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Key Points:
- East Asian dust affects regional and global climate by typical transport paths
- East Asian dust aerosols are more absorptive than those from Saharan Desert
- Dust-cloud-precipitation interaction over arid regions may aggravate drought

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Climate effects of dust aerosols over East Asian arid and semiarid regions

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Abstract East Asia is a major dust source in the world. Mineral dusts in the atmosphere and their interactions with clouds and precipitation have great impacts on regional climate in Asia, where there are large arid and semiarid regions. In this review paper, we summarize the typical transport paths of East Asian dust, which affect regional and global climates, and discuss numerous effects of dust aerosols on clouds and precipitation primarily over East Asian arid and semiarid regions. We hope to provide a benchmark of our present understanding of these issues. Compared with the aerosols of Saharan dust, those of East Asian dust are more absorptive of solar radiation, and its direct radiative forcing at the top of atmosphere is nearly positive or nil. It means that aerosols of East Asian dust can influence the cloud properties not only by acting as cloud condensation nuclei and ice nuclei (via first indirect effect, second indirect effect, and invigoration effect) but also through changing the relative humidity and stability of the atmosphere (via semidirect effect). Converting visible light to thermal energy, dust aerosols can burn clouds to produce a warming effect on climate, which is opposite to the first and second indirect effects of aerosols. The net dust aerosol radiative effects are still highly unclear. In addition, dust can inhibit or enhance precipitation under certain conditions, thus impacting the hydrological cycle. Over Asian arid and semiarid regions, the positive feedback loop in the aerosol-cloud-precipitation interaction may aggravate drought in its inner land.

1. Introduction

As one of the four major terrestrial sources of atmospheric aerosols (desert dust, biomass burning, and biogenic and anthropogenic air pollutions), desert dust aerosols are responsible for significant climate forcing through their direct effects on solar and thermal radiation [e.g., *Sokolik and Toon*, 1996; *Li*, 2004; *Shi et al.*, 2005], and their indirect/semidirect effects on cloud and precipitation processes [e.g., *Sassen*, 2002; *K. Huang et al.*, 2010]. Furthermore, dust aerosols have a critical effect on regional meteorology and climate because of the large amount of mineral dust particles emitted from arid and semiarid regions in the world [*Twomey et al.*, 1984; *J.-P. Huang et al.*, 2006a, 2006b, 2010; *J.-F. Huang et al.*, 2009a, 2009b, 2009c; *Shao et al.*, 2011]. These regions account for one third of the Earth's continental area and are the primary sources of dust storms. Every year, about 1000 to 3000 trillion grams (Tg) of dust are entrained into the atmosphere from arid and semiarid regions [*Penner et al.*, 2001; *J.-F. Huang et al.*, 2010]. The largest contribution to the global dust load is emitted from the North African (50–70%) and Asian deserts (10–25%) [*Tegen and Schepanski*, 2009]. Over Asia, about 800 Tg yr^{-1} of dust is injected annually into the atmosphere, of which ~30% is redeposited onto deserts, 20% is transported over regional scales, and the remaining ~50% is transported to the Pacific Ocean and beyond [*Zhang et al.*, 1997; *Liu et al.*, 2008; *Uno et al.*, 2009; *C. Li et al.*, 2010a]. Therefore, Asian dust not only plays an important role in the energy budget and hydrological cycle but also is critical in global biogeochemical cycle by acting as a potential source of iron to marine ecosystems in the North Pacific [*Bishop et al.*, 2002; *Gao et al.*, 1997; *Hsu et al.*, 2009; *Zhang and Gao*, 2007; *Mahowald et al.*, 2010].

Dust particles are produced by disintegration of aggregates following creeping and saltation of larger soil particles over deserts and other arid/semiarid surfaces [*Kok*, 2011; *Zhao et al.*, 2006]. Mineral aerosol may be composed of iron oxides (e.g., hematite and goethite), carbonates (e.g., calcite and dolomite), quartz, and clays (e.g., kaolinite, illite, and montmorillonite) [*Chou et al.*, 2008; *Coz et al.*, 2009; *Lafon et al.*, 2006; *Twohy et al.*, 2009]. The rising number of dust storms is due to increasing desertification, which is fed, in turn, by dust events that exacerbate drought conditions over semiarid areas [*Han et al.*, 2008; *Wang et al.*, 2008; *J.-P. Huang et al.*, 2010].

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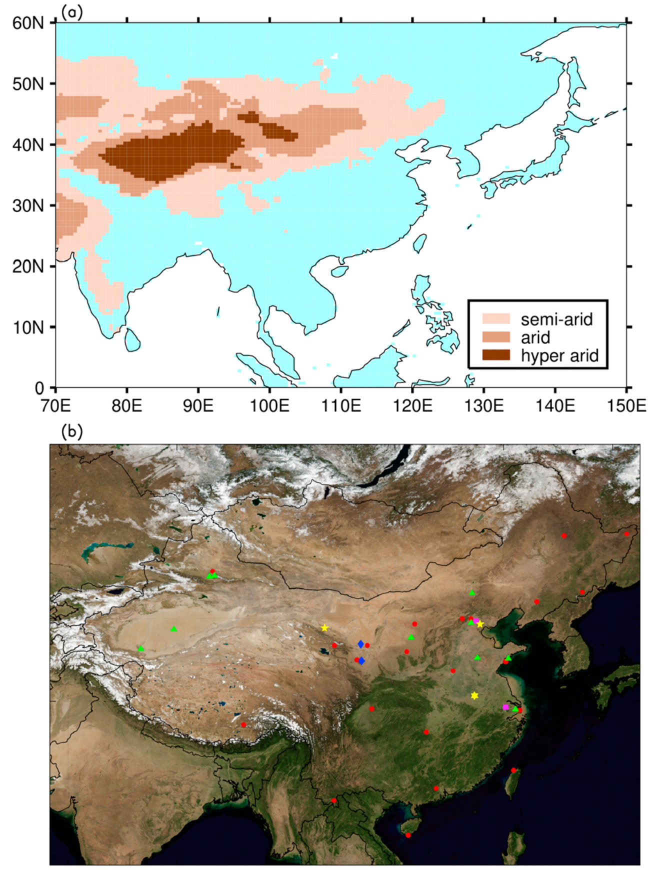

Figure 1. (a) Spatial distribution of arid and semiarid regions in Asia. The arid index is from *Feng and Fu* [2013]. (b) Aerosol observation networks established over China under EAST-AIRE, ARM Mobile Facility (AMF)-China, and by Lanzhou University, Fudan University, and the Nanjing University of Information Science and Technology (NUIST). Magenta squares and red solid circles stand for the baseline stations and haze-meter sties of the EAST-AIRE aerosol observation network operated by the Chinese Academy of Sciences and NUIST. Yellow five- and six-pointed stars represent the deployment locations of ARM Ancillary Facility (AAF) and AMF in 2008, respectively. Blue diamonds are the SACOL permanent and mobile facilities established by Lanzhou University. Green triangles are the aerosol sampling stations operated by the Fudan University under the Atmospheric Aerosols of China and their Climate Effects project. After *Li et al.* [2011a, Figure 1a].

In northwestern China, the Gobi Desert and Loess, the major source regions of Asian dust, are expanding [*Wang et al.,* 2008; *Fu et al.,* 2008; *J.-P. Huang et al.,* 2010].

Precipitation is one of the most important water resources in arid and semiarid regions. Small variations or changes in the amount, altitude, physical thickness, and/or microphysical properties of clouds due to natural and human influences can alter the surface radiation budget and the hydrological cycle [*J.-P. Huang et al.,* 2006a, 2006b, 2006c, 2010; *Chen et al.,* 2008, 2010]. As aerosol particles serve as a necessary ingredient for cloud droplet formation, any change in the amount and/or composition of aerosols can affect clouds in a variety of ways. Studying aerosol-cloud-radiation-precipitation interactions over these arid and semiarid regions is of great importance and possibly more urgently needed than for any other regions on Earth.

Many observational and modeling studies have focused on aerosol direct, indirect, and semidirect effects and found that small changes in cloud microphysical properties (e.g., droplet size, phase, and water path) could induce large changes in cloud radiative forcing (RF) and precipitation [*Intergovernmental Panel on Climate Change (IPCC),* 2007]. It is important to understand and quantify these aerosol effects; particularly how these forcing fields act together to alter the surface budget (heat flux), cloud properties, atmospheric dynamics, as well as the hydrological cycle (modification of precipitation regimes).

In this review paper, we summarize recently acquired knowledge about the climate effects of dust aerosols over East Asian arid and semiarid regions. This paper contains six sections. It begins in section 2 with a brief introduction about the dust source regions and its transports in East Asia. The direct effect of dust aerosols in these interested areas is reviewed in section 3. Sections 4 and 5 summarize recent studies on the indirect effect and semidirect effect of dust aerosols, respectively. Other studies concerning theories, mechanisms, and model simulations are also discussed in these sections, together with some comparative investigations. Section 6 presents the interaction among aerosol, cloud, and precipitation and then deduces a possible feedback loop in the arid and semiarid regions. In section 7, we summarize the current understanding about dust aerosol climate effects and discuss some obstacles to assess these effects.

2. Sources and Transports of East Asian Dust

Figure 1a shows the distribution of dryland in East Asia. The dryland is defined by the arid index, ratio of precipitation to potential evapotranspiration, and the threshold values to classify the hyperarid, arid, and

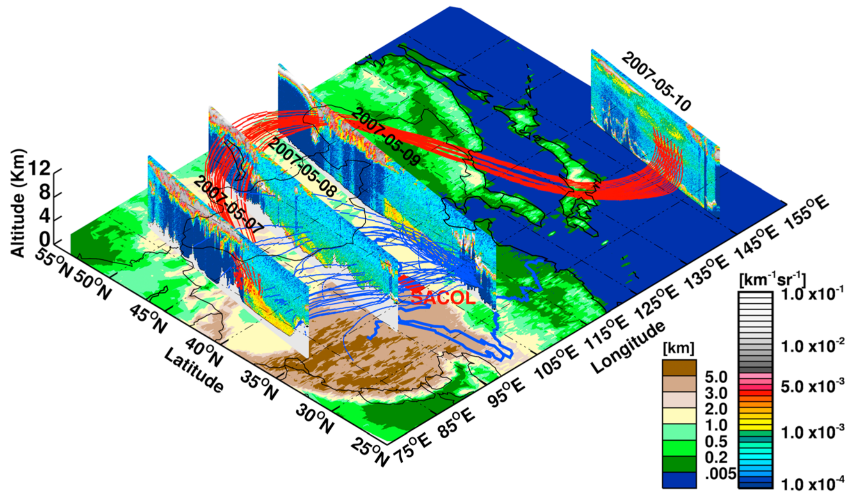

Figure 2. Illustration of a dust event, which originated in the Taklamakan Desert on 7 May 2007 and transported the dust to the Pacific Ocean. Red lines represent back trajectories initialized over the Eastern Pacific. Blue lines represent back trajectories initialized over central China. Total attenuated backscatter profiles at 532 nm from CALIPSO are shown in the vertical images. Color scales on the left represent the topographical elevation in km, while color scales on the right represent the 532 nm total attenuated backscatter in km^{-1} sr^{-1}. After *Huang et al.* [2008b, Figure 2].

semiarid is according to *Feng and Fu*'s [2013] method. The Taklimakan Desert and Inner Mongolian Gobi Desert are the two primary dust sources in East Asia. Dust storm events occur more frequently in the Taklimakan Desert than in the Gobi Desert [*Kai and Tokuno*, 1997]. The floating dust aerosols mostly occur in the Taklimakan Desert, while the blowing dust aerosols mostly occur in the Gobi region.

For globally, *Tegen and Lacis* [1996] reported that the dust mass loading is 36.6 mg m^{-2} for particles in the range of 0.1 to 10 mm, which corresponds to a burden of 18.7 Tg. The dust burden was estimated to be 13.8 Tg by *Takemura et al.* [2000], 31–40 Tg by *Ginoux et al.* [2001], and 18.1 Tg by *Liao et al.* [2004] for the 0.1–10 mm size range. *Huneeus et al.* [2011] compared 15 global aerosol models within the AeroCom project and found that there are large differences among those global models. And for Asia, emission ranges from 27 Tg to 873 Tg yr^{-1}. The large differences between these studies are a result of large uncertainties in emissions and of the different meteorological fields and schemes used for deposition calculations.

Through a few ground-based measurements, satellite retrievals and several field campaigns, the vertical distribution and long-range transport pattern of typical dust aerosols in East Asia are well illustrated by *J.-P. Huang et al.* [2007, 2008b, 2010]. Generally, dust aerosols are present in the boundary layer in and near the dust source regions [*Kai et al.,* 2004], while they can also reach the upper troposphere by strong convective updraft and then forming dust plumes, which are often entrained into the westerlies and transported from the continent to the open seas near Korea and Japan, and even reaching eastern North America [*Biscaye et al.,* 2000; *Liu et al.,* 2013], impacting atmospheric hydrological and radiative budgets along the way [*Husar et al.,* 2001; *Zhang et al.,* 2003; *Huang et al.,* 2007, 2008b]. The jet stream controls the path of dust aerosols aloft at high altitude, but trajectories in the lower atmosphere are determined by regional weather systems and topography. *Huang et al.* [2008b] gave an example of a dust event that took place on 7 May 2007 (Figure 2). Images from the CALIPSO and back trajectory analysis indicated that the dust event originated over the Taklimakan Desert on that day (or perhaps somewhat earlier) and merged with dust over the Gobi Desert; the plume was then advected over the Pacific. There are two major paths for dust aerosols to be transported to downwind regions: (1) the transport from the Qilian Mountain, Mongolia to the Pacific Ocean can take 3–4 days and (2) the path from the Qaidam Basin through Qinghai and Gansu provinces to reach the Pacific Ocean takes longer time and impacts more regions than the first path.

In addition to the two paths, dust originated in the Taklimakan Desert in northwestern China can travel across the Hexi Corridor and Loess Plateau to reach southeastern China, such as the Yangtze River Delta [*J. Liu et al.,* 2011a] and even Hong Kong and Taiwan [*Wang et al.,* 2011]. According to the altitude of transportation, three main types are identified [*Tsai et al.,* 2008]: upper level type (U type), lower level type (L type), and descending type (D type). For the U- and D-type dust trajectories, dust particles are first lifted to the middle

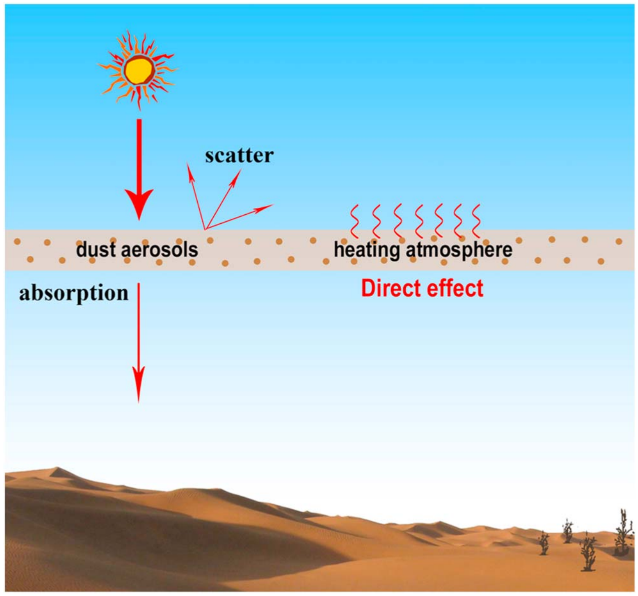

Figure 3. A schematic depiction of the direct radiative effect of dust aerosols.

troposphere through ascending motion, accompanied by a quick descending. For example, a dense dust layer was observed above 5 km during a dust event from 14 to 17 March in the Yangtze River Delta [*J. Liu et al.,* 2011a]. Dust aerosols originated in northwestern China can significantly enhance seasonal mean aerosol extinction near the surface and at higher altitude in spring in the Yangtze River Delta [*Liu et al.,* 2012].

To capture direct evidence of the impacts of dust aerosols over Asian arid and semi-arid regions, dust samples collection along dust pathways and observations have been made at a few highly instrumented supersites and mobile facilities (see Figure 1b). The Semi-Arid Climate and Environment Observatory of Lanzhou University (SACOL) established in 2005 [*J.-P. Huang et al.,* 2008a; *Guan et al.,* 2009; *Z. Huang et al.*, 2010; *Bi et al.,* 2010; *G. Wang et al.,* 2010] is an international long-term climate observatory located in the inland of loess and close to the major source regions of Asian dust—the Gobi Desert. In addition, extensive studies concerning Asian dust and anthropogenic aerosols were conducted during two field experiments: the East Asian Studies of Tropospheric Aerosols: An International Regional Experiment (EAST-AIRE) [*Li et al.,* 2007a] and the East Asian Studies of Tropospheric Aerosols and its Impact on Regional Climate (EAST-AIRC) [*Li et al.,* 2011a]. In the EAST-AIRE, the properties and climate effects of dust aerosols over their source regions in northern and western China [*J.-P. Huang et al.,* 2010; *W. Wang et al.,* 2010] and over downstream regions as far as across the Pacific [*Guo et al.,* 2010; *C. Li et al.,* 2010a, 2010b; *Liu et al.,* 2010; *Logan et al.,* 2010] were investigated. Based on these measurements, several studies were carried on concerning dust properties [*Wang and Huang*, 2009; *Sun et al.,* 2010; *K. Huang et al.,* 2010; *Fu et al.,* 2010; *Hansell et al.,* 2012], radiative forcing of dust aerosols [*Ge et al.,* 2010], and direct and indirect effects of mixtures of dust and anthropogenic aerosols on regional climate [*Qian et al.,* 2009; *Gu et al.,* 2010; *S.-H. Wang et al.,* 2010; *L. Zhang et al.,* 2010; *J. Liu et al.,* 2011a; *Fan et al.,* 2012].

As will be discussed below, the vertical distribution of dust aerosols can be an important factor determining their radiative effects. Besides observations using lidar from ground and space, however, there are very few aircraft measurements of aerosols over the deserts in East Asia. The data from the EAST-AIRE aircraft campaign over northern China during the dust season in spring 2005 helped to reveal elevated dust layers above the boundary layer downwind of the major dust source regions [*Dickerson et al.,* 2007; *Li et al.,* 2012].

3. Direct Radiative Effect

Direct radiative effect (DRE) of dust is defined as the effect of all anthropogenic and natural dust aerosols on the radiative flux at the top of the atmosphere and at the surface, and on the absorption of radiation within the atmosphere (Figure 3), which are determined by the aerosol optical depth (AOD) or extinction coefficient, the single-scattering albedo (SSA), and the asymmetry parameter or phase function. In addition, heating rate of dust aerosols is also used as an index to estimate dust DRE.

3.1. Radiative Forcing

Figure 4 summarizes dust RFs from several studies in different climate regions. On the global scale, estimate of the annual mean DRE of mineral dust is $-0.1\ \mathrm{W\,m^{-2}}$ with the uncertainty range from $-0.3\ \mathrm{W\,m^{-2}}$ to $0.1\ \mathrm{W\,m^{-2}}$ [*IPCC,* 2013]. A study by *H. Zhang et al.* [2010] and *L. Zhang et al.* [2010] reported that the global annual mean direct forcing is -0.4, -1.46, and $+1.86\ \mathrm{W\,m^{-2}}$ at the top of atmosphere (TOA), the surface and within the atmosphere, respectively. *Reddy et al.* [2005] estimated that the global annual mean all-sky

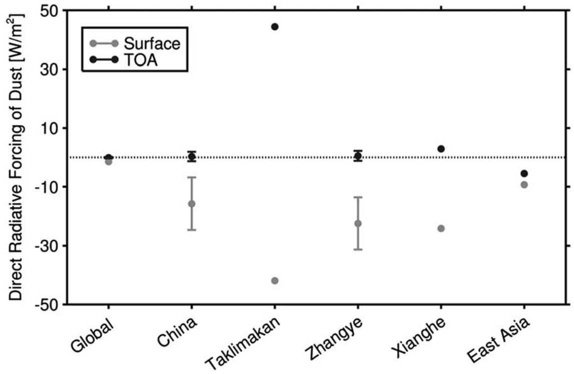

Figure 4. The daily mean direct radiative forcing of dust aerosols at the TOA and surface in different regions. The TOA value for the globe is from the *IPCC* [2013], and the corresponding surface value is from *H. Zhang et al.* [2010] and *L. Zhang et al.* [2010]; the values over China region, the Taklimakan Desert, Zhangye, Xianghe, and East Asia (20–50°N, 100–150°E) are from *C. Li et al.* [2010b], *J.-P. Huang et al.* [2009], *Ge et al.* [2010], *Li et al.* [2007b], and *Seinfeld et al.* [2004], respectively.

shortwave RF by dust at the TOA is $-0.28\ \mathrm{W\,m^{-2}}$. Simulated globally averaged RF values of dust at the TOA for the last glacial maximum, preindustrial, and doubled-carbon dioxide climate are -1.0, -0.4, and $+0.14\ \mathrm{W\,m^{-2}}$, respectively, relative to the current climate [*Mahowald et al.,* 2006]. Generally speaking, the shortwave RF values at both the TOA and the surface are negative. The longwave RF is positive, which results from the absorption and emission of longwave radiation by dust aerosols [*Markowicz et al.,* 2003]. However, the magnitude of cooling at the TOA is smaller than that at the surface, because cooling due to the scattering of solar flux is greatly offset by warming due to the absorption of solar and thermal radiation [*Miller and Tegen*, 1998; *Shi et al.,* 2008; *Bi et al.,* 2013]. For the diurnal cycle of dust DRE, the surface RF is negative during the daytime because of the loss of sunlight by absorption and backscattering of dust aerosols but is positive during the nighttime as a result of the continuous thermal emission from dust aerosols [*Cautenet et al.,* 1991; *Claquin et al.,* 1998].

Various human activities (such as land use practices and construction) can result in significant dust loading, with the anthropogenic fraction of dust estimated to be as much as 30% to 50% of total dust production, and its direct RF on both global and regional scales may be comparable to or exceed the forcing by other anthropogenic aerosols [*Tegen et al.,* 1996; *Sokolik and Toon*, 1996].

Claquin et al. [1998] studied the sensitivity of these mineral forcings to different treatments of the aerosol complex refractive index and size distribution. Aerosol complex refractive index has been identified as a critical parameter given the paucity and the incertitude associated with it. Furthermore, the imaginary part of the refractive index is inadequate if spectrally averaged. Its natural variability (linked to mineralogical characteristics) leads to variations of up to ±40% in aerosol forcing calculations. A proper representation of the size distribution when modeling mineral aerosols is required since dust optical properties are very sensitive to the presence of small particles. *Wang et al.* [2013] estimated the radiative effect of a heavy dust storm over northwestern China and found that the most important factors for longwave RF in the atmosphere are the AOD and its vertical distribution, whereas SW is dependent on the surface albedo, AOD, and SSA.

Dust DRE shows great regional characteristics as it largely depends on dust source and surface condition. Over bright surfaces like deserts, the shortwave RF at the TOA can be positive because more radiation, which should be reflected into the atmosphere, could be absorbed by dust, especially at shorter wavelengths [*Fu et al.,* 2009]. For Saharan dust outbreaks over the Mediterranean, *Santese et al.* [2010] found that the daily mean shortwave forcing is offset by a longwave forcing of ~30% at the surface and of ~50% at the TOA. *Mallet et al.* [2009] calculated the RF of dust aerosols for a major dust event over West Africa and found that the presence of dust particles induced a large instantaneous reduction of surface incoming shortwave ($-127\ \mathrm{W\,m^{-2}}$), while the net effect at the TOA was just a small cooling ($-12\ \mathrm{W\,m^{-2}}$). Similar positive shortwave RF at the TOA has been found over the Taklimakan Desert. *J.-P. Huang et al.* [2009] calculated that the daily mean net RF for a 5 day dust event in July 2006 over this region was 44.4, -41.9, and $86.3\ \mathrm{W\,m^{-2}}$ at the TOA, the surface, and within the atmosphere, respectively. Among these forcing estimates, about two thirds of the warming effect at the TOA is related to the longwave radiation, while about 90% of the atmospheric warming arises from solar radiation. At the surface, about one third of the dust solar radiative cooling effect is compensated by its longwave warming effect. Over Badain Jaran Desert, *Bi et al.* [2013] examined cloudless days with background aerosol loadings and found that the dust shortwave direct RF ranges from -4.8 to $0.4\ \mathrm{W\,m^{-2}}$ at the TOA, -5.2 to $-15.6\ \mathrm{W\,m^{-2}}$ at the surface, and 5.2 to $10.8\ \mathrm{W\,m^{-2}}$ in the atmosphere. Along the dust pathways, the dust aerosol RF at the TOA and the surface in Zhangye are

0.52 ± 1.69 and -22.4 ± 8.9 W m^{-2}, respectively [*Ge et al.,* 2010]. Over SACOL, *Y. Liu et al.* [2011] calculated the aerosol RF during the dust season (March, April, and May) and found the shortwave RFs are 0.68, −70.02, and 70.70 W m^{-2}, respectively, at the TOA, surface, and in the atmosphere. Using the CloudSat and Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) satellite data, *Kuhlmann and Quaas* [2010] found that aerosol plumes reduced premonsoon seasonally averaged (March–May) surface shortwave radiation throughout the Asian monsoon region by 20 to 30 W m^{-2}. In Xianghe, the downward region of the dust, *Li et al.* [2007b] showed that the heavy dust loading compounded with pollutants exerted an exceptionally strong direct forcing of about −15 to −25 W m^{-2} (daily mean) at the surface, but the forcing at the TOA was virtually nil. At Taihu in southeastern China with very high anthropogenic pollution, the mixed dust can reduce shortwave radiation by 48 W m^{-2} at surface and increase atmospheric absorption of shortwave radiation by 27 W m^{-2} [*J. Liu et al.,* 2011a]. Across China, the daily mean net RF of dust is 0.3 ± 1.6, -15.7 ± 8.9, and 16.0 ± 9.2 W m^{-2} at the TOA, the surface, and within the atmosphere, respectively [*Z. Li et al.,* 2010]. *Qian et al.* [2003] used a regional climate model to present direct aerosol RF effects, which were almost negative over China. Based on the Asian Pacific Regional Aerosol Characterization Experiment data, *Seinfeld et al.* [2004] estimated that the dust RF at the TOA and surface was −5.5 and −9.3 W m^{-2}, respectively, for the area of 20–50°N, 100–150°E during 5–15 April 2001.

3.2. Single-Scattering Albedo

Single-scattering albedo (SSA), the ratio of scattering efficiency to total extinction efficiency, ranges from 0 to 1. A smaller value indicates the particle with higher absorption of solar radiation. When dust particles travel from the source regions to remote areas, they are often coated with pollutants from both Asia and Europe along their pathways, resulting in much stronger absorption by these particles than those from other major deserts such as the Sahara [*Dubovik et al.,* 2002; *Eck et al.,* 2010]. Dust aerosols have a negative TOA forcing for large SSA and a positive TOA forcing for small SSA.

Loeb and Su [2010] did a sensitivity study about the uncertainty of direct aerosol RF and found that perturbing individual aerosol optical properties (AOD, SSA, asymmetry parameter, scale height, and anthropogenic fraction) from their base value, the perturbation of direct aerosol RF shows great sensitivity to SSA in cloudy columns. And the direct aerosol RF uncertainty in all-sky conditions is greater than in clear-sky conditions, even though the global mean clear-sky direct aerosol RF is more than twice as large as the all-sky direct aerosol RF.

By combining satellite-measured reflectance and surface measurements of transmittance, *Lee et al.* [2007] derived the SSA at 25 stations across China, including some dust-laden regions, and noted that the nationwide mean of SSA at 0.5 μm is 0.89 ± 0.04. In northwestern China [*Ge et al.,* 2010] and in India [*Pandithurai et al.,* 2008], the SSA of Asian dust ranges from 0.74 to 0.84 at 0.5 μm. *Costa et al.* [2006] found that the SSA of Asian dust can be as low as 0.76, which is much smaller than the value found for African dust. For example, *Fouquart et al.* [1987] reported a mean value of 0.95 for the SSA at 0.55 μm of African dust. *Haywood et al.* [2003] found that the SSA at 0.55 μm ranged from 0.95 to 0.99 during the Saharan Dust Experiment. *McFarlane et al.* [2009] derived a mean value of 0.94 at 0.5 μm during January–April 2006 in Niamey, Niger. The smaller values of SSA for Asian dust suggest that these particles are strongly absorbing aerosols. During a recent experiment near a major desert in rural western China, high concentrations of anthropogenic species were found [*C. Li et al.,* 2010c].

3.3. Heating Rate

Vertical distribution of aerosols has a great impact on heating rate profile, which provides more detail about dust aerosol DRE on different atmospheric levels than RF. Over the Taklimakan Desert, dust aerosols can heat the atmosphere by up to 1, 2, and 3 K d^{-1} (daily mean values) under light, moderate, and heavy dust conditions, respectively; the maximum daily mean of radiative heating rate can reach 5.5 K d^{-1} at 5 km, at the location of the dust layer [*J.-P. Huang et al.,* 2009]. Other study [*Lemaître et al.,* 2010] indicated that during the daytime, dust aerosols can warm the atmosphere between 0.3 and 4.0 K d^{-1} on average, depending on altitude and latitude. Strong warming (i.e., heating rate as high as 8 K d^{-1}) was observed locally within a limited portion of the dust plume. The uncertainty in heating rate retrieval in the optically thickest part of the dust plume was estimated to be between 0.5 and 1.4 K d^{-1}. During nighttime, much smaller values of heating/cooling rates were retrieved (less than ±1 K d^{-1}). The vertical range and the duration of the dust layer

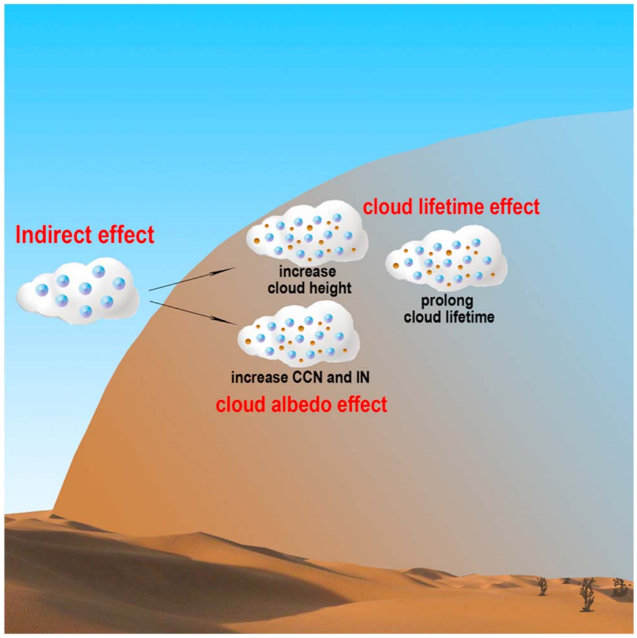

Figure 5. A schematic depiction of the indirect radiative effect of dust aerosols.

determine the effect of dust particles on the surface temperature [*Helmert et al.,* 2007; *Stanelle et al.,* 2010]. When a dust layer touches the ground and lasts for several days, an increase in surface temperature can occur, even during daytime. In the event of an elevated dust layer, there is a decrease in surface temperature. These temperature changes caused by dust DRE may result in horizontal temperature gradient, which can modify near-surface winds. Since surface wind threshold determines the uptake of dust from the surface, a feedback on total emission flux could be established.

Ramanathan et al. [2007] indicated that local pollution alone can enhance the solar heating of the lower atmosphere by ~50%. Other studies [*Tegen et al.,* 2006; *Milton et al.,* 2008; *Tulet et al.,* 2008; *Mallet et al.,* 2009] pointed out that dust aerosols can cool the surface by 0.5 to 10.0 K d^{-1}. Model studies by *Lau et al.* [2006, 2008] showed that the highly elevated surface air over the Tibetan Plateau may act as an "elevated heat pump" and alter the regional climate significantly through the absorption of solar radiation by dust coupled with black carbon emissions from industrial areas in northern India. As a result, a tropospheric temperature anomaly may be induced in late spring and early summer over parts of northern India and Tibet, leading to an earlier onset and intensification of the Indian summer monsoon.

In summary, dust aerosols have a warming effect within the atmosphere but a cooling effect at the surface. Unlike dark surface, the total effect of dust aerosols is usually warming or nil at the TOA in Asian arid and semiarid regions as its higher surface albedo and dust aerosols are more absorptive than those from any other dust source regions, such as the Sahara Desert. The largest forcing values are found over or near dust source areas, e.g., the Taklimakan Desert.

4. Indirect Radiative Effect

Dust aerosols can also change the energy fluxes in the Earth-atmosphere system by modifying cloud macrophysical and microphysical properties, such as the cloud liquid water content, cloud fraction, cloud top temperature, droplet number concentration, cloud particle size, and so on [*Twomey*, 1977; *Albrecht*, 1989; *Ackerman et al.,* 2000; *Liu et al.,* 2003; *Andreae et al.,* 2004; *Huang et al.,* 2006a], as shown in Figure 5.

4.1. Fundamental Physical Processes

As we all know that aerosols are necessary ingredients for cloud formation for their roles as condensation nuclei (CN) in forming cloud droplets and ice crystals. Dust aerosols are one of the most common types of ice nuclei (IN), and emerging evidence suggests that mineral particles can reach high into the upper troposphere to serve as IN for cirrus and mixed-phase cloud formation [*Heintzenberg et al.,* 1996; *Sassen*, 2002; *Sassen et al.,* 2003; *DeMott et al.,* 2003; *Field et al.,* 2006; *Teller and Levin*, 2006; *Barahona et al.,* 2010]. Dust may also interact with sea salt, anthropogenic pollutants, and secondary organic aerosol, forming particles that consist of a "core" of insoluble mineral dust with coatings of soluble material [*Gibson et al.,* 2007; *Levin et al.,* 2005; *Seisel et al.,* 2005]. Dust particles with a soluble coating are typically very efficient cloud condensation nuclei (CCN), often maintaining their activity as IN [*Levin et al.,* 2005]. In microphysical process, the ability of an aerosol particle to take up water and subsequently activate cloud condensation is determined by its size and composition. It has been suggested that the size of CCN is more important than its chemical composition [*Dusek et al.,* 2006]. Larger particles are more readily activated than smaller particles because they require a lower critical supersaturation. While CCN generally increases with CN for dusty events, the activation ratio tends to decrease sharply with increasing CN, implying that dust particles do not increase CCN concentration freely, despite mixing with other anthropogenic aerosols [*J. Liu et al.,* 2011b]. Moreover, due to their large size, dust particles can act as giant CCN that can form efficient collector drops and initiate the onset of drizzle and precipitation [*Feingold et al.,* 1999; *Levin and Cotton*, 2009]. When more aerosol particles are competing for the uptake of a fixed amount of liquid water content, the resulting cloud droplets should be more but smaller, which have a larger total surface area and higher cloud albedo (cloud albedo effect) [*Twomey*, 1977]. Therefore, a dust-polluted cloud reflects more solar radiation back to space, resulting in a negative RF at the TOA [*Lohmann and Feichter*, 2005; *Huang et al.,* 2006c; *G. Wang et al.,* 2010; *W. Wang et al.,* 2010, 2013].

In addition, these more numerous but smaller cloud droplets collide less efficiently with each other, which reduces the precipitation efficiency of polluted clouds, presumably prolonging their lifetime and ability in perturbing atmospheric circulation, thus leading to redistribution of clouds and precipitation (cloud lifetime effect) [*Albrecht*, 1989]. It also implies more solar radiation is scattered back to space, thus reinforcing the cloud albedo effect [*Lohmann*, 2006; *Takemura et al.,* 2007].

When it comes to warm-based mixed phase, delaying precipitation would allow more small cloud particles to ascend above the freezing level and initial ice process, which will release more latent heat and hence invigorate the vertical development of clouds. This is the so-called aerosol invigoration effect [*Andreae et al.,* 2004; *Koren et al.,* 2005]. The climate impact of this effect is likely a positive forcing [*Koren et al.,* 2010; *Rosenfeld et al.,* 2013].

4.2. Aerosol Effect on Cloud Properties

Due to scarce ground-based observations in Asia arid and semiarid regions, satellite products, especially data collected by the A-Train constellation, are wildly applied to study the effects of dust on clouds [*J.-P. Huang et al.,* 2006a, 2006b, 2006c, 2010; *Su et al.,* 2008; *W. Wang et al.,* 2010]. *Huang et al.* [2006a, 2006b] compared the cloud properties under dusty and dust-free conditions in the same meteorological environment over northwestern China and found that dust aerosols injected into clouds would decrease ice cloud effective particle size and optical depth of cirrus clouds by 11% and 32.8%, respectively [*Huang et al.,* 2006a]. *W. Wang et al.* [2010] also studied the dust aerosol effect on cloud properties over the dust source region (the Taklimakan) and dust transported region (the West Pacific) and found similar results. Other results [*Kawamoto et al.,* 2004; *Su et al.,* 2008; *J-P. Huang et al.,* 2010] in Asia arid and semiarid regions also showed similar trends and further confirmed the conclusion that dust aerosols can reduce cloud effective particle size. A study based on general circulation model simulation indicated that globally the ice crystal number and mass in cirrus are reduced by 10% and 5%, respectively, whereas the ice crystal size is increased by 3% [*Kuebbeler et al.,* 2013]. This difference between observation and modeling implies a large uncertainty in our understanding of the influence of dust on cloud properties.

As theoretical hypothesis and model simulations showed that aerosols have completely different influences on convective clouds and stratiform clouds [see *Tao et al.,* 2012 for a comprehensive review]. *Rosenfeld* [1999] and *Rosenfeld and Woodley* [2000] analyzed aircraft and satellite data and suggested that pollution aerosols suppress warm shallow clouds but invigorate deep convective clouds. *Koren et al.* [2005, 2010] obtained the invigoration effect of Saharan dust over the eastern Atlantic Ocean. *Fan et al.* [2012] used a cloud-resolving model fitted with extensive observations data collected during the EAST-AIRC campaign [*Li et al.,* 2011a] to simulate convective and frontal cloud systems over China and then found significant aerosol invigoration effect as expected.

4.3. Indirect Effect of Dust

Dust aerosols influence the radiative energy budget through both direct and indirect effects on cloud microphysical properties. The *IPCC* [2007] considered the aerosol indirect effect as the largest uncertainty in climate simulation. It is thus important to quantify the aerosol indirect forcing over Asian arid and semiarid regions. *Lohmann and Feichter* [2005] summarized various aerosol effects on cloud and climate. They found that the range of model estimates for cloud albedo indirect RF of all kinds of clouds varies from -0.5 to -1.9 W m^{-2}.

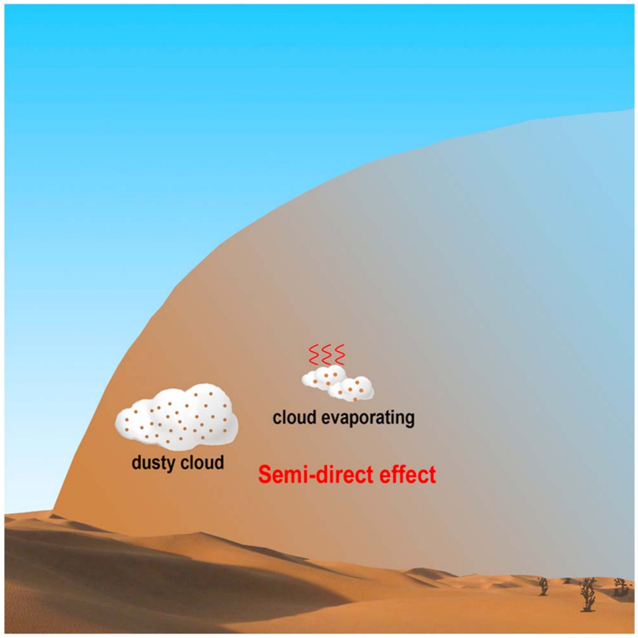

Figure 6. A schematic depiction of the semidirect radiative effect of dust aerosols.

All models indicated a negative forcing, which is -1 W m^{-2} on average, with a standard deviation of 0.4 W m^{-2}. The cloud lifetime effect varies considerably between different models (ranging from -0.3 to -1.4 W m^{-2}), resulting in an average forcing of -0.7 W m^{-2} with a standard deviation of 0.5 W m^{-2}. Semidirect effect estimates range from +0.1 to -0.5 W m^{-2}. The location of black carbon with respect to clouds gives rise to difference in the estimates. Generally, the aerosol indirect forcing is negative in most regions. As dust aerosols usually act as IN, some studies showed that ice clouds in dust-laden air were found at significantly warmer temperatures than in dust-free conditions [*Choi et al.,* 2010; *Sassen et al.,* 2003; *Seifert and Stevens*, 2010]. Increasing outgoing thermal radiation would amplify the cooling effect from aerosol indirect effect. *Su et al.* [2008] found that absolute values of instantaneous TOA RF of polluted clouds were significantly smaller than those of unpolluted clouds. *W. Wang et al.* [2010] found that dust aerosols caused an instantaneous net cloud cooling effect of 43.4% in northwestern China. Other studies also indicated that dust aerosols caused a cloud cooling effect at the TOA [*Huang et al.,* 2006a, 2006c].

Usually, the aerosol cloud interaction is quantified as ACI = $-\ln r_e/\ln\alpha$, where r_e is the drop size and α is aerosol burden. Based on *Twomey* [1977], it can be shown that ACI is ranging from 0 to 0.33. *McComiskey and Feingold* [2008] found that depending on anthropogenic aerosol perturbation, RF ranges from -3 to -10 W m^{-2} for each 0.05 increment in ACI. *Su et al.* [2008] estimated the contribution of dust direct, indirect, and semidirect effects to cloud RF using a combination of satellite observations and Fu-Liou model simulations for 16 instantaneous cases during 4 years over China and Mongolia in cloudy sky. The mean value of the combined indirect and semi-direct shortwave RF is 82.2 W m^{-2}, or 78.4% of the total dust effect. The direct effect amounts to only 22.7 W m^{-2} or 21.6% of the total effect. Because both first and second indirect effects enhance cloud cooling effect, the dusty cloud warming effect is mainly due to the semidirect effect of dust aerosols.

So far, the RF of aerosol invigoration effect is still unclear, as the thickening of the convective core may lead to cooling, but the expansion of anvil clouds associated with deep convective clouds can induce positive RF during the invigoration process [*Koren et al.,* 2010]. Additionally, the invigoration effect would enrich the upper troposphere and lower stratosphere with water vapor and bring in positive forcing. During the EAST-AIRC campaign [*Li et al.,* 2011a], the RF of aerosol invigoration effect was slightly positive at the TOA but negative at the surface [*Fan et al.,* 2012].

The cloud albedo effect cannot be easily separated from other effects. The processes that reduce the cloud droplet size per given liquid water content also decrease the chance of precipitation formation, presumably prolonging cloud lifetime (the cloud lifetime effect). In turn, an increase in cloud lifetime also contributes to a change in the time-averaged cloud albedo. Therefore, whether the cloud lifetime or the cloud albedo effect is more important is still an open question.

5. Semidirect Effect

Dust aerosols, due to their absorption of solar radiation, can influence cloud formation and development not only by acting as CCN or IN but also through changing environmental conditions. The schematic depiction of the semidirect effect is shown in Figure 6. Dust can absorb solar radiation and generate local heating, which in turn changes the relative humidity and the stability of the atmosphere, and thereby influences the cloud lifetime and cloud liquid water content. Dust aerosol layer may either stabilize or destabilize the boundary layer depending on where the aerosol layer is located [*Hansen et al.*, 1997; *Ackerman et al.,* 2000; *Koren et al.,* 2004]. Over the arid and semiarid regions, dust aerosols are likely to evaporate the clouds when aerosols appear within or above the clouds [*J.-P. Huang et al.,* 2006b, 2010; *W. Wang et al.,* 2010; *Sakaeda et al.,* 2011]. This effect is a rapid response associated with the direct effect.

The causal effect between aerosol atmospheric heating (and surface cooling) and cloudiness has been examined by modeling studies on eddy resolving, regional, and global scales, revealing a more complicated picture than initially conjectured [e.g., *Hill and Dobbie*, 2008; *Sakaeda et al.,* 2011; *Zhuang et al.,* 2010].

5.1. Modeling of Semidirect Effect

The semidirect effect has been simulated using the general circulation models (GCMs) and high-resolution cloud-resolving models because it is implicitly taken into account whenever absorbing aerosols are included in the radiation scheme [*Hansen et al.,* 1997, 2005; *Lohmann and Feichter*, 2001; *Jacobson*, 2002; *Menon et al.,* 2002; *Penner et al.,* 2003; *Cook and Highwood*, 2004]. When diagnosed within a GCM framework, the semidirect effect can also include cloud changes due to circulation effects and/or surface albedo effects. This is because the GCM response allows for large-scale circulation changes that can have significant additional effects that are not captured in the large eddy simulation models [*Johnson*, 2005]. Moreover, the semidirect effect is not exclusive to absorbing aerosols. Cumulus and stratocumulus case studies also showed semidirect effects, indicating a similar relationship between the height of the aerosol layer relative to the cloud and the sign of the semidirect effect [*Ackerman et al.,* 2000; *Ramanathan et al.,* 2001; *Johnson et al.,* 2004; *Johnson*, 2005].

Using a global circulation model, *Lohmann and Feichter* [2001] compared the magnitudes of all these competing effects and found that the semidirect effect can be important locally, despite the fact that indirect effects dominate globally. *Perlwitz and Miller* [2010] used the Goddard Institute for Space Studies (GISS) climate model to investigate how dust affects cloud coverage and found a contrary to expected decrease in low cloud cover due to heating by absorbing aerosols. Their experiments used a slab ocean and dust aerosols with varying optical properties. They showed that for sufficiently large dust aerosol optical depth and absorption, the total net effect is enhanced cloud cover for all but winter season. In some regions, cloud cover decreased, such as Eurasia, western North America, and South America. The largest cloud cover enhancement occurred over the oceans, central Africa, the Arabian Peninsula, India, and southeastern Asia. Over land, the cloud cover increased where the absorbing dust enhanced specific humidity due to increased moisture convergence driven by dust heating. Overall, this effect exceeds the reduced humidity that results from dust absorption, which enhances atmospheric evaporation. For mainly scattering dust particles with low AOD during wintertime, cloud cover enhancement responses were weak or even reversed.

Several simulations and observations about aerosol semidirect effect have been conducted over southern Africa [*Brioude et al.,* 2009; *Wilcox*, 2010]. *Miller et al.* [2004] used the GISS climate model and found that dust loading caused an increase in low-level cloud cover and precipitation. The author argued that in an arid region, where adiabatic heating was overwhelmed by longwave cooling and was balanced by subsidence, dust absorption in an aerosol layer aloft could reverse the atmospheric circulation, resulting in ascent and precipitation. *Sakaeda et al.* [2011] used 20 year run of the Community Atmospheric Model coupled to a slab ocean model. Over the ocean, where the aerosol layers are primarily located above clouds, negative TOA semidirect radiative effects associated with increased low cloud cover dominated over a weaker positive all-sky direct radiative effect. In contrast, over the land where the aerosols are often below or within cloud layers, reductions in cloud liquid water path led to a positive semidirect radiative effect that dominated over a near-zero DRE.

The effect of absorbing aerosols on cloud cover mainly depends on the altitude of the absorbing aerosols relative to the cloud and on cloud type [*Koch and Del Genio*, 2010]. Cloud cover decreases when aerosols are embedded within the cloud layer. Aerosols below a cloud layer may enhance convection and cloud cover. An aerosol layer above a cloud top stabilizes the underlying layer and tends to enhance stratocumulus clouds but may reduce the development of cumulus clouds. For example, *Takemura et al.* [2005] found that the areal coverage of shallow clouds increased by 20%–40% from clean to dusty condition. Absorbing aerosols can also increase cloud cover in convergent regions because they enhance deep convection and lower level convergence as moisture flows from the ocean to the land. Most global model studies indicated a regional variation in cloud response. In general, there is increased cloud cover over oceans and some land regions, with a net increase in lower level cloud cover and/or a reduction in upper level cloud cover. The result is a net negative semidirect effect feedback from the clouds in response to absorbing aerosols.

5.2. Observation of Semidirect Effect

Although model simulations addressed the potential importance of the semidirect effect, observational evidence for the semidirect effect strengthened the findings by using models. There are few reports discussing the Asian dust aerosol semidirect effect using observational data. Based on satellite observations, *Huang et al.* [2006b] studied the semidirect effect of dust aerosols on cloud water path over East Asia. They found that the water path of dust-contaminated clouds is considerably smaller than that of dust-free clouds. The mean ice water path (IWP) and liquid water path (LWP) of dusty clouds are less than their dust-free counterparts by 23.7% and 49.8%, respectively. The long-term statistical relationship derived from the International Satellite Cloud Climatology Project (ISCCP) data also confirms that there is a significant negative correlation between dust storm index and ISCCP cloud water path. This study showed some evidence of the semidirect effect of Asian dust aerosols on cloud properties. Analysis of satellite observations indicated that on average both local anthropogenic and natural dust aerosols transported into a region can significantly reduce water cloud particle size, optical depth, and LWP.

W. Wang et al. [2010] found that the LWP and IWP for dusty clouds were 31.2% and 21.9% less than those for unpolluted clouds over northwestern China. These results suggested that dust aerosols could warm clouds, increase the evaporation of cloud droplets, and further reduce cloud water path. *J.-P. Huang et al.* [2010] chose two semiarid regions in the same latitude with similar meteorological environments, one in China and another in the United States, to compare their aerosol effects. The authors found that the aerosols over China's semiarid region had a larger optical depth and higher absorption than those over the United States' semiarid regions and showed more significant semidirect effect. The semidirect effect may be the dominating factor in dust aerosol-cloud interaction over the arid and semiarid regions in East Asia and may contribute to the reduction of precipitation via a significantly different mechanism compared to that in Africa [*J.-P. Huang et al.,* 2006a, 2006b, 2010].

6. Dust-Cloud-Precipitation Interactions

Qian et al. [2009] analyzed the precipitation data collected across China over nearly half a century and found a persistent pattern of decrease in the occurrence of drizzle and light rain but of increase in the occurrence of heavy rain. Recent studies have suggested that these trends of precipitations were probably the results of aerosol effects, which could inhibit the rainfall in stratus clouds but invigorate the convective cloud to produce heavy rain.

6.1. Suppression of Drizzle and Light Rain

Aerosol indirect and semidirect effects are expected to reduce cloud particle size and even burn clouds, thus would suppress warm rain processes in stratocumulus clouds, small cumulus, topographic clouds, and so on [e.g., *Andreae et al.,* 2004; *Rosenfeld et al.,* 2001; *Khain et al.,* 2008; *J.-F. Huang et al.,* 2009a, 2009b, 2009c]. These hypotheses have been tested by results from both observations and model simulations. *Dai et al.* [2008] observed that as more aerosols entered clouds, precipitation was suppressed in the higher altitude of the clouds near the high altitude of Mounts Hua and Xi'an, China. *Yin and Chen* [2007] simulated the effects of mineral dust particles on the development of cloud microphysics and precipitation over northern China through a two-dimensional spectral-resolving cloud model and found that the heating caused by increasing dust aerosol loadings would suppress precipitation when dust particles acted as CCN and IN simultaneously during the development of a cloud. This is because the enhancement of CCN is nearly overwhelmed by the strong suppressing effect of IN. *Kawamoto* [2008] compared the amount of precipitation and nonprecipitation water cloud properties over China in the midlatitude frontal zone and found that the obtained relationships between precipitation and effective particle size/number density of droplets are consistent with the Twomey effect and aerosol scavenging by precipitation. In another dust prevalent arid region of West

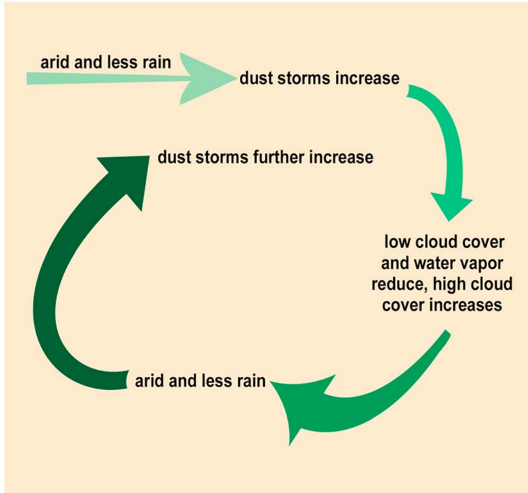

Figure 7. A schematic depiction of the feedback mechanism between precipitation and dust storm.

Africa, the suppression of precipitation induced by dust aerosols was also found [*Hui et al.,* 2008; *Solmon et al.,* 2008]. *Min et al.* [2009] focused on precipitation in a deep convection cloud system over eastern Atlantic and revealed that the consequence of microphysical effects of the dust aerosols was to shift the precipitation size spectrum from heavy precipitation to light precipitation.

Cloud droplets form on CCN and reduce the water vapor saturation barrier necessary for droplets to form. Higher aerosol concentrations result in more cloud droplet embryos competing for available water vapor. Changes in the number and size distribution of cloud droplets create thermodynamic and microphysical feedbacks that have the potential to change cloud evolution and properties. A significant feature of dust-cloud-precipitation interactions over the arid and semiarid areas is that it creates a positive feedback loop. The feedback loop begins with a decrease in rainfall and results in deficit in soil moisture. This leads to an increase in the occurrence of dust storms. Consequently, dust aerosols in the atmosphere warm clouds, increase the evaporation of cloud droplets and further reduce the cloud water path (the semidirect effect). This decreases the low cloud cover and water vapor amount, leading to less rainfall. The occurrence of dust storms would then increase, which could lead to even less rainfall. Figure 7 shows a schematic diagram of this feedback loop.

6.2. Increase of Heavy Rain

When it comes to deep clouds or storms, the stories about aerosol-cloud-precipitation interactions are completely different. Aerosol invigoration effect would enhance precipitation [*Petters et al.,* 2006; *Rosenfeld et al.,* 2008]. Microphysical processes associated with dust-cloud interactions also affect cloud hydrometeor profile and cause phase change, which, in turn, alter cloud dynamics and thermodynamics through latent heat release. Vertical precipitation profile can reflect the combined effects of dynamic, thermodynamic, and microphysical processes in cloud system [*Min et al.,* 2009]. *Koren et al.* [2012] examined the relationship between aerosol abundance and rain rate. The authors found that, for a range of conditions, increase in aerosol abundance was associated with local intensification of rain rate as detected by the microwave radiometer on board the Tropical Rainfall Measuring Mission satellite. This relationship was apparent over both ocean and land and in the tropics, subtropics, and midlatitudes. *Rosenfeld et al.* [2011] found that desert dust and heavy air pollution over East Asia possessed the ability to glaciate the top of developing convective clouds, creating ice precipitation instead of suppressing warm rain. *Li et al.* [2011b] used 10 years' worth of extensive ground-based and global A-Train (CloudSat, CALIPSO, and Moderate Resolution Imaging Spectroradiometer) space-borne satellite measurements to reveal strong climatic effects of aerosols on clouds and precipitation. The authors found that precipitation increased with aerosol concentration in deep clouds that had high liquid water content but decreased in clouds that had low liquid water content. Simulations using a cloud-resolving model confirmed these observations [*Zhang et al.,* 2007; *Tao et al.,* 2012].

7. Conclusions and Discussion

By reviewing a large number of studies, we reach the following conclusions:

1. Asian dust has two primary sources: the Taklimakan Desert and Inner Mongolian Gobi Desert. About 30% of the dust emission is redeposited onto the deserts and 20% is transported over regional scales. Parts of the dust travel across the Hexi Corridor and Loess Plateau to reach southeastern China. And the remaining 50% is transported over the global scale by entering the westerlies and flowing out from the continent to the open seas near Korea and Japan, and even reaching eastern North America.
2. Over the Asia arid and semiarid regions, dust aerosols have a cooling effect at the surface, a nearly warming effect or nil at the TOA, and a warming effect within the atmosphere. The largest forcing

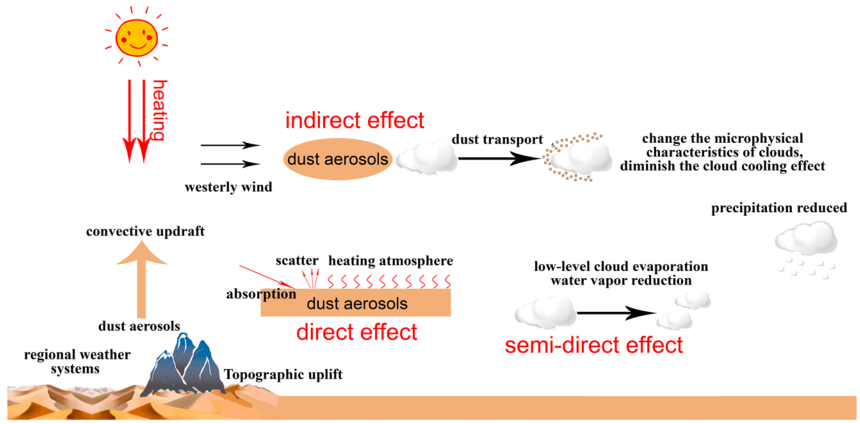

Figure 8. A schematic diagram summarizing dust-cloud-precipitation interactions in arid/semiarid and downwind (wet) regions.

values are located over or near dust sources. The dust aerosols from Asia are more absorptive than those from the Saharan Desert, as the former are easily polluted along their pathways.

3. Dust aerosols could modify cloud macrophysical and microphysical properties in many subtle ways. The first and second indirect effects are generally negative forcing on climate, while the invigoration effect and semidirect effect are positive. The quantification of those opposite effects is still highly uncertain.
4. Dust aerosols can suppress or enhance precipitation, depending on many factors, such as the vertical distribution of dust aerosols, humidity in the atmosphere, and cloud type. The amount of rain increases with aerosol concentration in deep clouds that have high liquid water content but decreases in clouds with low liquid water content. As for suppression of precipitation, there exists a positive feedback in dust aerosol-cloud-precipitation interactions (Figure 7). More dust storms can occur in the arid and semiarid regions, which may have contributed to the desertification observed in recent decades as well as accelerated occurrence of more arid conditions over the dryland of Asia. Figure 8 shows a schematic diagram summarizing dust-cloud-precipitation interactions in the arid/semiarid and downwind (wet) regions.

Recent studies have suggested that dust from arid regions can influence the microphysics of warm and cold cloud formation worldwide [*Twohy et al.,* 2009; *Stith et al.,* 2009; *Min et al.,* 2009; *J.-P. Huang et al.,* 2010]. However, dust aerosol effects and associated feedbacks that modulate cloud properties, rain formation, and cloud lifetimes are still the subject of much debate. Some studies claimed that the effects are fairly strong, while others suggested that the system is well buffered [*Stevens and Feingold*, 2009]. For example, large uncertainties exist in current estimates of aerosol RF because of our incomplete knowledge about aerosol-cloud interactions, as well as the distribution and physical and chemical properties of aerosols [*Yu et al.,* 2006]. The contribution of anthropogenic mineral dust to global dust emission has never been directly evaluated due to the lack of data; large uncertainties still exist in estimating relative contributions from natural and anthropogenic dust sources [*Ginoux et al.,* 2010]. Some aspects of the linkages between aerosols, clouds, and climate have been qualitatively confirmed (e.g., the suppression of precipitation by aerosols in warm rain), but some are still highly uncertain in terms of their magnitude (relationship between aerosol and surface precipitation), even their qualitative relationships (the effects of aerosol on cloud fraction, cloud LWP, and cloud lifetime). Untangling the interactions between aerosols and clouds remains a daunting task and hinders our ability to understand and predict the effects that anthropogenic activities have on regional and global climate change [*IPCC,* 2007].

Measurements conducted thus far to tackle this issue are unable to separate feedback processes arising from aerosol-cloud-climate interactions from total aerosol effects because real climate includes all microphysical and feedback mechanisms [*Takemura et al.,* 2007]. Using numerical weather prediction models, *Kischa et al.* [2003] and *Haywood et al.* [2005] suggested that a correct treatment of mineral dust and its radiative effects would improve the representation of the radiation budget and the accuracy of weather prediction itself. Studies on dust-cloud-precipitation interactions over the arid and semiarid regions provide insightful information for model diagnosis and improvements that are essential for projecting future climate change.

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