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Global Climate System (MO7003)

Impact on Climate of a Huge Scale Nuclear Assault

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Contents

1 Introduction

In the film Terminator an artificial intelligence named Skynet takes over and conducts a nuclear assault against the human race. In this report, the question of what would happen to the climate after such an incident is investigated.

The scenario of a fictional AI taking over and diminishing mankind by detonation of it's nuclear arsenal is very unlikely. However, a huge stock of these kinds of warheads exists and humans are very capable of making bad choices themselves. What would happen to the climate after a nuclear war?

Similar questions have been topic of scientific research in the past. In [1] a potential nuclear conflict between India and Pakistan was assessed. The paper [2] focuses on the consequences of a nuclear war between Russia and the USA. The work of [3] provides a very detailed estimation of the aftermath of such a scenario. Some of them, such as [1] and [2] predict a so-called nuclear winter, others like [3] are more optimistic in that respect.

In this report, we try to find the right setup for MPI-ESM1.2 to model the aftermath of a huge amount of nuclear weapons detonating. The goal is to run set model and observe changes in values such as surface temperature, precipitation and stratospheric temperature. Only effects on climate are relevant, so impacts like radioactive fallout, starvation and death are ignored.

2 Theory

In this section the theoretical background information that is necessary to choose the right model setups and interpret the output is provided.

2.1 Nuclear Weapons

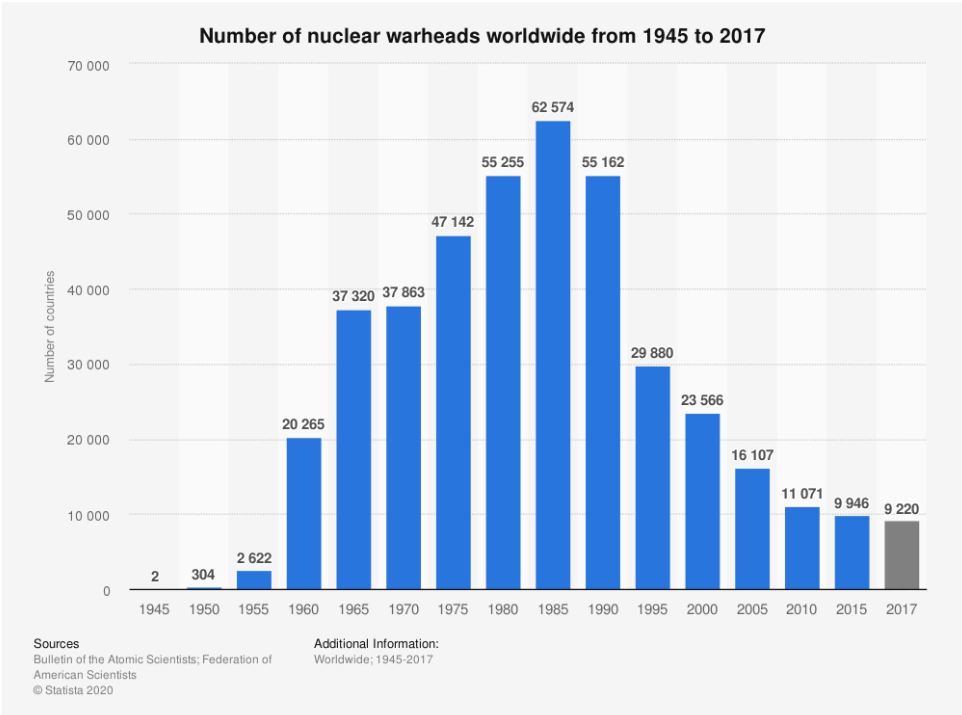


Figure 1: Development of nuclear arsenal over time [4].

Even though dismantlement is happening since 1985, as can be seen in figure 1, there was still an estimated amount of nearly 30 0000 warheads in the world in the mid 1990's. There are various kinds of nuclear weapons. The effects of an explosion depend not only on the blasting power but also the altitude of the explosion and the composition of the soil beneath [3]. The blast yield of a nuclear weapon is usually measured in the amount of TNT that would release the same amount of energy. The warhead called "Little Boy" that was dropped on Hiroshima by the USA in 1945 is one of the smaller bombs. Its blasting power only reaches about 15 kilotons TNT. The most powerful bomb that was ever tested is the so-called "Zar Bomb" with a blasting yield of about 50 megatons TNT. Figure 2 shows the radius of direct impact such a bomb would have if dropped in the middle of Stockholm.

The key mechanism in all types of bombs is the rapid release of a high amount of energy. It will not be discussed here how this works physically. The sudden increase in temperature causes the evaporation of all matter in a certain radius, the so-called fireball. The magnitude of that radius depends on the altitude of the explosion and the blasting power of the bomb. The gas reaches temperatures over 10 000 °C. This has various consequences. One of them is that a high pressure is initiated on the surrounding matter and therefore causes a shock wave. Here it is assumed that the bomb is dropped near the surface. The vaporised matter rises and drags various particles and water vapour with it. This causes the characteristic mushroom cloud. The heat also results in a firestorm. This means air is sucked in from the surrounding atmosphere near the surface. The particles originating from the vaporisation and burning, called soot, are dragged upwards. [6]

Of course there are other consequences following the release of ionising radiation and radioactive particles

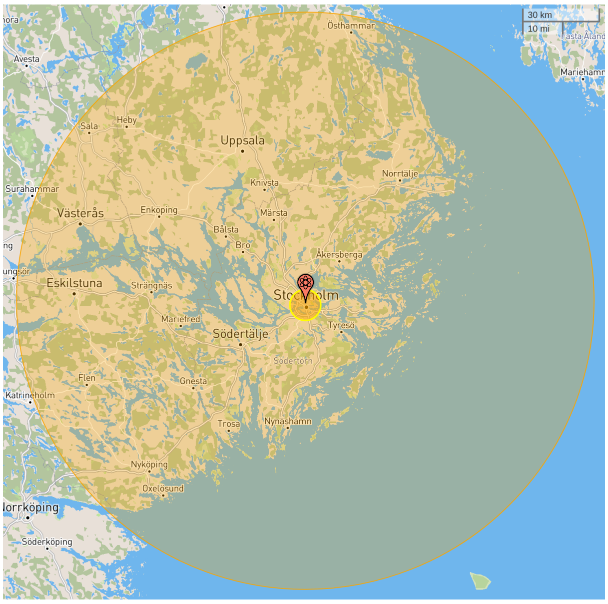


Figure 2: Big circle: minimal radius where no burns are expected for a person outside of it. Small circle: fireball. Generated with [5].

(fallout). Also the radius where humans will suffer from third degree burnings is much larger than just the fireball. However, this doesn't directly affect the climate, so it will not be discussed here.

2.2 Effects on Climate

A detonation of a nuclear bomb is similar to a volcanic eruption in the sense that in both scenarios a huge amount of particles is injected into the atmosphere. In both cases they are quickly distributed around the globe by the wind. However in course of a volcanic eruption large amounts of sulfur are released, while the explosion of an atomic bomb generates mainly so-called soot. Soot is a result of the firestorms described above that follow a nuclear explosion. It's exact composition depends on the material burned on the surface. However, the main component is black carbon. [2].

In contrast to sulfur, black carbon is absorbing short wave radiation and heats up the air around. This induces an upward drag. Therefore soot aerosols are transported up to the lower stratosphere where they remain much longer than particles induced by a volcanic eruption. How large that effect is, depends also on current weather conditions such as temperature, humidity and cloud coverage. Not all of the particles originating from the fires elevate to higher altitudes [3].

Once in the stratosphere the soot burden becomes evenly spread around the globe with time. Black carbon particles change in size and concentration due to coagulation and sedimentation. As will be discussed later both properties are important for the interaction with incoming solar radiation. The decline is believed to be logarithmic, but depends on various factors as particle size, humidity and temperature [2].

2.2.1 Nuclear Winter

A concept often predicted to follow a huge scale nuclear war is the so-called nuclear winter. The term describes a significant drop in surface temperature for several years. It means that in large regions of the world temperatures usually experienced in winter occur in the middle of summer. In [2] the effect of 150 Tg of black carbon injected into the atmosphere is modelled. Surface temperature was so low that temperatures below freezing were observed in the northern hemisphere all year long. This effect is caused by the obstruction of sunlight in the stratosphere. Less short wave solar radiation reaches the surface.

In [3] a more detailed aerosol modelling was conducted. Though still a significant drop in surface temperature was predicted, no nuclear winter effect could be observed.

2.2.2 Ozone

A huge amount of aerosols, such a soot, that absorb shortwave radiation result in a significant heatup of the stratosphere. This higher temperature catalyse chemical reactions that split up stratospheric ozone. Therefore the ozone layer is depleted [7].

2.3 Optical Aerosol Properties

This section is based on the book [8]. If radiation meets a particle there are essentially three things that can happen. Parts of the light can be either absorbed, scattered or not effected at all. To characterise an aerosol's interaction with radiation the four optical parameters *single scattering albedo (ssa)*, *asymmetry parameter (asy)*, *extinction coefficient (ext)* and *aerosol optical depth (aod)* are sufficient.

The single scattering albedo is the ratio between the scattered and extinct radiation. It is common to denote *asy* as g, *ext* as σ^{ext}, *ssa* as ω and *aod* as τ. To explain the role of the above parameters, two more variables have to be introduced, the scattering coefficient σ^{sca} and the absorption coefficient σ^{sca}. Both are given per meter along the atmospheric height. In a nutshell, the first one indicates how much of incoming radiation with a fixed wavelength is scattered, while the second parameter provides information about how much of this radiation is absorbed. The sum of these two yields the extinction coefficient

$$\sigma^{\text{ext}} = \sigma^{\text{sca}} + \sigma^{\text{abs}}.$$

By integrating *ext* over the height of the layer of aerosols the *aod* is obtained

$$\tau = \int_{h_0}^{h_1} \sigma^{\text{ext}}(h)\,\mathrm{d}\,h. \tag{1}$$

Intuitively, *aod* is what defines how much radiation of what wavelength is blocked. A high *aod* for short wavelengths means that not much sunlight can reach the earth's surface.

The *ssa* indicates how the radiation is mainly obstructed, by scattering or absorption. If all aerosols would be perfect spheres with the same radius, the calculation of *ssa* would simplify to $\omega = \frac{\sigma^{\text{sca}}}{\sigma^{\text{ext}}}$.

An *ssa* close to zero implies that nearly all of the extinct radiation is absorbed and transformed to infrared radiation. On the other hand, if the value is close to one most of the radiation that is affected by the aerosol is scattered.

The asymmetry parameter describes in which angle the solar radiation is scattered relative to the incoming light. If $g = 1$ the entire scattering happens in forward direction, if $g = -1$ in backward direction.

All of the above parameters depend on chemical composition and radius of the particle. The values vary with wavelength of incoming radiation. *aod* and *ext* also change with aerosol concentration.

3 Modelling

To model the influence of a nuclear assault on earth's climate the model MPI-ESM1.2 is used. In contrast to other climate models [2] the aerosols are prescribed in MPI-ESM1.2. More precisely the properties *aod*, *ext*, *asy* and *ssa*, described in 2.3, must be given for each month, latitude, pressure level (except *aod*) and wavelength. Same is true for the ozone concentration. As in [1], [2] and [3] no affect on the CO_2 concentration is assumed.

3.1 Soot

As already established the total amount of soot S (unit: Tg) in the atmosphere declines logarithmically. Like in [2] an e-folding lifetime of 4.6 is assumed. This yields

$$S(t) = S_0 e^{-4.6\frac{t}{12}} \tag{2}$$

where t is given in months after detonation and S_0 is the initial amount of soot reaching the stratosphere.

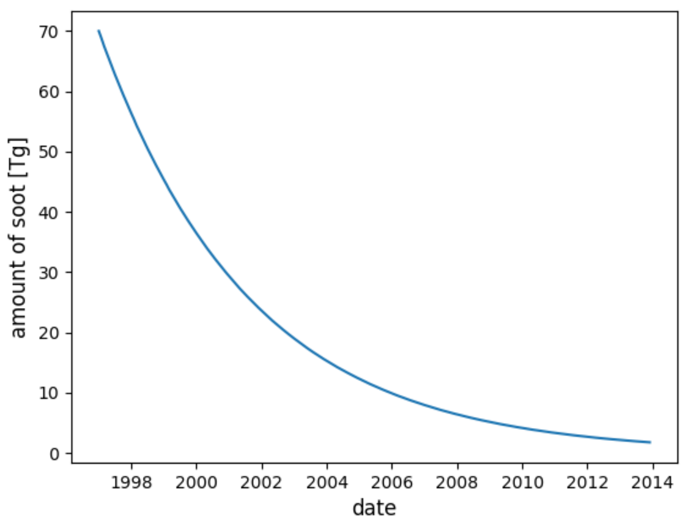


Figure 3: Soot in the stratosphere

To estimate S_0 previous studies are considered. In [1] it was predicted that the detonation of 20 000 nuclear warheads would result in a release of 150 Tg soot into the atmosphere. If we assume that Skynet is holding back some nuclear weapons for potential further use and only detonates less than 10 000, a value of 70 Tg is reasonable for S_0. Also it turned out that the model runs reliably with that amount of forcing and doesn't crash.

3.2 Ozone

The ozone is also prescribed in MPI-ESM1.2. It is assumed that the concentration of ozone ρ follows the formula

$$\rho(t) = \rho_{\text{control}}(t)e^{-kS(t)}$$

where ρ_{control} denotes the ozone concentration of the control run and S is the mass of soot in the atmosphere, calculated in 3.1. The motivation for that relation is, that the probability of an ozone molecule decomposing increases with the concentration of ozone in the stratosphere.

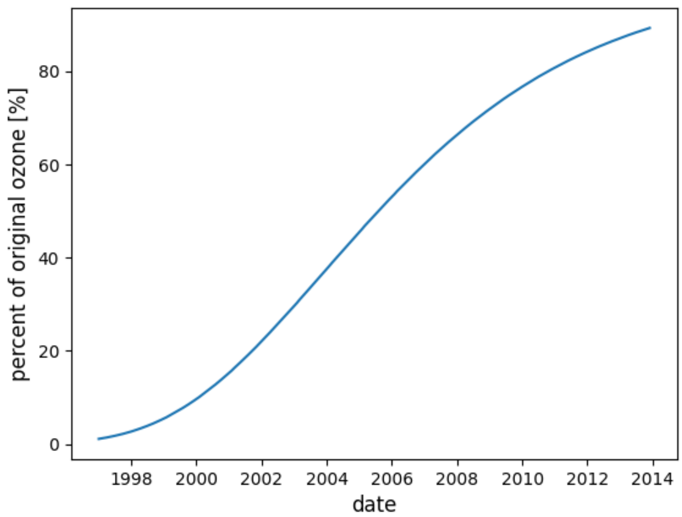


Figure 4: Percentage of the control run's stratospheric ozone with an initial amount of 70 Tg soot in the stratosphere

It remains to determine k. According to [7] an amount of 1 Tg soot in the stratosphere reduces the ozone concentration by 8 %. An injection of 5 Tg soot causes a depletion of 20%. Both cases result in different values for k. In this model an average of both results is used and k is set to 0.064. Figure 4 illustrates the depletion of stratospheric ozone over time for an initial amount of soot of 70 Tg.

3.3 Aerosols

The first idea was to estimate the values for *ssa*, *ext* and *aod* based on previous studies, as it is described in 3.3.1. Unfortunately, this was not possible because the model kept crashing. The observation that all optical parameters are highly dependent on the wavelength lead to the modelling approach described in 3.3.2. However this was not successful either, the model still didn't run through.

Several different setups for aerosol modelling were tested. The problematic parameter turned out to be the *ssa*. Possible reasons for that are discussed in section4. No setup with the estimated parameter of *ssa*, or similar parameters, ran even for one year. The maximum amount of soot that could be injected for the calculated *ssa* was 5 Tg. Assuming less absorbing soot particles, precisely choosing a refractive index of $m = 1.8 - 0.2\,\mathrm{i}$, resulted in a maximum amount of 10 Tg before the model crashed. An even lower imaginary part would result in a completely different behaviour. Therefore all of the experiments with 70 Tg of injected soot were tested with the same *ssa* as in the control run.

Three of the experiments are presented here. To save computation time not all were conducted for the same amount of years. The simulation always starts in the year 1997. (Note: This is the case because I should have watched the Terminator films before conducting the experiments. I thought Skynet would take over in 1997.). There are no differences in the ozone modelling. In all three setups the variables *ssa*, *asy*, *ext* and *aod* are only evaluated in the pressure levels where soot aerosols are assumed to accumulate. According to Figure 9 in [3] most of the black carbon released after a nuclear assault can be located in altitudes from 20 km to 50 km. To estimate the approximate pressure figure 1 in [9] was used. There exist 40 pressure levels in MPI-ESM1.2. The soot aerosols are assumed to be located at level 8 to 27.

Table 1 provides an overview over the different setups.

parameter/model	I	II	III
initial soot [Tg]	70	70	5
end year	2014	2014	2003
ssa	as control	as control	calculated
asy	as control	calculated	calculated
aod	estimated	calculated	calculated
ext	estimated	calculated	calculated

Table 1: Parameters of different modelling setups.

3.3.1 Estimation based on previous studies

The first attempt was to estimate the optical properties based on values used in previous research. No wavelength dependence was taken into consideration. The values were uniquely distributed over all coordinates, except altitude.

According to [1] soot has a mass extinction coefficient σ_m^{ext} of 5.5 m^2g^{-1}. This parameter can be used to calculate the extinction coefficient

$$\sigma^{\text{ext}} = \frac{\sigma_m^{\text{ext}} m_{\text{soot}}}{\frac{4}{3}\pi\left((R_A+h_1)^3-(R_A+h_0)^3\right)}$$

where $R_A = 6.378 \cdot 10^6$m denotes the mean earth radius, $h_1 = 5 \cdot 10^4$m is the upper boundary of the soot layer above sea level and $h_0 = 2 \cdot 10^4$m the lower boundary. To get a value for the *aod* data from [1] is used by assuming that there is a factor b such that *aod* amounts to $S(t)b$, with S from (2). The factor is estimated to be $b = 0.0051\ \text{Tg}^{-1}$.

The asymmetry factor remains the same as in the control run. Originally an *ssa* of 0.64 as mentioned in [2] was planed to use. However, the model crashed for reasons investigated in section 4. Therefore the *ssa* is not altered either in that setup.

3.3.2 Calculation

A more accurate approximation is to compute the relevant optical properties. The following description of how this can be done is from [8]. The key point here is to take into account that *ssa*, *asy*, *ext* and *aod* are dependent on the wavelength of the radiation the particle is interacting with.

The concentration of particles N_0 is actually given by a distribution dependent on the particle radius r, meaning $N_0 = \int n(r)\,\mathrm{d}\,r$. However, for simplification, we assume that all particles have the same radius. Formally this corresponds to defining n as the delta distribution concentrated around a radius r_e. According to [2] it is reasonable to assume an effective radius r_e of 0.1μm for soot aerosols.

The concentration N_0 is estimated by

$$N_0 = \frac{a}{\frac{4}{3}\pi\left((R_A+h_1)^3-(R_A+h_0)^3\right)} \tag{3}$$

where again $R_A = 6.378 \cdot 10^6$m denotes the mean earth radius, $h_1 = 5 \cdot 10^4$m is the upper boundary of the soot layer above sea level and $h_0 = 2 \cdot 10^4$m the lower boundary. Now the amount of soot particles a has to be estimated. Black carbon has a density ρ of about $2 \cdot 10^6$ g·m^{-3} [10]. So one molecule with radius r_e has a mass of $\rho\frac{4}{3}\pi r_e^3$. The total mass of soot is given in formula 2. This yields

$$a = \frac{3S(t)}{4\rho\pi r_e^3}.$$

A next step is to calculate the optical properties needed for our model. Therefore a theory called Mie theory is used. It proposes that only the radius of the aerosol, the wavelength of the light and the so-called refractive index m is needed to describe the relationship of a perfectly spherical particle with short wave radiation. The refractive index is a complex number that has to be determined experimentally. It varies depending on the exact chemical composition of the particle. As in [2] $m = 1.8 - 0.67\,\mathrm{i}$ is used for soot in this work. Now the dimensionless scattering and extinction efficiency factors Q^{sca} and Q^{ext} as well as the asymmetry factor are calculated by an open source implementation of the Mie Algorithm [11]. From these parameters σ^{sca} and σ^{ext} can be obtained by

$$\sigma^{\text{sca}} = \int_0^\infty \pi r^2 Q^{\text{sca}}(r) n(r)\,\mathrm{d}\,r = \pi r_e^2 N_0 Q^{\text{sca}} \tag{4}$$

$$\sigma^{\text{ext}} = \int_0^\infty \pi r^2 Q^{\text{ext}}(r) n(r)\,\mathrm{d}\,r = \pi r_e^2 N_0 Q^{\text{ext}}. \tag{5}$$

The single scattering albedo ω is defined as

$$\omega = \frac{Q^{\text{sca}}}{Q^{\text{ext}}} \tag{6}$$

which is in this case equivalent to $\omega = \frac{\sigma^{\text{sca}}}{\sigma^{\text{ext}}}$.

According to 1 the aerosol optical depth τ is obtained by integrating the extinction coefficient over the atmospheric heights where soot can be found. Assuming again a uniform distribution this simplifies to

$$\tau = \int_{h_0}^{h_1} \sigma^{\text{ext}}(h)\,\mathrm{d}\,h = (h_1 - h_0)\sigma^{\text{ext}}. \tag{7}$$

The optical parameters above are summarised in table 2 for a better overview.

Table 2: Optical properties needed for modelling.

name	unit	calculation	description
r_e	μm	given as 0.1	effective particle radius
m	–	given as $1.8 - 0.67\,\mathrm{i}$	refractive index
N_0	m^{-3}	equation (3)	concentration of soot
Q^{sca}	–	Mie Algortihm	scattering efficiency
Q^{ext}	–	Mie Algorithm	extinction efficiency
g	–	Mie Algorithm	asymmetry factor
ω	–	equation (6)	single scattering albedo
σ^{sca}	m^{-1}	equation (4)	scattering coefficient
σ^{ext}	m^{-1}	equation (5)	extinction coefficient
τ	–	equation (1)	aerosol optical depth

4 Results and Discussion

Before going into detail, a few general remarks about the setup. Model I and II have in common that the *ssa* of the control run in the respective year was used. This is reasonable, because the variance of *ssa* in time is relatively low in the first 5 years of the simulation as can be seen in figure 5. The biggest drop in temperature occurs in these years, so the focus lies there. However, the distribution is not uniform around the globe as it is assumed for soot particles from a nuclear explosion. This is visualised in figure 5. It cannot be said what effects are attributed to that difference in distribution. To answer that question another setup would be needed where all parameters remain the same and only the *ssa* distribution changes.

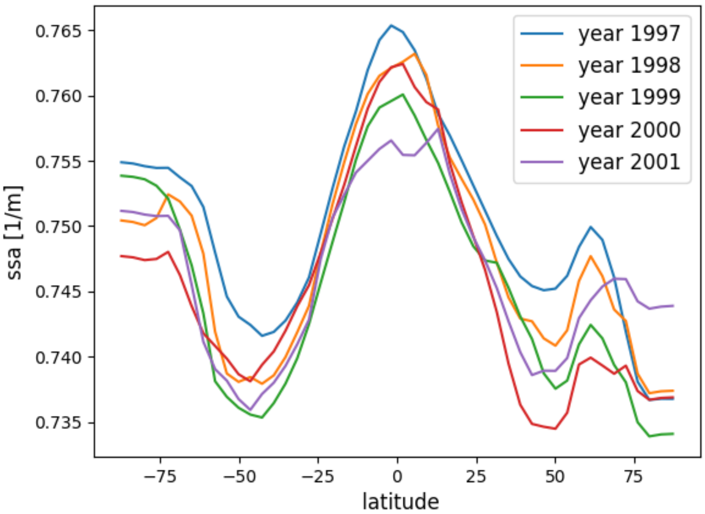


Figure 5: Distribution of *ssa* over latitude in models I and II for the fist five years of simulation.

Model III was only run for 5 Tg of soot. The smaller change in surface temperature makes it harder to distinguish natural fluctuations from phenomenons caused by the aerosol forcing. Especially regional changes are hard to account for in that small setup.

4.1 Changes near the Surface

In this section the change of temperature and subsequently sea ice cover caused by a huge amount of additional aerosols is investigated. As expected all models show an immediate drop in surface temperature. Additionally the change in precipitation is briefly discussed.

4.1.1 Surface Temperature, Sea Ice Coverage and Wind

As can be observed in figure 6 a significant drop in temperature occurs in model I and II. The absolute value of the maximal temperature drop is higher in model II. This can be explained by a significantly higher aerosol optical depth in model II. While in model I the *aod* in the first year is approximately 0.36 for all wavelengths, it's value averaged over wavelength is 0.56 for model II. The maximal change in annual mean temperature of model I is reached after 5 years and amounts to a drop of 4.8 K in comparison to the control run. In model II the greatest change, a 5.7 K cooling, can be observed after 3 years.

This is in line with previously conducted studies. In [2] the injection of 150 Tg soot into the atmosphere resulted in a maximal temperature drop of nearly 10 K after 3 years.

17 years later the mean surface temperature is nearly back to normal in both model I and II.

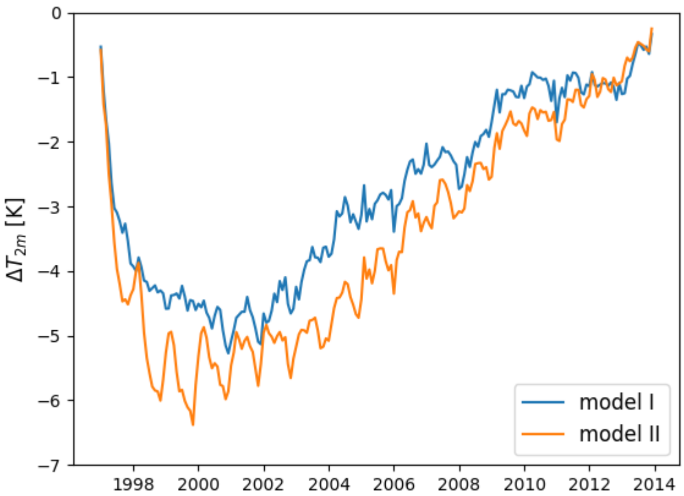


Figure 6: Temperature difference between control run and model I and II two meter above surface over time.

Higher seasonal differences can be observed in model II. One possible explanation lies in the wavelength dependence of *aod*. While the parameter is uniformly distributed over all wavelengths in model I, this is not the case for model II as can be seen in figure 7. So seasonal changes in incoming solar radiation in the lower stratosphere would be observable in modelling results of model II. The seasonal dependence diminishes over time. One explanation by [2] is that the oceans take time to adapt to cooling, while land masses respond instantly. This makes for a higher seasonal difference in forcing. However, that doesn't account for the fact that there is next to no seasonal difference in model I.

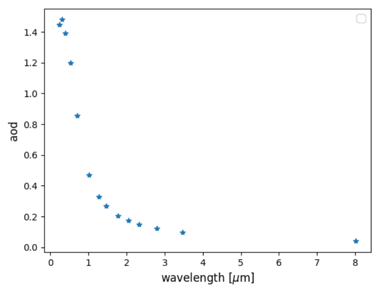
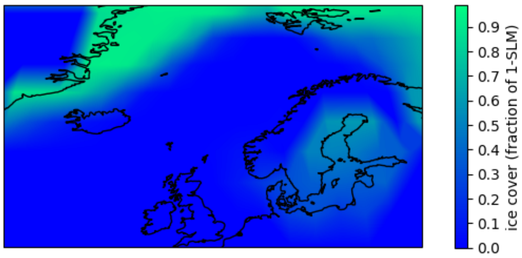


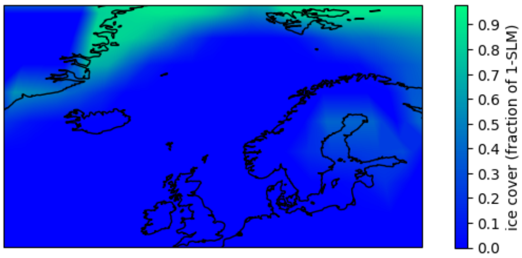
Figure 7: Distribution of *aod* over wavelength model II in 1997.

Model II is assumed to be more accurate. Not only are all optical properties but *aod* calculated. More importantly dependence on wavelength is taken into account. To investigate other changes only the output of II is considered.

A decreasing surface temperature leads to a general increase in sea ice coverage. Figure 8 visualised how this would effect northern Europe. Stockholm for example would be likely to experience frozen areas all around the year. That Iceland would still remain ice free the whole year, at least at the southern coast, suggests that the Gulf Stream remains intact.



(a) control



(b) model II

Figure 8: Sea ice coverage of northern Europe in year 1999.

In figure 9 a more global perspective is shown. It is apparent that without warming caused by the Gulf Stream, North America experiences a greater increase of permanent ice surfaces. Positive change in sea ice is happening in latitudes were Europe remains unaffected.

Meanwhile sea ice coverage even decreases in North East Asia. This suggests a warming in that area that will be investigated next.

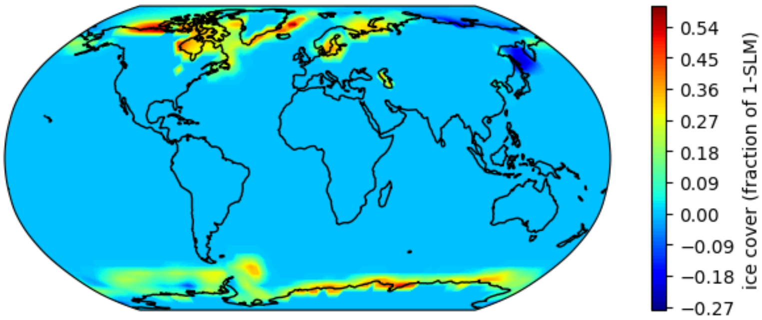


Figure 9: Difference of sea ice coverage between control run given in percent points in year 1999.

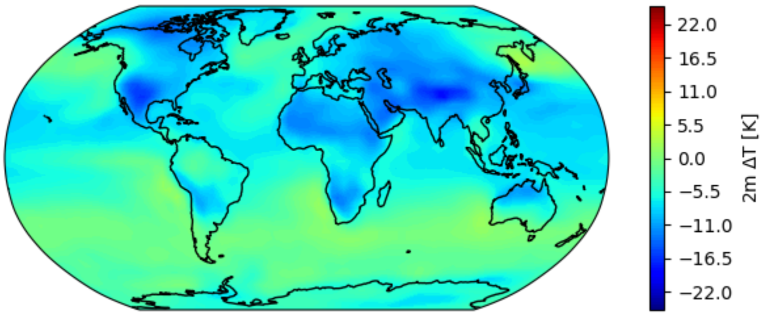
As can be seen in figure 10a change in temperature is location-dependent. The southern hemisphere has considerable less land mass. Due to the fact that oceans have a high heat capacity and therefore take longer to respond to forcing, it is not surprising that the cooling effect is greater in the northern hemisphere.

As it seams to be true in general that land mass cools significantly more than oceans, there are exceptions to that pattern. As can be seen in figure 10a the most significant anomalies are the northern half of South America, the Pacific Ocean and Antarctica. Looking simultaneously at change in wind speed shown in figure 10b, it is noticeable that in the first two regions considerable changes in wind speed can be observed. That strongly suggests a correlation between these two parameters. However, initially it is not clear if the local temperature difference causes a change in wind pattern or if change in wind speed is responsible for diverging temperature behaviour.

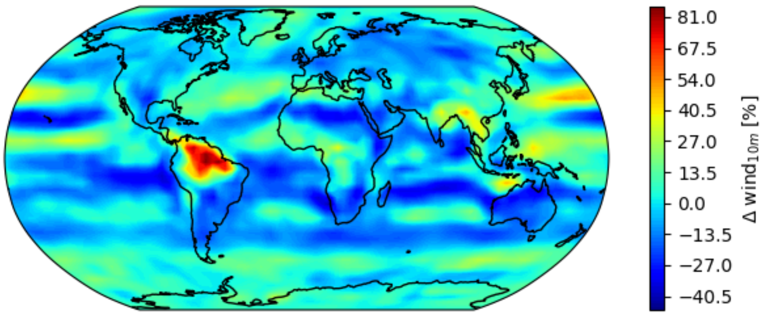
I would argue, that a mixture of both could be true. The temperature gradient caused by rapid cooling of land mass while the ocean is still relatively unaffected is very likely to trigger a change in the global wind pattern. For some reason this changes are especially prominent in North America and the Pacific. This could have something to do with two big mountain chains in these regions: the Andes and the Himalaya. Looking closely at figure 10b the wind difference changes along that lines.

There is thus a strong suspicion that cold air from the Eurasian continent is blown out on the Pacific in latitudes of about -10 °N to 50 °N. Meanwhile, it looks like warm air from the ocean causes a higher temperature in North East Asia, South America, North Australia and East Africa.

That there is no great change in Antarctica, I would attribute to the fact that the albedo is very high there anyways. Even under normal circumstances solar radiation doesn't induce a great heating of the land. So lack of it doesn't matter so much. However, in [2] even a warming could be observed, which is not the case here. It is caused by the stratosphere heating up significantly and therefore triggering a change in wind patterns. As will be discussed in section 4.2 it matters a lot for the stratospheric response if sulfur or soot is released. This is very likely the reason why local changes in this model are different than the result in [2].



(a) Change in temperature two meter above surface.



(b) Relative change in wind speed ten meter above surface.

Figure 10: Surface temperature difference and relative change in surface wind speed between control run and model II in year 1999.

It has to be kept in mind that in our setup only one control run and one forced run is conducted. So by comparing them, internal variability is added up. The effect of that is even higher if the changes attributed to aerosol forcing are expected to be smaller. This is the case if a small amount of soot is injected.

Nevertheless, even for a forcing induced by 5 Tg additional soot in the atmosphere, a definite drop in temperature can be observed as is shown in figure 11. A minimum in mean temperature difference over one year can be found after 5 years and amounts to -0.95 K. However, there is not much change in mean temperature over the year from the years 1999 (-0.93 K) and 2001 (-0.95). It is save to say that a temperature drop of about 1 K can be observed in course of the fist few years. This are the same results as presented in [1] for a forcing induced by 5 Tg additional soot in the stratosphere.

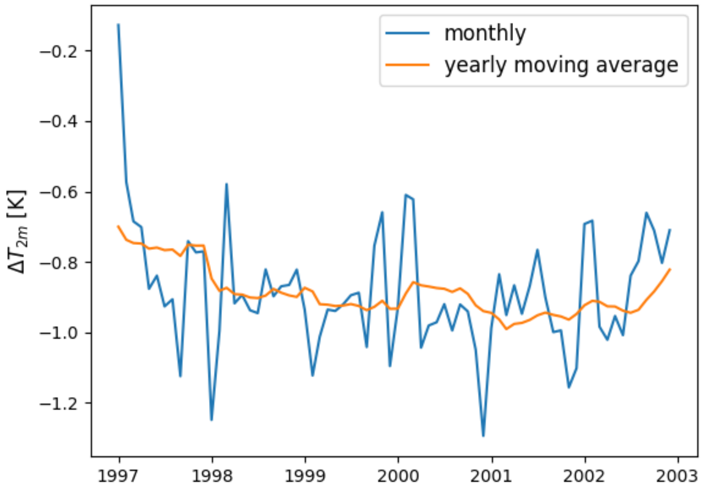
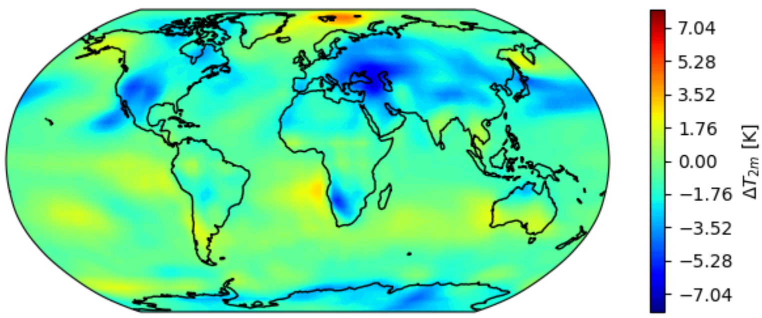
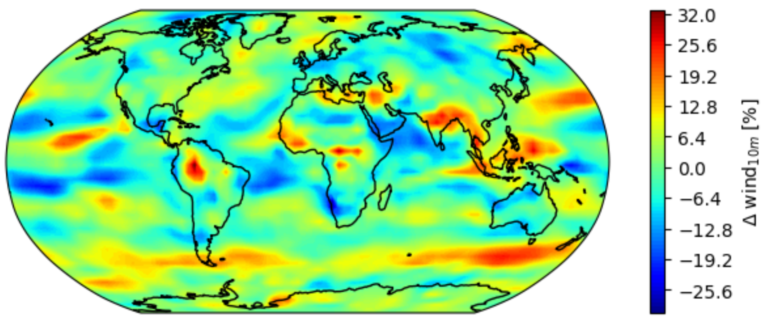


Figure 11: Temperature difference between control run and model III two meter above surface over time.

The spacial distributions are quite different for III as can be observed in figure 12a. There is still no warming of Antarctica as described in [2]. However, there is a clear warming in the northern polar circle. Both the changes in temperature and wind pattern, depicted in figure 12b, are much more heterogeneous as it was the case in model II. It has to be kept in mind that the scales in figure 12a and 10 are not the same. This is due to the fact that the amount of injected aerosols was very different. However, it is possible to observe an overall trend. Wind changes appear to be much more chaotic in model III than they do in model II. The same goes for temperature. Again it can only be suggested that there is a correlation between wind change and temperature change, but it cannot be said if one causes the other, or if both are the outcome of at least one other factor. Most likely the answer is a mixture of all of the above.



(a) Change in temperature two meter above surface.



(b) Relative change in wind speed ten meter above surface.

Figure 12: Surface temperature difference and relative change in surface wind speed between control run and model III in year 1999.

Altogether it can be said, that all presented setups result in a drop in the mean surface temperature. However the spacial distribution of both wind and surface temperature difference turns out to be very different. Apart from the amount of injected aerosols, the distinction between model II and III is the single scattering albedo. It's influence will be analysed in section 4.2.

4.1.2 Precipitation

There is another effect caused by less shortwave radiation that reaches the surface leading to a reduction in solar heating. It induces a decrease in evaporation. A consequence is a considerable drop in precipitation around the globe. As can be seen in figure 13 the maximal reduction amount to about 12 % after 4 to 5 years. The same effect, but much stronger was observed by [2]. As in our setup the precipitation minimum occurs a little later than the minimum in surface temperature.

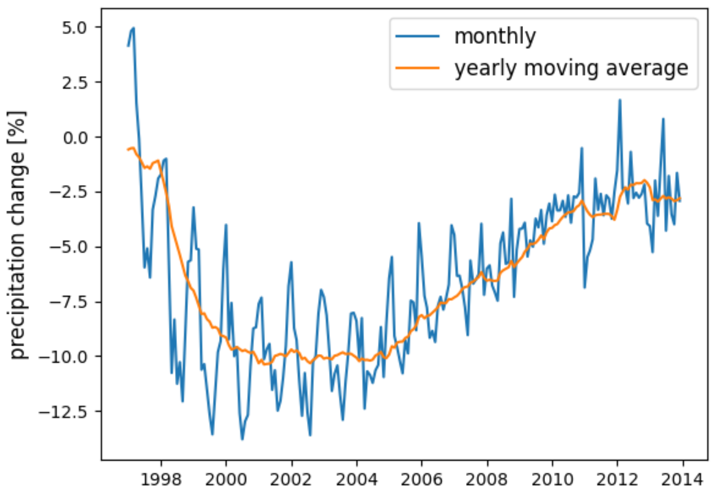


Figure 13: Change in precipitation of model II compared to the control run.

A similar observation of less magnitude can be made for model III. The results visualised in figure 14 are in line with what was found in [1] for an injection of 5 Tg soot.

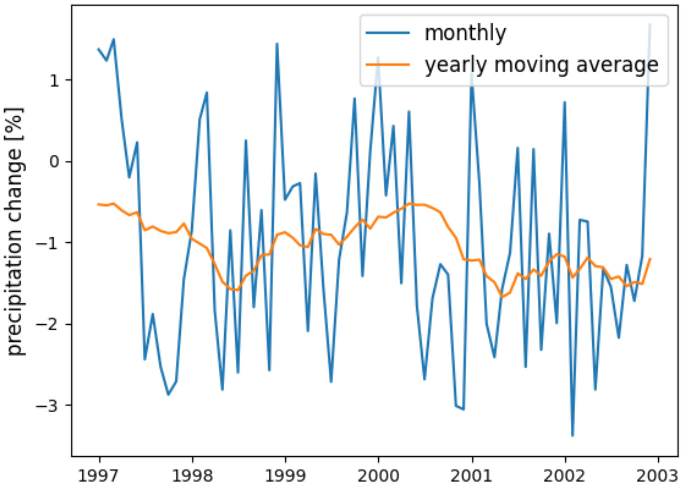


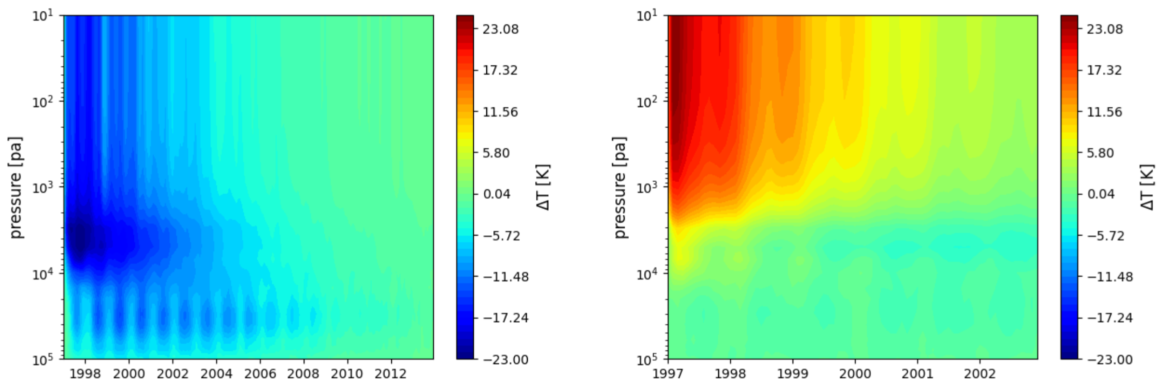
Figure 14: Change in precipitation of model III compared to the control run.

4.2 Stratospheric Temperature

In contrast to effects on surface temperature, changes in stratospheric temperature are vastly different for soot and sulfur particles.

The single scattering albedo is the same for models I and II. As already established model II is believed to be more accurate. Therefore only the results from the latter are discussed here, the trend is alike in model I.

In both [1] and [2] a stratospheric heating was observed. This is not the case for our model II as can be seen in figure 15a. In fact, even a cooling effect occurs. On the other hand, even 5 Tg soot cause a heating up to 23 K as can be seen in figure 15b. Note that both colour scales are the same in figure 15 even though the left plot shows a case with 14 times as much injected aerosols! This gives strong suspicion as to why numerical overflow issues kept appearing with a higher soot concentration. It also gives reason to a lot of the behaviour discussed above. For example why wind patterns are more affected relative to the magnitude of the conducted experiment.



(a) model II

(b) model III

Figure 15: Temperature distribution in the atmosphere.

The explanation for the contrasting heat signatures lies in different *ssa* values, which are shown in figure 16 for one year of the simulation. The variance in non-zero *ssa* values from volcanic eruptions is very small over the years. This is plausible because volcanoes release sulfur particles no matter in what year they erupt. Therefore it is reasonable to consider only one year as an example.

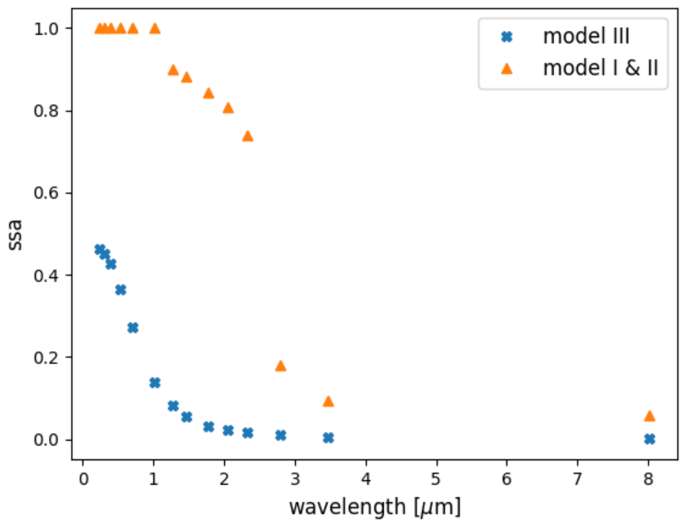


Figure 16: Orange triangles: mean *ssa* of all non zero values over altitude, latitude and time in year 1997 as used in model II. Blue crosses: calculated non zero *ssa* values for soot particles as used in model III.

It is apparent that the sulfur particles' interaction with shortwave radiation consists almost entirely of scattering, while soot aerosols absorb most of the radiation. Both scattering and absorption block sunlight and therefor reduce surface temperature. The effects on the stratosphere however are very different. The mechanism is comparable to a white and a black sunshade. Both will keep a person lying underneath cool, as most of the direct sunlight is blocked. But I don't want to be the one who has to touch the black sunshade to dismantle it.

4.3 Ozone

In theory a reduction of the ozone layer leads to a cooling of the stratosphere [12]. This could be a reason for the extreme cooling observed in figure 15a. However, it cannot be concluded what impact the change in ozone concentration really has. Therefore another experiment would be needed with the same configurations but the ozone reduction.

In our model ozone depletion depends exponentially on the amount of soot in the atmosphere. This arises the question if for a higher injection of soot, the diminishing ozone concentration could compensate some of the atmospheric heating observed in figure 15b. This cannot be answered here. It appears that depletion of ozone is caused by a heatup of the stratosphere. This however causes a negative temperature feedback, therefore a model with prescribed ozone may not be suitable for this investigation anyways.

5 Conclusion and Outlook

A lot of time and computation power was spent on figuring out how to get the model to run with the desired configurations. Unfortunately, no results were generated that model the initial scenario. Subsequently, the goal to find a realistic setup and run it on MPI-ESM1.2 could not be reached. Based on the results the question of what would happen to the climate after a huge nuclear assault cannot be answered. However, the conducted experiments still provide some insight into the topic.

5.1 Findings

All conducted experiments show a drop in surface temperature. Assuming the stratospheric ozone concentration has no significant impact on the surface temperature, the only other changes in all models are *aod* and *ext*. These are proportional to each other, so it essentially comes down to an increase in aerosol optical depth. It is safe to conclude that set parameter determines the mean temperature reduction. This is also in line with our theoretical understanding of the matter. A higher *aod*, means a higher proportion of solar radiation is blocked, which leads to less heat generation on the surface.

However, the special distribution of surface temperature reduction depends on what kind of particles are injected. This is reflected in the parameters *asy* and *ssa*. Especially the single scattering albedo seems to have a huge impact. A low *ssa*, or in other words highly absorbing particles, lead to a significant heatup of the stratosphere. This however could initiate atmospheric turbulence, which could lead to strong winds. These could contribute to the somehow unintuitive surface temperature distribution observed in the output of model III. However, this topic needs further investigation before anything can be concluded.

A reduction in precipitation is noticeable in all setups. This strengthens the theory that it is caused by a decrease in solar heating on the surface.

It is also worth mentioning that the setup for model III seems to be fairly good. Comparisons with an experiment that also investigated a release of 5 Tg soot [1] show very similar results.

5.2 Problems and Improvements

An obvious problem is that no experiment with the desired parameters was conducted successfully. This is most likely caused by numerical issues due to too high changes in time. A solution to that could be choosing smaller timesteps, but we didn't think of that until it was too late.

Another possibility would be to use a different climate model. This could also help overcoming the drawback that aerosols are prescribed and not modelled. In studies such as [3] this approach it taken. An advantage of modelling the aerosols instead of prescribing them is that effects like coagulation can be taken into account. These lead to different effective radii, which entail different aerosol optical properties. It would also be very interesting to not just assume a uniform distribution of soot around the globe, but investigate how soot particles spread over time.

At the moment altitudes where soot particles are assumed to accumulate are estimated. With a setup including dynamic aerosol modelling the elevation of the particles could be observed. This could change the heat forcing in the stratosphere more precisely over time. It would be very interesting to investigate if that would maybe lead to a warming of Antarctica which is described in literature but not observed here.

Ozone is prescribed too. In my opinion this is especially problematic, because theory leads to believe that the ozone concentration could be a feedback loop. Also there is no evidence that the currently used formula for ozone concentration is correct.

Another problem of the setup is that the control run as well as the experiments themselves are only conducted once and compared point wise (in time). The model is not deterministic, so the accuracy of the results are dependent on the inner variance of the model. However, it is save to say that at least a trend is observable.

The issues with model I are manifold. Looking at real life data shows that the assumption that aerosol properties are independent of the incoming wavelength is just not reasonable. Additionally the approximation of parameters is conducted in a very simplified manner. Last but not least *ssa* of the control run was used.

The last problem remains also for the improved model II. The control *ssa* differs in two ways from the calculated one. First its distribution is not constant across all latitudes. Maybe this could also contribute to the different temperature distributions observed for II and III. Though I doubt that the effect is too high compared to the second difference between the sulfur and soot particles, the *ssa*.

It was also not taken into account that a nuclear assault of that magnitude would decimate the earth's population significantly. People would not only die because of the explosions themselves, also of starvation caused by a reduction in agriculture. In the long run, this would decimate anthropogenic forcing. On the other hand, it may be worth investigating if CO_2 production caused by the fires and perish of plants in the aftermath would compensate that.

In my opinion it is not reasonable to use output generated for mainly scattering aerosols to draw conclusions for mainly absorbing aerosols. Still I learned a lot in course of that project and I think that's a nice outcome on itself.

A Code for Model II and III

Listing 1: Script to calculate optical properties and generate ncl script.

```
##########
# file: optical_properties.py
# requirements: template file 'template_change_aerosols.ncl'
#   in directory 'file_dir'
# output: ncl script to use on volcanic aerosol input files
##########

import numpy as np
import miepython as mie
from numpy import pi

# earth radius in m
R = 6.378*1e6

# pressure levels between 0 and 39
""" p_mid =
    1, 3, 7, 13, 22, 35, 52, 76, 108, 150,
    207, 283, 383, 516, 692, 922, 1224, 1619, 2133, 2802,
    3670, 4793, 6236, 8066, 10362, 13220, 16748, 21059, 26192, 32082,
    38675, 45908, 53672, 61799, 70056, 78139, 85673, 92219, 97287, 100368
"""
# intervall of soot height above sea level in m
h = np.array([2.,5.])*1e4
# pressure level
plev = np.array([27, 8])
# initial amount of soot in g
S0 = 70.*1e12
# refractive index
m = 1.8-0.67j
# density of black carbon in g*m^-3
rho = 2.*1e6
# radius of particle in m
r = 0.1*1e-6

# wavelengths
wl_lo = np.array([3077, 2500, 2151, 1942, 1626, 1299, 1242,
    778, 625, 442, 345, 263, 200, 3846])
wl_up = np.array([3846, 3077, 2500, 2151, 1942, 1626, 1299,
    1242, 778, 625, 442, 345, 263, 12195])
wl_mid = (wl_up + wl_lo)/2*1e-3 # scale for mie package

# define mie size parameter
x = 2*pi*r*1e6/wl_mid

# calculate optical parameters with Mie algorithm
Qext, Qsca, _ , asy = mie.mie(m,x)

# single scattering albedo
ssa = Qsca/Qext

# concentration: S*N0
N0 = 9./(16*pi**2*rho*r**3*((R+h[1])**3-(R+h[0])**3))
```

```
# extinction coefficient: S*sigext
sigext = pi*r**2*N0*Qext

# aerosol optical depth: S*aod
aod = (h[1]-h[0])*sigext

# print for control
print("initial values")
print("aod: ", aod*S0)
print("ext: ", sigext*S0)
print("ssa: ", ssa)
print("asy: ", asy)

# generate file change_aerosols.ncl
# from template template_change_aerosols.ncl
file_dir = '/home/ams/Studium/GCS/project_work/output/'
name = 'change_aerosols.ncl'
tmp_file = open(file_dir+'template_'+name, 'rt')
data = tmp_file.read()
data = data.replace('#PY_S0', str(S0))
data = data.replace('#PY_plev0', str(plev[0]))
data = data.replace('#PY_plev1', str(plev[1]))
data = data.replace('#PY_ext', '(/' + np.array2string(
    sigext, separator=',', max_line_width=10000)[1:-1]+'/)')
data = data.replace('#PY_ssa', '(/' + np.array2string(
    ssa, separator=', ', max_line_width=10000)[1:-1]+'/)')
data = data.replace('#PY_asy', '(/' + np.array2string(
    asy, separator=', ', max_line_width=10000)[1:-1]+'/)')
data = data.replace('#PY_aod', '(/' + np.array2string(
    aod, separator=', ', max_line_width=10000)[1:-1]+'/)')

tmp_file.close()
tmp_file = open(file_dir+name, 'wt')
tmp_file.write(data)
tmp_file.close()
```

Listing 2: ncl template

```
begin

year_start = 1997
year_stop = 2014
S0 = #PY_S0
plev0 = #PY_plev0
plev1 = #PY_plev1
ext = #PY_ext
ssa = #PY_ssa
asy = #PY_asy
aod = #PY_aod

year = year_start
do while(year.le.year_stop)
    print("year " + tostring(year))
    f = addfile("strat_aerosol_sw_T31_"+tostring(year)+".nc", "w")

    ; this variant seems to be quicker than doing 3 for loops
    ; even if values have to be written twice
```

```
    print("set non-zero ext and aod")
   m = 1 ; counter month
   do while(m.le.12)
       S = S0 * exp(-(tofloat(m)/12.0+(year-year_start))/4.6)
       w = 0 ; counter wavelength band
       do while(w.le.13)
           f->ext(m-1,:,:,w) = ext(w)*S(:)
           f->aod(m-1,:,w) = aod(w)*S(:)
           w = w+1
       end do
       m = m+1
   end do

    print("set non-zero ssa and asy")
    ; independent of amount of soot
   w = 0 ; counter wavelength band
   do while(w.le.13)
       f->ssa(:,:,:,w) = ssa(w)*1.0 ; to avoid a warning
       f->asy(:,:,:,w) = asy(w)*1.0
       w = w+1
   end do

    print("set zero values")
   p = 0 ; counter pressure level
   do while(p.le.plev1)
       f->ext(:,p,:,:) = 0
       f->ssa(:,p,:,:) = 0
       f->asy(:,p,:,:) = 0
       f->aod(:,p,:) = 0
       p = p+1
   end do
   p = plev0
   do while(p.le.39)
       f->ext(:,p,:,:) = 0
       f->ssa(:,p,:,:) = 0
       f->asy(:,p,:,:) = 0
       f->aod(:,p,:) = 0
       p = p+1
   end do

year = year+1
end do

end
```

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