Mount Agung Eruption Provides Test of a Global Climatic Perturbation

Abstract. *The Mount Agung volcanic eruption in 1963 provides the best-documented global radiative perturbation to the earth's atmosphere currently available. Data on stratospheric aerosols produced by this eruption have been used as input to a model for the atmospheric thermal structure. The computed magnitude, sign, and phase lag of the temperature changes in both the stratosphere and the troposphere are in good agreement with observations, providing evidence that the climatic response to a global radiative perturbation is significant, as well as support for the use of theoretical models to predict climatic effects.*

Radiative perturbations of global climate represent a very broad class of mechanisms for both natural and anthropogenic climatic change. Examples are changes in the composition of atmospheric gases, such as CO_2, water vapor, ozone, Freons, and other trace constituents; changes in the atmospheric aerosol content, such as by volcanic emissions or anthropogenic pollution; changes in the surface albedo of the earth; and changes in the flux of incoming solar radiation.

A large volcanic eruption provides an excellent opportunity for a case study of the response of the climatic system to a global radiative perturbation, including a chance to test our ability to model and understand the nature of the climatic response. Volcanic aerosols, spread globally by stratospheric winds, provide a perturbation for which the forcing function is reasonably well known and for which the climatic response should be rapid enough and strong enough to be measurable, at least in the cases of the largest eruptions.

The eruption of Mount Agung on the island of Bali (8°S, 115°E) in 1963 was one of the most spectacular volcanic eruptions in recent history (*1*). The eruption column extended into the stratosphere, where it deposited aerosols and gases, including sulfur-bearing compounds. The latter are believed to lead to the formation of sulfate aerosols, a major component of the stratospheric aerosols, through gas-to-particle chemical reactions (*2, 3*). Although there have been larger volcanic eruptions than Agung during the last few centuries—for example, Tambora in 1815 and Krakatoa in 1883 (*4*)—Agung occurred at a time when tropospheric temperature measurements were being obtained, accurate measurements of the aerosol optical depth were being made at several observatories in both hemispheres, and there was direct sampling of the composition of the stratospheric aerosols. Thus, even though there have already been numerous studies concerning the possible effect of volcanic eruptions on climate (*4–7*), it is worthwhile to compute in detail the expected temperature response of the atmosphere after the Agung event.

The global spread of the aerosols due to Agung could be tracked by means of their optical effects. As a result of the prevailing stratospheric winds the aerosols girdled the globe east to west within a few weeks, and within several months they had spread to all latitudes (*8, 9*). The most reliable estimate of the aerosol optical depth is provided by the increased extinction of starlight measured at astronomical observatories (*10*), which indicated that in the subtropics and mid-latitudes of the Southern Hemisphere the added aerosol optical thickness (*11*) for wavelengths ~ 0.5 μm had a peak value of 0.2 to 0.3 from August to November 1963 and was still substantial (~ 0.1) by mid-1964. Thus to test the impact of the aerosol on atmospheric temperatures we chose as a representative increase in the optical depth

$$\Delta\tau = 0.2 f(t) \qquad (1)$$

$$f(t) = \begin{cases} t/120 & 0 < t < 120 \\ 1 & 120 < t < 240 \\ e^{-(t-240)/240} & 240 < t \end{cases}$$

where t is the time in days after the eruption (*12*).

As a first approximation for the expected thermal response of the atmosphere to the increased amount of aerosol, we computed the vertical temperature profile with a one-dimensional radiative-convective model. The basis of the method is to compute the local radiative heating and cooling rates for solar and thermal radiation at each altitude to determine the changing temperature profile with a time-marching procedure (*13*). At any altitude where the computed temperature lapse rate is steeper than a preassigned maximum value (−6.5°K per kilometer), it is assumed that convection occurs with a vertical energy flux just sufficient to yield that preassigned maximum lapse rate. The relative humidity is kept fixed (*14*). Averaging over clear and cloudy regions is performed at each time step before computing the energy balance; climatological percentages of high, medium, and low clouds are employed, and it is assumed that the cloud-top altitude remains fixed (*14*). A thermal inertia appropriate for the upper ("mixed") layer of the ocean is included at the surface.

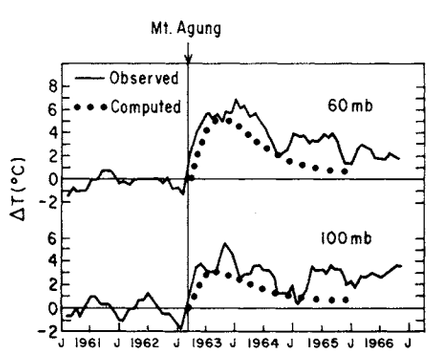

Fig. 1. Observed stratospheric temperatures over Australia (*17*) and computed temperatures after the eruption of Mount Agung, assuming that the added stratospheric aerosols are sulfuric acid and the average depth of the mixed layer of the ocean is 70 m.

This one-dimensional model was used to compute the expected temperature effect at low latitudes, where the amount of perturbed aerosol was largest. The atmospheric temperature profile was computed for a "control" case with the normal aerosol model proposed by Toon and Pollack (*15*), which has sea salt, soil (basalt), and sulfate particles in the troposphere and sulfate (75 percent H_2SO_4 and 25 percent H_2O in solution) in the stratosphere with a stratospheric aerosol optical depth of 0.005. Aerosols with an optical depth specified by Eq. 1 were then added to this background model, the vertical distribution of the additional aerosols being 20 percent in the region from 22 to 26 km, 60 percent from 19 to 22 km, and 20 percent from 16 to 19 km, in order to approximate the distribution of the Agung aerosols (*8*). The added aerosols were assumed to be sulfuric acid, except as indicated below. The size distribution of the added aerosols was taken to be the same as that of the normal stratospheric aerosols, with the effective radius of the distribution 0.2 to 0.3 μm; initially a substantial amount of larger "dust" particles was present, but in the several weeks required to produce aerosols over a large fraction of the globe the mean size decreased to a few tenths of a micrometer (*16*). North-south horizontal transport of energy by atmosphere and oceans was included by using specified climatological values. The solar insolation employed was the mean for low latitudes (30°N to 30°S). The heat capacity of the surface of the model was taken as that of 70 m of ocean water, cor-

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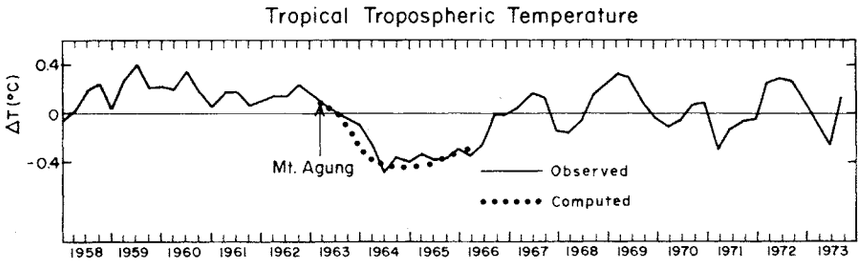

Fig. 2. Observed tropospheric temperatures between 30°N and 30°S (*19*) and computed temperatures after the eruption of Mount Agung, assuming that the added stratospheric aerosols are sulfuric acid and the average depth of the mixed layer of the ocean is 70 m.

responding to the average depth of the mixed layer, for the fraction of that latitude zone covered by water and zero for the remaining fraction.

Figure 1 shows the computed and observed temperatures in the stratosphere. The observations (*17*) were made over Port Hedland, Australia, and are smoothed with a 3-month running mean; at altitudes near 20 km (pressures ~50 mbar) the amplitude of the temperature increase was 4° to 8°C throughout the region from 10°N to 30°S. Part of the observed temperature increase may be due to the quasi-biennial oscillation, but it seems clear that the aerosols did result in stratospheric warming of a few degrees with a time scale for the increase of the order of several weeks (*18*).

Figure 2 shows the computed and observed (*19*) temperatures in the troposphere for the latitude range 30°N to 30°S. A more recent analysis (*20*) of the observations is in substantial agreement with the illustrated observations. It appears that after the Agung eruption the average tropospheric temperatures did indeed decrease by a few tenths of a degree with a time scale of the order of 1 year, in agreement with the theoretical result (*21*).

The physical explanation for the computed effects is straightforward. The overall effect of the added aerosols on the bulk of the atmosphere and the surface is cooling, because sulfuric acid is highly reflective to solar radiation and thus tends to decrease the amount of solar radiation absorbed by the earth-atmosphere system. The aerosols also interact with the thermal radiation, and in fact warm the earth-atmosphere system through a "greenhouse" blocking of radiation from the surface and troposphere; however, aerosols of the assumed size have too small an optical thickness in the infrared for the greenhouse effect to exceed the albedo effect (*7*). The effect of the aerosols on the local (stratospheric) temperature is heating, because (i) they absorb thermal radiation from the warmer atmosphere and surface below more effectively than they release heat to space, and (ii) they absorb a small amount of solar radiation because they have broad absorption bands in the near-infrared. A small amount of energy is involved in the stratospheric heating, but the local temperature is substantially influenced because of the low density of the atmosphere in that region.

It is therefore clear that the thermal effect of the aerosols must depend on their composition and size distribution, properties which are not very accurately known for the first few weeks after the volcanic eruption. Evidence from direct sampling (*2, 3, 15, 16, 22*) indicates that the volcanic material initially injected into the stratosphere contained a large amount of dust, which may have been silicate in composition and could thus absorb a significant part of the solar radiation incident on it. Within 1 month after the eruption most of the coarse dust had settled beneath the 20-km region (*3, 16, 22*). The sulfate concentration reached its maximum several months after the eruption (*3, 22*); the dominant composition was sulfuric acid and ammonium sulfate, materials which do not significantly absorb solar radiation. To examine the potential effect of aerosol composition, we made computations for the two extreme cases of pure silicate material and pure sulfuric acid (75 percent concentration by weight in water), with appropriate solar and thermal optical properties (*23*).

Table 1 shows that the silicate dust results in a substantial increase in the stratospheric heating with some decrease in tropospheric cooling. Because of the long time constant for the tropospheric cooling, the result for pure sulfuric acid should be appropriate for the troposphere. Although the silicate dust may significantly affect the stratospheric warming during the first several weeks, it is difficult to be quantitative because of the very nonuniform distribution of aerosols at that time. More reliable modeling of the expected thermal effects will require accurate knowledge of the composition, size distribution (*24*), and location of the aerosols.

The model results also depend on the assumed thickness of the ocean mixed layer and on our restriction of the model domain to latitudes between 30°N and 30°S, as indicated in Table 1, but in ways which are easily understood. With a 100-m mixed layer, instead of 70 m (*25*), the maximum tropospheric cooling is 0.44°C at 660 days after the eruption instead of 0.58°C at 590 days, and there is no significant change in the stratospheric warming. The date of maximum tropospheric cooling is little changed because in either case the time constant for cooling the mixed layer is longer than that for changing the aerosol optical thickness. The stratospheric temperature change depends mainly on the assumed stratospheric aerosol properties, but also on tropospheric temperature since the effect is basically due to absorption of upwelling thermal radiation. If global average properties are used instead of those for latitudes 30°N to 30°S, the tropo-

Table 1. Temperature change in the troposphere and stratosphere computed with the one-dimensional radiative-convective model. The column headings give the mixed-layer depth, aerosol composition, and model domain. The tropical model is limited to latitudes between 30°N and 30°S.

Δt (mo)	Temperature change (°C)							
	Troposphere				Stratosphere (55 mbar)			
	70 m, sulfate, tropical	100 m, sulfate, tropical	70 m, sulfate, global	70 m, silicate, tropical	70 m, sulfate, tropical	100 m, sulfate, tropical	70 m, sulfate, global	70 m, silicate, tropical
1	−0.02	−0.02	−0.01	−0.02	1.25	1.25	0.38	5.5
3	−0.08	−0.06	−0.03	−0.05	3.65	3.65	1.50	10.8
6	−0.25	−0.18	−0.11	−0.13	5.29	5.31	2.13	12.7
9	−0.40	−0.28	−0.17	−0.18	5.01	5.05	2.02	12.3
12	−0.50	−0.36	−0.22	−0.21	3.72	3.77	1.48	10.2
18	−0.57	−0.43	−0.24	−0.22	1.86	1.91	0.80	7.2
24	−0.56	−0.44	−0.24	−0.20	0.75	0.80	0.32	4.9
30	−0.52	−0.42	−0.22	−0.17	0.16	0.18	0.03	3.2
36	−0.47	−0.39	−0.21	−0.14	−0.13	−0.10	−0.03	2.2

spheric cooling is reduced to about half of that illustrated in Fig. 2, primarily because of the reduction of the average aerosol optical thickness (*26*). With a one-dimensional model perhaps the most that can be said is that the best estimate is somewhere between the results computed for the global average and the isolated tropical cases. Despite this uncertainty, the results are in excellent agreement with the observations. We believe that the observations are fairly conclusive in indicating tropospheric cooling by a few tenths of a degree Celsius 1 or 2 years after Agung, but the very close fit to the theoretical curve in Fig. 2 is almost certainly fortuitous, particularly in view of the noisy appearance of the observed temperatures in earlier and later years (*27*).

We believe that the greatest weakness of our model computations is the absence of interactions of the computed heating with the atmospheric dynamics. Table 1 suggests that the heating by aerosols is so large that it could be expected to influence large-scale dynamic phenomena, and thus future analyses should include special efforts to properly model three-dimensional atmospheric motions. A related defect of the computations is the omission of potential cloud cover feedbacks; the radiative perturbation due to aerosols could conceivably induce a change in the clouds which would counteract (or reinforce) the direct climatic effect of the aerosols. Proper analysis will require a model in which the cloud cover is computed from first principles and is allowed to realistically interact with the radiation field and the atmospheric dynamics. However, the extent to which the observed climatic effect agrees with that obtained from a simple radiative model in fact provides some evidence that such potential feedbacks do not overwhelm the direct radiative effect.

We cannot resist the temptation to point out the analogy between the terrestrial climate problem and the task of understanding the nature of other planetary atmospheres. Several years ago, ground-based telescopic observations of Venus were used to obtain remarkably precise values for the refractive index and size distribution of the aerosols forming the Venus clouds (*28, 29*)—more precise, in fact, than those properties are known for terrestrial aerosols today—and this led to identification of the Venus aerosols as sulfuric acid (*29, 30*). Models for the atmospheric thermal structure indicate that these sulfuric acid aerosols, which are large enough to cause a greenhouse effect, together with the abundance of CO_2 gas are responsible for the basic climatic state on Venus (*30, 31*), where the surface is hot enough to bake bread (or melt lead). Now it has become clear that the same atmospheric constituents, CO_2 and sulfate aerosols, are of great interest for their possible effects (both natural and anthropogenic) on terrestrial climatic change. Clearly a great deal stands to be gained by simultaneous studies of the earth's climate and the climate on other planets.

In summary, the magnitude, sign, and time delay of the temperature changes computed with a simple one-dimensional climate model for both the stratosphere and troposphere are in excellent agreement with those of the temperature changes observed after the injection of stratospheric aerosols by the eruption of Mount Agung. Despite the simplicity of the model, uncertainties about the input data, and severe limitations of the observations, the results are encouraging with regard to the question of whether global-scale climatic perturbations can be reliably modeled and thus predicted.

It is important to be cautious against overinterpretation of either the observations or the model results. The observations of aerosol properties and atmospheric temperatures are not nearly as detailed and accurate as desired, and we cannot be certain that the observed temperature variations are not in large part due to other mechanisms or simply to climatic noise. The model for the climate system is grossly oversimplified and particularly fails to properly account for three-dimensional transports and cloud feedbacks. Nevertheless, we believe that the several aspects involved in the agreement of observations and theory are too extensive to be coincidental, and that they provide strong evidence that radiative perturbations of the magnitude of such volcanic emissions do noticeably affect the climate, contrary to some recent opinions (*21*). Since potential anthropogenic radiative perturbations are similar in magnitude (*14*), the results also provide evidence that we should be concerned about possible inadvertent modification of global climate by humans.

Finally, we emphasize the enormous potential scientific value of the natural climatic experiment provided by a large volcanic eruption. One very useful strategy for investigating the global climatic system would be to make detailed observations and analyses of the next large volcanic explosion. The observations should include global monitoring of the spread, growth, and decay of the aerosols, sampling of the aerosol properties in situ, and accurate global monitoring of the climatic response. These data would permit testing of global climate models and aid in-depth analyses of radiative, chemical, and dynamical processes, which are essential for obtaining an improved understanding of the physical mechanisms and interactions involved.

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References and Notes

1. P. W. Booth, S. W. Matthews, R. E. Sisson, *Natl. Geogr.* **124** (No. 10), 436 (1963); M. T. Zen and D. Hadikusumo, *Bull. Volcanol.* (Ser. 2), **27**, 269 (1964).
2. J. P. Friend, R. Leifer, M. Trichon, *J. Atmos. Sci.* **30**, 465 (1973); R. D. Cadle and G. W. Grams, *Rev. Geophys. Space Phys.* **13**, 475 (1975); P. Hamill, C. S. Kiang, R. D. Cadle, *J. Atmos. Sci.* **34**, 150 (1977).
3. A. W. Castleman, H. R. Munkelwitz, B. Manowitz, *Tellus* **26**, 222 (1974).
4. H. H. Lamb, *Philos. Trans. R. Soc. London Ser. A* **266**, 425 (1970).
5. W. J. Humphreys, *Physics of the Air* (McGraw-Hill, New York, 1940).
6. H. W. Wexler, *Sci. Am.* **186** (No. 4), 74 (1952).
7. J. B. Pollack, O. B. Toon, C. Sagan, A. Summers, B. Baldwin, W. Van Camp, *J. Geophys. Res.* **81**, 1071 (1976); S. H. Schneider and C. Mass, *Science* **190**, 741 (1975).
8. A. J. Dyer and B. B. Hicks, *Q. J. R. Meteorol. Soc.* **94**, 545 (1968).
9. R. D. Cadle, C. S. Kiang, J. F. Louis, *J. Geophys. Res.* **81**, 3125 (1976).
10. H. Moreno and J. Stock, *Publ. Astron. Soc. Pac.* **76**, 55 (1964); A. Przybylski, *Astronaut. Acta* **14**, 285 (1964); W. M. Irvine and F. W. Peterson, *J. Atmos. Sci.* **27**, 62 (1970).
11. The optical thickness τ is defined such that a fraction $\exp(-\tau)$ of radiation impinging on the layer perpendicularly is transmitted unscattered.
12. Measurements of the intensity of solar radiation at the earth's surface (*8*) are in good agreement with this, and also indicate that the increase in the optical depth in the Northern Hemisphere was two to five times smaller than in the Southern Hemisphere. Similar optical depths are obtained from analysis of the amount of solar radiation in the earth's shadow measured during lunar eclipses [F. E. Volz, *J. Geophys. Res.* **75**, 5185 (1970)].
13. The time-marching procedure and convective adjustment are basically the same as those employed by S. Manabe and R. T. Wetherald [*J. Atmos. Sci.* **24**, 241 (1967)] and W. C. Wang and G. A Domoto [*J. Appl. Meteorol.* **13**, 521 (1974)]. We specify gaseous absorption coefficients within a frequency interval $\Delta\nu$ by means of a probability distribution function, which is thus used for an implicit integration over wavelength. This function is obtained for each frequency interval and each gas by using the band model of W. Malkmus [*J. Opt. Soc. Am.* **57**, 323 (1967)]. The number of frequency intervals (59 for solar radiation and 49 for the thermal spectrum) is sufficient to permit the spectral dependence of the cloud particle properties, which is relatively slowly varying, to be treated explicitly. The radiative calculations are similar to those described by Wang *et al.* (*14*). However, the present calculations were performed for a particular band of latitudes and the north-south horizontal transport of energy by the atmosphere and oceans was included by means of specified climatological transports [C. W. Newton, *Meteorol. Monogr.* **13**, 215 (1972)]; with this procedure the model yields a realistic surface temperature for any latitude band, but the feedback effect of temperature change on the horizontal transports is not included. A detailed description of the model will be given elsewhere.
14. W. C. Wang, Y. L. Yung, A. A. Lacis, T. Mo, J. E. Hansen, *Science* **194**, 685 (1976).
15. O. B. Toon and J. B. Pollack, *J. Appl. Meteorol.* **15**, 225 (1976).
16. S. C. Mossop, *Nature (London)* **203**, 824 (1964).
17. R. E. Newell, *J. Atmos. Sci.* **27**, 977 (1970); *Sci. Am.* **224** (No. 1), 32 (1971).
18. R. M. McInturff, A. J. Miller, J. K. Angell, J. Korshover, *J. Atmos. Sci.* **28**, 1304 (1971); A. B.

Pittock, *Q. J. R. Meteorol. Soc.* **100**, 39 (1974).
19. R. E. Newell and B. C. Weare, *Science* **194**, 1413 (1976).
20. J. K. Angell and J. Korshover, *Mon. Weather Rev.* **105**, 375 (1977).
21. These results contradict the conclusions of B. J. Mason [*Nature (London)* **264**, 608 (1976)] that the Agung volcanic aerosols produced neither an observed tropospheric response nor detectable effect on the troposphere in models. The observed decrease in the tropospheric temperature is considered to be significant (*19, 20*). The temperature change computed in the model is a straightforward radiative effect of demonstrable accuracy (*13*), and the absence of it in Mason's analysis must be due to deficiencies in the models he employed.
22. A. W. Castleman, H. R. Munkelwitz, B. Manowitz, *Nature (London)* **244**, 345 (1973).
23. For silicates we use the optical properties of basalt given by J. B. Pollack, O. B. Toon, and B. H. Khare [*Icarus* **19**, 372 (1973)], which, according to Pollack *et al.* (*7*), has an amount of absorption close to the limit that can be expected from silicate dust. For sulfuric acid we use the optical properties tabulated by K. F. Palmer and D. Williams [*Appl. Opt.* **14**, 208 (1975)].
24. The size of the sulfuric acid aerosols is crucial for determining their net radiation effect, since their impact on the troposphere through their effect on solar radiation is one of cooling while in the thermal infrared they cause a greenhouse warming, as pointed out by Pollack *et al.* (*7*). For the size distribution we have assumed, which is representative of the normal stratospheric aerosol (*15*), the thermal warming effect is only about one-fourth as large as the solar cooling. For a larger mean size, with the shape of the size distribution held fixed, the relative contribution of the greenhouse warming increases. On Venus the aerosols are large enough to yield a net warming and contribute substantially to the strong greenhouse effect on that planet (*31*). On the other hand, if the size distribution falls off toward larger particles faster than the log-normal distribution used in (*15*), the greenhouse warming will be decreased. In future measurements special efforts should be made to obtain not only the mean particle size, but also accurate data on the large-particle tail of the size spectrum.
25. On the basis of the graphs of K. H. Bathen [*J. Geophys. Res.* **77**, 7138 (1972)], we estimate the annual average global mixed-layer depth as about 70 m. The value 100 m is approximately the global average for the maximum mixed-layer depth during the year.
26. For the global average case we use an aerosol optical depth 0.625 times as large as that for the tropical case; that is, the maximum optical depth is 0.125 instead of 0.2. Observations (*8*) indicate that the maximum of 0.2 was applicable to about half the globe, with the maximum in the Northern Hemisphere about four times smaller. The temperature change ΔT in the troposphere for the global average case is reduced by more than the factor 0.625 for several reasons: (i) the average solar insolation is about 20 percent larger for the tropics than for the global average; (ii) the water vapor feedback ΔT is larger in the tropics than for the global average; and (iii) the inclusion of horizontal energy tranports out of the region of interest, with the absolute value of the transports kept fixed, slightly increases ΔT over that in the case of no transports.
27. Because of the difficulty in measuring the small climatic signal in the presence of substantial noise, or natural climatic fluctuations, a valuable complementary approach to the one we have taken is to statistically analyze the observed temperature changes afer all the large eruptions in recent historical times [C. Mass and S. H. Schneider, *J. Atmos. Sci.* **34**, 1995 (1977)]. Although the aerosol properties are not well known for the older eruptions, it is clear from Table 1 that a tropospheric cooling should be expected. Mass and Schneider do find statistical evidence for a posteruption cooling.
28. J. E. Hansen and A. Arking, *Science* **171**, 669 (1971).
29. G. T. Sill, *Commun. Lunar Planet. Lab. No. 171* (1972), p. 191; A. T. Young, *Icarus* **18**, 564 (1973).
30. A. P. Ingersoll, *J. Atmos. Sci.* **26**, 119 (1969); S. I. Rasool and C. De Bergh, *Nature (London)* **226**, 1037 (1970).
31. J. B. Pollack and R. Young, *J. Atmos. Sci.* **32**, 1025 (1975).
32. We thank P. Stone, R. Jastrow, Y. Yung, M. Cane, S. Manabe, and two anonymous referees for useful suggestions.

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Anilines: Selective Toxicity to Blue-Green Algae

Abstract. *The blue-green alga* Agmenellum quadruplicatum *(strain PR6) was very sensitive to aniline and* p-*toluidine (potential environmental toxicants) in an algal lawn assay (the growth of the algal lawn was inhibited with as little as 1 microgram of* p-*toluidine per disk). Assays with seven other species of blue-green algae showed that they had varying sensitivities ranging from 1 to 100 micrograms of* p-*toluidine. Under comparable conditions, 0.5 milligram or more of* p-*toluidine was needed to inhibit a green alga, a diatom, or two species of bacteria.* p-*Toluidine had no immediate effect on the photosynthesis or respiration of* A. quadruplicatum, *although growth was arrested and viability declined.*

Aniline and its methylated derivatives are commonly used in the manufacture of dyes and other organic chemicals. Although these types of compounds are known to be toxic to humans (*1*), very little is known about whether they are toxic to other living things. Thus, it is of interest that the toxicity of water-soluble extracts of fuel oils to blue-green algae is traceable to their content of toluidines (methylanilines) (*2*). In this report we further characterize the response of blue-green algae to toluidines and other methylated anilines.

The blue-green algae tested here can be divided into two groups on the basis of their resistance to *p*-toluidine (Table 1). Four species were completely inhibited by 10 μg of *p*-toluidine per disk. The sensitivity of three of these species persisted to as little as 1 μg of *p*-toluidine per disk, with an evident effect seen as smaller colonies beyond the zone of inhibition. Four other blue-green algae required 100 to 500 μg of *p*-toluidine per disk for complete inhibition. Using a different assay technique, Fitzgerald *et al.* (*3*) reported a 50 percent kill of the blue-green alga *Microcystis aeruginosa* with aniline at 20 parts per million (ppm); a 100 percent kill occurred with *p*-aminodimethylaniline (2 ppm) and *p*-aminodiethylaniline hydrochloride (1 ppm). By comparison, the green alga *Chlorella autotrophica* (strain 580) was not inhibited at 500 μg of *p*-toluidine per disk. Similarly, Smirnova *et al.* reported that *Scenedesmus obliquus*, also a green alga, was not affected by *o*-toluidine (*4*). A diatom, *Cylindrotheca* sp. (strain N1), showed only very slight inhibition at 500 μg of *p*-toluidine per disk. Two test bacteria showed no inhibition, even at 1000 μg per disk.

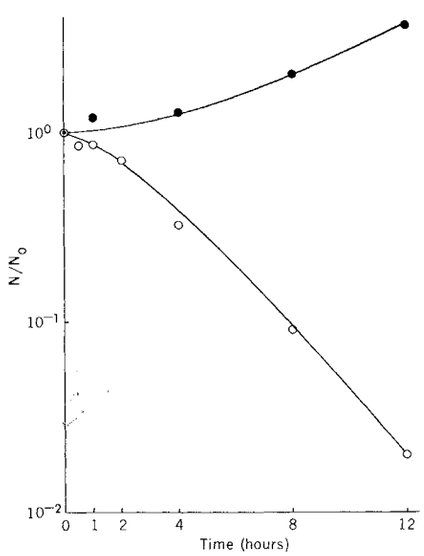

Fig. 1. Viability of *Agmenellum quadruplicatum* (strain PR6) after exposure to *p*-toluidine (500 ppb) in medium ASP-2; N is the cell number in suspension at time t; N_0 is the original cell number in suspension. *p*-Toluidine was added to the culture (1×10^7 cells per milliliter) at zero time; incubation continued at 30°C, with F40CWX lamps, under continuous aeration with 1 percent CO_2 in air. One-tenth of a milliliter of appropriately diluted cell suspensions was plated out on medium ASP-2 plus 1 percent agar and was incubated as described in Table 1. (Closed circles) Control culture; (open circles) culture plus *p*-toluidine.

We tested aniline, toluidines, and other substituted anilines further in order to better establish the spectrum of toxicity of this group of compounds toward a sensitive blue-green alga, *Agmenellum quadruplicatum* (strain PR6), a marine coccoid form. The data (Table 2) show that strain PR6 was about as sensitive to aniline as to *p*-toluidine. Both compounds completely inhibited growth at 10 μg per disk. Of the other compounds tested, complete inhibition of the alga occurred at 100 μg except for those compounds with substituents on the nitrogen atom. *N*-Methylaniline and *N*-ethylaniline were much reduced in toxicity, causing only slight inhibition at 500 μg per disk. *N,N*-Dimethylaniline was not toxic even at 1000 μg per disk. This result suggests that the toxicity of the *p*-toluidine and aniline molecules is associated with the amino group. The reduced toxicity of 6-ethyl-*o*-toluidine may simply be due to steric hindrance of the amino group. By comparison *p*-cresol, commonly found in water extracts of fuel oils (*2*), caused little inhibition of strain PR6. Biphenyl, Aroclor 1242 and Aroclor 1254 (Aroclor compounds are polychlorinated biphenyls), and dimethyl-

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