, for instance, by Humphreys (1940, p. 591), have generally followed Pernter (1889) in indicating a prevailing particle diameter of about 1.85 μm when the ring was observed in the Krakatau dust veil.)

Exceptionally beautiful sunsets and twilight glows have accompanied many volcanic dust veils and attracted much comment. Most prominent in the paintings, photographs and descriptive accounts is simply the brilliant rosy or fiery (also described as 'lurid' or 'flame') red coloured glow which lingers near the horizon unusually long after the Sun has set, when a volcanic dust veil is present.† Within the dusk and dawn displays a phenomenon fairly often reported is a purplish patch, which may be seen at elevations broadly about 20° when the Sun is between 3 and 7° below the horizon (20 to 50 min after sunset in middle latitudes) but, in the best occurrences, is repeated when the Sun is about twice as far (10 to 13°) below the horizon (60 to 90 min after sunset): the second purple patch may be the more clearly seen against the fainter illumination of the background sky at that time. The optics have been explained by Gruner (1942). The angular distance of the first purple patch from the sum is the same as that of Bishop's ring, and the two may be essentially the same phenomenon (Gruner, p. 518). After

(a)

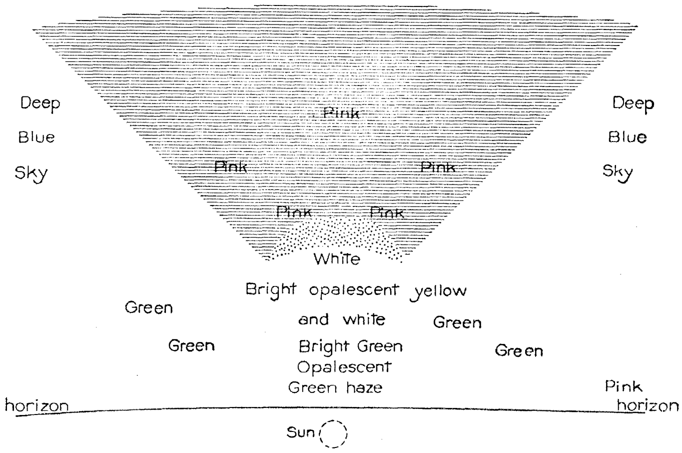


FIGURE 1 (*a*). Sketch of 'sunset colours' recorded by F. A. R. Russell at Richmond, Surrey, about 17h to 17 h 30 min on 9 November 1883. Sunset had been at 16 h 23 min. The pink patch was in view from about 16 h 40 min to 17 h 58 min, descending from about 45° elevation till it disappeared below the horizon. There were two maxima of illumination, about 17 h and 17 h 32 min. The earlier maximum was very bright, the later one was described as 'bright and deep purple coloured'.

Krakatau 1883 Bishop's ring was more widely noticed many months later when the twilights were becoming fainter. Purple patch observations from England in 1963–4, 1883, 1855–7, 1818 and 1814–15 suggest dust veils of similar structure in all those years. The same may be deduced from observations in Scandinavia in 1680, 1661, 1636 and 1553. In all these years therefore particle sizes and heights of the dust layer at points far removed from the eruption which produced the veil seem to have been similar to those in the Krakatau case. There is indeed a remarkable similarity between the observations sketched by F. A. R. Russell about an hour after sunset on 9 November 1883 and by the present writer in December 1963, in both cases in Surrey, England, reproduced here as figures 1*a* and *b*. A similar observation was made at Guildford an hour before sunrise on 17 January 1964.

The brilliant whiteness of the sky, noticeable for much of the day, within about 15 to 20° of the Sun was probably the most frequently observed symptom of the volcanic dust veils in the cases here discussed; though where the dust concentration was particularly dense (as over the equatorial and tropical zones at times in 1883 and 1963 and over northern temperate latitudes in 1783) the

† This colour was much used by J. M. W. Turner in his paintings from about 1807 onwards and especially in the 1830s. Though these works did not pretend to accuracy of form, the colour for which his work is famous may well have been suggested by the volcanic dust veils of those years (see the observations summarized in Appendix I).

sky often had a 'dirty' or 'muddy' look. For some time immediately before sunrise and immediately after sunset the white glare replaced the unusually prolonged red glow. It seems that Bishop's ring must be regarded as a rare phenomenon occurring only when and where the dust particles happen to have been sorted as to size and yielded a strong predominance of the required size: this probably limits it to association with just one of the usually overlapping dust layers in the stratosphere produced probably at one explosive stage of a given volcanic eruption.

With the densest volcanic dust veils the dimming of the Sun, which in southern France in June 1783 was not visible till it reached 17° above the horizon, and the brightness of the dusk lasting well on into the night (also particularly in 1783), have attracted more notice than any ring or colour phenomena. Part of the explanation probably lies in the wider than usual range

(b)

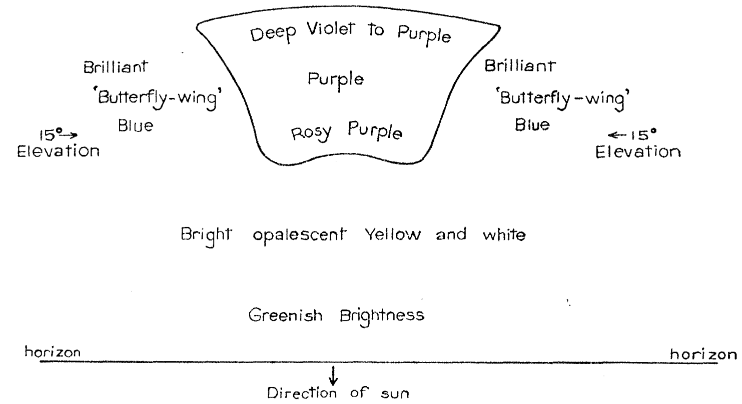


FIGURE 1 (b). Sky colours observed from Guildford, Surrey at 17 h 05 min to 17 h 10 min G.M.T. on 24 December 1963, as sketched at the time by H. H. Lamb. Sunset was at 15 h 56 min. The purple patch was softly illuminated, but marked off sharply at its edges from the much more brilliantly illuminated parts of the sky on either side and below, though with somewhat less distinction at its upper edge. The patch was in view from about 16 h 30 min for three-quarters of an hour or more, descending towards the horizon. In the earlier stages, before the sketch was made, the sky on either side of the purple patch was brilliant white and below the patch bright, yellowish white with a rose-coloured band along the horizon: the purple patch itself was brighter then and higher in the sky.

of particle sizes present in those veils. The dimming of the Sun was also remarked on by observers without instruments in 1601 (in Scandanavia), in 1821 (in France), in 1831 (in Africa) and in 1963 (in Africa and Arabia).

The persistence of the dense veil in 1783 after the eruptions in Iceland in May–June and Japan in August is made clear by the following passage written by Benjamin Franklin, then living in Paris as the first diplomatic representative and plenipotentiary of the newly formed United States of America:

During several of the summer months of the year 1783, when the effects of the Sun's rays to heat the Earth in these northern regions should have been the greatest, there existed a constant fog over all Europe and great part of North America. This fog was of a permanent nature; it was dry and the rays of the sun seemed to have little effect towards dissipating it, as they easily do a moist fog...They were indeed rendered so faint in passing through it that, when collected in the focus of a burning glass, they would scarce kindle brown paper. Of course, their summer effect in heating the Earth was exceedingly diminished.

Hence, the surface was early frozen.

Hence, the first snows remained on it unmelted....

Hence, perhaps the winter of 1783–84 was more severe than any that happened for many years.

The cause of this universal fog is not yet ascertained. Whether it was adventitious to the Earth, and merely a smoke preceeding from consumption by fire...or whether it was the vast quantity of smoke, long continuing to issue during the summer from Hekla, in Iceland, and from that other volcano which arose out of the sea near the island, which smoke might be spread by various winds over the northern part of the world...

It seems, however, worthy the inquiry whether other hard winters, recorded in history, were preceded by similar permanent and widely extended summer fogs...

Parts of this quotation anticipate the type of investigations which it is the object of the present survey of volcanic dust veils to facilitate.

During the existence of the most persistent volcanic dust veils white glare and prolonged lurid red twilight glows appear to become regular daily occurrences over wide regions of the Earth. Bishop's ring and the bright purple patch were always of more intermittent occurrence, up to ten times a month at any one place, if they occurred at all. Blue Sun and Moon have been still more localized in place and time, chiefly near the erupting volcano or broadly within the same latitude zone. This order of frequency appears to be also the order of size of the particles responsible, the bigger the particles the rarer, more localized and short-lived the occurrence. The biggest particles must tend to fall out before the patch of abundant dust with which they were ejected has time, after being carried round the Earth, to diffuse into a veil of fairly uniform texture covering one or more latitude zones.

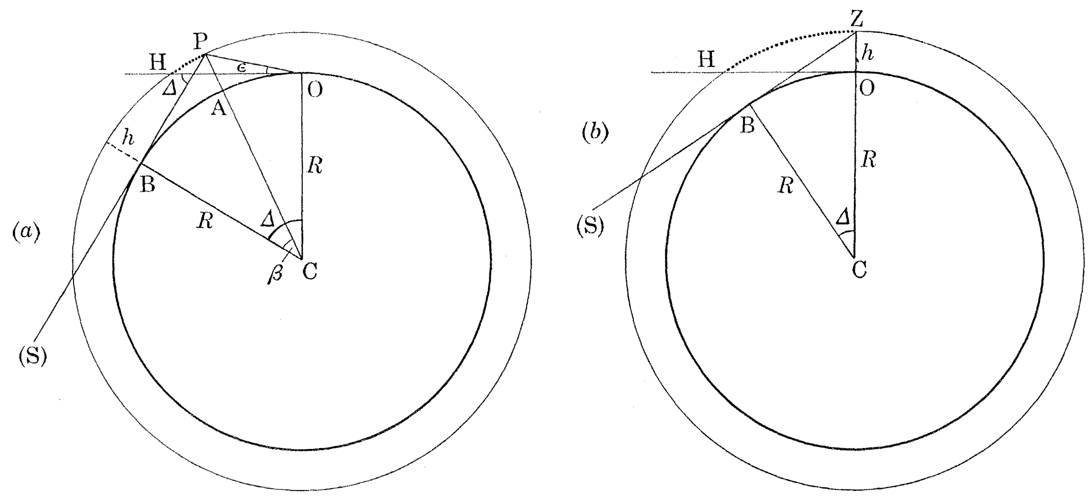


FIGURE 2 (*a*) Geometry of a layer of dust surrounding the Earth, illuminated by the Sun (S) from the horizon H up to P, as seen by an observer on the Earth's surface at O. Angle of elevation of P, ϵ; angle of depression of the Sun below the horizon, Δ; radius of the Earth, R; height of the dust layer, h; illuminated portion of the dust layer as seen from O shown by dots.

(*b*) Geometry of a layer of dust surrounding the Earth, illuminated by the Sun (S) from the horizon H up to the zenith Z, as seen by an observer on the Earth's surface at O.

It appears from this summary that dust clouds with particles commonly of 1 to 5 μm cross section have remained in existence for several months, but that those dust veils which spread over much of the Earth more characteristically consist of particles of various sizes from $\leqslant 0.5$ to about 2 μm across. Occasionally the denser dust clouds within the veil are size-sorted sufficiently to give Bishop's ring with angular radius corresponding to prevalence of particles of near 1 μm cross-measurement. Confirmatory evidence that the sizes mentioned are most abundant may also be deduced from the lifetimes of several dust veils authenticated over various parts of the globe by visual and actinometric observations (see later, pp. 441, 457–458 and appendix I). After the Bali 1963 eruption it was confirmed by direct measurement of volcanic dust particles collected at a height of 20 km over Australia between April 1963 and April 1964 (Mossop 1964).

Day-to-day variations in the phenomena observed near, or within the same latitude zone as, the volcano are doubtless due to inhomogeneities in the dust cloud attributable to variations in the ejection rate and to subsequent lateral and vertical turbulence of the wind flow in the dust layer. Farther from the volcano, and later in time, day-to-day variations become less and less noticeable: such as do occur may be attributed either to some continued inhomogeneity of dust concentration or, possibly, to condensation effects.

(*b*) *Heights of the dust layers*

The greatest heights reached by the smoke and other solid ejecta from various eruptions have been estimated in at least three ways:

(i) By measurement of the angle of elevation of the top of the column of smoke, steam and debris over the volcano observed from a known distance during the eruption.

(ii) Following a proportionality rule, noticed in the case of the 1883 Krakatau eruption, between heights observed by method (i) and the greatest distance at which the sounds of the particular explosions were heard. The maximum height attained by the debris from the greatest explosion on that occasion was estimated as about 50 km.

(iii) By observation of the times of appearance and disappearance of the solar illumination of the dust layer near sunrise and sunset or, as a variant of this, by noting the angle of elevation (ϵ) of the upper limit of the illuminated part of the dust layer when the Sun's depression (Δ) below the horizon is known.

The geometry of the methods grouped under (iii), applicable to sightings on the dust layer at any distance whatever from the volcano, is made clear in figures 2*a* and *b*. Applying the sine rule to triangle OPC in figure 2*a* we obtain

$$h = R\left\{\frac{\cos \epsilon}{\cos\,(\Delta+\epsilon-\beta)}-1\right\}. \tag{1}$$

β is obtained from the expression

$$\tan \beta = \frac{\cos \epsilon - \cos(\Delta+\epsilon)}{\sin(\Delta+\epsilon)} \tag{2}$$

derived from the geometry of the figure.

The special cases of most interest are: (i) when $\epsilon = 0$, $\beta = \frac{1}{2}\Delta$—i.e. when the illuminated portion of the dust layer is just rising or setting and (ii) when $\epsilon = 90°$, $\beta = \Delta$—i.e. when the dust layer is illuminated just to the zenith. These can be reliably observed with no equipment other than a timepiece.

In case (i), the expression (1) for the height of the dust layer reduces to

$$h = R\,(\sec \tfrac{1}{2}\Delta - 1). \tag{3}$$

In case (ii), we use the simpler geometry of figure 2*b*, which gives

$$h = R\,(\sec \Delta - 1). \tag{4}$$

Figure 3 provides curves for obtaining the height of an illuminated dust layer near sunrise and sunset. The heights so derived may need correction for:

(*a*) Refraction of the Sun's rays through the atmosphere. (This is equivalent to reducing the solar depression by about half a degree (cf. Humphreys 1940 p. 466).)

(*b*) Extinction, due to atmospheric turbidity and scattering, of the Sun's rays which pass nearest to the Earth—near B in figure 2*a*. The lowest rays which effectively illuminate the dust layer pass at some height above B. This is probably as high as the tropopause when deep cloud development is present beyond the horizon in the direction of the rising or setting Sun. But on most occasions when the sky is clear enough for good observation of the optical effects in the dust layer, the dust may be effectively illuminated by rays which have passed only a kilometre or so above the Earth's surface at B. In the case of the 1963 dust layer, which was observed by various methods and in many places, including sightings from aircraft (e.g. Flohn & Henning 1964) flying above the tropospheric turbidity, surface observations by this method indicated heights for the dust layer which agreed with the other observations to within 1 to 5 km. No such great correction appears necessary therefore as reported by Paton (1964, pp. 162, 175–6) in the case of the much more faintly lit noctilucent clouds at 80 to 85 km which are never seen with the sun less than 6° below the horizon because of the general brightness of the background sky.

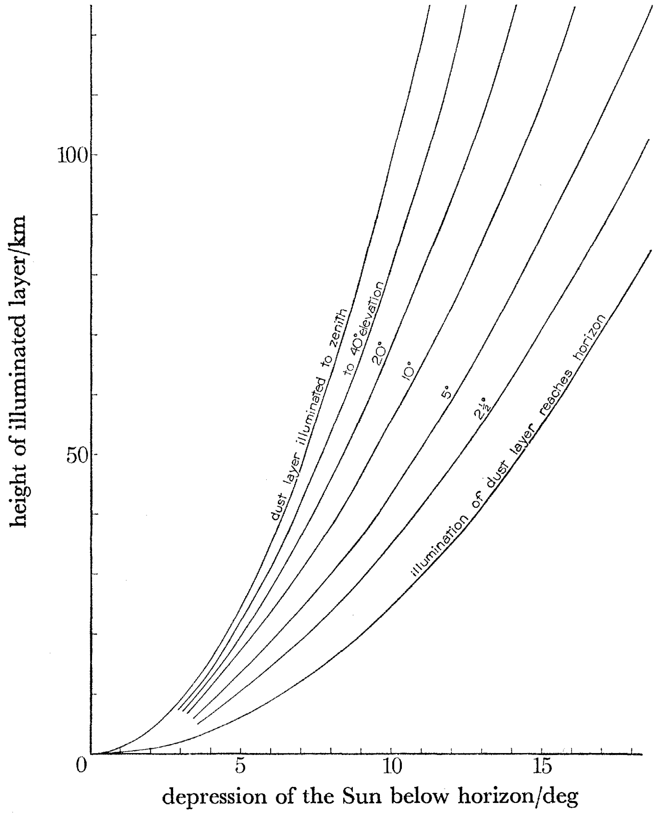


FIGURE 3. Nomogram for finding the (maximum) height of the dust layer.

It seems safe, in conditions where the lower atmosphere is clear over an area extending 150 km or more in the direction of the rising or setting Sun (usual when optical effects of volcanic dust layers are clearly seen), to accept the heights observed by the methods discussed here under (iii) as correct to within 2 to 5 km. This is all the more reasonable since the tentative corrections (*a*) and (*b*) apply in opposite directions and tend to cancel out. The observations (on the height of the upper—i.e. last illuminated—part of the dust layer) should however be repeated on as many nights as possible.

Figure 4 displays the various observed heights of haze and dust layers (usually of the tops of such layers), and of the columns of dust, smoke and vapour over the erupting volcano during the main stages of activity, in the case of fifteen eruptions for which such observations are available. Even from this small sample some points are evident:

(1) Dust etc. is thrown to different heights at different stages of the same eruption, and dust clouds or layers at several different heights result.

(2) In several cases it is clear that the greatest heights were observed only during, or soon after, the eruption and that there was some lowering of the dust top over the months that followed.

(3) None of the eruptions which failed to put dust into the stratosphere produced a persistent cloud except in so far as the Irazu and Surtsey eruptions continued to produce fresh dust over many months.

(4) Though these height observations are far too few to establish any tendency for the dust to become concentrated about certain preferred heights, presumably by the vertical circulation of

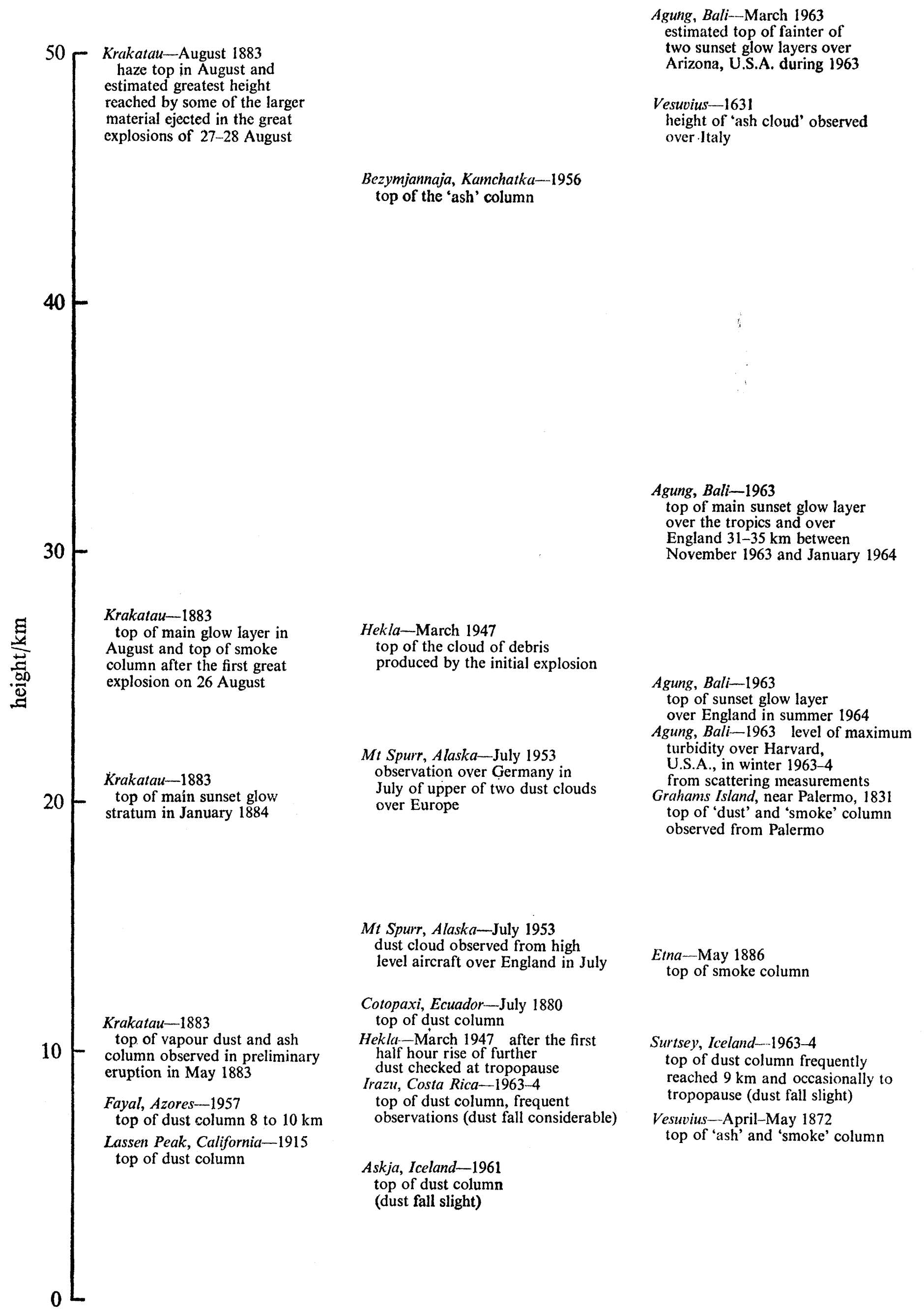


FIGURE 4. Greatest heights derived for the dust after particular volcanic eruptions. The height is indicated by the level on the diagram of the (*italicized*) name of the volcano. In the Krakatau eruption report (Royal Society 1888) it is stated that the 'vapour column' over Vesuvius commonly reaches 7 to 8 km during 'great' eruptions.

the atmosphere, the observations do appear more or less bunched into three height ranges: (*a*) top of the dust at 7 to 15 km, in the upper troposphere; (*b*) top at 20 to 27 km; (*c*) top at 45 to rather over 50 km, where some of the dust and any associated water vapour, etc., would be liable to get caught up by convective motions in the mesosphere and carried up to the mesopause (80 to 82 km).†

Those dust concentrations which occur in the lower stratosphere ((*b*) in the above list) must be the ones that account for the most persistent and dense veils. Dust in the troposphere (the (*a*) cases) will be soon—within a few weeks at most—washed to the ground by rain. On the other hand, the extreme heights (*c*) occasionally reached presumably involve only a minute fraction of the volcanic material sent up. In contrast to the troposphere which is a region of rapid vertical air transfer and exchanges, the stratosphere is characterized by very slow vertical transfer and exchanges of air except during a few sudden warming events which take place over the higher latitudes each year in late winter and spring.

Occurrence of a maximum frequency of volcanic dust at heights around 20 km has parallels (i) in the observed distribution of nuclear fission products after bomb tests in the 1950s and 1960s (e.g. figures 5*a*, *b*), (ii) in an apparently permanent and world-wide layer between about 16 and 23 km of high aerosol content (figure 5*b*) of the order of 100 particles per litre. Concentrations of matter in the lower stratosphere naturally depend on the levels of injection, or in the case of (ii), of formation of the particles concerned. There may, however, be an additional factor: patterns of mean drift of the atmosphere in a vertical north–south plane derived by Faust & Attmannspacher (1961), Faust (1962) from other considerations (observed temperature gradients) imply some general tendency to confluence of the vertical motions between about 10 and 30 km height over middle latitudes, and this could tend to concentrate dust etc. at those heights. It is noticeable that the three levels (*a*), (*b*), (*c*) noted in connexion with volcanic dust are just those of the three layers of minimum vertical motion suggested by Faust.

The supposedly permanent aerosol layer at heights around 25 km, consisting mainly of particles—radii 0.1 to 1.0 μm—of ammonium sulphate and ammonium persulphate, is attributed (e.g. Junge, Chagnon & Manson 1961; Cadle & Powers 1966, p. 181) to oxidation at these heights, by encounter with the atomic oxygen increasing with height, of H_2S and SO_2 diffused up from the lower atmosphere. These gases are also prominent in volcanic emissions. It has been suggested therefore that volcanic eruptions may be effective in creating veils in the stratosphere not only by the dust thrown up but by increasing the concentration of sulphate particles in the permanent aerosol layer. Mossop (1963, 1964) collected particles from this aerosol layer over Australia before and after the Bali eruption of March 1963. Those collected before the eruption were typically water-soluble sulphate particles with diameters up to 1 μm, encasing much smaller insoluble particles thought to be of extraterrestrial origin. In the collections in 1963 and 1964 after the eruption the main constituent was volcanic dust, which was present as angular insoluble material initially larger (cross-measurements up to 4 μm) than the normal

† Luminous night clouds (noctilucent clouds), cirrus-like cloud structures sometimes called ultra-cirri because of their exceptional height, characteristically about 82 km, which accounts for their illumination by the sun far into the night, were first noted after the Krakatau 1883 eruption. Reports of them seem to have been mainly clustered in years after great volcanic eruptions: reports were most numerous in 1884–92 and 1964–7, but some were also observed in 1900, 1902–4, 1932–7, 1949–52, more frequently in 1953–7, and again in 1959–62 (Arakawa 1959; Gruner 1942, pp. 519, 521; Paton 1964, pp. 167, 169). The association of most, but not all, with eruption years may be suggestive. Sampling by rocket techniques of the air in such clouds over Sweden suggests, however, that they consist of ice crystals: so any association with volcanic eruptions is probably through injection of water vapour into the mesosphere.

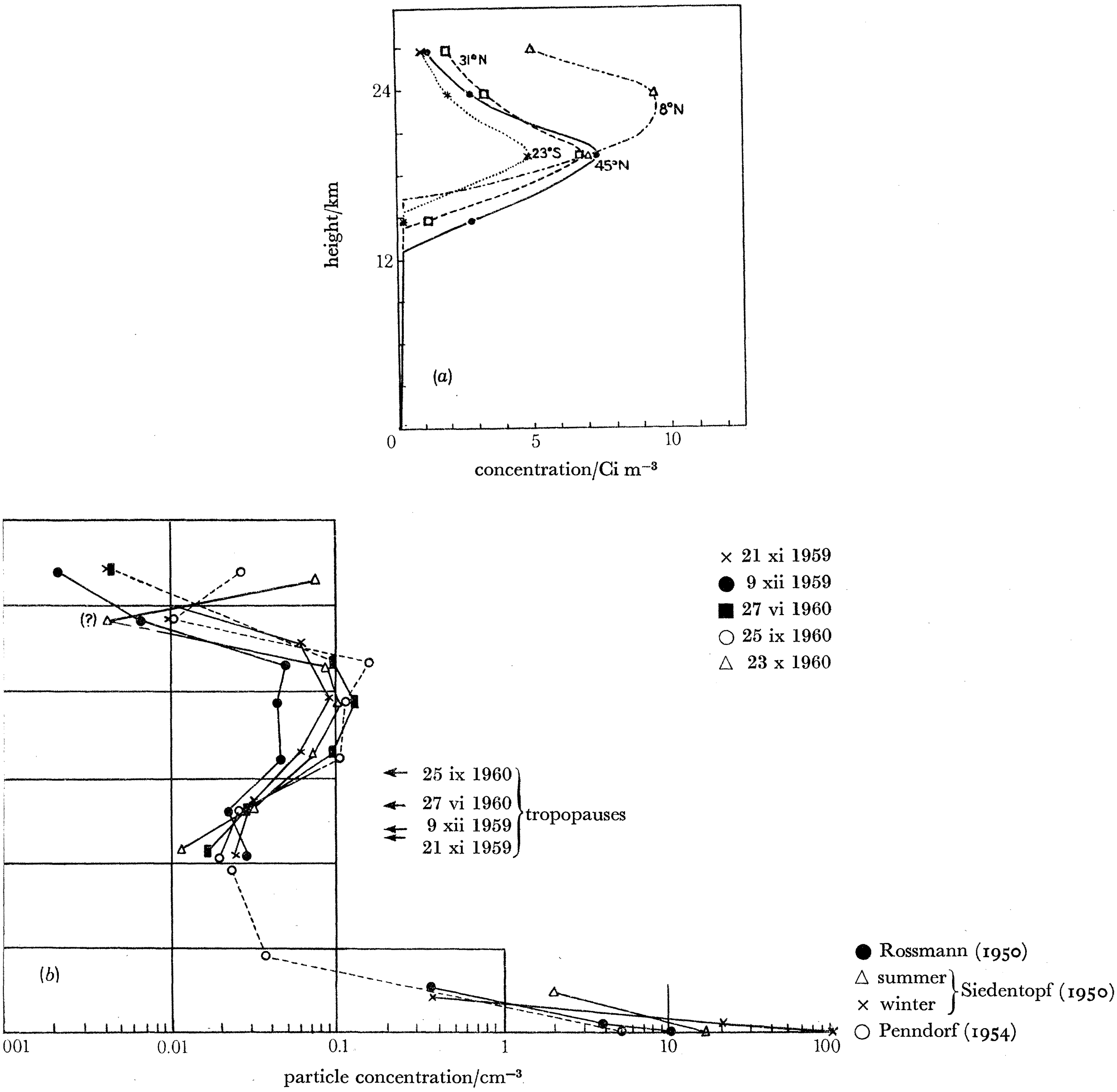


FIGURE 5. (*a*) Height distribution of strontium-90 in different latitudes, averaged for the period June 1957–June 1958, from balloon measurements (after Flohn & Reifferscheid 1961).

(*b*) Vertical distribution of aerosols of radius 0.1 to 1.0 μm. The curves about 10 km were all obtained over the U.S. Middle West, those at lower levels over central Europe. The well-marked aerosol layer between 16 and 23 km is believed to be of world-wide occurrence. (After Chagnon & Junge 1961).

aerosols. The volcanic dust fragments were also coated with soluble material, which further increased their size to an unknown extent and which may mean that even one year after the eruption the total size of the average particle was $\geqslant 1\ \mu$m although most of the solid fragments then present were $\leqslant 0.5\ \mu$m. Mossop noted that the soluble material was more abundant after the eruption; the material coating the dust fragments was reported as fluid, strongly sulphurous and itself apparently a product of the Bali eruption.

With increasing passage of time after an eruption the slow fall of the particles in the stratosphere must in any case tend to concentrate them in the lowest layers above the tropopause, before entering the troposphere and being soon rained out of the atmosphere.

(c) *Fall speeds and atmospheric residence times of the dust particles*

Stokes's law gives the terminal velocity w of a spherical particle of radius r and density ρ falling through still air (density ρ_a) as

$$w = \tfrac{2}{9}gr^2\left(\frac{\rho-\rho_a}{M}\right), \tag{5}$$

where M is the viscosity of the air and g is the gravitational acceleration.

In the case with which we are concerned this is corrected for slip by multiplying by Cunningham's factor $(1+A\,l/r)$, where A is a constant and l is the mean free path of the gas molecules. Humphreys (1940, p. 592) rewrites Cunningham's factor in terms of atmospheric pressure p, to obtain

$$w = \tfrac{2}{9}gr^2\left(\frac{\rho-\rho_a}{M}\right)\left(1+\frac{B}{rp}\right), \tag{6}$$

where B is a constant for any given temperature. When p is measured in millibars, B has values of about 0.0062 at -55 °C and 0.0084 at $+23$ °C.

The viscosity of air M also depends on the temperature, which (following Humphreys) is given approximately by

$$M = \frac{150.38\ T^{\frac{3}{2}}}{T+124} \times 10^{-7} \tag{7}$$

in *grammes per centimetre per second*. Its value is about 1416×10^{-7} at -55 °C and about 1822×10^{-7} at $+23$ °C. T in this expression stands for the absolute temperature in kelvins.

Adapting Humphreys' calculations, we may take ρ as 2.3 for volcanic dust, a substance similar to glass. ρ_a can be treated as 0 in equation (6) because it is negligible by comparison with ρ: the density of air is about 0.0012 g cm^{-3} at average surface temperature and pressure and about 0.000017 g cm^{-3} at average conditions at 30 km altitude. Table 2 gives the resulting terminal velocities and corresponding times taken for a spherical particle to fall through 1 km of still air at various heights and for several particle sizes. The table is not continued to heights below 10 km because once a particle falls below the stratosphere it is likely to be soon washed out of the atmosphere in rain or snow. Its residence time in the troposphere is likely to be similar to that of the water vapour molecules themselves, for which Flohn (1963, p. 43) estimates 9 days is the average in the equatorial rainbelt and 11 days the global average.

Table 3 gives the corresponding total times taken to fall from various starting heights to the tropopause at either (*a*) 17 or (*b*) 12 km, typical heights for (*a*) the tropics and (*b*) temperate or polar latitudes, ignoring any vertical motion of the air.

The actual fall times will differ somewhat from those indicated in table 3 (*a*) because, as pointed out by Humphreys, the particles of volcanic dust are not as a rule spherical but consist very largely of thin-shelled bubbles or fragments thereof with smaller teminal velocities than solid spheres; (*b*) because the solid particles are commonly enlarged into flattened spheres by a coating of liquid matter (e.g. H_2SO_4 which may be also ultimately derived from the volcanic eruption—see p. 438, preceding section); (*c*) because of vertical and horizontal transports of the air in the stratosphere.

Table 2. Terminal velocity for various particle diameters

height	5 μm terminal velocity	5 μm time to fall	2 μm terminal velocity	2 μm time to fall	1 μm terminal velocity	1 μm time to fall	0.5 μm terminal velocity	0.5 μm time to fall
km	cm s^{-1}	1 km	cm s^{-1}	1 km	cm s^{-1}	1 km	cm s^{-1}	1 km
40	2.1	13 h	1.0	28 h	0.25	4.7 days	0.063	19 days
30	0.51	54 h	0.24	5 days	0.06	20 days	0.015	11.5 weeks
20	0·17	7 days	0·08	15 days	0·02	8.6 weeks	0.005	34.4 weeks
15	0.12	10 days	0.055	21 days	0.014	12 weeks	0.0035	48 weeks
10	0.10	12.5 days	0.045	27 days	0.011	15.5 weeks	0.0030	62 weeks

Table 3. Residence times in the stratosphere for various particle diameters

Total times taken to fall to tropopause at (*a*) 17 km, (*b*) 12 km

initial height	5 μm		2 μm		1 μm		0.5 μm	
km	(*a*)	(*b*)	(*a*)	(*b*)	(*a*)	(*b*)	(*a*)	(*b*)
40	11 weeks	18 weeks	25 weeks	41 weeks	1.9 a	3.1 a	7.8 a	12.5 a
30	9 weeks	17 weeks	21 weeks	37 weeks	1.6 a	2.8 a	6.5 a	11.3 a
25	7 weeks	14 weeks	16 weeks	31 weeks	64 weeks	2.4 a	5.0 a	9.7 a
20	3 weeks	10 weeks	7 weeks	23 weeks	29 weeks	1.7 a	2.2 a	6.9 a
17	—	7 weeks	—	15 weeks	—	61 weeks	—	4.7 a
15	—	4½ weeks	—	10 weeks	—	39 weeks	—	3.0 a

The figures in table 3 nevertheless appear, when compared with observation, to be reasonable as regards the order of time that dust remains suspended after an eruption—especially if one assumes that the optical effects of the smallest-grain dust, about half a micrometre or less across, gradually become indistinguishable from those of ordinary scattering by the atmosphere. Bishop's ring, produced by particles of the middle sizes in table 3, after the August 1883 Krakatau eruption, was last seen 2.8 years later in Europe, 3.1 years later in Colorado. Over subtropical latitudes in the northern and southern hemispheres, where the tropopause is sometimes as high as 15 to 17 km, it was seen until 1 to 1.3 years after the eruption. In lower latitudes, nearer where the eruption occurred, the ring was apparently masked by the general glare and obscurity caused by the larger particles; after this had gone, the ring was not observed, perhaps because of some net drift of the air in the lower stratosphere towards the poles. Residence times of 10 to 20 years are clearly possible (if speeds of fall through the air were alone responsible for removing debris) for non-spherical volcanic fragments of dimensions about the wavelength of maximum solar energy around 0.5 μm, and possibly up to 10 years for many of the fragments of 0.5 to 1 μm size; though poleward transport might mean that such long residence times effectively occur only over the polar regions, if anywhere. Owing to 'leakage' of stratospheric air through the tropopause gaps associated with frontal zones and jet streams in high latitudes, however, the overall average residence time in the stratosphere of the air itself is estimated as about 10 years (Bolin 1965).

Greater persistence of the dust veils over middle than lower latitudes, even in the case of eruptions near the equator, is fairly clear from the reports—especially after the Krakatau 1883 eruption as explained in the last paragraph. That greatest pesistence should occur over the polar regions seems likely to give rise to important climatic effects through more prolonged cooling there and production of more sea ice.

Another implication is that examination of the annual layers of the ice caps in Greenland and Antarctica for volcanic dust might produce an improved chronology of volcanic dust in the atmosphere, including perhaps an indication of the number of years over which deposition continued after each great eruption.†

3. Spread and transport of the dust by the upper winds

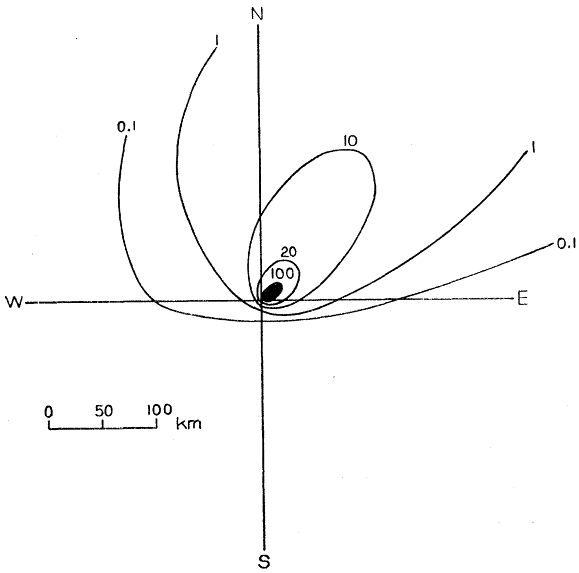
This section proceeds from consideration of some of the greatest, quantitatively surveyed dust deposits from past volcanic eruptions to evidence of how the dust is spread, including the breadth and rate of growth, and later decline, of the dust veils. A short summary of the main features of the upper wind circulation through the range of relevant heights is included, as well as observational evidence of various kinds of the transport not only of volcanic matter but also nuclear fission products from atomic weapon tests. The main streams of the upper winds over the Earth are quasi-zonal, and particles are carried rapidly eastwards or westwards around the latitude zone into which they are first injected. The spread is broadened by the meanderings of the upper westerlies over middle and higher latitudes. Cyclonic and anticyclonic eddies, meridional motions associated with the major seasonal changes, and a slow net poleward drift of the air in the lower stratosphere at least between the equator and latitudes 30° N and S, are the processes principally concerned in the extension of the volcanic dust veil to the widest range of latitudes that it ultimately covers. The result of these motions is that dust from volcanic eruptions in the higher latitudes of each hemisphere only creates veils dense enough to affect the radiation balance over latitudes 30 to 90° in that hemisphere, whereas eruptions near the equator produce worldwide dust veils. With the latter the dust is progressively lost from its main reservoir over the latitude of origin, partly by the tendency to drift away to higher latitudes and partly by falling into the high-reaching tropical troposphere where it is soon washed out in rain: hence, after the first two years, or thereabouts, the effective dust veil remaining in the stratosphere from an equatorial eruption seems increasingly confined to the two polar regions before finally disappearing.

(a) *Evidence from volcanic dust deposits*

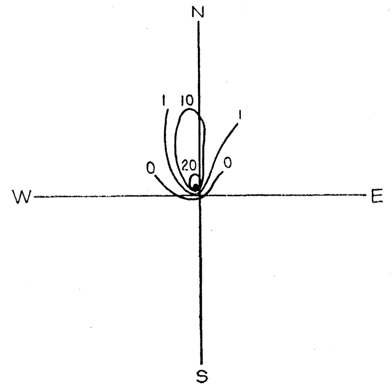
Figure 6 illustrates the depths of tephra deposited over areas around and down-wind from four eruptions in Iceland investigated by Thorarinsson *et al.* (1959). Two, or possibly three, of these must be accounted great eruptions on any reckoning. The authors say that the Öraefajökull 1362 eruption produced the biggest tephra layer in Iceland in historical times (i.e. since A.D. 870); Thorarinsson (1958) has estimated the tephra production of the 1362 eruption as 6.3 km^3, or possibly 10 km^3 when freshly fallen. The 1362 eruption is believed to have been the third greatest in Iceland since the Ice Age as regards quantity of solid matter thrown up and the greatest in Europe since the eruption of Vesuvius in A.D. 79 which destroyed Pompeii. The solid matter thrown up during the enormous lava eruption in Iceland (Laki) in 1783, the biggest lava flow on the Earth in historical times (Thorarinsson *et al.* 1959; Sapper 1917), emitted from

† So far this application of glaciology is in its infancy. Volcanic dust layers have been found in the great ice sheets both in Greenland and Antarctica, but have so far been used only to correct errors of counting of annual layers (due to lack of definition) in the stratigraphy against eruptions of known date. (Dust from the Katmai, Alaska, 1912 eruption found in the north Greenland ice revealed an error of one year.)

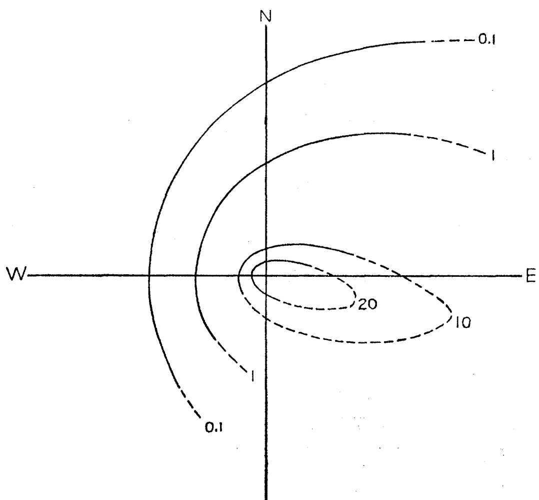
a fissure 25 km long, produced a tephra layer over Iceland that does not appear to have been equally thoroughly surveyed. This is doubtless because the bulk of the deposit must lie at the bottom of the Atlantic, the eruption having occurred 50 km and less from the coast in the direction in which the dust was carried. The tephra deposit over an area of 4600 km^2 in Iceland, equivalent to a circle of 77 km diameter, was estimated by Thoroddsen at 1.6 km^3 (Thorarinsson 1944, p. 119), corresponding to an average depth of 35 cm over this area; the total solid matter ejected, presumably including matter deposited over a far wider area, was estimated by the same writer as 2 to 3 km^3 (Sapper 1917, p. 139). Dust fell 'in great quantities' as far away as northern Scotland (Caithness), 1000 km from the eruption, and in the Faeroe islands. From the



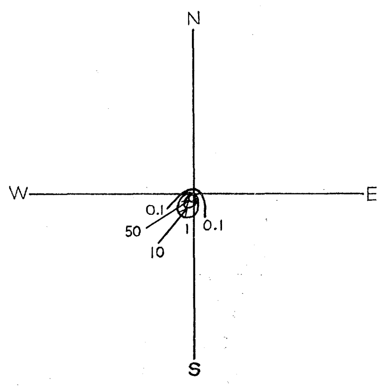
(*a*) Hekla. Dust from an eruption about 750 B.C.



(*b*) Hekla A.D. 1104



(*c*) Öraefajökull A.D. 1362



(*d*) Hekla A.D. 1947

FIGURE 6. Distribution of observed depths of tephra layers produced by various volcanic eruptions in Iceland. Depths in centimetres. Area covered to more than 100 cm depth shown solid black. Broken lines indicate extrapolations out to sea. (Adapted from maps and diagrams by Thorarinsson in various papers, cited in this work.)

evidently appreciable deposit in Caithness, from the numerous reports of the exceptional density of the dust veil (see appendix I), and on the basis of dust veil index values calculated from the apparent effect upon the temperature of the northern hemisphere to be explained later on pp. 470 to 476 (see also appendix I), a total dust ejection of the order of 10 to 12 km^3 from Iceland in 1783 appears more likely than the estimates quoted. The diagrams in figure 6 afford a basis for comparison of the magnitudes of the dust clouds in other cases where the depths deposited over an area, or even at one or two points, at known distances around and down-wind from the volcano are available. Similar maps of the dustfall from recent eruptions in Japan and prehistoric eruptions in Alaska–Yukon and New Zealand have been published by Eaton (1963) and add further to our knowledge of comparative magnitudes.

One New Zealand eruption (Lake Taupo in central North Island, near 38° S, 176° E), tentatively dated about A.D. 1000 to 1250, apparently deposited 7 to 8 km^3 of dust. The undated 'ancient' Yukon eruption in the Rockies near 61° N, 142° W, supposedly between about 500 B.C. to A.D. 500 (Capps 1915), produced a tephra deposit, mapped in Eaton (1963), amounting to 35 to 40 km^3 in its present compressed state–so probably over 50 km^3 when freshly fallen. Hence, for this eruption, an initial dust output from 2 to 10 times as great as from the thoroughly investigated Krakatau 1883 eruption may be deduced. The other cases mapped by Eaton fall well within the range of magnitudes of eruptions assessed in this report. Judged similarly just by the tephra deposit within a few hundred kilometres of the volcano, the Hekla eruption about 750 B.C. and the Öraefajökull eruption in 1362 (both in figure 6) must have been of about the same order of magnitude as Krakatau 1883. But in all these other cases the dust was injected into the atmosphere so far from the equator that, as will be shown later in this section, only a quarter to a half the Earth was likely to be much affected. The greatest effect over a period of years would be expected to be on temperatures and sea ice production in the polar regions of one hemisphere only.

The fact that the main dust dispersal was in a different direction for each Iceland eruption illustrated in figure 6 must mean that the bulk of the material was thrown up in instantaneous explosions or over short periods during which the wind that happened to be blowing did not vary much. Some of the biggest injections of dust into the atmosphere, especially those that go high enough into the stratosphere to produce the most persistent veils, are produced in great explosions, often lasting only minutes or less.

Despite the evidence of variable winds typical of Iceland in figure 6, Thorarinsson *et al.* (1959, p. 152) state that the tephra layers found in Iceland—there may be up to 100 in a single subsoil profile—are generally thickest to the east of the volcanoes: this is in accord with the fact that upper winds with westerly components are commonest (and evidently have been so over post-glacial times).

In figure 7 *a* and *b*, reproduced from Thorarinsson (1944), the dust deposit from the eruption of Askja in Iceland on 28–29 March 1875 is shown, with depths over Iceland and hourly isochrones on its passage to Scandinavia, as indicated in an investigation soon after the occurrence by Mohn. This shows the dust spreading out over an area 600 km (over 5° of latitude) broad during the course of its transport to the west coast of Norway, 1400 km from the volcano, spending just 16 h on the way. The dust is thought to have been carried by the winds in the upper troposphere in this case, and the narrowing of the dust trail as it crossed Scandinavia may have been due to partial washing out by rain and snow over the mountains. Deposits of dust from 1 to about 10 mm deep over Scandinavia were confined to a strip about 200 to 250 km broad.

Figures 8*a* and *b* indicate how the tephra was carried by the upper winds during the first 2 to 3 days after the Hekla 1947 and Krakatau 1883 eruptions. At the points reached by the dust cloud after this lapse of time tephra was being deposited in observable quantities over a strip of the Earth's surface which was 100 to 140 km (approximately 1 to $1\frac{1}{2}°$ latitude) broad after the Hekla eruption in 1947 and 500 to 600 km (approximately 5° latitude) wide after the Krakatau explosion in 1883. The breadth of the initial dust pall dense enough to affect the solar radiation was doubtless a good deal wider than the trail of measurable dust deposit—judging by the Askja case (figure 7*b*), probably $1\frac{1}{2}$ to $2\frac{1}{2}$ times as wide. The distances travelled by the dust

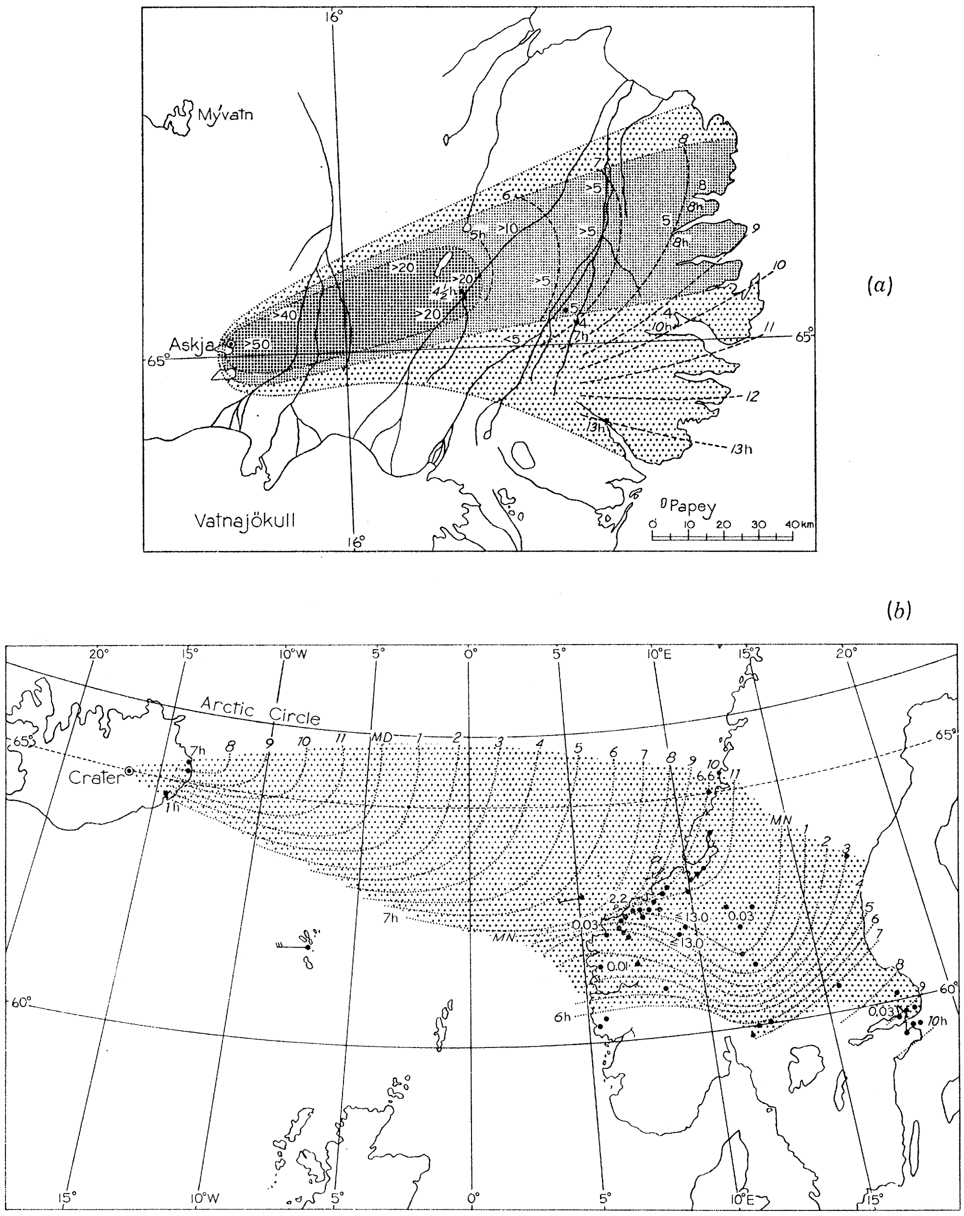


FIGURE 7. (*a*) Spread of dust over Iceland from Askja on 29 March 1875. Depths of deposit in centimetres. Dotted lines are isochrones, times G.M.T. (after Thorarinsson). (*b*) Spread of dust from the Askja volcano in Iceland to Scandinavia on 29–30 March 1875. Point depths of dust deposit in Scandinavia given in millimetres. (After Mohn 1877). Note: *MD* means midday, *MN* midnight.

along its path from the volcano in the 2- to 3-day periods covered by figures 8*a* and *b* were about 4000 and 2000 km respectively. Transport in the upper troposphere and in the stratosphere was plainly involved in both cases; though the quantity of dust and the range of height through which the dust was spread were much greater in the Krakatau case, accounting for the much wider fan of significant dust deposit.

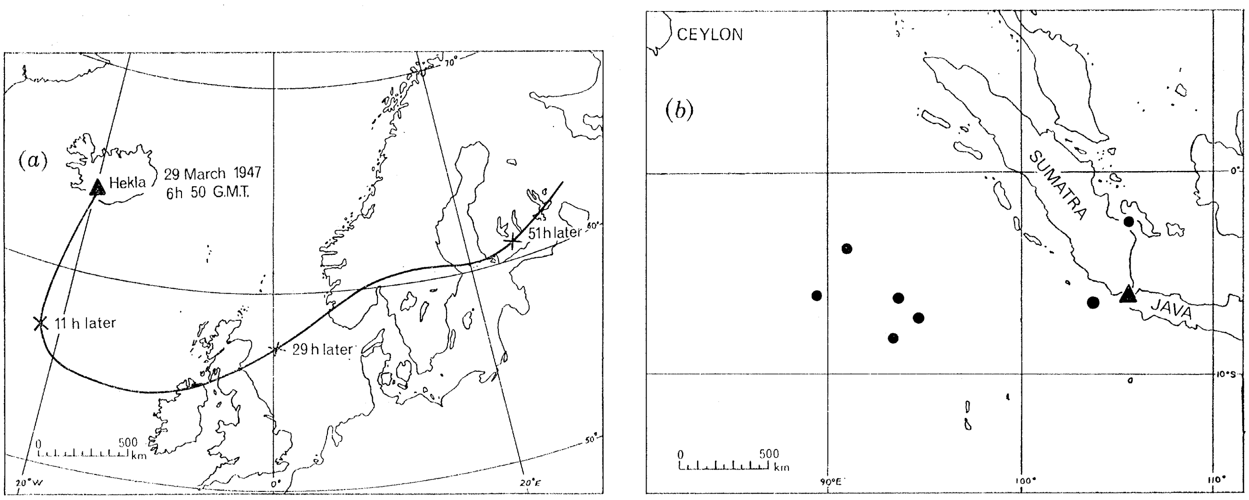


FIGURE 8. (*a*) Track of the dust cloud in the upper troposphere from the Hekla eruption in March 1947 (after Thorarinsson 1954). Dust fell in observable quantities at points along the track marked, including over a belt 100 to 140 km wide across Finland. Dust which fell in northern Scotland may have been carried by the wind in the stratosphere.

(*b*) Points where dust or pumice was observed on the surface of the sea within 3 days after the main explosion of Krakatau on 26–27 August 1883 (after Royal Society 1888).

(*b*) *Formation of veils in the atmosphere*

Because most volcanic dust appears to be thrown into the upper atmosphere in more or less instantaneous explosions, the problem of its spread by the atmospheric circulation is essentially the same as that of the spread of nuclear bomb debris at similar heights after test explosions. The progress of nuclear fission products through the atmosphere is readily traced, and has been much studied; the results will be presented in §§3 (*c*) and (*d*) below. Rates of fall-out and persistence in the atmosphere of bomb material and volcanic dust veils will, of course, only be similar in cases where the particle sizes and abundance of solid matter are alike. Certainly no atomic weapon test so far conducted has put enough dust into the atmosphere for the general optical evidence that follows great volcanic eruptions to be reported (see, however, Staff members of the Forecast Research Laboratory, Tokyo 1955; Arakawa & Tsutsumi 1956).

Volcanic explosions are not uncommonly repeated at varying strengths over a period ranging from hours to months. Volcanic dust clouds are therefore likely to be formed in several, or even in numerous, layers. Later, the fine dust may be moved by the vertical circulation of the atmosphere towards the commonly observed heights noted in the previous section; but the dust veil is likely to consist in detail of several overlapping layers, each drawn out horizontally by the great horizontal eddies in the wind circulation and spreading vertically downwards through some depth because of the tendency of the larger particles (with the greatest ratio of weight to air resistance) to fall.

(*c*) *The zonal wind circulation*

The dominant features of the mean wind circulation over the globe are great, more or less westerly and easterly, i.e. 'zonal', upper windstreams girdling the Earth at various latitudes shifting with the seasons. These are illustrated in figures 9*a* and *b* by average pole to pole cross-sections for January and July, which show mean flows of up to 50 m s^{-1} (100 knots), or rather more, in the westerly jet stream zones over the middle and higher latitudes. The shaded areas on the diagrams indicate upper easterly winds, the strongest mean flow indicated in these areas being of the order of 30 m s^{-1} (60 knots). Mean drifts in the transverse, meridional directions are of the order of a hundredth part of the strength of these zonal components. (The mean vertical components are about three orders of magnitude less again.) Hence, the upper winds should quickly carry any suspended matter around the latitude zone concerned; but the spread to other latitudes is likely to be a much slower process and to depend a great deal on eddy diffusion.

The prevailing zonal winds at different latitudes and heights may be summarized as follows:

(i) *Equatorial and tropical zones*

(*a*) In the bottom 10 km rather light, mainly easterly winds.

(*b*) In the upper troposphere, between 10 and 17 km, in general a somewhat irregular régime of rather light winds, mainly easterly in the lower part but often westerly in the upper levels. An easterly jet stream with speeds over 25 m s^{-1} (50 knots) occasionally reaching 50 m s^{-1} (100 knots), near the core, occurs in the upper troposphere (10 to 17 km) in the northern hemisphere summer near 10° N all along the equatorial flank of the subtropical high-pressure ridge from Tibet to the Sahara.

(*c*) Above the tropopause (about 17 km) at latitudes within about 10° of the equator prevailing easterly winds alternate with prevailing westerlies in a roughly 2.2 year cycle, the easterlies at their maximum being stronger than the westerly maximum. These are rather steady zonal winds of strengths commonly over 20 m s^{-1} (40 knots), and perhaps 33 m s^{-1} (65 knots) at the core. The Krakatau dust in 1883 encircled the Earth westwards at a mean speed of 32.5 m s^{-1} (63 knots). Each new cycle appears first as westerly or easterly winds setting in at levels about 30 km (or above) and usually takes about 13 months to work its way down to the tropopause. This means that there is commonly a layer of zonal westerly winds overlying a layer of easterlies, or vice versa, and sometimes a layer of one or other sandwiched in between winds of opposite sense above it and below it, all between 17 and 30 km (see Ebdon 1961; Veryard & Ebdon 1961; Belmont & Dartt 1964). The period length and the rate of descent of the easterly and westerly wind régimes in this cycle are not constant. Though about 26 months (2.2 years) has been the commonest length for a complete cycle as seen in the observations between 1954 and 1968, individual cycles have ranged from 21 to 35 months. There is some reason (from similar and probably related fluctuations in surface weather) to suppose that the slower cycles were typical in the seventeenth, eighteenth and early nineteenth centuries as in the most recent years 1963–8 (e.g. Wright 1968).

Figures 9*a* and *b* show the principal seasonal changes of the prevailing windstreams and the latitudes of their boundaries, also the relative wind speeds prevailing at different heights and seasons. They cannot, however, reproduce the 2- to 3-year alternating cycle of easterly and westerly winds in the equatorial stratosphere.

(ii) *Extratropical latitudes*

(*a*) Throughout the troposphere, mostly westerly winds prevail (especially in the upper troposphere), apart from the low level Trade Winds in the subtropics and the usually much smaller areas of low-level easterlies over and around the polar cap and occasionally between 45 and 65° latitude in connexion with blocking anticyclones.

(*b*) The upper westerly winds, meandering over a zone 20 to 40° latitude wide, are strongest in the upper troposphere, where speeds of 40 to 60 m s^{-1} (80 to 120 knots) are characteristic at the heart of the main jet stream over much of the year and winds of over 100 m s^{-1} (200 knots) are occasionally observed. At the level of maximum wind in the upper troposphere (*ca.* 8 to 12 km)

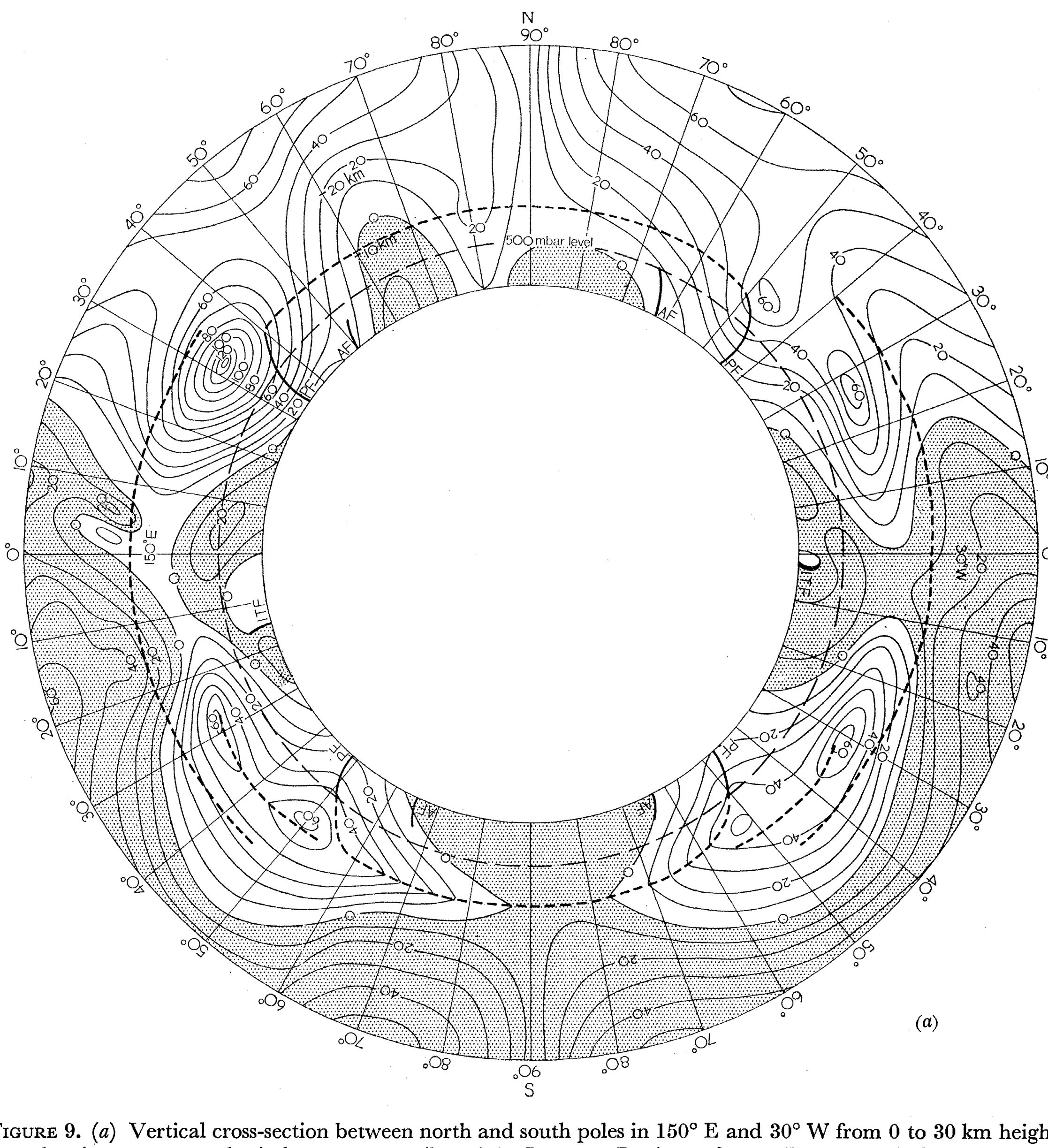
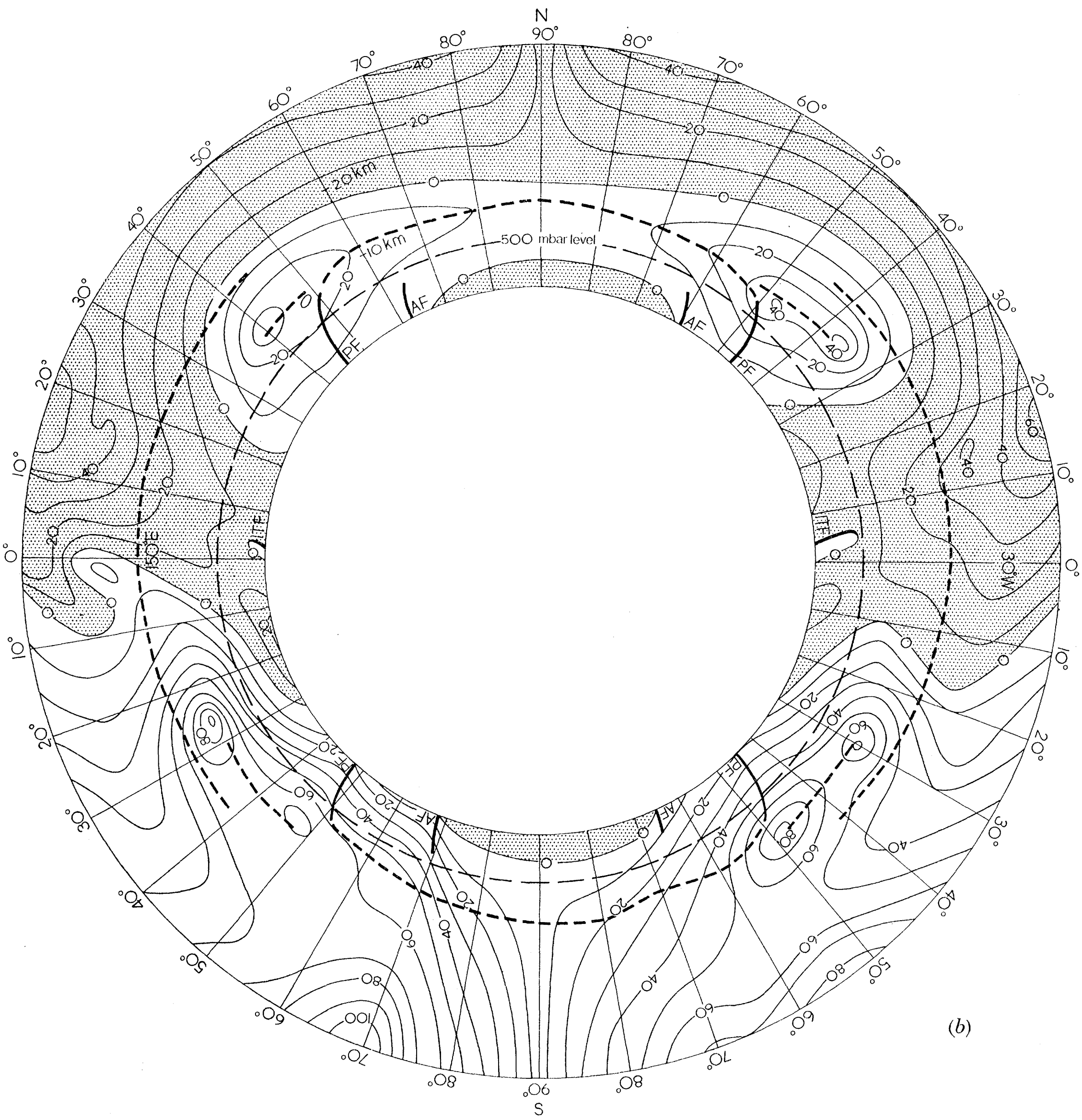


FIGURE 9. (*a*) Vertical cross-section between north and south poles in 150° E and 30° W from 0 to 30 km height, showing mean zonal wind components (knots) in January. Regions of prevailing east winds shaded, west winds clear. The height scale is linear.

wind speeds are generally over 13 m s^{-1} (25 knots) over all the middle latitudes for most of the year.

(*c*) The meandering upper westerlies extend up through the lower stratosphere, from the tropopause at 10 to 12 km up to 15 to 20 km at all seasons, though the wind speeds decrease with height and the meanderings range rather less widely. At the level of minimum wind, about 20 km, speeds are commonly between 0 and 25 m s^{-1} (0 and 50 knots).

(*d*) Above about 20 km westerly winds, increasing with height, prevail in the winter half year and are strongest around the fringe of the region of polar night; easterly winds prevail in the summer half year. At heights of 30 to 35 km wind speeds of 25 to 50 m s^{-1} (50 to 100 knots) are again common and about the 60 km level very strong zonal winds from 50 to 100 m s^{-1} (100 to 200 knots) are believed to be often present (Murgatroyd 1957).



(*b*) Vertical cross-section at 150° E and 30° W between north and south poles from 0 to 30 km height, showing mean zonal wind components (knots) in July. Regions of prevailing east winds shaded, west winds clear. The height scale is linear.

Sudden breakdowns of the winter stratospheric westerlies affecting a broad range of latitudes between about 30° and the pole, starting on one or other side of the hemisphere, are observed in the northern hemisphere at dates between about December and April, and are accompanied by subsidence and warming in the centre of the disturbance leading to a summer-type easterly régime in the stratosphere (Hare 1962; Scherhag 1959). This summer-type régime may or may not persist. Early breakdowns are usually followed by reversions to a winter-type circulation—i.e. renewal of a circumpolar westerly vortex—and this alternation may be repeated several times, though more and more feebly. In these cases final establishment of the summer régime comes late (late April–May). If no breakdown of the winter stratospheric westerlies occurs until late March or April, the summer régime is more likely to set in finally at that time (Gaigerov 1967). The changes in the stratospheric circulation over the Antarctic and the southern temperate zone during the southern hemisphere spring are less sudden and come later in the season, typically over the months September to November (Lamb 1960; Phillpot 1964).

As the moving air passes through the area affected by one of these winter (or spring) stratospheric warmings it first descends, commonly by 1 to 4 km, and then regains approximately its former level. These events may therefore, through the medium of horizontal eddy motion and mixing with neighbouring air at each stage, constitute a process which tends towards vertical expansion of any dust layer present.

(*d*) *Evidence of transport and spread of material in the prevailing zonal windstreams*

The probable times taken for material travelling with the general westerly winds in the upper troposphere over middle latitudes to complete one circuit of the globe were deduced by Angell (1960, 1961), and Giles & Angell (1963), studying (i) the course of constant height balloons ('transosondes') around the northern hemisphere at the 300 and 150 mbar levels (10 to 14 km) from release over Japan; and (ii) geostrophic wind trajectories at 500 mbar (5 to 6 km) from various points in the southern hemisphere temperate zone (figures 10*a*, *b*).

Figure 10*a* shows the mean progress during 10 days of flight from releases near 35° N over Japan from September onwards through the winter months and on to June in 1957–8 and 1958–9. Figure 10*b* shows the modal trajectories around the southern hemisphere for 14 days from hypothetical releases at Perth, Christchurch, Stanley (Falkland Isles), Cape Town and the South Pole in two winter months and one summer month. Dust in multiple layers should be expected to show a greater scatter of distances covered and to spread rather more rapidly across the wind zone into which it was injected.

From figures 9 and 10 it appears that volcanic dust should take on average 2 to 4 weeks to circuit the globe in the upper troposphere over temperate and higher latitudes and on average probably 3 to 6 weeks to do so over the equator (where winds are lighter and the circumference of the Earth is greater but the air tracks meander much less and are nearly zonal). Times required at 20 km are likely to be twice as long over middle and higher latitudes, but over the equator often to differ little from the times taken in the upper troposphere. At greater heights the transport is faster again and at 30 km or above is likely to be similar to, or faster than, in the troposphere. An exception to this faster transport in the middle and upper stratosphere occurs over the equatorial zone whenever volcanic dust is caught in a level where the air is stagnating in the course of a change-over from the easterly to the westerly régime or vice versa. The Krakatau dust cloud in fact circuited the equatorial zone in about 2 weeks in August to September 1883 at heights perhaps mainly between 20 and 45 km, carried by the easterly winds then prevailing in the equatorial stratosphere. The Bali eruption dust seems to have taken about 4 to 5 weeks to make the first circuit in 1963. In both cases the equatorial stratospheric winds were from the east. From these considerations taken together with Pasquill's diagnosis of lateral spread (summarized below) it appears that the dust from one volcanic explosion is likely to spread in the course of its first circuit of the globe as a veil across the whole breadth of the wind zone in which it is embedded, and it is likely to become a fairly uniform veil simultaneously covering the whole latitude zone within 1 to 4 months after the eruption.

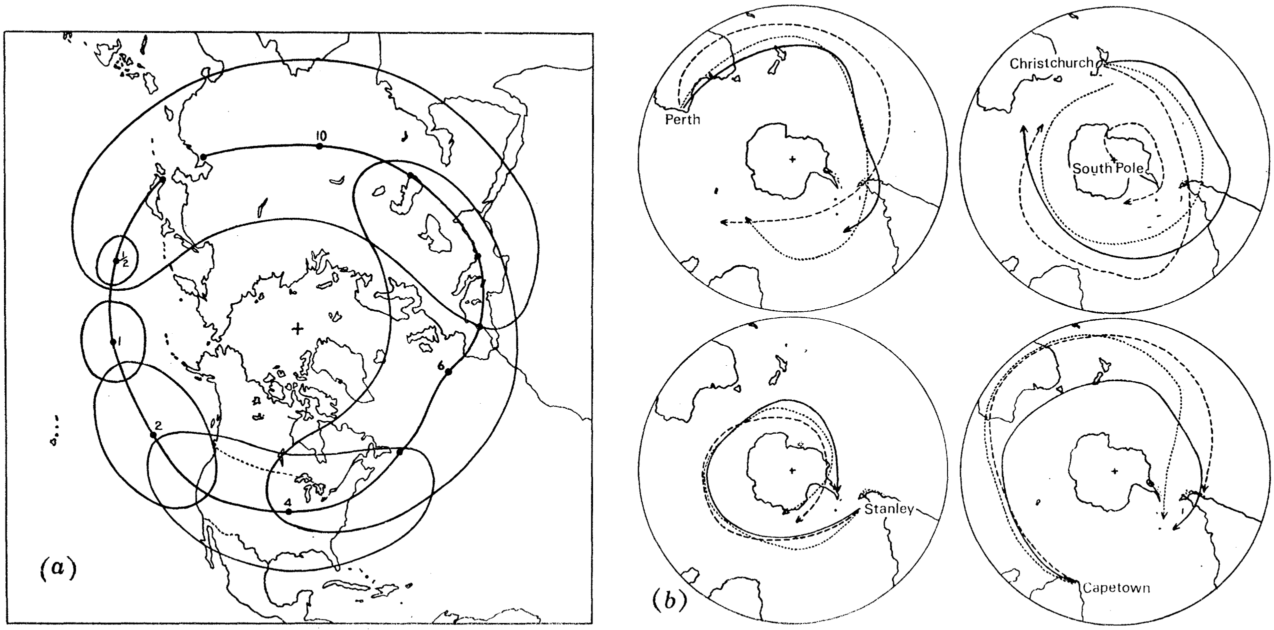


FIGURE 10. (*a*) Mean trajectory of constant-pressure-level balloon transosondes at the 300 mbar level from Iwakuni, Japan in 1957–58 and areas (the ellipses) within which observations and computations indicate that 50% of the transosondes would be found a given number of days after release (after Angell 1961).

(*b*) Modal 500 mbar trajectories from Perth, Australia; Christchurch, N.Z.; Port Stanley, Falkland Islands; and Cape Town for July 1957, January and July 1961; and from the South Pole for July 1957 (after Giles & Angell 1963). – – –, July 1957; ——, January 1961; . . . . ., July 1961.

The Krakatau dust cloud in 1883 was observed to take thee to four days to pass over places in the equatorial zone after its second circuit. This implies that it had developed an extent of about 10000 km along the windstream (about 0.13 of the distance travelled) and that it covered a quarter of the length of the initial latitude zone at any given instant by one month after injection.

Pasquill (1962, p. 267 *et seq.*) has discussed the spread of nuclear bomb test debris across the prevailing windstreams, partly on the basis of aircraft observations (Machta *et al.* 1957) of the pollution surveyed along the 95 and 80° W meridians over north America during the first 3 days after, and up to 3200 km or so down-wind from, tests in Nevada at 37° N, 116° W in 1951, and partly by reference to diffusion theory. The quantities of debris found at these two north-south sections across the prevailing westerly winds around the 10 km level in the upper troposphere showed a nearly Gaussian distribution. The width of the debris cloud, taken between the points on either side where concentration fell to one-tenth of the peak concentration in the centre, was found to average 0.15 of the distance travelled from the point of origin. This lateral spread was an order of magnitude greater than that observed in the case of the constant height balloons, possibly because of the depth of the debris layer and the variation of wind with height within it. The rate of lateral spread expressed as a ratio of the distance travelled is of the same order of magnitude as in the three volcanic dust instances already discussed (Askja 1875; Krakatau 1883; Hekla 1947). It may therefore apply, without too much adjustment, to spread of the dust both in the upper troposphere and perhaps in the stratosphere also. Probably a spread across the wind of between 5 and 20 % of the distance travelled may be expected, at least as long as this implies no spreading of the dust beyond the particular zone of the world's wind circulation into which it was injected. A similar scatter of distances covered in the line of the windstream, fore and aft of the centre of the dust concentration, should also be expected. Lateral spread may commonly be somewhat slower within the equatorial zone owing to less horizontal eddy motion in the upper windstreams there than in other latitudes. This short of difference may also apply to the stratosphere generally as compared with the troposphere. W. Schuepp (personal communication 1965) reported that after the volcanic eruption in Bali in February to March 1963 optical effects were not noticeable in the Congo until after April, although other observers had seen the effects farther south in Africa as early as 19 March.†

Eaton (1963) has mapped the axial directions of tephra deposits from 24 recent volcanic eruptions all over the world and the frequencies of different directions in the case of 244 dust deposits from the volcano, Asama, Japan. These constitute a sort of 'fossil' record of the upper winds at the time of each eruption. Though the wind layer most effective in producing the deposits is somewhat indeterminate, it is probably the wind in the upper troposphere and lower stratosphere: the prevailing circumpolar upper westerlies of extratropical latitudes in both hemispheres are the most prominent feature of Eaton's world survey and secondly the alternating easterlies and westerlies near the equator. Eaton points out that as the volcanic dust deposits remain substantially intact, easily recognized and in many cases approximately datable in the ocean-bed sediments and in geological strata of all ages, this offers an indicator of upper wind régimes in prehistoric times and in the geological past.‡

There is evidence that most progress of volcanic dust into other latitude zones is made during the great seasonal circulation changes. These are normally accompanied by meridional motions entailing great shifts of mass of air into and out of the region of the polar caps in both stratosphere and troposphere about autumn and spring respectively, though some vacillating movements occur in the course of the changes and the dates of the major shifts in the Arctic stratosphere may be advanced or delayed by a month or more in some years. In the troposphere 'blocking' of the middle latitudes westerlies by large amplitude waves with marked cross-latitude ('meridional') flow aloft, and stationary anticyclones or cyclones at the surface, occurs most frequently in March to May and September to November; the lower stratosphere shares in these patterns. About the same time there is a gradual shift of mass of the atmosphere from the northern to the southern hemisphere, and from the continents to the oceans, in the northern

† The winds in the lower stratosphere over the equatorial zone, as represented by Ascension Island (8° S, 14½° W), between February and April 1963, were from the east at all heights up to about 30 km. At, or just above, this height a changeover to westerly winds occurred soon afterwards.

‡ Schwarzbach (1961, pp. 70–72) quotes a number of examples, e.g. the general indications of north-westerly upper winds given by the dust fans over distances up to 500 km resulting from the volcanic eruptions in western Germany (Eifel) about 10000 to 8000 B.C. (See also Eaton 1963 and Gill 1961, p. 341.)

Survey of the dust and tephra deposits from all these known eruptions of both modern and prehistoric times makes it appear highly improbable that dust from a single volcanic eruption ever covered more than 1 to 2 % of the surface of the Antarctic ice cap to a depth of over 1 cm. And this would be soon buried by fresh snow. Thus, volcanic dust seems unlikely ever to have affected the melting of the ice cap enough to influence world sea level (Lamb 1968).

spring and early summer, and back again in the northern autumn and early winter. (On the other hand, the Asian monsoons involve much interchange of the lower air across the equator in that sector, culminating about January and July, but this seems to be mainly in the lower troposphere.)

(e) *Mean meridional drifts—transfer to other latitude zones*

(i) *General*

The meridional components in the mean upper wind circulation are difficult to observe directly, because they are weak. Knowledge has therefore been built up partly by the observation of any substances carried which serve as tracers and partly from theoretical derivations using conservative properties of the air itself such as potential temperature and angular momentum.

Net poleward drift of the air in the lower stratosphere was mooted by Brewer (1949), following a much earlier suggestion by Dobson, to explain the observed distribution of water vapour and ozone at the levels concerned. It has been supported by some, but not all, more recent studies of other tracers (see, for example, Kulkarni 1966) as well as by some theoretical approaches (see, for example, Murgatroyd & Singleton 1961; Faust & Attmannspacher 1961). It is clear, however, from the many studies of nuclear fission products in recent years that eddies diffuse material in both directions from the latitude of origin. It is also clear that improbably large west wind velocities relative to the Earth would be acquired, through conservation of angular momentum about the Earth's axis, by air carried from the equator to the pole in any continuous circulation. A convenient survey of the existing state of knowledge of meridional transport and exchanges within the stratosphere, excluding the evidence of volcanic dust, was given by Sheppard (1963).

More recent theoretical work by Murgatroyd (1969) adds the following details. The mean circulation in the lower stratosphere is probably in three cells: (1) from the equator towards middle latitudes, (2) from higher latitudes towards middle latitudes, and (3) from the higher latitudes towards the pole—the average horizontal components in each case probably a few kilometres a day, the vertical components of few tens of metres a day (generally upwards near the equator, downwards over middle latitudes and near the pole). There are preferred directions for eddy diffusion within the stratosphere, namely downward and poleward or upward and equatorward. The proposed three-cell structure of the mean meridional circulation and the preferred paths of eddy transfers are illustrated in figures 11*a*, *b* both from Murgatroyd's work.

Thus our emerging understanding of meridional transports in the stratosphere, partly from the study of tracers, including volcanic dust, and partly from theoretical considerations, allots a place both to eddy transports and to a slow mean vertical meridional circulation, as illustrated in figures 11*a* and *b*. The direction of the mean circulaion in the stratosphere below 30 km is generally poleward over low and high latitudes but equatorward over middle latitudes, between 30 and 60° according to Murgatroyd but rather between 50 and 75° from the diagrams derived from a different partly theoretical approach by Vincent (1968). There seems, however, from Vincent's study to be a complex sequence of month by month changes, in the course of which the mean meridional drift is poleward over all latitudes in some months in late summer and autumn and is equatorward over all latitudes in the northern hemisphere in May: on the whole, it appears to be poleward over most latitudes during most of the year, though usually in three cells as indicated by Murgatroyd. Outside the tropics the eddy transfers, which operate in both directions, are, however, at present thought to be more important than the mean circulation.

(ii) *Examples from tracers other than volcanic dust*

Numerous studies of the spread to other latitudes, outside the initially affected wind zone, of the products of nuclear fission such as the radioactive isotopes of tungsten ^{185}W, strontium ^{90}Sr, manganese ^{54}Mn, carbon ^{14}C, hydrogen ^{3}H (tritium) etc. in the fall-out after bomb tests

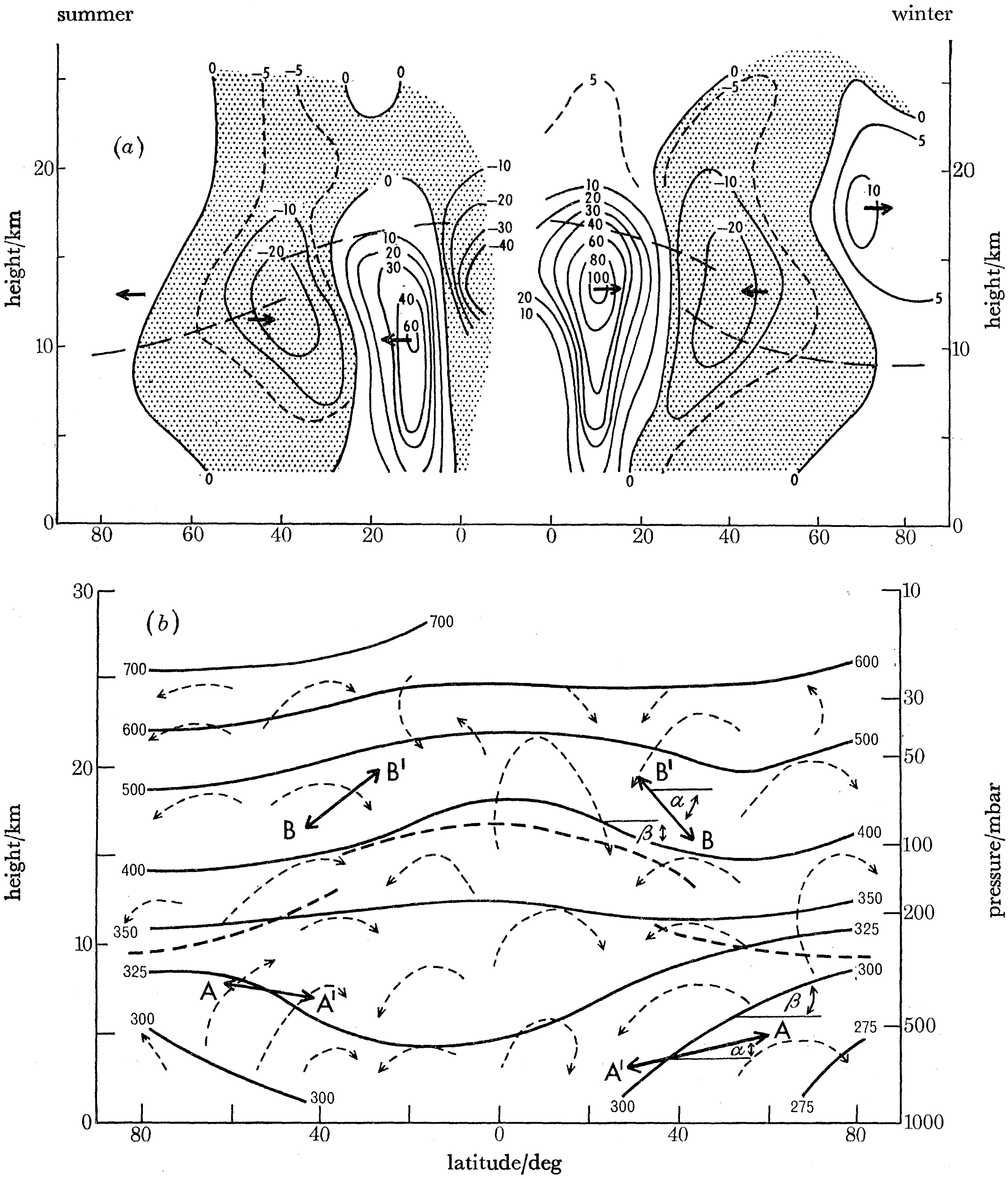


FIGURE 11. (*a*) Mean meridional drift in summer and winter in the atmosphere between the 3 and 25 km levels (as calculated by Murgatroyd 1969). Shaded areas—equatorward drift; meridional velocities in centimetres per second.

(*b*) Vertical-meridional structure of the atmosphere in summer and winter, showing prevailing tropopause heights, potential temperature isopleths (in kelvins), directions of the mean circulation and of preferred eddy motions. (After Murgatroyd, unpublished. Reproduced by kind permission.) —— Potential temperature isopleths (K); – – –→, mean circulation streamlines; ↔ preferred eddy transfers; — — —, tropopause.

in the 1950s and 60s, as well as natural tracers such as ozone and water vapour in the stratosphere, have contributed greatly to our knowledge of the general wind circulation in the stratosphere (see, for example, Feely & Spar 1960; Goldsmith & Brown 1961; Davidson, Friend & Seitz 1966; Feely, Seitz, Lagomarsino & Biscaye 1966; Martell 1968). Consideration of the shape of the isentropic surfaces (e.g. Sheppard 1963) also aids understanding of the probable paths of eddy motion, but may not be fully applicable in regard to ozone, water vapour and volcanic dust which absorb solar radiation.

The effectiveness of both eddy transport and some mean poleward drift could be sensed from the latitude distribution of the tungsten isotope ^{185}W in November to December 1959, a year and a half after the only occasion when this element had been put into the atmosphere. The greatest concentration remained near the equator, at 21 km height over latitudes 0 to 5° N, not far from the latitude of the original injection of 12° N over the western Pacific; but there were secondary maxima at heights of 15 to 18 km over latitudes 50 to 90° in both hemispheres, amounting in each case to about one third of the concentration then present over the equator.

Evidence of poleward transfer of nuclear fission products may be seen here in figures 12*a* and *b*. Most rapid decline of the concentration of bomb test pollution occurs in each of these cases in the equatorial zone, presumably through washing out of the bottom 17 km in the equatorial rains. This does not, however, touch material suspended in the stratosphere. There is no rain in the stratosphere, and transference of air from the stratosphere to the troposphere takes place mainly over middle and higher latitudes in the tropopause breaks near the jet stream.

Figure 12*c* illustrates a situation brought about by injection of material at a high northern latitude. The strontium-90 distribution early in 1962, following injection in Soviet tests near 75° N the previous year, shows the main concentration between 60° N and the pole but substantial amounts well mixed everywhere north of latitude 30° N; much smaller amounts have spread, at higher levels, to the equator. Figure 12*d*, showing the strontium-90 concentration at all latitudes from pole to pole over the three following years, suggests that there may have been a small seepage of this material into the southern hemisphere; though the diagnosis is complicated by further ^{90}Sr injected near 12° N in United States bomb tests after April 1962. The main equatorward spread of the Soviet test material from 75° N is undoubtedly accounted for by eddy transport within the range of latitudes reached by the meandering westerlies. Seepage of much smaller amounts as far as the equator and beyond may be partly due to mixing by such eddy transports as occur in low latitudes and partly to the limited zone of trans-equatorial flow of the mean circulation seen in figure 11*b*. The seepage into the southern hemisphere could also be due to a mean drift in the *upper* stratosphere, which above about 30 km may be directed all the way from the polar regions of the summer hemisphere to the polar regions of the winter hemisphere: a mean drift in this sense was derived theoretically by Murgatroyd & Singleton (1961, fig. 3, p. 132) and is supported by studies (List, Salter & Telegadas 1966) of tracers (rhodium ^{102}Rh and cadmium ^{109}Cd) experimentally injected at the levels concerned.

(iii) *Volcanic dust*

The evidence of the volcanic dust veils listed in this paper seems entirely consistent with the above indications obtained from other tracers. Volcanic dust injected into the stratosphere in low latitudes produces veils which, if dense enough for easy detection (and probably for significant meteorological effects), slowly spread over the whole Earth, though the advances into latitudes poleward of about 35° notably seem to take place in steps and particularly in (late) autumn. Poleward drift with a mean circulation is also indicated by an apparent tendency for the veils to clear first over low latitudes. Dust injected into the stratosphere in high latitudes tends to spread as a veil over all latitudes between the pole and about 30° in the hemisphere concerned, but has never so far been reported as a readily detectable veil or capable of dimming the Sun anywhere farther from the source than that.

The report on the Krakatau 1883 eruption (Royal Society 1888) was much less specific about the clearing of the dust veil from the low latitudes in which it originated than about its first sightings in higher latitudes. The durations of that veil here quoted (p. 441) have been derived from the dates of the individual observations given in the report. That the analysis of all the evidence by the original investigators did indicate a drift away from the equator appears from passages on pp. 323 and 325 of the report, where the course of the drift is tentatively described and the cessation of observable effects near the equator is mentioned.

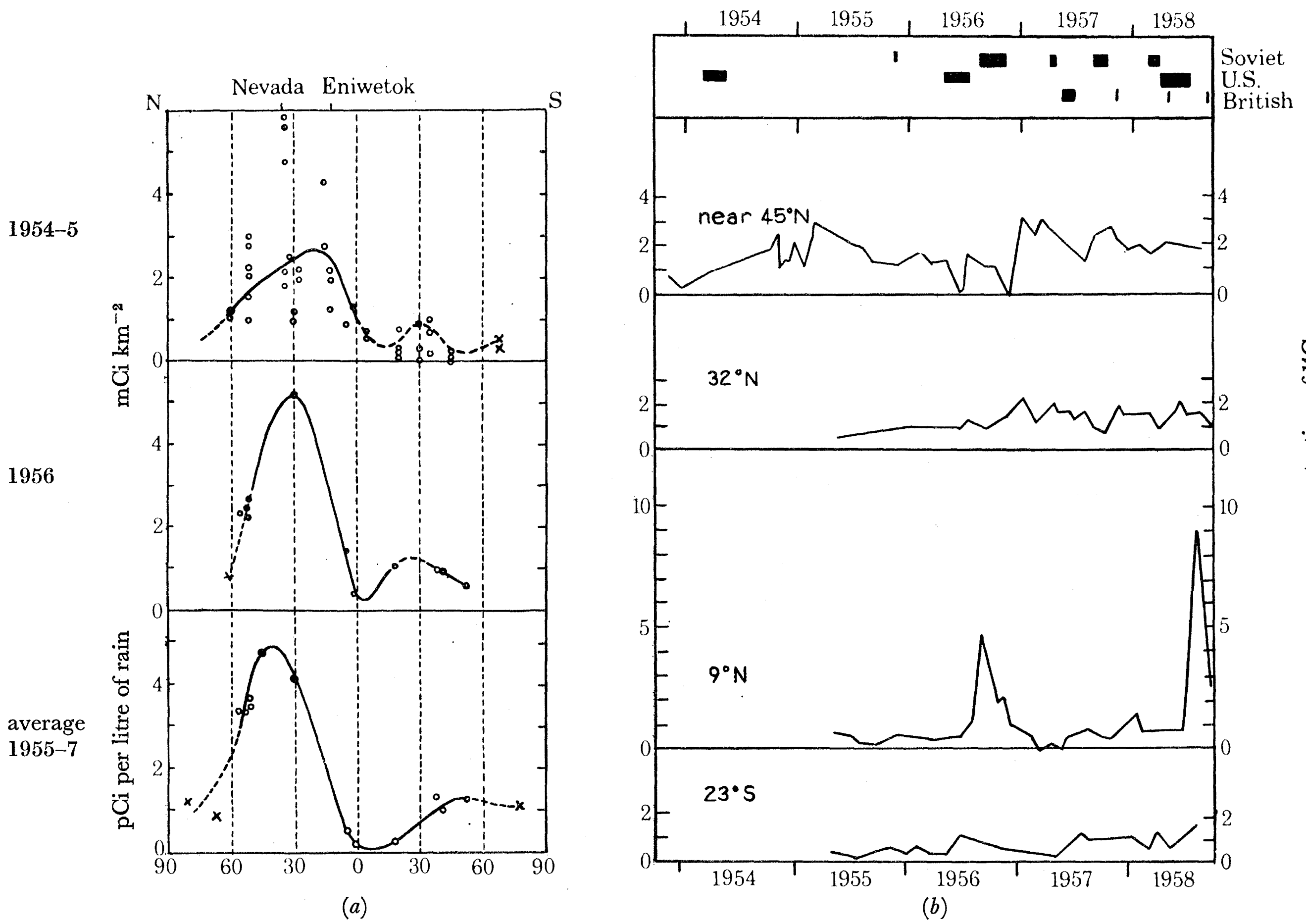


FIGURE 12. (*a*) Latitude distribution of strontium-90 fall-out (after Flohn 1959 and partly from Stewart *et al.* 1957); the top two graphs show total deposition in mCi km^{-2}: the bottom graph shows the concentration of strontium-90 in picocuries per litre of rain. Comparable profiles of total nuclear fission product concentrations by latitude in surface air in 1959 and 1960, after further bomb test explosions, have been given by Newell (1961).

(*b*) Excess atmospheric concentration (above the normal concentration produced by cosmic ray bombardment) of radioactive carbon-14 in units of 10^7 atoms per gramme of air sampled at heights near 20 km. The injections by bomb test series are indicated by solid bars at the top of the figure. The Russian injections were north of 40° N, the American injections at 10–12° N, the British near the equator (2° N). (From a figure given by Hagemann *et al.* 1959.)

Mossop's (1964) collections by aircraft at 20 km of stratospheric dust over Australia along the 145° E meridian after the 1963 Bali eruption showed in the first year greater concentrations at 15 to 35° S than farther south, but in April 1964 the greatest concentration was at 40 to 45° S.† The dust from the eruptions of Krakatau in May and August 1883 and Agung (Bali) in

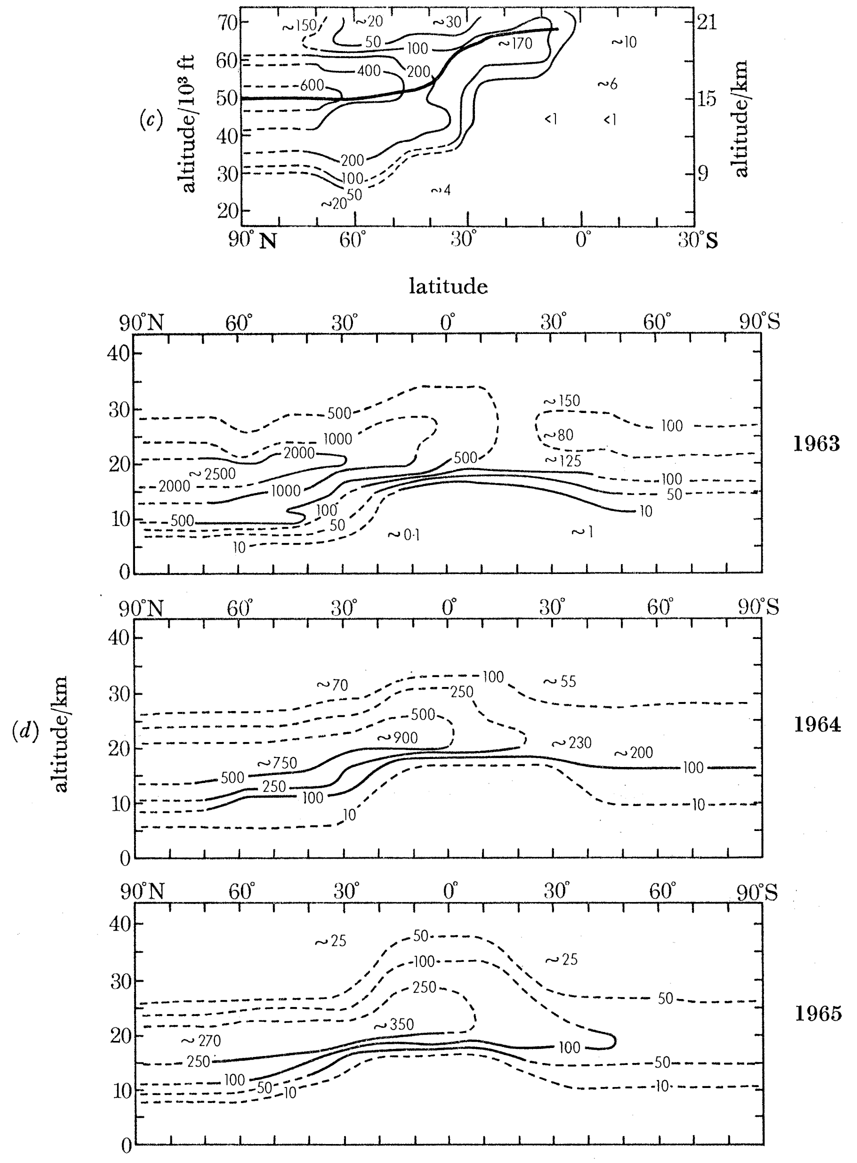


Figure 12. (c) Distribution of strontium-90 in the atmosphere by height and latitude, January to April 1962, about half a year after Soviet tests near 75 °N (after Davidson *et al.* 1966). Units: disintegrations of ^{90}Sr per minute per 1000 s.c.f. (standard cubic feet of air, i.e. cubic feet of air at 1013 mb and 15 °C).

(*d*) Distribution of ^{90}Sr in the atmosphere by height and latitude in January to April 1963, 1964 and 1965 (after Feely *et al.* 1966). Units: as in 12 (*c*). The ^{90}Sr particles have an estimated median residence time in the atmosphere of about 10 months. The distributions here shown are attributed partly to Soviet tests near 75° N in 1961 and 1962 and partly to American tests near 12° N in 1962.

† The Bali dust was first picked up in sub-stratospheric air samples at 15–35° S over eastern Australia in April 1963, but observations at Melbourne (38° S) indicated increasing numbers of particles after that. Particles with diameter > 1 μm decreased after July 1963 but those with diameter about 1 μm and smaller were increasing in number up to April 1964, the main concentration being at height 20 to 22 km (Harris 1964).

February and March 1963, both in the equatorial zone, seems to have spread rather quickly as far as about 35° N and 35° S where the first traces of it were seen within 8 to 12 weeks, but was much delayed in reaching higher latitudes than that. It was first obvious over Europe about 50° N late in the following autumn in both cases, though Volz (1965) reports evidence of a slight amount of Bali dust over Europe as early as mid-May 1963. Figure 13, taken from Wexler (1951), indicates how in 1883 (the Krakatau case) the spread of volcanic dust originating in low latitudes to regions north of about 35° N is likely to have been influenced by the normal

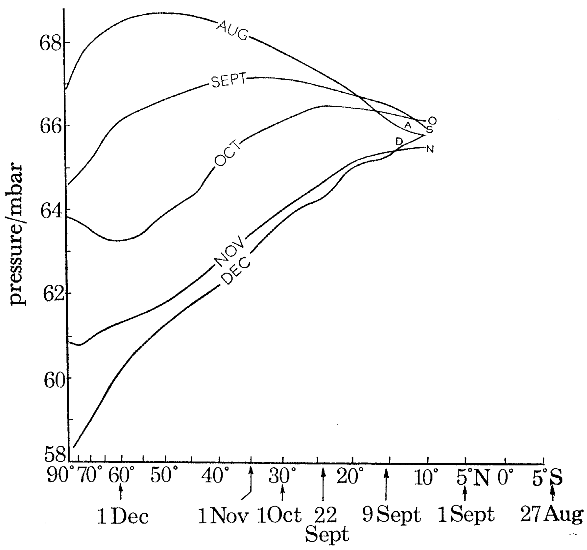


FIGURE 13. Average pressure profiles by latitude, 80–10° N, at 19 km in the stratosphere month by month, August to December. Arrows beneath the latitude scale indicate northernmost limits of the dust from the Krakatau eruption in 1883 from August to December. (As given by Wexler 1951.)

seasonal change of the stratospheric circulation in autumn. Measurements by Dyer & Hicks (1965, 1968 and verbal communication, October 1966) of direct, diffuse and total solar radiation, and particularly study of the contribution of the diffuse radiation to the total, at Aspendale, Melbourne (38° S) indicated the arrival of the Bali volcanic dust in April to June 1963, followed by some decline of concentration and then further abnormal turbidity (dust) arriving there in each of the following southern autumns—i.e. about April in 1964, 1965 and 1966. Presumably, transfer of dust from lower latitudes was repeated in 'pulses' each autumn, though each 'pulse' was weaker than the one before it. Observations of the direct solar radiation each southern summer at the South Pole (Flowers & Viebrock 1967) indicated the presence of some abnormal turbidity by November 1963 and a sharp increase (decrease of radiation) in the first days of December 1963; the loss of direct radiation indicated that greatest concentration of the dust over the South Pole was attained by about October 1964—allowing for impossibility of observation during the winter darkness, say 12 to 18 months after the eruption. After that a slow recovery was under way, but the measured radiation values were still below the 1961–2 level in January 1966.

Dyer reports (personal communication 1968) that 5 years after the Bali 1963 eruption the dust still left in the stratosphere had become more or less uniformly distributed over all latitudes, though this would mean that depletion of the solar radiation would be greatest over high latitudes because of the low elevation angle of the incident beam. Indeed, any significant effect might be confined to high latitudes. Despite the near uniformity of the total amounts of dust in the air columns, it was spread more thinly through a greater range of heights over high latitudes. This situation, the product of 5 years of the mean meridional drifts and eddy diffusion, had been reached despite presumably greater losses of dust from the stratosphere through the tropopause gaps near the jet streams over middle and high latitudes than any losses from the stratosphere nearer the equator.

By contrast, there seems to be no recorded case of a volcanic dust veil from an eruption in Iceland or elsewhere north of 55 to 60° N ever having been noticeable south of about 30° N. The dust has been known to spread over the whole breadth of the zone of meandering circumpolar westerlies as far south as north Africa and Syria, but not in any readily observable quantity nearer the equator. Similarly, eruptions of Antarctic volcanoes seem to go undetected in terms of any visible dust cloud or significant radiation effect north of 40 to 45° S, as shown by the known activity over the last 150 years of Erebus, Deception Island and the South Sandwich Islands and the suspected great eruption of Thompson Island in 1895 or 1896 (Lamb 1967).

4. Other types of dust carried by the atmosphere

Other types of dust and solid particles, including sea salt and smoke, are transported long distances by the atmosphere. Saharan sand has been deposited in observable quantities on ship's decks as far away as 46° W in the tropical Atlantic and on other occasions in southern Scandinavia—i.e. after going at least one-twelfth of the way round the world or 3000 to 3500 km by the shortest air tracks. Smoke trails from forest fires, and from great centres of population and industry, can also be traced over great distances (cf. p. 431). Blowing dry soil, especially light loess soils in Eurasia and north America south of the ice margin during great ice ages, and all sorts of particulate pollution including the dust of modern wars may also come in question. Wexler (1936) paid some special attention to this aspect of the years of rapid soil erosion by wind in the United States in the early 1930s, finding that although particle counts (including smoke) in the surface air were greatest in autumn and winter total atmospheric turbidity was greatest in spring when the prevalent instability allowed dust to be readily carried up to high levels and distributed throughout the troposphere. Ångström (1929) quotes measurements of several independent investigators which indicate the most prevalent dust and smoke particle sizes as from about 0.2 to 1.5 μm in the lower atmosphere, much the same range as for volcanic dust. But these dust and smoke trails are all trivial in extent and duration compared with a persistent stratospheric dust veil covering many latitudes.

None of the natural agencies mentioned can have raised dust into the stratosphere. Any approach to comparability with a volcanic dust veil can only have arisen when the supply of dust within the troposphere was sustained over long periods—e.g. loess driven by strong winds near the margins of the great inland ice domes of the ice age. And even then the effect on radiation should differ from, and be less than, that of volcanic dust in the stratosphere. Solar radiation intercepted in the troposphere must be to some extent effective in warming the surface layers, and only that radiation which is reflected back to space may be altogether lost.

5. Radiation effects of volcanic dust, including indirect evidence from surface air temperatures and polar ice

In this section available theoretical and observational evidence of the effects of a volcanic dust veil, on both short wave and long wave radiation, are discussed. Depletion of the direct, solar beam is to a considerable extent made good by forward scattering, which increases the intensity of diffuse radiation received at the surface of the Earth from the sky: the ratio depends on the angle of incidence of the solar beam, so that the net depletion is greatest in high latitudes. It is shown that the presence of known volcanic dust veils at various times in the last 100 to 300 years can be clearly traced in the available records of intensity of the direct solar beam and of temperatures averaged over a large part of the Earth. The number of eruptions which can be examined in this connexion is sufficient to indicate the characteristic magnitudes of the effects and their duration.

Any veil of dust or other material in the atmosphere must tend to screen (i.e. intercept, partly absorb and partly reflect) both solar and terrestrial radiation. The Earth's radiation balance and temperatures prevailing in the lower atmosphere are likely to be affected by the difference in the screening power of the dust veil with respect to incoming short-wave radiation from the Sun, with its maximum intensity about 0.5 μm, and outgoing long-wave radiation from the Earth, with its maximum intensity about 12 μm.

There is some direct evidence suggesting abnormal interception of solar radiation in the stratosphere after the Bali eruption in February to March 1963. Monthly mean temperatures at the 20 and 25 km levels over Ascension Island (8° S, 14½° W), which undergo a 26-month oscillation like the stratospheric winds over the equatorial zone, but have a marked annual variation superposed with highest values each September, rose in 1963 to values that had never been reached since the observations began in 1958 and were 3 °C warmer in September 1963 than any previous or subsequent maximum. The coldest month at those heights during the following 12 months, March 1964, produced a monthly mean temperature 4 °C above any previous minimum and has not been approached since (written June 1968). Positive anomalies seem to have continued, fading away, till about February 1965 (R. A. Ebdon personal communication).

As regards the lower atmosphere after the Bali 1963 eruption, Scherhag (1965) found a temperature anomaly of -0.24 °C from the 1931 to 1960 averages for the whole northern hemisphere atmosphere from the surface up to 16 km and -0.40 °C for surface temperatures over the whole world for the 12 months December 1964 to November 1965. If part of this deficit be attributed to the downward climatic trend in progress since about 1940 (Mitchell 1961), the remainder cannot be statistically significant: the standard deviation of annual values of surface temperature averaged over the world is probably between 0.2 and 0.3 °C, when due allowance is made for the great areas of ocean. It seems likely however, that the deficit in 1963 to 1964 below that of the latest dust-free year was rather bigger, about 0.5 °C.

Humphreys (1940, p. 594) gives the intensity I_x of the solar beam after passing through a thickness x cm of a layer of sparsely distributed dust particles of radius r cm numbering n per cubic centimetre as

$$I_x = I \exp(-2n\pi r^2 x), \tag{8}$$

where I is the initial intensity outside the atmosphere.

The reduction of the direct solar beam occurs because the effective dust particles are quite large in comparison with the wave lengths represented, over three-quarters of the solar intensity being at wave lengths less than 1 μm. From Humphreys's formula, a mass of 10 km^3 split into spheres of 1 μm diameter, as in a big volcanic eruption, should deplete the solar beam by several parts per cent.

The sizes of the particles in any persistent volcanic dust veil are, however, much smaller than the wavelengths of most of the outgoing terrestrial radiation. The dust particles therefore effectively cannot reflect the outgoing radiation, but only scatter it. Following Rayleigh's laws of scattering, Humphreys (1940, p. 596) gives the intensity E_y of the terrestrial radiation after passing through y cm of the dust layer by an expression that reduces to

$$E_y = E \exp(-6000\, nr^6 y/\lambda^4) \tag{9}$$

where E is the initial intensity and λ the wave length.

By comparing the thickness of dust layers which should reduce the solar and terrestrial radiation in the ratio 1/e, using expressions (8) and (9), Humphreys estimated that volcanic dust was 30 times as effective in depleting the solar radiation as it was for terrestrial radiation. No adequate observational verification of this estimate has so far been possible. It appears from this, however, that volcanic dust must have a powerful 'reverse greenhouse' effect, favouring the transmission of outgoing radiation rather than the incoming.

Humphreys's calculations were based on the assumption that the volcanic dust particles in the veil are generally about 1.85 μm in diameter. If now we try the effect of substituting other diameters in his formulae, we find that the ratio of depletion of incoming to outgoing radiation varies as in table 4 below.

Table 4. Depletion ratio for different particle sizes

particle diameter (μm)	1	1.85	4.3
depletion ratio: solar/terrestrial radiation	300	30	1

These figures indicate a strong 'reverse greenhouse effect' for the most frequently 'observed' sizes of volcanic dust, whether or not they are enlarged by a fluid accretion (see p. 440). However, at sizes of about 1 μm and less the situation alters because another process, namely the scattering rather than stopping of all the longer wave lengths in the solar radiation must become important.

Another complication has been pointed out by Fritz (1949). Scattering in all directions evenly, as envisaged by Rayleigh, the forward and backward components in particular being equal, only applies to particles which are very small in comparison with the wave length—under $\frac{1}{10}\lambda$ approximately. With larger particles forward scattering exceeds the energy sent back. Thus, energy in the longer wave lengths in the solar beam, which exceed the size of the dust particles, is more than half sent forwards towards the Earth.

It does not seem possible therefore to express the effects as simply as in Humphreys's brief theoretical treatment. Depletion of the solar beam must be caused mainly by the larger sizes of dust and must in total be somewhat less than indicated by Humphreys's expression (equation (8), above) which ignores the considerable forward scattering of the red and infrared by dust particles of sizes less than about 0.7 μm. Nevertheless, the expectation of a 'reverse greenhouse effect' remains, and it seems likely that there is a most effective dust size for this at particle diameters about 1 to 2 μm.

Varying sizes and abundance of the dust particles mean that no very simple theoretical rule can readily be given to express the depletion of net radiation—i.e. depletion of incoming direct radiation *less* the enhanced sky radiation, and *less* the very slight (probably negligible) back-scattering and reflexion of the radiation given off by the Earth. In the circumstances, further understanding must first be sought from observation.

Budyko (1968*a*) quotes the following average values for different latitudes of the ratio of the depletion of the total incoming short-wave radiation received at the surface to depletion of the direct solar beam:

latitude	90°	60°	30°	10°	0°
ratio to depletion of direct beam	0.24	0.21	0.16	0.13	0.10

These figures refer to particles of some tens to some hundreds of micrometres in diameter, constituting a layer of silicate dust, but are considered to be a reasonable guide to the volcanic dust situation so far as it also concerns particles larger than the wave lengths of the incoming radiation.

Depletion of the incoming solar beam must be greatest in high latitudes, especially over the polar regions, since oblique rays take much longer paths through any horizontal dust layer. There can be no corresponding latitude difference in the way that a dust veil affects outgoing terrestrial radiation. Therefore depression of surface temperatures is likely to be greatest in high latitudes. Also, as noticed in a previous section, dust veils in the stratosphere probably drift towards high latitudes and persist longest there.

The longest series of relevant observations are Köppen's (1873, 1914) computations of average surface air temperature (*a*) over Europe and New England (U.S.A.) for each year from 1750 to 1871, and (*b*) over the northern hemisphere temperate zone from 1811 and over the tropical zone from 1813 to 1910. These are still the only available series of yearly values of average temperature over any large area of the Earth covering the years in question. Budyko (1968*a*) has lately published a similar series (in the form of a graph) for the northern hemisphere for the years 1881 to 1960, and this appears to be in reasonable agreement with Köppen's portrayal of the year-by-year changes between 1881 and 1910. Data for (*a*) were available from the eighteenth century onwards from many stations in most parts of Europe and four or five places in U.S.A. Data for the tropical zone in (*b*) began with Madras only and made use of various short station records of unknown reliability in the early years. Comparison of Köppen's figures with the successive 5-year means from 1840 onwards computed for various latitude zones by Mitchell (1961) from much ampler observation material makes it clear that Köppen's work is to be regarded only as a first approach and that upward or downward tendencies over periods of more than about 5 years are not reliably depicted in it. Nevertheless, it seems certain that where Köppen indicates great temperature differences from one year to the next one, two or three following years, affecting many stations, such differences must be real, at least as regards sign, and some indication of magnitude. These data are reproduced in figure 14: in the case of (*a*), to reduce effects of any unreliability of the early temperature records, Köppen has used a $\frac{1}{4}(a+2b+c)$ smoothing, where b is the value for the year in question, a that for the previous year and c the following year's value. The values indicated by the lower graphs (*b*) are unsmoothed. Even so, the standard deviation of the temperate zone yearly temperatures in figure 14*a* is 0.55 °C, compared with 0.38 °C in figure 14*b*: it seems desirable therefore to reduce temperature departures read off figures 14*a* to $\frac{2}{3}$ of their value, this being the ratio of the standard deviations. This has been done in the computations for the dust veil index assessments in appendix I.

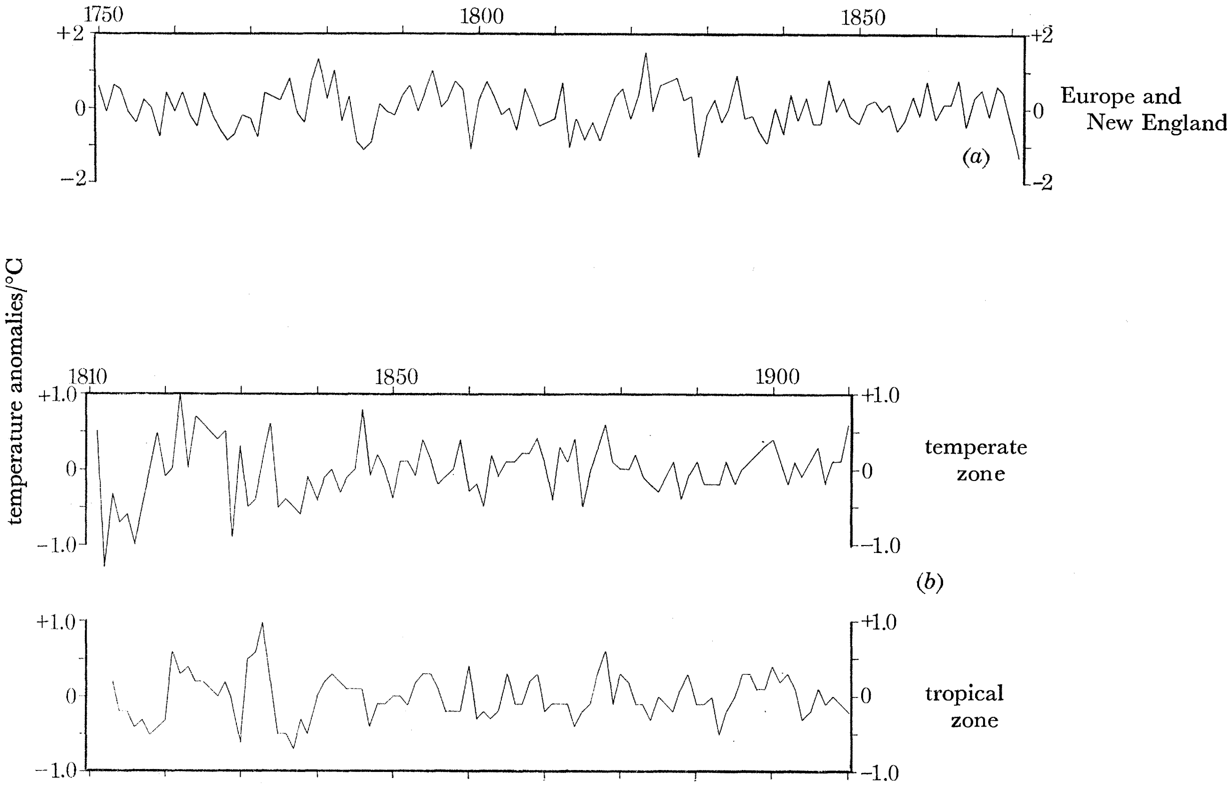


FIGURE 14. Anomalies of yearly mean surface temperature averaged for certain areas (after Köppen 1873, 1914). (a) From 1750 to 1871 for Europe and the New England states in U.S.A., $\frac{1}{4}(a+2b+c)$ smoothing. (b) From 1811 to 1910 for the northern hemisphere temperate zone and for the tropical zone, unsmoothed yearly values. Standard deviations of the yearly values here plotted: (a) Europe and New England, 0.55° C; (b) Northern temperate zone (land stations), 0.38 °C; (b) Tropical zone (land stations), 0.29 °C.

The real—i.e. all around the hemisphere—variations of temperature over both temperate and tropical zones are probably rather less than indicated in figure 14*b* because the observing stations used represented principally the continents: reasonable estimates of the standard deviations of the all-round yearly values would be about 0.3 and 0.2 °C respectively. The strongest features of these curves are the lower average temperatures lasting a year or two after a number of great volcanic eruptions whose dust veils have been recorded by observation, notably those in 1783, 1811–13, 1815, 1835–37, 1875, 1883, 1888–92 and 1902. Temperature drops bringing the average for the whole year down by 0.5 to 1 °C apply to most of the cases mentioned. After 1783 the temperature difference probably exceeded 1 °C. The great Tambora eruption in 1815 was followed by an anomaly of annual mean temperature for 1816 averaging −1 °C in middle latitudes of the northern hemisphere and −0.7 °C for 'the whole Earth' (actually northern hemisphere land data between 13 and 65 °N—Madras to Archangel and some in New England), according to Köppen's estimates. The year 1816 became famous in much of Europe and North America as 'the year without a summer' (Brooks 1949, p. 118; Hoyt 1958). Cyclonic activity seems to have been abnormally concentrated in positions near Newfoundland and from central Ireland across England to the southern Baltic. In Merionethshire there were only 3 or 4 days without rain in the 6 months May to October 1816, and the average temperatures of the summer months in London and other parts of England ranged from 2 to 3 °C below normal. In eastern Canada and New England there was widespread snow between 6 and 11 June 1816 and frosts in each month. Dearth ensued in many countries, amounting to famine in places, e.g. in parts of Wales and northern Ireland. (Some crops did not ripen, others rotted or sprouted in the fields; stocks of flour were exhausted by the end of the year. This led to numerous small farms being abandoned in Co. Tyrone and to begging, vagrancy and food riots in Wales also.) In the 1690s in Scotland there had been much more severe and widespread famine, with deaths and depopulation of the country-side on a massive scale: this resulted from failures of the grain harvest in a run of seven cold summers out of eight between 1693 and 1700. Reported winds and weather suggest that cyclonic activity was concentrated close to southern Scotland and over the North Sea. A volcanic dust veil may well have been involved (see appendix I); the Arctic drift ice was abnormally far south, surrounding Iceland in 1695, and the very cold winds brought Scotland sunless, drenching summers, with 'easterly haars', and sharp frosts and deep snows as the storminess increased early in the autumns. Additional data used by Köppen after 1820 suggested that temperature drops of 0.5 to 1.0 °C probably do apply to the whole Earth in some cases. Mitchell (personal communication) estimates that second-order eruptions are on average followed by a 5 year temperature anomaly of about -0.1 °C.

A study was made of the degree of association between the dates of major dust eruptions (dust veil index values in appendix I $\geqslant 300$) and average temperatures over the relevant latitude zones (figure 14) remaining for 5 years or more continuously below the 'undisturbed level' represented by a $+2\sigma$ deviation i.e. runs of years with no occurrence of $+2\sigma$. Contingency tables using various criteria were drawn up: two out of six χ^2 values calculated appeared statistically significant at about the 4 % level and two others had about a 10 % probability of occurring by chance.

Gentilli (1948) has objected to any supposed effect of volcanic dust on world temperature that the years 1913 and 1922, respectively following reported great eruptions in Alaska and the Andes, were not colder than normal over the world. But these were rather high latitude eruptions. The years concerned were, moreover, within the generally warm climatic period of the early twentieth century: and 5 year average temperatures computed for 10° latitude zones by Mitchell (1963, table 5, p. 177) indicate a progressive cooling by 0.9 °C from 1905–09 to 1915–19 north of 70° N—i.e. against the long-term trend and in opposition to the concurrent changes in almost all other latitudes. Latitudes 50–70° N also shared in the temporary cooling, apparently more briefly (seen in the 1915–19 values), but nowhere south of 50° N. The secular melting of the pack-ice in the highest latitudes on the Arctic Ocean north of 72 to 75° N—perhaps the best known effect of the Arctic warming—was interrupted by a run of seven summers 1912–18 in which the total extent of sea ice at the end of the melting season (August to September) remained on average about 10 % greater than in the previous decade and about 12 % greater than in the 1920s (Lamb & Johnson 1959, p. 124). It is quite possible, however, that the excessive ice of these years should rather be looked on as a delayed result of the dust veils from the eruptions of the previous decade—i.e. in 1902–3 and 1907. Temperatures in 1921 and 1922 after the Andes eruption seem to have shown small deficiencies (anomalies averaging overall -0.1 to -0.2 °C) below the decade means for the 1920s at representative southern hemisphere stations between 5 and 35 °S. (Information is lacking about higher southern latitudes.) The eruptions discussed took place in positions such that the dust veils must only have affected one hemisphere and probably only extratropical latitudes in that hemisphere. In both cases there seems to have been some slight effect upon surface temperatures in the sense expected, but it is clear that the 1921–2 dust veil over the southern hemisphere was not of the

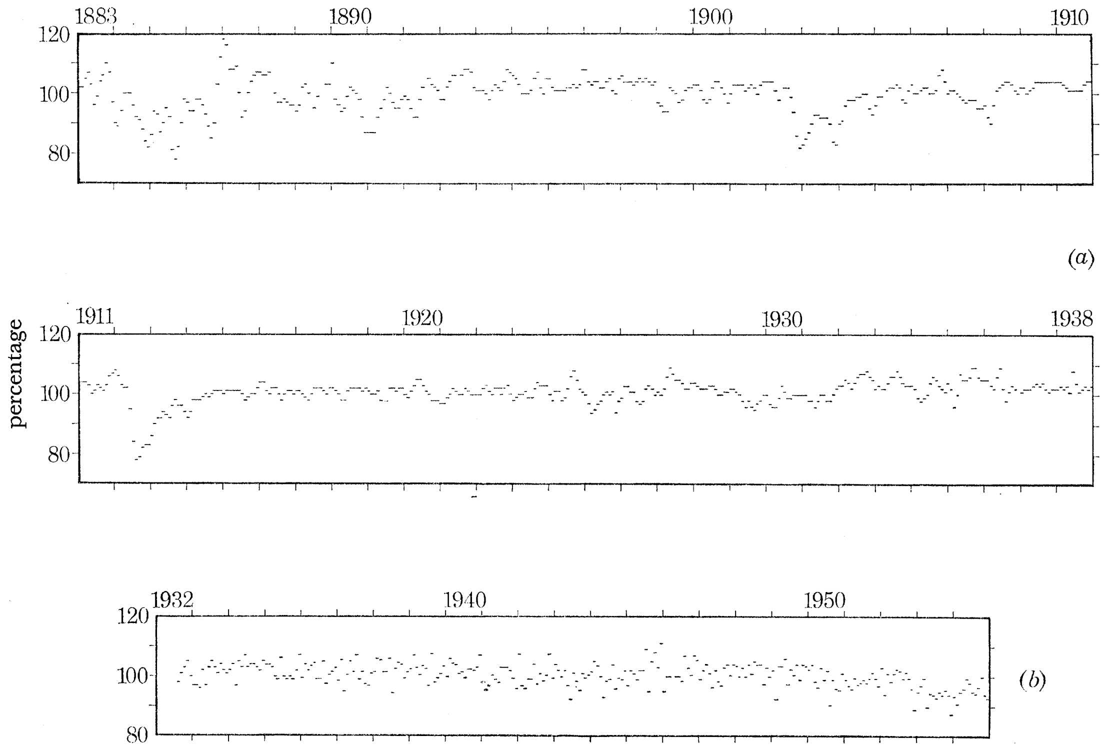


FIGURE 15. (*a*) Average monthly values of the strength of the direct solar radiation derived from observations at mountain observatories, from 1883 to 1938, as percentage of the overall mean. (For distribution of observatories contributing see text.)

(*b*) Average monthly values of the strength of the direct solar radiation derived from observations at Matsumoto (36° 15′ N, 137° 58′ E) and Shimizu (32° 47′ N, 132° 58′ E) in southern Japan from 1932 to 1954, as percentage of the overall mean.

first magnitude. The greatest effect of the Alaskan 1912 eruption may well have been upon the ice on the Arctic seas.

From 1883 onwards measurements of the strength of the direct solar radiation at one or more observatories, mostly sited to minimize the adjustments necessary for weather and cloudiness, are available. The average values for each month from 1883 to 1938 are reproduced here in figure 15*a*, from Kimball (1924) and Hand (1939), as percentages of the overall mean. The observing stations represented in this figure were all in the northern hemisphere between about 30° and 50° N. After 1923 only places in the United States are represented. The coverage may be listed as follows:

Montpellier, France	1883–1900	Mount Weather, Virginia	1907–14
Pavlovsk, Russia	1892–1913	Kew, England	1908–21
Lausanne, Switzerland	1896–1904	Madison, Wisconsin	1910–38
Warsaw	1901–18	Santa Fe, New Mexico	1912–22
Washington, D.C.	1905–38	Helwan, Egypt	1914–23
Simla, India	1906–15	Lincoln, Nebraska	1915–38
Paris	1907–13	Blue Hill, Harvard	*ca.* 1930–8

As to how the observations of direct solar beam intensity were handled, Kimball (1918) has given only the briefest account: most weight was given to observations at the highest solar elevations available. The monthly values have all been adjusted by $\frac{1}{4}(a+2b+c)$ smoothing, where *a* and *c* represent the averages for the preceding and following months and *b* for the month as shown. Even so excessive scatter mars the record of the period 1883–92 when only one observatory was operating, and for this reason the excessively high insolation values obtained for a few months in the winter of 1886–7 may safely be disregarded. The plotted values indicate depletion of the direct solar beam by up to 20 % for a large part of the year immediately following the great volcanic eruptions in 1902 and 1912 and for a good deal longer after the Krakatau eruption of 1883. Several other cases of important depletion appear before 1921. Minor depletions of 2 to 5 % (below preceding values) affecting individual months in 1931, 1934, 1935 and 1936 were tentatively attributed by Hand to dust storms in the United States, that being the time of the so-called 'dust bowl' erosion of the grain lands in the Middle West.

The course from 1932 to 1954 of the monthly mean values of strength of the direct solar radiation as derived from observations in Japan, at Matsumoto (36° 15′ N, 137° 58′ E) and Shimizu (32° 47′ N, 132° 58′ E) away from urban pollution of the atmosphere, may be seen here in figure 15*b* (after Arakawa & Tsutsumi 1956). The scatter of the monthly mean values should no doubt be ascribed largely to the difficulties of selection and reduction of observations to equivalence with clear sky conditions. The values are essentially clustered about the overall average of the series until 1949, after which a falling tendency may be suspected. The sharp reduction of general level of the values in 1953 and 1954 (by about 5 % overall, and 16 % in the extreme month) cannot be entirely or even mainly due to the eruption of the Mt Spurr volcano in Alaska on 9 July 1953, because the low values start about December 1952. Arakawa and Tsutsumi incline to the view that they are largely attributable to atomic bomb tests in the equatorial Pacific in November 1952 and spring 1954 and a test in central Asia in August 1953, which being ground bursts might have thrown up dust on a scale equivalent to volcanic eruptions: these were not hydrogen bombs, however. In the absence of direct observation of such a dust veil, other causes—such as observational difficulties, a real solar variation or pollution of the tropopause layer by the exhaust gases and vapours of high flying aircraft—seem likelier.

All the radiation measurements discussed were made between $29\frac{1}{2}$ and $52\frac{1}{2}$° N. No equivalent long series monitoring the radiation received in the southern hemisphere seems to have been published. Reductions by Abbot, Aldrich & Hoover (1942), and Aldrich & Hoover (1954, and personal communication) of observations made at the Montezuma observatory, Chile (22° 40′ S, 68° 56′ W 2710 m above sea level (a.s.l.)) from 1920 to 1957 to obtain the solar constant indicated a 1 % deficiency (12-month average) just in the year 1922 (lowest monthly value 98.3 % of normal, July 1922) following the 1921 Andean eruption. Radiation observations at the Commonwealth Observatory (Mount Stromlo), Canberra, Australia (approx. 35.1° S, 149.1° E) from 1926 onwards were reported by Rimmer (1937) to show depletion of the direct beam throughout the remainder of 1932 and 1933 after another eruption in the Andes on 10 April 1932.† The greatest depletion of the monthly values was 3.6 % in some months of 1932 whilst the sky radiation reached abnormally high values, especially in September to October 1932 and January 1933.

The reductions of strength of the direct solar beam observed at the surface after volcanic eruptions are largely compensated by increased diffuse radiation from the sky glare. Nevertheless, they provide an excellent indicator of the presence of a dust veil and probably give reliable comparisons of dust veil density from one case to another. All the big reductions of

† The frequency of observation of luminous night clouds in the northern summer of 1932 and several following years (see p. 438) may suggest that traces of the dust and vapours ejected entered the mesosphere and were transported to high northern latitudes at altitudes of 50 to 80 km.

direct solar beam intensity shown by figure 15 stand in obvious relation to the known great volcanic eruptions of the period.

The reduction of total radiation received is thought to be no more than a quarter to a tenth part of the deficiency of the direct beam, (see p. 462) thanks to net forward scattering. However, the margins of error applying to all radiation measurements so far make it difficult to determine this balance reliably.

The measurements of total incoming radiation, and the direct and diffuse components of it, made at the Goetz Observatory, Bulawayo (20° 09′ S, 28° 37′ E, 345 m a.s.l.) from January 1963 to March 1964, and for comparison the average monthly values of the previous 6 years, have been made available to the writer by the kindness of Mr Archer, the officer in charge, and of the Director of the Meteorological Service, Salisbury, Rhodesia. These data are reproduced here as figure 16*a*. The eruption of Mount Agung, Bali (9° S, 114° E), believed to have produced the dust veil, took place from 19 February till 17 March 1963, when the greatest explosion was reported. Unfortunately the Goetz radiation figures are monthly totals of measurements made under all weather conditions, and the departures shown in figure 16*a* must owe something to weather anomalies as well as to the dust veil. The period from mid-1961 to 1964 (or later) was one of highly anomalous weather over Africa, drier and less cloudy than usual over the region of the Goetz Observatory and the Rhodesias and with a serious and prolonged drought in the Transvaal, but excessive rains in the equatorial belt and in the extreme north and south of the continent. No special significance should be attached therefore to the curious chance that the observed reduction of the direct solar radiation (12-month average March 1963 to February 1964 just 90.3 % of the 1957–62 'normal' figure) was exactly compensated by the excess of diffuse radiation, making the total radiation (direct plus diffuse) for the same 12 months just 100.0 % of the 1957–62 average.† With rainfall only two thirds of normal for those 12 months at Bulawayo clearer skies should have produced an excess of insolation; so the total radiation reported probably was depressed by the dust veil.

The Goetz observations show a significant positive anomaly of the diffuse (sky) radiation lasting 8 to 12 months after the eruption. Abnormal brilliance of the sky near the Sun was reported by W. Schuepp observing in the Congo as long as 22 months after the eruption (i.e. till December 1964). More prolonged effects, continuing at least until 1966, were reported in higher southern latitudes, over Australia (see below) and the South Pole (p. 458).

The course of the direct, diffuse and total radiation before and after the Bali 1963 eruption, as shown by the routine observations at noon on cloudless days at Aspendale, Melbourne (38° 02′ S, 145° 06′ E) has been presented by Dyer & Hicks (1965) and is reproduced here in figure 16*b* by kind permission. The elements of the radiation received are expressed as fractions of the averages ($\bar{I}$, $\bar{D}$ and $\bar{T}$) of the previous 4 years since the observations began in 1959. At the time of strongest effects in July to August 1963 the direct radiation was reduced by 24 %, the diffuse radiation was doubled and the total radiation appeared deficient by about 7 %. However, the difficulty presented by high 'noise' ratio in the total radiation calculations is plain, and the indicated 7 % reduction of total radiation must be less reliable than the other figures quoted. It is noticeable in figure 16*b* that the direct radiation was most affected in the winter seasons—i.e. with low angles of the midday sun—though Dyer & Hicks believe that arrival of additional dust

† Similar radiation effects were observed at Pretoria (26° S) after the arrival of the stratospheric dust about 20 April 1963 (Bosua 1963). The direct radiation measured at Pretoria showed an abrupt drop of about 5 % between 20 and 22 April.

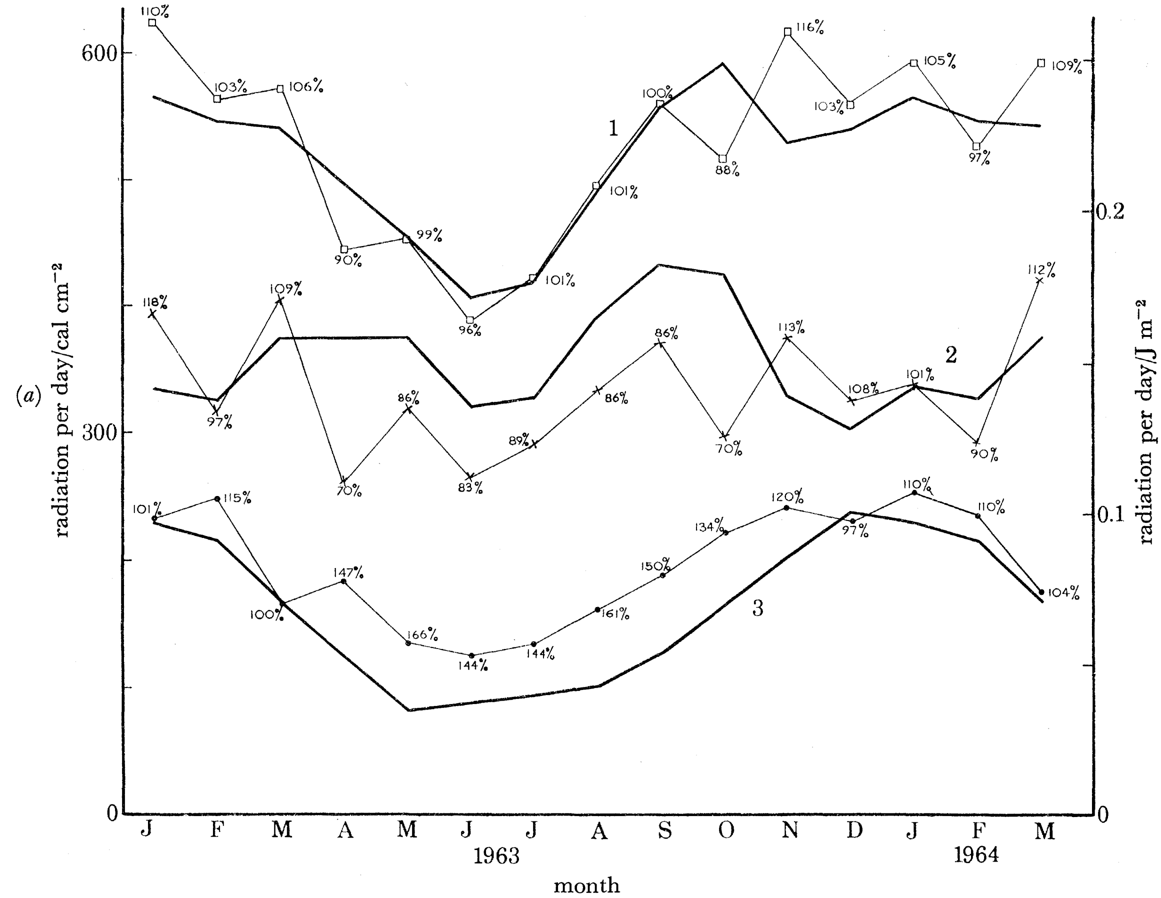
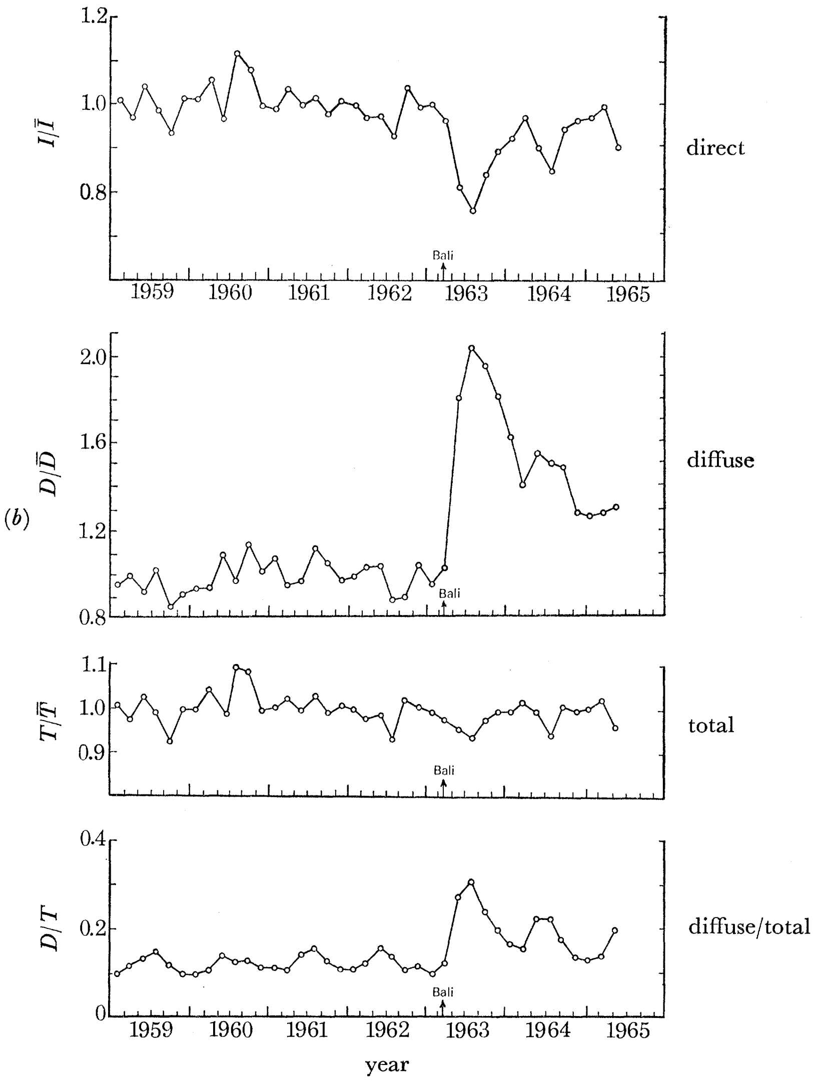


FIGURE 16. (*a*) Radiation observed at the Goetz Observatory, Rhodesia (20° 9′ S, 28° 37′ E) before and after the eruption of Mt Agung in Bali in February to March 1963. 1, total radiation, monthly average values; 2, direct solar beam, monthly average values; 3, diffuse radiation, monthly average values. Bold lines averages for 1957–1962. Thin lines 1963–4; monthly percentages of the 1957–1962 average indicated.

over 38° S each southern autumn contributed to this: the direct radiation reduction was again as great as 16 % in July 1964 and about 10 % in July 1965. Volz (1964) reported a reduction by 0.4 stellar magnitude (equivalent to 30 % of the intensity) of the direct solar radiation (monthly values) at the wave length of maximum intensity (0.5 μm) observed in Australia and South Africa at various times in the first 6 to 7 months after the Bali 1963 eruption—the same figure as reported by the Goetz Observatory, Rhodesia, in April and October 1963. It remains unknown whether the depletion of the solar beam averaged for observatories distributed similarly to those operating after 1883 and 1902, in temperate and lower latitudes of the northern hemisphere, would have been as great as 30 % in any one month.

Volz (1965), summarizing global observations of the variation of stratospheric turbidity after the eruption of the Agung (Bali) volcano in 1963, indicates that the maximum turbidity about October 1963 exceeded that observed in spring 1903, though the subsequent fall-away was initially more rapid and the total effects of the 1902–4 and 1963–6 dust veils over the hemisphere of origin may be about equal. They clearly both exceeded the integrated effect of the 1912–14 veil by a factor of 2 or thereabouts and that of the Andes 1932 eruption by a factor of 5 or more.

The Bali 1963 dust veil was clearly the most effective one since 1902–3 and possibly since 1883–6.



(*b*) Direct (I), diffuse (D) and total (T) solar radiation, and the contribution of the diffuse (sky) radiation to the total, measured at Aspendale, Melbourne (38° S, 145° E) after the Bali 1963 eruption. Fractions of the 1959–62 averages.

Apart from the overall reduction of incoming radiation, and, with the greatest dust veils, of temperatures at the surface, in the first year after an eruption, continuance of similar effects for up to several more years over the highest latitudes should produce two other important consequences.

(1) Increased production and reduced melting of ice in the polar regions due to the lower temperatures prevalent there for 1 to 10 years after a great dust veil producing eruption.

(2) Increased poleward gradients of temperature, accompanied by increased zonal winds in the upper atmosphere, over a zone near the edge of the dust veil. This effect should progress polewards if there is a net poleward drift of dust in the lower stratosphere at a rate which would probably vary from case to case. Evidence of it might be expected at some time between 1 and 3 years after the formation of a world-wide dust veil, probably following an initial phase in which decreased energy transmission should produce a lowering of temperature everywhere, lessened temperature gradients and a weakened circulation.

6. Assessment of the volcanic dust in the atmosphere: dust veil index

In this section simple formulae are developed for calculation of a dust veil index (d.v.i.) for any eruption, using observations either of the depletion of the solar beam or the temperature lowering in middle latitudes or the quantity of solid matter dispersed as dust. Examples are given of the calculation of the index in cases where all these approaches could be used. The extent of agreement and discrepancy between the different approaches is seen and the manner of arriving at the figure finally adopted is explained. Rules of thumb used for estimating the fraction of the Earth effectively covered by a volcanic dust veil at its maximum extent are also included: this fraction and the duration of the dust veil contribute to the magnitude of the d.v.i.

Meteorology requires an assessment of volcanic dust veils in terms of:

(i) *density*—i.e. *opacity*—especially *initial density*, which must be related to the quantity of fine tephra (i.e. dust) ejected within a short time during explosive phases of the eruption.

(ii) *extent*—especially the *maximum extent* attained by the veil. This can be taken as depending upon the latitude of injection and its relation to the circulation of the upper atmosphere over the following 3 to 6 months.

(iii) *duration*—i.e. the *total life* of the dust veil. Longevity clearly depends upon entry of the dust into the stratosphere, and must be enhanced in those cases when further explosions of the volcano produce repeated injections over some time.

Such information is needed for comparative treatment of observed (or apparent) effects of different eruptions upon surface weather, on lower and upper atmospheric temperatures, and on the large-scale wind circulation. It would be convenient if all eruptions could be rated according to a single numerical index. But, for some purposes, whatever is known about the opacity, extent and duration of each veil should be specified separately.

These considerations have guided the compilation here presented in appendix I, which gives in chronological order all known eruptions from A.D. 1500 to 1965 that might be thought to have had meteorological effects. Various lists of volcanic eruptions are available, but none of them apart from Sapper's have hitherto attempted a systematic assessment of their importance as 'ash' eruptions. The principal lists here consulted and the letters that identify them in appendix I are:

(H) Humphreys (1940, pp. 615, 616) from A.D. 79 to 1913

(K) Royal Society (1888, pp. 384–401). 'List of principal ascertained volcanic eruptions from 1500 to 1886.'

(S) Sapper (1917, 1927) from 1500 to 1914.

(Sh) Shaw (1936, pp. 21–25, 408). List of 'important eruptions' from 1500 to 1932.

Besides these, other sources too numerous to mention have given details of this or that eruption. For the years since 1913 every known, or accidentally discovered, report has been followed up for possible further details; but it is clear that till 1953 or later there was less volcanic activity, at least in terms of rate of output of dust, than in the preceding period. The most useful sources of information for recent years and decades appear to be

Gutenberg & Richter (1954, pp. 289–303: table 19 'List of active volcanoes and last eruptions')

Bulletin volcanologique (1924–60)

Bulletin of volcanic eruptions (1961 onwards).

From the varied information obtained an attempt has here been made to define a dust veil index (d.v.i.), which should be related to the total loss of incoming radiation to the Earth occasioned by each eruption. The index should provide a comparison of the sizes of the '*bites*' out of the radiation supply curves as illustrated in figure 15*a*—i.e. the more or less triangular areas between the curve representing observed values of in-radiation (which dips after an eruption and then recovers) and the imaginary continuous line corresponding to undisturbed radiation supply. The radiation deficits which we would like to compare differ from those in figure 15*a* by being losses of total (not just direct) radiation and in being integrated over the whole Earth or a whole hemisphere (not just representative of the middle latitudes of one hemisphere). However, it is thought that the loss of total radiation should approximate to some fixed fraction of the loss of direct radiation (in most cases the only information reported hitherto) and that the radiation curves observed in middle latitudes (the only ones available so far) are from the best latitude zone to indicate a middle value fairly representative of the hemisphere—subject to the provisos that: (i) the latitude of injection probably indicates whether the equatorial zone was ever covered, and whether there is likely to have been any appreciable spread of the dust into the other hemisphere; and (ii) the dust may have persisted much longer over the small area of the polar cap beyond latitudes 70 to 80° than elsewhere. An approximation to the desired radiation deficit comparison may apparently be obtained in the case of eruptions before the start of the radiation record in 1883 by comparing the sizes of the troughs in Köppen's middle latitudes temperature curves (figure 14).

The shape of the troughs in the radiation curve, and (if some smoothing were adopted) of those in the temperature curves, is characteristically a steep fall followed by a gentler, but fairly even, upward slope during the recovery. We might have expected some nearer approach to an exponential shape of the recovery curve of any parameter which registers the presence and gradual disappearance of the dust. The more rapid cut-out of the dust veils (except perhaps over the polar regions) as implied by the triangular shape of the areas of interest on the available radiation curves (which are for middle latitudes—figure 15*a*) may be a result of poleward drift of the dust. It is reasonable to treat the area which measures the overall loss (integrated depletion) of radiation as triangular—i.e. proportional to the maximum depletion multiplied by the total time elapsed between the eruption and the regaining of the initial radiation, or temperature, value.

This permits the d.v.i. to be defined by three simple alternative formulae, different ones being applicable according to what information is available for different eruptions but in many cases providing alternative estimates of the same dust veil and thus affording a reliability check. Any more elegant formulae—e.g. adopting exponential decay rates—would be inappropriate in relation to the coarseness of most of the data available hitherto and might not constitute a closer approximation to reality. Each formula includes a numerical factor designed to bring the index to 1000 for the Krakatau 1883 eruption, chosen for obvious reasons as the basis of comparison.

Dust veil index formulae

$$\text{d.v.i.} = 0.97\, R_{\mathrm{D\,max}}\, E_{\max}\, t_{\mathrm{mo}} \qquad (1)$$

where $R_{\mathrm{D\,max}}$ is the greatest percentage depletion of the direct radiation as registered by the monthly averages in middle latitudes of the hemisphere concerned after the eruption. (A considerable lag must be allowed for the maximum depletion to be realized, since spread of the dust to other latitudes appears to be associated with seasonal changes of the circulation occurring just once or twice a year. After the Krakatau 1883 eruption the densest part of the veil appears to have taken 2 years to reach middle latitudes.) E_{max} stands for the maximum extent attained by the dust veil, rated according to the following crude scale:

(i) Whole Earth = 1 (from eruptions between about 20 °N and 20 °S).

(ii) Extratropical latitudes in one hemisphere plus entire tropical zone (applied to eruptions near the subtropical jet stream in latitudes about 20 to 35°) = 0.7.

(iii) For eruptions at latitude 35° to about 40 to 42° an intermediate value of 0.5 is applied.

(iv) Extratropical latitudes in one hemisphere = 0.3.

t_{mo} is the total time in months between the eruption and last observation of the dust veil or of its effect upon monthly radiation (or temperature) values in middle latitudes.

$$\text{d.v.i.} = 52.5\ T_{\mathrm{D}max}\ E_{max}\ t_{mo} \qquad (2)$$

Here $T_{\mathrm{D}max}$ stands for the estimated lowering of average temperature in degrees Celsius over the middle latitude zone of the hemisphere affected, in this case the anomaly for the *year* most affected.

$$\text{d.v.i.} = 4.4qE_{max}\ t_{mo} \qquad (3)$$

In this expression, the estimated total volume in cubic kilometres of solid matter dispersed as dust (q) in the atmosphere, is used. The best estimates are presumably those obtained from the measured dust deposit over a wide area about, and down-wind from, the volcano. Such estimates must always be difficult, and subject to a much wider margin of uncertainty, than the measured radiation and temperature effects (even if the temperature variation after volcanic eruptions be sometimes obscured or confused by independent variations of the atmospheric circulation and possibly of solar output). Estimates based on the amount of solid matter removed from the volcano must be a great deal less reliable still, because by far the greatest proportion of the material removed by great explosions goes as blocks and boulders, some of them enormous, and falls within the first hour.

Even when the reckoning of the dust veil index has been systematized by the use of these routine formulae, there is still a good deal of unavoidable estimation involved—in particular, in the choice of values for q, for $T_{\mathrm{D}max}$ and t_{mo} indicated by various curves, and for E_{max}. It must simply be recognized that this is about as far as one can go towards objectivity in the assessment of past eruptions. The method seems reliable at least as an indicator of order of magnitude. Index values must be rounded to avoid false impressions of precision. Clearly formula (3) uses the least reliable data, and not too much should be expected of it. In many cases the chief value of formula (3) may be to use it to work back from the d.v.i. derived from the other formulae to obtain an indication of how much solid matter was probably carried in the atmosphere as dust. In the case of some early eruptions, observers' descriptions of the density of the dust veil, when they viewed the Sun and Moon through it, may be compared with the later cases for which the dust veil index can be computed from at least two of the above formulae, and this may provide the basis for an estimate of the d.v.i. more reliable than formula (3) alone.

The way in which the maximum probable spread of the dust veil over the Earth, E_{max}, is used in all the formulae implies that, for the same quantity of solid matter blown up, d.v.i. of eruptions in the polar and temperate zones of either hemispheres are only 30 % as great as those of eruptions within the tropics. Subtropical eruptions (in latitudes 23 to 40°) score d.v.i. 50 to 70 % as great as eruptions within the tropics for the same amount of solid matter ejected. The magnitudes of the eruptions considered just as dust producers may be compared by taking the figures for d.v.i./$E_{\max}$ given in appendix I: these figures may also be used if it is desired to eliminate the assumed poleward drift. Apportionment of the dust between northern and southern hemispheres must be important in connexion with the effects of the dust upon the radiation balance over the respective polar regions. For study of cumulative effects over the northern hemisphere (figure 23 later and Appendix II) the rule of thumb has been adopted that dust originating within about 15° of the equator becomes equally divided between the hemispheres; dust originating from eruptions in latitudes 15 to 20° is counted two-thirds to that hemisphere (and one-third to the other); dust from eruptions poleward of latitude 20° is reckoned as effectively all staying within the same hemisphere.

Consistency checks between d.v.i. values computed by the different formulae for the most fully observed great dust veils since 1800 are informative. Comparisons of the results obtained from the formulae for five eruptions for which all three formulae could be used are plotted in figures 17*a* and *b*. Details of the computation results and the figures and estimates used can be seen in table 5. The average temperature lowering in middle latitudes is taken from Köppen's estimates except in 1912, when it is based on data for 15 well separated stations around the northern hemisphere, and in 1963–6, when world maps of temperature anomaly could be used. In 1912 and 1963–6, owing to longer term climatic trends then in progress, it is important to use estimates of temperature lowering (during the occurrence of the dust veil) from values for the immediately preceding years rather than from any long-period averages. The same doubtless applies to other eruptions in our chronology (appendix I), e.g. in the 1760s, and should be taken as a general working rule.

Alternative estimates of the amount of dust and/or total tephra dispersed by the 1912 eruption of Katmai in Alaska are seen in figure 17*b* (open circles); these indicate (i) the unrepresentativeness of the total tephra estimate, which seems to have been about four times the probable quantity of atmospheric dust judged from the graph, and (ii) the apparently gross underestimate of the dust given by the estimate of the deposit. (Difficulty of estimation in the case of the eruption in the Valley of Ten Thousand Smokes—i.e. Katmai—in 1912, writes Dr P. E. Baker, arises from the unusual character of the eruption, described as a 'sand flow' or 'pyroclastic flow'. This was a glowing hot avalanche in which perhaps about 4 km^3 of broken material and dust moved under the influence of gravity; very little of it would be likely to escape as dust into the atmosphere.)

One may reasonably conclude from figure 17 that the radiation and temperature formulae give results that correspond with each other satisfactorily, at least in cases of well-marked dust veils. The procedure adopted in the assessments in appendix I is to use these two methods wherever possible, take the average of the two figures obtained and quote the nearest appropriate round figure as the index value.

The estimates of total dust in figure 17*b* behave reasonably, if we except the widely divergent dust estimates for Katmai 1912. However, even for the Krakatau 1883 eruption it was necessary to choose the estimate of 6 km^3 of dust from among other figures up to 18 km^3 (the latter for total tephra, of which it may be an underestimate because the deposit was so largely into the ocean) quoted in the literature. For two earlier very great eruptions, in 1815 and 1835, the only available estimates of the total tephra (q in table 5) gave d.v.i. values respectively 7 and 4 times as great as those computed from the temperature lowering. This indicates either grossly excessive estimates of the solid matter blown up or that no more than $\frac{1}{7}$ to $\frac{1}{4}$ of it was dust small

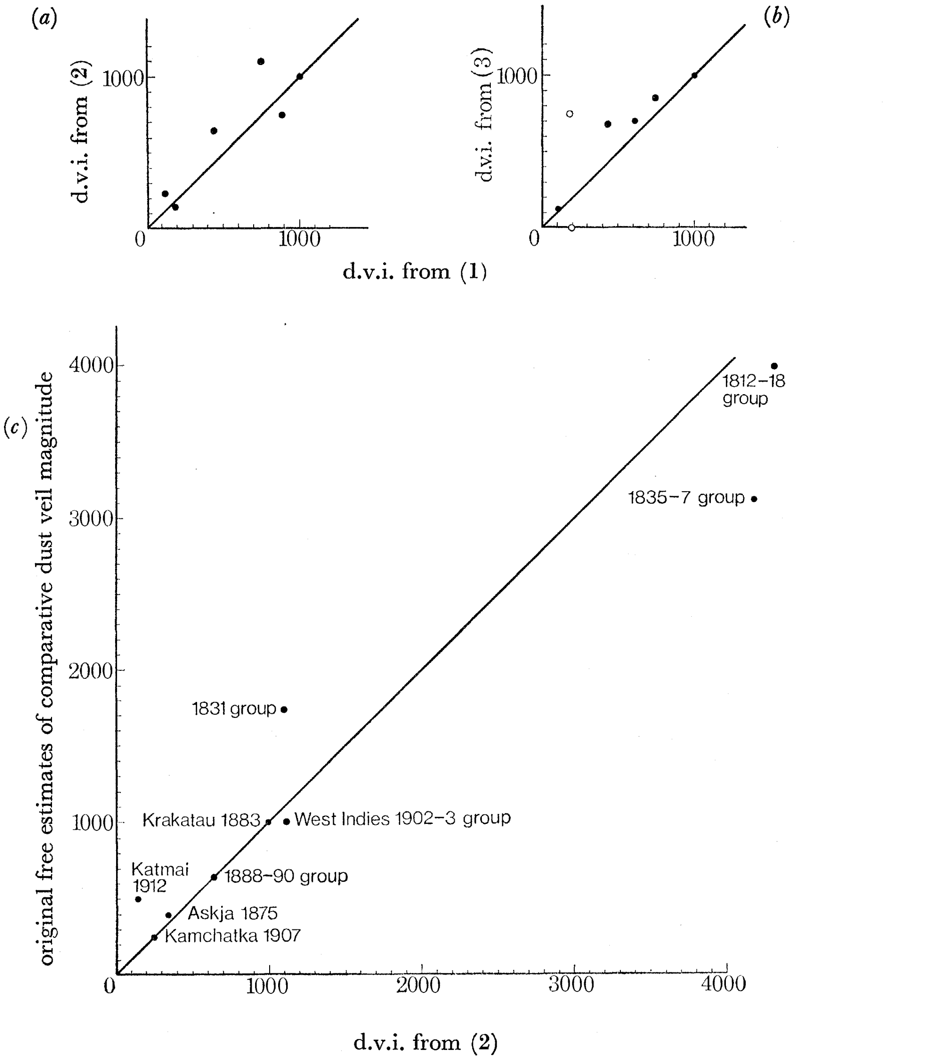


FIGURE 17. Comparisons of dust veil index values obtained by different methods and formulae (see text).

enough to be carried by the atmosphere. The latter explanation may well be the true one; these have been reported as exceptionally violent explosions, Although the sound of the explosion of Krakatau in 1883 was heard as far away as Alice Springs (Australia) and the island of Rodriguez, distances of 3500 and 4500 km, compared with only half as far in the case of Tambora in 1815, the difference may well be due to the very much slenderer chances of reportage in those parts of the world in 1815. Most approaches agree in assessing the 1815 and 1835 eruptions as greater in terms of tephra than Krakatau 1883, but the dust estimates cannot be used as more than a very coarse indicator of magnitude. It may be safe to rely more closely on dust estimates in cases where the dust layer is known to have been thoroughly surveyed, as with those illustrated in figure 6 and in Eaton's paper (1963): only these dust surveys have been accepted in the d.v.i. assessments printed in appendix I.

A fourth source of dust veil index values is represented by the first 'free' estimates made several years earlier by the present author from all the miscellaneous information assembled in appendix I, playing particular attention to optical observations of the opacity of each veil, to its extent and duration, and to comparability of the assessments throughout the chronology. These 'free' estimates were done afresh six months later and reviewed once more after another year. They were all completed before any of the formulae were used. Figure 17*c* shows how these estimates compare with those computed from the temperature lowering in the case of nine veils, including the very great ones in 1815 and 1835. The crude estimates appear usable; they are accordingly given in appendix I (in brackets) in cases where they are likely to be useful—e.g. as the only available assessment of some early eruptions and to indicate the probable contributions of individual eruptions which occurred in or about the same year as each other to a total veil for which a dust veil index value has been calculated from temperature or radiation measurements.

Table 5. Comparison of estimates and computations of dust veil index by various methods, and figures used

		dust veil index calculated by formula			figures used in the formulae				
eruption and year	initial free estimate of d.v.i.	(1) radiation loss	(2) temperature lowering	(3) dust quantity estimates	E_{max}	R_{Dmax}	T_{Dmax} °C	q	t_{mo}
Agung, Bali 1963	—	875	750	—	1.0	25	0.4	—	36
Katmai, Alaska 1912	500	190	140	750 total tephra 0.2 another estimate of dust	0.3	24	0.3	21 or 0.006	27
Kamchatka 1907	250	110	240	120	0.3	13	0.5	3	30
West Indies and Central America 1902–3	1000	750	1100	850	1.0	21	0.6	5.4	38
Bandaisan, Ritter Is. and Bogoslov 1888–90	200 150 300 } total 650	440	650	680	0.5 1.0 0.3	18	0.5	4.7	50
Krakatau 1883	1000	1000	1000	1000	1.0	27	0.5	6.0	38
Askja, Iceland, 1875	400	—	380	110	0.3	—	1.0	3.5	24
Coseguina 1835 other Chilean eruptions 1835 Avachinsky 1837	2500 500 250 } total 3250	—	(4200)	(17500 Coseguina only)	1.0	—	1.0	50 Coseguina only	(80)
Mediterranean, Philippines and Ecuador 1831	1750 (total)	—	1100	—	1.0	—	1.0	—	36
Tambora 1815 All eruptions 1811–18	4000	—	4400	32000	(1.0) (1.0)	—	1.0	150 ?	48

When eruptions of different volcanoes occur in or about the same year, it is generally the total dust veil produced that matters to meteorology.

For most of the early eruptions listed the d.v.i. assessments have been guided by a careful study of the range of the present investigator's initial d.v.i. estimates that corresponded to each of Sapper's ratings b_1, b_{1-2}, b_2, b_{2-3}, b_3, etc. in all cases where information at his disposal made possible an independent assessment. The comparison of computed d.v.i. values with Sapper ratings (included here in appendix I) reveals, however, unmistakably that Sapper shifted his standards abruptly about 1900 and badly overestimated the eruptions of the early years of the twentieth century. Therefore no use has been made of his estimates for 1900–14. Where no further evidence is available, as for most of the earliest eruptions, the average d.v.i. value for the Sapper assessment has been adopted.

The consistency studies suggest that reasonable confidence can be had in dust veil index values derived from two or more methods, formulae (1) and (2) being preferred, when the results are averaged and rounded to the nearest 10 up to 100; to the nearest 50 up to 500; to the nearest 100 up to 1000; to the nearest 250 up to 2500 and to the nearest 500 above that. When only one method can be used, the coarser rounding procedure for the next grosser range should be applied. The full list in appendix I indicates 25 cases between 1500 and 1963–6 of dust veils comparable with that after the 1883 Krakatau eruption including some that were probably rather greater. 1815 and 1835, probably also 1601, 1660 and 1693–4, as well as some doubtful cases depending on temperature evidence in the 1760s and 1820s, all give some evidence of greater magnitude. In addition, Appendix I includes over 50 cases of dust veils which were of only rather less magnitude (d.v.i. values from 100 to 800).

Where the dust veil has been assessed on evidence of temperature variation, there is a manifest danger of arguing in a circle if one proceeds to use the total dust veil index values over a period of years to show how volcanic dust affects climate. In 1763, 1826–30 and 1879–81 abrupt falls of temperature of the order of 2σ on the curves in figure 14 indicate large dust veil index values, though no direct observations of a dust veil in those years are known to the writer. Such cases must be omitted from numerical sums used to correlate volcanic dust and climate (as later in this report, p. 484 *et seq.*), even though it is known that scantily reported eruptions did occur which may very well have been responsible and even though the omission of some genuine cases for this reason may weaken any relationships appearing. Radiation measurements after the eruptions of 1888–92 makes it quite certain that these eruptions produced and maintained a major dust veil, and previous writers are agreed on this, although no reports of optical effects or direct observations of this dust veil appear to have been preserved—probably because to most observers they represented merely a continuation of some of the effects of the Krakatau dust veil seen in the immediately preceding years.

7. The geographical distribution of volcanic activity

The world distribution of volcanoes active since A.D. 1500 is mapped in figure 18. The spatial distribution is of only secondary interest to meteorology because of the ability of the upper atmospheric circulation to spread the dust. However, certain regions are clearly more liable to great volcanic explosions than others, the greatest frequencies being first and foremost in the Pacific island groups—Indonesia, the Moluccas, Celebes and Phillipine Islands—and secondly in the region described by Mexico, the West Indies and northern Andes. Frequencies are high elsewhere in a ring bordering the Pacific—e.g. Kamchatka and the Aleutians.

The distribution of the greater eruptions (d.v.i. $\geqslant$ 100) by latitude zone, and the latitude distribution of dust production (totals of (d.v.i.)/$E_{\max}$), are shown in figure 19. It will be seen that the inclusion of smaller eruptions makes no difference to the distribution, which is dominated by the equatorial zone eruptions whether d.v.i. of (d.v.i.)/$E_{\max}$ be considered. Some other regions, however, make important contributions: Iceland, the northernmost Pacific group (Kamchatka-Aleutians-Alaska), and possibly the Antartic, attract notice. Sapper's distribution illustrated in figure 19*a* agrees well with that here obtained.

The activity in the main regions since 1500 is indicated in table 6 by the frequency per century of eruptions producing various dust yields ((d.v.i.)/$E_{\max}$). The preponderance of Indonesia and the islands in the equatorial zone (latitude $\leqslant$ 15°) of the western Pacific stands out. This preponderance is even more marked if the extent of the dust veils is taken into account (d.v.i. values unadjusted). Odd cases of eruptions which were as great dust producers as any since 1500, and in some cases greater, are known to have occurred elsewhere in earlier times—

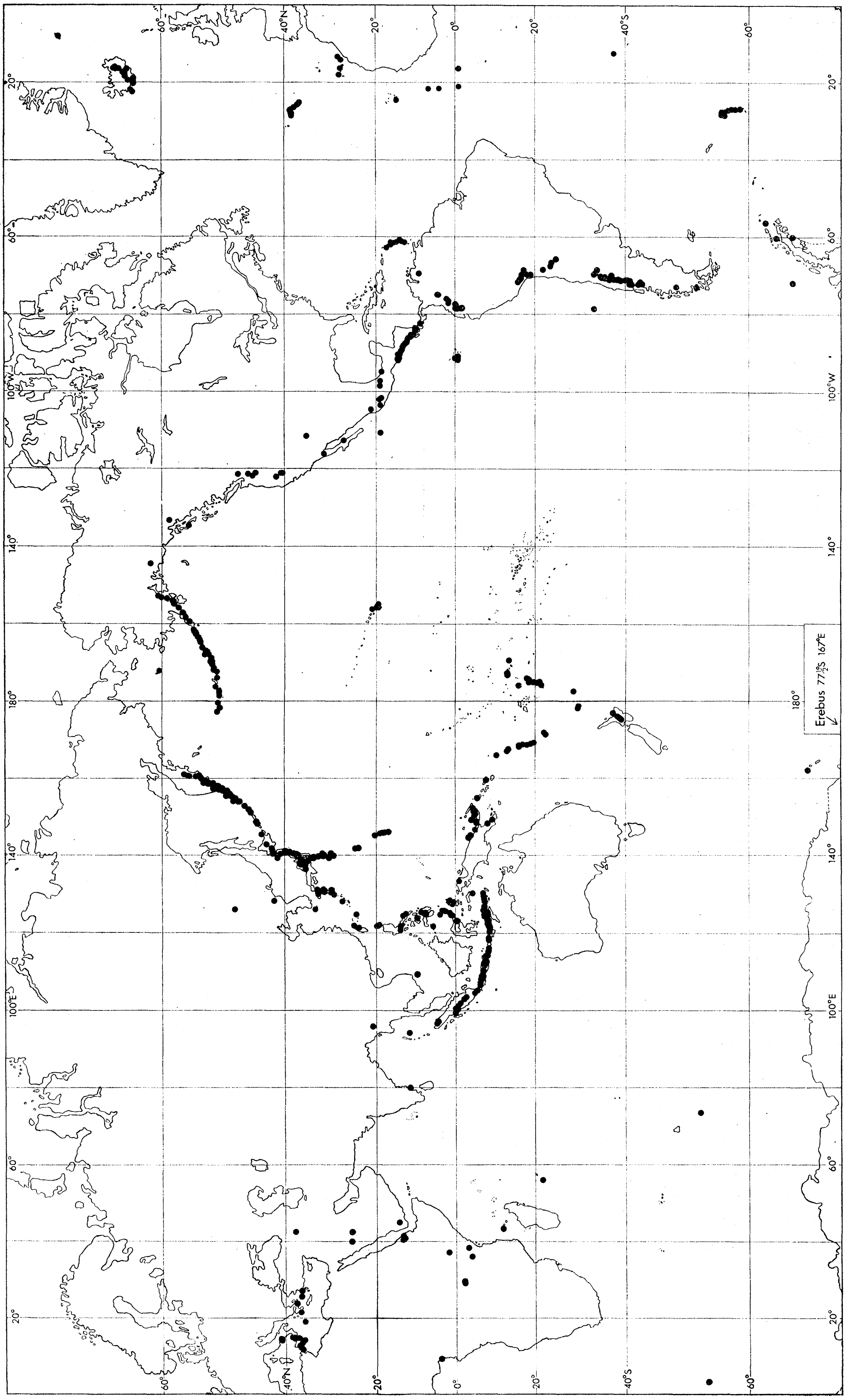
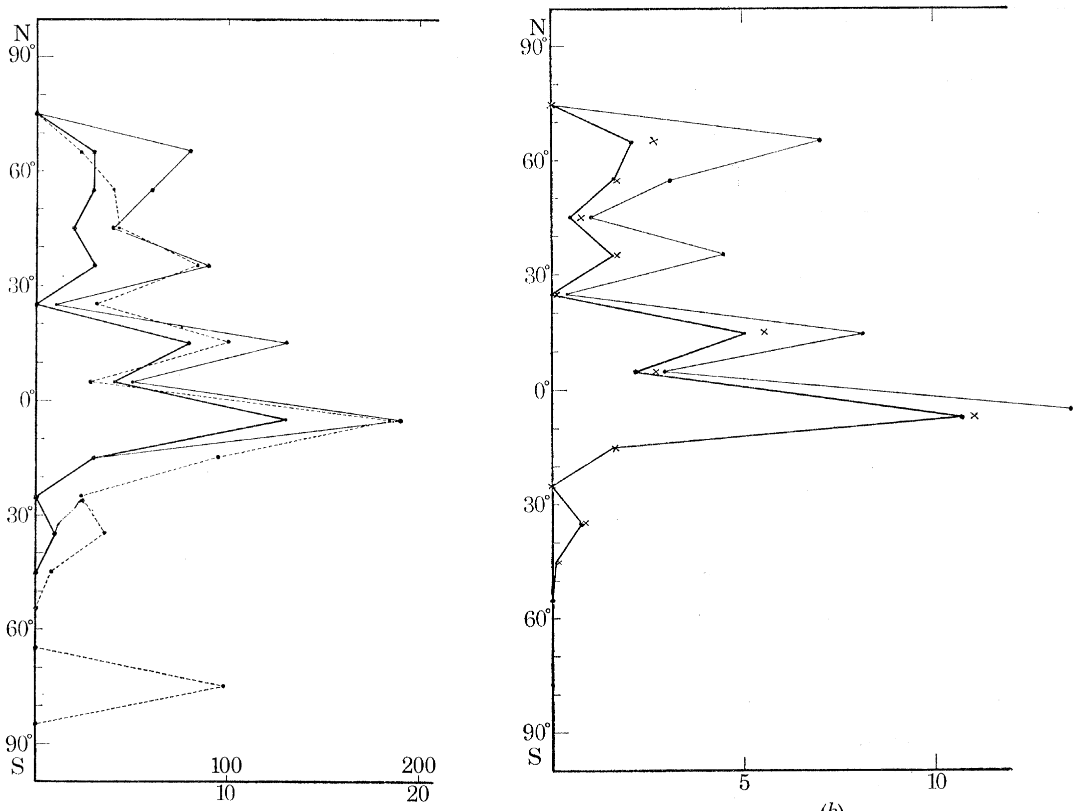


FIGURE 18. World distribution of active volcanoes 1500–1964.



(*a*)
inner scale: Sappers total eruptions
outer scale: d.v.i. $\geqslant$ 100

(*b*)
$10^{-3} \times$ total of (d.v.i.)/E_{max}

FIGURE 19. (*a*) Numbers of eruptions producing d.v.i. $\geqslant$ 100 in each 10° zone of latitude. ——, totals 1700–1964 (information inadequate to continue curve south of 20° S since New Zealand eruptions before 1850 likely to be unknown); ——, totals 1800–1964 (no information from New Zealand in early years): Eruptions south of 50° S unassessable; – – –, Sappers total of eruptions (all types) 1800–1914 (omits some eruptions on sub-Antarctic islands 60–70° S).

(*b*) Totals of (d.v.i.)/E_{max} produced by eruptions in each 10° zone of latitude. ——, since 1700; ——, 1800–1964; the curves use only those eruptions which gave (d.v.i.)/E_{max} $\geqslant$ 100; $\times$, totals for 1800–1964 using all known eruptions with (d.v.i.)/E_{max} $\geqslant$ 10.

e.g. Öraefajökull, Iceland in 1362, ca. 10 km³ dust when freshly fallen (see p. 444); the eruption of Hekla, Iceland about 750 B.C. estimated (Thorarinsson *et al.* 1959, p. 154) to have produced 12 km³ when freshly fallen; the northern Rockies eruption supposed to have occurred between 500 B.C. and A.D. 500 (probably about 50 km³—see p. 444); the eruptions of Santorin in the Aegean about 1500 and 1450 B.C. (which archaeologists now associate with the disaster that overwhelmed Minoan Crete) and of Vesuvius in A.D. 79 (dust fall 3 m deep near the volcano and some deposits as far away as north Africa, Egypt and Syria); and certain eruptions in New Zealand (see p. 444) and the far south, the last mentioned identified by dust layers in Patagonia-Tierra del Fuego and on the bed of the Ross Sea near Mount Erebus. It seems probable that when account is taken of the effect of latitude of origin on the spread of the dust no known individual eruption in the equatorial zone gave a d.v.i. value of > 4000 to 5000 and the greatest ones in other latitudes no more than 1000 to 2000.

The greatest explosions in the latter stages of the Krakatau 1883 eruption were attributed to the sea getting into the active crater and, by its cooling power, forming temporary crusts of solidified lava which were soon blown away. This may well be a factor in the explosiveness of other volcanoes among the Pacific islands. However, other cases of very great explosions throwing up great quantities of dust and rocks have occurred when a long dormant, and supposedly extinct, volcano awakes to new life—as with Vesuvius in A.D. 79 (when the crater had become thickly wooded).

TABLE 6. VOLCANIC DUST EJECTED IN THE MAIN PRODUCING REGIONS

region	first reported eruption	average number of eruptions per century since 1500 (or later start of information) (d.v.i.)/E_{max} $\geqslant 100$	$\geqslant 250$	$\geqslant 1000$
Iceland	A.D. 1104	2.6	2.6	0.6
Mediterranean	*ca.* 1500 B.C.	1.9	1.3	0
Azores–Canary Islands	A.D. 1558	1.0	0.7	0
Mexico–West Indies–Andes (north of 20° S)	1539	4.0	3.8	0.7
Andes (south of 20° S)	1520 (1800 taken as datum, reports thought incomplete earlier)	1.8	0.6	0
Antarctic and Sub-Antarctic (south of 60° S)	1821	(1 to 4?)	?	?
New Zealand and SW Pacific Islands south of 15° S	1846	2.5	2.5	0.8
Indonesia, Moluccas, Celebes, Philippine Is. etc.	1500	8.8	8.8	2.6
Japan	*ca.* A.D. 800	1.5	1.5	0
Kamchatka, Aleutians, Alaska	1737	3.9	3.5	0

Sapper (1917, p. 133) pointed out that the volcanoes of the Pacific perimeter accounted for the vast majority of the purely explosive, 'ash' eruptions in the world (87 % of all eruptions of the years 1895–1913, according to his estimate). By contrast, the activity of the mid-ocean ridge volcanoes is mainly effusive and produced most lava. In this respect Iceland leads, having produced the greatest lava flows on Earth in post-glacial time—the country is the youngest volcanic land of so great extent, supposedly only 13 Ma old—and estimated to account for one third of all the lava produced on the Earth (16–17 out of about 50 km^3) since A.D. 1500. However, the production of tephra in Iceland is also great.

8. TIME DISTRIBUTION OF VOLCANIC DUST VEILS

Variations in time of the incidence of volcanic dust in the atmosphere must be considered important to meteorology because of the probable cumulative effects of in-radiation deficit in periods of frequent dust veils. These seem likely to operate most upon the radiation balance over the polar regions, where the insolation is always at low angles and the rays travel obliquely through the dust layers, and because of longer persistence of the dust over high latitudes than elsewhere. (There is some suggestion that poleward drift of the dust in the lower stratosphere proceeds with spurts each autumn (cf. p. 468).) Depletion of the incoming radiation in high latitudes must tend to increase the extent of ice surface, particularly sea ice and snow on land, owing to reduced summer melting, thus increasing the albedo and tending to maintain a deficit of radiation absorbed for some years after the disappearance of each dust veil. Repetition of this process might be expected to lead to a still longer-lasting build-up of sea ice and ultimately to lower sea temperatures in other latitudes.

The time and latitude distribution of injections of volcanic dust since A.D. 1500 is displayed in figure 20.

Evidence of the occurrence of epochs differing in total extent of volcanic activity over the world is not lacking. And suggestions of an association between the frequency of great volcanic eruptions and the extent of polar ice over the last thousand years or so are not hard to find. This is apparent in the records of both phenomena as they have affected Iceland (figure 21), where natural disasters from these causes have all along been the one threat overhanging the country's history: the records appear correspondingly reliable.

Records of Antarctic ice on the Southern Ocean are by no means so long or so complete as those of the Arctic ice about Iceland: something like continuous reportage only begins with the surge of exploration and seal-hunting activity by many nations about 1820. Since then the

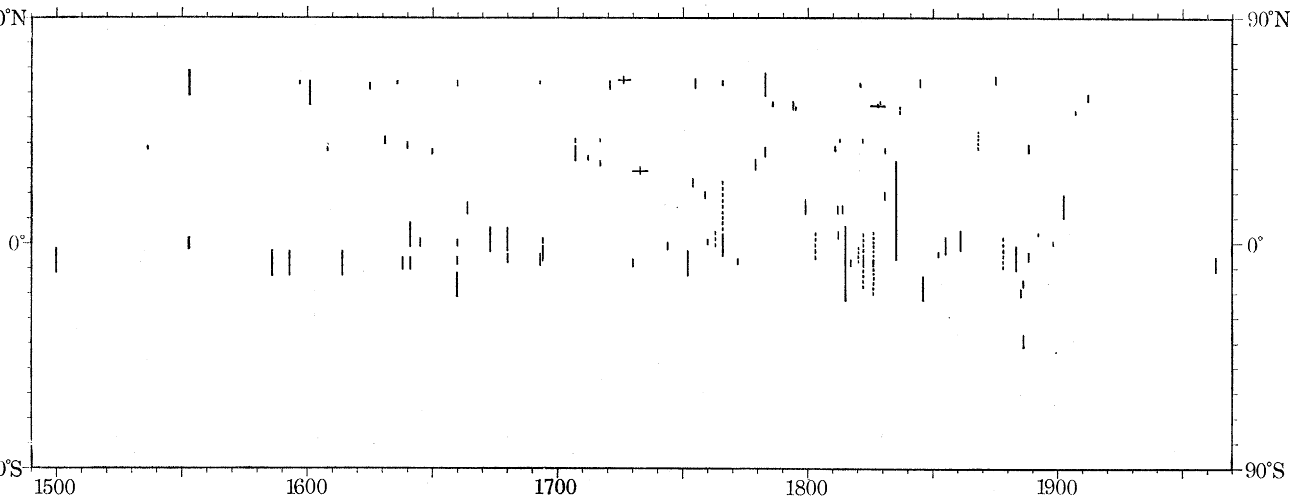


FIGURE 20. Injections of volcanic dust (d.v.i. ⩾ 100) since A.D. 1500. D.v.i. values entered in the year of the eruption at a position centred on the latitude of injection of the dust: vertical scale d.v.i. (1883) = 1000. Dust veils before about 1800–50 which were mainly south of 20° S likely to have remained unknown. The northern hemisphere record is probably complete as regards major dust veils: those of 1553–4 and 1601–2 are known from direct reports of the veil, though the eruptions concerned cannot be identified with certainty. Broken lines indicate d.v.i. values derived from temperature data alone, though in each of these cases the eruption concerned is believed to be identifiable.

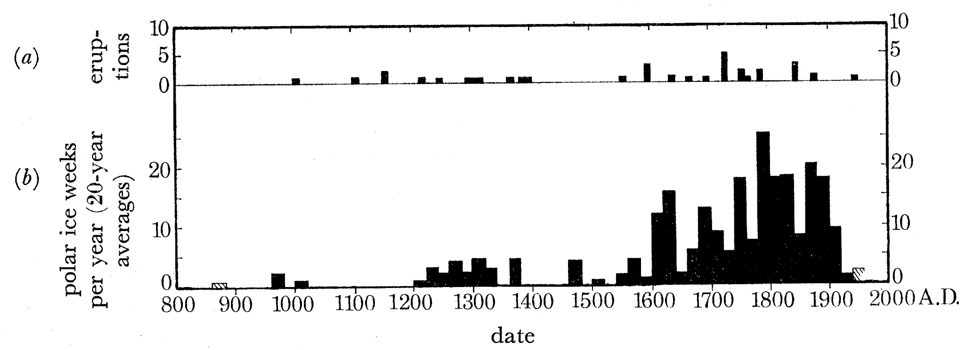


FIGURE 21. (a) Great volcanic eruptions in Iceland. (By decades. An eruption is counted twice if it extended into a second year. The eruptions of 1104, 1362, 1693, 1766, 1783–4, and 1875 should perhaps be further weighted.) (b) Incidence of Arctic sea ice in that area since A.D. 870. ▨, data incomplete for 20-year period and the incidence shown is the minimum possible.

several runs of extraordinarily bad ice years when large tabular bergs (broken off pieces of the floating ice-shelves where the inland ice of Antarctica reaches the sea), as well as pack-ice, were sighted in positions as far north as 35 to 40° S in various sectors—1828, 1832–4, 1840, 1844, 1850, 1854–7, 1888–90, 1892–6 (one small berg sighted as far north as $26\frac{1}{2}$° S in the South Atlantic in April 1894), 1898-9, 1904–7—seem to show some correspondence (allowing appro-

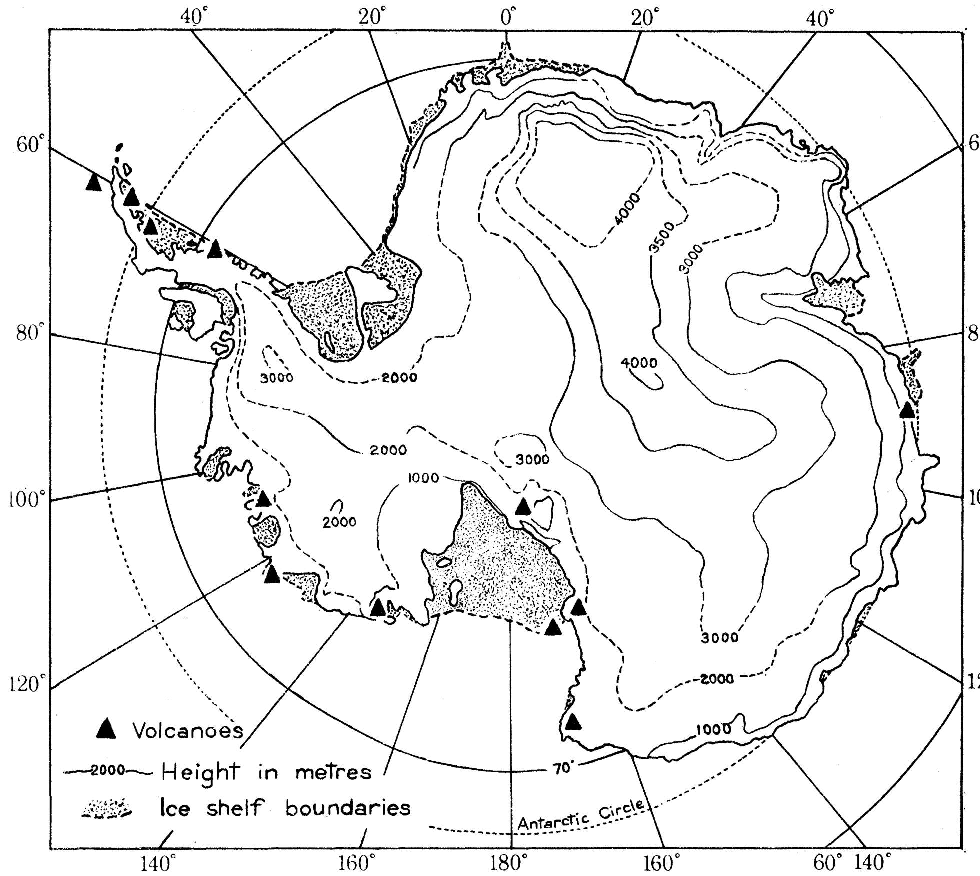


FIGURE 22. Volcanoes in Antarctica.

priate lags) with the phases of greatest volcanic activity in our chronology (cf. appendix I). Not all aspects of this can be explained simply by more sea ice forming (and less melting) when the sun is veiled by volcanic dust (as may be the case in the Arctic): for enormous bergs of what was originally land ice up to 160 km long figure amongst the Southern Ocean reports. It is possible that this unsolved riddle points to volcanic activity in about those years at, or off, the coast of Antarctica where at least ten volcanoes are known to exist south of the polar circle (map figure 22). Eruptions and associated submarine earthquakes, and tidal waves, in that area seem as yet the likeliest physical agencies for breaking up the ice shelves.

(*a*) *Fossil evidence of the last 20 ka*

The stratigraphy found in numerous profiles of bogs and lake sediments in southern South America examined by the Finnish geologist, Professor Auer (1956 (the names and positions of the places where the stratigraphy was studied are given by Auer on a map p. 8), 1958, 1959, 1965), reveals a sequence of layers of volcanic tephra that characterizes the whole region examined, between 38° and 55° S. The dates of the post-glacial waves of volcanic activity in the Andes indicated by a few carbon-14 tests upon the organic material next the dust layers in the Patagonian stratigraphy may have counterparts elsewhere in the world: this suggestion of Auer's has started a good deal of activity in tephrochronology, since it needs testing by many more radiocarbon dating experiments on dust layers in surface deposits on land and in the ocean bed all over the world, as well as by dating volcanic dust found in the ice-caps in the Antarctic and Arctic. Auer believes that his Patagonian results indicate some world-wide waves of enhanced volcanic activity, possibly brought about by stresses in the Earth's crust associated with post-glacial isostatic movements and changes of world sea level. This suggestion carries the implication that there may be a long-term influence of climate on volcanic activity as well as vice versa.

The waves of volcanic activity in the southern Andes during post-glacial time defined by Auer are:

I around 7000 B.C.
II about 3500 to 3000 B.C.
III about 500 to 200 B.C.
IV about A.D. 1500 to 1900.

Details are given in appendix IV, which also contains Auer's identification of four older ash layers in late glacial times between 15000 and 9000 B.C.

The only widespread volcanic ash layers found in the peat stratigraphy on the islands Tristan de Cunha (37° S, 12° W) and Gough (40° S, 10° W), situated on the mid-Atlantic ridge, have been radiocarbon dated to approximately 2345 years ago (a thin layer) and 4720 years ago (a thick layer at the base of the peat) (Wace 1961; Dickson, Holdgate *et al.* 1965). The ash is believed in both cases to be from volcanic outbreaks on the islands, though the dates agree with Auer's waves of Andean activity III and II respectively. It is thought that the volcanic activity around 5000 years ago may have been sufficient to destroy the previous vegetation of the islands.

There is some evidence (Auer 1959, p. 208, reporting work by S. Thorarinsson and P. Nejstadt) that eruption waves II, III and IV affected Iceland and Kamchatka as well as Fuego-Patagonia. In Europe (Mediterranean) III and IV seem to be identifiable as periods of enhanced volcanic activity, though little is known about any activity that might correspond to I or II.† The last period of volcanic activity in Europe north of the Alps, in the German Eifel (Laacher See and other craters and crater lakes), dated from 10000 to about 8000 B.C. (Averdieck & Döbling 1959; Straka 1954, 1956) centred about the Alleröd warm climatic period seems to coincide with one of Auer's earlier waves of volcanic activity in Pagatonia O_4 (see appendix IV). Auer reports that the same rhythm of volcanic activity may be traced in Alaska, II being prominent in the stratigraphy examined by Heusser. Of 14 important eruptions in Japan within the last 10000 years that have been radiocarbon dated (e.g. Kigoshi *et al.* 1962, 1963, 1964, 1965), four were within 1000 years centred about 2000 B.C. and another four in the last millenium B.C., mainly after 500 B.C. There were equal concentrations of the dates of

† It may be that the great explosions of the island of Santorin ($36\frac{1}{2}$° N, $25\frac{1}{2}$° E) in the Aegean, ascribed to 1500 and 1450 B.C. and the latter estimated by geologists from the height of the tidal waves as at least four times as violent as Krakatau in A.D. 1883, should be regarded as a late manifestation of Auer's wave II, dated about the end of the post-glacial warm climatic epoch. Another great Mediterranean eruption is thought to be indicated by the 'rain of fire and brimstone' which overwhelmed Sodom and Gomorrah, probably between 2000 and 1700 B.C. (The sites have since been submerged by a rise of the water of the Dead Sea, though this was much higher in the ice age. The Sodom and Gomorrah disaster, interpretation of which has always been controversial, may be the earliest surviving narrative description by witnesses of a volcanic eruption.)

earlier eruptions around 19000 and 11000 B.C. Nevertheless, it is as yet premature to regard these supposed world-wide waves of volcanic eruptivity as proven. Professor P. E. Damon (University of Arizona Geochronology Laboratories, personal communication 10 September 1966) writes that it is true that four of the five large ash falls in the western United States, which have been dated (Wilcox & Powers 1965), fall within the epochs suggested by Auer; but the latter are given such wide latitude as regards duration that there is a more than one in three probability that the coincidences are fortituous.

(*b*) *The last* 400 *years*

The incidence of reported eruptions in later times offers some further support to the concept of world-wide variations of eruptive frequency. It seems that great eruptions were rarer both in Iceland and the Mediterranean for at least 700 to 1000 years before A.D. 1600 than between 1600 and 1900. That there has been less frequent activity of the magnitude required to produce d.v.i. values $\geqslant$ 100 in all parts of the world since about 1900 stands out from the list of eruptions given in appendix I. The well-watched and reported eruptive activity of Vesuvius displays this pattern of variation over the past 2000 years, and it is believed to apply to the Aegean also. The Antarctic explorers of the period 1800–50 likewise found many volcanoes in a state of activity that they have not maintained since—though it is not apparent that there has been any diminution of activity in the South Sandwich Islands.

Before proceeding to more detailed examination one must consider the varying probability of reportage in different regions.

The dates from which documented reporting of great eruptions in the most active volcanic regions begins may be summarized as follows (largely after Sapper 1917):

Mediterranean and SW Asia	since Classical times
Japan	Mt Fuji and some other volcanoes from similar antiquity, but other parts of the Japanese region and Kuriles first reported at various later times and no likelihood of complete survey before the 18th century A.D.
Iceland	from 9th century A.D.
Azores and Canaries	from 15th century
Cape Verde Islands	from 16th century
Central America, West Indies and northern Andes	from 16th century
Chilean and Argentine Andes	from 16th to 17th century
Hawaii	from 1832
North American Rockies	from 1843
Java and Sumatra	from about 1500
Philippine Islands, Moluccas and Celebes	from about 1600
Other parts of East Indies	mainly by 1700
Samoa	from 1700
Kamchatka	from about 1725
Aleutians and Alaska	from about 1780

New Zealand and Polynesia	effectively from about 1800
Antarctic and sub-Antarctic islands	beginning about 1820, though no precision about moments of maximum activity before I.G.Y. (1957–8), and no certainty that an important dust veil over the southern temperate and polar zones might not have passed unobserved before this date—as may be involved in the problem of Thompson Island (appendix I and Lamb 1967).

There are just two regions, Iceland and the Mediterranean, from which our knowledge is probably complete as regards great eruptions (such as produce d.v.i. $\geqslant$ 100, (d.v.i.)/$E_{max} \geqslant$ 250–300) for the last 1000 years or more. Hence the interest of figure 21. Unfortunately Sapper's table (1917, p. 104 *et seq.*) of the number of eruptions reported each century (over longer time-intervals before A.D. 1500), registers only the increasing abundance with time of reports of lesser eruptions. By confining our attention to eruptions producing d.v.i. > 100 we shall obtain a more reliable survey of the variation of activity with time, and these are the only eruptions likely to be of meteorological importance through the radiation effects of the dust veil.

Reporting from the most active volcanic regions in the equatorial zone was established before 1700, and even largely begun before 1600, so summation of the cases which produced d.v.i. > 100 may provide a reasonable index of volcanic dust in the atmosphere over the northern half of the world over all this time, and particularly of the variations of its incidence from decade to decade. Though some known eruptions, for which all details fail, particularly in the equatorial zone, in the period 1500–1780 may have produced unreported dust veils—these have been entered in the list in appendix I without d.v.i. assessment—it seems likely most remarkable dust veils would have attracted comment and so have been reported. From 1780 the list should be substantially complete and at least the 25-year totals of d.v.i. from major eruptions must give a good representation of cumulative effects. The later the date the more unlikely it becomes that any eruption not assessed because of paucity of information produced any significant dust veil. Any defects in the series of index values after 1780 bear witness to the basic difficulty of deriving a numerical index from such diverse and partly inadequate reports. Before 1780 d.v.i. summations quoted in this work must be regarded as minimum ($\nless$) values, the approximation probably being the poorer the farther back in time. In particular, it may well be that if d.v.i. could be computed from world temperature variations at that time, the values of dust veil index for the 1680s and 1690s should be bigger.

For the southern hemisphere no complete summation would be possible before recent years, though contributions to the volcanic dust present in the atmosphere originating between about 0 and 20° S may be complete enough and must be taken into account in any northern hemisphere assessments.

Table 7 and 7*a* in appendix II give yearly values of d.v.i. for the northern hemisphere, obtained by confining attention to eruptions giving d.v.i. > 100 and allocating 40 % of the d.v.i. to the year of the eruption, 30 % to the next year, 20 % to the third year and 10 % to the fourth year, except in cases where a longer duration of the dust veil is known to apply—when the d.v.i. value has been apportioned at a linearly declining rate over the number of years concerned. A different, more uniform apportionment has been applied in a few cases of (rather minor) eruptive activity continuing over several years. Table 7, designed to facilitate investigation of possible relations between d.v.i. and weather, including temperature relations, presents a series of yearly d.v.i. values omitting all those assessments in appendix I where 'dust veils' have been deduced from the observed occurrence of a temperature deficit: the cases concerned are 1763, 1766, 1775, 1798, 1803, 1807, 1820, 1826, 1852, 1868 (in part) and 1878. Table 7*a* gives the yearly values that include these cases; in all these years eruptions occurred which may well have produced important dust veils, of which the temperature curve provides our only evidence.

Figure 23 displays the course of eruptive activity since the year 1500 in terms of volcanic dust in the atmosphere over the northern hemisphere, as given by 10-year means of d.v.i. (top curve – values plotted against the middle of the decade referred to) and by cumulative total of d.v.i. over the previous 25 years (second curve). The latter was chosen as a device for watching possible cumulative effects, e.g. on the Arctic ice or temperatures affected thereby. The lower curves in figure 23 present a number of climatic indicators, chosen as the longest available records representative of the world and of Europe or parts thereof. A still longer record of the occurrence of sea ice at the coasts of Iceland has been presented in figure 21 side by side with the occurrence of great volcanic eruptions in Iceland.

A degree of parallelism between the course of these temperature indicators and volcanic activity (dust veil index scale inverted) is apparent both in figures 21 and 23, particularly in the case of Iceland sea ice with its culmination in the 1780s. There can be no doubt about the trustworthiness of this date. The period of cold climate, known as the Little Ice Age, that produced its sharpest phases in Europe between about 1430 and 1850, clearly overlapped with the wave of volcanic activity here studied (Auer's IV) that was decreasing after 1840 and over by 1920. The mountain glaciers in some parts of the world reached their maximum extent about the time of maximum cumulative d.v.i.—in the Savoy Alps in 1820, Vatnajökull (Iceland) about 1850, in British Columbia about 1860. Nevertheless most indicators, including most glaciers, show the cold climate culminating earlier, generally in the seventeenth century, and the beginnings of it as early as 1200–1300. Thus, although the frequency of volcanic dust veils may have contributed to the low temperatures prevailing, or kept temperature low when it was inclined to recover, it can hardly be suggested that volcanic dust was alone responsible for the cold climate.

Moreover, the down-turn of temperatures about 1940 ± 15 years occurred at a time when volcanic dust cannot have played any part.

9. Effects of volcanic dust

A limited investigation to explore the degree of relationship between volcanic dust and weather or climate, in a number of cases where physical reasoning suggested there might be an association, produced some results here to be discussed which may serve to open up the subject. It was thought that the occurrence of any pattern of relationships with the d.v.i. numbers must depend upon general trustworthiness of the d.v.i. assessments, and give ground for believing that the assessments and chronology (appendix I) provide a basis for further research.

(*a*) *Temperature relations*

These were only studied in terms of decade means, because year to year variations of Köppens estimates of world temperature have already been used in the assessment for eruptions after

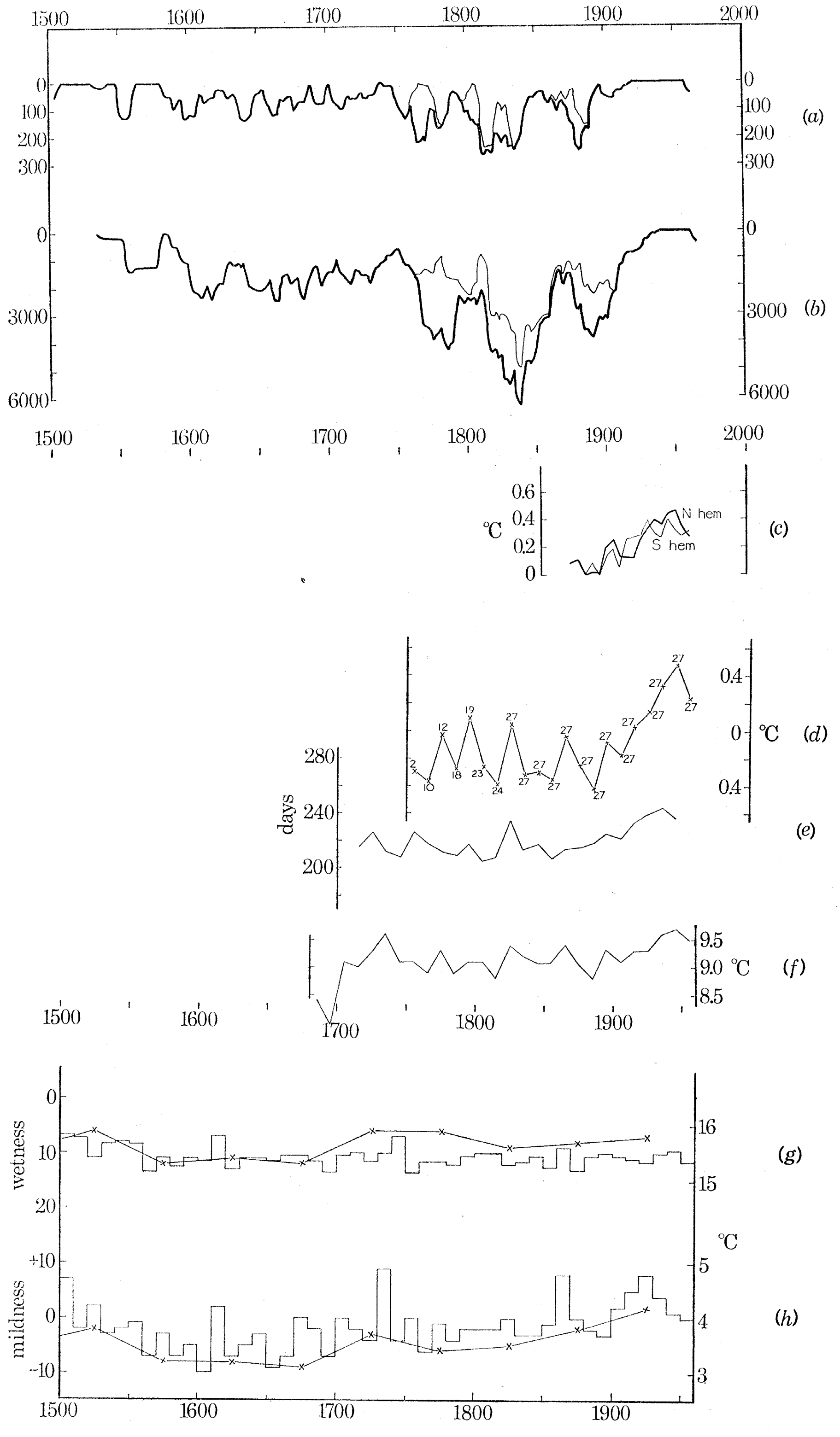


FIGURE 23. Volcanic dust over the northern hemisphere and various climatic indicators. (*a*) 10-year running mean values of d.v.i. plotted against the middle of the decade. (*b*) 25-year cumulative d.v.i. In (*a*) and (*b*) the finer line indicates values obtained by ignoring cases of dust veils assessed solely on evidence of temperature anomaly. There is some reason to suppose that if the d.v.i. values for the 1680s and 1690s could be computed from year to year variations of temperature the values for those decades should be much greater (see appendix I). (*c*) The computations of World Temperature averages, northern and southern hemispheres (bold line northern hemisphere), are 5-year means due to Mitchell (1961). (*d*) The variations of annual mean temperature over Europe are decade means for stations in the area 40–60° N, 0–30° E due to Callendar (personal communication), the numbers of stations used being shown against each point. (*e*) Length of ice free season on R. Neva at Leningrad by decades. (*f*) The average temperatures of the year in central England are taken from Manley's (1958, 1964) series for central England. (*g*) The summer wetness index and (*h*) winter mildness index, decade values for England, come from Lamb (1966, pp. 97–98, 217–221) and the corresponding 50-year average temperature values (×——×) for central England, derived by regression equations from the index values for periods before the beginning of Manley's series of homogenized thermometric observations, come from Lamb (1966, p. 186).

1750 and between that date and 1882 were in most cases the main plank on which the assessment was based. Most interest may therefore be attached to relations established outside those years. D.v.i. values for dust veils evidenced solely by temperature deficit have been omitted from all the tests here described.

(i) *Correlation coefficient connecting average temperature of the northern hemisphere* 0–60° N (Mitchell 1961) *with d.v.i. decade values* 1870–1959:

$$r = -0.94 \quad (9 \text{ pairs}).$$

This appears significant beyond the 1 % level, but is accepted with reserve because the short record available was known to cover the period of closest parallelism and because of the downward tendency of temperature after 1940.

One or two probes for relations with cumulative d.v.i., and with change of d.v.i. from the average value of the previous 50 years, produced lower values of r, not significant.

(ii) *Correlation coefficient connecting average temperature of the European region* 40–60° N, 0–30° E (Callendar, personal communication) with d.v.i. decade values 1751–1960:

$$r = -0.70 \quad (21 \text{ pairs}).$$

This appears significant at the 0.1 % level, but again the trends of the last 20 years do not fit the relations.

One or two probes for relations with changes of d.v.i. or changes of temperature from the average values of the previous 50 years indicated lower values of r, not significant or barely significant at the 5 % level. It seems therefore that the main relation found is between the general course of the decade to decade variations of temperature and d.v.i.

(iii) *Correlation coefficients connecting average temperature in central England* (Manley 1958, 1964) and d.v.i. decade values 1670–1959:

Summer temp. June to August $r = -0.29$ (29 pairs)

Not statistically significant (approaches significance at the 10 % level).

Winter temp. December to February $r = -0.49$ (from 1680, 28 pairs)

Significant at the 1 % level.

Average temperature of the year $r = -0.42$ (from 1680, 28 pairs).

Significant at the 5 % level.

These figures indicate temperature relations far beyond the decades in which year-to-year temperature variations played much, or even any, part in the d.v.i. assessments.

The correlation coefficients quoted, connecting temperatures and d.v.i. values in the same decade, would have been materially higher but for an apparently (half a decade) delayed reaction of the temperatures to the great dust veils of the 1830s and by the cool westerly summer weather in Britain of the period 1900–30 when the general circulation seems to have been at a very long-term maximum of strength, associated with the twentieth-century climatic optimum.

Probes for relations with changes of temperature or changes of d.v.i. from the averages of previous decades, produced only lower values of r, not significant. Importance was attached to testing with d.v.i. taken as departure from the 100-year mean of the period in which the decade was situated, as a way of allowing for possible incompleteness of the dust veil record in the early decades used. The fact that this produced only lower values of the correlation coefficients seems to indicate that the record may be substantially complete as regards dust veils great enough to have any climatic effect.

(iv) *Correlation coefficients connecting the amount of Arctic sea ice at the coast of Iceland* (*Koch's index*— Lamb & Johnson 1961) *with d.v.i. decade values* 1780–1959:

$$r = +0.61 \quad (18 \text{ pairs}).$$

Significant at 1% level.

An equally high correlation coefficient (+0.61) was found when the decade values of the ice were compared with 25-year cumulative d.v.i. totals averaged over half a century ending with the decade in question. In respect of ice on the polar seas, therefore, there is some ground for supposing that the effects of successive eruptions not too far apart may be cumulative.

Despite the high values of r, the association apparently failed in the remarkably ice-free years 1840–54. This period of remarkably open water has been tentatively attributed to the effects of the extraordinary wind circulation patterns of the 1830s (extraordinarily much blocking and rather strong, though variable, circulation patterns in high latitudes) in breaking up and removing ice from the Arctic.

(*b*) *Relations with the atmospheric circulation* (*long term*)

Since there were some indications in the temperature correlations of a limited degree of association between volcanic dust and the decade to decade climatic trend, decade values of circulation indices were explored in the same way as for temperature using the longest available sequences of charts and indices. For a small area of northwest Europe sufficiently reliable isobars of monthly mean pressure each January and July go back to 1750, provisional isobars (pressure values unreliable) to 1680 (Lamb & Johnson 1959, 1961, 1966).

(i) *Correlation coefficients connecting strength of the general circulation over the North Atlantic in January* (*overall range of pressure between regions of highest and lowest monthly mean pressure*) *and d.v.i., decade values* 1790 *to* 1959:

$$r = -0.41 \quad (17 \text{ pairs}).$$

This figure has about a 10% probability of occurring by chance. When, however, the same circulation index values were compared with 25 year cumulative d.v.i. in the same decade measured as departure from the average 25-year cumulative d.v.i. of the half century ending with that decade

$$r = -0.64 \quad (17 \text{ pairs}),$$

which appears significant at the 1% level. The latter result suggests an association between long-term cumulative effects of volcanic dust in the atmosphere and weakened atmospheric circulation; though, if standing alone, this might be treated as just one symptom of the near coincidence of the Little Ice Age cold epoch in all its manifestations and the wave of volcanic activity of A.D. 1500 to 1900. Yet it clearly tends to support Wexler's (1956) contention that repeated volcanic dust veils must weaken the atmospheric circulation—a contention that found some support in another analysis of the general circulation in the Little Ice Age (Lamb 1963, pp. 137–41, 146).

(ii) *Correlation coefficients connecting the strength of the general circulation in July with d.v.i. decade values*:

Correlation coefficients between North Atlantic pressure gradient indices in July since 1790 and contemporary d.v.i. values were effectively zero. When compared with the same measure of cumulative d.v.i. as for January the values rose to $r = +0.28$ for northwesterly winds over northwest Europe and $r = -0.30$ for southwesterly winds over the Newfoundland Banks, both however far from being significant. The high-summer (July) wind circulation in the North

Atlantic region before 1850 is, however, known to have been strong, possibly because of the excess of Arctic ice emerging towards low latitudes in the sector (Lamb & Johnson 1961), and its long-term changes may therefore be anomalous.

Decade values from 1860 to 1959 of an index of the pressure gradient for southerly winds in July over the east Asia–Japan region, an aspect of the Asian summer monsoon, and d.v.i. gave correlation coefficients for contemporary values

$$r = -0.29 \text{ (not significant)}$$

and

$$r = -0.64 \text{ (significant at the 5\% level)}$$

in the case of 25-year cumulative d.v.i. taken as departure from its average value over the previous half century.

(iii) *Correlation coefficients connecting the longitudinal position of key features of the North Atlantic wind circulation and d.v.i. decade values.*

Lamb & Johnson (1959, 1961) have shown that about 45–55° N the longitudes occupied by lowest pressure over the western Atlantic, by maximum pressure in the northern extension of the 'Azores' anticyclone and by minimum pressure in the European sector (in winter the col between oceanic and continental anticyclones) all tend to be farther east when the general circulation is strong than when it is weak. This provides something that can be measured on the earliest charts, based mainly on wind observations, before monthly mean pressure values were reliable.

Comparisons were therefore made between decade values of d.v.i. and the longitude of the January col normally over western Europe, using the provisional charts still under compilation by the author, to extend the survey back to 1680.

The correlation coefficient connecting simultaneous values was effectively zero.

When the longitude of the col was compared with 25-year cumulative d.v.i. expressed, as before, as departure from the average of the previous half century

$$r = -0.28 \quad \text{(28 decades).}$$

The negative sign implies that a western position of the col tends to accompany large amounts of volcanic dust. This just approaches the 10 % level of significance, but it may be important to notice that the correlation is only 'spoilt' by the 1830s the decade already mentioned for its highly abnormal circulation and in connexion with slower than normal response of the climate to the very great dust veils. If this one decade were ignored, a correlation coefficient significant at the 1 % level would be obtained.

There is therefore some further ground for pointing to an association between cumulative effects of volcanic dust and weakening of the long-term averages of the wind circulation in winter over the North Atlantic.

Similar comparisons of d.v.i. with the July circulation produced no correlation coefficients approaching significance—as in the case of indicators of the strength of the summer circulation in the North Atlantic sector since 1790, already discussed. A contributory reason could, however, be that neither the mean position of highest pressure in the Atlantic ridge nor of lowest pressure over northern Europe in July lies within the part of the map that is firmly grounded upon observation in the years before about 1750.

(*c*) *Relations between great volcanic eruptions and year-to-year variations of the atmospheric circulation*

Defant (1924) studied the variations of the average pressure difference over the North Atlantic between 30 and 65° N over the year as a whole from 1881 to 1905. He found that in the years in which the fourteen greatest volcanic eruptions of that period (not all here assessed as giving d.v.i. > 100) took place, including one in 1886 at $38\frac{1}{2}°$ S, all assessed as b_1 or b_2 eruptions by Sapper (1917, 1927), the pressure difference was on average 11 % less than the overall mean figure. Singling out the four greatest eruption years of that period—1883, 1886, 1888 and 1902—Defant found an average 18 % weakening of the atmospheric pressure gradient in the eruption year, followed by values averaging 9 % *above* normal in the next year and 17 % stronger than normal in the second year after the eruption. Defant attributed the abnormal strength of the circulation about two years after an eruption to abnormal equator–pole temperature difference due to polar cooling permitted to develop by reduced meridional exchanges of air in the period of weak circulation occurring while the dust veil was general and dense.

There appeared to be a random element in the month to month variations, but most months were affected by anomalies in the same sense as the relevant year after an eruption. Defant also pointed out that the amplitude and period of the variation of circulation strength after an eruption must be expected to depend upon the quantity of dust thrown up and the height reached by it.

The dust veil assessments and chronology in the present work, considered in relation to the long series of monthly mean pressure and wind charts now available, make it possible to repeat Defant's studies in somewhat greater detail and in relation to many more eruptions. A separation of the eruptions by latitude thereby becomes possible. This is certainly advisable since our discussion of what is now known of the spread of dust in the stratosphere by the high-level wind circulation (pp. 453–459) indicates that the intensity and duration, even perhaps the very nature, of the effects of the dust on prevailing temperatures and south–north temperature gradients may differ with the latitude of injection of the dust.

Figure 24 displays the average strength of the general circulation over the North Atlantic, measured by the overall range of monthly mean pressure between the regions of highest and lowest pressure in every month of the year, from 2 years before a great volcanic eruption to 4 years after the eruption. The vertical scale is measured in percentage of the overall average value of this circulation index for each month of the year from 1873 to 1964. The curves in figure 24*a* show how the circulation varied after the three great eruption years in low latitudes—1883, 1888 and 1902—and (thin line) with the inclusion of the more uncertain dust veil of 1878. This circulation index followed the same course as described by Defant, falling below normal strength in the eruption year and then rising to above normal values in the next two years, but only the weakening in the eruption years appears as a statistically significant anomaly with just a 2 to 3 % probability of occurring by chance.

The curves in figure 24*b* show the average course of the index in relation to the seven great eruption years in equatorial or northern latitudes 1875, 1878, 1883, 1888, 1902, 1907 and 1912 (thin line) and in relation to the three considerable eruptions in high northern latitudes in 1875, 1907 and 1912 (bold line). None of the circulation anomalies attain statistical significance in this case. What is interesting, however, is that with these high latitude eruptions there is no trace of weakening of the circulation in the eruption year: rather it appears that some tendency for stronger circulation, as would correspond to enhanced equator–pole thermal difference (e.g. arising from the effect of the dust being confined to high northern latitudes), sets in immediately.

For the circulation in the months of January and July in the neighbourhood of the British Isles the long series of monthly charts going back to 1750 (fully analysed), and to 1680 (incomplete analysis with only provisional isobars which have been shown by test to give a reliable indication of the pattern over northwest Europe but must not be used for pressure values), make it possible to study the variations after 43 eruption years (35 since 1750) with northern hemisphere d.v.i. values > 100.

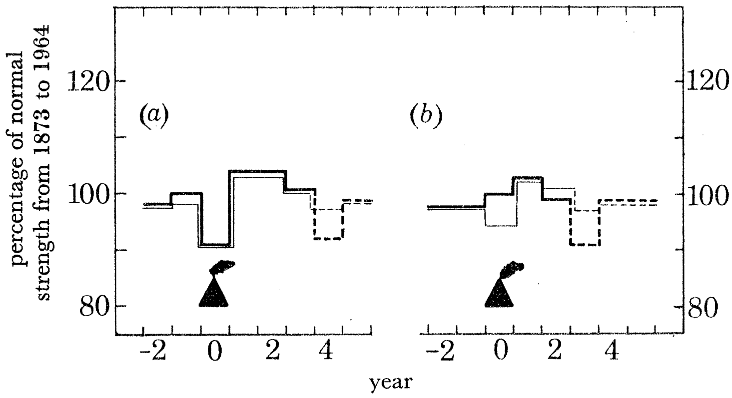


FIGURE 24. Course of the annual mean values of the North Atlantic overall range of monthly mean pressure (between regions of highest and lowest mean pressure) index of circulation strength from 2 years before to 5 years after years of great volcanic eruptions in various latitudes. (The standard deviation of the individual yearly values 1873 to 1964 is 7.74 % of the overall average value of the index.) (*a*) ——, Average for 1883, 1888 and 1902 eruptions in low latitudes; ——, average for 1878, 1883, 1888 and 1902 eruptions in low latitudes. (*b*) ——, Averages for 1875, 1907 and 1912 eruptions in high northern latitudes; ——, averages for great eruptions in all northern and equatorial latitudes between 1873 and 1912: namely, 1875, 1878, 1883, 1888, 1902, 1907 and 1912. Broken lines indicate portions of the curves complicated by possible overlapping effects of eruptions in different years.

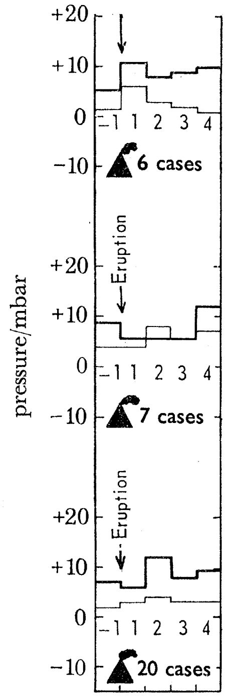
Figure 25 shows the average course of the pressure gradients 50–60° N, 0° E (bold line) and 10–0° E, 55° N (thin line), respectively favouring westerly and southerly winds in the British Isles and North Sea region, in January. The cases are separated according to the latitude of the eruption. Two years in which great eruptions occurred in more than one of the chosen latitude zones were left out. Few of the differences of the circulation indices from one year to the next within these still fairly small samples of up to 23 comparable eruption years attain statistical significance. The anomalies at some lags after an eruption do, however, appear to represent statistically significant departures from the 215-year average values of the indices over the period 1750–1964.

The negative anomaly of the pressure gradient for westerly winds in the first January after an equatorial zone eruption appears possibly significant, at about the 5 % level; the positive anomaly of this gradient in the second January after an equatorial zone eruption also appears significant at, or beyond, the 5 % level (on the basis of 20 great eruption years after 1750). There were no significant anomalies of westerly gradient in the years following high latitude eruptions, but in general the pressure gradient tended to be rather above normal right from the first January after the eruption. The variation of this circulation index after eruptions in the subtropical zone appeared rather similar to the case of equatorial zone eruptions, but in general more years elapsed before the positive anomaly occurred; the average anomalies of this group of cases barely reached statistical significance at any point.

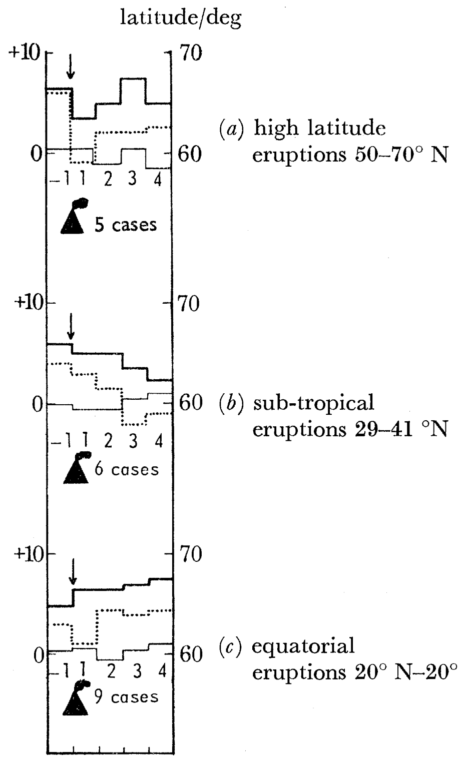
Positive anomalies of the pressure gradient for southerly winds in the North Sea region in January appear significant at or near the 1 % level in the second January after eruptions in the equatorial and subtropical zones (cases since 1750). After eruptions in high latitudes this index also shows a different course: the positive anomalies of the gradient for southerly winds in the first January after the eruption appear significant at the 5 % level (cases since 1750). From the first year on the tendency to excess winter southerly declines and soon vanishes.

To make possible a fuller appreciation of the similarities and the differences between the atmospheric behaviour after different eruptions, the course of the January circulation indices used after every individual eruption year studied since 1680 (provisional isobars before 1750) is reproduced in figure 27 in appendix III of this report. Often further eruptions occurred within the first year or two over which the atmospheric effects after one eruption were being followed. Such years were counted twice, or more, in the statistics—e.g. as the second, third or fourth year after one eruption and the first year after another eruption.

Figure 26 displays the corresponding behaviour of the circulation in July in the British Isles–North Sea region after eruption years. In addition, the latitude of lowest pressure at



25. Januarys after eruption



26. Julys after eruption

FIGURE 25. Differences of average pressure in the British Isles–North Sea region in January, from the last January before to the fourth January after a volcanic eruption (restricted to great eruptions producing d.v.i. > 100 over the northern hemisphere). All cases since 1750. ——, 50–60° N, 0° E average pressure difference in January 1750–1964 + 8.7 mbar, standard deviation of the values for the individual Januarys 6.5 mbar; ——, 10–0° E, 55° N average pressure difference in January 1750–1964 + 2.5 mbar, standard deviation of the values for the individual Januarys 3.3 mbar. The changes of average pressure difference here shown are clearly reduced by the cases of a few great equatorial zone eruptions, which were followed by the characteristic drop and rise of the 50–60° N index 1 or even 2 years later than in the average sequence.

FIGURE 26. Differences of average pressure in the British Isles–North Sea region and of latitude of the lowest mean pressure at 0° E in July, from the last July before to the fourth July after a volcanic eruption (restricted to great eruptions producing d.v.i. > 100 over the northern hemisphere). All cases since 1750. ——, 50–60° N, 0° E average pressure difference in July 1750–1964 + 7.0 mbar, standard deviation of the values for the individual Julys 3.0 mbar, ——, 10–0° E, 55° N average pressure difference in July 1750–1964 + 1.3 mbar, standard deviation of the values for the individual Julys 2.3 mbar; . . . . ., average latitude 1750–1964 of lowest mean pressure at 0° E in July 63.5° N, standard deviation of the positions for the individual Julys 4.4° latitude.

0° E—a parameter of importance to the character of the summer in Britain—is indicated by a dot and dash line. There are fewer cases than for January, because when the exact date of an eruption within the year is not known it is impossible to say which is the first July after the eruption.

There is on average a southward displacement of the pressure minimum in the first July after a great eruption towards, or over, the British Isles, represented by an anomaly that measured against the whole population of the 215 Julys since 1750 appears significant beyond the 1 % level in the case of equatorial zone eruptions and at the 5 % level in the case of high latitude eruptions. In the Julys after the subtropical zone eruptions, as in January, the circulation underwent a slower variation of somewhat different character: the southward displacement in the third July might in these cases be considered statistically significant.

The strength of the gradient for westerly winds seemed to show no significant anomalies in the Julys after major eruptions, except in the case of the weakening observed in the first July after high latitude eruptions (significance about the 1 % level suggested by the six cases since 1750).

Examination of the other July pressure gradient index suggests a tendency for more than usual northerly components in the British Isles–North Sea region developing over the years after high latitude eruptions and attaining significance at about the 5 % level in the second and fourth Julys after the eruption. The same tendency is apparent in the second July after equatorial zone eruptions (significance of the anomaly also about the 5 % level for the nine cases after 1750).

The course of the July indices after individual eruptions is shown in figure 28 in appendix III.

All the fixed time-lag relations indicated by the studies described appear weaker than might be obtained by considering a variable time taken for the effects to mature and equating the years of weakest circulation, or southermost depression tracks in summer near the British Isles, regardless of whether they fell one, two or more years after the eruption. Such a method of investigation has been rejected here because of its element of subjective judgement. Yet it must be expected that the effects take more or less time to mature, and then to die away, according to the magnitude of the dust output in the eruption, the latitude of the eruption, height of the maximum dust concentration in the stratosphere, initial state of the atmospheric circulation, etc. Specifically, it appears that the effects indicated by these investigations as normal for the latitude of the eruption occurred after a longer than normal lapse of time (i.e. one or two years late) in the case of the eruptions in 1694–5, 1815–16, 1835–8, 1868–71, 1883–4, among which very great eruptions in low latitudes (years with northern hemisphere d.v.i. > 500) are prominent.

10. Concluding comments

Some associations between great volcanic dust veils (d.v.i. > 100 in one hemisphere) weather and climate are indicated by these brief preliminary investigations. But they differ according to the latitude of injection of the dust. They also take more or less time to mature in different cases, probably depending on the initial state of the atmospheric circulation and its current reaction to other influences. They seem to take longest in the case of very great eruptions (world d.v.i. ⩾ 1000) in low latitudes.

That many of the fixed time lag relations should appear statistically significant, even though we observe some variation in the time taken for the effects to develop and die away, presumably indicates that (possibly for reasons connected with the summer–winter changes of the general circulation) there is a preferred or most-frequent time-scale. Nevertheless, the relations must appear weaker than they would if the rate of development were really the same in all cases.

The level of statistical significance of many of the results mentioned is rather low. This is not surprising in view of the crudity and diversity of the original, often purely descriptive, observational material from which numerical assessments of dust veil index (d.v.i.) had to be made. The atmospheric pressure fields prevailing in the Januarys and Julys before 1750 were also derived from observations among which the only fully trustworthy numbers were the frequencies of different wind directions. Moreover, no one individual month can be relied upon to show the weakening or strengthening of the circulation characteristics of most of the year in which it falls. When great eruptions occurred in too quick succession, their effects may have overlapped and sometimes may have been out of phase, thus tending to obscure any general relation. Remembering these shortcomings, the high proportion of possibly significant results indicated by a very brief investigation, which picked on volcanic dust–weather associations that might be expected to occur on some or other physical grounds (e.g. reduced inradiation, reduced temperature and reduced temperature gradients or, if the reduction of incoming radiation were in high latitudes only, increased equator to pole temperature gradient) in itself constitutes a reason for believing the general proposition that great volcanic dust veils affect the atmospheric circulation, weather and climate.

Cumulative effects of repeated dust veils appear to be important in some ways, particularly as regards the amount of ice on the polar seas and, possibly because of that, has an effect on some aspects of the general atmospheric circulation.

Despite the approximate contemporaneity of the recent period of cold climate, the so-called Little Ice Age of 1430 to 1850, and a wave of volcanic activity in which great eruptions seem to have been abnormally frequent in many parts of the world between 1600 and 1900 or thereabouts, it is obvious that volcanic dust is not the cause of all decade-to-decade and century-to-century climatic differences. There seems (figure 23 and appendix II) to have been less volcanic dust in the atmosphere since about 1915 than for a long time before that. In particular, volcanic dust cannot be held responsible for a number of the individual severest winters and most abnormal winter circulation patterns affecting Britain and Europe: 1962–3, those of the 1940s and 1739–40 are obvious cases, but this comment is probably also true for 1794–5, 1657–8 and 1607–8. Nor does the greatest cumulative d.v.i. (*ca.* 1840) coincide with the period of lowest average winter temperature (in the seventeenth century), or the minimum of cumulative d.v.i. (continuing from 1940 to 1962) with the period of highest average winter temperature in Britain and Europe (about the 1920s). It is less easy to be sure that volcanic dust veils have not played a part in most of (possibly all) the wretchedest summers in Britain e.g. 1912, 1903, 1879, the 1840s, 1816, the 1760s, 1725, 1695, 1594, 1586 and 1587, 1555—possibly even 1954 was a similar case owing to high level debris from a ground explosion of hydrogen bomb in the equatorial Pacific early that year (Staff members of the Forecast Research Laboratory, Tokyo 1955.)* Similarly, in New England the three coldest years and the three coldest summers in the

* *Note added in proof* 23 *March* 1970. Professor Takeo Yamamoto of Yamaguchi University, Japan reports (personal communication, February 1970) that the four greatest famine-producing bad harvest years in the record from 1599 of the Iwate prefecture in the northeastern part of the main island of Japan were 1695, 1755, 1783 and 1838, all suggested in this report as years with major volcanic dust veils. (However, 1560 seems to have been at least comparably bad in Kyoto.) In the summer of 1755 the snow on a mountain less than 900 m high failed to melt, suggesting that the mean air temperature in July and August remained more than 4 °C below the modern average.

180-year temperature record for New Haven, Connecticut, from 1780 to 1960 (Landsberg 1967) were all notable volcanic dust veil years with northern hemisphere d.v.i. ranging from 525 to 135 (average value 361) compared with the 180-year average value of 68. Von Rudloff (1967) has pointed out that the period of highest average summer temperatures and most sunshine in central Europe during the available instrumental record was 1942–53: this agrees fairly closely with the minimum of volcanic dust indicated by cumulative d.v.i. zero over the previous 30–40 years and allowing increasing transparency of the atmosphere as the finest dust in the stratosphere settled out.

Wagner's (1940) treatment of climatic variations indicated that either this settling out of the volcanic dust or some small increase in the solar constant were the likeliest causes of the increased vigour of the general wind circulation from the nineteenth to the early twentieth century; though neither could be conclusively demonstrated at the time. The present study may improve our understanding of the role of volcanic dust in the climatic changes of the last thousand and of the last hundred years, but indicates that it is not operative in some important cases of temperature drop. Volcanic dust seems to have something to do with some anomalous years and decades, but cannot be invoked as the cause of the differences between the most unlike climatic epochs within the last millenium.

Similarly, Auer's waves of volcanic activity over the last 10 ka and more, even if they be substantiated, suggest no case in which frequent occurrence of volcanic dust veils characterized a whole climatic era; therefore the character of none of these is likely to have been determined by volcanic dust. There are indications rather that waves of volcanic activity tended to occur as a complication in the latter part of several eras of fairly stable (i.e. prolonged) climatic character and, in the case of warm climates, the dust may have been instrumental in disturbing the equilibrium and bringing the epoch to an end. (It may be that in the course of any long climatic era shifts of stress in the Earth's crust are ultimately produced by progressive changes of glaciation and sea level that are liable to cause a wave of volcanic activity, even though the rigidity of the uppermost part of the crust is considerable). In the case of the cold climatic period after A.D. 1500, covered by this work, it seems likely that the frequent occurrence of dense volcanic dust veils, particularly between 1780 and 1840, tended to prolong a cold climate which from 1700 onwards was tending to give way to warmer conditions.

Budyko (1968*b*) has lately taken up again the suggestion that volcanic dust may have caused the Quaternary ice ages. His presentation starts with the assumption (1968*b* p. 11) that the frequency of great volcanic eruptions in the last 100 years (taken as four) is a fairly close approximation to the overall average frequency for the entire Quaternary (last million years); and, secondly, that eruptions are independent of each other and subject to purely random variations of frequency, so that an estimate of the maximum frequency of great eruptions in any short time span such as a century or less to be expected to occur once in a million years can be calculated by statistical methods. The same would apply to the probable number of occurrences of various lower frequencies. On this basis he concludes that 40 great eruptions in one century should occur once in 10 ka, 130 great eruptions in a century once in 100 ka, and that a frequency of 100 great eruptions within 5 years might be expected once in 1 Ma. Even 50–100 great eruptions in one century would, according to Budyko, reduce the direct radiation by 10–20 % over the whole century and lower the average air temperature by 1–3° C. In the light of the investigations with which this report is concerned, Budyko's first assumption appears to be no better than a (very insecure) guess. His final conclusion seems, moreover, to imply a more random distribution of the ice ages in time than most dated evidence suggests.

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Appendix I. Chronology of volcanic eruptions and dust veils in the atmosphere A.D. 1500 to 1968

Bold print indicates eruptions that have generally been described as 'great' by previous compilers and for which a dust veil index $\geqslant$ 250 has been computed or appears probable.

Dust veil index computation methods are described in text, pp. 470–476. Suffix numbers indicate formula used. Suffix 'av' indicates average d.v.i. for Sapper *b* category, adjusted for E_{max} appropriate to the latitude of the eruption. Suffix 'adj' indicates d.v.i. further adjusted from the Sapper *b* category mean value. (d.v.i.) values in brackets are estimates.

Italics distinguish *total veil* assessments from the figures given for individual contributory eruptions.

Braces beside the date column indicate which eruptions and eruption years are supposed to belong to each total.

The column headed (d.v.i.)/E_{max} indicates the magnitudes of the eruptions as dust producers without regard to the area over which the dust veil may have been spread by the general circulation of the atmosphere.

Letters in brackets indicate sources of information (see text, p. 470).

Sapper's assessments are explained in the text (p. 429).

'?' against unassessed eruptions indicates some suspicion of an important eruption.

Against completely unassessable eruptions no mark is entered.

			Sapper's assessment of magnitude:	estimated or assumed total dust for	dust veil index								
					total veil in groups of	from individual eruptions							
date	volcano	position	*a*, lava *b*, tephra	d.v.i. / km³	eruption years	World	Northern Hemisphere	Southern Hemisphere	R_{Dmax}	T_{Dmax} / °C	t_{mo}	E_{max}	d.v.i. / E_{max}
1500	**Java** (S, Sh)	6–8° S, 106–114° E	b_1	—	—	$\mathbf{1000}_{av}$	500	500	—	—	—	1.0	**1000**
		Hennig (1904) quotes a report of 1505 in central Europe that a high-level haze, or smoke, for long dimmed the Sun in that year											
1510	Trölladyngja, Iceland (K)	65° N, 17° W	—	—	—	—	—	—	—	—	—	—	—
		Erupted also in 1389											
1510	Hekla, Iceland (K)	64° N, 19½° W	—	—	—	—	—	—	—	—	—	—	—
ca. 1520	Tierra del Fuego	53–55° S, 65–70° W	—	—	—	—	—	—	—	—	—	—	—
		Despite conflicting interpretations in the various English language accounts, it seems most probable that it was volcanic fires, since extinct, that caused Magellan to call the island Fireland when first seen in 1520: unlikely however that he happened to encounter the most vigorous phase of the eruption. (For this reason the eruption date is given as '*ca.* 1520')											
1535–7	Etna (K, Sh)	38° N, 15° E	a_2b_2	—	—	150_{av}	150	—	—	—	—	0.5	**300**
		Classed in (K) as great eruption											
1538	Vesuvius (K)	41° N, 14° E	—	—	—	—	—	—	—	—	—	—	—
		Not mentioned in a history of the eruptions of this volcano by Lister 1960											
1539	Pichincha, Ecuador (K)	0° S, 78½° W	—	—	—	—	—	—	—	—	—	—	—
		Sapper gives Pichincha as in a state of continual eruption in the 16th century, the total product (in terms of lava?) equivalent to one 1st or 2nd order eruption											
1539	Popocatepetl, Mexico	19° N, 98½° W	—	(0.5)	—	50_3	33	17	—	—	(24)	1.0	50
		Tephra deposit observed at places 40 to 50 km distant to NW, NE and SE											
1550	Tala (S, Sh)	?	b_{2-3}	—	—	70_{av}	35	35	—	—	—	1.0	70
		(Sh) gives position in the Molucca Is. near 2° N, 128° E, but the name suggests Talaga, in the same island group, near 2° S, 126° E (or possibly Taal, Luzon 14° N, 121° E)											
1553	Pichincha, Ecuador (K)	0° S, 78½° W	—	—	—	($\nless$ 500)	($\nless$ 250)	($\nless$ 250)	—	—	—	1.0	($\nless$ **500**)
		Remarkable purple sunset afterglows in Scandinavia summer 1553. (Royal Society 1888, p. 384)											
1554	Hekla (K)	64° N, 19½° W	—	—	—	—	—	—	—	—	—	—	—
		Repeated observation of a high level haze over central Europe: 'the sun often veiled about in darkness as if threatening the doom of mankind'. (K) records extensive 'dry fog' in 1557											
		Classed in (K) as great eruption of Hekla. But the dense and prolonged dust veil may imply more than one source											
		Total veil 1553–54:	—	—	($\nless$ *1000*)	—	($\nless$ *1000*)	—	—	—	—	0.3?	$\nless$ *1000*
1558	Teneriffe (K)	28° N, 17° W	—	—	—	—	—	—	—	—	—	—	—
1565	Pacaya, Guatemala (K)	14° N, 91° W	—	—	—	—	—	—	—	—	—	—	—
1569	Citlaltepetl, Mexico	19° N, 97° W	—	—	—	10_3	7	3	—	—	—	1.0	10
		Tephra fell 20 km away to WSW											

1577	Pichincha, Ecuador (K)	0° S, 78½° W	—	—	—	—	—	—	—	—	—	—	—
		(K) quotes contemporary report of unusually red skies in Venice in 1576, the year of Titian's death											
1580	Katla (Kötlugia), Iceland (K)	63½° N, 19° W	—	—	—	—	—	—	—	—	—	—	—
1581	Fuego, Guatemala (K)	14½° N, 91° W	—	—	—	—	—	—	—	—	—	—	—
1583	near Reykjanes, Iceland (K)	64° N, 23° W	—	—	—	—	—	—	—	—	—	—	—
1586	**Kelud (Keloet), Java** (Sh)	8° S, 112½° E	b_1	—	—	**1000**$_{av}$	500	500	—	—	—	1.0	**1000**
1587	Thingvalla Hraun, Iceland (K)	64° N, 21° W	—	—	—	—	—	—	—	—	—	—	—
1587	Pichincha, Ecuador (K)	0° S, 78½° W	—	—	—	—	—	—	—	—	—	—	—
1590	Colima, Mexico	19½° N, 104° W	—	(0.5)	—	50$_3$	33	17	—	—	(24)	1.0	50
		Tephra fell 40–50 km away to NE											
1593	**Ringgit (Roung), Java** (S, Sh)	8° S, 114° E	b_1	—	—	**1000**$_{av}$	500	500	—	—	—	1.0	**1000**
		There are known references in various places to the Sun and Moon going pale over Europe in the 1590s (including e.g. Shakespeare, *Midsummer Night's Dream*, act 2, scene 1)											
1595	Tinakula, Santa Cruz Is., Melanesia	9½° S, 166° E	—	—	—	—	—	—	—	—	—	—	—
		Explosive eruption with ash cloud seen (but no details known)											
1597	Hekla, Iceland (K, S, Sh)	64° N, 19½° W	b_2	—	—	100$_{av}$	100	—	—	—	—	0.3	300
		The Hekla 1597 eruption was violent. The other two Iceland eruptions near or from under the Vatnajökull ice cap											
1598	Öraefajökull, Iceland (S, Sh)	64° N, 17° W	b_{2-3}	—	—	20$_{av}$	20	—	—	—	—	0.3	70
1598	Grimsvötn, Iceland (S, Sh)	64½° N, 17½° W	b_{2-3}?	—	—	20$_{av}$	20	—	—	—	—	0.3	70
		These three Iceland eruptions in 1597–8 all classed as great in (K)											
1601	—	The Sun greatly dimmed by a constant haze over southern Scandinavia in 1601, possibly the culminating effect there of the eruptions of the 1590s but likelier an unknown eruption, e.g. in Kamchatka, Alaska or elsewhere. The Sun and Moon appeared 'reddish, faint and lacked brilliance' in central Europe all through the year 1601 and up to the end of July 1602.											
		Total veil 1601 (minimum estimate, by comparison with 1783 and 1883)			(*$\not<$ 1000*)	—	(*$\not<$ 1000*)	—	—	—	—	—	(*$\not<$ 1000*)
1603–10	Etna	38° N, 15° E	—	—	—	**(300)**	(300)	—	—	—	—	0.5	**500–600**
		Long continued activity of Etna 1603–10 with climax about 1606–7 d.v.i. probably applies almost wholly to 1607–9											
1608	Ternate, Molucca Is. (K)	1° N, 127° E	—	—	—	—	—	—	—	—	—	—	—
1614	**Little Sunda Islands** (S, Sh) believed to be in the Sunda Sea just north of Sumbawa, but possibly Tambora	8° S, 118° E	b_1	—	—	**1000**$_{av}$	500	500	—	—	—	1.0	**1000**
		The Sun observed as 'shining quite red and bloody' at Linz, Austria in August 1614											
1615	Gunung Api, Banda, Molucca Is. (K)	4½° S, 130° E	—	—	—	—	—	—	—	—	—	—	—

APPENDIX I (*cont.*)

date	volcano	position	Sapper's assessment of magnitude: *a*, lava *b*, tephra	estimated or assumed total dust for d.v.i. km³	dust veil index: total veil in groups of eruption years	dust veil index from individual eruptions: World	Northern Hemisphere	Southern Hemisphere	R_{Dmax}	T_{Dmax} /°C	t_{mo}	$E_{\max}$	d.v.i./$E_{\max}$
1619	Hekla, Iceland (K)	64° N, 19½° W	—	—	—	—	—	—	—	—	—	—	—
2–14. ix. 1625	**Katla, Iceland** (K)	63½° N, 19° W	—	—	—	**(250)**	(250)	—	—	—	—	0.3	**(800)**
		Ash fell in quantity up to 1000 km away in Bergen, Norway, also on ships and Faeroe Islands. Magnitude judged similar to Askja 1875											
1627	Caraballos, Luzon, Philippine Is. (K)	?	—	—	—	—	—	—	—	—	—	—	—
16. xii. 1631	**Vesuvius** (K, S, Sh)	41° N, 14° E	a_2b_2	—	—	**300_{adj}**	300	—	—	—	—	0.5	**600**
		Great explosion. Tephra and lava a 20 Mg boulder was thrown 5 km, another of similar size 13 km. Dust fell at Taranto 240 km away, and in some places collected to a depth of 4 m. The top 75 m of the mountain blown away. Dust cloud observed rising to 50 km											
1635	Ternate (K)	1° N, 127½° E	—	—	—	—	—	—	—	—	—	—	—
18. v. 1636	Hekla (H, K, S, Sh)	64° N, 19½° W	b_2	—	—	100_{av}	100	—	—	—	—	0.3	**300**
		Dust cloud carried away to SE. Sky glows in Scandinavia in 1636. Seamen arriving (in Britain?) reported that for weeks the sky had seemed on fire after sunset (K). The eruption continued till the following winter (H). Classed in (K) as a great eruption											
1637–8	Timor (K)	9–10° S, 124–7° E	—	—	—	—	—	—	—	—	—	—	—
1638	**Roung (Raoen), Java** (S, Sh)	8° S, 114° E	b_{1-2}	—	—	**500_{av}**	250	250	—	—	—	1.0	**500**
1640	**Komagatake, Japan** (S, Sh)	39° N, 141° E	b_{1-2}	—	—	**250_{av}**	250	—	—	—	—	0.5	**500**
1641	**Awu (Awoe), Great Sangihe Island, Celebes** (S, Sh)	3½° N, 125½° E	b_1	—	—	**1000_{av}**	500	500	—	—	—	1.0	**1000**
1641	Taal, Luzon, Philippine Is. (K)	14° N, 121° E	—	—	—	—	—	—	—	—	—	—	—
1641	**Gunung Adiksa, Java** (S, Sh)	8° S, 112° E	b_{1-2}?	—	—	**500_{av}**	250	250	—	—	—	1.0	**500**
1646	**Makjan (Makyan), Molucca Is.** (K, S, Sh)	½° N, 127½° E	b_2	—	—	**300_{av}**	150	150	—	—	—	1.0	**300**
1646	Palma, Canary Is. (K)	28° N, 18° W	—	—	—	—	—	—	—	—	—	—	—
1650	**Santorin (Thira), Aegean** (S, Sh)	36½° N, 25½° E	b_{1-2}?	—	—	**250_{av}**	250	—	—	—	—	0.5	**500**
1651	Pacaya, Guatemala (K)	14° N, 91° W	—	—	—	—	—	—	—	—	—	—	—

1653	Ternate (K)	1° N, 127½° E	—	—	—	—	—	—	—	—	—	—	—
1660	Sinchulagu (K), possibly stands for an eruption near Sincelejo, Colombia	(9° N, 75° W)	—	—	—	—	—	—	—	—	—	—	—
1660	**Pichincha, Ecuador** (S, Sh)	0° S, 78½° W	b_2	—	—	**300**$_{av}$	150	150	—	—	—	1.0	**300**
1660	**Omate**, near Arequipa, **Peru** (S, Sh)	16½° S, 71° W	b_1	—	—	**1000**$_{av}$	300	700	—	—	—	1.0	**1000**
1660	**Teon (Teun), Banda Sea** (S, Sh)	7° S, 129° E	b_2?	—	—	**300**$_{av}$	150	150	—	—	—	1.0	**300**
3–12. xi. 1660	**Katla, Iceland** (K, S, Sh)	63½° N, 19° W	b_2	—	—	**250**$_{adj}$	250	—	—	—	—	0.3	**800**
		Similar amount and distribution of dust to 1625; produced nearly as much as the Katla 1755 eruption. Dust fell in quantity on ships south of the Faeroe Is., blackening the sails completely to the colour of thick black soot. Dust probably fell both in Faeroe and Shetland Is. In January 1661 in Denmark the sky 'seemed a bath of lurid fire', causing much public alarm											
1664	Pacaya (K)	14° N, 91° W	—	—	(500)	**(500)**	(330)	(170)	—	—	—	1.0	**(500)**
		In 1665 red twilights were remarked in London during the great plague. 'At this fatal season the sunsets were unusually splendid ... Long after the sun had set the sky was stained with crimson' (K) Description similar to 1963–4 and other cases											
1667	Omate (K)	16½° S, 71° W	—	—	—	—	—	—	—	—	—	—	—
1668	Pacaya (K)	14° N, 91° W	—	—	—	—	—	—	—	—	—	—	—
1669	Etna (K)	38° N, 15° E	—	—	—	—	—	—	—	—	—	—	—
1671	Pacaya (K)	14° N, 91° W	—	—	—	—	—	—	—	—	—	—	—
1673	Ternate (K)	1° N, 127½° E	—	—	—	—	—	—	—	—	—	—	—
1673	**Gamma Kunorra, Halmahera, Moluccas** (S, Sh)	1½° N, 127½° E	b_1?	—	—	**1000**$_{av}$	500	500	—	—	—	1.0	**1000**
1674	Amboina, Moluccas (K)	4° S, 128° E	—	—	—	—	—	—	—	—	—	—	—
1677	Palma, Canary Is. (K)	28° N, 18° W	—	—	—	—	—	—	—	—	—	—	—
1680	**Tongkoko, Celebes** (H, K, S, Sh)	1½° N, 125° E Violent	b_1	—	—	**1000**$_{av}$	500	500	—	—	—	1.0	**1000**
April–May 1680	**Krakatau, Sunda Strait** (K, S, Sh)	6° S, 105½° E	b_{2-3}	—	—	**400**$_{adj}$	200	200	—	—	—	1.0	**400**
		A long-continued moderate ash eruption. Vast quantities of pumice ejected, covering the island of Krakatau and the surrounding seas, and continuing with little intermission until about November 1681. The ash destroyed the island's rich forest growth. Blood red light filling the sky before sunrise on 22. v. 1680 alarmed the populace in Denmark											
1682	Etna (K)	38° N, 15° E	—	—	—	—	—	—	—	—	—	—	—
1690–1691	Gunung Api, Banda, Moluccas (K)	4½° S, 130° E	—	—	—	—	—	—	—	—	—	—	—

APPENDIX I (*cont.*)

date	volcano	position	Sapper's assessment of magnitude: *a*, lava *b*, tephra	estimated or assumed total dust for d.v.i. / km^3	dust veil index: total veil in groups of eruption years	dust veil index from individual eruptions: World	Northern Hemisphere	Southern Hemisphere	R_{Dmax}	T_{Dmax}/°C	t_{mc}	E_{max}	d.v.i./E_{max}
13. ii. 1693	Hekla, Iceland (H, K, S, Sh)	64° N, 19½° W	b_{2-3}	—	—	100_{adj}	100	—	—	—	—	0.3	**300**
		Dust fell in Scotland and Norway and on ships at sea. The eruption continued until August 1693. Classed in (H) and (K) as a great eruption											
1693	**Serua (Seroea), Moluccas** (H, K, S, Sh)	6½° S, 130° E	*a*? b_{1-2}	—	—	$\mathbf{500}_{av}$	250	250	—	—	—	1.0	**500**
1694	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1694	**Amboina, Moluccas** (H, K, Sh)	4° S, 128° E Violent	—	—	—	($\nless$ **250**)	($\nless$ 125)	($\nless$ 125)	—	—	—	1.0	($\nless$ **250**)
1694	**Celebes** (H, K, Sh)	1° N–6° S, 119°–125° E Violent	—	—	—	($\nless$ **250**)	($\nless$ 125)	($\nless$ 125)	—	—	—	1.0	($\nless$ **250**)
20. xi. 1694	**Gunung Api, Moluccas** (H, K, S, Sh)	4½° S, 130° E	b_2	—	—	$\mathbf{400}_{adj}$	200	200	—	—	—	1.0	**400**
		Violent. The eruption continued till April 1695 (Sapper gives it as 1696) d.v.i. adjusted for prolonged eruption											
1696	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	(*1.0*)	*60*	—	—
8. v. 1698	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
20. i. 1698 and 19. vii. 1698	Carguairazo (Carhuairazo), Chile (K)	? Possibly Corcovado 43° S, 72½° W	—	—	—	—	—	—	—	—	—	—	—
		If the low temperatures in England and a wide region surrounding over the years 1694–8 were taken as representative of a world-wide anomaly, and their departure from the immediately preceding and following years were attributed entirely to volcanic dust, d.v.i. for the *total veil* should be **3000** to **3500**											
1700	Savaii (Sawaii), Samoa (S, Sh)	14° S, 172° W	a_2	—	—	(< 100)	(< 50)	(< 50)	—	—	—	1.0	(< 100)
		(S) gives date as approximately 1700											
1706	Palma, Canary Is. (K)	28° N, 18° W	—	—	—	—	—	—	—	—	—	—	—
20. v. 1707	Vesuvius (H, K, Sh)	41° N, 14° E	—	—	—	(150)	(150)	—	—	—	—	0.5	**(300)**
		Violent (K). Eruption continued till August (H, K) Classed in (H) and (K) as a great eruption											
25. v. 1707 and especially **25. vii. 1707**	**Santorin** (H, K, Sh)	36½° N, 25½° E Violent	—	—	—	**(250)**	(250)	—	—	—	—	0.5	**(500)**
		Marked higher d.v.i. than Vesuvius 1707 because of evidence of at least two explosive phases											

1707	**Fuji (Fujiyama), Japan** (H, K, S, Sh)	35° N, 139° E Violent	b_{1-2}	—	—	**350$_{av}$**	350	—	—	—	—	0.7	**500**
1712	**Miyakeyama,** on Island Miyake Shima, off Japan (S, Sh)	34° N, 139½° E	b_2	—	—	200$_{av}$	200	—	—	—	—	0.7	**300**
1712	Siao (Siaoe), near Celebes (K)	2½° N, 125½° E	—	—	—	—	—	—	—	—	—	—	—
1716	Taal, Luzon (K)	14° N, 121° E	—	—	—	—	—	—	—	—	—	—	—
1716	Hofs Jökull, Iceland (K)	65° N, 19° W	—	—	—	—	—	—	—	—	—	—	—
1717	Eyjafjallajökull, Iceland (K)	63½° N, 19½° W	—	—	—	—	—	—	—	—	—	—	—
iv.–vi. 1717	Vesuvius (K)	41° N, 14° E	—	—	—	(100)	(100)	—	—	—	—	0.5	(200)
		Intense darkness 160 km from volcano (K) This Vesuvius eruption is not recorded by Lister (1960); evidently the dust cloud was considerable, but there may have been no serious lava flow or destruction near the volcano d.v.i. estimate takes account of the reported density of the dust cloud and duration of the eruption											
1717	**Kirishima Yama, Japan** (S, Sh)	32° N, 131° E	b_2	—	—	200$_{av}$	200	—	—	—	—	0.7	**300**
1717	Fuego, Guatemala (S, Sh)	14½° N, 91° W	b_{2-3}	—	—	70$_{av}$	35	35	—	—	—	1.0	70
11. v. 1721	**Katla, Iceland** (H, K, Sh)	63½° N, 19° W Violent (K)	—	—	—	**(250)**	(250)	—	—	—	—	0.3	**(750)**
		The Katla eruption continued till autumn 1721. For 2 months the Sun was seen red in Europe, as through a lofty haze of considerable intensity. Strange appearance of the Sun and atmospheric glows reported from France to Persia The estimate of d.v.i. could not be much lower, when the reported effects are compared with Katla 1660, though the Katla 1721 dust deposit has received less notice in the literature (cf. Thorarinsson 1958, p. 46)											
1723–6	Irazu, Costa Rica (K)	10° N, 84° W	—	—	—	—	—	—	—	—	—	—	—
17. v. 1724	**Krafla (Krabla), Iceland** (K, S, Sh)	65½° N, 17° W	a_1	—	—	(30+15 per year till 1729)	(30+15 per year till 1729)	—	—	—	80–100	0.3	**(350)** (spread over 1724–9)
		Explosive eruption of new crater on SW slope of Krafla marked the beginning of a series of fissure (mainly lava) eruptions in the Myvatn neighbourhood (the famous 'Myvatn fires') which lasted until 1730; the great lava flow (estimated at 1 km^3) from the Leirhnukur-Bjarnaflag fissure reached the lake Myvatn in August 1729, destroying the settlement there											
1727	Öraefajökull, Iceland (K, S, Sh)	64° N, 17° W	b_{2-3}	—	—	20$_{av}$	20	—	—	—	—	0.3	70
		Eruption, not so violent as in 1362, produced only (basaltic) tephra. Two farms had to be abandoned											
1728	Hekla, Iceland (K)	64° N, 19½° W	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1728	Antisana, Ecuador (K)	½° S, 78° W	—	—	—	—	—	—	—	—	—	—	—
1730	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—

APPENDIX I (*cont.*)

date	volcano	position	Sapper's assessment of magnitude: a, lava b, tephra	estimated or assumed total dust for $\frac{\text{d.v.i.}}{\text{km}^3}$	dust veil index: total veil in groups of eruption years	dust veil index, from individual eruptions: World	dust veil index, from individual eruptions: Northern Hemisphere	dust veil index, from individual eruptions: Southern Hemisphere	R_{Dmax}	$\frac{T_{D\text{max}}}{°\text{C}}$	t_{mo}	E_{max}	$\frac{\text{d.v.i.}}{E_{\text{max}}}$
1730–6	**Lanzarote, Canary Is.** (K, S, Sh)	29° N, 13½° W	$a_1\, b_{1-2}$	—	—	(**250**$_{\text{adj}}$ + 10 per year till 1733)	(250 + 10 per year till 1733)	—	—	—	—	0.7	(**400**) (spread over 1730–3)
		Estimated to have produced 1.5 km³ lava. Extensive 'dry fog' reported (in Europe?) in 1733											
1730	Roung (Raoen), Java (S, Sh)	8° S, 114° E	b_2	—	—	**300**$_{\text{av}}$	150	150	—	—	—	1.0	**300**
1732	Etna (K)	38° N, 15° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1732	Jan Mayen	71° N, 8° W Fire and dust emitted for 28 h	—	—	—	—	—	—	—	—	—	—	—
1735	Etna (K)	38° N, 15° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1737	Kamchatka (K)	Probably *ca.* 54° N, 160° E	—	—	—	—	—	—	—	—	—	—	—
1737	Avachinskaya, Kamchatka (K)	53½° N, 159° E	—	—	—	—	—	—	—	—	—	—	—
20. v. 1737	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1739	Tubachinskaya (probably Sopka Tolbachik, near Kluchev), Kamchatka	(55° N, 160½° E?)	—	—	—	—	—	—	—	—	—	—	—
1742	Sangay, Ecuador (K)	2° S, 78° W	—	—	—	—	—	—	—	—	—	—	—
		(S) describes Sangay as one of the most frequently active volcanoes in the world, yet has produced only the equivalent of about one major eruption per century in terms of solid matter issued											
v. and xi. 1744	**Cotopaxi, Ecuador** (K, S, Sh)	1° S, 78° W Violent	$a_3 b_2$?	—	—	**300**$_{\text{av}}$	150	150	—	—	—	1.0	**300**
		(S) and (Sh) indicate that the eruption began in 1742											
1748–52	Sandfellsjökull, Iceland (K)	64° N, 17° W	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1749	Taal, Luzon (S, Sh)	14° N, 121° E	b_{2-3}	—	—	70$_{\text{av}}$	35	35	—	—	—	1.0	70
19. x. 1751	Vesuvius (K)	41° N, 14° E Eruption continued till 9. xi. 1751	—	—	—	(10)	(10)	—	—	—	—	0.5	(20)
1752	**Little Sunda Is.** (S, Sh): possibly Tambora (see also 1614, 1815)	8° S, 118° E	b_1	—	—	**1000**$_{\text{av}}$	500	500	—	—	—	1.0	**1000**

1753	Skeidararjökull, Iceland (K)	64° N, 17½° W	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1754	**Taal, Luzon** (S, Sh)	14° N, 121° E	b_2	—	—	**300_{av}**	150	150	—	—	—	1.0	**300**
19. xi. 1754	Hekla (K)	64° N, 19½° W	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
2. xii. 1754	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
iii. 1755	Etna (K)	38° N, 15° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
17. x. 1755	**Katla** (H, K, S, Sh)	63½° N, 19° W	b_2	5	—	**$400_{2,3}$**	400	—	—	0.7	48	0.3	**1200**
		Violent till 1. xi. 1755, the eruption did not end till viii. 1756. Produced the thickest Katla dust layer of historical times; enormous amount of tephra in the first phase of the eruption carried eastwards and deposited 20–30 cm deep over southern-southeastern Iceland, destroying whole farming districts. Volcanic dust fell on a ship 120 km NW of Shetland on 23–24 October. Haloes (corona-Bishop's ring?) around the Sun and Moon, and 'igneous meteors' were reported all over Europe before and after the great Lisbon earthquake of 1. xi. 1755. Fiery red sunrises and sunsets reported in various parts of Europe and over the Atlantic from November 1755 to January or February 1756. At Lisbon on 30. x. 1755 the light of the Sun dim; in Cornwall, England, on 1. xi. 1755 the Sky full of fiery red clouds, in the afternoon becoming 'a very odd coppery colour in places'. Flame-coloured glow till 2½ h after sunset in western Ireland on 1. i. 1756											
		530_2, 315_3											
24–27. iv. 1756	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
26. xii. 1756	Luzon, Philippine Is. (K)	13–18° N, 120–124° E	—	—	—	?	?	—	—	—	—	—	—
		Cold year in 1758 in middle latitudes might be taken to indicate an additional dust layer subsequent to that produced by Katla 1755 estimate Luzon 1756 at **400_2** for total veil summations(?)											
i. 1758	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
18. iv. 1759	Etna (K)	38° N, 15° E	—	—	—	—	—	—	—	—	—	—	—
		Eruption repeated in May 1759											
21. ix. 1759	**Jorullo, Mexico** (K, S, Sh)	19° N, 102° W	a_2b_2	—	—	**300_{av}**	200	100	—	—	—	1.0	**300**
		Twilight observations on a dust layer high in the atmosphere over Augsburg made by Lambert on 18. ix. 1759											
xi. 1759 and 24. xii. 1760	Vesuvius (K)	41° N, 14° E	—	—	—	—	—	—	—	—	—	—	—
1760	**Makjan (Makyan), Molucca Is.** (S, Sh)	½° N, 127½° E (cf. b_2 av. 300)	b_2	—	—	**250_2**	125	125	—	0.4	12	1.0	**250**
3. xii. 1760	Peteroa, Andes (K)	35½° S, 70½° W	—	—	—	—	—	—	—	—	—	—	—
1762	Peteroa (K)	35½° S, 70½° W	—	—	—	—	—	—	—	—	—	—	—
ii. 1763	Etna (K)	38° N, 15° E	—	—	—	—	—	—	—	—	—	—	—
		No known report of a dust veil, but 1763 was a cold year in middle latitudes (fig 14*a*; Köppen 1873), with a severe harvest failure in Japan, and further cold years followed 1765–71											
1. ix. 1763	Molucca Is. (K)	2° N to 3° S, 125–131° E	—	—	—	—	—	—	—	—	—	—	—
		Total veil 1763 on basis of temperature variation only			*600_2*	?	?	—	—	*0.6*	*(18)*	*1.0*	*600*

APPENDIX I (*cont.*)

date	volcano	position	Sapper's assessment of magnitude: *a*, lava *b*, tephra	estimated or assumed total dust for d.v.i. km³	dust veil index: total veil in groups of eruption years	dust veil index from individual eruptions: World	Northern Hemisphere	Southern Hemisphere	R_{Dmax}	T_{Dmax} °C	t_{mo}	E_{max}	$\frac{\text{d.v.i.}}{E_{\text{max}}}$
iii. 1766	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1766	Etna (K)	38° N, 15° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
5. iv. 1766	Hekla (H, K, S, Sh)	64° N, 19½° W Violent	$a_4 b_2$	—	—	200_{adj}	200	—	—	—	—	0.3	**650**
		Dust in initial stages carried NW from the volcano. The tephra layer does not seem to have been thoroughly surveyed. Eruption continued with violent phases until 7. ix. 1766 and some activity until 1768 Classed in (H) and (K) as a great eruption											
1766	**Mayon, Luzon** (H, K, Sh)	13½° N, 123½° E Violent	—	—	—	(**2300**)	(1150)	(1150)	—	—	—	—	(**2300**)
viii. and x. 1767	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
		Temperature deficit over Köppen's middle latitudes observing stations culminating with values in 1767 and 1771 about 1.3° C (equivalent adjusted anomaly for whole zone 0.7 to 0.9 °C) below the level ruling in apparently undisturbed years about that time (figure 14 *a*)											
iv. 1768	**Cotopaxi, Ecuador** (K, S, Sh)	1° S, 78° W	$a_3 b_2$	—	—	$\mathbf{900}_{\text{adj}}$	450	450	—	0.4	45	1.0	**900**
		$(900\text{–}950)_2$ estimate based on apparent halt in recovery of temperatures after the 1766 eruptions											
1768	Tinakula, Santa Cruz Is., Melanesia	9½° S, 166° E Ash eruption. (No details)	—	—	—	—	—	—	—	—	—	—	—
1770	Colima, Mexico (K)	19½° N, 104° W	—	—	—	—	—	—	—	—	—	—	—
		Total veil 1766–71 on evidence of temperatures only			$\mathit{3400}_2$	—	—	—	—	*0.9*	*72*	*1.0*	*3400*
1772	**Gunung Papandajan (Papandayan), Java** (K, S, Sh)	7½° S, 108 °E	b_1	1	—	$\mathbf{250}_{2,\,3}$	125	125	—	0.2	30	1.0	**250**
		Violent and 'accompanied by the extrusion (including lava?) of much larger quantities of material than that thrown out of Krakatau in 1883' (K). (S) quotes an estimate of 'over 1 km³' of solid matter ejected: this estimate suggests that the earlier accounts may have been exaggerated as regards effect upon the atmosphere. (The figure also represents the lower limit for Sapper's first order eruptions.) Any temperature effect in middle latitudes seems to have been very slight (figure 14*a*) 315_2, 130_3											
14. ii. 1775	Vulcano, island N. of Sicily (K)	38½° N, 15° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
vii. 1775	Pacaya, Guatemala (K)	14° N, 91° W	—	—	$\mathbf{1000}_2$	?	?	—	—	—	—	—	—
		Köppen indicates cold years in 1776 and especially 1777 (figure 14 *a*)											
28. iii. 1776	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
27. iv. 1776	Etna (K)	38° N, 15° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
		Total veil till 1777 on evidence of temperatures only			*1000*	—	—	—	—	*0.8*	*24*	*1.0*	*1000*
1779	Vesuvius (K)	41° N, 14° E Violent according to (K)	—	—	—	(10)	(10)	—	—	—	—	0.5	(20)

1779	**Sakurashima, Japan** (K, S, Sh)	31½° N, 131° E	$a_1\,b_{1-2}$	—	—	**450_2**	450	—	—	0.7	18	0.7	**650**
1780	Etna (K)	38° N, 15° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1780	Vulcano (K)	38½° N, 15° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
24. iv. 1781	Etna (K)	38° N, 15° E Eruption continued all May 1781	—	—	—	(< 10)	(< 10	—	—	—	—	—	—
v. 1783	**Eldeyjar,** small islands off Reykjanes, Iceland (K, S, Sh)	63½° N, 23° W	b_{1-2}	—	—	—	—	—	—	—	—	—	—
			—	—	—	**(700)**	(700)	—	—	—	50	0.3	**2300**
		d.v.i. **700**, d.v.i./E_{max} **2300** estimated as combined effect of Eldeyjar and Laki-Skaptar Jökull eruptions in Iceland in 1783											
end of v. and vi. 1783	**Laki (and Skaptar Jökull), Iceland** (H, K, S, Sh)	64° N, 18° W	a_1b_1	(3)	—	—	—	—	—	—	—	—	—
		Violent explosion on 8 and 18. vi. 1783. The eruption continuing over the next 7 month sproduced the greatest lava flow on Earth in historical times, variously estimated at from 12 to 27 km³ 190_3: see page 443, q = 12 km³ might be a more reasonable estimate, implying 750_3 Dust from this and the Eldeyjar eruption fell over all Iceland, the Faeroe Is. and northern Scotland, where it was enough to destroy crops in Caithness. A thick dry haze spread over Europe, first reported at Copenhagen on 29 May, in France from 6 June onwards, noted in northern Italy from 18 June, reaching Syria and the Altai in central Asia by 1 July, when it stretched from N. Africa to Scandinavia. Despite reports of a sulphurous smell all over western Europe (and unpleasant action on the eyes), and damage to plants in Holland 18–24 June, the haze over Europe and Asia and N. America must have been largely in the upper atmosphere because it was always present that summer regardless of low level wind directions and it was not washed out by rain. Visibility at the surface in Europe was however often reduced to 5 km in the months June–September 1783 The dust veil must have been exceptionally dense (opaque): in southern France the sun was wholly obscured by it at elevations less than 17°. The sun was dimmed and shone 'red, or bluish-white and rayless', most of the day. No stars were seen below 40° elevation. The moon was similarly affected, seen as a dull red or blue object. Even in Italy in June the 'rayless' sun was easily looked at. In England (Norfolk) Parson Woodforde noted that in July the sun remained coppery coloured until it was 20° above the horizon. The haze was reported as lasting till end of September–early October 1783 over Europe. Coloured sunsets and the unusual glare of the sky at night, giving at new moon as much light 'as a full moon even at midnight' also attracted notice											
2–5. viii. 1783	**Asama Yama (Asama), Japan** (H, K, S, Sh)	36½° N, 138½° E	a_2b_2	—	—	**300_{av}**	300	—	—	—	50	0.5	**600**
		According to an account quoted in (K) 'the most frightful eruption on record'. Rocks 10 to 25 m thick were hurled in all directions by the violence of the explosions. Towns and villages buried. Lava stream 60 km long Köppen estimated that the temperature over much of the world was reduced almost 2 °C for 3 years after the 1783 eruptions; figure 14 *a* indicates that there may have been some effect for 5–6 years											
1784–85	Hekla (K)	64° N, 19½° W	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1785	Cerro Quemado (volcano of Quezaltenango), Guatemala	15° N, 91½° W 'Great eruption' (*Encycl. Brit.*)	—	—	—	—	—	—	—	—	—	—	—
1785	Vesuvius (K, Sh)	41° N, 14° E Violent (K)	—	—	—	—	—	—	—	—	—	—	—
		Total veil 1783–6 on evidence of temperature curve. E_{max} taken as 0.4, the dust originating partly near 64° N, partly near 36° N, gives d.v.i. 1000_2 for total veil			*1000_2*	—	—	—	—	1.3	*50*	*0.4*	*2500*
1786	Pavlof (Pavlovskaya or Paulow) Alaska (S, Sh)	55½° N, 162° W	$b_{2(-1)}$	—	—	150_{av}	150	—	—	—	—	0.3	**500**
1786	Amukta, Aleutian Is. (S, Sh)	52½° N, 171° W	b_{2-3}	—	—	20_{av}	20	—	—	—	—	0.3	70

Appendix I (*cont.*)

date	volcano	position	Sapper's assessment of magnitude: *a*, lava *b*, tephra	estimated or assumed total dust for d.v.i. (km³)	dust veil index: total veil in groups of eruption years	dust veil index from individual eruptions: World	Northern Hemisphere	Southern Hemisphere	R_{Dmax}	T_{Dmax} (°C)	t_{mo}	E_{max}	d.v.i./E_{max}
1787	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1787	Etna (K)	38° N, 15° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1789	Kilauea, Hawaii (Sandwich Islands) (K)	19½° N, 155½° W	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1790	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
11. v. 1792	Etna (K)	38° N, 15° E	—	—	—	(10)	(10)	—	—	—	—	0.5	(20)
		Continued erupting till 1793											
1793	Alaid, Kurile Is. (K)	51° N, 155½° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1. iv. 1793	Illigiyama (probably Unzentake)	32½° N, 130½° E	—	—	—	—	—	—	—	—	—	—	—
		Positions rather uncertain. Lava eruption of Unzentake reported											
	and Kyushu Mts, Japan (K)	32–33° N, 131° E	—	—	—	—	—	—	—	—	—	—	—
1793	Tuxtla, Mexico (S, Sh)	18° N, 95° W	b_{2-3}	—	—	70_{av}	47	23	—	—	—	1.0	70
1794	Vesuvius (K, S, Sh)	41° N, 14° E	$a_3\, b_{2-3}$	—	—	35_{av}	35	—	—	—	—	0.5	70
		Violent. Town of Torre del Greco destroyed for fourth time											
1795	Colima, Mexico (K)	19½° N, 104° W	—	—	—	—	—	—	—	—	—	—	—
1795	Pogrumni (Pogrumnoy), Umanak Is., Aleutians (S, Sh)	55° N, 165° W	b_1	—	—	**300_{adj}**	300	—	—	—	—	0.3	**1000**
1795	Kluchev (Klyuchevskaya Sopka, Klyuchev), Kamchatka (K)	56° N, 160½° E	—	—	—	—	—	—	—	—	—	—	—
1796	Bogoslof (Bogoslov), Aleutians (S, Sh)	54° N, 168° W	$a_2 b_2$	—	—	100_{av}	100	—	—	—	—	0.3	**300**
		Total veil 1794–6			*500_2*	—	—	—	—	*0.7*	*30*	*0.5*	*1000*
1798	Chahorra, Teneriffe (K)	28° N, 17° W	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1798	Izalco, Salvador (K)	14° N, 89½° W	—	—	—	?	?	—	—	—	—	—	**(300)**
		(S) considers this frequently erupting volcano produces in all only the equivalent of one major eruption in a century											
1799	Fuego, Guatemala (H, K, S, Sh)	14½° N, 91° W	$b_{3(-2?)}$	—	—	**600_{adj}**	300	300	—	—	—	—	**600**
		Reddish brown 'vapour' observed in the sky by Humboldt at Cumana (10½° N, 64° W), Venezuela from 10. x. 1799 to 3. xi. 1799. Sky the colour of fire each evening											
		(H) suggests Fuego 1799 might be classed as a great eruption											
		Total veil 1798–9			*900_2*	—	—	—	—	*1.1*	*15*	*1.0*	*900*

1803	Cotopaxi, Ecuador (K)	1° S, 78° W	—	—	—	—	—	—	—	—	—	—	—
11. viii. 1805	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
		Total veil 1803–5 on evidence of temperature alone			*1100*$_2$	—	—	—	—	*0.7*	*30*	*1.0*	*1100*
1805–7	Izalco, Salvador (K)	14° N, 89½° W	—	—	—	—	—	—	—	—	—	—	—
1807	Gunung Merapi (Merapi), Java (K)	7½° S, 110½° E	—	—	—	—	—	—	—	—	—	—	—
1. v. 1808	São Jorge (St George), Azores (H, K, Sh)	38½° N, 28½° W	—	—	—	(20)	(20)	—	—	—	—	—	—
		Abnormal brilliance of twilight noticed on 6. v. 1808 at Plaistow, London by Luke Howard (H) suggests that these São Jorge and Etna eruptions might be classed as great											
26. iii. 1809	Etna (H, K, Sh)	38° N, 15° E	—	—	—	(20)	(20)	—	—	—	—	—	—
		Rose-coloured haze over the sky at twilight noted at Plaistow, London on 1 and 2. iv. 1809 by Luke Howard											
		Total veil 1807–10 on evidence of temperature alone			*1500*$_2$	—	—	—	—	*0.7*	*42*	*1.0*	*1500*
31. i. 1811	Sabrina, Azores (K, S, Sh)	38° N, 25° W	b_2	—	—	200$_{adj}$	200	—	—	—	—	0.5	**400**
		Violent (K). A temporary islet of cinders thrown up off the coast of São Miguel											
		Unusually highly (red) coloured skies noted over London by Howard on 3 and 8. ii. 1811 and again from mid-June in every month to December 1811											
11–16. vi. 1811	Azores (K)	*ca.* 38° N, 25° W	—	—	—	—	—	—	—	—	—	—	—
12. x. 1811	Vesuvius (K)	41° N, 14° E	—	—	—	—	—	—	—	—	—	—	—
		Red twilights were particularly noticed by Howard on 12, 15 and 19. x. 1811											
25 and 27. x. 1811	Etna (K)	38° N, 15° E	—	—	—	—	—	—	—	—	—	—	—
		'Cinders carried a great distance'											
30. iv. 1812	Soufrière, St Vincent, West Indies (H, K, S, Sh)	13½° N, 61° W	b_2	—	—	**300**$_{av}$	150	150	—	—	—	1.0	**300**
		Violent. Classed in (H) and (K) as a great eruption. Ash fell at Barbados about 170 km to the east											
		Luke Howard noted unusually luminous coloured twilight on 8. v. 1812 and similar observations in every month to the end of October in London											
12. vi. 1812	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1812	Awu (Awoe), Great Sangihe Is., Celebes (S, Sh)	3½° N, 125½° E	b_2	—	—	**300**$_{av}$	150	150	—	—	—	1.0	**300**
1813	Vesuvius (K)	41° N, 14° E	—	—	—	(100)$_2$	(100)	—	—	—	—	0.5	(200)
		Continued erupting from May till December 1813, with violent discharge of ash on 25. xii. 1813											
		Luke Howard noted unusual orange-red twilights over London (Tottenham) from 31. v. 1813 to the end of the year. Paleness of the sun reported at Forli, northern Italy in September											
1814	Kamchatka (K)	50–60° N, 156–163° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
		Submarine eruption											
1. ii. 1814	**Mayon, Luzon** (H, K, S, Sh)	13½° N, 123½° E	a_2b_2	—	—	**300**$_{av}$	150	150	—	—	—	1.0	**300**
		Unusually brilliant twilights, with 'lemon surmounted by purple' on 10. xi. 1814, and much orange and rose-red coloration reported over London by Luke Howard from 5. xi. 1814 to 1. i. 1815											

Appendix I (*cont.*)

date	volcano	position	Sapper's assessment of magnitude: *a*, lava *b*, tephra	estimated or assumed total dust for d.v.i. km³	dust veil index: total veil in groups of eruption years	dust veil index from individual eruptions: World	Northern Hemisphere	Southern Hemisphere	R_{Dmax}	T_{Dmax} °C	t_{mo}	E_{max}	d.v.i./E_{max}
7–12. iv. 1815	**Gunung Tambora (Tambora, Tomboro), Sumbawa** (H, K, S, Sh) (sometimes referred to as Tambora in the Little Sunda Islands)	8° S, 118° E	b_1	(150)	—	$(\mathbf{3000})_2$	1500	1500	—	1.0	48	1.0	**3000**
		For 3 days there was darkness at 500 km from the volcano. 'Extrusion of much larger quantities of material than that thrown out of Krakatau in 1883' (K). Sapper quotes various workers' estimates of the solid ejecta ranging from 100 to 300 km³, doubtless mainly as solid blocks of great size and only a small proportion as dust											
		Many reports of remarkable sunsets in London and luminous twilight from 15. v. 1815 to the end of the year. The eruption continued in some degree until 1819 (S)											
		Estimate of Tambora veil by subtraction of figures for other eruptions from d.v.i. for total veil 1812–18. 3200_3. q = 15 km³ would give d.v.i. equivalent to that here deduced from other approaches, possibly implying that of the total solid matter blown away only about 15 km³ was carried as dust											
1817	Roung (Raoen), Java (S, Sh)	8° S, 114° E	b_2	—	—	$\mathbf{300}_{av}$	150	150	—	—	—	1.0	**300**
1818	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
		Richly coloured twilights noted by Luke Howard and other observers in London from 29. vi. 1818 to January 1819											
1818	Colima, Mexico (K)	19½° N, 104° W	—	—	—	—	—	—	—	—	—	—	—
11–21. x. 1818	Gunung (Goenoing), near Batavia (K)	*ca.* 7° S, 107–110° E	—	—	—	—	—	—	—	—	—	—	—
		Violent eruption (*Phil. Trans. Roy. Soc. Lond.* 1819)											
31. x. 1818	Hekla (K)	64° N, 19½° W	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1818	Beerenberg, Jan Mayen	71° N, 8° W	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
		Eruption from small crater											
		Total veil 1811–18			4400_2	—	—	—	—	*1.0*	*84*	*1.0*	*4400*
iv. to vi. 1819	Etna (K)	38° N, 15° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
20. vi. 1819	Denodur, Bhooj (Bhuj), western India (K)	*ca.* 23° N, 70° E	—	—	—	—	—	—	—	—	—	—	—
16. i. 1820	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1820	Amboina, Moluccas (K)	4° S, 128° E	—	—	—	—	—	—	—	—	—	—	—
11–12. vi. 1820	Gunung Api, Banda, Molucca Is. (K)	4½° S, 130° E	—	—	—	—	—	—	—	—	—	—	—
27. ii. 1821	Bourbon (K) believed to be St Paul Island	38° S, 78° E	—	—	—	—	—	—	—	—	—	—	—
		Eruption continued at least until 11. iv. 1821, when still much smoke											

iii.1821	Kluchev (Klyuchevskaya Sopka), Kamchatka (K)	56° N, 160½° E	—	—	—	—	—	—	—	—	—	—	—
		Several reports of exceptional sunset colours observed in London and Paris in July and August 1821. On 18. viii. 1821 the sun was seen blue or the colour of quicksilver in various parts of England and France, where it was also described as 'enfeebled by dense vapours and absolutely white'											
1821	Irazu, Costa Rica (K)	10° N, 84° W	—	—	—	—	—	—	—	—	—	—	—
		Total veil 1820–1 on evidence of temperature alone			*500$_2$*	—	*(250)*	*(250)*	—	*0.4*	*24*	*1.0*	*500*
19. xii. 1821	**Eyjafjallajökull, Iceland** (K, S, Sh)	63½° N, 19½° W	b_2	—	—	100$_{av}$	100	—	—	—	—	—	**300**
		Eruption continued 'very violent till 28. ii. 1822' (K)											
xii. 1821	Bridgeman Island, South Shetlands	62° S, 56½° W	—	—	—	—	—	—	—	—	—	—	—
		Smoke emerging from a small crater, when the island was first visited: unlikely that the most vigorous phase of eruptive activity was encountered (see also 1822, 1839 and 1880)											
From ii. 1822	Vesuvius (K, S, Sh)	41° N, 14° E	b_2	—	—	150$_{av}$	150	—	—	—	—	0.5	**300**
		Eruptions on 31. ii., 10. vii. and 22. x. to 4. xi. 1822, the last period being one of violence on 22 and 23. x. 1822: smoke and stones thrown to 'an enormous elevation, about 20000 ft' (6 km), reducing the sun to a blue spot and darkening the daylight											
22. vii. 1822	Gunung Ber Api, Sumatra (K) (possibly Marapi, ½° S, 100½° E)	?	—	—	—	?	?	—	—	—	—	—	—
8–12. x. 1822	**Galunggung (Galoeng-goeng), Java** (K, S, Sh)	7° S, 108° E	b_1	—	—	(**500**$_{adj}$)	(250)	(250)	—	—	—	1.0	(**500**)
		Estimated ejection of 1.5 km^3 of tephra 'Sunset skies near London very bright and the clouds frequently red' from March to September 1823. Unusual sky colours also in November 1822 in Britain (*Royal Society* 1888, p. 395) Galunggung 1822 300$_3$ (A d.v.i. value of 300 is low for a Sapper b_1 eruption and also in relation to the reported twilight effects near London: an adjusted value of 500 represents the lowest that seems likely. The temperature dip evidence is not strong because of unlike behaviour of the different curves in figure 14 at this date)											
29. xii. 1822	Gunung Merapi (Merapi), Java (K)	7½° S, 110½° E	—	—	—	—	—	—	—	—	—	—	—
1822–4	Bridgeman Island, South Shetlands	62° S, 56½° W	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
		Smoke issuing with great force through fissures in the rock (report by Weddell)											
26. vii. 1823	Katla, Iceland (K)	63½° N, 19° W	—	—	—	—	—	—	—	—	—	—	—
1823	South Sandwich Islands	56–59° S, 26–28° W	—	—	—	—	—	—	—	—	—	—	—
		'nine burning volcanoes...' (report by B. Morrell)											
31. vii. 1824	Lanzarote, Canary Is. (K)	29° N, 13½° W	—	—	—	?	?	—	—	—	—	—	—
		Eruption continued till October or November. 'Peculiar mist around the sun' as seen in central Italy in August 1824											
1824	Near Reykjanes, Iceland (K)	64° N, 23° W	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1824	Amboina, Molucca Is. (K)	4° S, 128° E	—	—	—	?	?	—	—	—	—	—	—
		Total veil 1822–4 (using Köppen's curve in figure 14*a* which is less erratic than 14*b* at this point)			*2200$_2$*	—	—	—	—	*(1.0)*	*(42)*	*1.0*	*2200*
1825–31	Isanotski (Isanotskaya) Unimak Island (S, Sh)	55° N, 164° W	$a_? b_2$	—	—	20$_{av}$ per year 1825–31	20 per year 1825–31	—	—	—	—	0.3	70 per year 1825–31
1826	Tolima, Colombia (K)	4½° N, 75½° W	—	—	—	—	—	—	—	—	—	—	—

APPENDIX I (*cont.*)

date	volcano	position	Sapper's assessment of magnitude: *a*, lava *b*, tephra	estimated or assumed total dust for d.v.i. $\frac{\text{d.v.i.}}{\text{km}^3}$	dust veil index: total veil in groups of eruption years	dust veil index from individual eruptions: World	Northern Hemisphere	Southern Hemisphere	R_{Dmax}	$\frac{T_{Dmax}}{°C}$	t_{mo}	E_{max}	$\frac{\text{d.v.i.}}{E_{max}}$
1826	**Kelud (Keloet), Java** (S, Sh)	8° S, 112½° E	b_2	—	—	**300**$_{av}$	150	150	—	—	—	1.0	**300**
		Kelud 1826 possibly greater than this?											
1827	Perace (Puracé), Colombia (K)	2½° N, 76½° W	—	—	—	—	—	—	—	—	—	—	—
1828	Atitlan, Guatemala (K)	14½° N, 91° W	—	—	—	—	—	—	—	—	—	—	—
1829	Kluchev (Klyuchevskaya Sopka, Kljutschew), Kamchatka (S, Sh)	56° N, 160½° E	a_1b_2	—	—	100$_{av}$	100	—	—	—	—	0.3	**300**
		Estimated emission of 3.7 km³ of lava (S)											
1829	Deception Island	63° S, 60½° W	—	—	—	—	—	—	—	—	—	—	—
		Steam issuing from crater: unlikely that the visitors witnessed the maximum activity											
16. iv. 1830	Etna (K)	38° N, 15° E	—	—	—	(10)	(10)	—	—	—	—	0.5	(20)
		Violent (K)											
ca. 1830	Visokoi Is., South Sandwich Islands	56½° S, 27½° W	—	—	—	—	—	—	—	—	—	—	—
		'A burning mountain with smoke issuing in various places' (report by J. Brown, who is unlikely to have encountered the most vigorous phase of the activity on his visit in 1830)											
		Total veil 1826–30: this estimate lacks confidence because of the absence of direct evidence of a dust veil, and because the main feature of the temperature curves which yields the high d.v.i. figure is a very sharp minimum in just one year 1829. Nevertheless the Java eruption (Kelud 1826) may have been underestimated by Sapper											
					(2500)$_2$	—	—	—	—	*(1.0)*	*(48)*	*1.0*	*(2500)*
17. ii. 1831	Etna (K)	38° N, 15° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
Summer 1831	Vesuvius (K)	41° N, 14° E	—	—	—	(10)	(10)	—	—	—	—	0.5	(20)
		Explosion cleared the crater of a mass of *scoriae* (cinders) (K)											
10. vii. 1831	**Giulia, Graham's Island**—also known as Siarra or Sciacca, Nerita or Ferdinandea) (H, K, S, Sh)	37° N, 12–13° E	b_2	—	—	200$_{adj}$	200	—	—	—	—	0.5	**400**
		Explosive submarine eruption produced a temporary island of cinders which existed for 3 months in the Mediterranean between Sicily, Malta and Pantellaria. Eruptive activity continued till early August 1831. Smoke and dust were observed rising in a column to 20 km height											
1831	Pichincha, Ecuador (H, K, Sh)	0° S, 78½° W	—	—	—	?	?	—	—	—	—	—	—
		In full activity (K) Pichincha 1831 classed in (H) and (K) as a great eruption											
vii. 1831	Barbados (K)	13° N, 60° W	—	—	—	?	?	—	—	—	—	—	—
		'Followed by great obscurity and blue sun at Bermuda on 11. viii. 1831'											
vii–viii. 1831	**Babuyan (Babujan), Philippine Is.** (H, K, S, Sh)	19° N, 122° E	b_2	—	—	**300**$_{av}$	200	100	—	—	—	1.0	**300**

		Remarkably red and yellow skies, and lurid sunsets, first noticed in England on 1. vi. 1831, spread eventually to Sweden and N. Africa (Algiers) and across Asia and N. America. The dust veil was observed as an 'extraordinary dry fog' over all Europe, N. Africa, Siberia and in the U.S.A. in August. The sun in N. Africa became visible only when 15–20° or more above the horizon, its light so much diminished in many countries that it could be looked at all day with the naked eye. In the West Indies the sky appeared overcast in August, 'but of a decided bluish colour'. In Germany and in New York also it was reported that the clear sky had not its normal blue. Blue or green sun and moon observed in many countries. The twilight was so luminous as to lengthen the daylight: in Berlin small print could be read out of doors when the sun was 19° below the horizon. (*Royal Society* 1888, pp. 396–398.) (Summer weather in Japan produced disastrous harvest failures in 1832 and 1833) The temperature evidence suggests (figures 14*a*, *b*) that the dust covered the higher latitudes only, presumably therefore that it came from the Mediterranean and not from Babuyan. But the descriptive evidence indicates a strong effect over North America and North Africa as well as the Far East. (Köppen's tropical zone temperatures, still based on only one or two isolated places, may perhaps be disregarded.) The Babuyan eruption is generally accepted as a great one. The situation appears similar to 1783											
1833	Atitlan, Guatemala (K)	14½° N, 91° W	—	—	—	(< 10)	(< 10)	(< 10)	—	—	—	—	—
		Total veil 1831–3. E_{max} taken as 0.7 since some of the dust originated at 19° N and the rest at 37° N			1000_2	—	—	—	—	*1.0*	*(30)*	*0.7*	*1400*
viii. 1834	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1834	Merapi, Java (K)	7½° S, 110½° E	—	—	—	—	—	—	—	—	—	—	—
20. i. 1835	**Coseguina, Nicaragua** (H, K, S, Sh)	13° N, 87½° W	b_1	(50)	—	**(4000)**$_{adj}$	2000	2000	—	—	—	1.0	**(4000)**
		Great explosion blew the whole top of the mountain away. Estimates of total solid ejecta 50–150 km³, doubtless mainly as solid blocks 17500_3: q = 12 km³ would give d.v.i. equivalent to other methods											
20. i. 1835	Aconcagua, Chile (K)	33° S, 70° W	—	—	—	—	—	—	—	—	—	—	—
20. i. 1835	Osorno, Chile (K)	41° S, 72½° W	—	—	—	?	—	?	—	—	—	—	—
		Erupted again on 20. ii. and 11. xi. 1835											
20. ii. 1835	'all the volcanoes of Chile' (K)	—	—	—	—	?	—	?	—	—	—	—	—
		Osorno (41° S, 72½° W), Minchinmavida (43° S, 72½° W), and Corcovado (43° S, 73° W) all erupting											
1835	Gunung Api, Banda, Molucca Is. (K)	4½° S, 130° E	—	—	—	—	—	—	—	—	—	—	—
1836	Guadaloupe, West Indies (K)	16° N, 61½° W	—	—	—	—	—	—	—	—	—	—	—
1837	Avachinskaya Sopka (Avachinsky, Awatska), Kamchatka (H, K, Sh)	53° N, 159° E	—	—	—	?	—	?	—	—	—	—	—
		Remarkable afterglows noted in various places (in central Europe?) in 1837, especially in October, when they lasted 1 h 36 min (K) Avachinsky 1837 classed in (H) and (K) as a great eruption											
21. viii. 1837	Southern hemisphere eruption (possibly New Zealand or Antarctic)? (K)	—	—	—	—	?	—	?	—	—	—	—	—
		'Red sky at Lasaya (?), Tasmania and terrible explosions' (K)											
1838–9	Bridgeman Island, South Shetlands	62° S, 56½° W	—	—	—	—	—	—	—	—	—	—	—
		Signs of activity noted when the island was visited: strong sulphurous smell. Unlikely that the activity continued always at a constant low level (see also 1821, 1822 and 1880)											
vi. 1840	Kilauea, Hawaii (K, S, Sh)	19½° N, 155½° W	a_2	—	—	(< 10)	(< 10)	—	—	—	—	—	—
		Total veil 1835–41			4200_2	—	—	—	—	*1.0*	*(80)*	*1.0*	*4200*

Appendix I (*cont.*)

date	volcano	position	Sapper's assessment of magnitude: *a*, lava *b*, tephra	estimated or assumed total dust for d.v.i. / km³	dust veil index: total veil in groups of eruption years	dust veil index from individual eruptions: World	Northern Hemisphere	Southern Hemisphere	R_{Dmax}	T_{Dmax}/°C	t_{mo}	E_{max}	d.v.i./E_{max}
ca. 1841	Erebus, Ross Is., Antarctic (S)	77½° S, 167° E	—	—	—	?	?	—	—	—	—	—	—
		When first seen by James Clark Ross's expedition on 22. i. 1841, Erebus was emitting dense smoke and spurts of flame to about 600 m above the summit. Though such great activity has never been seen again, it is unlikely that the expedition happened to witness the most violent phase											
ca. 1842	Deception Island	63° S, 60½° W	—	—	—	—	—	—	—	—	—	—	—
		The whole southern side of Deception Island on fire, thirteen craters erupting: unlikely that the phase of maximum activity was witnessed by the visitors in 1842. Boiling springs, steam and much noise had been reported on a visit in 1838											
1843	Mount Rainier, Rocky Mts, U.S.A. (K)	47° N, 122° W	—	—	—	—	—	—	—	—	—	—	—
		Erupted also in 1841											
1845	Merapi, Java (K)	7½° S, 110° E	—	—	—	—	—	—	—	—	—	—	—
1845	Antuco, Chile (K)	37½° S, 71½° W	—	—	—	—	—	—	—	—	—	—	—
2. ix. 1845 to iv. 1846	**Hekla** (K, S, Sh)	64° N, 19½° W	a_2b_2	—	—	$\mathbf{250}_{\mathrm{adj}}$	250	—	—	—	—	0.3	**800**
		On 2. ix. 1845 dust-fall in the Faeroe and Shetland Is; dust also covered a ship near the Orkney Is, 800 km from the volcano. Dust layer over SE Iceland 1 to 4 cm thick. Further explosive phases of the eruption on 15. iv. 1846 with three new craters sending pillars of fire up to 4 km high. (Dust believed to have fallen in Co. Sligo, NW Ireland in 1846 damaging the leaves of the potato crop, in addition to the blight of that year, and marking other objects in the open with 'black spots like ink'.) Reports of richly coloured sunsets and afterglows in Britain from 2. ix. 1845 through September and first half of October 1845 and in Switzerland from mid-April till end of May 1846											
		d.v.i. adjusted for prolonged eruption with at least two explosive phases. Dust layer marked in Iceland soil profiles, but apparently not thoroughly surveyed											
1846	**Amargura (Armagora) Island, S. Pacific** (K, S, Sh)	18° S, 174° W	$a_3?\ b_2$	—	—	$(\mathbf{1000})_{\mathrm{adj}}$	(330)	(670)	—	—	—	1.0	**(1000)**
		(S) gives this eruption as 1846–1847											
		As 1846 was a warm year in the northern temperate zone, followed by much colder years, there is some suggestion that dust from Armagora 1846 reached the northern hemisphere after a delay of rather more than a year											
1846	Zuqar Is., Red Sea (K)	14° N, 43° E	—	—	—	(< 10)	(< 10)	(< 10)	—	—	—	—	—
1847	Irazu, Costa Rica (K)	10° N, 84° W	—	—	—	—	—	—	—	—	—	—	—
ii. 1850	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
		From the 1850s onwards Vesuvius was a tourist attraction, the summit normally 'on fire'; the lava could usually be seen 'surging about in waves like an angry sea at a bar', but it did not flow out											
		Total veil 1846–50: E_{max} taken as 0.7 since the dust originated partly near 64° N, partly near 18° S			$\mathit{1300}_2$	—	—	—	—	*0.6*	*60*	*0.7*	*1800*
20. viii. 1852	Etna (K, S, Sh)	38° N, 15° E	a_2b_3	—	—	50_{av}	50	—	—	—	—	0.5	100
1852	Fuego, Guatemala (K)	14½° N, 91° W	—	—	—	—	—	—	—	—	—	—	—
1852	Mauna Loa, Hawaii (S, Sh)	19½° N, 156° W	a_2	—	$(\mathit{250})_2$	—	—	—	—	—	—	—	(250)

1852	Gunung Api, Banda, Moluccas (K)	$4\frac{1}{2}$° S, 130° E	—	—	—	(200)	(100)	(100)	—	—	—	1.0	(200)
		Total veil 1852–3: estimate doubtful because of discrepant course of Köppen's temperature curves at this point											
ii. 1854	Shiveluch, Kamchatka (K)	$56\frac{1}{2}$° N, $161\frac{1}{2}$° E	—	—	—	—	—	—	—	—	—	—	—
1–28. v. 1855	Vesuvius (K)	41° N, 14° E Violent (K)	—	—	—	?	?	—	—	—	—	—	—
1855	Mauna Loa (S, Sh)	$19\frac{1}{2}$° N, 156° W	a_2	—	—	—	—	—	—	—	—	—	—
1855–6	Cotopaxi, Ecuador (H, K, Sh)	1° S, 78° W	—	—	—	**(700)**	(350)	(350)	—	—	—	—	**(700)**
		In eruption in 1855 and 1856. Ash fell 50 km from Cotopaxi on 13. xii. 1856 Cone of pink light reported (seen from London?) after sunset in November and December 1855. Remarkable after-glows and cone of pink light repeatedly observed in England from November 1856 till February 1857											
1856	Awu (Awoe), Celebes (S, Sh)	$3\frac{1}{2}$° N, $125\frac{1}{2}$° E	b_{2-3}	—	—	70_{av}	35	35	—	—	—	—	—
2. iii. 1856	Sangihe (Sangir, Sanguir), Celebes (K)	$3\frac{1}{2}$° N, $125\frac{1}{2}$° E (possibly Awu)	—	—	—	?	?	—	—	—	—	—	—
ix. 1856	Japanese volcano (K)	?	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
v.–vi. 1858	Vesuvius (K)	41° N, 14° E Violent (K)	—	—	—	(10)	(10)	—	—	—	—	0.5	(20)
1859	Mauna Loa (S, Sh)	$19\frac{1}{2}$° N, 156° W (S) reports 2.7 km³ lava emitted	a_1	—	—	—	—	—	—	—	—	—	—
		Total veil 1855–8. (H) and (K) accept evidence of a major dust veil in 1856–7 but differ in attributing it to 'Cotopaxi and others' or mainly to Vesuvius. d.v.i. calculation suggests another great eruption: probably Sangihe 1856 (d.v.i. *ca.* **500** implied)											
					1250_2	—	—	—	—	*0.6*	*42*	*1.0*	*1250*
8–27. v. 1860	Katla, Iceland (K)	$63\frac{1}{2}$° N, 19° W	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
8. xii. 1861	Vesuvius (K)	41° N, 14° E	—	—	—	—	—	—	—	—	—	—	—
29. xii. 1861	Makjan (Makyan), Molucca Is. (K, S, Sh)	$\frac{1}{2}$° N, $127\frac{1}{2}$° E	b_2	—	—	$(\mathbf{800})_{adj}$	400	400	—	—	—	—	**(800)**
1862	Vesuvius (K)	41° N, 14° E	—	—	—	—	—	—	—	—	—	—	—
		Reports of fine sunsets with unusual colours in England and central Europe in 1862 and 1863, including during the eruption of Vesuvius *Total veil* 1860–2. (Estimates vary widely according to which eruption and date is taken as the origin of most of the dust—Makjan 1861 seems likeliest, unless another great eruption occurred in 1859–60)											
					1500_2	—	—	—	—	*0.9*	*30*	*1.0*	*1500*
12. viii. 1863	Island in the Mediterranean (K)	?	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1863	Merapi, Java (K)	$7\frac{1}{2}$° S, 110° E	—	—	—	—	—	—	—	—	—	—	—
1864	Merapi, Java (K)	$7\frac{1}{2}$° S, 110° E	—	—	—	—	—	—	—	—	—	—	—
ii. 1865	Etna (K)	38° N, 15° E Continued in eruption till July	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1866	Santorin (Thira), Aegean (K)	$36\frac{1}{2}$° N, $25\frac{1}{2}$° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—

APPENDIX I (*cont.*)

date	volcano	position	Sapper's assessment of magnitude: *a*, lava *b*, tephra	estimated or assumed total dust for d.v.i./km³	dust veil index: total veil in groups of eruption years	dust veil index from individual eruptions: World	Northern Hemisphere	Southern Hemisphere	R_{Dmax}	T_{Dmax}/°C	t_{mo}	E_{max}	d.v.i./E_{max}
1867	Icelandic volcano (K)	?	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
xi. 1867	Vesuvius (K)	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	0.5	(20)
		Eruption continued till March 1868											
1868	Mauna Loa (S, Sh)	19½° N, 156° W	a_2	—	—	(10)	(10)	—	—	—	—	1.0	(10)
8. x. 1868 and 19–20. xi. 1868	Vesuvius (K)	41° N, 14° E	—	—	—	?	?	—	—	—	—	—	—
		Unusual sunset colours, various reds and white, in Surrey, England, in December 1868											
28. xi. 1868	Etna (K)	38° N, 15° E	—	—	—	—	—	—	—	—	—	—	—
19. v. 1869	Izalco, Salvador (K)	14° N, 89½° W	—	—	—	?	?	—	—	—	—	—	—
		(See note against 1798 eruption)											
26. v. 1869 till 7. vi. 1869	Etna (K)	38° N, 15° E	—	—	—	?	?	—	—	—	—	—	—
		Remarkable coloured sunsets and afterglows in England and central Europe reported on many dates from mid-July to late September 1869											
1869	Icelandic volcano (K)	?	—	—	—	—	—	—	—	—	—	—	—
		From the temperature curve d.v.i. calculation suggests a great eruption, possibly Colima 1869, 600–650 by subtraction of other eruptions from total veil											
vi. 1869	Colima, Mexico (K)	19½° N, 104° W	—	—	—	?	?	—	—	—	—	—	—
23. ii. 1870	Izalco, Salvador (K)	14° N, 89½° W	—	—	—	—	—	—	—	—	—	—	—
		(See above)											
1870	Cebcruco, Mexico (S, Sh)	21° N, 105° W	a_3b_{2-3}	—	—	50_{av}	50	—	—	—	—	0.7	70
		Total veil 1868–70: E_{max} taken as 0.7 as a compromise value to allow for the dust originating in various altitudes			*700_2*	—	—	—	—	*0.8*	*24*	*0.7*	*1000*
1870	Icelandic volcano (K)	?	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
23. iv. 1872 till 3. v. 1872	Vesuvius (H, K, S, Sh)	41° N, 14° E	$a_?b_{2-3}$	—	—	70_{av}	70	—	—	—	—	0.5	150
		Ash and smoke rose in column to 7–8 km height; ash discharged on 29. iv. 1872. Pumice fell to a depth of a few metres at Pompeii, sand fell all over Italy on 6. v. 1872. (Many sightseers lost when the sudden eruption broke out and lava issued from two new craters) (H) and (K) class both Vesuvius and Merapi 1872 eruptions as great. At Madrid the sunsets in 1872 were later considered comparable with those of 1883 (K)											
iv. 1872	Merapi, Java (H, K, Sh)	7½° S, 110° E	—	—	—	(80)	(40)	(40)	—	—	—	1.0	(80)
		Violent (K)											
vi. 1872	Japanese volcano (K)	?	—	—	—	—	—	—	—	—	—	—	—
		Not listed among great eruptions in Japan											

1872	‘all the volcanoes of the New Hebrides’ (K)	13–21° S, 165–172° E	—	—	—	?	?	?	—	—	—	—	?
		Total veil 1872: 100–150, depending on source of dust, here taken as mainly East Indies			*150*$_2$	—	—	—	—	*0.2*	*12–15*	*1.0*	*150*
9–10. i. 1873	Grimsvötn, Iceland (K)	$64\frac{1}{2}$° N, $17\frac{1}{2}$° W	—	0.3	—	10_3	10	—	—	—	(24)	0.3	30
		Heaviest tephra fall in recent times from this volcano under the Vatnajökull ice-cap, which probably reduced the output of dust, estimated totalling 0.3 km^3 over Iceland when freshly fallen											
29. viii 1874	Etna (K)	38° N, 15° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
ca. 15. xii. 1874	Icelandic volcano (K)	65° N, 17° W	b_1	3.5	—	(< 10)	(< 10)	—	—	—	—	—	—
		First signs of activity of Askja, according to Thorarinsson											
29. iii. 1875	**Askja, Iceland** (H, K, S, Sh)	65° N, 17° W	b_1	3.5	—	**300**$_{2,3}$	300	—	—	1.0	24	0.3	1000
		3 to 4 km^3 solid ejecta estimated by Thoroddsen, apparently much of it as dust											
		Dust from Askja deposited in Norway and Sweden, and possibly in Germany also. The eruption continued into April. The air over the British Isles was hazy up to great heights in April											
		380$_2$, 110_3											
1875	Sveinagja, Iceland (S, Sh)	?	a_2	—	—	—	—	—	—	—	—	—	—
		Lava eruption											
1877	Cotopaxi, Ecuador (S, Sh)	1° S, 78° W	a_3b_{2-3}?	—	—	50_{av}	25	25	—	—	—	1.0	50
1877	Mauna Loa (K)	$19\frac{1}{2}$° N, 156° W	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
1878	Ghaie, New Ireland, Bismarck Archipelago (S, Sh)	4° S, 152° S	b_{2-3}	—	—	?	?	?	—	—	—	—	—
1880	Bridgeman Island, South Shetlands	62° S, $56\frac{1}{2}$° W	—	—	—	—	—	—	—	—	—	—	—
		Again smoking (see also 1821 and 1839); unlikely that the visitors happened to witness the greatest moment of activity. No signs of activity when visited in 1909											
3. vii. 1880	Cotopaxi, Ecuador	1° S, 78° W	—	—	—	?	?	?	—	—	—	—	—
		Dust column to 12 km height. 2 Tg (2 million tons) of dust fell over a wide area, perhaps equivalent to 3×10^6 m^3 of loose freshly fallen dust, i.e. b_4 on Sapper's scale. Green Sun reported											
1880	Lake Ilopango, Salvador	*ca.* $13\frac{1}{2}$° N, 89° W	—	—	—	?	?	?	—	—	—	—	—
		An island 50 m high raised in the lake											
		Total veil 1878–81. Direct evidence of dust and optical effects fails, but several eruptions, acknowledged to be of some importance, did occur in the equatorial zone. Ghaie 1878 listed by Shaw as a major eruption			*(1250)*$_2$	—	—	—	—	*0.6*	*40*	*1.0*	*(1250)*
1880–1882	Mauna Loa (S, Sh)	$19\frac{1}{2}$° N, 156° W	a_2	—	—	(< 10)	(< 10)	—	—	—	—	—	—
26–27. viii. 1883	**Krakatau** (H, K, S, Sh)	6° S, $105\frac{1}{2}$° E	b_1	6.0	1000	—	—	—	27	0.5	38	1.0	**1000**
		Great explosive eruption, preceded by other explosive phases starting on 20. v. 1883											
		Much or most of the island of Krakatau and its mountain disappeared; other small nearby islands in the Sunda Strait also disappeared, whereas some new islands were formed and the topography of the sea bottom was greatly altered. Estimates of the amount of solid matter blown up range from 6 to 18 km^3. Pumice floating on the sea blocked the passage of ships in the strait											
		Appreciable amounts of ash fell 500 km away as early as 21. v. 1883, and on 24. v. 1883 dust was noticed at Timor, 2000 km away to the east, probably carried by the tropospheric winds since the greatest height observed for the dust column at that stage was 11 km. Dust 50 cm deep on Krakatau before the final and greatest explosions on 26–27 August, which threw a dust column up to an observed height of 27 km. ‘Complete darkness’ for a period at noon on 27. viii. 1883 at Bandung, Java, 250 km away, a couple of hours after the greatest explosion. Darkness lasted 4 to 5 h at Batavia, 160 km from the volcano, and there was much rain											
		Pyrheliometric measurements of the direct solar beam, available for the first time before and after a great eruption, indicated a reduction of 20–30 % in the monthly mean values, deficiency of up to 20 % occurring as long as 2–3 years after the eruption; though this loss of incoming direct-beam radiation would be to a considerable extent made up by an increase of the diffuse (sky) radiation (see text											

Appendix I (*cont.*)

date	volcano	position	Sapper's assessment of magnitude: *a*, lava *b*, tephra	estimated or assumed total dust for d.v.i. km³	dust veil index: total veil in groups of eruption years	dust veil index from individual eruptions: World	Northern Hemisphere	Southern Hemisphere	R_{Dmax}	T_{Dmax} / °C	t_{mo}	$E_{\max}$	d.v.i. / $E_{\max}$
		and compare observational evidence of this noted against the 1783 and 1831 eruptions). Dust skies were indicated by optical effects observed over Europe and North America and Hawaii from autumn 1883, and over the southern temperate zone from South Africa (Transvaal) to Sydney, Australia and St Paul Island (39° S, 78° E) from September to November 1883 onwards											
6. x. 1883	St Augustine, Alaska (H, K, S, Sh)	59½° N, 153½° W	b_{2-3}	—	—	20_{av}	20	—	—	—	—	0.3	70
		(H) and (K) class this Alaskan eruption also as a great eruption											
1883	Bogoslof (Bogoslov), Aleutians (S, Sh)	54° N, 168° W	*a*? *b*?	—	—	—	—	—	—	—	—	—	—
1885	**Falcon Is.** (S, Sh)	20° S, 175° W	b_2?	—	—	$\mathbf{300}_{av}$	100	200	—	—	—	1.0	**300**
21. v. 1886	Etna (K)	38° N, 15° E	—	—	—	(10)	(10)	—	—	—	—	0.5	(20)
		Eruption continued for several days. Smoke column to 14 km height on 24 May											
		Haze, mist and fog following this eruption made the sun invisible in Sicily. The haze spread over all Italy from 29. v to 3. vi. 1886 so that the sun was grey at 30° elevation and red at all lower elevations. There were reddish yellow afterglows that summer											
10. vi. 1886	**Tarawera, North Island, New Zealand** (H, K, S, Sh)	38½° S, 176½° E	b_1	1.5	—	(**400**)	—	(400)	—	—	(30)	0.5	(**800**)
		(S) estimated 1.5 km³ solid matter ejected											
		100_3, 500_{av}											
1886	**Niafu, Tonga Is.** (S, Sh)	16° S, 175½° W	b_2	—	—	$\mathbf{300}_{av}$	100	200	—	—	—	1.0	**300**
1888	**Bandai san, Japan** (S, Sh)	38° N, 140° E	b_1	3	—	(**250**)	(250)	—	—	—	—	0.5	(**500**)
		One side of the peak Kobandai blown away, estimated as approximately 3000 Tg (3 × 10⁹ tons) of material, probably equivalent to 3 km³ in volume when freshly fallen. Eruption brief; one violent explosion											
		330_3 Bandai san											
1888	**Ritter Island, Bismarck Archipelago** (S, Sh)	5½° S, 148° E	b_1	1.7	—	(**250**)	(125)	(125)	—	—	—	1.0	(**250**)
		(S) estimates 1.7 km³ solid matter blown up											
		370_3 Ritter Is.											
ii. 1890	Bogoslof (Bogoslov), Aleutians (H, Sh)	54° N, 168° W	—	—	—	(50)	(50)	—	—	—	—	0.3	(170)
		(H) classes Bogoslof 1890 as a great eruption											
		Direct solar beam reduced in certain months 1888–91 by 15–20 % from its preceding and succeeding values											
		Total veil 1888–90; *440*₁, *650*₂, *680*₃. t_{mo} taken from radiation curve because temperature curves of different zones disagree											
					$\mathit{550}_{1,2}$	—	—	—	*18*	(*0.5*)	(*50*)	(*0.5*)	*1100*
1891–9	Vesuvius (S, Sh)	41° N, 14° E	a_2b?	—	—	?	?	—	—	—	—	—	—
		d.v.i. about 100_2 in 1895–6 possibly partly associated with Vesuvius											
7. vi. 1892	Awu (Awoe), Great Sangihe Is. (H, S)	3½° N, 125½° E	—	—	—	100_1	—	—	8	—	15	1.0	100
		(H) classes Awu 1892 as a great eruption											
		Direct solar beam strength reduced by 4 to 8 % in 1893–4 and in 1895–6, compared with the immediately preceding and succeeding periods, presumably in the first place due to the Awu 1892 eruption and possibly some contribution from Vesuvius											

1895 or 1896?	**'Thompson Island'** (island supposed destroyed by great volcanic eruption—see Baker 1967; Lamb 1967)	*ca.* 54° S, 5° E	—	—	—	**(400$_{2,3}$)**	—	(400)	—	(0.2)†	(120)	(0.3)	**(1300)**
		Sharp lowering of temperatures at the few observing stations then existing south of 45° S, which took from 3 to (farther south) 10 years to recover. No direct report of dust veil known (possibly confined to almost uninhabited regions south of 40 to 45° S). Quantity of matter blown up can be estimated from reported size and sketched appearance of the island before its disappearance											
		Whole entry in brackets because the evidence of temperature lowering and disappearance of the island cannot be supported by any direct observation of the presumed eruption or the dust cloud											
1898	Una-Una (Oena-Oena), Celebes (S, Sh)	0° S, 122° E	$b_{3\,(-2?)}$	—	—	140$_{adj}$	70	70	—	—	—	1.0	140
4. vii. 1899	Mauna Loa (S, Sh)	19½° N, 156° W	a_2	—	—	10	10	—	—	—	—	1.0	10
		Total veil 1898–9			*150$_1$*	—	—	—	*10*	—	*15–18*	*1.0*	*150*
8. v. 1902	**Mont Pelée, Martinique** (H, S, Sh)	15° N, 61° W	a_2b_2	—	—	(100)	—	—	—	—	—	1.0	(100)
		Hot dust rain disaster in Martinique											
17. v. 1902	**Soufrière, St Vincent** (S, Sh)	13½° N, 61° W	b_1	1	—	**(300)**	—	—	—	—	—	1.0	**(300)**
		Some dust fell 3000 km away; dust cloud estimated to have carried millions of megagrammes (tons), over 1 km³, of solid matter (S)											
		170$_3$											
24. x. 1902	**Santa Maria, Guatemala** (H, S, Sh)	14½° N, 92° W	b_1	5.4	—	**(600)**	—	—	—	—	—	1.0	**(600)**
		Santa Maria continued in eruption till 1906, producing estimated total of 5.4 km³ of solid ejecta (S)											
		900$_3$											
		Direct solar beam strength weakened by about 20 % in certain months from late 1902 till early 1904, recovering its previous level only in mid-1905. Twilight phenomena reported till 1904											
ii.–iii. 1903	Colima, Mexico (H, Sh)	19½° N, 104° W	—	—	—	—	—	—	—	—	—	—	—
1904	Minami Iwoshima (Iwo, Iwojima) island (S, Sh)	24° N, 141° E	a_2b_2?	—	—	30$_1$	30	—	—	—	—	0.7	40
		Total veil 1902–4: *750$_1$, 1100$_2$, 850$_3$*			*900$_{1,2}$*	—	*450*	*450*	*22*	*0.6*	*36*	*1.0*	*900*
1905–6	Savaii (Sawaii), Samoa (S, Sh)	14° S, 172° W	a_1b_4	—	—	(10)	(5)	(5)	—	—	—	1.0	(10)
		Great lava eruption, much more than 1 km³ of lava											
1906	Vesuvius (S, Sh)	41° N, 14° E	a_3b_3	—	—	20$_{av}$	20	—	—	—	—	0.5	40
1906–7	Bogoslof (Bogoslov), Aleutians (S, Sh)	54° N, 168° W	a_2b_{2-3}	—	—	20$_{av}$	20	—	—	—	—	0.3	70
28. iii. 1907	Shtyubelya Sopka (Ksudatch, Sjadutka)	52° N, 157½° E	b_1	3	—	150$_{1,2,3}$	150	—	13	0.5	30	0.3	**500**
		(S) estimates 3 km³ of solid matter blown into the atmosphere											
		Twilight phenomena 1907 to 1909											
		110$_1$, 240$_2$, 120$_3$											
1907	Mauna Loa (S, Sh)	19½° N, 156° W	a_2	—	—	(10)	(10)	—	—	—	—	1.0	(10)
1910	Heard Island	53° S, 73½° E	—	—	—	—	—	—	—	—	—	—	—
		No details known											
1911	Leskov Island, South Sandwich Islands	56½° S, 28½° W	—	—	—	—	—	—	—	—	—	—	—
		Signs of activity reported by Filchner											

† Area adjusted average of regions affected and those unaffected.

Appendix I (*cont.*)

date	volcano	position	Sapper's assessment of magnitude: *a*, lava *b*, tephra	estimated or assumed total dust for d.v.i./km³	dust veil index: total veil in groups of eruption years	dust veil index from individual eruptions: World	Northern Hemisphere	Southern Hemisphere	R_{Dmax}	T_{Dmax}/°C	t_{mo}	E_{max}	d.v.i./E_{max}
1911	Taal, Luzon (S, Sh)	14° N, 121° E	b_2	—	—	30_1	15	15	4	—	9	1.0	30
1912	Erebus, Ross Island, Antarctic	77½° S, 167° E Pumice and ash given out	—	—	—	?	—	?	—	—	—	—	—
6. vi. 1912	Katmai, Alaska (H, S, Sh)	58° N, 155° W Katmai 1912 classed in (H) as a great eruption Estimates of solid ejecta irreconcilable: (S) estimates 21 km³ solid ejecta, though the eruption was brief. Another estimate was only 0.006 km³ The sky over England never acquired its normal blue even in the clearer periods of the wet summer of 1912. Evidence of the dust cloud over Algeria by 19. vi. and over California by 21. vi. 1912. Twilight phenomena till 1914. Direct solar beam weakened by about 25% from its preceding values in certain months in 1912, approximately regaining its former level first in late 1914 Widely discrepant estimates of solid ejecta (see p. 473) 190_1, 150_2	b_1	—	—	$150_{1,2}$	150	—	24	0.3	27	0.3	**500**
1913	Colima, Mexico (S, Sh)	19½° N, 104° W	b_2	—	—	(10)	(7)	(3)	—	—	—	1.0	(10)
1914	Sakurashima, Japan (S, Sh)	31½° N, 131° E Dust survey indicates about 0.5 km³ of dust deposit Sapper's estimates of magnitude creep upwards towards the last years of his survey and seem, at this date, unreliable 35_1, 7_3	$a_1 b_2$	0.5	—	$20_{1,3}$	20	—	(6)	—	6	0.5	40
1914	Minami Iwoshima (Iwo, Iwojima) (S, Sh)	24° N, 141° E No major eruptions in 1914 reported in Japanese meteorological texts concerned with volcanic effects	b_1	—	—	(< 10)	(< 10)	—	—	—	—	—	—
19 and 22 v. 1915	Lassen Peak, California	40½° N, 121½° W Two explosions with dust column up to 6–7 km as climax of minor activity 1914–16. No dust veil seen more than 300 km from the volcano	—	—	—	< 1	—	—	—	—	—	—	—
29. ix. 1919	Mauna Loa	19½° N, 156° W Brief, but much lava	—	—	—	—	—	—	—	—	—	—	—
1921	Andes (frontier of Chile and Argentina)	— Direct solar beam as observed in the southern hemisphere (Chile) weakened by not less than 1 to 1.7% in certain months in 1922, regaining its former level in 1923 It has been alleged that world temperature was lowered by this eruption for some time afterwards. The temperature anomaly over the zone 5 to 35° S, presumably more affected than any other zone except those to the south of 35° S, appears to have been only −0.1 to −0.2° C in 1921 and 1922 95_2	—	—	—	100_2	—	100	—	0.2	18	0.5	200
1927–30	Visokoi Island, South Sandwich Islands	56½° S, 27½° W Much vapour produced	—	—	—	—	—	—	—	—	—	—	—
i. 1929	Candlemas Island, South Sandwich Islands	57° S, 27° W Activity observed	—	—	—	—	—	—	—	—	—	—	—

1929	Asama, Japan	$36\frac{1}{2}$° N, $138\frac{1}{2}$° E	—	0.1	—	4_3	4	—	—	—	18	0.5	8
		Two explosions, estimated to have thrown up about 0.1 km³ of dust in all											
1930 or 1931	Gorely, Kamchatka	53° N, 158° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
		Dust eruption											
iii. 1931	Kluchev, Kamchatka	56° N, $160\frac{1}{2}$° E	—	0.1	—	2_3	2	—	—	—	12	0.3	5
		About 0.12 km³ of solid material blown away											
10. iv. 1932	Quizopu, Chile	$35\frac{1}{2}$° S, $70\frac{1}{2}$° W	—	—	—	35_1	—	35_1	3.3 to 3.6	—	20	0.5	70
		Altogether about 5 volcanoes in the Chilean Andes, all about 36° S, erupting for some days											
		Dust fell in Montevideo and Buenos Aires, 1300 km to the east. Magnitude appears from this rather similar to Hekla 1947 from radiation obs. 1931–36 at Mt Stromlo, Canberra, Australia (reported by Rimmer 1937)											
3. xii. 1935	Bristol Is., South Sandwich Islands	58° S, 27° W	—	—	—	—	—	—	—	—	—	—	—
		Violent eruption reported, but no dust cloud observed											
1936–7	Kluchev, Kamchatka	56° N, $160\frac{1}{2}$° E	—	—	—	(3)	(3)	—	—	—	—	—	(3)
		Eruptive activity began in November 1936 and built up to an explosive climax in July 1937											
i. 1937	Mount Darnley, Bristol Is., South Sandwich Islands	59° S, $26\frac{1}{2}$° W	—	—	—	(10)	—	(10)	—	—	—	—	—
		Lava flowing into the Southern Ocean											
i. and iii. 1937	Sopka Tolbachik, Kamchatka	55° N, $160\frac{1}{2}$° E	—	—	—	(2)	(2)	—	—	—	—	—	—
		Ash eruption											
9–11 ii. 1937	Llaima, Chile	$38\frac{1}{2}$° S, 72° W	—	—	—	(< 10)	—	(< 10)	—	—	—	—	—
		'Much ash and lava'; flooding caused by melted snow											
29. v.– 10. vi. 1937	Raluan and Ghaie, Bismarck Archipelago	4° S, 152° E	—	—	—	(10)	5	5	—	—	—	1.0	(10)
		Produced new volcanic island (Raluan), 243 m high											
		Total (*Northern Hemisphere*) *veil* 1936–7			*10_1*	—	—	—	*3*	—	*10*	*0.3*	*30*
1937–8	Tjerimay, Java	7° S, $108\frac{1}{2}$° E	—	—	—	< 5_1	—	—	—	—	—	—	—
		Long-lasting ash eruption (24. vi. 1937 to 7. i. 1938) and some lava											
5. vi. 1938	Mayon, Luzon	13° N, 124° E	—	—	—	< 5_1	—	—	—	—	—	—	—
		'Much lava and widespread ash fall'											
1939–41	Santorin (Thira), Aegean	$36\frac{1}{2}$° N, $25\frac{1}{2}$° E	—	—	—	(< 1)	(< 1)	—	—	—	—	—	—
		New crater cones formed in eruptive activity which lasted from 20. viii. 1939											
		Many small explosions: clouds of ash and steam up to 500–1500 m											
i.–iii. 1944	Vesuvius	41° N, 14° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
		Smoke and ash eruption only. The ash column observed carried up to 6 km height. Dust fell 700 km away in the Balkans 18–26. iii. 1944. Darkness in Bari 200 km from the volcano											
1944–6	Kluchev, Kamchatka	56° N, $160\frac{1}{2}$° E	—	—	—	(< 10)	(< 10)	—	—	—	—	—	—
		Frequent rain of ashes from December 1944 all through 1945 and 1946											
29. iii. 1947	Hekla	64° N, $19\frac{1}{2}$° W	—	—	—	(20)	(20)	—	—	—	(24)	0.3	(70)
		Ash column to 27 km height briefly in explosion on 27. iii. 1947 during the first hour of eruption, which lasted till September; there was probably no further injection of material into the stratosphere after the first morning. Dust layer 10 cm thick over part of southern Iceland, dust fell on ships in the Atlantic 800–1000 km away to the south and east and three days later in Finland. Comparison with the surveys of the ash layers in Iceland from other eruptions indicates a total of only 0.18 km³ of solid matter put into the atmosphere on this occasion (b_3 Sapper rating). Estimation is however difficult because the greater part of the dust must have been deposited in the Atlantic south and east of Iceland. Lava flow until 1948											
		6_3 if 0.18 km³ estimate of dust be accepted											
1–23. vi. 1950	Mauna Loa	$19\frac{1}{2}$° N, 156° W	—	—	—	(< 1)	(< 1)	—	—	—	—	—	—
		'One of this mountain's greatest eruptions', estimated to have produced 0.5 km³ of lava (Sapper's a_2 rating); but no ash mentioned											

Appendix I (*cont.*)

date	volcano	position	Sapper's assessment of magnitude: *a*, lava *b*, tephra	estimated or assumed total dust for d.v.i. km³	dust veil index: total veil in groups of eruption years	dust veil index from individual eruptions: World	Northern Hemisphere	Southern Hemisphere	R_{Dmax}	T_{Dmax} °C	t_{mo}	E_{max}	d.v.i./E_{max}
i. 1951	Mount Lamington, Papua, New Guinea	9° S, 148° E	—	—	—	20_1	10	10	0.6	—	30	1.0	20
		First known eruption of this mountain. 4000 killed. Largely an eruption of hot ash; summit cone raised 600 m. Eruptive activity continued until 1952											
9. vii. 1953	Mount Spurr, Alaska	61° N, 153° W	—	—	—	$2_{1,3}$	2	—	*ca.* 1?†	—	—	0.3	7
		Explosion blew dust column to 23 km height. Darkness for 2 h at Anchorage, Alaska, 125 km away, where dust fell 0.3 to 0.6 cm deep. Dust cloud travelled around the hemisphere over N. America and Europe, where optical effects were briefly observed. (The dust fall appears much less than from Hekla in 1947)											
		3_1(?), 1_3. Any reduction of transmitted solar radiation after this eruption appears masked (because swamped by a 5 % reduction evidently due to other causes that characterized 1953 and 1954 from beginning to end) (see p. 466)											
vi. 1955	Ranco and Puyehue, Chile	40–41° S, 72° W	—	0.3	—	20_3	—	20	—	—	(24)	0.5	30–40
		Estimated 0.3 km³ of tephra blown into the atmosphere (b_2 Sapper rating). Ash fell at Santiago, Chile 800 km north of the eruption											
30. iii. 1956	Bezymjannaja, Kamchatka	56° N, 160½° E	—	—	—	(10)	(10)	—	?	—	6	0.3	(30)
		Activity began in 1955; explosive phase on 30. iii. 1956 blew dust column up to 45 km height. Dust-fall observed 400 km away NE of the volcano											
		The dust cloud from Bezymjannaja travelled across the Arctic and was observed by aircraft over Britain. Most of the dust is believed to have been coarse and to have settled out quickly, but some abnormal sky effects were reported up to 6 months after the explosion											
		Assessed as greater than Hekla 1947 because the explosion carried debris to a greater height. *q* estimated as 1.0 km³ (Gorshkov 1959) gives d.v.i. = 8_3											
1956	Bristol Is., South Sandwich Islands	58° S, 27° W	—	—	—	—	—	—	—	—	—	—	—
		Violent eruption reported, but no dust cloud observed											
9. x. 1957	Fayal, Azores	38½° N, 29° W	—	—	—	(1)	(1)	—	—	—	—	—	(1)
		Submarine eruption produced temporary island offshore, near Capelinhos lighthouse and covered all Fayal island with grit and ashes. Column of dust observed rising to 8–10 km height											
		no twilight effects reported											
1955–7 (exact date unknown)	Bouvet Island (Bouvetöya), Southern Ocean	54½° S, 3½° E	—	—	—	—	—	—	—	—	—	—	—
		Small lava eruption, but no dust cloud observed											
21–24. v. 1960	Puntiagudo volcano, and several new volcanic vents in southern Chile	39–45° S, 72–73° W	—	—	—	(50)	—	(50)	—	—	—	0.5	100
		Black ash fell in Bahia Blanca, 800–1000 km from source, forming a uniform deposit about 1 mm thick											
		Eruption of fire, smoke and ash blown up to about the tropopause for several days following great earthquakes. Reports reaching Europe dwelt most on the earthquakes, but photographs of the dust cloud appear to indicate much greater magnitude than the Hekla 1947 eruption and some penetration of the stratosphere probable. The deformation of the local topography appears to have been greater than for many years in Chile											
28. iii. 1961	Bezymjannaja, Kamchatka	56° N, 160½° E	—	—	—	(< 1)	(< 1)	—	—	—	—	—	—
		Dust blown up to 7 km height. Dustfall blackened the snow 50 km away. (This volcano was considered extinct until 1955)											
10. x. 1961	Tristan da Cunha	37° S, 12½° W	—	—	—	(< 1)	—	(< 1)	—	—	—	—	—
		A new volcanic cone formed near sea level. Smoke continued to issue forth for some months. (Believed to be the first eruption on Tristan for about 2500 years)											
5. iii. 1962	South Sandwich Islands	55.9° S, 27.9° W	—	—	—	?	—	?	—	—	—	—	—
		British Antarctic Survey ship found evidence of an important submarine eruption in the South Sandwich Islands; position and precise											

† Possibly 10 based on tentative indications of duration of a radiation effect at U.S. observatories, as interpreted by Wexler.

		date arrived at from supposed association with the epicentre of a submarine earthquake, the sea surface being 'choked with pumice over an area of thousands of square miles', thick enough to halt the vessel. (Not known whether any dust or tephra ejected into the atmosphere or whether the eruption was at such depth in the ocean that only pumice floated up to the surface)											
19. ii.– 17. iii. 1963 (especially 17. iii. 1963)	**Gunung Agung (Agoeng), Bali**	$8\frac{1}{2}°$ S, $115\frac{1}{2}°$ E	—	—	—	**$800_{1,2}$**	400	400	25	0.4	(36)†	1.0	**800**
		Explosive eruptions in February and March 1963, especially 17. iii. 1963; many hundreds or thousands of people killed Dust veil encircled the Earth and spread to all latitudes, causing 'overcast skies' in Arabia and the Persian Gulf in summer 1963, remarkable sunsets, white glare around the Sun, prolonged twilight glows and other optical effects over N. America and Europe from autumn 1963 onwards to about January 1965: some abnormality of sunset colours detectable at least until 1966. Dust obtained in air samples from the stratosphere over Sydney, Australia, in May 1963 and still present in quantities detectable by radiation measurements in 1966. Reduced intensity of direct solar beam measured over southern Africa from March 1963 onwards, and over Australia, amounting to 30 % deficit in some places in some months. Measurements also showed compensatory increase of sky radiation Surface temperature averages for those months in 1964 for which an overall world survey appeared possible indicated a deficit of about 0.4 °C below the averages of the last 12 months before the eruption: the margin of error to which this estimate is liable is indicated by mean departures ranging from −0.2 to −0.7 °C obtained for different months. As late as May 1965 a departure of −0.2 to −0.3 °C was indicated. (Some part of the predominance of negative temperature anomalies with respect to the average values for 1930–60, and the continuance of these anomalies late in 1966, should probably be set down to an independent climatic trend which started 20 years before the eruption) 875_1, 750_2											
1963–5	Surtsey, Iceland	63° N, $20\frac{1}{2}°$ W	—	0.7	—	15_3	15	—	—	—	5	0.3	45
		Eruption began with cloud of steam rising out of the sea 15 km southwest of Westman Islands on 14. xi. 1963 and continued (after April 1964 mainly effusion of lava) through 1964, by which time a new island 170 m high and 1.7 km long by 1.3 km broad had been established. Prolonged emission of smoke, steam, ash and lava, the ash column frequently reaching to 9 km height, occasionally to the tropopause, but doubtful if it ever entered the stratosphere. Dust dispersed in all directions but scarcely 1 cm depth of dust-fall over the neighbouring islands 'In spite of the small amount of ash we have had red sunsets and sunrises' in Iceland (Thorarinsson personal communication, May 1964): these may however have been attributable to the Bali dust in the stratosphere. By June 1965 the Surtsey eruption had ceased and the island was cool enough for the first plants to gain a foothold: continued, though weaker eruptive activity nearby was producing another new island just east of the one formed in 1963–4 Marked up because of prolonged eruption of tephra over 5 months though injection(s) into the stratosphere uncertain											
28. ix. 1965	Taal, Luzon	14° N, 121° E	—	0.09	—	$(10–15)_3$	(7)	(6)	—	—	(24–36)	1.0	(10–15)
		Six hours of paroxysms with ash clouds rising to 16–20 km											

ADDENDUM: Preliminary notes of eruptions 1966 to 1968

12. viii. 1966	Awu, Great Sangihe Island, Celebes	$3\frac{1}{2}°$ N, $125\frac{1}{2}°$ E	—	~3 km³ dust	—	possibly 150 to 200	—	—	—	—	~12‡	1.0?	possibly 150 to 200
		Explosive eruption, ash, boulders and lava ejected. Ship reported total darkness in mid-afternoon at a point 370 km SW of the volcano. Dustfall on the ship 1 to 1.5 cm deep, but evident that a very dense cloud was overhead. Winds of the required direction and speed to carry the dust cloud over the ship's position existed only at heights between about 12 and 18 km. Probable therefore that most of the dust concentration was within the troposphere Uncertain what amount of dust may have entered stratosphere, though likely that some did. No direct reports of stratospheric dust from this source, perhaps because Bali 1963 dust was still present. No firm assessment possible but may have contributed to the apparent longevity of the Bali veil (very slight optical effects still detected over northeastern U.S.A. in January 1968)											
4–7. xii. 1967	Deception Island	63° S, $60\frac{1}{2}°$ W	—	~0.1	—	$< 1_3$	—	—	—	—	—	—	—
		Ash eruptions formed a new island in Telefon Bay, 65 m high and 1 to 2 km long by 0.8 km broad. Top of ash column during the eruption observed at one time rising to 10 km on 5. xii. 1967, but by 7. xii. 1967 not above 6 km Duration believed short. Uncertain whether any dust entered stratosphere. Clearly a minor eruption leading to no significant dust veil											
11–12. vi. 1968	Fernandina Island, Galapagos	$\frac{1}{2}°$ S, 92° W	—	1 to 2(?) km³	—	possibly 50 to 100	—	—	—	—	~10	1.0?	possibly 50 to 100
		Explosive eruption. Central part of island in old crater blown up. Ash fall 1 cm deep over area up to 300 km W of volcano, where Sun blotted out by high level veil Few reports so far received. No firm assessment possible at date of writing											

† Provisional estimate of total duration of significant effects in middle latitudes.
‡ Provisional estimates of Awu 1966 and Fernandina 1968 dust veil durations from radiation observed by Dyer at Melbourne.

Appendix II. Tables of yearly values of d.v.i.: Northern Hemisphere

Table 7 (a)

Omitting dust veils deduced solely from evidence of temperature behaviour. Limited to eruptions producing d.v.i. ⩾ 100. (See text, pp. 484–485, regarding apportioning of the index over the years.)

year ...	1500	1510	1520	1530	1540	1550	1560	1570	1580	1590
0	200	—	—	—	—	—	—	—	—	—
1	150	—	—	—	—	—	—	—	—	—
2	100	—	—	—	—	—	—	—	—	—
3	50	—	—	—	—	100	—	—	—	200
4	—	—	—	—	—	500	—	—	—	150
5	—	—	—	50	—	350	—	—	—	100
6	—	—	—	50	—	200	—	—	200	50
7	—	—	—	50	—	100	—	—	150	40
8	—	—	—	—	—	—	—	—	100	30
9	—	—	—	—	—	—	—	—	50	20
year ...	**1600**	**1610**	**1620**	**1630**	**1640**	**1650**	**1660**	**1670**	**1680**	**1690**
0	10	40	—	—	150	100	340	—	280	—
1	400	30	—	120	400	75	255	—	210	—
2	300	20	—	90	275	50	170	200	140	—
3	210	10	—	60	175	25	85	150	70	140
4	110	200	—	30	75	—	130	100	—	285
5	10	150	100	—	—	—	100	50	—	205
6	20	100	75	40	60	—	65	—	—	105
7	50	50	50	30	45	—	30	—	—	45
8	50	—	25	120	30	—	—	—	—	—
9	50	—	—	85	15	—	—	—	—	—
year ...	**1700**	**1710**	**1720**	**1730**	**1740**	**1750**	**1760**	**1770**	**1780**	**1790**
0	—	75	30	160	—	—	110	—	135	—
1	—	—	100	130	—	—	77	—	90	—
2	—	80	75	90	—	200	45	50	45	—
3	—	60	50	50	—	150	13	37	400	—
4	—	40	55	—	60	160	—	25	300	—
5	—	20	15	—	45	255	—	13	200	120
6	—	—	15	—	30	150	—	—	160	130
7	300	120	15	—	15	95	—	—	45	90
8	225	90	15	—	—	40	—	—	30	50
9	150	60	15	—	—	80	—	180	15	130
year ...	**1800**	**1810**	**1820**	**1830**	**1840**	**1850**	**1860**	**1870**	**1880**	**1890**
0	90	—	15	50	150	—	—	80	—	85
1	60	80	—	200	75	—	160	40	—	45
2	30	180	200	130	—	—	120	—	—	20
3	—	170	150	80	—	—	80	—	400	15
4	—	170	100	40	—	—	40	—	300	10
5	—	695	70	525	100	—	—	120	240	5
6	—	490	80	450	205	140	—	90	170	—
7	—	375	65	375	140	105	—	60	50	—
8	—	195	50	300	90	70	160	30	170	30
9	—	30	75	225	30	35	120	—	125	25
year ...	**1900**	**1910**	**1920**	**1930**	**1940**	**1950**	**1960**			
0	15	15	—	—	—	—	—	—	—	—
1	5	—	—	—	—	—	—	—	—	—
2	180	60	—	—	—	—	—	—	—	—
3	135	45	—	—	—	—	160	—	—	—
4	90	30	—	—	—	—	120	—	—	—
5	45	15	—	—	—	—	80	—	—	—
6	—	—	—	—	—	—	40	—	—	—
7	60	—	—	—	—	—	—	—	—	—
8	45	—	—	—	—	—	—	—	—	—
9	30	—	—	—	—	—	—	—	—	—

TABLE 7 (*b*)

Including those dust veils deduced and assessed solely on evidence of temperature anomaly. 1500–1699 and 1900 onwards as table 7 (*a*).

year ...	1700	1710	1720	1730	1740	1750	1760	1770	1780	1790
0	—	75	30	160	—	—	110	90	135	—
1	—	—	100	130	—	—	77	45	90	—
2	—	80	75	90	—	200	45	50	45	—
3	—	60	50	50	—	150	133	37	400	—
4	—	40	55	—	60	160	90	30	300	—
5	—	20	15	—	45	255	60	213	200	120
6	—	—	15	—	30	150	540	150	160	130
7	300	120	15	—	15	95	405	100	45	90
8	225	90	15	—	—	40	450	50	30	50
9	150	60	15	—	—	80	270	180	15	130
year ...	1800	1810	1820	1830	1840	1850	1860	1870	1880	1890
0	90	75	115	100	150	—	—	140	250	85
1	60	80	75	200	75	—	160	70	125	45
2	30	180	250	130	—	40	120	—	—	20
3	220	170	175	80	—	30	80	—	400	15
4	165	170	100	40	—	20	40	—	300	10
5	110	695	70	525	100	10	—	120	240	5
6	55	490	500	450	205	140	—	90	170	—
7	300	375	375	375	140	105	—	60	50	—
8	225	195	250	300	90	70	280	530	170	30
9	150	30	175	225	30	35	210	375	125	25

Appendix III. Diagrams showing the course of various circulation parameters for January and July near northwest Europe after great volcanic eruptions since 1680

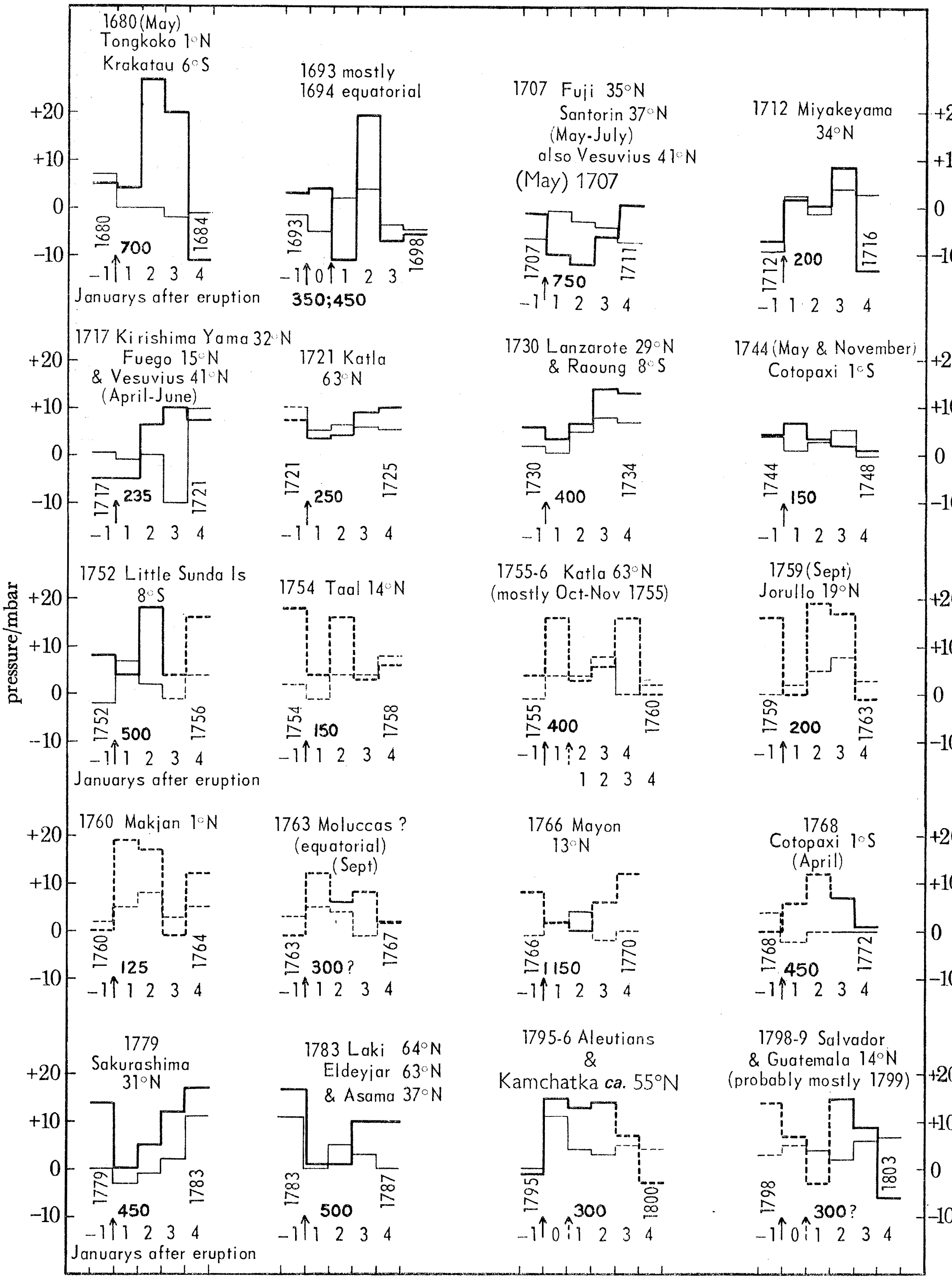

FIGURE 27*a*. January mean pressure differences British Isles–North Sea region in relation to individual volcanic eruptions, d.v.i. values for the northern hemisphere quoted in bold numbers. Crude pressure differences from provisional isobars before 1750. Broken lines indicate portions of the curves complicated by possible overlapping effects of other eruption years. See legend of figure 28 for explanation of curves.

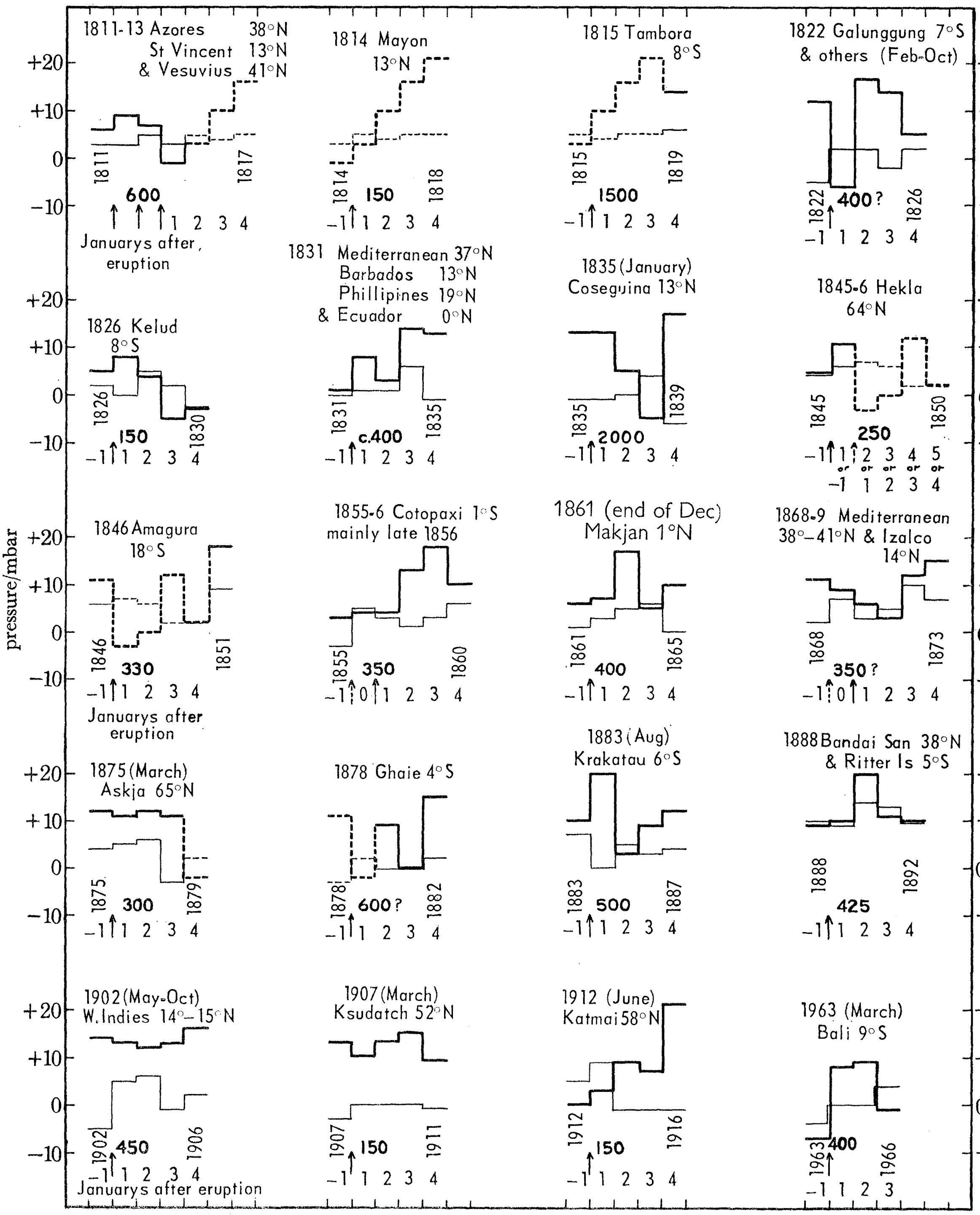

FIGURE 27 *b*. For legend see facing page.

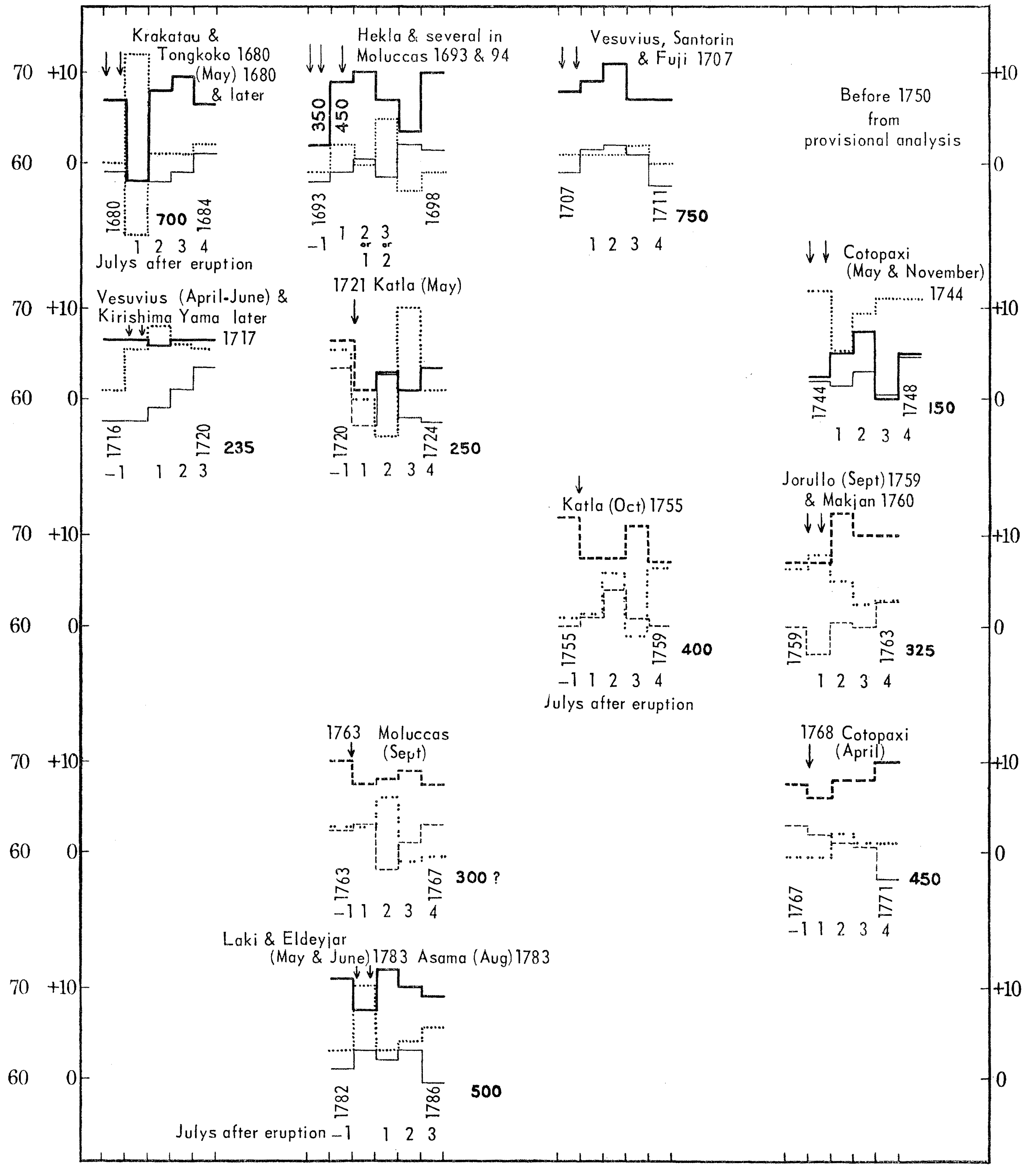

FIGURE 28*a*. July mean pressure differences British Isles–North Sea region in relation to individual volcanic eruptions, d.v.i. values for the northern hemisphere quoted in bold numbers. Crude pressure differences from provisional isobars before 1750. Broken lines indicate portions of the curves complicated by possible overlapping effects of other eruption years. ——, pressure difference 50–60° N, 0° E; ——, pressure difference 10° E – 0°, 55° N; . . ., latitude of lowest pressure at 0° E.

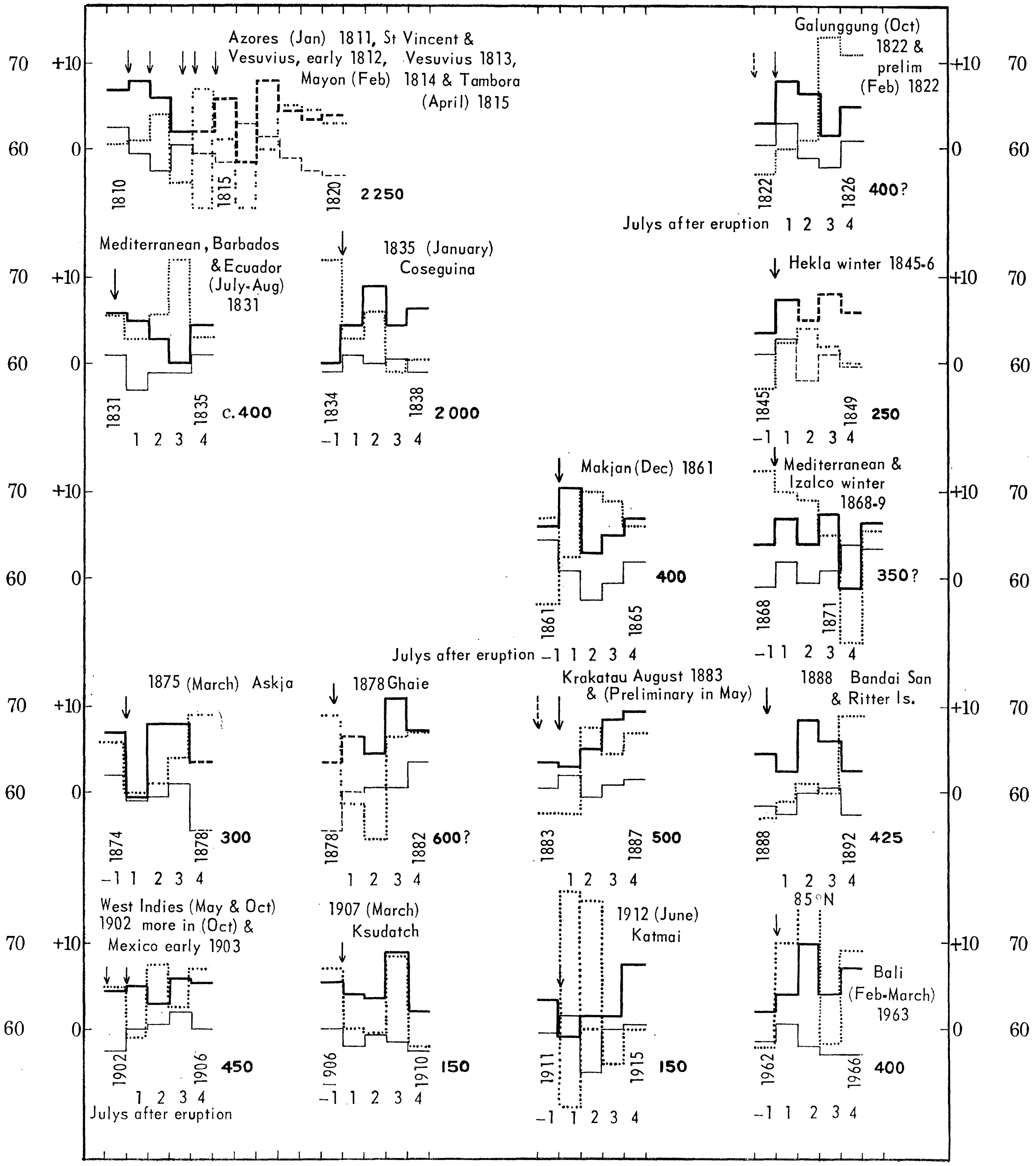

FIGURE 28*b*. For legend see facing page.

Appendix IV. Waves of volcanic activity suggested by Auer

Auer's tephra layers in southern South America
(*Patagonia–Tierra del Fuego*, 40–55° S)

Late-glacial eruption waves

O_1 (Stratum of pumice particles, coarse bubbles of lava) dated by pollen profiles as belonging to the Oldest Dryas period,† about the end of the last glacial maximum, presumably about 15000–14000 B.C.

O_2 (Stratum of pumice particles) dated by pollen profiles as more or less concurrent with, or at the end of, a rather short, very dry climatic phase in Fuego–Patagonia, thought by Auer to correspond to the dry and mild Bölling climatic period† in Europe 12000–11500 B.C.

O_3 (Ash stratum) placed by pollen profiles in the latter part of the Older Dryas,† before the warm Alleröd period; if Auer's assumed equivalence of Patagonian and European dates of the post-glacial climatic sequence be accepted, about 10000 B.C.

O_4 (Ash stratum) placed by pollen profiles at the end of the warm Alleröd period†, about the sharp onset of the cool and damp Younger Dryas, presumably therefore about 9000 B.C.

The late-glacial volcanic strata in Chile are described by Auer (1959, p. 205) as (altogether) several metres thick.

It is noticeable that the pollen profiles in the above-mentioned late-glacial millenia imply non-uniform rates of sedimentation and vegetation growth, both increasing greatly as the ice age waned. Ages of these layers cannot therefore be established by assuming a constant time/depth ratio in the stratigraphy. Radiocarbon dating is needed to establish the suggested dates, which so far mostly rest upon assumed correspondence of events in Patagonia to known dates of similar botanical climatic sequences in Europe and north America.

Post-glacial eruption waves

I (A white rice-like tephra layer, commonly 1 to 3 cm thick) ^{14}C dated‡ about 7000 B.C., thus placed within what is called in Europe the Boreal period† of rapidly increasing warmth: the pollen profiles, however, indicate (Auer 1959, p. 204) a retreat of the forest—hence a dry period—just before this eruption wave and a great increase of the forest immediately afterwards—hence moister conditions—continuing well into the early part

† The sequence of climatic stages recognized in Europe from vegetation history near the Baltic (here referred to) since the last ice age maximum, and the conventional names of the Blytt–Sernander system, with later additions, is summarized by Flint (1957) from 'most probable ^{14}C dates' as follows with approximate dates:

Earliest Dryas tundra stage	13000–12000 B.C.
Bölling warm stage	12000–11500 B.C.
Older Dryas	11500–10500 B.C.
Alleröd warm stage	*ca.* 10000– 9000 B.C.
Younger Dryas	8800– 8200 B.C.
Pre-Boreal	8200– 7500 B.C.
Boreal	7500– 5500 B.C.
Atlantic	5500– 2700 B.C.
Sub-Boreal	2700– 500 B.C.
Sub-Atlantic	since 500 B.C.

This chronology is essentially the same as that given by Godwin (1956, p. 62) for the history of the climate in Britain since 10000 B.C. on botanical and other evidence.

‡ Error margins applicable to the dates in table 7, which are supported so far by a rather small number of radiocarbon age determinations, may be envisaged by 95 % confidence limits about ± 150 to 200 years at 0–3000 B.C. about ± 300 to 400 years at 7000–10000 B.C. (cf. Auer 1958, p. 229).

of the climatic optimum. In so far as the sites used lie east of the main Andes divide, moist climatic periods may imply either high sea temperatures producing high atmospheric moisture content or cyclonicity and occasional easterly winds in the latitudes concerned.

II (A greenish, or grey-yellow, brown tephra 5 to 30 cm thick: this is the coarsest of all the post-glacial tephra layers in Patagonia, chemically resembling the earlier late-glacial eruption layers,) ^{14}C dated‡ about 3500–3000 B.C., placed by pollen profiles and by the end of the period of maximum sea level at the end of the post-glacial climatic optimum,† the latter stages of which had produced dry climates at the sites used. Disturbance of the pollen profiles about the time of the II eruptions indicates greater instability of climate, but another very dry climatic period followed (apparently corresponding to the European 'Sub-Boreal').

III (A white tephra layer commonly 3 to 10 cm thick) ^{14}C dated‡ about 500–200 B.C., thus approximately coincident with the very wet period of the early Sub-Atlantic† climatic recession in NW Europe. More numerous datings of this volcanic activity by reference merely to pollen stratigraphy in Patagonia and Tierra del Fuego suggest it spread over a longer time, possibly from about 900 B.C. to A.D. 100 and the increase of both forest and bog in different parts of that region indicates a generally chillier, moister climate than in the preceding epoch. A progressive reversion to drier vegetation types (increasing steppe-land) over most subsequent time suggests a very gradual recovery towards warmer, drier climates.

IV (A light grey ash stratum.) Found in the top layers of pollen profiles in Patagonia at depths indicating occurrence within the last 400–500 years. This stratum is however less frequently reported than the earlier ones in Auer's studies—perhaps not noteworthy, or even not present, south of about 50° S in South America. Its date is clearly concurrent with the increased activity in Iceland (see figure 20) in recent centuries. Lamb & Johnson (1961, p. 391) have noted that the southern hemisphere does not seem to have fully shared the cold climatic epoch of 1550–1800, though it did so after 1800, and it may be that something similar applied even to the volcanic activity in the two hemispheres.

† and ‡ Apply to footnotes on facing page.