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Black carbon absorption effects on cloud cover, review and synthesis

D. Koch[1,2] **and A. Del Genio**[2]

[1]Columbia University, New York, NY, USA
[2]NASA GISS, New York, NY, USA

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Correspondence to: D. Koch (dkoch@giss.nasa.gov)

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Abstract

Absorbing aerosols (AA's) such as black carbon (BC) or dust absorb incoming solar radiation, perturb the temperature structure of the atmosphere, and influence cloud cover. Previous studies have described conditions where AA's either increase or decrease cloud cover. The effect depends on several factors, including the altitude of the AA relative to the cloud and on the cloud type. Cloud cover is decreased if the AA's are embedded in the cloud layer. AA's below cloud may enhance convection and cloud cover. AA's over cloud-level stabilize the underlying layer and tend to enhance stratocumulus clouds but may reduce cumulus clouds. AA's can also promote cloud cover in convergent regions as they enhance deep convection and low level convergence as it draws in moisture from ocean to land regions. Most global model studies indicate a regional variation in the cloud response but generally increased cloud cover over oceans and some land regions, with net increased low-level and/or reduced upper level cloud cover. The result is net negative radiative forcing from cloud response to AA's. In some of these climate model studies, the cooling effect of BC due to cloud changes was strong enough to essentially cancel the warming direct effects.

1 Introduction

Black carbon (BC), the light-absorbing component of carbonaceous aerosols, warms the atmosphere where it is suspended and is therefore thought to contribute to global warming. Pollution sources of BC include incomplete combustion of fossil fuels such as diesel and coal, and burning of biofuels and open biomass. BC also absorbs radiation after it is deposited on snow and promotes snow-melt (e.g. Flanner et al., 2007), further contributing to warming.

BC also affects clouds in ways that are poorly understood. These effects may be warming or cooling and are potentially similar in magnitude to the better-understood warming direct and snow-albedo effects. Like other aerosols, BC contributes to the number of cloud condensation nuclei and therefore affects cloud cover and lifetime. BC may also act as ice nuclei and therefore change ice or mixed-phase clouds. Finally, absorbing aerosols perturb the temperature gradient in the atmosphere and therefore affect cloud distributions. This study focuses on these latter BC radiative effects on clouds rather than microphysical (indirect) effects.

Atmospheric BC suspended near clouds has been thought to contribute to cloud evaporation, originally termed the "semi-direct effect" (Hansen et al., 1997). This loss of cloud cover exacerbates the warming impact of BC. However there are numerous studies that describe additional mechanisms whereby BC may either reduce or increase cloud cover. Reduction/enhancement of low-level cloud cover has a positive/negative forcing and therefore a warming/cooling effect. The reverse applies to high-level clouds.

As efforts are under consideration for using BC as a global warming mitigation target, it is important that we have a better understanding of its impacts on clouds. Since radiative forcing from cloud changes is quite large, the temperature changes associated with cloud redistribution from BC radiative effects are potentially much larger than those due to BC direct effects.

The purpose here is to review literature on how BC and other absorbing aerosols (AA's), such as soil dust, affect clouds through their absorptive heating of the atmosphere. Since many of these studies are regional in scale, we attempt to find common conditions under which absorbing aerosols either enhance or reduce cloud cover. Finally we discuss results from global model studies and uncertainties in these model predictions.

2 Cloud burn-off

Absorbing aerosols embedded in or near a cloud layer heat the layer and promote cloud evaporation. This is the original "semi-direct effect" as first described by Hansen et al. (1997). This effect has been reproduced and documented in several studies, especially in cloud-resolving models.

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Ackerman et al. (2000) performed large-eddy-simulation (LES) studies of trade cumulus clouds for conditions appropriate for the Indian Ocean Experiment (INDOEX). They showed that boundary layer AA's reduced the relative humidity, increased the rate of cumulus cloud detrainment, and stabilized the boundary layer, resulting in decreased daytime cloud fraction by 5–10%.

Hill and Dobbie (2008) used the UK Met Office's LES model with bin-resolved cloud microphysics to study the impact of an absorbing layer of aerosols in a boundary layer on non-precipitating marine stratocumulus. They showed that the BC layer reduced liquid water path (LWP), reduced cloud-top altitude, increased cloud-base altitude and caused a positive semi-direct radiative forcing change. They also showed that increased cloud condensation nuclei, not necessarily due to the absorbing aerosols but due to increased particles overall, reduced droplet size and further increased evaporation, enhanced boundary-layer dynamics, cloud-top entrainment and cloud evaporation.

3 Altitude dependence of AA's and cloud layers

3.1 AA below cloud

Although BC within the cloud layer enhances cloud evaporation, BC located above or below cloud can enhance cloud cover. BC below cloud level enhanced cloud cover in the 3-dimensional Eulerian cloud-resolving model studies of McFarquhar and Wang (2006) for trade wind cumuli under INDOEX conditions. For experiments with the AA's in the cloud layer, the clouds dissipated as in the studies in Sect. 2. However for AA's below the cloud level, the heating below cloud enhanced vertical motions and increased cloud cover and liquid water path (LWP).

Similar results were obtained over land in LES experiments of Feingold et al. (2005) for the effect of Amazon smoke on clouds in September. Smoke emitted at cloud level decreased cloud cover mostly due to its stabilization of the cloud layer. Smoke also re-

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duced cloud cover due to decreased surface and sensible heat fluxes. However smoke emitted at the surface could destabilize the surface layer and increased convection and cloud cover. Thus, for these experiments, the cloud cover response depends upon the relative strength of destabilization convective enhancement that increases cloud cover and surface flux reduction that reduces cloud cover. Feingold et al. (2005) point out that surface fluxes are less variable over ocean; thus over land the surface conditions play a critical role in determining cloud response.

3.2 AA above stratocumulus clouds

Absorbing aerosols aloft increase atmospheric stability. Increased stability over stratocumulus clouds strengthens the inversion and reduces cloud-top entrainment of overlying dry air and tends to enhance the underlying clouds.

The altitude influence of AA's on clouds was demonstrated clearly in the LES experiments of Johnson et al. (2004), designed to study the effect of AA on subtropical marine stratocumulus clouds. As in Sect. 2, AA's within the boundary layer where the clouds reside decreased LWP, and resulted in positive semi-direct forcing. AA's both within and above cloud layer also resulted in cloud reduction, although to a lesser extent. However, an AA layer above cloud increased cloud cover, as it increased the contrast in potential temperature across the inversion, decreased entrainment rate, and caused a shallower, moister boundary layer with higher LWP. The AA's above cloud also reduced downwelling solar flux, possibly also increasing LWP. The aerosol cloud-radiative effect in this case was nearly equal and opposite to the direct effect. The authors note that global models do not simulate stratocumulus clouds well and may therefore inadequately reproduce these effects.

Brioude et al. (2009) presented a field example of biomass burning smoke over marine stratocumulus off the coast of California in summertime. They analyzed cloud data from Geostationary Operational Environmental Satellite (GOES) and MODIS, and derived biomass burning aerosol using a passive biomass burning tracer in the FLEXPART model. They found that biomass burning aerosols enhance cloud cover, especially for high-humidity conditions and for low lower tropospheric stability conditions (LTS), since the aerosols have the effect of increasing LTS. Higher LTS decreases vertical entrainment of dry air from above, increasing boundary layer (BL) relative humidity. Most of the region had AA's above the BL and increased cloud cover. However farther from the continent, at the edge of the study domain, AA's occurred within the BL and had decreased cloud cover.

3.3 AA above cumulus clouds

While the stabilizing effect of AA's aloft can enhance stratocumulus cloud cover, they may on the other hand suppress cumulus cloud development. Over land the AA's may also reduce evaporation from the surface and therefore moisture available for cloud formation.

Fan et al. (2008) simulated cloud reduction by AA's in Houston in August due mostly to the stabilizing effects of the AA's. They used the two-dimensional cloud-resolving Goddard Cloud Ensemble model with spectral-bin cloud microphysics coupled to a land-surface model. The aerosols cooled the surface, increasing underlying relative humidity, but the heating aloft decreased the relative humidity there. The heating decreased the temperature lapse rate, leading to a more stable atmosphere and decreased convection. Decreased precipitation also resulted from the shallower clouds. Compared to simulation with no aerosols, cloud fraction and cloud optical depth decreased by about 20%; LWP, IWP and precipitation also decreased. The cloud forcing change was about 10 $\mathrm{W\,m^{-2}}$, and direct effect 2.2 $\mathrm{W\,m^{-2}}$.

MODIS observational studies for the biomass burning season of the Amazon by Koren et al. (2004, 2008) also demonstrated cumulus cloud cover reduction due to increased smoke. They argued that the smoke plumes stabilized the BL, reducing convective activity and BL cloud formation. The smoke also reduced radiation penetration to the canopy, therefore decreasing evapotranspiration, an important source of moisture in this region. In addition, the smoke particles may compete for available moisture, decreasing the ability of the air to reach supersaturation. Koren et al. (2008)

reported a transition from smoke enhancement of cloud cover at low aerosol optical depth (AOD), which they argued was microphysical invigoration, to smoke radiative reduction of cloud cover as AOD increased. They also showed that cloud inhibition due to radiative absorption effects was stronger for smaller initial cloud cover.

4 Enhanced low-level regional convergence over land

Several studies have described increases in clouds and/or precipitation in regions where lofted AA's have enhanced upper level convective activity, and this has promoted larger scale circulation that includes low-level convergence carrying in moist air. In many cases the studies describe reductions in moisture or cloud cover in adjacent regions, so it is not clear how much the effect increases cloud cover overall. Three regions where this has been studied include Africa, south Asia and Southeast Asia. Most of the studies are performed with global models but with focus on particular regional changes.

Over Africa, Stephens et al. (2004) used a 2-D cloud-resolving model to study the effect of a lofted dust layer on tropical convection for either dry or moist atmospheric conditions. The experiments with lofted dust produced low-level convergent flow toward the dust region and enhanced convection. The non-dusty adjacent regions experienced reduced convection. The enhanced convection was greater for an experiment with moist mid-troposphere conditions compared to one initialized in a drier atmosphere, demonstrating the importance of initial hydrologic conditions.

Miller et al. (2004) also focused on dust effects on clouds in Africa, in the Western Sahara. They used the GISS climate model and found that dust loading caused increased low-level cloud cover and precipitation. They argued that in an arid region, where diabatic heating is overwhelmed by longwave cooling and is balanced by subsidence, dust absorption in an aerosol layer aloft could reverse the circulation, resulting in ascent and precipitation.

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Rudich et al. (2003) analyzed AVHRR satellite observations of smoke and clouds from the Kuwait oil fires of March 2001. They found that the absorbing aerosols in the heaviest smoke plumes cooled the underlying surface, but heated the plume and induced convective clouds above the layer of the smoke plume.

Many studies have considered the effects of AA's on the South Asian monsoon. Lau et al. (2006) used the NASA finite volume GCM with aerosol optical depth from the GOCART model, to study dust and black carbon effects on the Indian monsoon. They showed that lofted dust accumulates beside the Tibetan Plateau and creates an elevated heat pump that draws in moisture from the Indian Ocean. Black carbon pollution may contribute to intensification of the Indian monsoon and the sea level pressure anomaly pattern, with consequent weakening of the East Asian monsoon.

A similar result was obtained by Randles and Ramaswamy (2008) using the GFDL GCM with offline aerosols from MOZART-2 and fixed SSTs. They found that for sufficiently high aerosol extinction optical depths, cloud amount increased as absorption increased. The aerosols warmed the atmosphere but cooled the surface and reduced latent and sensible heat fluxes. Low-level convergence and increased vertical velocity overcame the stabilization effects of the aerosols so that monsoon circulation and precipitation were enhanced over northwestern India.

Chung et al. (2002) also modeled the INDOEX region, with the NCAR CCM3 and fixed SSTs. They found that pollution haze cooled the land surface, warmed the atmosphere, stabilized the boundary layer, and reduced evaporation and sensible heat flux from the land. The forcing weakened the north-south temperature gradient and caused a northward shift of the ITCZ. This resulted in enhanced convective precipitation, latent heat release, and a further increase in convergence. The shift in precipitation also caused reduced rainfall in adjacent regions, such as southwest Asia. Decreased evaporation over the oceans from the haze caused decreased precipitation over the rest of the Tropics.

However, Ramanathan et al. (2005) found a different impact of AA's on the south Asian monsoon, using the NCAR coupled ocean-atmosphere model and transient

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simulations from 1930 to 2000. Absorbing aerosols followed emission trends scaled to observed present-day levels. As in the previous studies, they found that the AA's in the region reduced surface radiation, surface temperatures, evaporation from the surface and rainfall. The reduced evaporation resulted from the reduction in vertical temperature gradient (warming aerosols aloft and cooling at the surface), which inhibited convection, increasing relative humidity near the surface but decreasing RH aloft; the higher relative humidity near the surface reduced evaporation. The aerosols over northern India reduced the meridional temperature gradient and in this study, weakened the monsoon. Little change in cloud cover occurred in the simulations. Compared with the other studies in this region, this model included a full climate response; it also focused on BC and not dust as in Lau et al. (2006). Even though Ramanathan et al. (2005) found reduced precipitation in south Asia, a follow-up study by Chung and Ramanathan (2006) showed that the weakened SST gradient and meridional circulation also resulted in enhanced precipitation over sub-Saharan Africa. This second study also confirmed some of the results of other models for the region, that AA's in Southeast Asia cause diabatic warming, upwelling and enhanced precipitation locally; however this effect was overwhelmed by the SST gradient effect that weakened the monsoon in India.

In general these studies on the Indian monsoon region focused more on changes in precipitation rather than cloud cover. However Norris (2001) analyzed surface-based cloud cover observations and found an overall increase of low-level cloud cover in the Indian Ocean during January to April from 1952 to 1996. The mechanism for this increase however was not apparent.

In the Southeast Asian region, Menon et al. (2002) used the GISS GCM with fixed SSTs to see how regional AA affects that region's climate. They found that the region with greatest pollution had enhanced convective upwelling, increased precipitation and increased cloud cover. The precipitation changes in southern China agree with observations while the enhanced cloud cover apparently does not. Zhang et al. (2009) performed a similar experiment with the NCAR CAM3 model and obtained opposite results for the regional distribution of rain and cloud. It is not clear how the single scattering albedo or vertical distribution of the aerosols compare in these studies.

To some extent the conflicting results from different global models for particular regions may result from natural variability within their noisy climate systems as well as differences in their general cloud responses to forcing changes as discussed below.

5 Summary of regional variations in cloud response to AA's

From the studies on how clouds respond to AA's, some general patterns have emerged, summarized in Fig. 1. AA's embedded within cloud layers or potential cloud layers reduce cloud cover due to their heating and reduction of relative humidity (case 4). The altitude of AA's relative to a cloud or potential cloud layer plays a critical role in determining the cloud response. AA's below cloud generally promote convective activity and enhance cloud cover (case 6). AA's aloft stabilize the boundary layer and may promote cloud cover for some conditions such as for marine stratocumulus as they stabilize the boundary layer and reduce mixing with dry air above (e.g. Johnson et al., 2004; Brioude et al., 2009; case 2). However AA over shallow cumulus clouds inhibits cloud development (e.g. Koren et al., 2004) as they stabilize the surface layer and reduce surface evaporation (case 1). On the other hand, in some land regions, lofted AA's may enhance upper level convection, promoting low-level convergence that could carry moisture and increase clouds (case 3). An additional case is the reduction of high-level clouds found in some global models and is discussed below (case 5); this would also cause a negative forcing.

6 Impacts of AA on clouds in global models

Considering the results of the regional studies, with AA's leading to cloud cover increase for some conditions but decrease for others, it is not surprising that global model studies find both cloud reduction and enhancement.

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The global BC model study of Wang (2004) found that the net cloud-cover change was increased by BC. This study included both Qflux climate and fixed SST simulations, using the CSM-NCAR model, both with and without BC. The BC effects on clouds were much stronger for the Qflux simulations. He found that BC enhanced convective activity and cloud cover in the northern branch of the ITCZ, with smaller magnitude of reduction in clouds and convective activity in the Southern Hemisphere. The enhancement is especially large in the Pacific and over India. Significant changes were also found for precipitation, meridional heat transport, surface heat fluxes and boundary layer height. He found a much larger reduction in surface radiation for the Qflux simulation due mostly to increased low-level cloud cover. Furthermore, the top-of-atmosphere BC all-sky forcing was 30% less, or $-0.10\,\mathrm{W\,m^{-2}}$ for the Qflux compared with the SST simulation. The cloud cover changes varied greatly with region, with increases occurring e.g. over northern Eurasia as well as the northern branch of the ITCZ mentioned above. Overall, BC was not found to cause significant warming in the climate experiments apparently due to the compensatory cloud-cover and hydrological changes.

A similar result was found in the transient twentieth-century coupled climate experiments of Koch et al. (2010) for a sensitivity experiment in which pollution BC was set to zero from 1970 to 2000 and the climate compared to the case with all sources. Pollution BC includes all fossil and biofuel sources and changes in biomass burning since the year 1890, mostly tropical and African-grassland. In the BC-reduction experiments, the climate cooled only an average of about $-0.03\,^{\circ}\mathrm{C}$ during the three decades, in part apparently due to concurrently increased cloud cover. Figure 2 shows results from the experiments, for the changes in radiation, BC concentrations and cloud between the experiment with all sources and the BC-reduction experiment. BC reduction caused a small +0.03% increase in cloud cover (Fig. 2a) with larger increase of +0.15% for low-level clouds (Fig. 2b) and decreased high-level clouds (Fig. 2c). Both the increased low and decreased high cloud cover contribute a negative TOA radiative forcing change. As shown in Fig. 2d, the net forcing is close to zero, so that the cloud negative forcing essentially negates the positive direct BC forcing change (calculated from the radiation change due to BC at each radiation timestep) of +0.3 Wm^{-2}. Figure 2f, g shows the changes in BC concentrations at low and low-mid levels of the troposphere. Near source regions, the concentrations are larger at the surface, but in the remote oceanic regions the concentrations increase aloft. Low-level cloud cover increases over these oceanic regions, qualitatively consistent with the studies showing cloud cover increase beneath lofted AA's over the oceans. Upper-level clouds are reduced especially in biomass burning regions, an effect seen in some other global model studies discussed below. Figure 2h shows the relative humidity in model layer 1 as an indication of surface moisture conditions. To some extent, over land, moist regions have increased low-level cloud cover (Fig. 2b, h) including the eastern US, western Europe and northeastern Russia, while dry regions have reduced cloud cover, such as eastern Europe.

Roeckner et al. (2006) performed future transient climate simulations in the ECHAM5 model, beginning in the 20th century and going to 2050. Two futures were considered, the first had increased carbonaceous aerosol emissions (37% BC and 25% particulate organic matter) using an SRES A1B estimate; the second had no change in carbonaceous aerosol emissions after year 2000. For the A1B, carbonaceous aerosols decreased from Europe and China but increased in many low-latitude regions, especially African biomass burning. Similar to Wang (2004) and Koch et al. (2010), A1B did not warm compared with the control, and actually cooled in regions where BC increased strongly, such as the African biomass burning region, the Indian subcontinent and the Atlantic Ocean biomass burning outflow region. Hydrologic changes were strong in these regions, with enhanced precipitation, increased soil moisture, and increased liquid water path over India and Africa. Over the Atlantic Ocean off the coast of Africa, increased cloud cover and liquid water path also occurred as BC above the boundary layer stabilized the boundary layer and cooled the surface. Roeckner et al. (2006) noted that these hydrological changes are much more likely in a future warming climate and are not necessarily expected for 20th century conditions.

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In another GISS climate model investigation, Perlwitz and Miller (2010) studied how dust affects cloud cover. These experiments used a slab ocean and dust with varying optical properties. They showed that for sufficiently large dust AOD and absorption, the net effect is enhanced cloud cover for all seasons except winter. Cloud cover decreased over Eurasia, western North America and South America. The largest cloud cover enhancement occurred over oceans, central Africa, the Arabian peninsula, India and southeastern Asia. Over land, cloud cover increased where the absorbing dust enhanced specific humidity due to increased moisture convergence driven by the dust heating. Overall this effect exceeded the reduced humidity that results from dust absorption enhancing atmospheric evaporation. For scattering (low absorption) dust, for small dust AOD, and for winter conditions, the cloud cover enhancement responses were weak or even reversed.

Allen and Sherwood (2010) found a net +0.1 $\mathrm{W m^{-2}}$ positive aerosol-radiative cloud effect in another study using the NCAR CAM3 model with mixed-layer ocean. In this study both scattering and absorbing aerosols are prescribed from Chung et al. (2002) and aerosol heating is uniformly placed in the lowest 3 kilometers of the model. The study found a strong land-sea contrast in cloud response, with increased cloud cover over oceans and cloud loss over land, mostly at mid-level. Over land the dominant aerosol effect is reduction of relative humidity, while over oceans the aerosol stabilization of the boundary layer tends to enhance cloud cover. The GCM used here is the same as in Wang (2004), however here AA's are not isolated from other aerosols types and the aerosol effects are placed uniformly in the lower troposphere.

Some global model studies report reduction of upper level clouds due to lofted AA. The resulting forcing is also negative. Presumably this results from cloud burn-off or from reduction of relative humidity at high altitudes, associated with lofted AA.

For example, Penner et al. (2003) found a net negative forcing cloud response to carbonaceous aerosols in the GRANTOUR GCM due mostly to loss of high-level clouds. They calculated forcing by taking the difference between relaxed forcing, or forcing change taken at TOA averaged over the simulation, and the instantaneous TOA

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radiative forcing; the former includes the cloud response while the latter does not. They found that aerosols injected at mid-tropospheric levels enhanced low-level and also reduced upper-level cloud cover, with both contributing to a negative cloud-radiative effect. The radiative cloud effect due to biomass burning aerosols is $-0.37\,\mathrm{Wm}^{-2}$ in the experiments, with small response for non-biomass burning aerosols. Aerosols injected near the surface on the other hand reduced cloudiness.

Roberts and Jones (2004) performed Qflux climate experiments in the Hadley Centre Climate model, isolating the effects of BC on climate. They found the climate sensitivity of BC to be less (62%) than for CO_2. The reason seemed to be reduction in high-altitude clouds caused by BC. Although not discussed in their paper, their Fig. 8a also suggests an increase in low-level clouds due to BC.

Menon and Del Genio (2007) also report a negative semi-direct effect of $-0.08\,\mathrm{Wm}^{-2}$ in their (fixed SST) GISS simulations due to decreased long-wave cloud forcing change associated with loss of high-level clouds mostly in biomass burning regions.

There are several global climate model studies that have reported BC efficacy, or the temperature change per forcing relative to that for CO_2, less than 1. In some cases the studies indicate that cloud changes are responsible for the low efficacy. Hansen et al. (2005) showed that BC efficacy decreases with higher BC altitude in the GISS model. They argued that low-altitude BC reduces cloud cover, however BC above the boundary layer inhibits underlying convection and therefore enhances large scale cloud cover. This cloud enhancement greatly reduces the warming direct effect of the BC. Yoshimori and Broccoli (2007) also found BC efficacy less than 1 due to negative cloud response since BC caused increased low-level cloud cover but decreased mid-upper level cloud cover.

If we assume, as an upper limit, that the reduced BC efficacy is due to cloud cover changes, we may infer maximum BC radiative-cloud forcing estimates for these models. Hansen et al. (2005) found BC efficacy of 0.78 for fossil fuel BC and 0.58 for biomass burning BC for forcings of $0.49\,\mathrm{Wm}^{-2}$ and $0.19\,\mathrm{Wm}^{-2}$. If the reduced efficacy were due to cloud changes, the cloud forcing changes are -0.11 and $-0.08\,\mathrm{Wm}^{-2}$

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respectively. Yoshimori and Broccoli (2007) had efficacy of 0.59 for BC forcing of 0.99 Wm^{-2} giving maximum cloud forcing of $-0.4\ Wm^{-2}$. Jones et al. (2007) report a BC efficacy of 0.71 for BC forcing of 0.39 Wm^{-2}, leading to inferred upper-limit cloud effect of $-0.11\ Wm^{-2}$. Chung and Seinfeld calculated a 0.70 efficacy for 0.33 Wm^{-2} forcing, giving a $-0.18\ Wm^{-2}$ upper limit cloud response. From these we have a range of -0.1 to $-0.4\ Wm^{-2}$ for maximum cloud forcing response to AAs.

7 Discussion

We have considered several mechanisms by which absorbing aerosols may either increase or decrease cloud cover. Although the original "semi-direct" effect was associated with reduction of low-level clouds and positive forcing, there are many studies describing cloud-cover increase from absorbing aerosols and negative radiative forcing. Global model studies have both enhanced and decreased cloud cover depending on region and conditions.

The sign of the cloud change depends on several factors. First is the altitude of the AA's relative to the cloud or potential cloud level. For AA's within the cloud layer, absorptive heating burns off the clouds and moisture. AA's below cloud level can enhance convective activity and increase cloud cover. AA's above cloud-level stabilize the underlying layer and can result in either decreased or increased cloud depending on cloud type and underlying conditions. AA's above stratocumulus clouds tend to enhance cloud cover. AA's above shallow cumulus in oceanic regions might also strengthen the inversion and promote transition to a stratocumulus regime with increased cloud cover (e.g. Stevens et al., 2001), although this was not demonstrated in the AA-cloud studies presented here. Since AA's are derived from land-sources and are lofted before transport over the oceans, they tend to reside above cloud level over oceans (e.g. Fig. 2f, g) and would therefore typically promote clouds. However for AA's above low cumulus clouds over land, the enhanced stability can reduce convective cloud formation. Over land the AA's blocking of incoming solar radiation also

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reduces surface evaporation that can reduce moisture available for cloud formation. Finally, AA's advected into a convergent region can enhance deep convection and low-level convergence, potentially drawing in moist air from oceanic regions. An additional effect found in some global model studies is reduction of upper-level clouds, especially due to lofted biomass burning aerosols. This is apparently due to decreased upper-level humidity and/or cloud burn-off (e.g. Penner et al., 2003). Since the clouds are high-altitude the radiative forcing is negative.

The cloud response to AA's also depends upon the aerosol optical properties. The AOD must be sufficiently large, and the cloud response increases for lower single scattering albedo, i.e. for more absorbing particles. In some studies, replacing AA's with scattering aerosols would reduce or reverse the influence on the cloud cover (e.g. Johnson et al., 2004; Perlwitz and Miller, 2010; Wang, 2004). Koch et al. (2009) showed that most global BC models have underestimated their BC absorption optical depth; this would cause the models to underestimate effects of BC on clouds. Note that most of the cloud-scale models discussed here used simplified assumptions about the aerosol optical properties, such as specifying single scattering albedo and AOD and may therefore achieve a stronger cloud response compared to the global models. Recently many global models are including aerosol mixing, with enhanced absorption as BC mixes with other chemical species. This will enhance the AA effects on clouds. It should also be noted that BC is co-emitted with other mostly scattering aerosol species such as organic carbon and sulfur dioxide that oxidizes to sulfate, so the single scattering albedo of a given source will be larger than that of BC alone. From a mitigation perspective, the most relevant issue ultimately is the cloud response to particular sources containing BC.

Accurate simulation of absorbing aerosols and simulation of clouds are both challenging for global models. The models have coarse vertical resolution and must accurately capture both the vertical transport and removal of partially hydrophilic particles in convective plumes. Furthermore accurate simulation of biomass burning plumes is generally not parameterized in global models. The vertical distribution of AA's in global

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models has undergone little constraint due to limited observational information, however this distribution is especially diverse among models (e.g. Textor et al., 2006; Koch et al., 2009). In those regions where models have been tested, the performance is often not very good. Compared with a few aircraft BC profiles in the North American region, most models overestimated BC aloft (Koch et al., 2009). Recently available CALIPSO satellite retrievals are beginning to help models validate their vertical aerosol distributions. However this product must be combined with an indicator of aerosol absorption, such as the OMI aerosol index, to identify the presence of absorbing aerosols.

We may also question whether global model cloud schemes are able to reproduce the AA-cloud relations captured by cloud-scale models. For example, studies focused on the impacts of AA's on marine stratocumulus (e.g. Johnson et al., 2004; Hill and Dobbie, 2008) have noted that these clouds are not well simulated by global models. Small-scale cloud changes, such as cloud layer thickness, cloud-top entrainment, cloud fraction and the tendency to drizzle are affected by the scale of interactions among radiation, turbulence and moist physics on small horizontal and vertical scales. Johnson (2005) compared the marine stratocumulus cloud response from the NCAR single column model with simulations using an LES and found the latter had a response larger by a factor of five. On the other hand, cloud-scale models are limited in their ability to include climate feedbacks.

Indeed, some of the global model studies emphasized the importance of using a model with climate response to accurately include AA effects on clouds. For example, Wang (2004) found much larger cloud changes when using a Qflux model compared to a model with fixed SSTs. Several of the climate model studies (Wang, 2004; Penner et al., 2003; Roeckner et al., 2006; Koch et al., 2010) found that BC did not warm climate due to compensatory cloud cooling. It may also be the case that several model studies found BC efficacy less than 1 due to increased cloud cover for lofted BC (Hansen et al., 2005). Model studies that do not include climate response would also lack full cloud response.

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However, global climate models are already known to be diverse in the sign of their response to the climate warming from CO_2, with some models producing more cloud and some less cloud, even though nearly all models have positive cloud climate sensitivity due to other related feedbacks such as water vapor changes (e.g. Soden and Held, 2006). Therefore we may expect similar diversity in cloud cover response to AA's simply because of basic differences in the model cloud feedback behaviors. It is interesting to note that in the GISS model, low-level low-mid-latitude cloud cover generally decreases in response to increasing CO_2 (Del Genio et al., 2005a, b), however low-level cloud cover generally increases in these regions in response to absorbing aerosols (Perwitz and Miller, 2010; Koch et al., 2010). Therefore in this case the AA's would have to counter the tendency of the model to decrease low-level clouds under the warmer conditions associated with aerosol absorption. Modeling AA effects on clouds is probably more challenging compared to greenhouse gas effects, due to their short lifetime and heterogeneous distribution in the atmosphere.

Several global models studies (Perlwitz and Miller, 2010; Wang, 2004; Hansen et al., 2005; Penner et al. 2004; Roberts and Jones, 2004; Roeckner et al. 2006, Koch et al., 2010) using different global climate models (GISS, CSM-NCAR, Grantour, Hadley Center, ECHAM5) indicate a negative cloud-climate forcing response to absorbing aerosols. Four of these studies (Wang, 2004; Penner et al., 2003; Koch et al., 2010; Roeckner et al., 2006, for future climate conditions) indicate the effect may be strong enough to nearly eliminate BC warming effects on climate. Therefore as we consider BC as a global warming mitigation target, we need to improve our understanding and estimation of these cloud responses. Future research should include ongoing comparison of model BC vertical distribution using aircraft data and CALIPSO retrievals. Study of BC absorption and cloud parameterizations in global models would help test the models' abilities to accurately simulate cloud responses to AA. Performance of parallel AA-cloud relation experiments in global and cloud scale models may also help test the ability of global models to capture the robust features of the cloud-scale models. Finally, field and satellite based studies, such as Koren et al. (2004),

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Mahowald and Kiehl (2003), and Brioude et al. (2009), in a variety of conditions and regions are important for testing the models.

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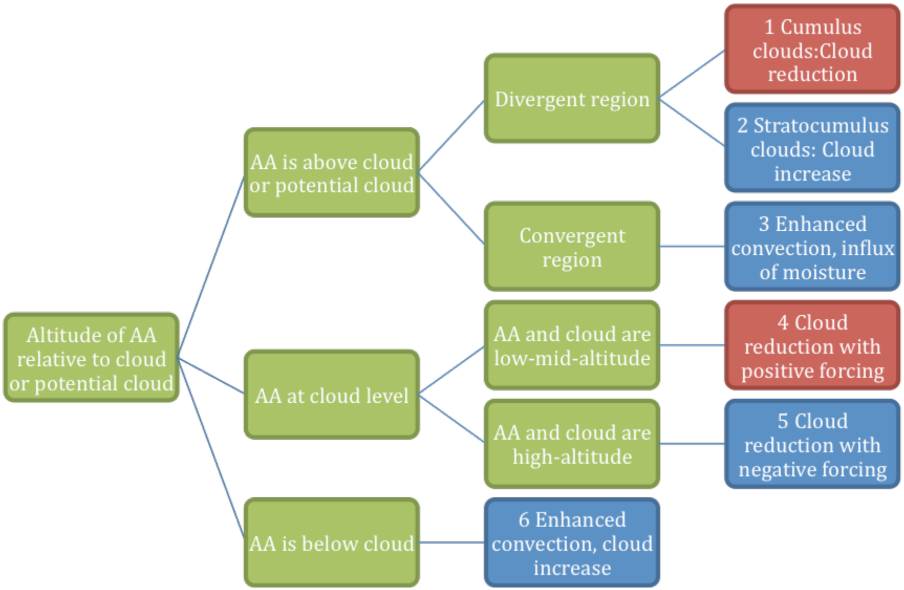

Fig. 1. Summary of aerosol absorption effects on cloud cover. Red and blue indicate positive and negative forcing effects due to cloud changes.

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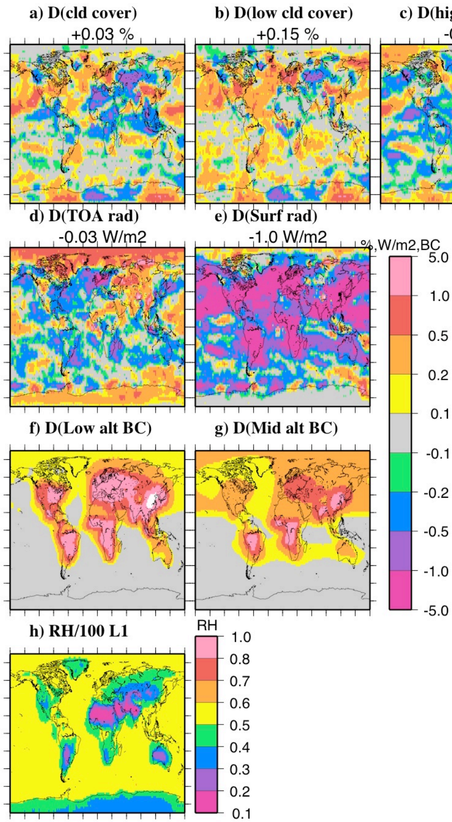

Fig. 2. Average changes due to BC during 1970–2000 in Koch et al. (2010) transient climate simulations for **(a)** cloud cover %, **(b)** low-level cloud cover %, **(c)** high-level cloud cover %, **(d)** top of atmosphere radiative forcing (Wm^{-2}), **(e)** surface forcing (Wm^{-2}), **(f)** average concentration for model levels 1 and 2 (ng/kg/100), **(g)** average concentration for model levels 3–5 (ng/kg/100); **(h)** Relative humidity (%/100) in lowest model level for run with all sources.

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