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Northern Hemisphere Tropospheric Mid-Latitude Circulation after Violent Volcanic Eruptions

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Abstract

The strengths of the polar stratospheric vortex and geopotential height anomalies of the 500 hPa layer are studied that are observed after recent violent volcanic eruptions. After all tropical eruptions the polar stratospheric vortex was intensified. The tropospheric anomaly patterns after tropical eruptions are very similar to those of winter months with a very strong stratospheric vortex, irrespective whether volcanically forced or not. Hence, if they have any effect on the wintertime tropospheric circulation, tropical eruptions seem to force a natural mode of the stratospheric winter circulation which is associated with a specific response of the tropospheric circulation with maximum amplitude over the North Atlantic and adjacent continental regions. Therefore, it remains difficult to give statistical evidence of volcanic impact on climate on the basis of the few observations after these rare events. A combination of observational studies and model experiments may help to overcome these difficulties in the future.

Zusammenfassung

Die troposphärische Zirkulation in mittleren Breiten der Nordhemisphäre nach starken Vulkanausbrüchen

Die nach den jüngsten starken Vulkaneruptionen beobachteten Stärken des polaren Stratosphärenwirbels und der Anomalien der geopotentiellen Höhe der 500 hPa Fläche werden untersucht. Nach allen tropischen Eruptionen war der polare stratosphärische Wirbel verstärkt. Die troposphärischen Anomalieverteilungen nach tropischen Eruptionen sind sehr ähnlich zu jenen in Wintermonaten mit einem sehr starken stratosphärischen Wirbel, unabhängig ob dieser vulkanisch angetrieben ist oder nicht. Das bedeutet, daß wenn sie überhaupt einen Effekt auf die winterliche troposphärische Zirkulation haben, tropische Eruptionen einen natürlichen Mode der stratosphärischen Zirkulation anregen, der mit einer spezifischen Reaktion der troposphärischen Zirkulation mit besonders starker Amplitude über dem Nordatlantik und den angrenzenden kontinentalen Gebieten verbunden ist. Deshalb bleibt es schwierig, auf der Grundlage der wenigen Beobachtungen nach diesen seltenen Ereignissen den statistischen Nachweis des Einflusses von Vulkanen auf das Klima zu erbringen. Eine Kombination von Beobachtungsstudien und Modell-Experimenten kann bei der Überwindung dieser Schwierigkeiten in der Zukunft helfen.

Introduction

The recent eruption of the tropical volcano Mt. Pinatubo in June 1991 revived the public interest in the effects of violent volcanic eruptions on climate. Concerning wintertime surface temperature effects, Groisman (1992) and Robock and Mao (1992, RM hereafter) described a stable anomaly pattern after volcanic eruptions. While RM suggested a mechanism according to which heating of the aerosol-containing stratosphere in lower latitudes results in stronger westerlies near the polar circle, Groisman suggested that the westerlies were caused by polar cooling. Following RM, the westwind anomalies "penetrate down" into the troposphere and bring warm and moist air from the Atlantic over continental Eurasia. Here, and over the western part of North America positive temperature anomalies are recorded during winters after violent volcanic eruptions. Temperatures decline over the east of

North America and Greenland due to the advection of Arctic air masses, and in the Northern Hemisphere subtropics due to reduced solar radiation. Such observation is not new (see e.g. Spirina, 1973; Loginov, 1984; Groisman, 1985; Graf, 1986), but it never found much resonance in the literature.

Here we shall investigate the tropospheric geopotential height anomaly fields, observed for the few cases of recent violent volcanic eruptions for which data have been available. We want to examine whether the mechanism of advection of warm moist air to the continents due to "downward penetrating westerly winds", as proposed by RM, worked in any of these cases. In addition, the observed patterns of geopotential anomalies following the volcanic eruptions shall qualitatively be compared with model simulations.

The process of volcanic aerosol forcing was modelled in a perpetual January General Circulation Model experiment (Graf et al., 1992) including long- and short-wave radiative effects of the aerosol. In these simulations, differential stratospheric heating due to the absorption of thermal and solar radiation by the stratospheric aerosol led to significant temperature anomaly patterns in the lower troposphere of the Northern Hemisphere. These pattern were very close to those found by RM and comparable to those observed during the winters after the Mt. Pinatubo eruption. The mechanism can be explained with linear theory (e.g. Geller and Alpert, 1980; Schmitz and Grieger, 1980). In case of a strong polar vortex, vertically propagating tropospheric planetary waves are trapped more effectively in the troposphere than in periods with a weak vortex. Other numerical simulations with a similar forcing type (i.e. strong versus weak polar night jet) have the same effects (e.g. Boville, 1984) on the planetary wave patterns.

Following RM, we used those volcanoes as key events which had either a Dust Veil Index (DVI) of at least 250 or a Volcanic Explosivity Index (VEI) larger than 4. This leaves only six cases for which data are available (see Table 1). Observational data cover the period from January 1957 (for the 500 hPa level) and from January 1963 (for the other levels) through December 1989.

We use monthly means of the geopotential heights of lower tropospheric pressure layers (850, 700 and 500 hPa) north of 20° N. We calculated anomalies of the geopotential heights relative to the 1982–1989 mean, since for the case of the winter of 1991/92 we only have anomalies for this reference period. The data are based on the analyses of the National Meteorological Center of the U.S. Weather Service and were transformed to a $5° \times 5°$ grid at Max-Planck-Institut für Meteorologie. The stratospheric data are monthly means of the height of the 50 hPa layer on a $10° \times 10°$ grid. We obtained these data from the Stratosphere Group of the Meteorological Institute of the Free University Berlin.

Four of the eruptions that we considered in our study are tropical ones, and only two (Bezymianny and St. Helens) exploded in higher northern latitudes. RM studied Bezymianny's climate impact not before the winter of 1957/58, arguing that aerosol from a high latitude source is transported but slowly into lower latitudes, where it can absorb solar radiation and, thus, increase the meridional temperature gradient during winter. St. Helens was not considered by RM because of the low SO_2 concentration in the eruption plume. However, according to McCormick and Trepte (1987), an increased optical depth with a maximum of $\tau = 0.06$ was observed over the North Pole during the year after this eruption. In addition, with its VEI of 5 and the DVI of 500, it complies with the rule of RM and was therefore also considered in our study as an example of high latitude eruptions. Otherwise there would be no reason to include the 1956 Bezymianny eruption in this study. For this eruption the relevant data are simply unknown, and the DVI is very small (see Table 1).

Table 1 Volcanic eruptions considered in this study. DVI and VEI are taken from Tobock and Mao (1992), SO_2 data are from different sources ([1]Millan et al., 1985; [2]Cadle, 1980; [3]Rose et al., 1982; [4]Bluth et al., 1991).

Volcano	Latitude	Longitude	Date of Eruption		DVI	VEI	SO_2-Emission
Bezymianny	56° N	161° E	March 30	1956	30	5	unknown
Agung	8° N	116° E	March 17	1963	800	4	12,0 Mt[2]
Fuego	14° N	91° E	October 10	1974	250	4	3,1 Mt[3]
St. Helens	46° N	122° E	May 18	1980	500	5	2,1 Mt[1]
El Chichón	17° N	93° E	April 3	1982	800	5	7,0 Mt[4]
Mt. Pinatubo	15° N	120° E	June 15	1991	1000	6	20,0 Mt[4]

Another aspect might be of interest concerning high latitude volcanos: The aerosol built up from such eruptions is believed to remain in the stratosphere over the polar cap for a long time. This was observed, for instance, after the St. Helens eruption in 1980 (McCormick and Trepte, 1987). If most of the aerosol remains over the pole, it may there absorb thermal radiation. Thus, it will warm the polar stratosphere, leading to a weaker polar vortex, i.e. to an effect opposite to the one caused by tropical volcanos. Therefore, we also included in our analysis the first winters after these eruptions.

Results

We examined the height anomalies of the 850, 700 and 500 hPa pressure levels observed during the winters after strong volcanic eruptions. All anomaly patterns found are similar throughout the troposphere, suggesting a barotropic response. The amplitudes of the anomalies increase with height. Here, we shall discuss the 500 hPa level results only, since the data set is longest for this level; it also contains some of the possible effect of the 1956 Bezymianny eruption. As a reference period we took the winters from 1982 to 1989 because this is also the reference period for the 1991/92 winter for which only anomalies were available.

The observed height anomalies of the 500 hPa level are shown in Figure 1 for the winters that immediately followed the tropical eruptions of Agung, El Chichón and Pinatubo. For the extratropical eruptions we studied the winters one year after the eruptions. We decided to use the second winter after the Fuego eruption, as that eruption was late in 1974, and normally it takes several months before the initially gaseous SO_2 is converted into sulphate aerosol (which interacts with radiation) and distributed quasihomogeneously in the tropics. Moreover, the stratosphere responds but slowly to diabatic heating anomalies produced by additional aerosol. The pattern of the geopotential anomalies for the winter of 1974/75 (not shown here) has the same basic features as the one for the winter of 1975/76, but with smaller amplitude.

The 500 hPa height anomalies observed during the winters immediately after the tropical eruptions are very similar over the Atlantic and Eurasian regions, while those anomalies differ greatly between the four winters over the North Pacific and the western half of North America. Table 2 contains the pattern correlation coefficients (PCC) between the 500 hPa geopotential height anomaly fields of the four winters displayed in Figure 1a. To the right of the matrix diagonal the PCC are given for the 90° W to 145° E sector, i.e. for the Atlantic and Eurasian area, and to the left of the matrix diagonal the North Pacific and western North America (145° E to 90 °W) area is used. In order to account for the natural variability of planetary wave patterns we computed the PCCs allowing the patterns to shift up to 10° in latitude and longitude. The maximum PCCs ("optimal PPC") are then given in Table 2.

Table 2 Pattern correlation coefficients (area weighted) between the anomaly patterns shown in Figure 1a. To the right of the matrix diagonal the area 90° W to 145° E, and to the left the sector 145° E to 90° W is used.

	63/64	75/76	82/83	91/92
63/64	1.0	0.52	0.30	0.65
75/76	0.62	1.0	0.76	0.75
82/83	– 0.56	– 0.70	1.0	0.48
91/92	0.47	0.37	0.39	1.0

The differences found in the PCCs for the different regions can readily be explained by the impact of El Niño on the geopotential field in the mid-latitudes of the Northern Hemisphere: During the very strong El Niño of 1982/83 the Aleutian low was deep, while in the winter of 1975/76, when equatorial Pacific sea surface temperatures were low, an anticyclonic anomaly developed in this region. The two other winters had weak (1963/64) or moderate (1991/92) positive sea surface temperature anomalies in the El Niño region and, therefore, negative and weak positive PCC, respectively. Thus the differences in the pattern correlations between the single cases of tropical volcanic eruptions over the North Pacific and western North America are basically due to the impact of El Niño on the observed geopotential fields. For the Atlantic and Eurasian region the El Niño influence is much smaller, and therefore the anomaly patterns are more similar for the single cases. Again, the PCC are weakest for any combination with the "Super El Niño"-winter of 1982/83.

The anomaly patterns over the North Atlantic and Eurasia agree well with perpetual January simulations (Figure 2, right panel) of the climatic impact of tropical volcanos (Graf et al., 1992). These patterns are characterized by a low geopotential of the 500 hPa layer in very high latitudes with a trough over Greenland during the winters after the Agung, Fuego, El Chichón and Pinatubo eruptions. Over

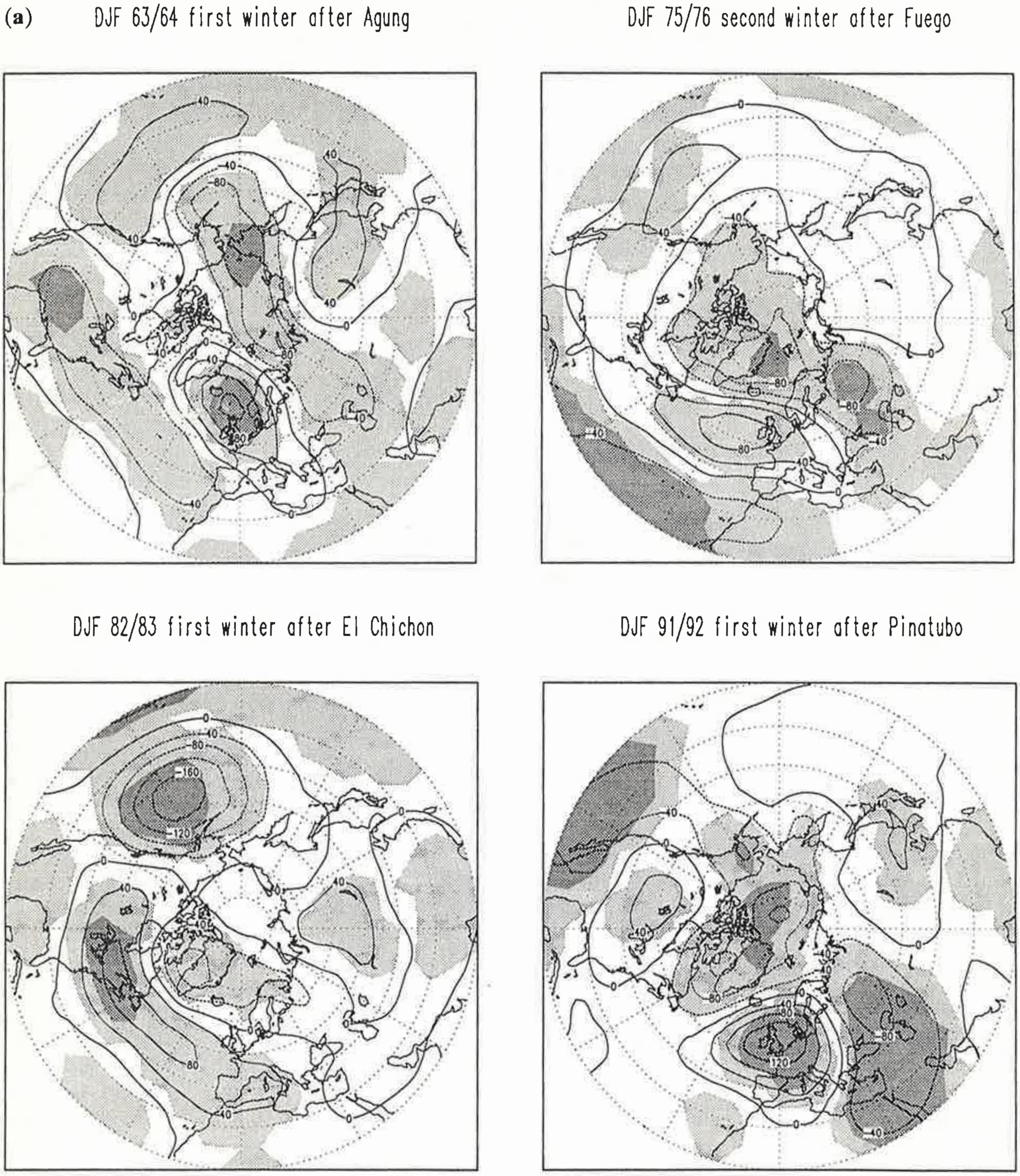
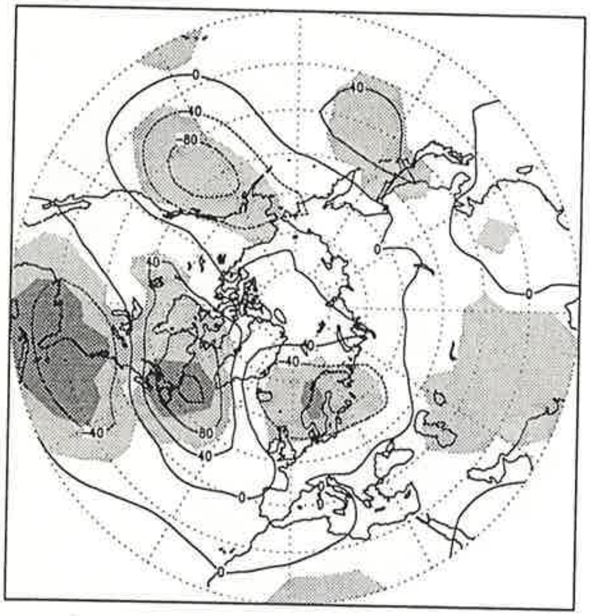
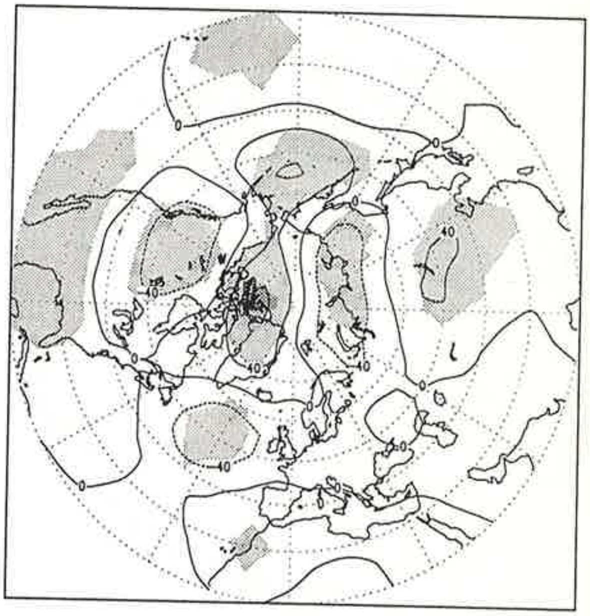


Figure 1 Anomalies (mean 1982–89) of the height of the 500 hPa layer during winter (DJF) after volcanic eruptions. The heavy line indicates zero anomaly.
(a) First winter after tropical eruptions (for Fuego eruption second winter)
(b) Second winter after Northern Hemisphere higher latitude eruptions.
(c) First winter after Northern Hemisphere higher latitude eruptions.
Light (heavy) shading indicates anomalies exceeding ± 1 (± 2) standarddeviation(s).

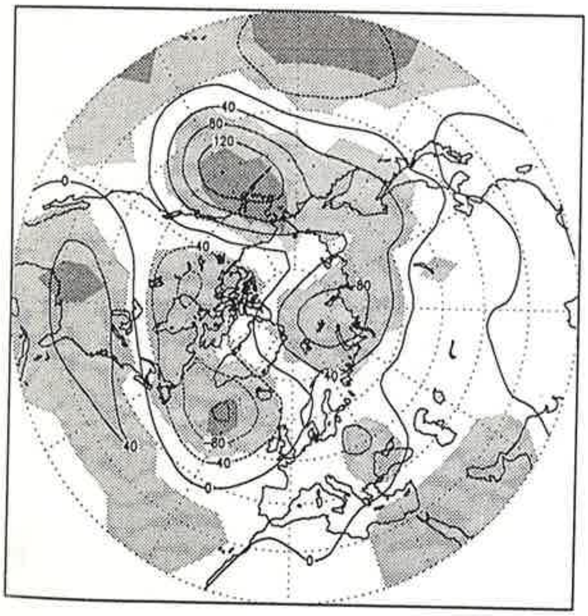
(b) DJF 57/58 second winter after Bezymianny



DJF 81/82 second winter after St. Helens



(c) DJF 56/57 first winter after Bezymianny



DJF 80/81 first winter after St. Helens

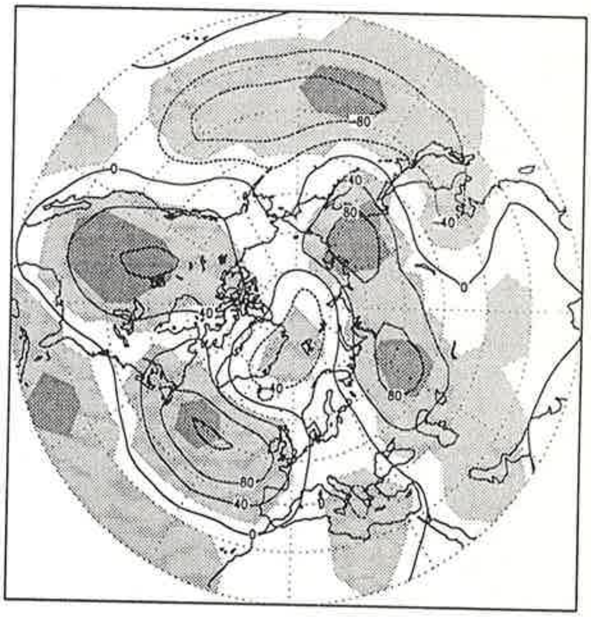


Figure 1 (continued)

the eastern North Atlantic and West Europe, an anticyclonic anomaly is prominent in all cases. This anomaly extended all over the North Atlantic to the eastern part of North America in the winter of 1982/83. This is in good agreement with model calculations of Kirchner and Graf (1993) who conducted GCM experiments combining an El Chichón type aerosol with the observed SSTA of the El Niño winter of 1982/83 (Figure 2, left panel). The Fuego cloud is by far the weakest of the tropical eruptions under consideration. It was selected by RM because it fits their rule of DVI >= 250 or VEI >= 4. An inspection of observed stratospheric temperatures (Figure 3) shows that the tropical temperature at

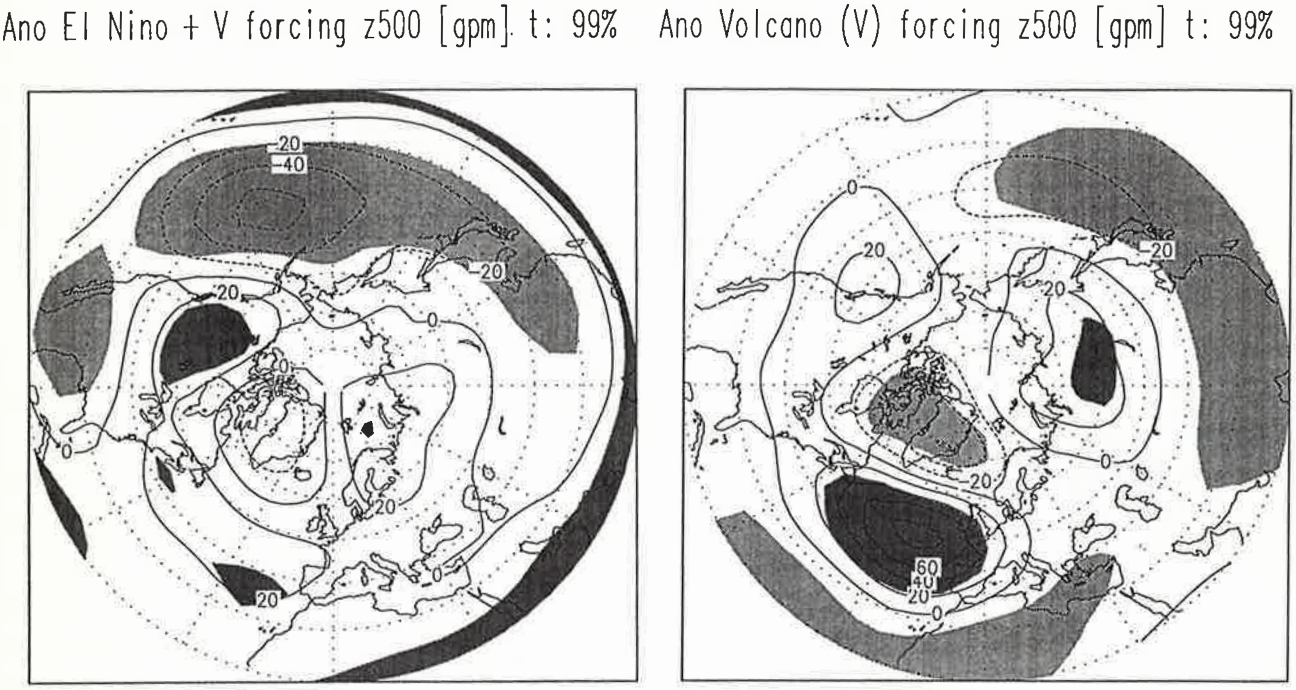


Figure 2 Mean anomalies of the height of the 500 hPa level over the Northern Hemisphere for 58 permanent January simulations with El Niño and volcanic forcing (left panel, Kirchner and Graf, 1993), and with volcanic forcing only (right panel, Graf et al. 1992). The shaded areas indicate anomalies exceeding the 99 % confidence level for a local t-test.

the 30 hPa level was slightly above normal, and the North Pole was very cold. Thus the meridional temperature gradient was enhanced producing a stronger than normal polar night vortex (Figure 4). In this case the proposed dynamic mechanism can work independent of whether or not it is volcanically forced.

The 500 hPa layer height anomalies of the second winter following the northern hemispheric mid-latitude eruptions (1957/58 and 1981/82) are completely different from those for the tropical eruptions (Figure 1b). The same holds true for the winters immediately after the high-latitude eruptions (Figure 1c) with only one exception: the winter of 1980/81. This result either suggests an aerosol behaviour different from the one anticipated by RM (i.e. a stronger tendency of the aerosol to remain in high latitudes) or it is due to a very weak aerosol. Observations of an increase of the optical depth at the North Pole from 0.0001 in 1979 to 0.06 after St. Helens (McCormick and Trepte 1987), besides the obviously weak aerosol, support also the first view. If this is true, it will be necessary to reexamine the climatic effects of violent historic eruptions at least for the cold season. By doing so, strong differences in the climate effect for the cold season can be expected between tropical and high-latitude eruptions. This, on a regional scale, was reported also by Lough and Fritts (1987).

Discussion

Winters following the few recent violent tropical volcanic eruptions have had stronger than normal polar stratospheric vortices. The observations of the mid-tropospheric geopotential field after those eruptions do not contradict the hypothesis of RM that tropical volcanoes have a significant effect on wintertime tropospheric circulation over the Atlantic and Eurasian regions. However, a statistical prove of this relationship based on observations can not be given because of the small number of events. After strong eruptions in high northern latitudes a different geopotential anomaly pattern develops, which is not consistent with RM's hypothesis. The positive temperature anomalies in middle and high northern latitudes found by RM also after high latitude eruptions therefore require another explanation than tropospheric advection of maritime air due to primary stratospheric dynamic forcing.

It is suggested by GCM experiments (Boville, 1984; Graf et al., 1992) that the basic mechanism underlying the response of tropospheric circulation to tropical volcanic forcing is a strengthened stratospheric polar winter vortex resulting from enhanced differential heating by the aerosol-containing layers. In conformity with linear theory (Geller and Alpert, 1980; Schmitz and Grieger, 1980), the change of the tropospheric planetary wave structure

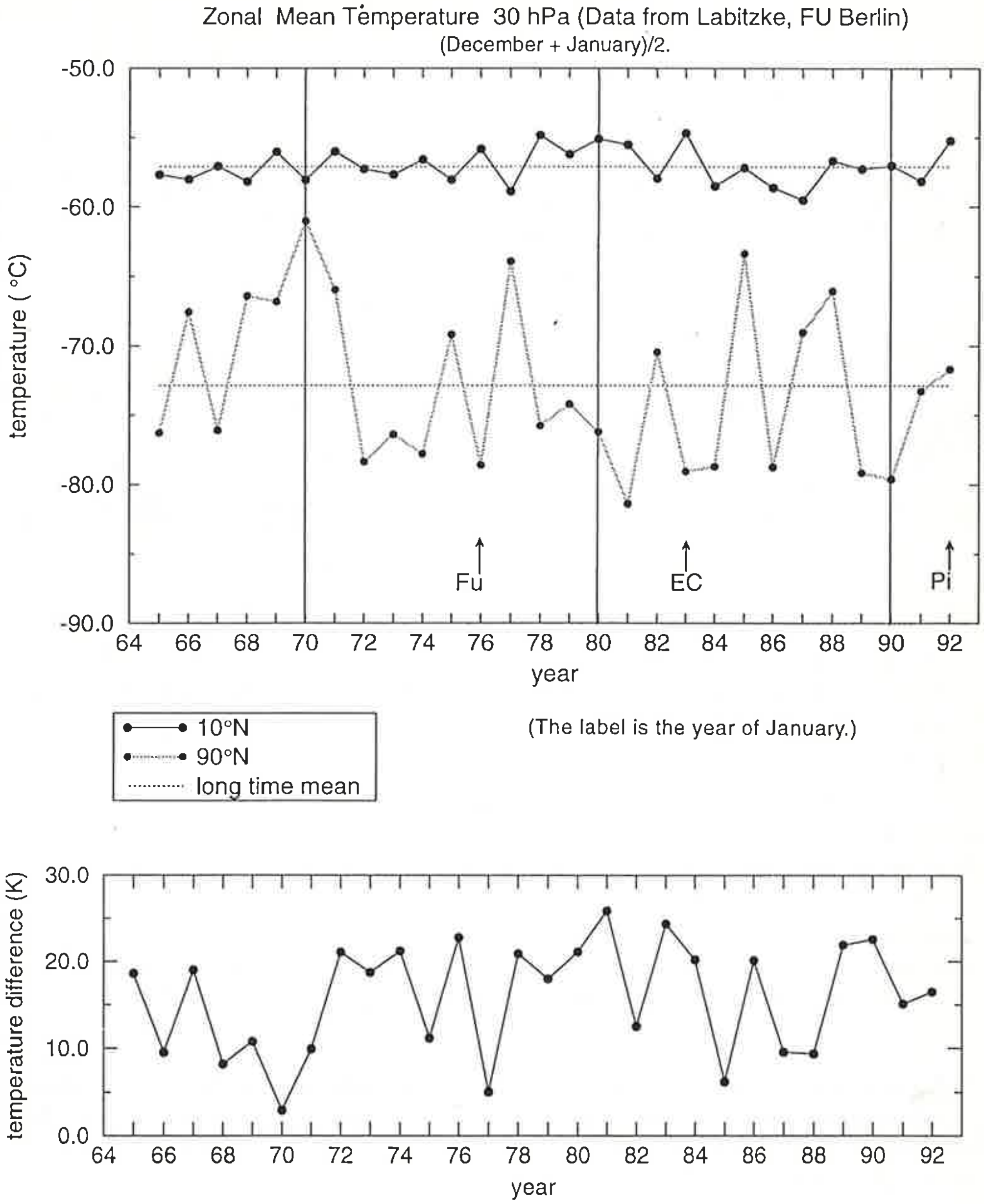


Figure 3 (a) Zonal mean temperature at the 30 hPa level at 10° N and over the North Pole, December/January mean. The tropical eruptions covered by the data are indicated by arrows (Fu = Fuego, EC = El Chichón, Pi = Pinatubo). The Fuego eruption is expected to have influenced the winter 1975/76 because of its late eruption date in 1974 (see text). The abscissa labels indicate the year of the January.
(b) Temperature difference 10° N minus North Pole at the 30 hPa level, December/January mean. The abscissa labels indicate the year of the January.

is due to the trapping of planetary wave energy in the troposphere as a result of the strengthened polar night jet. Since changes in the strength of the stratospheric polar winter vortex may occur also without volcanic forcing, volcanos may just stimulate a natural mode of the stratospheric winter circulation.

In order to study the relationship between stratospheric and tropospheric circulation, we can easily increase the number of cases in our investigation just by comparing the winters with strong polar vortex to those with a weak one.

To select these winter months, we made an Empirical Orthogonal Function (EOF) analysis of the

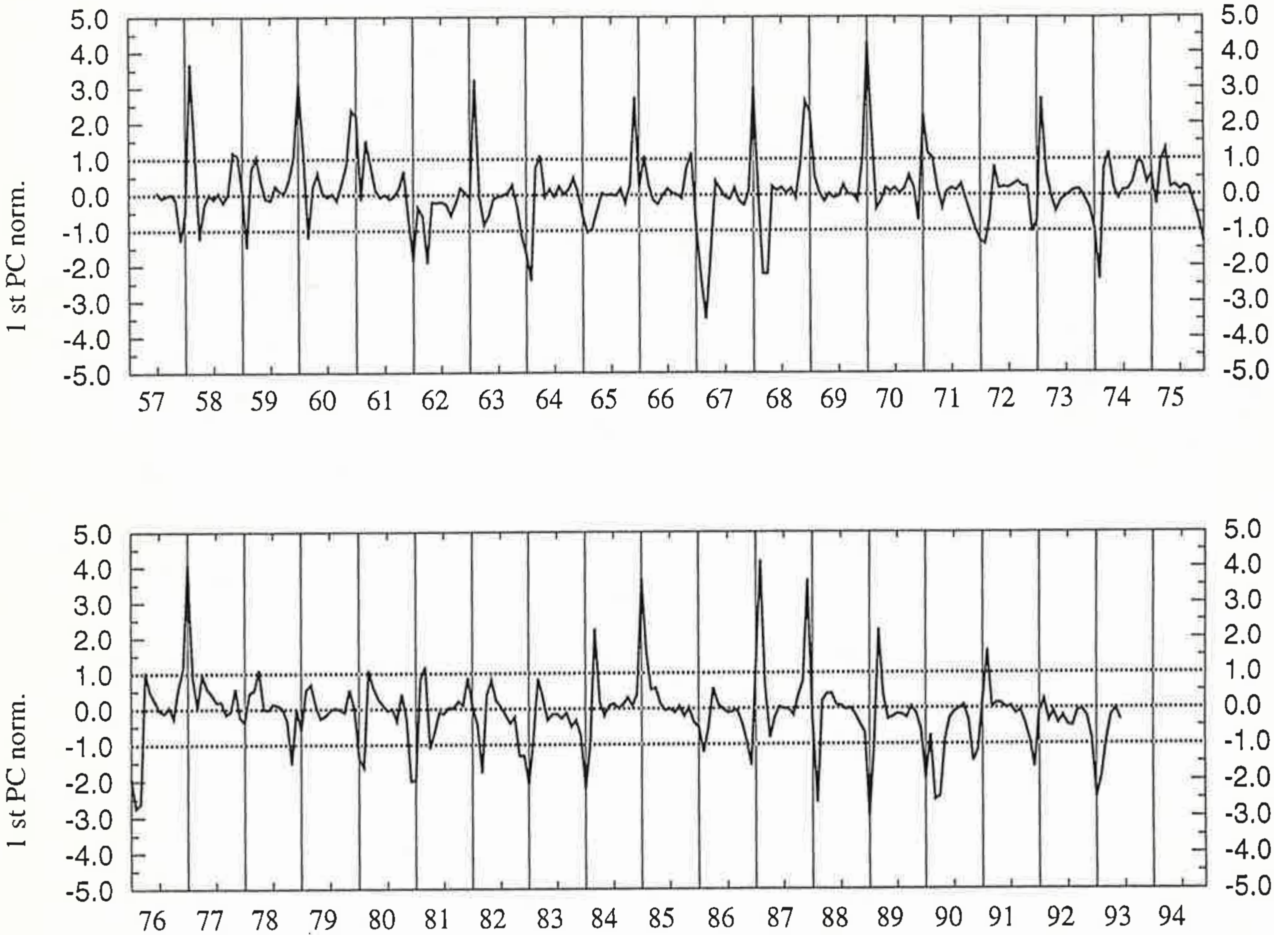
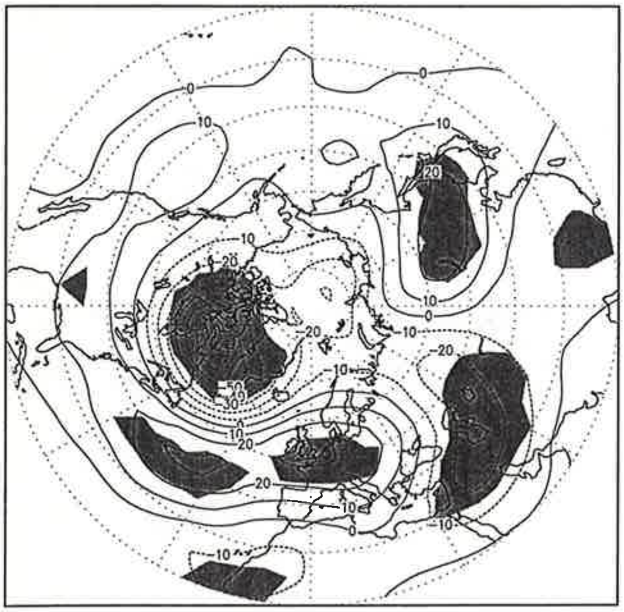


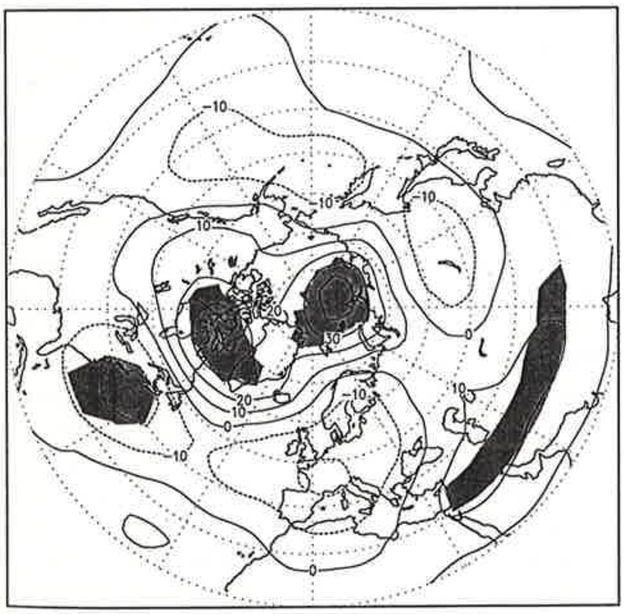
Figure 4 Values of the first Principal Component of the 50 hPa layer geopotential for all months July 1957 through June 1993, normalized to the standard deviation. Negative values denote a stronger polar vortex and positive values indicate a weak vortex. Note that during summer months this mode practically does not exist. Vertical lines denote January.

height of the 50 hPa layer. The first EOF (explaining 52.4 % of the total variance (area weighted) for winter months (DJF), and 44 % for all months after extraction of the mean annual cycle) describes variations in the polar vortex. Its Principal Component (PC) is highly correlated (r = 0.97) with the anomalies in the zonal mean geostrophic wind at the polar circle. Starting out from the time series of the first PC, normalized by its standard deviation (Figure 4), we chose those months with values larger than ± 1 standard deviation. Most of these strong anomalies occur during the winter months of December through February, only few are found in the transitional seasons and none during the period between May and September. We found 46 winter months with a very strong, and 41 months with a weak stratospheric vortex between 1957 and 1993. The composite anomaly patterns for the 500 hPa geopotential and for the surface air temperature for months with strong and weak polar vortex, and for the difference of the two are shown in Figures 5 and 6 respectively. In these Figures the statistical significance (99.5 % confidence level) of the amplitudes is indicated by shading. The composite anomalies of the 500 hPa layer height from the 1957–1989 mean of the 38 months with a strong polar stratospheric vortex represent a wave structure (Figure 5a), with troughs over North-East America and the Northwest Atlantic, over the Ural mountain range and, fairly weak, over the Bering Strait. Ridges are found over the East Atlantic and West Europe, and over East Siberia. A pattern that is just the opposite in many respects, but with smaller amplitudes, is found for the 41 months with a weak stratospheric vortex (Figure 5b), with the strongest signal over the western North Atlantic. The difference pattern (strong minus weak vortex, Figure 5c) indicates a deep polar low centered over the Davis Strait and an anticyclonic anomaly over the North-East Atlantic resulting in enhanced westerlies over the North

(a) z500 57/7–89/6 (strong vortex – all months)



(b) z500 57/7–89/6 (weak vortex – all months)



(c) z500 57/7–89/6 (strong – weak vortex)

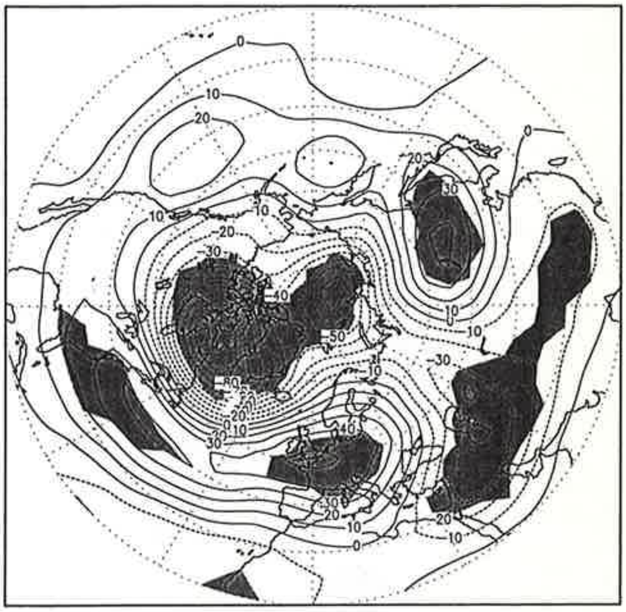
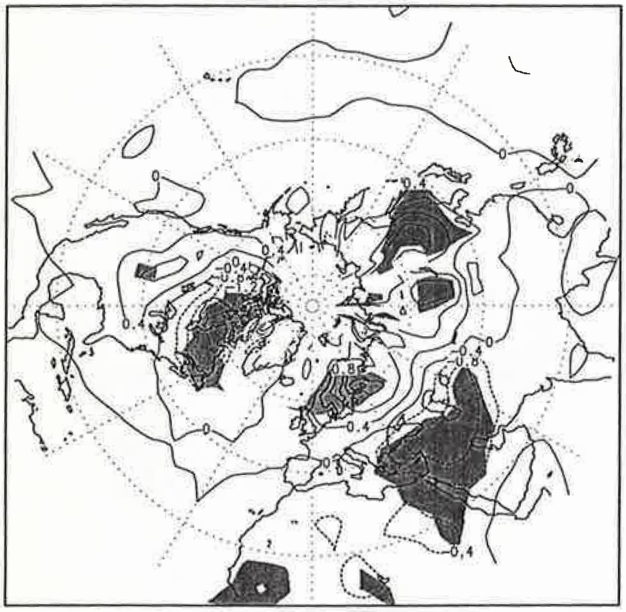


Figure 5 Mean Northern Hemisphere anomaly (mean 1957–1989) of the height of the 500 hPa layer for 38 month with strong (5a), 41 months with weak stratospheric polar vortex (5b) and difference strong minus weak vortex (5c). Shading indicates statistically significant amplitudes (local t-Test, 99.5 % confidence level).

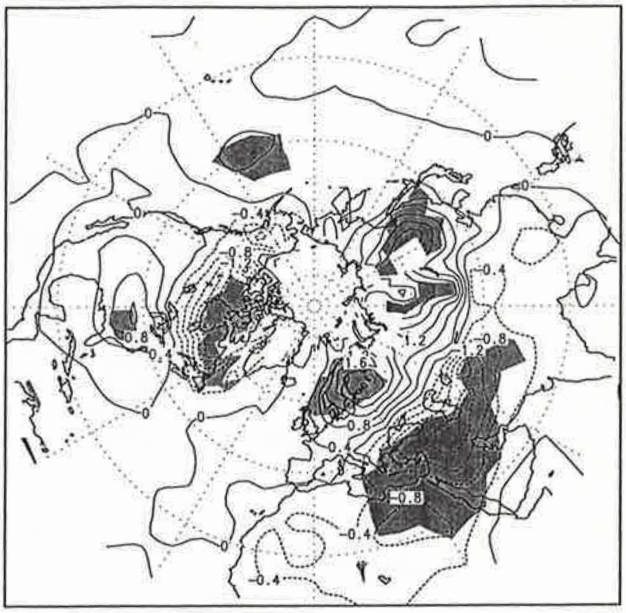
Atlantic. Troughs develop east of the Rocky Mountains and over Central Russia. This observed pattern is consistent with the mean signal simulated with a GCM for the Pinatubo aerosol in the winter of 1991/92 (Figure 2, right panel). This is indicated by a pattern correlation coefficient, area weighted, of r = 0.65 for the Northern Hemisphere north of 20° N.

The anomaly pattern of the surface air temperature (Jones et al., 1988) for months with a strong (44 months) polar vortex shows a very clear positive temperature anomaly over middle and high latitudes of Eurasia (mainly Northern Europe and East Siberia), while negative temperature anomalies prevail over North-East America and over Arabia and North Africa (Figure 6a). This pattern is similar to RM's surface air temperature anomalies for the 12 strongest volcanic eruptions of the past century, and also similar to the GCM results of Graf et al. (1992). Again for the 42 months with weak polar vortex (Figure 6b) a similar anomaly pattern with opposite sign develops, leading to quite substantial and statistically significant temperature differences (Figure 6c) between months with a large negative First Principal Component (strong stratospheric polar vortex) and a strong positive one (weak vortex). The patterns in Figures 5 and 6 are not dominated by the winter months after strong volcanic eruptions. This was tested by excluding these months (not shown).

(**a**) Tsfc 57/8–92/7 (strong vortex – all months)



(**c**) Tsfc 57/8–92/7 (strong – weak vortex)



(**b**) Tsfc 57/8–92/7 (weak vortex – all months)

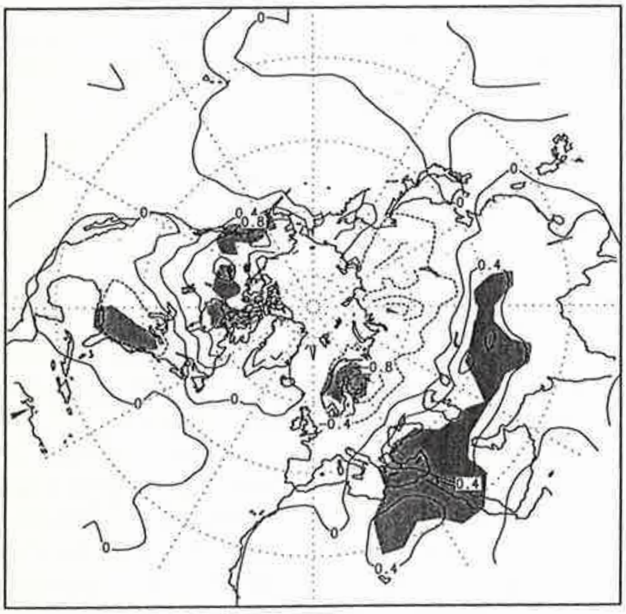


Figure 6 Surface air temperature for 44 months with strong (6a), and 42 months with weak (6b) stratospheric polar vortex (see Figure 3) and difference of strong minus weak vortex (6c). Shading indicates statistically significant amplitudes (local t-Test, 99.5 % confidence level).

Conclusions

Though it is impossible to give statistical prove because of the limited availability of free atmosphere data and the rareness of events, the mid-tropospheric geopotential anomaly patterns over the Northern Hemisphere observed after violent tropical volcanic eruptions do not contradict the hypothetical mechanism suggested by RM to explain the observed temperature pattern after violent tropical volcanic eruptions. The analyzed tropospheric circulation anomalies may explain the surface air temperature anomalies in middle and higher latitudes as found by RM and others. The response is less clear and may even have opposite sign for eruptions in high latitudes.

The possible mechanism leading to stronger westerlies over the North Atlantic in winter and to positive temperature anomalies over large parts of Eurasia becomes evident when comparing GCM experiments with observations. It consists in exciting a natural mode of the stratospheric winter circulation. Differential heating between lower and higher latitudes by the aerosol-containing stratospheric layers increases the meridional temperature gradient and strengthens the polar stratospheric winter vortex. Stratospheric circulation affects the tropospheric circulation via vertical wave propagation. These mechanisms have been imposed already more than ten years ago with simple barotropic models (Schmitz and Grieger, 1980; Geller and Alpert, 1980).

The anomalous surface air temperature patterns for winter months with strong polar stratospheric vortex for the years from 1957 to 1993 are nearly the same as those analyzed by RM for the 12 strongest volcanic eruptions of the last century. This pattern is not dominated by the volcanically disturbed months. If they have any effect on the tropospheric winter circulation in middle and high northern latitudes, tropical volcanic eruptions enhance an inherent natural mode of the northern hemispheric winter circulation. Therefore, it is generally difficult to prove the statistical significance of any observed volcanic impact on global climate in winter. The type of tropospheric response is the same for all winter months with a strong polar vortex, irrespective of whether or not an anomalous aerosol concentration was observed. Further, since volcanic eruptions are rare events the number of cases is too small to apply statistical tests. However, we can make use of GCM experiments (see e.g. Boville, 1984; Rind et al., 1992; Graf et al., 1992) and study the stability of the tropospheric response found there. Such simulations support the idea that the correlations between the strength of the polar stratospheric vortex and the tropospheric geopotential field, or the surface air temperature, are due to the stratosphere forcing the troposphere, and not vice versa.

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