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博士学位论文

中国西北地区沙尘气溶胶辐射强迫效应的研究

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摘 要

沙尘气溶胶对太阳辐射能量强烈的吸收作用, 不仅会改变地表能量平衡, 而且还会通过改变大气的辐射加热结构, 加速大气低层水汽和云的蒸发, 从而抑制和减弱局域降水。因此研究西北地区气溶胶的辐射强迫作用, 对了解沙尘气溶胶对这一区域大气能量平衡和水份循环的影响, 特别是对西北地区干旱的影响, 有着重要意义。本论文利用 CERES、MODIS 和 CALIPSO 等卫星遥感资料及地面观测资料, 借助于辐射传输模式, 着重详细研究了沙尘气溶胶的直接辐射强迫效应, 定量计算了沙尘的直接辐射强迫值及其对大气辐射加热率的影响; 初步揭示了沙尘与云的相互作用对辐射能量平衡的影响, 并在此基础上提出了分离沙尘气溶胶直接辐射强迫效应与间接/半直接综合辐射强迫效应的方法。

西北沙尘气溶胶一直缺乏详细可靠的光学性质数据库, 区域辐射强迫的计算存在着很大的不确定性。本论文首次利用 2008 年中美联合观测的张掖站数据反演得到的沙尘粒子的光学特性, 特别是单次散射反照率, 计算了西北地区的辐射强迫。利用辐射闭合试验验证了反演的沙尘粒子参数的可靠性, 在试验成功的基础上计算了沙尘气溶胶在大气层顶和地表的日平均辐射强迫值。结果表明虽然沙尘气溶胶在大气层顶的辐射强迫值都接近于零, 但是在地表却表现出很大的负的辐射强迫值, 这种地表与大气层顶很大的差值说明, 大气中的沙尘气溶胶有效阻挡了太阳光到达地表, 增加了大气中太阳短波的吸收, 这势必会影响大气的垂直温度分布, 对大气层结的稳定和水分循环产生影响。

气溶胶的垂直廓线对大气加热率和辐射强迫的计算至关重要, 但由于观测手段的限制, 区域和全球尺度上的气溶胶垂直廓线的问题一直没有解决。本论文利用最新发射的 CAPLISO 卫星数据得到了塔克拉玛干地区沙尘气溶胶的垂直廓线, 并输入 Fu-liou 模式, 计算了沙尘辐射强迫值及加热率的垂直分布。结果表明由于沙尘的强吸收性, 使得很大一部分太阳辐射被吸收留在了大气中, 从而进一步加热大气。在强沙尘出现时, 伴随着沙尘过程的加强发展, 沙尘层大气的加热率可以从 1K day^{-1} , 2K day^{-1} 增加到 3K day^{-1} , 甚至沙尘最强时可以达到 5.5K day^{-1} 。这项工作首次定量评估了在塔克拉玛干沙漠中强沙尘暴发生时, 沙尘层对大气的加热作用。

本文还通过分析 4 年的 CERES 卫星资料比较了纯云区 (CLD) 和被沙尘污染过的云区 (COD) 云微物理特性和辐射强迫的变化, 再次证实了当云下有沙尘出现时, 沙尘粒子对云光学特性有显著的改变; 并发现沙尘的存在会明显减弱云在大气层顶的冷却效应, 云冷却效应的减弱可认为是由于沙尘增暖作用抵消的结果。而沙尘在 TOA 处的这种增暖效应, 不仅包括直接效应也包含了由云物理特性改变所引起的间接和半直接效应, 但是无法只利用卫星资料将直接效应和其他两种效应分开。本文首次提出了一种将卫星资料分析结果与模式模拟结果相结合来分离直接与间接效应的新方法, 来达到这一目的。利用此种方法估算的在 TOA 处沙尘短波瞬时直接辐射强迫值为 22.7 Wm^{-2} , 间接与半直接辐射强迫和为 82.2 Wm^{-2} , 分别占总辐射效应的 21.6 %和 78.4%。该方法的提出为定量研究云与气溶胶的相互作用提供了一个新的思路。

关键词: 沙尘气溶胶、直接辐射强迫、间接辐射强迫、西北地区、光学特性

Abstract

Dust aerosols have strong absorption of solar radiation. They not only modulate the surface energy, but also affect the heating rate of atmosphere which would enhance cloud evaporation, and weaken and suppress regional precipitation. The investigation of the dust aerosol radiative forcing is therefore important to understand the dust aerosol effects on the regional energy balance and hydrological cycle, over Northwestern China with the arid climate. In this study, we quantify the dust direct radiative forcing, and its influence on atmospheric radiative heating rate by using the CERES, MODIS and CALIPSO satellite data, ground-based observation and radiative transfer model. We also examine the radiation energy budget change due to the interaction of dust and cloud, where we propose a comprehensive approach to separate the dust aerosol direct radiative effect and indirect/ semi-direct radiative effect.

There are large uncertainties in estimation of regional radiative forcing of dust over northwestern China because of lacking detailed and reliable observations of dust optical properties. Herein we use the dust aerosol optical properties, retrieved from ground-based observations at Zhangye site during the 2008 China-US Joint Field Experiment, to calculate the dust aerosol radiative forcing. After verifying the reliability of the retrieved dust aerosol optical and microphysical properties through the radiation closure experiment, we calculate the daily mean radiative forcing of dust aerosol at both the top of atmosphere and surface. It is shown that dust aerosol has a strong negative forcing at the surface, but the magnitude of dust ARF at the TOA is very small. The large ARF difference between TOA and surface means that solar radiation is kept within the atmosphere. This would reduce the vertical temperature gradient, enhance the stability of the atmosphere, and thus slow down hydrological-cycle.

The vertical profile of dust aerosols is important to estimate the atmospheric heating rate, and is also important to calculate the dust ARF. However, it is difficult to

obtain the aerosol vertical profiles on the regional and global spatial scales until the active lidar observation becomes available from CALIPSO satellite. In this study, the CALIPSO lidar measurements are used to derive the vertical distribution of dust aerosols. We use the Fu-Liou radiation model along with the input of dust aerosol vertical profile to estimate the dust radiative forcing and atmospheric radiative heating rate. It is found that dust aerosols can maintain a large portion of solar energy in the atmosphere because of the strong absorption, thereby heating the atmosphere. During an evolution of a dust storm case, the dust aerosols heat the atmosphere (daily mean) by up to 1, 2, and 3Kday⁻¹, respectively. The maximum daily mean radiative heating rate reaches 5.5Kday⁻¹. This work quantitatively estimates the warm effect of dust layer during a dust storm in the Taklimakan desert for the first time.

In our study, four years of CERES data are also used to quantify the differences of cloud microphysical properties and radiative forcing between CLD (pure cloud) and COD (cloud over dust). We show that the analysis of satellite observations, dust aerosols do change cloud microphysical properties. It is shown that when dust exists under a cloud it has a warming effect at TOA, which combines direct, indirect and semi-direct effect. However, it is difficult, using only satellite observations, to separate the dust aerosol direct effect from those caused by altered cloud properties. We propose a new simple method that uses the satellite observations along with the radiation model calculations to separate these effects. It is shown that the averaged direct, and the combined indirect and semi-direct instantaneous shortwave radiative forcing are 22.7 and 82.2Wm⁻², respectively, which correspond to 21.6 and 78.4% of the total RF value. This method provides a new way to quantitatively study the interaction between clouds and aerosols.

Keywords: Northwest of China, Dust aerosol, direct radiation forcing, indirect radiation forcing, aerosol optical properties

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第一章 绪论

1.1 研究目的与意义

气候系统是极其复杂的多圈层系统,影响气候变化的因子多种多样。在诸多影响气候变化的因子中,云和气溶胶是两个非常重要,但又最不确定的影响因子(Haywood et al.,1999; Higurashi et al., 2002)。云和气溶胶对气候影响的不确定性,主要是由其辐射强迫机制的复杂性造成的。大气中气溶胶的辐射强迫机制远比 CO_2 等微量温室气体的温室效应复杂。与 CO_2 等温室气体相比,气溶胶在以下几方面与之有很大的不同。第一:与温室气体在大气的存活时间较长不同,气溶胶的生命周期较短,这就造成了气溶胶的时空分布不均的特性;第二:气溶胶对地气系统的强迫作用,并不像温室气体是均匀作用在全天和四季的,而是呈现出四季中夏季最强,而白天又强于夜晚的特性。第三:与其它分子吸收不同,即对某一种特定的种类,一个分子与另一个分子的吸收不会变化,但气溶胶单个粒子的光学性质会随粒子大小发生变化。第四:各种气溶胶粒子之间相互作用与气溶胶的浓度、粒子大小以及成分组成都非线性关系。所以,由于上述的原因,描述气溶胶对气候辐射强迫注定比温室气体的作用要复杂的多。其次,不同种类气溶胶的特性变化很大,这会影响它们吸收和散射辐射的程度,所以这就使得有的类型的气溶胶对地气系统起加热作用,有的反而起冷却作用。另外,气溶胶可作为云凝结核,影响云的微物理特性和宏观云量。不同的云也存在不同的辐射强迫。低云起冷却作用,高云则通常起加热作用。所以由于上述的云和气溶胶影响机制的复杂性以及云和气溶胶物理光学性质的系统资料的缺乏,对气溶胶和云的辐射强迫和气候效应的问题,至今并未得到很好的解决。但也因为云和气溶胶在气候变化中的重要作用,有关云和气溶胶的研究近年来一直受到高度重视,被列为全球变化研究中优先的项目(NRC,2001)。

沙尘是大气气溶胶的主要种类之一,对区域甚至全球的生态环境和气候变化具有重大影响。由于中国北方沙漠及土地的荒漠化,这些起源与中国塔克拉玛干沙漠和戈壁以及蒙古荒漠地区,常发于晚冬和初春的沙尘暴,使得我国西北地区成为世界上仅次于撒哈拉沙漠的第二沙尘暴频发地。每年中国北方输入大气的沙尘达 800TgYr^{-1} , 约占全球总量的 $1/3$ 。大量沙尘粒子被强风吹卷入大气,并随高

空气流向下游陆地及海洋地区远程传输,使沙尘影响范围扩大到全球尺度,沙尘长距离的输送也被多个观测结果所证实 (Haywood et al.,1999;Higurashi & Nakajima 2002, Huang et al.,2008)。沙尘气溶胶不但可以将入射的太阳辐射反射到外太空,可作为云的凝结核影响云滴的数浓度,进而改变云的光学厚度和生命期,而且由于沙尘粒子的吸收特性,沙尘气溶胶还可以影响大气的辐射加热结构,从而改变大气环流并对气候造成影响。近年来,对云和沙尘气溶胶的相互作用已经引起了国内外学者的广泛关注,并且一些结果已经发现了非洲撒哈拉沙尘气溶胶与云相互作用的一些性质。Rosenfeld [2001] 等通过卫星资料分析发现含有沙尘气溶胶的云滴很少通过碰并过程而产生降水;在另一篇文章 (Rosenfeld [1996]) 撒哈拉沙尘气溶胶的冰核作用来解释以色列的播云试验。并与 Levi (Levi and Rosenfeld, 1996) 指出以色列沙尘多发季节冰核浓度也有增加的现象。虽然塔克拉玛干和戈壁沙漠也是沙尘气溶胶的主要源区,然而对中国这个重要沙尘区,沙尘气溶胶与云和气候反馈的相互作用研究甚少。黄建平等 (Huang et al., 2006a; 2006b; 2006c) 通过 MODIS 和 CERES 资料的分析发现,在沙尘区,沙尘粒子和云粒子相互作用,使得气溶胶通过减少云水含量和光学厚度,对云冷却作用的抑制可能会导致 $1\pm0.6\text{Wm}^{-2}$ 的增暖,这将对西北地区乃至东亚地区的气候产生显著的影响。最近,葛颀铭等发 (Ge et al., 2009) 现中国西北沙尘气溶胶与非洲撒哈拉地区的沙尘相比,其吸收性更强。而这一特性将会使得沙尘气溶胶对大气低层产生加热作用,加速大气低层水汽和云的蒸发,从而抑制和减弱降水,这一过程或许将会加剧西北地区的干旱。而上述的这些结论都只是个例分析,还需大量观测数据和数值模拟结果进行验证。所以,进行西北地区沙尘气溶胶的辐射特性的深入研究以及沙尘气溶胶对云和降水影响的数值模拟研究,特别是对云的光学物理特性的改变,对探明沙尘气溶胶-云-陆面过程的反馈相互,进一步揭示其对西北干旱气候的影响至关重要。

1.2 国内外相关研究进展

气溶胶影响地气的辐射平衡,主要是通过三种作用方式:直接辐射强迫效应、间接辐射强迫效应以及半直接辐射强迫效应。图 1.1 是第四次 IPCC 报告给出的 2005 年各种因子构成的全球平均的辐射强迫值。从图中我们可以看出目前气溶

胶的辐射强迫值的误差范围最大,无论直接效应还是间接效应,特别是间接效应,在所有因子中都是最不确定的,是造成估算人为净辐射强迫时误差较大的原因。这表明,目前气溶胶以及与云相互作用的各种研究结果差别很大,气溶胶对气候系统的影响至今还处于很低科学认知水平。如要减小气溶胶辐射强迫的不确定性,就必须解决以下两个问题:1.各种关键气溶胶的时空分布及光学性质。2.各种相互作用强迫机制 (Penner et al.,1994)。图 1.2 列出了大气种 4 中关键的气溶胶的来源,及其不同的强迫机制。

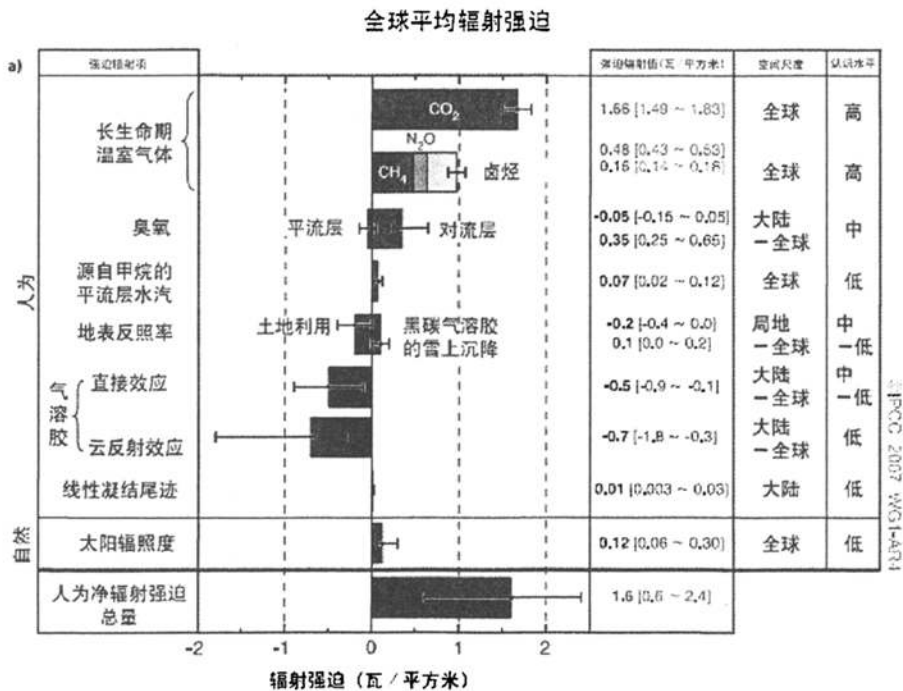


图 1.1 2005 年各种因子和物理构成的全球平均的辐射强迫 (引自 IPCC, 2007)

表 1.1 主要气溶胶种类、来源与相关的物理强迫机制（引自 Penner et al., 1994）

主要种类	主要来源	强迫机制
水溶性无机物 (如：硫酸盐、硝酸盐、铵盐以及煤炭燃烧产生的硅)	传输、当地产生、工业污染物、施肥过程（铵盐）、当地动物排泄	A. 直接作用：晴空下对太阳辐射的散射 B. 间接作用：使得 CCN 改变从而影响云反照率 C. 间接作用：使得 CCN 改变从而影响云滴寿命
浓缩有机物	传输、当地产生、工业污染物以及农作物燃烧	A, B, C
黑碳	传输、当地产生、工业污染物以及农作物燃烧	D. 吸收太阳辐射，改变大气加热率
矿物沙尘	荒漠化引起的沙土随大气移动加剧	A, B, C, D 及吸收和放出长波辐射

1. 2. 1 气溶胶直接辐射强迫的研究现状

虽然人们认识到气溶胶粒子会通过散射和吸收太阳辐射来直接的影响地球辐射平衡（McCormick and Ludwig, 1967; Charlson and Pilat, 1969; Atwater, 1970; Mitchell Jr., 1971; Coakley et al., 1983）这个现象，已经超过了 60 年，但是量化这个影响值的过程却进展缓慢，并且至今并未很好的得到解决. 区域气溶胶直接辐射强迫主要是由其粒子的光学特性、区域地表反照率以及沙尘垂直分布所决定的，当气溶胶类型为吸湿性粒子，还和当地水汽含量有关（Haywood and Boucher, 2000）。气溶胶的光学性质，特别是单次散射反照率，决定粒子对太阳辐射的散射吸收，是区域直接辐射强迫值的主要决定因子。但一些研究发现对于吸收性气溶胶，当在气溶胶粒子单次散射反照率一定的情况下，在地表反照率较大的地区为正的辐射强迫而在低反照率地区则为负的辐射强迫（Haywood and Shine, 1997; Chylek and Wong, 1995）。全球年平均的气溶胶直接辐射强迫则是由此种气溶胶光学性质和其在全球的时空分布共同影响的（Haywood and Boucher,2000）。第四次 IPCC 报告指出，目前模式估算全球平均的硫酸盐的辐射强迫约为 $-0.4\pm0.2\text{Wm}^{-2}$ ；来自化石燃料的有机碳的辐射强迫为 $-0.05\pm0.05\text{Wm}^{-2}$ ；来自化石燃料的黑碳的辐射强迫为 $+0.2\pm0.15\text{Wm}^{-2}$ ；生物质燃烧的辐射强迫为 $+0.03\pm0.12\text{Wm}^{-2}$ ，硝酸盐的辐射强迫为 $-0.1\pm0.1\text{Wm}^{-2}$ ；矿物粉尘的辐射强迫为

$-0.1 \pm 0.2 \text{ Wm}^{-2}$ 。虽然以上数据表明单一气溶胶的直接辐射强迫仍不确定,但 IPCC 指出由于大气模式的改进,许多模式现在能够描述所有重要的气溶胶分量。在北半球许多大陆地区,可以清楚地辨别出主要由硫酸盐、有机碳、黑碳、硝酸盐和工业粉尘组成的工业气溶胶。实地观测、卫星和地基观测的改进已经能够验证全球气溶胶模式的模拟结果。这些改进允许首次对气溶胶的总直接辐射强迫进行量化。现在首次可以得到所有气溶胶的总直接辐射强迫为 $-0.5 \pm 0.4 \text{ Wm}^{-2}$ 。这是第四次 IPCC 报告与第三次相比主要的一个进展 (IPCC, 2007)。

IPCC 报告中估算的沙尘全球年平均的直接辐射强迫范围为 -0.3 Wm^{-2} 到 0.1 Wm^{-2} , 相比较硫酸盐等其它气溶胶有较大的不确定性。这主要是因为沙尘粒子光学性质及分布的不确定性而造成的。图 1.3 为 Sokolic 等 (Sokolic et al., 1996) 总结的沙尘气溶胶各物理量的代表值及不确定范围。从表中可以看出无论是每年沙尘量的排放量以及生命周期及各种光学性质都有很大的不确定性。以沙尘粒子的单次散射反射率为例,多个非洲撒哈拉地区沙尘研究结果表明撒哈拉沙漠源区沙尘气溶胶粒子的单次散射反射率在 $0.55 \mu\text{m}$ 波段上为 0.95 到 0.99 之间 (Fouquart et al., 1987; Haywood et al., 2003; MacFarlane et al., 2008), 但最新结果显示来自我国西北沙尘源的沙尘粒子的单次散射反照率远小于非洲撒哈拉源区的粒子, 吸收性较强, 在 $0.5 \mu\text{m}$ 波段上, 数据范围在 0.73-0.85 间 (Ge et al., 2009; Pandithurai et al., 2008)。虽然近几年地面观测试验的增多及卫星数据反演精确性的提高, 沙尘气溶胶全球辐射作用的认知水平有所提高, 但由于缺乏不同地区沙尘粒子光学特性的数据库, 以及区域地表反照率的复杂性, 使得区域沙尘气溶胶的辐射强迫值较全球平均值有更大的不确定性。本论文则利用近年地面加强观测试验得到的我国西北沙尘的主要光学特性及卫星反演的地表特性, 主要研究了我国西北地区沙尘的直接辐射强迫。结果表明我国西部沙尘气溶胶在大气层顶的辐射强迫接近于零, 这说明我国西北地区强吸收性的沙尘气溶胶吸收太阳辐射, 使很大一部分能量都留在了大气中, 从而改变了大气的垂直加热结构 (详见第四章)。

表 1.2 沙尘各参数的代表值和不确定范围（引自 Sokolik et al., 1996）

沙尘对于太阳辐射强迫的贡献		
参数变量	代表值	不确定范围
沙尘年产量 (TgYr^{-1})	3000	1000-5000
矿物气溶胶生命周期	4	3-7
矿物气溶胶质量消光系数 (m^2g^{-1})	1	0.2-2
认为产生的矿物气溶胶光学厚度的全球平均估算值	0.014	0.003-0.05
500nm 上矿物气溶胶的 单次散射反照率 (SSA)	0.85	0.65-0.95
500nm 上矿物气溶胶的 不对称因子 (g)	0.75	0.65-0.88

1.2.2 气溶胶间接辐射强迫的研究现状

如图 1.2 所描述, 气溶胶间接效应通常分为两类: 由于气溶胶粒子增加, 在水汽含量一定时, 使云滴增加而云滴粒子减小的作用, 称之为第一类间接效应 (Twomey, 1974); 在低层大气中, 水汽凝结在悬浮在空气中的气溶胶粒子上, 就形成了云滴。所以, 气溶胶粒子 (云凝结核 CNN) 的浓度、大小以及吸湿性都会对云滴的浓度和大小有直接影响。这些影响进一步改变云的短波辐射效应。特别应指出, 当增大凝结核的浓度时, 云滴的浓度会随之增大, 从而增大了云的短波反照率 (Twomey et al., 1977, 1984; Coakley et al., 1987; Durkee et al., 1988)。所以第一类间接又可以被称为“云反照率”作用。第二类间接效应是指由于云滴粒子减小伴随的云滴浓度增大, 而可能会减少降水概率, 从而改变云的厚度 (Pincus and Baker, 1994) 及生命周期 (Albrecht, 1989)。这种延长云的生命周期以及增大云覆盖率的作用不仅会增强云的短波冷却辐射作用, 也会增强长波增暖辐射作用。但第二类作用主要还是影响低云, 所以低云的冷却效应成为主要作用, 使得最终的净辐射作用为一种冷却作用。另外, 降水概率的减少可能会进一步改变大气中水汽含量和热量的垂直分布, 这将会改变地球的水份循环 (Charlson et al.,

1992)。另外一些吸收性气溶胶，例如：黑碳，沙尘，吸收太阳辐射加热大气边界层，蒸发低层云和水汽使得地表增温的半直接作用（Hansen et al., 2000），也会对水汽含量和分布产生影响。

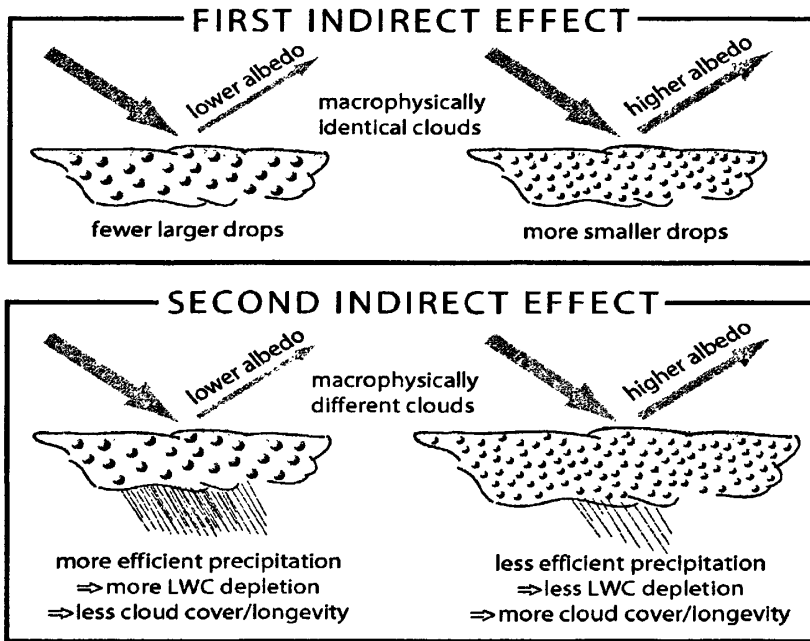


图 1.2 气溶胶间接作用示意图（引自 Albrecht, 1989）

第一类间接效应一直是气溶胶间接作用的研究重点。气溶胶第一类间接效应已经被多个研究结果证实：一些飞机穿云试验（Martin et al., 1994; Guillepe et al., 1996; Pawlowska et al., 2000; Taylor et al., 1994; Mcfarquhar et al., 1987）结果表明气溶胶会增加云的凝结核和云滴的数密度并减小云滴粒子的大小。卫星资料也在较大尺度上增强了局地穿云试验的结果（Coakley et al., 1987; Kaufman et al., 1997; Nakajima et al., 2001）。然而到目前为止，并没有可靠的观测数据发现气溶胶粒子会使云生命周期增长的第二类间接作用。前三次的 IPCC 报告（IPCC, 1994, 1996, 2001）都只考虑了第一类间接效应，最新 IPCC 报告也只是估算了第一类间接效应的辐射强迫值，并指出第二类间接作用应该被看成气候响应的一部分，而不是辐射强迫（IPCC, 2007）。后文中气溶胶间接效应都指第一类间接作用。

气溶胶间接辐射强迫的估算是直接强迫估算不稳定性的 2 到 3 倍 (IPCC, 2001; Haywood and Boucher, 2000)。自第三次 IPCC 报告以来, 估算气溶胶间接效应的全球模式数量已经大大增加, 并且以更为严格的方式对这些估值进行了评估。第四次 IPCC 指出首次估算了气溶胶对水云的影响 (第一类间接效应) 产生间接辐射强迫值为 $-0.7[-1.8\sim-0.3]\text{Wm}^{-2}$, 这一结果来自多个模式的研究, 更详细地描述了气溶胶-云的相互作用过程。包括更多气溶胶种类或受卫星观测资料约束的模式研究。尽管此次估算的结果已经比第三次 IPCC 报告结果有了进展, 缩小了强迫估计值的范围, 但在观测和过程模拟两方面仍然存在很大的不确定性。表 1.3 为总结的最新 IPCC 和其他文章发表的各种气溶胶直接和间接辐射强迫的估算值的对比。

我国西北沙尘气溶胶作为一种常见的强吸收性气溶胶, 其与云的相互作用, 近年来已经引起了国内科学家的广泛关注。黄建平等 (2006a, 2006b, 2006c) 利用 MODIS 和 CERES 资料发现当云被沙尘污染后形成的沙尘云, 其光学性质和物理特性都有了显著变化, 沙尘云云滴的云水路径及粒子半径明显减小, 云的辐射强迫减弱, 同时由于沙尘气溶胶的加热作用, 抑制了降水, 直接影响陆-气间水分交换。本论文结合 CERES 卫星观测资料和模式模拟结果, 分析了 11 个沙尘云个例, 估算出间接和半直接辐射强迫约占总辐射效应的 78.4%, 是沙尘气溶胶的主要作用 (详见第五章)。但这些结果都是由个例分析得到的, 还需要大量数据和模式模拟结果得以验证。

表 1.3 气溶胶直接和间接辐射强迫估算值的对比

强迫机制	气溶胶种类	2007 年 IPCC (Wm^{-2})	Haywood 等人结果 (Wm^{-2})
直接效应	硫酸盐	-0.6~-0.2	-0.82~-0.26
	黑碳	0.05~0.35	0.16~0.42
	有机碳	-0.1~0	-0.04~-0.02
	生物质燃烧	-0.09~0.15	NA
	矿物沙尘	-0.3~0.1	-0.46~0.09
	硝酸盐	-0.2~0	NA
间接效应	云反照率	-1.8~-0.3	-1.8~-0.3
	云生命周期	NA	-0.4~2.3
	飞机轨迹云	0.003~0.03	0.005~0.06

1.3 论文主要内容安排及创新点

1.3.1 论文主要内容和章节安排

本论文主要研究内容为我国西北地区沙尘气溶胶的直接和间接辐射强迫效应。论文第一步进行了各影响因子对辐射强迫估算的敏感性试验,以期了解各因子对估算的结果的重要性。第二步在计算沙尘直接辐射强迫值时,首先选取了2006年塔克拉玛干沙漠地区的一次强沙尘暴过程,计算了此区域的沙尘辐射强迫并利用最新的CALIPSO卫星资料得到沙尘气溶胶的垂直分布廓线,首次计算了沙尘的垂直加热率。然后利用2008中美联合观测试验张掖站反演的沙尘粒子的光学特性,输入模式计算大气层顶及地表辐射通量,并将模式模拟值与观测值进行比较,做辐射闭合试验。在辐射闭合的情况下估算沙尘气溶胶在大气层顶及地面的辐射强迫值。第三步则进行了沙尘间接辐射效应的研究。首先利用4年CERES卫星资料分析了沙尘对云光学性质的改变,其次利用观测数据计算了沙尘总的辐射强迫值,然后将模式计算与观测结果相结合估算了间接辐射强迫值与直接辐射强迫值的比率。

依照以上的论文的主要研究内容,本论文章节安排如下:

第一章:引言。主要阐述了气溶胶特别是沙尘气溶胶的研究的重要性,并介绍了气溶胶直接和间接辐射效应最新的研究进展。

第二章: Fu_Liou 与 SBDART 辐射传输模式的简介。主要介绍了本论文中主要使用的两种辐射传输模式的主要原理和优点。

第三章:模式输入参数对辐射强迫值的敏感性试验。主要测试了在计算辐射强迫值时,各个输入变量对计算结果的重要性。

第四章:沙尘气溶胶的直接辐射强迫效应。此章为本论文的一个主体,主要计算了张掖及塔克拉玛干地区沙尘在大气层顶和地表的直接辐射强迫值以及在塔克拉玛干地区的垂直加热率。

第五章:沙尘气溶胶的间接辐射强迫效应。此章为本论文的另一个主体,主要研究了当云被沙尘污染后云的光学特性的变化,计算了沙尘总的辐射强迫值,并估算了直接和间接效应所占的百分比。

第六章:论文总结和下一步的工作计划。

1.3.2 论文的主要创新点

本论文的研究利用了多个最新的卫星及地面观测试验的数据，主要有以下几个创新点。

第一：我国西北沙尘粒子一直缺乏可靠详细的光学性质数据库，对区域辐射强迫的计算产生很大的不确定性。本论文首次利用最新的 2008 年中美联合地面观测数据得到的沙尘粒子的光学特性，特别是单次散射反照率，计算了西北地区的辐射强迫。

第二：气溶胶的垂直廓线对大气加热率的计算至关重要，对辐射强迫的计算也有一定的影响，但由于观测手段的限制，垂直廓线的问题一直没有解决。2006 年 4 月发射的 CALIPSO 卫星上装载的激光雷达，为我们提供了云和气溶胶的垂直分布给了我们一个前所未有的机会去研究气溶胶辐射强迫和加热率的垂直结构。本论文利用了从 CAPLISO 卫星数据得到的塔克拉玛干沙漠的沙尘气溶胶的垂直廓线，并输入 Fu-liou 模式，计算了沙尘辐射强迫值及加热率的垂直分布。

第三：气溶胶与云的相互作用的研究一直是一个难点。本论文利用多年的 CERES 卫星资料再次证实了沙尘粒子对云光学特性的改变，并与模式模拟结果结合，首次提出了一种分离直接辐射效应和间接辐射效应的方法，估算了直接和间接作用分别占总辐射强迫值的百分比。

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第二章 辐射传输模式

2.1 引言

在计算辐射强迫时,我们需要使用辐射传输模式来进行模拟计算。本文在进行辐射通量闭合试验和计算辐射强迫时,我们根据 Fu-Liou 辐射传输模式和 SBDART 辐射传输模式不同的特性,在不同情况下分别使用了这两种辐射传输模式。这一章对这两个模式进行了详细介绍。

2.2 Fu_Liou 辐射传输模式

Fu-Liou 辐射传输模式是由美国华盛顿大学付强教授与加州大学廖国男教授发展建立的(Fu and Liou, 1992, 1993),此模式建立的目的是为大气数值模式的积分过程提供一个精确而快速的辐射通量和加热率的计算方法,此模式随后得到 Rose 和 Charlock 以及 Kato 等人完善发展(Rose and Charlock 2002; Kato et al., 2005),现在已经被 NASA-langley 研究中心用于 NOAA 和 NASA 卫星观测的大气和地表通量的数据的日常业务反演。Fu-Liou 辐射传输模式是最早在模式中用统一的方法处理太阳和红外辐射的模式之一。该模式的有以下几个最大优点:首先模式中的辐射传输方案以 delta 四流近似为基础,同时提供可选的 delta 二流近似和 delta 二流/四流近似。Delta 二流和四流近似都具有很高的计算效率,但二流与四流相比,虽然二流计算速度更快,但在有云的情况下辐射通量和加热率的计算会产生显著的误差(Fu et al., 1997)。其次模式中运用了相关 K-分布方法进行非灰气体吸收的参数化。另外一个突出的优点是模式中云辐射特性的参数化考虑了卷云包中的非球形冰粒子。

Fu-Liou 模式主要计算大气层层顶和地表处及大气各层、向上和向下方向上、太阳及红外波段的辐射通量。但由于模式计算采用 4 流或者 2 流近似,所以只能计算辐射通量,并不能计算某个特定角度的辐射亮度。模式还可以计算辐射通量的垂直分布以及垂直加热率。计算的垂直层数可以根据不同的计算要求来确定。

表 2.1 列出了 Fu-Liou 模式计算波段的划分。Fu-Liou 模式共划分为 18 个波段:其中短波波段 6 个,并且又将第一个短波波段 0.2-0.7 μm 的可见光波段分为了 10 个小波段。短波波段上主要考虑了 2500-14500 cm^{-1} 波段上水汽的吸收带; 50000-14500 cm^{-1} 波段臭氧的吸收带; 2850-5200 cm^{-1} 波段的二氧化碳吸收带;

以及氧气的 A, B 和吸收带。长波波段 12 个, 长波波段上考虑了 0-2200cm⁻¹ 波段上的水汽吸收带, 540-800cm⁻¹ 波段的二氧化碳吸收带; 980-1100cm⁻¹ 波段臭氧的吸收带; 1100-1400cm⁻¹ 波段的甲烷吸收带和同样波段上的一氧化二氮吸收带。

表 2.1 Fu-Liou 模式计算波段波长范围

	短波波段 (um)	可见光波段 (um) (0.175-0.7)	长波波段 (cm ⁻¹)
1	0.175-0.7	0.1754-0.2247	2200-1900
2	0.7-1.3	0.2247-0.2439	1900-1700
3	1.3-1.8	0.2439-0.2857	1700-1400
4	1.8-2.5	0.2857-0.2985	1400-1250
5	2.5-3.5	0.2985-0.3225	1250-1100
6	3.5-4.0	0.3225-0.3575	1100-980
7		0.3575-0.4375	980-800
8		0.4375-0.4975	800-670
9		0.4975-0.5950	670-540
10		0.5950-0.6896	540-400
11			400-280
12			280-0

Fu-Liou 模式中输入的大气廓线包括气压、温度、水汽和臭氧 4 个参数的垂直分布。模式中自带有 5 种大气廓线, 分别是: 热带地区标准大气廓线; 中纬度地区冬季、夏季的标准大气廓线; 副极地地区冬季、夏季的标准大气廓线。计算时可根据不同情况选择使用, 或者可以自己输入计算地区特定的大气廓线。

模式中地表反照率的输入有两种方法。第一种是在假设地表均一的情况下, 认为地表反照率是不随波长变化的, 只需输入宽谱的地表反照率。第二种则是要考虑不同地表类型, 地表反照率随波长的变化, 在输入宽谱地表反照率的同时, 要给定一种地表类型, 使宽谱的地表反照率按此类地形反照率的波谱分布, 得到各波段的地表反照率; 或者也可以直接输入每个波段的地表反照率。模式中给出了 18 种不同地表类型的地表反照率随波长变化的波谱分布, 这是根据 IGBP (International Global Biosphere Product) 的分类得到的, 分别是: 常绿针叶林、常绿阔叶林、落叶针叶林、落叶阔叶林、混合林、茂密灌木丛、稀疏灌木丛、亚

热带树林、亚热带草原、草地、湿地、农田、城市、相间农田、永久雪地、荒原/沙漠、海洋、苔原、新雪和海冰。这 18 种类型基本包括了所有基本的地表类型，计算时可以从 MODIS 卫星可以得到宽谱的地表反照率，再从这 18 种基本类型中选择一种要计算的相似地表类型，就可以得到每个波段的地表反照率的值。

模式中可以选择分别计算云的辐射影响以及各类气溶胶的辐射影响。在计算有云时的辐射通量时，模式中可放置任意层云，如两层云就是即可以计算既有高云又有低云时的情况。模式中云粒子的单次散射特性按照云水含量 (LWC) /冰水含量 (IWC) 以及有效粒子半径 (Re/De) 进行了参数化 (Fu et al., 1993; Fu 1996; Fu et al., 1998)，所以在原版模式输入时需要输入这两个参数量。因为利用云的光学厚度和尺度，也可以通过转换公式得到云水/冰水含量，所以在新版中，也可以通过输入云的光学厚度、云顶、云底高度以及粒子有效半径来计算。水云的有效粒子范围通常在 4 到 30 μm ，而冰云的有效粒子半径通常都大于 20 μm ，所以模式中水云和冰云的有效粒子半径范围分别为 5-30 μm 和 20-180 μm 。当冰云有效粒子半径小于 20 μm 时就会出现不合理的计算结果。

在计算气溶胶的辐射作用时，模式可设置一种单一成分气溶胶，也可以设置成由两种不同成分组成的气溶胶。模式中提供了 18 种气溶胶类型，计算时当选择了其中某种后，就确定了此种气溶胶的吸收和散射特性。表 2.2 给出了这 18 种粒子在 0.67 μm 上的单次散射反照率和不对称因子。其中除了不溶性粒子 (Insoluble Mode)、集结性矿物沙尘 (Nuclearion Mode)、堆积性矿物沙尘 (Accumulation Mode)、粗糙性矿物沙尘 (Coarse Mode)、传输性矿物沙尘 (Transported Mode), 这五种类型外，其他 13 种气溶胶的光学特性与水汽有关，会随相对湿度而变化。表 2.2 列出的这 13 类气溶胶的单次散射反照率和不对称因子是在相对湿度为 (50%) 时的参考值。对于气溶胶的垂直分布，计算者即可输入每层的光学厚度，但需要注意出入的层数应该和大气廓线层数一致；又可以直接利用模式中的指数分布，只要给定高度，则从地面到这一高度的气溶胶会按指数分布递减。

表 2.2 Fu-Liou 模式中 18 种气溶胶类型的单次散射反照率及不对称因子

气溶胶类型	单次散射反照率 (SSA)	不对称因子 (ASY)
海洋性粒子(Marine)	1.0	0.8113
大陆性粒子(Continental)	0.9565	0.6293
城市性粒子(urban)	0.9327	0.6270
0.5 μm 沙尘	0.9653	0.6622
1.0 μm 沙尘	0.9255	0.6931
2.0 μm 沙尘	0.8715	0.7725
4.0 μm 沙尘	0.7949	1.0
8.0 μm 沙尘	0.6996	0.8795
不溶性粒子(Insoluble)	0.7486	0.8163
水溶性粒子(water soluble)	0.9730	0.6570
烟尘 (soot)	0.1789	0.3079
堆积性海盐(Accumulation)	1.0	0.7724
粗糙性海盐(Coarse)	1.0	0.8464
集结性矿物沙尘(Nuclearion)	0.9740	0.6521
堆积性矿物沙尘(Accumulation)	0.9093	0.7162
粗糙性矿物沙尘(Coarse)	0.7118	0.8741
传输性矿物沙尘(Transported)	0.8763	0.7532
硫酸盐粒子(Sulfate Droplets)	1.0	0.7634

2.3 SBDART 辐射传输模式

SBDART (Santa Barbara DISORT Atmospheric Radiative Transfer) 辐射传输模式是由美国加利福尼亚大学地球系统计算学院的地球空间研究组在 1998 年开发的, 用来计算在晴空和有云条件下地球大气和地面的辐射传输。模式中考虑了能影响紫外、可见和红外辐射的所有重要的吸收和散射过程。SBDART 模式其实主要是集合了精确的离散坐标辐射传输模式 DISORT (Stamnes et al.,1988)、低

分辨率大气透射模式 (Kneizys et al., 1983) 和水滴和冰晶对光散射的 Mie 散射结果 (Wiscomb's, 1980; Stackhouse, 1991), 这三部分内容而成的。此模式不仅适用于大气辐射能量平衡方面的研究而且也可用于遥感问题的研究。

与其他模式相比, SBDART 有计算波长范围长, 波段分辨率高的优点。SBDART 模式可计算波长范围为 0.2-100 μm , 在可见光波长范围内波段分辨率为 5nm, 在热红外波长范围内波段分辨率为 200nm。SBDART 波段划分的高分辨率给模式结果与观测数据的对比提供了很大的便利, 使得模式计算结果可以与不同波长范围的地面辐射仪器观测结果进行验证比较。关于气体的吸收, SBDART 采用了 LOWTRAN 7 中采用的低分辨率带模式, 并且考虑了大气中所有相对活跃的气体分子的吸收作用。此种方法是基于逐线计算方法的, 与 Fu-Liou 模式中基于的相关 K 分布法计算方法不同。

SBDART 中使用了 DISORT (Discrete Ordinate Radiative Transfer; Stamnes et al., 1988) 即离散坐标系解法来数值求解辐射传输方程。此种方法提供了一种稳定的数值方法来求解垂直不均一的平面水平大气的辐射传输方程。此种方法可以不同高度上, 计算任意角度上长、短波的辐照亮度。DISORT 方法是用来处理水平平行模式大气中的辐射传输问题, 这就假设大气是水平均匀而垂直变化的。但实际中很难准确的找到大气是水平平均一的这一假设成立的状况。但总的来说, 当水平大气的变化尺度远大于垂直大气的变化尺度时, 水平方向的辐射传输并不是那么重要。所以实际大气可以假定为平面水平大气。在 DISORT 中, 计算时使用了平面地球投影坐标。当计算较大的太阳高度角时, 一般的平面地球投影的假设会使得计算结果出现显著误差, 但在 DISORT 中补充了一阶修订, 即用正确的球面几何法来计算直接光的消光。Dahack 和 Stamnes (Dahack and Stamnes, 1991) 证明, 即使当太阳高度角到达 90°后, 利用此种近似方法也可以得到足够精确的计算结果。

SBDART 模式中大气的垂直分层可根据计算者的要求进行划分, 模式中自带的 6 种大气廓线的垂直分层为 35 层。与 Fu-Liou 模式相比, SBDART 多加了一种廓线: US62, 这是一条可以代表典型美国大陆状况的大气廓线。这六种廓线 (McClatchey et al., 1972) 都是在大气科学领域得到广泛应用的大气廓线, 包

括气压、温度、水汽、臭氧的标准垂直廓线。另外，计算者也可以自行输入大气廓线。

SBDART 模式中用 5 种基本的地表类型来参数化各波段的地表反照率。这五种类型分别是：海水 (Tanre et al., 1990)，湖水 (Kondratyev et al., 1969)，植被 (Reeves et al., 1975)、雪 (Wiscombe and Warren., 1980) 和沙地 (Staetter and Schroeder 1978)。虽然，SBDART 中的地表类型没有 Fu-Liou 模式中的齐全，但对于一个特定的地表来说，其各波段的地表反照率都可以通过混合这些基本类型来很好的近似。SBDART 模式中，计算者可以给定水、植被、雪和沙地这四种基本类型的比例，来得到计算者所需要的地表类型。比如：将水、植被和沙地这三种类型结合，其混合得到的各波段的地表反照率可以很好的代表常绿林带的地表反照率波谱；如将水和沙地按一定权重比例混合，其地表反照率波谱就和湿地的各波段的地表反照率很近似。在 SBDART 模式中，假设地表反射辐射的角度是各向同性的，也就是说各波段地表反照率的值不随着太阳高度角的变化而变化。这种朗伯反射的假设虽然在很多情况下是适用的，也是在目前缺乏反照率随角度变化数据的情况下一个很好的选择，但是也应该注意，在某些角度下，这种假设会造成的很大的误差。

在计算有云和气溶胶情况下的大气辐射传输时，需要云和气溶胶粒子的消光系数，单次散射反照率和不对称因子这三个光学特性参数。在 SBDART 模式中，将云粒子光学性质按照有效粒子半径进行了参数化，预先利用 Mie 程序计算了有效粒子半径在 2-128 μm 范围内的所有云滴粒子的散射系数、单次散射反照率和不对称因子。图 2.1 所示为 SBDART 模式中云滴粒子有效直径分别为 2、8、32 和 128 时消光系数、单次散射反射率和不对称因子随波长的变化。所以只需在模式中输入云粒子有效半径，就可以直接得到相对应的消光系数、单次散射反照率和不对称因子，再输入另一个参数云水含量（或是云光学厚度）就可以计算有云情况下的辐射通量。

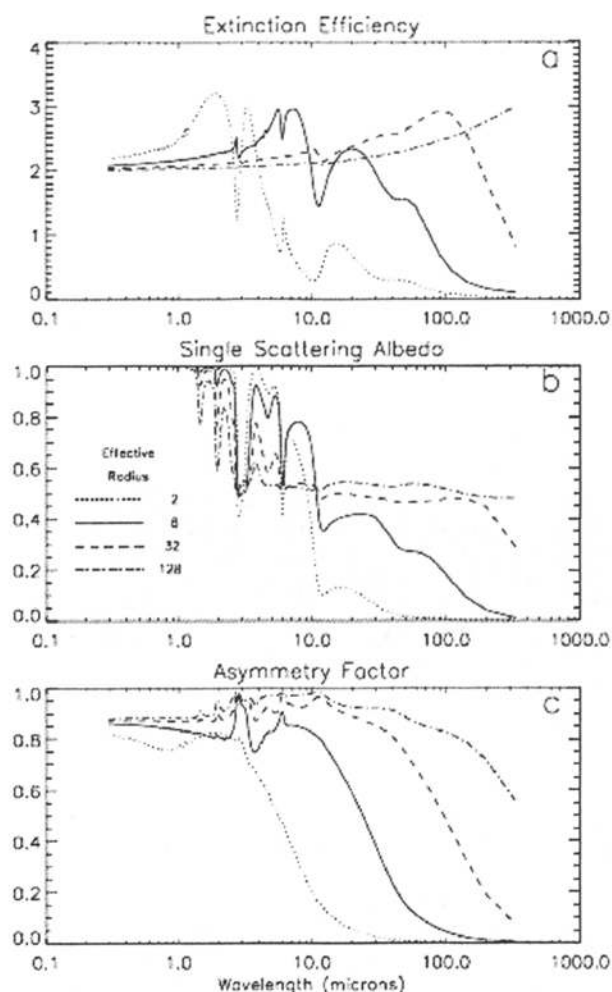


图 2.1 云滴有效粒子半径为 2、8、32 和 128 μm 时所对应的消光系数 (a)，单次散射反照率 (b) 和不对称因子 (c) (引自 Ricchiazzi et al., 1998)

关于气溶胶的计算，在 SBDART 中可以设置低层和高层两种气溶胶。对于边界层大气和高层大气，模式分别提供了 3 种标准气溶胶类型的消光系数、单次散射反射率和不对称因子。低层大气的 3 种类型是：乡村型、城市型和海洋型。高层大气的 3 种类型为：旧火山灰、新火山灰和气象沙尘。图 2.2 所示为模式中城市型气溶胶粒子在不同相对湿度条件下的光学特性。从图中可以看出城市型粒子在不同的相对湿度条件下的单次散射反照率变化很大。城市气溶胶是六种模式自带气溶胶类型中对水汽最敏感的一种。除了上述的这六种模式自带的种类，计算者也可设置其他类型的气溶胶类型，只需输入气溶胶粒子的消光系数，单次散

射反射率和不对称因子（或者是散射相函数）。对于气溶胶的垂直分布，模式中不但可以通过输入气溶胶层的高度及这层上气溶胶粒子数密度来自定义气溶胶

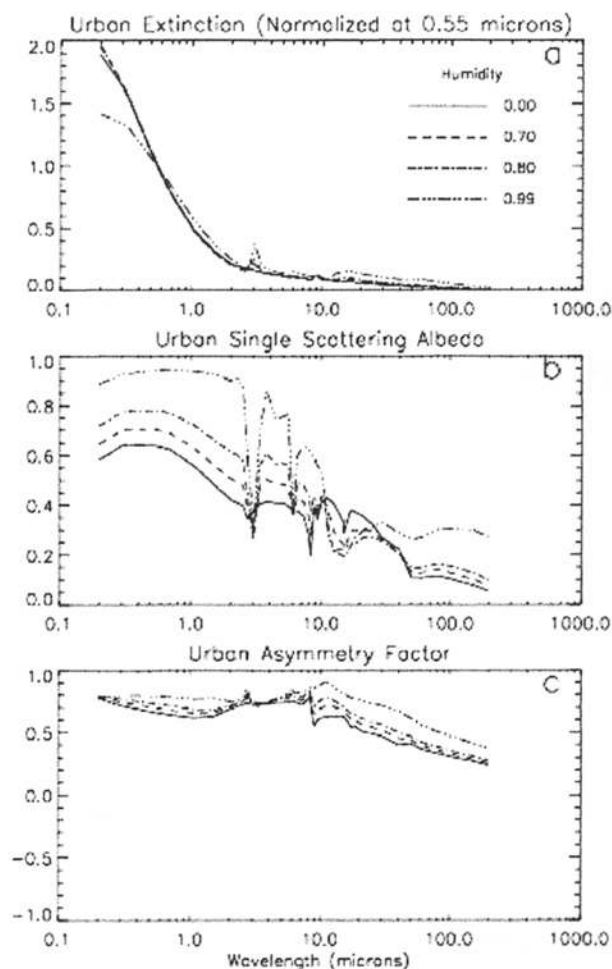


图 2.2 0.55 波段不同相对湿度下，城市型气溶胶粒子的消光系数 (a)、单次散射反照率 (b) 和不对称因子 (c) （引自 Ricchiazzi et al., 1998）

的垂直廓线，也可以使用模式中自带的两种垂直分布廓线：5km 和 23km 能见度模型 (MaClatchey et al., 1972)，如图 2.3 所示。这两种廓线分别代表了在低能见度 (5km) 和高能见度 (23km) 条件下，气溶胶粒子的垂直分布。从图中可以看出在 5km 以上，两种廓线是一样的，但是由于消光主要发生在 5km 以下，所以在 5km 以下两种廓线是不一样的，分别按照密度尺度因子 0.99 和 1.45 进行指数分布。在模式中设置不同的气溶胶垂直分布廓线需要通过设置参量“能见度

(Visibility)”来实现。“能见度”定义为 $VIS = 3.912/\delta_{ext}$, δ_{ext} 为 0.55 μm 波段地表的气溶胶消光系数, 因为气溶胶的消光系数和粒子数浓度是成比例的, 所以气溶胶光学厚度的垂直分布可以写成: $\tau_{aer}(z) = \frac{3.912}{Vis} \int_z^\infty \frac{n(0)}{n(z)} dz$, $n(z)$ 为气溶胶粒子的垂直分布, $n(0)$ 为地表气溶胶粒子数密度, 积分上限为 100km。

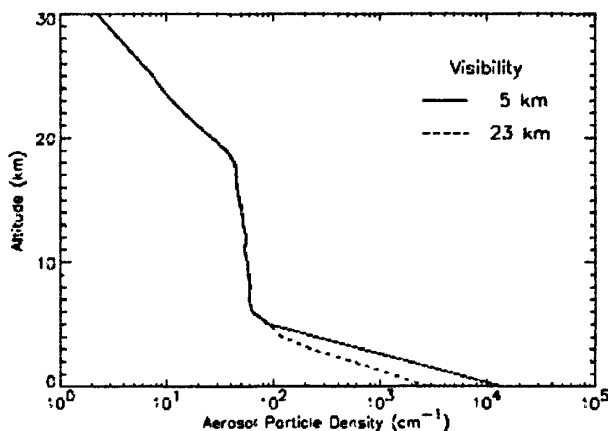


图 2.3 SBDART 模式中气溶胶的两种垂直分布廓线 (引自 Ricchiazzi et al., 1998)

2.3 小结

本章主要从模式辐射传输解法、大气廓线类型、地表类型和地表反照率输入、云滴粒子参数化、气溶胶参数化类型和垂直分布等几方面, 着重详细的介绍了论文中计算时所使用的 Fu-Liou 辐射传输模式和 SBDART 辐射传输模式。两种模式都是大气辐射研究领域所广泛使用的成熟的辐射传输模式, 各具有不同的优缺点。Fu—Liou 模式由于使用了四流/二流近似方法, 所以模式具有计算速度快的优点, 但却不能计算各角度的辐射亮度。而 SBDART 模式与其他模式相比, 具有计算波长范围长, 波段分辨率高的优点, 并且由于使用 DISORT 方法求解辐射传输方程, 所以模式也可以计算辐射亮度, 但由于波长分辨率高, 使得计算时间较长。在计算中, 可以根据不同的需要, 分别使用这两种模式。

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第三章 沙尘辐射强迫的敏感性试验

3.1 引言

利用模式计算气溶胶在 TOA 及地表的辐射强迫时, 因为目前对气溶胶粒子的分布和物理、化学性质, 地表性质, 及气溶胶和云的相互作用还不是十分清楚, 所以计算得到的气溶胶辐射强迫值有着很大的不确定性。在本节中, 为了定量了解这些关键模式输入量的不确定性对于计算辐射强迫值的影响, 对各关键影响因子做了敏感性试验。但由于云的性质及其和气溶胶的相互作用非常复杂, 而目前对此的认知水平还很低, 所以本节的气溶胶辐射强迫敏感性测试只为气溶胶直接辐射强迫, 并不包括气溶胶间接辐射强迫。

气溶胶粒子与太阳辐射的相互作用主要由以下几个因子综合决定: 1. 气溶胶粒子光学及微物理性质(光学厚度、消光系数、单次散射反照率、不对称因子、浓度、粒子分布、粒子形状等)。2. 地表性质(地表反照率)。3. 大气廓线(温度、水汽、臭氧) 4. 日地轨道(纬度、季节)。表 2.1 列出了此次敏感性测试中输入模式各参数的基准值。此处应说明, 敏感性试验中, 首先输入表中所列的各参数的基准值, 计算 TOA 和地表的向上和向下通量值, 作为辐射通量的基准值, 当进行某一个参数的敏感性测试时, 增大或减小这一参数值, 其他的输入参数值保持不变, 并将得到的结果与辐射通量基准值比较, 由此测试辐射通量结果对这一参数是否敏感, 以及这一参数是如何影响辐射通量结果的。

表中部分参数将在下文中描述其敏感性测试时做详细解释, 包括其定义和物理含义。此处需对表中其他一些固定参数做一简介。因为本文的研究主要为西北地区, 所以假设经纬度为[39.082°N, 100.276°E], 日期为 2008 年 3 月 22 日(春分), 这样便于大气廓线参数和日地距离参数的选取。大气廓线选取了中纬度地区夏季标准的大气廓线; 水汽含量则是根据 NOAA 的 NCEP 再分析资料得到的; 臭氧含量是根据 Total Ozone Mapping (TOMS) 卫星资料(Bhartia et al., 1993)得到的; 并根据地理纬度和日期计算了日平均太阳高度角为 58.68°, 代表日地距离的参数-偏心率为 1.0。

表 3.1 敏感性试验中模式输入参数的基准值

敏感性试验中模式输入参数的基准值	
大气廓线:	中纬度、夏季标准大气廓线
偏心率:	1.0
地形高度:	1.4km
水汽含量:	0.67 g/cm ²
臭氧含量:	0.315 atm-cm
地表反照率 (宽谱):	0.23
光学厚度 (500nm):	0.5
单次散射反照率 (500nm):	0.79
不对称因子 (500nm):	0.69
埃斯屈朗指数:	0.4
计算波长范围:	0.25-4.0 μm
SZA:	53°

3.2 气溶胶粒子光学特性对 TOA 及地面辐射通量影响的敏感性试验

气溶胶粒子的光学特性经常用以下三个变量来定义: 气溶胶光学厚度 (τ), 单次散射反照率 (ω), 散射相函数。

3.2.1 气溶胶光学厚度敏感性试验

气溶胶光学厚度是单位面积的大气柱内的气溶胶消光截面的总和, 用来表示由于气溶胶粒子对太阳光的吸收和散射而造成的消光。通常沙尘的光学厚度范围为 0-1.0 (500nm), 但发生强沙尘暴时在 500nm 波段的光学厚度也会超过 3。

(McFarlane et al., 2009)。在敏感性试验中, 我们取 $\tau(500\text{nm})=0.5$ 为基准值计算辐射强迫值, 然后将光学厚度分别增加 20% ($\tau(500\text{nm})=0.6$), 60% ($\tau(500\text{nm})=0.8$), 以及减少 20% ($\tau(500\text{nm})=0.4$), 40% ($\tau(500\text{nm})=0.2$), 计算不同光学厚度下的辐射通量值, 与基准值进行比较, 即可看出气溶胶光学厚度对 TOA 和地表的辐射通量产生了怎样的影响。

表 3.2 给出了当沙尘光学厚度变化时, TOA 和地表辐射通量的变化值和百分

比。从表中可以看出,当气溶胶光学厚度增大时,地表向上及向下的辐射通量都有显著的减小,当光学厚度增大 20%时,由于气溶胶的消光作用,地表向下的直接辐射通量减小 33.92 w/m^2 ,同时由于光学厚度增加使得散射辐射增大 11.97 w/m^2 ,最终造成地表向下总的辐射通量减少 21.95 w/m^2 。由此可以看出气溶胶光学厚度的增加会减少到达地面的太阳辐射通量。另外从表中数据可以看出,TOA 处向上的辐射通量对气溶胶光学厚度的变化并不十分敏感, τ 增加或减少 20-60%,其变化的百分比都在 2%内。这是由于在进行气溶胶光学厚度对辐射通量影响的敏感性试验中,所取的 500nm 波段上的气溶胶单次散射反照率为 0.79,此值的选取使得 TOA 处辐射通量值对气溶胶光学厚度的变化并不敏感。所以在此处应该注意,因为地表的反照率和气溶胶单次散射反照率分别决定着:由地面反射到 TOA 处的通量和由气溶胶散射辐射造成的 TOA 向上辐射通量,所以选取的这两个参数基准值的不同会使结果发生变化。

表 3.2 光学厚度增大/减小 20%和 60%后 TOA 和地表的辐射通量变化
(单位: w/m^2)

τ	TOA_up	Suface_up	Suface_down	Suface_down_dir	Suface_down_dif
Increase 20%	0.49 (0.3%)	-4.50 (-4.6%)	-21.95 (-5.1%)	-33.92 (-14.7%)	11.97 (6.0%)
Increase 60%	1.07 (0.7%)	-12.83 (-12.8%)	-61.98 (-14.3%)	-87.31 (-37.7%)	25.33 (12.6%)
Decrease 20%	-0.66 (-0.4%)	4.60 (4.6%)	22.88 (5.3%)	39.22 (17.0%)	-16.34 (-8.1%)
Decrease 60%	-2.80 (-1.7%)	14.75 (14.8%)	73.67 (17.1%)	142.59 (61.6%)	-68.92 (-34.3%)

3.2.2 单次散射反照率的敏感性试验

单次散射反照率(SSA, ω)计算定义为:气溶胶粒子散射系数与消光系数的比值;这是决定气溶胶粒子吸收、散射的一个重要参数。在试验中选取的气溶胶 500nm 的 SSA 基准值为 0.79。这个基准值的选择是根据,由 2008 年 4 月-6 月中美联合观测期间 MFRSR 数据反演得出的 11 个晴空例子气溶胶粒子 SSA 的平均值 (Ge et al., 2009)。

表 3.3 给出了当 SSA 增大或减小时,TOA 和地表辐射通量的变化。从表中所列得到,当 SSA 增大 20%,为 0.948 时,TOA 处向上的辐射通量增大 40.68 w/m^2 ,

即比当 SSA 为 0.79 时, 多了 25.1% 的能量散射出大气层; 当 SSA 减少 20% 时, 为 0.632 时, 粒子吸收性增强, 则散射出大气层的能量减少了 21.6%。由此看出, 气溶胶粒子的单次散射反照率对 TOA 的能量收支有着显著的影响。Ramanathan 等人的结果表明 (Ramanathan et al., 2001), 当 $SSA > 0.95$ 时, TOA 处为负的辐射强迫; 当 $SSA < 0.85$ 时, TOA 处为正的辐射强迫; 而当 SSA 在 0.95 到 0.85 之间时, TOA 处的辐射强迫值可正可负, 正负的变化取决于地表反照率, 气溶胶垂直分布等其他的因子的影响。对于地表来说, 因为 SSA 只与散射辐射有关, 与直接辐射无关, 所以当 SSA 改变时地表向下的直接辐射通量并没有变化, 散射辐射通量随着 SSA 的增大 (减小) 而增大 (减小)。而地表向上的辐射通量, 由于因为地表反照率不变, 所以只与向下通量有关, 与其变化一致。

表 3.3 SSA 增大/减小 10% 和 20% 后 TOA 和地表的辐射通量变化
(单位: W/m^2)

ω	Up_TOA	Up_Suf	Down_Suf	Down_Suf_dir	Down_Suf_dif
Increase 10%	21.69 (13.4%)	6.38 (6.4%)	29.44 (6.8%)	0.0 (0.0%)	29.44 (6.8%)
Increase 20%	40.68 (25.1%)	11.24 (11.3%)	55.54 (12.9%)	0.0 (0.0%)	55.54 (12.9%)
Decrease 10%	-18.9 (-11.7%)	-5.8 (-5.8%)	-26.66 (-6.2%)	0.0 (0.0%)	-26.66 (-6.2%)
Decrease 20%	-34.9 (-21.6%)	-10.81 (-10.8%)	-50.24 (-11.6%)	0.0 (0.0%)	-50.24 (-11.6%)

3.2.3 不对称因子的敏感性试验

散射辐射随角度的分布被定义为散射相函数(Phase Function)。理论上, 如果给定了球形气溶胶粒子的分布, 则可以利用 Mie 理论计算得到气溶胶粒子的散射相函数 (Van de Hulst, 1981; Bohren and Huffman, 1983)。但在实际计算中, 在大多数辐射传输模式中都近似用 Henyey-Greenstein (HG) 的散射相函数 (Henyey and Greenstein, 1941) 来代替精确的散射相函数。HG 散射相函数较之精确的散射相函数较为简单, 是因为 HG 函数中只用一个参数来定义各个方向的散射能量分布, 即一不对称因子 g (Asymmetry factor)。公式 (3.1) 即为用 g 计算散射相函数的表达式。

$$P_{HG}(\cos\theta) = \frac{1-g^2}{(1+g^2-2g\cos\theta)^{3/2}} \quad (3.1)$$

$(1+g)/2$ 代表了前向能量的分布。在图 3.1 中, 当横坐标 $\theta=0^\circ$ 时, 代表前向; $\theta=\pm 180^\circ$ 时, 代表后向。从图中可以看出, 当 g 增大时, 进入前向的能量增加, 当 g 等于 1 时, 则完全为前向散射; 当 $g=0$ 时则散射成对称分布, 如瑞利散射。一般 g 的取值范围在 0.5 到 0.8 之间。在本次试验中, g 在 500nm 波段的基准值为 0.69。与 SSA 的取值一样, 这也是根据由 2008 年 4 月-6 月中美联合观测期间 MFRSR 数据反演得出的 11 个晴空例子气溶胶粒子 g 的平均值得到的。

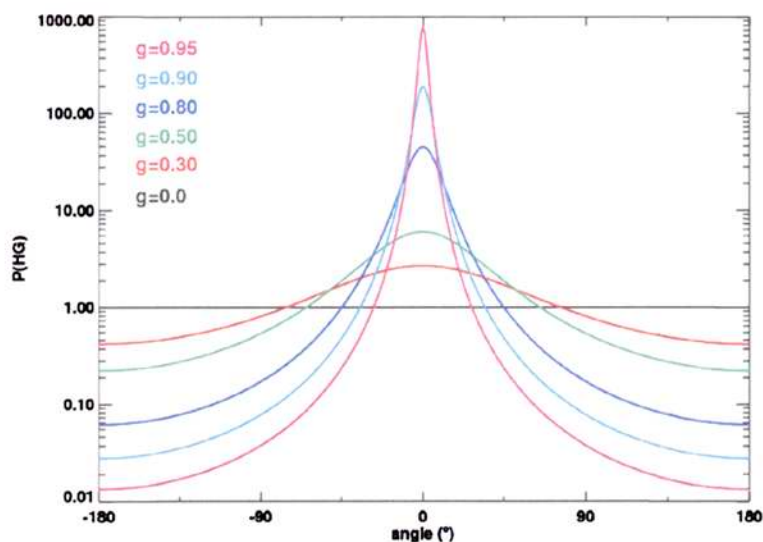


图 3.1 不同 g 值计算得到的 HG 相函数随角度分布的变化图

从表 3.4 所列数据可以看出不对称因子对 TOA 和地表辐射通量的影响并不十分显著, g 值增大, 前向散射增大, 所以地表向下辐射增加, 同时 TOA 处向上辐射减小。

表 3.4 g 增大/减小 10%和 20%后 TOA 和地表的辐射通量变化
(单位: w/m^2)

g	Up_TOA	Up_Suf	Down_Suf	Down_Suf_dir	Down_Suf_dif
Increase 10%	-7.37 (-4.6%)	1.9 (1.9%)	8.91 (2.1%)	0.0 (0.0%)	8.91 (2.1%)
Increase 20%	-14.52 (-9.0%)	3.76 (3.8%)	17.97 (4.2%)	0.0 (0.0%)	17.97 (4.2%)
Decrease 10%	6.87 (4.2%)	-1.69 (-1.7%)	-7.93 (-1.8%)	0.0 (0.0%)	-7.93 (-1.8%)
Decrease 20%	10.86 (6.7%)	-2.47 (-2.5%)	-12.31 (-2.8%)	0.0 (0.0%)	-12.31 (-2.8%)

3. 2. 4 埃斯屈朗指数的敏感性试验

气溶胶光学特性参数（光学厚度、单次散射反照率、不对称因子）都会随着辐射波长而变化。气溶胶光学厚度随波长的变化则由一个特定参数来表示，即埃斯屈朗指数（Angstrom exponent; Angstrom, 1929, 1930）。埃斯屈朗指数值较大时代表气溶胶粒子为小粒子；当埃斯屈朗指数值较小时代表粒子为大粒子。在试验变化 A 时，500nm 波长的 τ 值不变，其他各波长上的 τ 值将根据 A 的变化而变化。从表 2.5 中可以看出，A 值的变化对 TOA 处向上的通量影响很小。当 A 增大 75%值为 0.7 时，地表向下的直接辐射增大了 9.6%，同时散射辐射减小了 6.1%，这是由于太阳能量是按波谱分布的，A 值变大使得能量分布较大波长上的光学厚度减小，使得到达地面的直接辐射增加，而同时散射辐射减小。

表 3.5 Angstrom exponent 增大/减小 25%和 75%后 TOA 和地表的辐射通量变化
(单位: w/m^2)

A	Up_TOA	Up_Suf	Down_Suf	Down_Suf_dir	Down_Suf_dif
Increase 25%	-0.19 (-0.1%)	1.16 (1.2%)	3.74 (0.9%)	7.74 (3.3%)	-4.0 (-2.0%)
Increase 75%	-0.55 (-0.3%)	3.32 (3.3%)	9.86 (2.3%)	22.17 (9.6%)	-12.31 (-6.1%)
Decrease 25%	0.24 (0.1%)	-1.26 (-1.3%)	-3.89 (-0.9%)	-15.74 (-6.8%)	4.11 (2.0%)
Decrease 75%	0.77 (0.5%)	-4.17 (-4.2%)	-12.73 (-3.0%)	-25.11 (10.9%)	12.38 (6.2%)

3.3 地表性质对 TOA 及地面辐射通量影响的敏感性试验

在研究气溶胶对太阳能量收支平衡的影响过程中，地表性质的准确描述对结果至关重要（Atwater,1970; Mitchell Jr., 1971; Coakley et al., 1983）。地表对气溶胶

影响辐射系统的作用很复杂,这是因为地表与气溶胶之间的多次反射,会造成地表反照性质对气溶胶辐射的非线性影响。一般来说,对于同样的非吸收性气溶胶,当地表的反照较强时(例如沙漠或者冰雪),气溶胶的辐射作用较弱。但是,对于吸收性气溶胶则完全不一样,即使吸收性很弱的气溶胶粒子,在地表反照较强的情况下,也会在 TOA 处产生加热作用(辐射强迫为正值)。由此来看,如果对地表反照性质认识不清,会对气溶胶辐射强迫的估算造成另外的不确定性。

尽管大家已经认识到对地表性质的准确描述很重要,但是如何来准确描述地表的反照性质一直是个难点。由自然地表反射的太阳辐射是各向不均一的,通常用双向反射分布函数(BRDF, bidirectional reflectance distribution function)来描述地表反射的太阳辐射随角度在上半球的分布。BRDF 函数代表的是地表的固有性质,它是由组成地表的材料成分,质地和结构所决定的。但在辐射传输模式中,因为 BRDF 的复杂性,通常将地表对辐射的反照性质简化用一个较简单的参数来表示,即地表反照率(surface albedo)。地表反照率为反射光与入射光的比值,它是所有组成的 BRDF 在上半球的积分。地表反照率除了是由地表的性质决定的以外,它还与波长和入射光的角度有关。辐射传输模式中,海洋表面的地表反照率除了要考虑波长与入射角度影响外,还与风速和叶绿素的浓度有关。但陆地表面与海洋表面相比,要更为复杂,更不均一,以及随波长变化更大,所以陆地的表面性质则更难描述。MODIS 和 MISR 卫星由于在多角度和多波段进行测量,所以可以较好的描述陆地地表性质。图 3.2 为 MODIS 卫星数据反演的在全球亮背景(white-sky)和暗背景(Black-sky)下不同地表类型的反照率随太阳高度角(SZA)的变化。在 White-sky 和 Black-sky 下的反演是两种不同的反演方法,分别代表的是由反射的散射光造成的双半球反射率和由直接光造成的半球反照率。从图 3.2 中可以看出,对于不同的地表类型,在太阳高度角较小时(太阳位置较高)White-sky 的反照率大于 Black-sky 下的反照率,而当 SZA 大于 55°时,则反之。

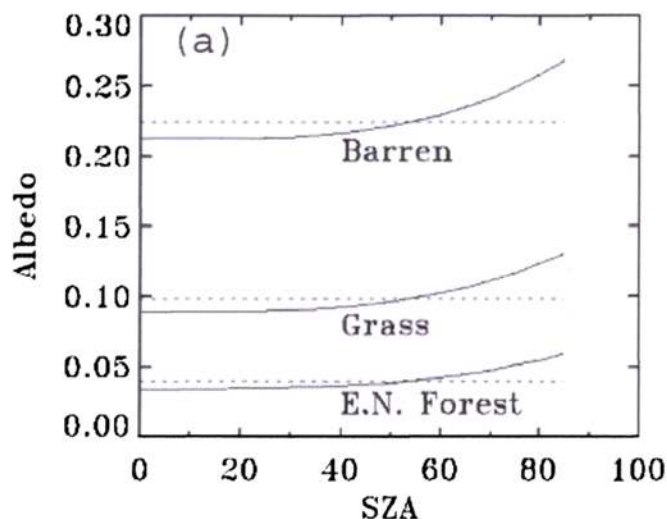


图 3.2 MODIS 反演的在 white-sky 和 dark-sky 下可见光波段荒漠、草地和森林的地表反照率随 SZA 的变化 (引自 Lewis P., 1994)

另外, 地表反照率还会随波长而变化。图 3.3 是不同地表类型下, 地表反照率随波长的变化。由图看出, 地表反照率随波长有很大的变化, 并且不同地表类型随波长变化的形状也不同, 所以在定义地表反照率时, 只给定宽谱的反照率是不恰当的, 应该给出波谱地表反照率。

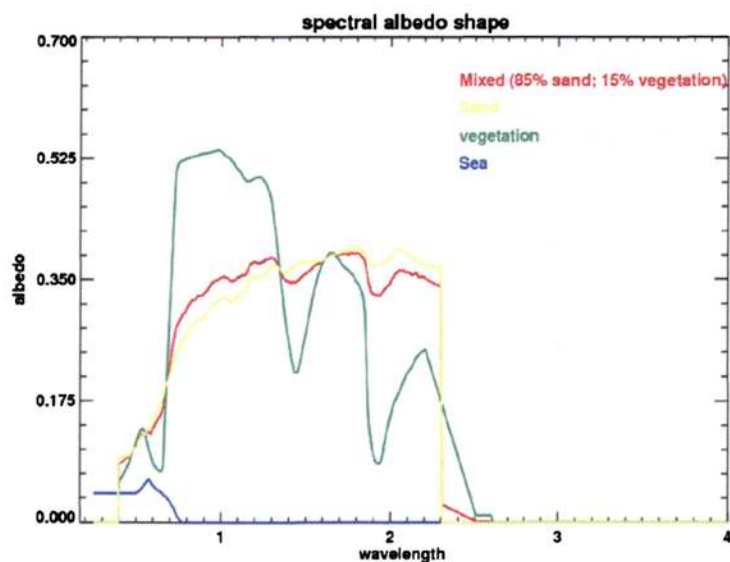


图 3.3 不同地表类型地表反照率随波长的变化。图中蓝线代表海洋，绿线代表农田，黄色代表沙地，红色代表混合地表（85%为沙地，15%为农田）

表 3.5 给出了在变化不同地表类型后，TOA 和地表辐射通量的变化。因为研究的主要地域为西北地区，所以本次试验地表的基本类型是沙地与农田的混合型地表，其宽谱的地表反照率为 0.23，随波长变化的形状为图 2.3 中所示的混合型

表 3.6 不同地表类型下 TOA 和地表的辐射通量变化
(单位: w/m^2)

Suface-Albedo	Up_TOA	Up_Suf	Down_Suf	Down_Suf_dir	Down_Suf_dif
Ocean_0.04	-67.1 (-41.4%)	-89.88 (90.0%)	-8.15 (-1.9%)	0.0 (0.0%)	-8.15 (-1.9%)
Vegetation_0.14	-27.4 (-17.0%)	-37.75(37.8%)	-3.69 (-0.9%)	0.0 (0.0%)	-3.69 (-0.9%)
Sand_0.35	196.1 (121.%)	289.9 (290.3%)	40.5 (9.4%)	0.0 (0.0%)	40.5 (9.4%)

（红色实线表示）波谱分布。表 3.5 中做敏感性测试的其他几种类型为：海洋-宽谱地表反照率为 0.04；农田-宽谱地表反照率为 0.14；沙地-宽谱地表反照率为 0.35。从表中可以看出地表反照率对 TOA 和地表向上的辐射通量起着决定性的作用。当将宽谱地表反照率从 0.23 增大到 0.35，波谱形状从混合变为沙地时（参看图 3.3），TOA 处向上的辐射通量可以增加 121%，地表向上的辐射通量增加 290%，并且由于地面向上通量的增加使得反射辐射与气溶胶的多次散射增加，所以地面向下的辐射通量由于散射辐射的增强也增大了 9.5%，所以在估算气溶胶的辐射强迫时，地表反照率是一个非常重要的因子。

3.4 水汽对 TOA 及地面辐射通量影响的敏感性试验

水汽作为大气成分中的可变因子，其体积比含量为 0-0.04%，并随时空分布变化很大。虽然水汽含量很小，但却是地球大气中主要的辐射和动力因素。它与二氧化氮等温室一样，吸收地球发射的长波辐射，加热地表及低层大气。尽管水汽是一种强烈的温室气体，但其在大气中的浓度随着对地表的气候变化的相应而变化，因此被认为是一种反馈效应，而不是辐射强迫（IPCC，2007）。

图 3.4 标出了，水汽的几个强吸收带。2.74 μm 和 2.66 μm 这两个吸收带彼此很接近，在太阳光谱中组合成一个强吸收带，即为 2.7 μm 带。在近红外谱区，这

些吸收带吸收低层大气中相当大一部分太阳辐射通量，它们是中心分别在 0.94, 1.1, 1.38, 1.87 μm 吸收带。另外，中心在 0.72 μm 和 0.82 μm 的吸收带，虽然强度较弱，但它们对太阳加热大气的贡献也并非不重要。在可见光区，还有一些水汽吸收谱线，不过它们对吸收和加热速率的贡献似乎很小。

表 3.7 给出了当水汽密度增大及减少 20%、40% 时，辐射通量的变化。大气层水汽密度的基准值为 $0.67\text{g}/\text{cm}^3$ 。这是根据 NOAA reanalysis 资料分析得到的，西北张掖地区春季 4-5 月水汽密度的平均值。从表中可以看出，当水汽增大 40%，大气层水汽密度为 0.94 时，到达地面的总的辐射通量减少 $6.29\text{w}/\text{m}^2$ ，其中由于水汽对直接光的吸收，使得到达地面的直接辐射减少 $2.46\text{w}/\text{m}^2$ ；另外部分散射光也被水汽吸收，所以到达地面的散射辐射也减少了 $3.83\text{w}/\text{m}^2$ 。

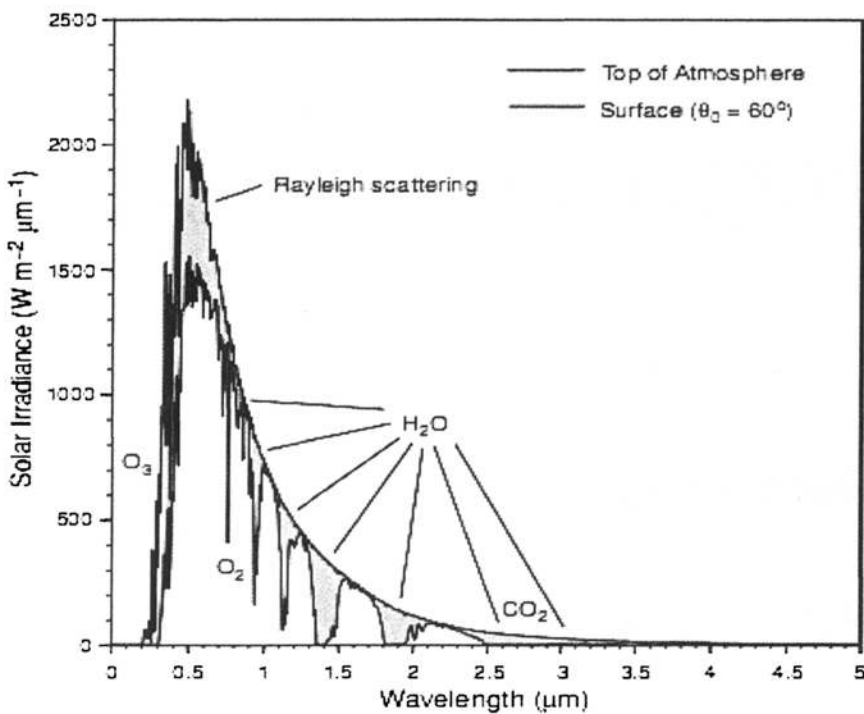


图 3.4 在大气顶处测得的太阳辐照度曲线，以及在无气溶胶或无云的大气中，当太阳高度角为 60° 时，在地面测得的太阳辐射度曲线。图中标出了吸收区和散射区。（引自 廖国男，《大气辐射导论》）

表 3.7: 水汽含量增大/减小 20%和 40%后 TOA 和地表的辐射通量变化
(单位: w/m^2)

	Up_TOA	Up_Suf	Down_Suf	Down_Suf_dir	Down_Suf_dif
Increase 20%	-0.88 (-0.5%)	-0.18 (-0.2%)	-1.45 (-0.3%)	-0.66 (-0.3%)	-0.79 (-1.0%)
Increase 40%	-1.7 (-1.0%)	-0.36 (-0.4%)	-2.86 (-0.7%)	-1.3 (-0.6%)	-1.56 (-0.8%)
Decrease 20%	0.94 (0.6%)	0.18 (0.2%)	1.51 (0.3%)	0.66 (0.3%)	0.85 (0.4%)
Decrease 40%	1.97 (1.2%)	0.36 (0.4%)	3.09 (0.7%)	1.34 (0.6%)	1.75 (0.9%)

3.5 臭氧对 TOA 及地面辐射通量影响的敏感性试验

臭氧是大气中的一中自然微量成分，在大气中的体积比为 $0-12 \times 10^{-4}$ ，并随季节和纬度而变化。臭氧主要出现在约 15-30km 的高度范围内，即称之为臭氧层。从图 3.4 中我们可以看出，在紫外波段，波长小于 $0.32\mu\text{m}$ 的 99% 太阳辐射都被臭氧吸收了，而这一波长范围内的高强度太阳紫外辐射，对几乎所有生物都有伤害，所以臭氧层吸收致命的太阳紫外辐射，这对地球上的生命至关重要。但目前由于人类活动使臭氧耗减，以成为一个重要的科学研究课题。

表 3.8 为改变臭氧浓度，TOA 和地表辐射通量的变化。本次试验臭氧浓度的基准值为 0.315atm-cm 。这是由装载于雨云 7 号卫星上的臭氧总量成像光谱仪 (TOMS) 资料得到的，张掖地区春季 (4-5 月) 臭氧浓度的平均值。从表中看当臭氧浓度减少 40%，为 0.189atm-cm 时，地表向下的辐射通量变化了 0.7%，增大了 3.09 w/m^2 。此处应该注意由于臭氧吸收主要是对生物有伤害的紫外辐射，所以到达地表的辐射通量增加的 0.7%，多为短波紫外波段，会对地表生物产生较大伤害。

表 3.8: 臭氧含量增大/减小 20%和 40%后 TOA 和地表的辐射通量变化
(单位: w/m^2)

	Up_TOA	Up_Suf	Down_Suf	Down_Suf_dir	Down_Suf_dif
Increase 20%	-1.12 (-0.7%)	-1 (-1%)	-3.24 (-0.7%)	-1.98 (-0.9%)	-1.26 (-0.6%)
Increase 40%	-2.18 (-1.3%)	-1.94 (-1.9%)	-6.29 (-1.5%)	-3.83 (-1.7%)	-2.46 (-1.2%)
Decrease 20%	1.32 (0.8%)	1.17 (1.2%)	3.78 (0.9%)	2.30 (1.0%)	1.48 (0.7%)
Decrease 40%	3.06 (1.9%)	2.71 (2.7%)	8.77 (2.0%)	5.35 (2.3%)	3.42 (1.7%)

3.6 本章小结

在本章中我们对气溶胶光学特性、下垫面性质以及大气水汽和臭氧含量对气溶胶辐射强迫计算的影响做了敏感性试验。在气溶胶光学性质参数中,分别对光学厚度、单次散射反照率以及不对称因子做了测试。结果表明气溶胶光学厚度对地表辐射通量的影响要大于对 TOA 处辐射通量的影响。当气溶胶增减 20%时,地表处向上直接辐射通量值可减少和增加 14.7%和 37.7%。气溶胶粒子单次散射反照率作为表示气溶胶粒子吸收散射特性的一个参数,其变化则是对 TOA 处通量值有显著影响,当 SSA 增大减少 20%时,对 TOA 处向上的辐射通量的影响可达到 25%左右。而不对称因子 g 主要代表前向散射的大小,所以其对 TOA 和地表处的通量值的影响都并不是很大,当其值变化 20%时,对各处的影响都小于 10%,相比较而言,对 TOA 处通量的影响要大于对地表处的影响。地表反照率作为模式计算辐射强迫时的一个重要因子,对 TOA 处的辐射通量有着非常大的影响,从敏感性计算来看当我们把地表类型从地表反照率为 0.24 的沙田混合型变为地表反照率为 0.04 的海洋型、0.14 的农田型和 0.35 的沙漠型,TOA 处向上的辐射通量可以分别变化 44.1%, 17%, 121%。由此看出地表反照率的设定对与 TOA 处辐射强迫计算的结果至关重要。我们还对大气中水汽含量和臭氧含量对辐射通量的影响做了测试,结果表明这两个参数对 TOA 和地表处的短波辐射通量影响并不大。由此看来,气溶胶粒子的光学特性会对辐射强迫的计算有很大影响,所以得到正确的光学特性参数就显得尤为重要,野外观测试验作为一种有效的方法,发挥着越来越重要的作用。

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第四章 沙尘气溶胶的直接辐射强迫效应

4.1 引言

尽管气溶胶通过散射和吸收太阳光而产生的直接辐射强迫效应已经被认知了很长时间,但是由于缺乏气溶胶粒子的光学特性和地表反照率等重要参数,特别是沙尘这种具有强吸收性的非球形型粒子,使得量化这个影响值的过程进展缓慢,目前对于沙尘气溶胶直接辐射强迫效应的研究还有很大的不确定性。本章充分利用了 CERES、CALIPSO 卫星数据,以及 2008 年中美联合观测期间的地面观测数据,详细研究计算了西北地区沙尘气溶胶直接辐射强迫效应。

4.2 塔克拉玛干沙漠地区沙尘辐射强迫值

4.2.1 引言

沙尘气溶胶是影响地气能量平衡的重要因子之一,它不仅可以直接的吸收和散射太阳辐射能量,也可以间接的通过改变云粒子分布和浓度来改变地气辐射能量(Twomey, 1977; Albrecht, 1989)。而且作为一种吸收性气溶胶,沙尘还可以对大气干绝热率产生很大影响,从而影响大气的水分循环(Ramnasa et al., 2004; Ackerman et al., 2000; Koren et al., 2004; Huang et al., 2006)。最近, Satheesh 等人研究发现在亚非地区由于沙尘粒子的吸收性,可以减少达到地表的辐射能量而使低层大气增温 0.3-0.5K/day。另外由于沙尘粒子半径通常较大,使得沙尘气溶胶还可以通过对长波辐射的影响而产生显著的温室效应(Zhang and Christopher, 2003)。

在计算沙尘气溶胶的辐射作用时,其在大气中的垂直分布是一个重要的影响参数,特别是在计算大气垂直加热率时,Carlson 和 Benjamin 模式结果表明改变气溶胶在大气各层的分布,会对大气加热率产生非常显著的影响。另外 Meloni 等人研究发现气溶胶在 TOA 处辐射的短波辐射强迫值与气溶胶的垂直分布廓线有紧密的联系,而且有其他结果指出晴空条件下 TOA 处的长波辐射强迫值也对沙尘的分布高度很敏感(Liao and Seinfeld, 1998)。2006 年发射的 CALIPSO (Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation) 卫星搭载了主

动式雷达观测仪器以及被动的红外和可见光成像仪,主要用于探测区域和全球尺度的气溶胶垂直分布廓线,这为我们提供了一个前所未有的机会研究沙尘气溶胶的辐射作用。与其他空间被动式遥感探测仪器不同,CALIPSO 可以有效探测亮地表上空的气溶胶甚至可以穿过薄的卷云探测到云下的气溶胶(Winker et al., 2006; Hu et al., 2006, 2007; Liu et al., 2004, 2008; Huang et al., 2007, 2008)。

本章将研究 2006 年发生在塔克拉玛干沙漠的一次沙尘暴过程中的沙尘的直接辐射强迫效应以及大气的垂直加热率。塔克拉玛干沙漠是大气悬浮沙尘粒子的重要的生成源,沙尘粒子被风卷入大气并向下风方向传输,使其不仅可以影响东亚地区,也可以影响到北太平洋地区,甚至北美地区(Haywood et al., 1999; Higurashi&Nakjima, 2002; Huang et al., 2008)。Fu-Liou 辐射传输模式将用于计算辐射强迫值与大气加热率(Fu and Liou, 1992, 1993)。模式计算中输入的沙尘垂直分布廓线是从 CALIPSO 卫星资料得到的。我们利用 CERES (Clouds and Earth's Radiant Energy System) SSF (Single Satellite Footprint) 卫星资料中 TOA 处的辐射通量来限定沙尘粒子的散射性质参数,从而确定模式中的沙尘气溶胶种类。本章将辐射传输模式结合 CALIPSO 和 CERES 卫星观测资料数据,以期能够得到可靠的沙尘辐射强迫和大气加热率的估算值。

4.2.2 塔克拉玛干沙漠地区夏季沙尘

塔克拉玛干沙漠是世界最大的纯沙沙漠之一,中国境内最大的沙漠,也是世界上最大的流动沙漠,位于亚洲中部,中国新疆维吾尔自治区境内。塔克拉玛干沙漠位于塔里木盆地中央,四周被高山包围,平均海拔 1km,沙漠东西长约 1000 公里,南北宽 400 公里,总面积达 330 000km²。沙漠系暖温带干旱沙漠,是典型大陆性气候,风沙强烈,温度变化大,全年将水少。全年有三分之一是风沙日,大风风速每秒高达 30 米。由于整个沙漠受西北和南北两个盛行风向的交叉影响,风沙活动十分频繁而剧烈,流动沙丘占 80%以上。据测算低矮的沙丘每年可移动约 20 米,近一千年,整个沙漠向南延伸了约 100 公里。

根据长期沙尘传输观测记录,沙尘暴事件多发于冬季后期以及春季,所以目前大部分研究工作都主要关注的晚冬初春的沙尘暴事件(Iwasaka et al., 1983; Zhang et al., 1997; Murayama et al., 2001; Uno et al., 2001; Sun et al., 2001, Wang et

al., 2005)。而对发生于夏秋两季的塔克拉玛干沙漠的沙尘暴事件, 几乎没有工作对其研究分析过。但是最新 CALIPSO 观测资料表明在塔克拉玛干地区, 沙尘事件的发生不仅只是在晚冬初春, 而是贯穿整年 (Liu et al., 2008a), 并且在夏季还会经常发生强沙尘暴事件 (Huang et al., 2007)。塔克拉玛干沙漠沙尘暴对辐射能量平衡的作用以及对区域气候的影响至今还没有得到一致的研究结论, 在本章中我们将量化分析塔克拉玛干地区 2006 年 7 月 26-31 日一次沙尘暴过程中, 沙尘气溶胶的直接辐射强迫效应及大气垂直加热率。

这次沙尘暴过程, 开始于 2006 年 7 月 26 日: 由于西伯利亚较强的由西北延伸到东南的低压槽系统, 在北纬 35° - 45° 的塔克拉玛干地区产生很大的气压梯度并形成每秒超过 20m 的强烈的西北风。7 月 26 日在这种天气背景下, 在塔克拉玛干北部, 伴随者局地一些小的沙尘龙卷, 这一次沙尘暴过程开始发展, 在有利的天气系统条件下, 迅速的向南方发展。此次沙尘暴过程从 26 日开始到 31 日缓慢结束, 历时共 6 天, 其中 27-29 日达到高峰。

4.2.3 沙尘粒子消光系数的垂直分布

CALIPSO 是颗太阳同步卫星, 每日在地方标准时 13: 30 (LST) 通过赤道, 白天和夜晚都近垂直观测 532nm 和 1064nm 波段上的总后向散射系数垂直廓线。另外, CALIPSO 除了观测两个波段的总的后向散射系数, 还提供在 532nm 波段上线性去极化的后向散射系数廓线。使用这种去极化的测量方法, 不仅可以辨别冰云和水云 (Hu et al.), 还可以识别非球形气溶胶粒子。目前已经公布的 CALIPSO level1B 数据主要包括: 经过校准的 532nm 和 1064nm 波段上半轨道(半天或是晚上) 高分辨率的后向散射消光垂直廓线和 532nm 波段的去极化比值廓线。其中其垂直分辨率在 8.2km 为 30m, 8.2km 以上为 60m。本章中我们使用了 CALIPSO level1B 数据来计算气溶胶消光系数的垂直廓线。

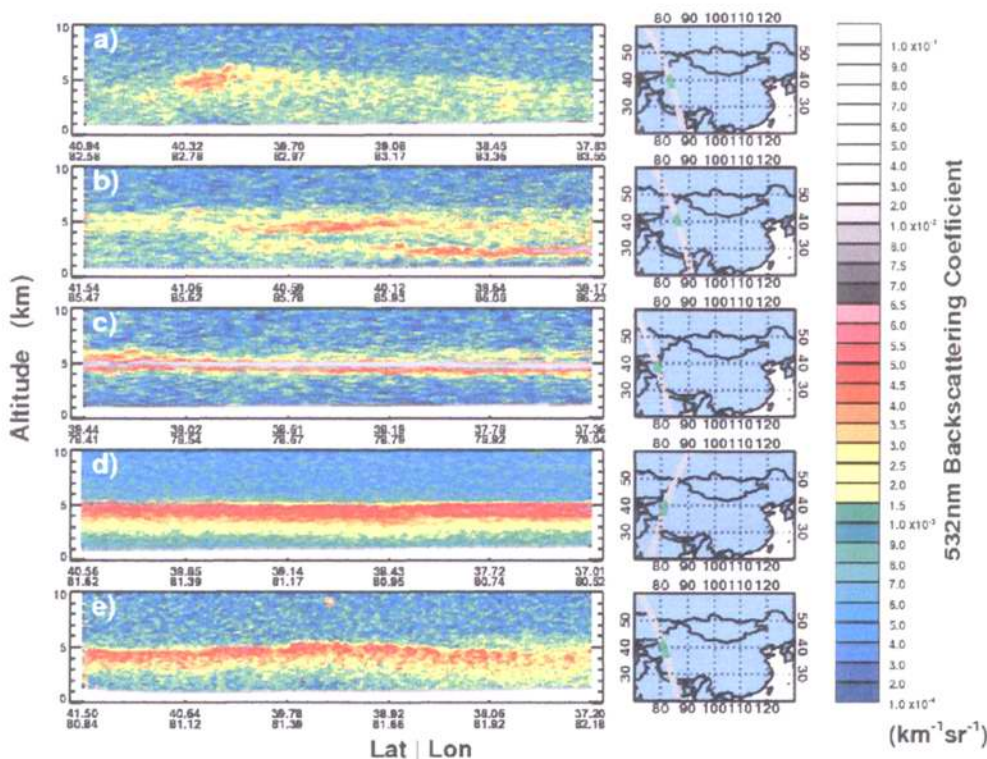


图 4.1 2006 年 7 月 24 (a 图)、26 (b 图)、29 (c 图)、30 (d 图)、31 (e 图) 日塔克拉玛干沙漠地区 (35°N - 45°N) 轨道高度上后向散射消光强度的垂直剖面图 (左图) 和卫星运行轨道图 (右图)

图 (4.1) 左图所示为 2006 年 7 月 24、26、29、30、31 日, 塔克拉玛干地区 CALIPSO 卫星探测得到的后向散射消光强度的垂直剖面图, 右图为与之相对应的卫星轨道运行位置图。其中 24、26、29、31 为卫星白天探测数据, 30 日为夜晚探测数据, 因为夜间没有太阳光的散射, 所以从图 (4.1d) 可以看出夜间卫星探测得到的信号噪音较小。图中的色彩条是 CALIPSO 数据分析所用的色彩条, 其中绿色、黄色和橘红色代表气溶胶, 从图中可以看出 24 日到 31 日的这次沙尘过程, 沙尘分布整个塔里木盆地; 橘红色代表的强沙尘层主要分布在 2.5km 到 5.5km 之间。但是 29 日的图中显示 5km 的高度上并不是通常代表强沙尘的橘红或是红色, 而是通常代表云的灰白色, 这是因为当沙尘光学厚度非常大时, 沙尘的一些光学性质与云的很相似, 比如图中所示的后向散射强度值以及色比值 (即

1064nm 与 532nm 波段上后向消光的比值), 目前的 CALISPO 产品中会将强沙尘误认为是云而显示为灰白色, 所以单凭图 4.1 我们无法确认 29 日 5km 上空是强沙尘层还是云层。

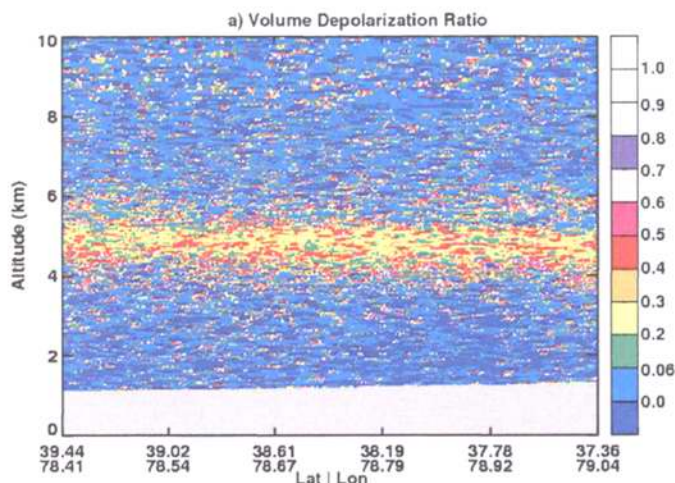


图 4.2 塔克拉玛干地区 (35°N - 45°N) 7 月 29 日轨道高度去极化比值垂直剖面图

为了区分确认是云还是沙尘, 我们利用去极化比值来做进一步认定。去极化比是指 532nm 波段上雷达接收的垂直方向信号与水平方向型号的比值, 通常水云和其他球形粒子气溶胶由于粒子形状为球形, 所以去极化比值接近与零, 而沙尘气溶胶由于其粒子的非球形性使其去极化比值较大。图 4.2 所示为 2006 年 7 月 29 日塔克拉玛干沙漠的去极化比值的垂直剖面图。从图中可以看出 5km 处去极化比值较大, 在 0.2-0.5 之间, 所以排除了是云层的可能性, 所以 7 月 29 日 km 处应该是强沙尘层。另外图 4.3 所示的 Aqua 卫星的 MODIS 真彩合成卫星图片也进一步证实了应该是强沙尘层。

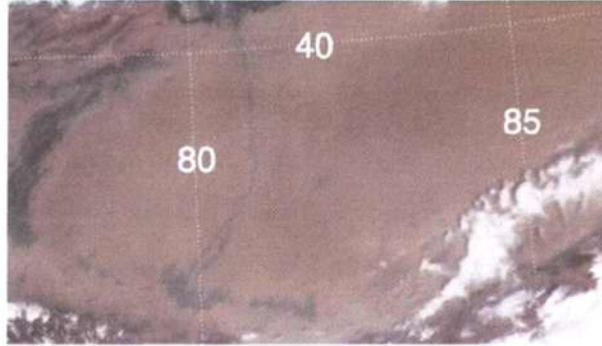


图 4.3 由 Aqua 卫星上 MODIS 0.65 μm , 0.56 μm 和 0.47 μm 三通道数据合成的
2006 年 7 月 29 日中国西北地区真彩卫星图片

在我们确定这一沙尘过程中并没有云层后, 我们就可以从 CALISPO 观测资料得到的后向散射系数廓线, 按照公式 (4.1) 直接计算气溶胶光学厚度的垂直廓线 (Hu et al., 2006)。

$$\tau(z) = \frac{1}{2\eta} \ln(1 - 2\gamma'(z)S_{a,eff}) \quad (4.1)$$

其中 η 为多次散射因子, 在本章计算中其取值为 0.7; $S_{a,eff}$ 为单次雷达散射比率, 即总的消光系数与后向散射系数的比值, Omar 等人的工作总结给出了 6 种主要气溶胶类型的单次雷达比率, 其中沙漠沙尘类型的气溶胶的 $S_{a,eff}$ 值在 532nm 和 1064nm 波段分别为 40 和 30 (Omar et al., 2004), 计算中我们将单次雷达散射率值定为 44。公式 (4.2) 为 $\gamma'(z)$ 的定义公式, 是指某计算层, 将后向散射系数从层顶积分到层底的垂直积分值。

$$\gamma'(z) = \int_{z_{top}}^{z_{base}} \beta_a(z) dz \quad (4.2)$$

式中 $\beta_a(z)$ 为 CALIPSO 垂直分辨率上的每层的后向散射系数。

当计算得到每层气溶胶的光学厚度后, 我们也就很容易得到消光系数 β 的垂直廓线, 如公式 (4.3) 所示。

$$\beta(z) = \frac{\tau(z)}{\Delta z} \quad (4.3)$$

图 4.4 即为计算得到的 7 月 24-31 日沙尘过程中沙尘气溶胶的分布。图 4.4 左图为 CPLIAPSO 运行轨道上, 沙尘气溶胶消光系数的垂直分布截面图; 右图为沙尘消光系数垂直廓线的卫星运行轨道平均值。

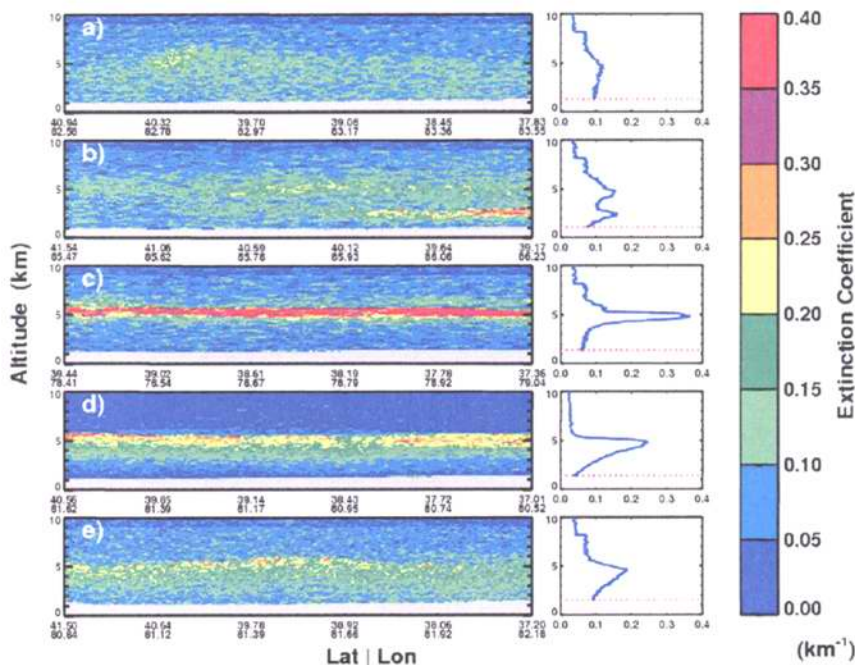


图 4.4 2006 年 7 月 24 (a 图)、26 (b 图)、29 (c 图)、30 (d 图)、31 (e 图) 日塔克拉玛干地区沙尘气溶胶消光系数在卫星轨道上的剖面图 (左图); 沙尘消光系数垂直廓线的卫星运行轨道平均值 (右图)

从图 4.4 中可以看出: 此次沙尘过程从 24 日开始发展, 沙尘主要分布在地面和大气层 5km 之间, 沙尘消光系数从地表处的 0.1km^{-1} 随着高度慢慢增加, 在 5km 处达到最大, 随后又随着高度递减。发展到 26 日时, 沙尘在卫星运行轨道的不同位置上, 出现两个不同高度的沙尘层, 一个位于 3km 处, 另一个位于 5km 处, 两个高度上的沙尘层消光系数都为 0.17km^{-1} 左右。随着沙尘过程不断发展, 到 29、30 日到达顶峰, 5km 处形成厚厚的一个沙尘层, 在这一层的消光系数最大可达到 0.35km^{-1} , 而由于沙尘层高度较高, 沙尘很容易向下游传送, 甚至爬上青藏高原, 就有结果证明在此次沙尘过程后期在青藏高原也有发现沙尘层

(Huang et al., 2007)。此次过程到 31 日后逐渐减弱。

为了获得此次沙尘过程中沙尘气溶胶整层光学厚度的信息和在整个塔克拉玛干地区的空间分布,我们计算了卫星运行轨道上垂直整层沙尘气溶胶的光学厚度,如图 4.5 所示。从图中可以看出,在沙尘发展初期(24-26 日)整个沙漠地区的沙尘分布并不均一,有明显的空间变化,沙尘光学厚度分布在 0.4-0.9 之间;而到了发展顶峰时期(29-30 日),沙尘在塔克拉玛干沙漠地区并无显著空间变化,均匀的分布着很厚的沙尘层,光学厚度可达到 0.8-0.9。

4.2.4 沙尘气溶胶粒子光学特性

在得到了沙尘气溶胶的光学厚度和垂直分布后,要计算辐射强迫值和加热率还必须得到沙尘粒子的散射特性参数:单次散射反照率和不对称因子。由于目前还没有任何有关塔克拉玛干沙漠中心的野外观测数据,沙尘粒子散射特性参数无法利用观测数据反演得到,所以我们利用 CERES 卫星大气层顶辐射通量的观测值以及 Fu-Liou 辐射传输模式的模拟计算来得到近似的气溶胶粒子的光学特性参数值。Fu-Liou 模式中定义了四种不同沙尘粒子类型的光学特性参数(模式的详细介绍请参看第二章内容),分别是:集结性矿物沙尘(Nuclearion)、堆积性矿物沙尘(Accumulation)、粗糙性矿物沙尘(Coarse)和传输性矿物沙尘(Transported)(Hess et al., 1998)。定义的光学特性参数主要包括:消光系数、单次散射反照率和不对称因子。其中消光系数是归一化的消光系数,只是用来计算光学厚度/消光系数的波谱分布,0.67 μm 波段上的单次散射反照率和不对称因子值可参见表 4.1。我们的思路就是利用 Fu-Liou 模式分别计算这四种不同散射性质的沙尘粒子在 TOA 处的向上短波辐射通量,并与 CERES 卫星数据进行比较,计算结

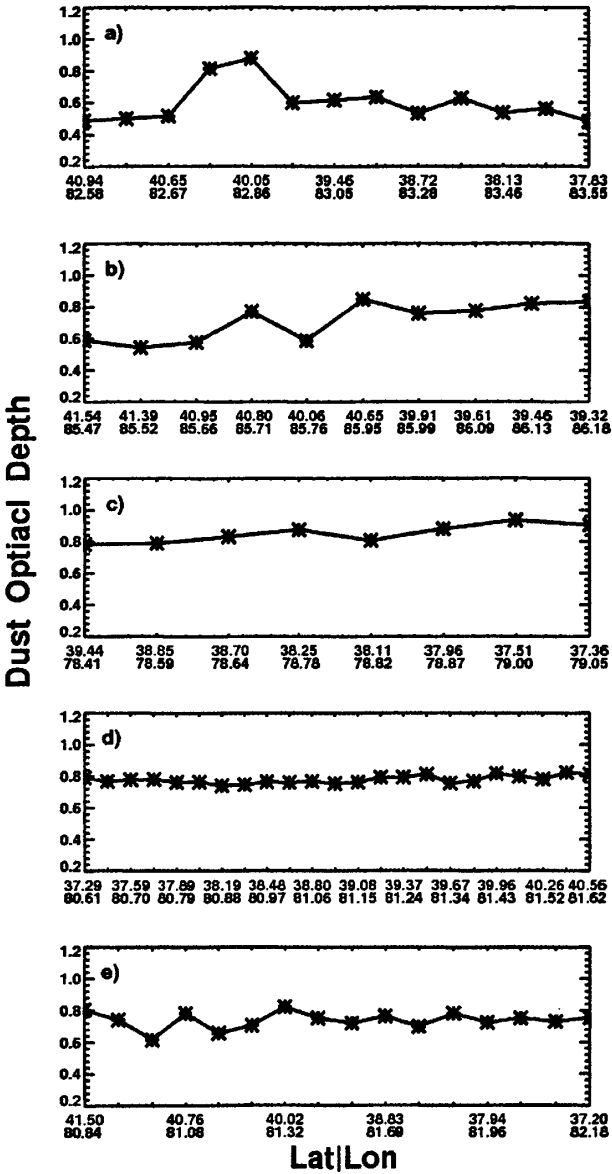


图 4.5 2006 年 7 月 24 (a 图)、26 (b 图)、29 (c 图)、30 (d 图)、31 (e 图)
日塔克拉玛干沙漠地区沿卫星轨道的沙尘气溶胶的整层光学厚度

果与之误差最小时使用的沙尘粒子类型,则认为是可以代表塔克拉玛干地区沙尘气溶胶的粒子类型。在这个过程中,因为地表反照率对 TOA 处的短波通量的影响很大,所以地表反照率参数的选取对于计算结果至关重要。为了在模式中输入准确的地表反照率,我们使用了 CERES 卫星反演数据组基于 MODIS 观测资料

表 4.1 Fu-Liou 模式中四种沙尘粒子类型在 0.67 μm 上的三次散射反照率和
不对称因子

	Nucleation mode	Accumulation mode	Transported mode	Coarse mode
Single-scattering albedo (ω)	0.9740	0.9093	0.8763	0.7118
asymmetry factor (g)	0.6521	0.7162	0.7532	0.8741

得到宽谱地表反照率 (Minnis et al., 2008), 并利用模式中的沙漠地表特性地表反照率的谱分布, 将其订正到各个波段上, 图 4.5 以 7 月 31 日一个计算点为例给出了订正后各个波段上地表反照率。由于是沙漠中心的地表几乎没有植被, 所以地表反照率时空变化很小, 其他几日和各个计算点的宽谱的地表反照率平均值为 0.346, 标准差仅为 0.005。另外, 我们利用美国空间系统研究实验室的 NCEP/NCAR 再分析资料得到模式计算所需的温度和水汽的大气廓线资料。

在得到了所有模式计算需要的参数值后, 我们首先利用 Fu-Liou 模式分别选用四种具有不同光学特性的沙尘粒子种类, 计算 TOA 处向上的短波辐射强迫值。结果发现只有选用传输型矿物沙尘时, 所得的计算结果才与卫星观测结果较为一致。图 4.6 所示即为这一次沙尘过程中沿 CALIPSO 卫星轨道的各个计算点 TOA

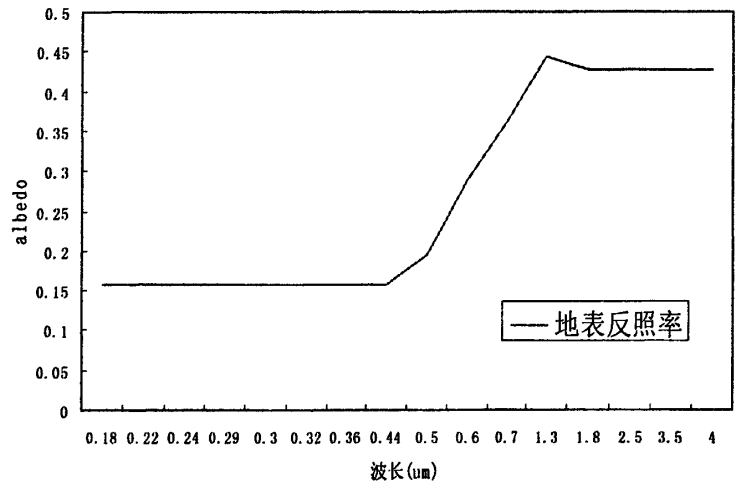


图 4.5 7 月 31 日塔克拉玛干地区 (41.50° N, 80.84° E) 各个波段的平均地表反照率

处向上模式计算的辐射通量值与卫星观测值对比的散点图。从图中可以看出,大部分计算点的两种计算结果具有很好的一致性,模式计算的各点平均值为 270.07 W/m^2 , 卫星观测平均值为 272.07 W/m^2 , 两者误差为 1.47 W/m^2 , 这种误差也在合理可接受范围内。所以可以认为传输型矿物沙尘 (Transported) 可以用来较为准确的代表塔克拉玛干地区的沙尘粒子特性, 这一区域沙尘粒子的单次散射率大概在 0.88 左右, 这也被后来 Ge 等人的工作所证实此取值是合理。Ge 等人利用 2008 年中美联合观测张掖站 MFRSR 观测数据, 反演得到起源于塔克拉玛干和内蒙古戈壁的沙尘气溶胶的光学特性, 其中反演的粒子单次散射反照率和不对称因子分别为 0.82 和 0.71 (Ge et al., 2009), 这与我们使用的传输型沙尘的 SSA 和 g 都只有 6% 的相对误差。由此我们认为利用模式模拟和卫星观测资料结合得到气溶胶的光学特性是一种合理可行的方法。

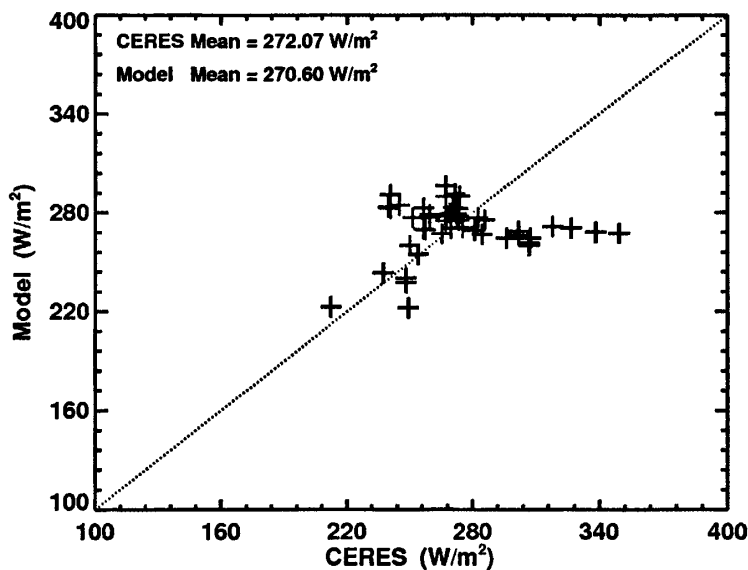


图 4.6 2006 年 7 月 24 (a 图)、26 (b 图)、29 (c 图)、30 (d 图)、31 (e 图) 日塔克拉玛干沙漠地区沿 CALIPSO 卫星轨道各计算点 TOA 处向上的模式计算的辐射通量值与 CERES 卫星观测值的对比散点图

4.2.5 沙尘气溶胶辐射强迫的计算

从上面的计算分析中, 我们已经得到了要计算沙尘辐射强迫值所需的重要参

数值, 如: 大气温湿廓线、地表反照率、沙尘光学厚度、沙尘的垂直分布廓线以及沙尘粒子的单次散射反照率和不对称因子。我们将这些参数值输入 Fu-Liou 模式来计算大气层顶、地表和大气内部的沙尘直接辐射强迫值。因为短波辐射也就是太阳辐射只发生在白天, 夜晚只有长波辐射, 所以我们将短波辐射进行 24 小时平均, 进行日平均短波、长波以及净辐射强迫计算。表 4.2(a-c)所列为计算的塔克拉玛干地区大气层顶、地表和大气内部的短波、长波和净辐射强迫值。从表中可以看出, 短波和长波在此处辐射强迫值分别为 14.11Wm^{-2} 和 30.30Wm^{-2} , 净辐射强迫可以达到 44.41Wm^{-2} , 这表明沙尘在大气层顶是显著的加热作用。一般硫酸盐等非吸收性粒子在 TOA 处都为负的辐射强迫作用, 而这里我们计算得到的 14.11Wm^{-2} 正的辐射强迫, 主要是由于沙尘的强吸收性以及沙尘层位于高反照率的沙漠地表之上。由于大气中的沙尘层阻挡了短波辐射到达地面, 所以对地表是显著的冷却作用, 在地表的短波辐射强迫为 -64.72Wm^{-2} , 但同时沙尘层由于其强烈的吸收性, 使得自身温度升高, 所以在此过程中, 沙尘也在不断的向地表辐射长波能量, 因此沙尘在地表的长波辐射强迫为 22.83Wm^{-2} , 减缓了短波辐射的冷却效应, 总的净辐射强迫为 -41.89Wm^{-2} 。得到了 TOA 和地表上的辐射强迫值以后, 将两者相减则得到了大气内部的辐射强迫值。大气内部短波的辐射强迫可以达到 78.83Wm^{-2} , 长波和净辐射强迫为 7.47Wm^{-2} 和 86.31Wm^{-2} 。从结果可以看出沙尘气溶胶 TOA 和地表的辐射平衡中都扮演着重要角色: 沙尘使整个地气系统向外空间释放的能量减少了, 又同时阻挡了很大一部分能量到达地表, 所以我们可以总结出在此次研究中沙尘的主要作用是将大部分能量留在了大气中, 从而改变大气的垂直加热结构。

表 4.2(a) 塔克拉玛干地区 TOA 处日平均辐射强迫值 (单位: Wm^{-2})

Date	July 24	July 26	July 29	July 30	July 31	Mean
SW	12.49	15.6	14.93	14.42	13.13	14.11
LW	28.64	28.86	32.73	27.64	33.65	30.30
Net	41.13	44.46	47.66	42.06	46.78	44.41

表 4.2(b) 塔克拉玛干地区地表上日平均辐射强迫值 (单位: Wm^{-2})

Date	July 24	July 26	July 29	July 30	July 31	Mean
SW	-49.70	-67.0	-73.53	-67.27	-66.12	-64.72
LW	19.92	23.37	21.93	22.58	26.36	22.83
Net	-29.78	-43.63	-51.60	-44.69	-39.76	-41.89

表 4.2(c) 塔克拉玛干地区大气内部日平均辐射强迫值 (单位: Wm^{-2})

Date	July 24	July 26	July 29	July 30	July 31	Mean
SW	62.19	82.6	88.46	81.69	79.25	78.83
LW	8.72	5.49	10.8	5.06	7.29	7.47
Net	70.91	88.09	99.26	86.75	86.54	86.31

4.2.6 沙尘气溶胶辐射加热率的计算

从上节辐射强迫的计算, 我们得出结论, 沙尘气溶胶的主要作用是使得大部分能量留在了大气内部, 所以这就必然使得大气内部垂直加热率的改变。为了进一步明确是这一效应以及沙尘对大气温度的改变, 我们利用 Fu-Liou 模式进行了大气垂直辐射加热率的计算。

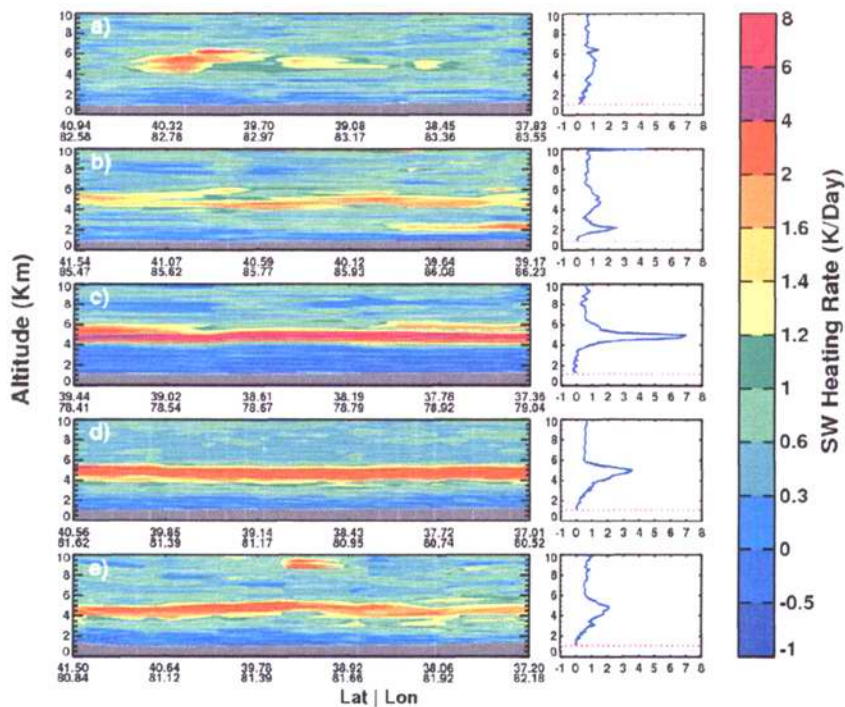


图 4.7 左图为卫星轨道上日平均短波辐射垂直加热率的剖面图,右图为轨道平均的日平均短波辐射垂直加热廓线图 (a)7月 24 日(b)7月 26 日(c)7月 29 日(d)7月 30 日(e)7月 31 日

图 4.7-4.9 为计算的 7 月 24 日到 31 日这一次沙尘过程中, 沙尘气溶胶对日平均的短波、长波和净辐射大气垂直加热率的影响。这是通过计算有沙尘影响下的大气加热率和无沙尘条件下大气加热率的差值而得到的。另外和辐射强迫值的计算一样, 加热率的计算也都是 24 小时平均后的日平均加热率。从图中可以看出, 沙尘对于短波辐射加热率有着显著作用。在沙尘层厚度较小 (图 4.7a)、中等 (图 4.7b 和 e) 和较大 (图 4.7c 和 d) 三种情况下, 沙尘层的短波辐射加热率可以分别达到 1, 2 和 3k/day, 这主要是由于沙尘粒子对短波辐射的吸收作用造成的。从图 4.7 中我们可以看出最大的日平均短波加热率出现在 7 月 29 日, 大概 5km 处, 达到了 7 k/day, 由此也可以看出短波加热率峰值的出现与沙尘消光系数的最大值是想对应的, 这说明沙尘粒子对短波的消光中, 吸收作用是其主要作用。

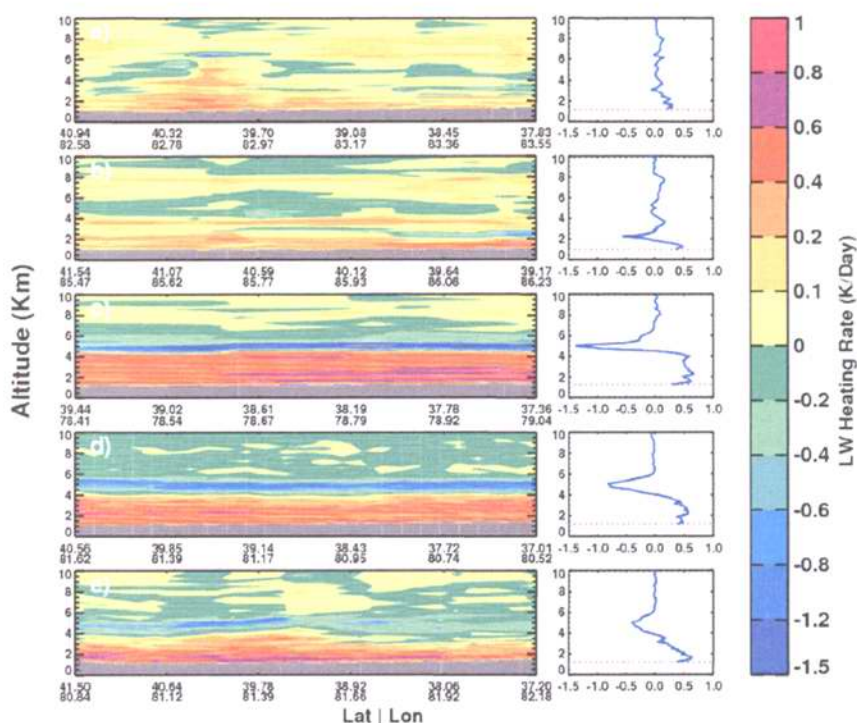


图 4.8 和图 4.7 一致，此图为长波辐射加热率

图 4.8 所示为长波辐射加热率，从图中可以看出沙尘造成的长波辐射加热率没有短波显著，但在沙尘层以下表现出明显的加热作用以及在沙尘层上部为冷却作用，这是因为沙尘气溶胶吸收短波能量使得沙尘层温度增高，所以会向下辐射能量，从而加热沙尘层以下大气，同时沙尘层因为向外辐射长波能量，所以为冷却作用。沙尘造成的沙尘层下大气的增暖作用在地表最为明显，可以达到 0.5 k/day 。沙尘层的冷却效应根据沙尘厚度的不同大概在 $0\sim 1.5\text{ k/day}$ 的范围内。这部分长波的冷却作用部分抵消了沙尘层的短波加热作用，所以沙尘层的净辐射加热率在 24 日为 1 k/day ，26 日为 $1.5\sim 2.0\text{ k/day}$ ，29 日为 5.5 k/day ，30 日为 3 k/day ，31 日为 1.5 k/day ；沙尘层下部的净辐射加热率的范围在 0.5 k/day 到 1.0 k/day 之间（如图 4.9 所示）。由以上分析我们可以看出，沙尘在沙尘层高度上短波的加热作用为主要作用，而对于沙尘层以下的大气则长波的加热作用占主导。

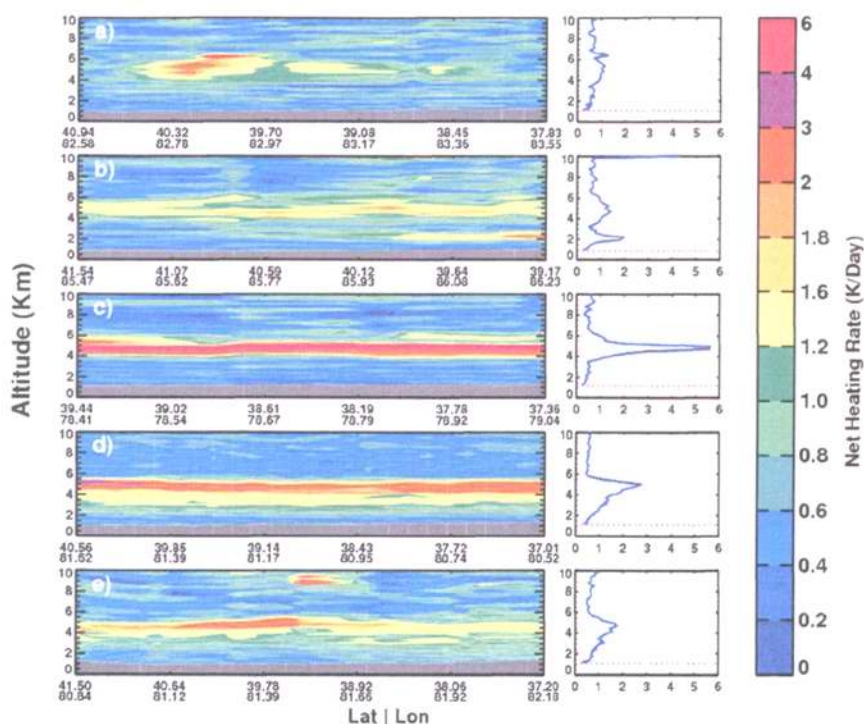


图 4.9 和图 4.7 一致，此图为净辐射加热率

4.2.7 辐射强迫和加热率计算的不确定性分析

虽然我们在利用模式计算辐射强迫和加热率时，尽量使用可靠的观测数据来减少计算结果的误差，但由于一些输入参数的不确定性还是不可避免的造成了计算结果的误差。本节量化给出了由于几个重要参数的不确定性对辐射强迫和加热率的影响。在本次计算中，重要的输入参数主要包括沙尘气溶胶的消光系数、地表反照率以及沙尘粒子的单次散射反照率。所以我们利用敏感性测试来估算这三个参数对辐射强迫值的影响。表 4.3 所示为敏感性试验结果。由于沙尘气溶胶的消光系数是根据 CALIPSO 卫星的后向散射比得到的，CALIPSO 雷达信号比的不确定性的范围在 20%左右 (Winker et al., 2006)，所以增大/减小雷达信号比 20% 会使得在 TOA、地表和大气的净辐射强迫值分别变化 ± 6.8 , ∓ 7.6 , $\pm 14.4 \text{ Wm}^{-2}$ 。地表反照率是致使计算结果不确定的另一个来源，我们将地表反照率增大或减少 10%，使得 TOA 处的短波辐射强迫值增大或减少了 4.59 Wm^{-2} ，地表上增大或减

少了 3.36 Wm^{-2} 。气溶胶粒子的单次散射反照率是目前计算气溶胶辐射强迫中最重要却又最不确定的因子。Dubovik 等人通过分析 AERONET 的反演结果指出，气溶胶粒子的单次散射反照率在变化 0.03，会使得估算的辐射强迫值产生 12% 的不确定性 (Dubovik et al., 2002)。通过表 4.3 也可以看出在我们的敏感性试验中，单次散射率造成的估算值的不确定值最大。SSA 变化正负 6% 会分别造成 TOA 处、地表和大气中净辐射强迫 $\mu 15.2, \pm 11.1, \mu 26.3 \text{ Wm}^{-2}$ 的变化。

表 4.3 辐射强迫值不确定性的估算值（单位： Wm^{-2} ）

	SSA $\pm 6\%$			Surface Albedo $\pm 10\%$			Lidar Ratio $\pm 20\%$		
	TOA	Surface	Atmos	TOA	Surface	Atmos	TOA	Surface	Atmos
SW	$\mu 14.815$	± 11.450	$\mu 26.265$	± 4.585	± 3.360	± 1.225	± 1.775	$\mu 11.155$	± 12.935
LW	$\mu 0.360$	$\mu 0.370$	± 0.010	0	0	0	± 4.820	± 3.480	± 1.335
Net	$\mu 15.175$	± 11.080	$\mu 26.255$	± 4.585	± 3.360	± 1.225	± 6.595	$\mu 7.675$	± 14.270

除了上述对辐射强迫不稳定性的分析外，我们还估算了粒子单次散射反照率、地表反照率和 CALISPO 雷达信号比这三个因子对辐射加热率造成的不确定性。在敏感性试验中我们发现，虽然上述的三个因子对辐射强迫的计算会有很大的影响，但对整个大气的垂直加热率的廓线形态不会有影响，只是会改变数值的大小。表 4.4 所列的就是三个因子变化时对辐射加热率数值的影响。从表中我们看出粒子的单次散射反照率依然是对结果影响最大的因子，当 SSA 增大或减小 6% 时，大气日平均加热率的数值会变化 0.28 k/day ，地表反照率对加热率影响最小，变化 $\pm 10\%$ 时，加热率仅改变了 ± 0.010 。并且这三个因子对加热率的影响主要是对短波的影响，对长波的作用很小。

图 4.4 辐射加热率不确定性的估算值（单位： k/day ）

	SSA $\pm 6\%$	Surface Albedo $\pm 10\%$	Lidar Ratio $\pm 20\%$
SW	$\mu 0.272$	± 0.010	± 0.138
LW	$\mu 0.004$	0	± 0.019
Net	$\mu 0.276$	± 0.010	± 0.157

4.3 中美联合观测期间张掖地区沙尘直接辐射强迫的研究

4.3.1 引言

目前对于沙尘直接辐射强迫的研究, 由于沙尘粒子光学特性参数数据的缺乏, 使得研究结果具有很大的不确定性。为了解决这一问题, 近些年来在亚洲地区开展了一系列的大型野外观测试验, 如亚洲气溶胶特性试验 (ACE-Asia), 亚洲大气环境气溶胶粒子试验 (APEX) 以及印度洋观测试验 (INDOEX) 等。虽然这些观测试验取得了很大的进展, 但是也有其局限性, 尤其是缺乏亚洲陆地内部的观测数据, 特别是中国西北内陆地区, 这样具有独特地理环境的气溶胶特性数据。2008 年兰州大学大气科学学院、中科院大气物理所、中国气象局、美国能源部和马里兰州等单位联合举行的西北地区野外观测试验, 以兰州大学半干旱气候与环境观测站为中心, 另设景泰、张掖国家气候观测台两个移动观测点, 历时三个多月, 获得了我国西北地区气溶胶粒子特性的第一手资料。Ge 等人通过分析张掖站点的观测数据反演了来自于塔克拉玛干沙漠和内蒙古戈壁的沙尘粒子的光学厚度、单次散射反射率、不对称因子、粒子谱分布和复折射指数 (Ge et al., 2009)。表 4.5 所列为反演的沙尘粒子的各个参数的平均值, 从表中数据可以看出反演的 0.5 μm 波段上粒子的单次散射反照率仅为 0.75, 这说明西北地区沙尘气溶胶与其他地区的粒子的光学特性相比有明显差别, 具有强烈的吸收特性。本节将在得到沙尘粒子反演参数值的基础上, 进行辐射闭合试验用以验证粒子光学参数值反演的可靠性, 并进一步计算沙尘的直接辐射强迫值。

4.3.2 辐射闭合试验

利用 MFRSR 观测资料反演得到气溶胶粒子的光学特性后 (反演的详细过程请参见 Ge et al., 2009), 我们通过比较地面观测得到的与模式计算模拟的总辐射通量、直接辐射通量和间接辐射通量值, 进行辐射闭合试验从而用于检验反演的粒子光学特性参数的准确性。在利用辐射传输模式计算辐射通量时, 最重要的输入参数是气溶胶的整层光学厚度、粒子的单次散射反照率和不对称因子, 表 4.5 给出了从联合观测试验期间挑选的 11 个沙尘例子反演得到的沙尘粒子的重要特性参数。除了上述几个重要的输入参量外, 计算时模式中还需输入地表反照

率、水汽含量和臭氧含量等值。其中大气温度和水汽的垂直廓线来自于 NCEP/NCAR 日平均资料；臭氧含量数据是从 TOMS 卫星观测资料中得到的；地表反照率的宽谱值是 CERES 数据中得到的。为了得到每个波段上地表反照率的值，我们首先假设下垫面是沙地和农田的混合地型，然后不断的变化这两种地型所占的百分比，直到在这种百分比混合下由各个波段地表反照率的值计算得到的宽谱的地表反照率与 CERES 观测值一致，由此我们得到了每个波段上地表反照率的值。11 个挑选个例的宽谱地表反照率的平均值为 0.21，沙地和农田的混合百分比分别为 85%和 15%。

表 4.5 从 MFRSR 反演的 11 个沙尘个例的 0.67 μm 波段上的气溶胶的垂直光学厚度、单次散射反照率、不对称因子和埃斯屈朗指数的平均值和标准差

	平均值	标准差
光学厚度 (AOD)	0.19	0.07
埃斯屈朗指数 (Angstrom)	0.58	0.19
单次散射反照率 (SSA)	0.82	0.01
不对称因子 (g)	0.71	0.02

从试验结果来看，模式计算结果与观测数据具有很好的一致性，这说明反演的沙尘粒子的参数值可以较好的代表真实的沙尘气溶胶光学特性。图 4.10 所示为 2008 年 4 月 24 日地面观测的瞬时短波的总辐射、直接辐射和散射辐射的值与模式计算值的对比。地面观测的总辐射通量值是通过精确波谱光度计 (Precision Spectral Pyrometer) 测得的；散射辐射通量是有 CM21 光度计得到的；直接辐射通量的测量是由两种独立的仪器 Eppley Normal Incidence 光度计和 CH1 太阳光度计同时进行的，两种数据可以进行相互检验。在 4 月 24 日个例的分析中，地面观测的日平均总辐射通量值、散射辐射通量值和垂直辐射通量值与模式模拟结果的差分别是 -3.6、0.9 和 0.8 W/m^2 。从 4 月 24 个例的试验结果来看，模式计算结果与观测数据具有很好的一致性，这说明反演的沙尘粒子的参数值可以较好的代表真实的沙尘气溶胶光学特性。

表 4.6 6 个完全无云例子的模式模拟值与地面观测值的平均误差

	Total SW Flux (Wm^{-2})	Diffuse SW Flux (Wm^{-2})	Direct SW Flux (Wm^{-2})
Mean observed Flux	609.82	136.94	674.66
Mean simulated Flux	601.28	134.04	672.59
Mean Flux difference Mod - Obs	-8.54	-2.90	-2.07

为了进一步验证反演的沙尘粒子光学特征参数值的准确性,我们挑选了 6 个完全无云的沙尘个例,进行辐射闭合试验。表 4.6 给出了这 6 个例子的平均计算结果。从表中我们可以看出,模式模拟计算的总辐射通量、散射辐射通量和直接辐射通量值与观测结果的差值分别是-8.5、-2.9 和-2.1 W/m^2 。从上述结果我们可以总结得出:模式计算的直接辐射通量与观测数据的一致说明我们所反演的气溶胶的光学厚度和大气中气体的吸收是准确的;观测与模式散射辐射通量计算值的吻合说明反演的粒子的单次散射反照率 and 不对称因子是合理的。另外,由于地表的总辐射和散射辐射通量对于地表反照率是不敏感的,所以我们无法从模式和观测得到的这两个量数据的一致性上,来说明模式中所用地表反照率值的可信性,所以我们通过对比模式和观测计算的大气层顶处(TOA)向上的辐射通量值来检测地表反照率值的准确性。图 4.11 给出了挑选的无云的 6 个例子的模式模拟的 TOA 处的向上通量值与 CERES 卫星观测值的对比。模式与观测的 TOA 处瞬时向上辐射通量平均值的差别为 8 W/m^2 ,这说明模式中所用的地表反照率可以较好的代表真实的下垫面条件。

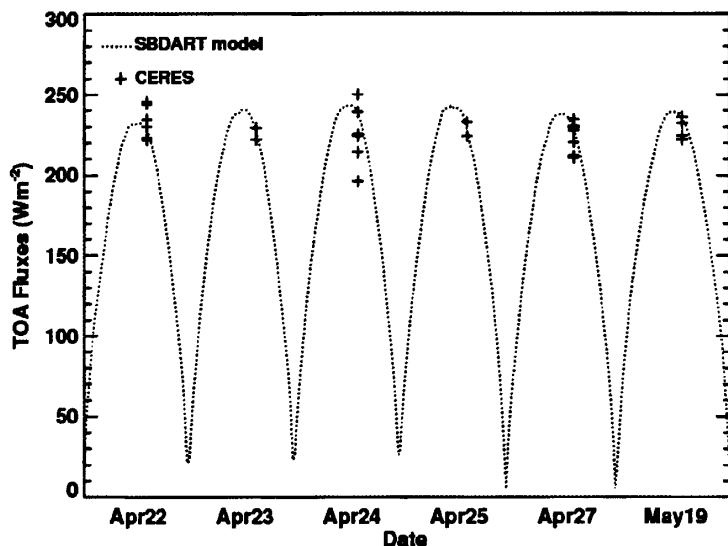


图 4.11 CERES 卫星观测的 TOA 处向上的辐射通量值（十字符号）和模式模拟值（点线）的对比。

4.3.3 直接辐射强迫计算

气溶胶粒子微物理特性是决定气溶胶直接辐射强迫的重要因子。上述辐射闭合试验的成功,使得我们对代表沙尘粒子微物理特性的各个参数的反演数据的准确性有了很大的信心,所以在此基础上我们计算了沙尘的直接辐射强迫,以期计算结果具有较高的准确性和代表性。我们将上述的粒子特性参数(光学厚度、单次散射反照率 and 不对称因子等)和大气及地表参数输入 SBDART 辐射传输模式计算了地表和大气层顶处的沙尘气溶胶短波辐射强迫值,计算公式如下所示:

$$T = F_{down} - F_{up}$$

$$C = T - T^{CLR}$$

其中 F 为辐射通量, T 为净辐射通量, C 为辐射强迫值; 下标 $down$ 和 up 分别代表通量的方向: 向上和向下; 上标 CLR 代表晴空下的净辐射通量。

图 4.12 所示为地表和大气层顶的日平均沙尘直接辐射强迫值。从图中可以看出, 沙尘气溶胶在地表具有很强的负辐射强迫作用, 最大值的出现伴随着气溶

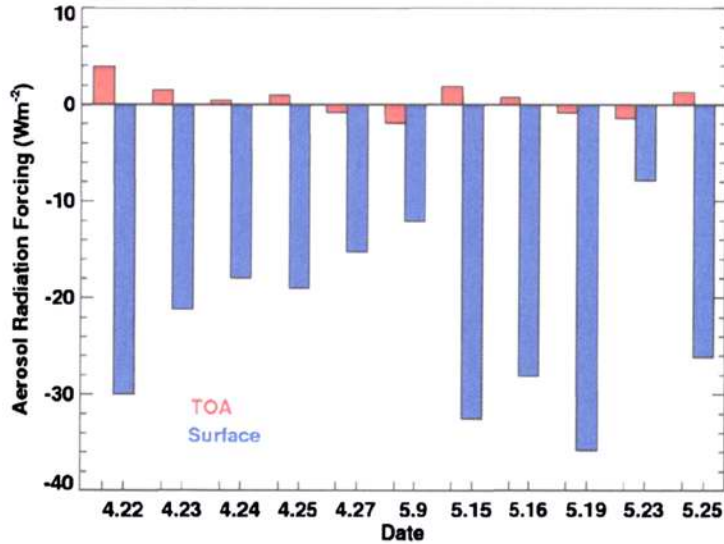


图 4.12 11 个沙尘个例大气层顶 (红色) 和地表 (蓝色) 处直接辐射强迫值。胶光学厚度的最大值发生在 5 月 19 日, 平均直接辐射强迫为 $-22.4 \pm 8.9 \text{ Wm}^{-2}$, 这会显著的导致地表温度的降低。McFarlane 等人计算的非洲 Niamey 地区地表的辐射强迫值为 $-21.1 \pm 14.3 \text{ Wm}^{-2}$, 这与我们的计算结果非常相似 (McFarlane et al., 2008)。

在全球辐射能量平衡中, 气溶胶在大气层顶的直接辐射强迫通常是负值 (Yu et al., 2006; Kim et al., 2008), 这说明气溶胶在整个气候系统中起到的是冷却作用。而对于区域尺度来说, 我们计算的沙尘气溶胶在大气层顶的直接辐射强迫值非常小 (小于 4 Wm^{-2}), 并且根据每个例子单次散射反照率 (SSA) 不同, 11 个个例辐射强迫的计算值有正有负, 这显示说明沙尘气溶胶对于半干旱地区的地气短波辐射能量平衡影响很小。但是我们应该注意到, 全球尺度上气溶胶辐射强迫效应对气候影响则是各个区域尺度上作用的综合体现, 而各个地区的辐射强迫效应, 则是根据各个区域的气溶胶种类和地表反照率的不同, 有着显著差别。

图 4.12 显示, 虽然沙尘气溶胶在大气层顶的辐射强迫值都接近于零, 但是

在地表却表现出很大的负的辐射强迫值, 这种地表与大气层顶很大的差值说明, 大气中由于气溶胶的存在使得很大一部分太阳辐射被吸收留在了大气中, 从而加热大气, 减少涡旋热汇聚 (Miller et al., 1998, 1999)。有研究指出, 由于气溶胶这种冷却地表、加热大气的作用, 使得大气边界层垂直温度梯度加大, 这可能会减少蒸发, 影响云的形成 (Hansen et al., 1997; Ackerman et al., 2000), 进一步减弱水分的循环 (Ramanathan et al., 2001)。

在计算了 TOA 和地表处的日平均气溶胶短波辐射强迫后, 我们还计算了地表处的日平均气溶胶辐射强迫效率 (ARFE), 用以进一步研究气溶胶辐射强迫与光学厚度的关系。日平均气溶胶辐射强迫效率的计算是通过将气溶胶日平均辐射强迫值除以日平均光学厚度值得到的。为了便于与其他结果进行比较, 我们利用 $0.5\mu\text{m}$ 波段上的光学厚度进行计算。地表处气溶胶短波平均辐射强迫效率为 $-95.1 \pm 10.3 \text{ Wm}^{-2} \tau^{-1}$ 。图 4.13 所示为计算的地表处日平均辐射强迫效率与单次散射反照率的关系, 从图中可以看出, 当 SSA 较小时, 就会在地表产生较强的负的辐射强迫效率, 这与我们预期一致, 因为 SSA 值越小, 就说明气溶胶的吸收性越强, 就会有越多的太阳辐射被气溶胶吸收从而无法到达地表。

与 2006 年春季在 Niamey 的计算结果相比, 由于我们反演的单次散射反照率较小, 所以估算得到的辐射强迫效率比 Niamey 地区的计算值大 $40 \text{ Wm}^{-2} \tau^{-1}$ (McFarlane et al., 2008)。Kim 等人的研究指出东亚沙尘气溶胶的辐射强迫效率应在 -55 到 $-106 \text{ Wm}^{-2} \tau^{-1}$ 之间, 我们的计算的平均结果基本位于此范围的上限, 有个别例子的结果甚至超过了上限值。但是, Kim 等人提出这个范围值时也曾指出, 当 ARFE 的值超过了范围上限时, 很有可能是分析的沙尘粒子混合了黑碳粒子而造成的。我们计算的 ARFE 值较大, 同时反演的单次散射率较小, 但这些结果是否是由于所研究的沙尘气溶胶中混合了黑碳粒子所造成的, 还需要通过进一步的工作来研究证明。

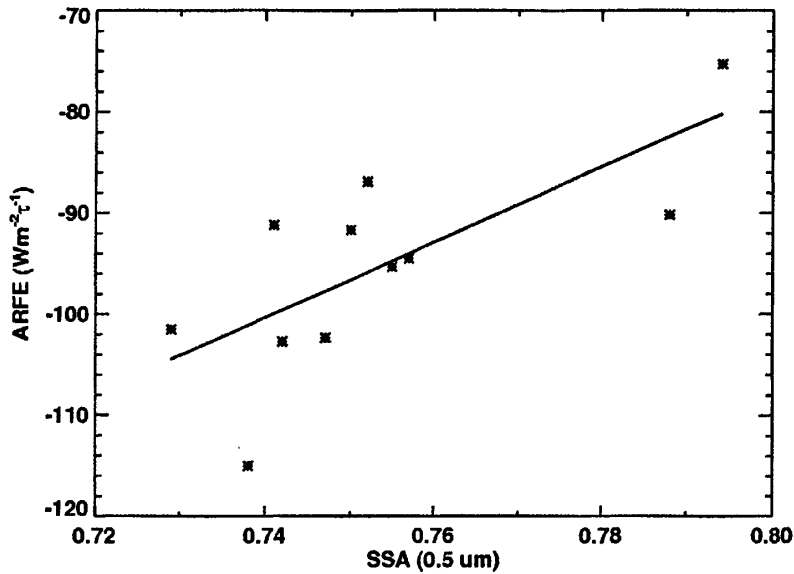


图 4.13 0.5 μm 波段上 11 个例子的日平均地表短波辐射强迫效率随 SSA 的变化图

4.4 本章小结

本章主要研究了沙尘气溶胶在晴空条件下对地气系统产生的影响,即直接辐射效应。本章工作主要包含两个主要的工作,第一项工作是对 2007 年夏季塔克拉玛干沙漠地区的一次强沙尘暴过程进行了研究。在此项工作中我们利用 CALIPSO 和 CERES 卫星资料,通过 Fu-Liou 模式研究计算了塔克拉玛干地区在此次沙尘暴过程中沙尘气溶胶在大气层顶、地表和大气中的短波、长波和净辐射强迫效应以及大气沙尘层对大气的加热作用。研究发现,沙尘暴过程中,大气中的沙尘层会根据不同的沙尘厚度产生 1-3 k/day 的净辐射加热,最强的辐射加热在沙尘暴发展的顶峰出现,可达到 5.5 k/day。在 TOA、地表和大气的日平均辐射强迫分别为 44.4、-41.9 和 86.3 W/m^2 。其中 TOA 处的辐射增暖作用有三分之二是由于沙尘对短波的吸收和长波的放射所造成的,而这其中沙尘对短波的吸收为主要作用可占 90%左右;在地表处由于长波的放射增暖作用抵消了三分之一的短波冷却效应。这些都显著表明沙尘气溶胶的短波和长波的辐射效应对于 TOA 和地表的能量平衡都起着重要的作用。由上面结果我们看到沙尘的短波直

接辐射强迫对地气的能量平衡有重要作用,光学厚度和单次散射反照率,特别是 SSA 作为重要的气溶胶参数,对短波直接辐射强迫计算起着决定性作用。在塔克拉玛干地区沙尘暴过程的计算中,因为没有地面观测资料可以用来反演 SSA,所以我们通过比较 CERES 和模式计算的 TOA 处辐射通量值的方法确定沙尘种类,从而得到 SSA。但是我们得到的沙尘气溶胶的 SSA 在 0.67 μm 波段上仅为 0.88,非洲撒哈拉地区的研究结果在 0.94 左右 (Dubovik et al.,2002; Sally et al., 2009) 这种结果的不一致使我们觉得应该对东亚沙尘粒子的单次散射反照率以及其他光学特性参数进行进一步的研究。所以在后续第二项工作中,我们利用 2008 年中美联合观测试验中在张掖地区的地面观测资料,反演了沙尘气溶胶的各种光学特性,其中反演的 0.67 μm 波段上的单次散射反照率为 0.82,与塔克拉玛干地区的计算结果基本一致。两种完全不同的方法得到了较为一致的结果,这增强了我们对 SSA 反演结果的信心。为了进一步检验沙尘粒子反演参数的可靠性,我们利用地面观测的辐射通量与模式模拟数据,进行了辐射闭合试验,地表处观测的辐射通量、散射辐射通量和直接辐射通量与模式的模拟值仅差 -8.5、-2.9 和 -2.1 W/m^2 ,这说明辐射闭合试验是成功的,沙尘粒子光学特性参数的反演、大气廓线和地表反照率的选取都具有一定的可靠性。在此基础上,我们计算了 TOA 处和地表处沙尘气溶胶的短波日平均辐射强迫值。地表处的日平均辐射强迫值随着沙尘气溶胶光学厚度的变化在 -7.9~-35.8 W/m^2 之间变化,而 TOA 处的辐射强迫值很小,且有正有负,平均值趋于零值。通过此项工作可以说明,在半干旱地区,沙尘气溶胶有效阻挡了太阳光到达地表,增加了大气中太阳短波辐射的吸收,这势必将会影响大气垂直温度分布,对大气层结的稳定和水分循环产生影响。

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第五章 沙尘气溶胶的间接辐射强迫辐射效应

5.1 引言

由于气溶胶会对全球及区域气候产生重要的影响,所以气溶胶辐射强迫的研究已经成为气候研究中最活跃的领域(IPCC, 2007)。气溶胶通过反射和吸收太阳辐射而产生的作用,称之为气溶胶直接辐射强迫,在上一章中我们已经通过个例分析详细阐述了沙尘气溶胶直接辐射强迫效应。气溶胶对地气辐射平衡的影响除了通过直接辐射强迫作用外,还可以通过改变云的微物理特性从而间接的产生影响,即间接辐射强迫效应(Twomey, 1977; Albrecht, 1989),以及通过蒸发低层云,即半直接辐射强迫效应(Grassl, 1975),来改变能量平衡。目前许多观测及模式研究都侧重于对直接辐射强迫作用的研究,但是最近一些工作指出云微物理特性和云量的微小改变都会导致云辐射强迫的很大变化。所以,理解以及定量计算气溶胶的间接辐射效应对气候变化的研究起着至关重要的作用(IPCC, 2007)。

影响西北地区的沙尘暴通常发源于塔克拉玛干沙漠和蒙古戈壁,而且高空沙尘可以伴随着沙尘暴迁徙数千公里到达韩国和日本,所以沙尘气溶胶已经成为东亚地区最重要的气溶胶种类之一(Huang et al., 2006a)。由于沙尘粒子不仅可以反射太阳光造成冷却效应,也可以吸收太阳短波和长波辐射造成增暖效应,使得东亚地区的沙尘粒子与其他地区的沙尘具有显著区别,所以研究具有独特性质的东亚沙尘气溶胶的辐射强迫效应具有特殊意义。

虽然有许多工作已经对东亚地区沙尘气溶胶在晴空条件下对辐射能量平衡的影响做了深入研究,但是很少有工作探索研究在有云条件下的沙尘作用。Huang 等人利用搭载在 Aqua 和 Terra 卫星上的 MODIS 和 CERES 卫星资料分析研究了东亚地区沙尘气溶胶对云的微物理特性和辐射强迫的影响。他们发现,平均而言,在云被沙尘粒子污染后,云的有效粒子直径、光学厚度和卷云的冰水路径比纯云条件下分别减少了 11%、32.8%和 42%。由于云的微物理特性发生了变化,而导致在有沙尘时云的瞬时净辐射强迫值从无沙尘状况下的 -161.6 w/m^2 增加到 -118.6 w/m^2 (Huang et al., 2006a)。但是上述结果是基于 10 个个例分析得到的。

本章将扩展上述研究个例,并利用辐射传输模式结合卫星资料分析结果将沙尘气溶胶直接辐射强迫和间接、半直接辐射强迫分开。以期通过计算研究沙尘对大气层顶和地表辐射能量的影响,能更好的理解沙尘气溶胶粒子与云的相互作用。

表 5.1 16 个挑选个例的详细信息

Case Image	Date	Lat. ° N	Lon. ° E
1	2003/03/26	44.0-47.5	123.5-124.5
2	2003/04/09	36.5-40.5	78.5-84.0
3	2004/03/09	44.0-46.0	104.0-113.0
4	2004/03/27	43.0-47.0	111.5-115.5
5	2005/03/09	39.5-45.5	108.5-111.5
6	2005/04/06	48.5-50.5	112.5-118.5
7	2005/04/08	40.0-41.5	102.5-105.5
8	2005/04/09	43.5-46.5	116.5-118.5
9	2005/04/13	45.5-49.5	112.0-117.0
10	2005/04/20	36.5-37.5	116.5-119.0
11	2006/02/17	43.0-48.0	113.0-115.0
12	2006/02/21	41.5-44.5	123.0-125.5
13	2006/04/17	36.5-40.5	100.0-106.5
14	2006/04/18	41.0-48.0	116.0-122.0
15	2006/04/22	42.0-48.0	116.0-117.0
16	2006/04/28	43.0-45.0	104.5-111.5

5.2 卫星资料介绍

在此次计算中我们分析了从 2003 年 6 月到 2006 年 6 月四年间,搭载于 Aqua 卫星上的 CERES 1B 版本的 SSF (Single scanner Footprint, Gerier et al., 2001) 数据资料。CERES SSF 卫星资料集结合了 CERES 辐射通量测量值、MODIS 云微物理参数反演值 (Minnis et al., 2008) 以及辅助的气象场资料,从而形成了一个合理的、高质量的集合了云和气溶胶卫星反演参数值以及辐射通量信息的,用于辐射平衡和气候研究的资料合集。SSF 数据合集中有 140 个参数值,在本次研究中我们使用了利用可见-红外-劈窗法 (VISST; Minnis et al., 2004,2008) 反演的冰云路径 (IWP)、水云路径 (LWP)、云顶有效温度 (T_e)、云滴有效半径 (R_e)、冰云有效直径 (D_e) 和光学厚度 (τ) 等参数值来研究沙尘气溶胶对云微物理特性的影响;另外使用 CERES 卫星探测的大气层顶 0.2-5.0 μm , 8-14 μm , 5-100 μm

三个波段范围的辐射通量值来估算沙尘对 TOA 处对云辐射强迫的影响。

为了估算东亚沙尘气溶胶对云微物理特性和辐射强迫的作用,我们选取了两个区域分别代表云在不同环境下的情况:CLD 区代表纯云区,即云处于无沙尘的环境中;COD 区代表沙尘云区,即云处于沙尘环境中,已经被沙尘污染。CLD 和 COD 区的选取是利用 MODIS 合成图像和地面测站观测资料来确定的(Huang et al., 2006a)。地面测站通常将观测到的沙尘分为四个档次进行报告,沙尘暴、吹沙、浮尘和无沙尘。如果一个区域由卫星观测表明为有云区域,且地面观测为无沙尘,则此区域定义为 CLD 区;但如果地面观测为沙尘暴、吹沙或者是浮尘,则定义为 COD 区。表 5.1 给出了本次计算中依据此条件挑选的 16 个研究个例的详细信息。虽然沙尘暴常常发生在相对湿度非常小的地区,特别是沙漠地区,但是大部分由冷锋或气旋系统造成的沙尘事件可以将云由系统将其带到湿度小较干的地区。例如,在本次挑选的 16 个例子中,除了第二个例子外其他 15 个例子的发生发展都是伴随着蒙古气旋发生的。在蒙古气旋系统中,云通常都生成在湿度较大较暖的冷锋前部,而沙尘暴也发生在冷锋前缘,所以沙尘云区也通常位于冷锋前部相同位置。

5.3 云辐射强迫

云辐射强迫是一个已经被广泛用于研究气溶胶-云相互作用的参数量,被定义为晴空条件和实际天空条件下净辐射通量的差(Charlock and Ramanathan, 1985; Hartmann et al., 1986; Ramanathan, 1987; Cess and Potter, 1987; Ramanathan et al., 1989)。

大气层顶(TOA)和地表的短波、长波辐射通量定义为:

$$\begin{aligned} T_{SW} &= F_{SW}^{down} - F_{SW}^{up} \\ T_{LW} &= F_{LW}^{down} - F_{LW}^{up} \end{aligned} \quad (5.1)$$

云辐射强迫定义为:

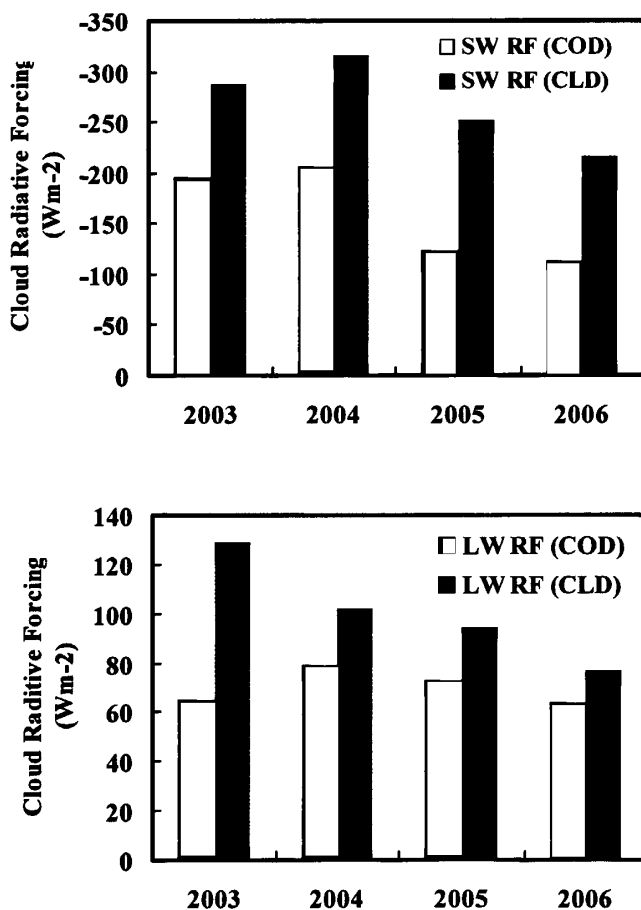
$$\begin{aligned} C_{SW} &= T_{SW} - T_{SW}^{CLR} \\ C_{LW} &= T_{LW} - T_{LW}^{CLR} \\ C_{NET} &= C_{SW} + C_{LW} \end{aligned} \quad (5.2)$$

其中, F_{SW}^{down} (F_{SW}^{up}) 和 F_{LW}^{down} (F_{LW}^{up}) 分别为向下(向上)短波和长波通量,上标 CLR

代表是无云，无沙尘的晴空条件。所以在计算 COD 区云的辐射强迫时，即包含了云的辐射作用，也包括了沙尘气溶胶的辐射作用。

估算辐射强迫时所用的辐射传输模式为 Fu-Liou 辐射传输模式，关于 Fu-liou 模式的介绍详见第二章。

图 5.1 所示的柱状图为通过分析 2003 年-2006 年卫星资料得到的 TOA 处 COD 区和 CLD 区辐射强迫的对比。在每个挑选个例的计算中，COD 区的短波、长波和净辐射强迫值的绝对值都小于 CLD 区。COD 区四年平均瞬时的短波、



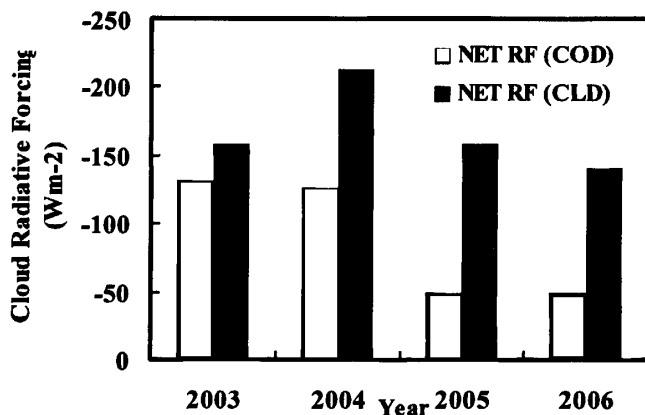


图 5.1 COD 和 CLD 区 TOA 处年平均瞬时短波、长波和净辐射强迫值的对比

长波和净辐射强迫分别为-138.9, 69.1 和-69.7 w/m^2 , 仅为 CLD 区值的 57.0, 74.2 和 46.3%。从图 5.1 我们可以看出, 沙尘有效减小了云在辐射平衡收支中的冷却作用, 这和 Huang 等人以前的结果一致, 他们的结果也都表明当沙尘存在于云层之上时, 沙尘粒子会有效减小 TOA 处云的冷却作用 (Huang et al., 2006a)。只是在 Huang 等人的结果中, COD 区的净辐射值较 CLD 减少 40%, 小于本次结果的 46.3%, 这也许是因为发生在 2005 年到 2006 年间的沙尘暴强度大于发生在 2003-2004 年的强度所造成的。所以沙尘在 COD 区对于 TOA 处云冷却作用的减弱可以看做是沙尘气溶胶的一种增暖作用。为了证明 COD 和 CLD 区云辐射强迫的变化不是由太阳入射能量和云性质的不同而引起的, 我们对比了 CLD 区和 COD 区云辐射强迫随太阳高度角 (SZA) 以及随云光学厚度的变化, 以及各种云微物理特性参数的对比, 例如云顶温度、光学厚度、粒子大小和云水路径。

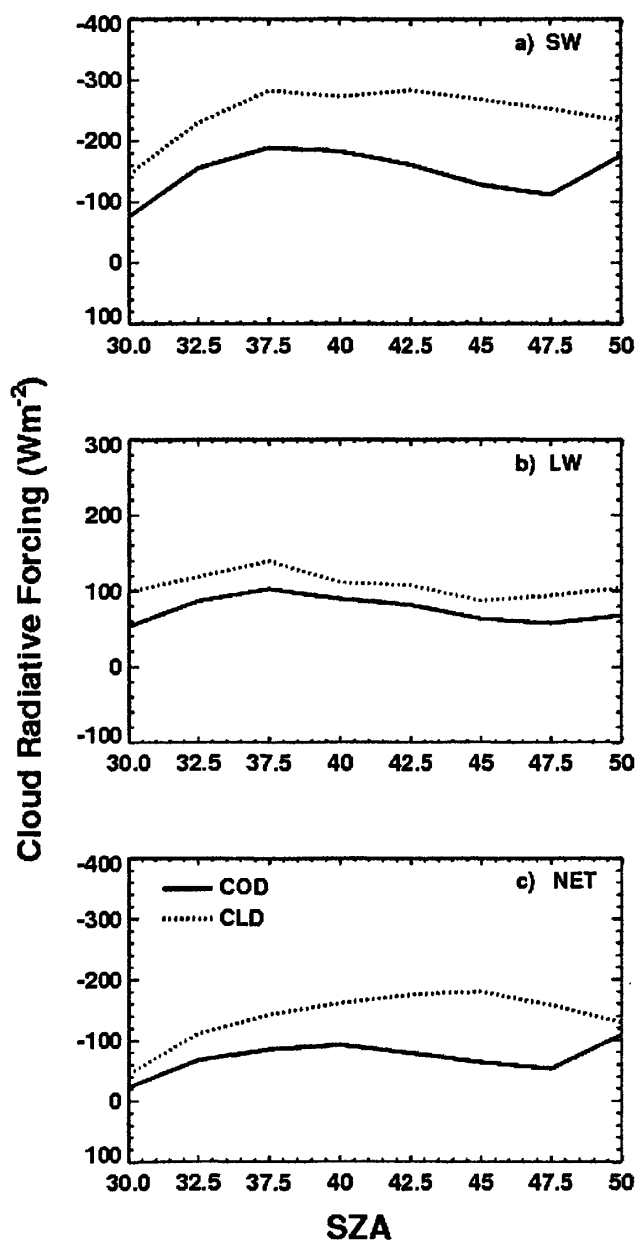


图 5.2 TOA 处 COD 和 CLD 区云短波、长波和净辐射强迫随太阳高度角变化的对比

图 5.2 和 5.3 分别对比了 COD 区和 CLD 区大气层顶瞬时短波、长波和净辐射随太阳高度角和云光学厚度的变化。从图 5.2 及图 5.3 的 a 和 c 图可以看出, COD 区云短波和净辐射强迫绝对值显著小于 CLD 区的值。但是, 图 5.2 和 5.3 的 b

图显示，对于长波来说，COD 和 CLD 区的云辐射强迫值并没有明显差别，特别是从 5.3b 图中可以看出，当云的光学厚度小于 35 时，COD 和 CLD 区的云

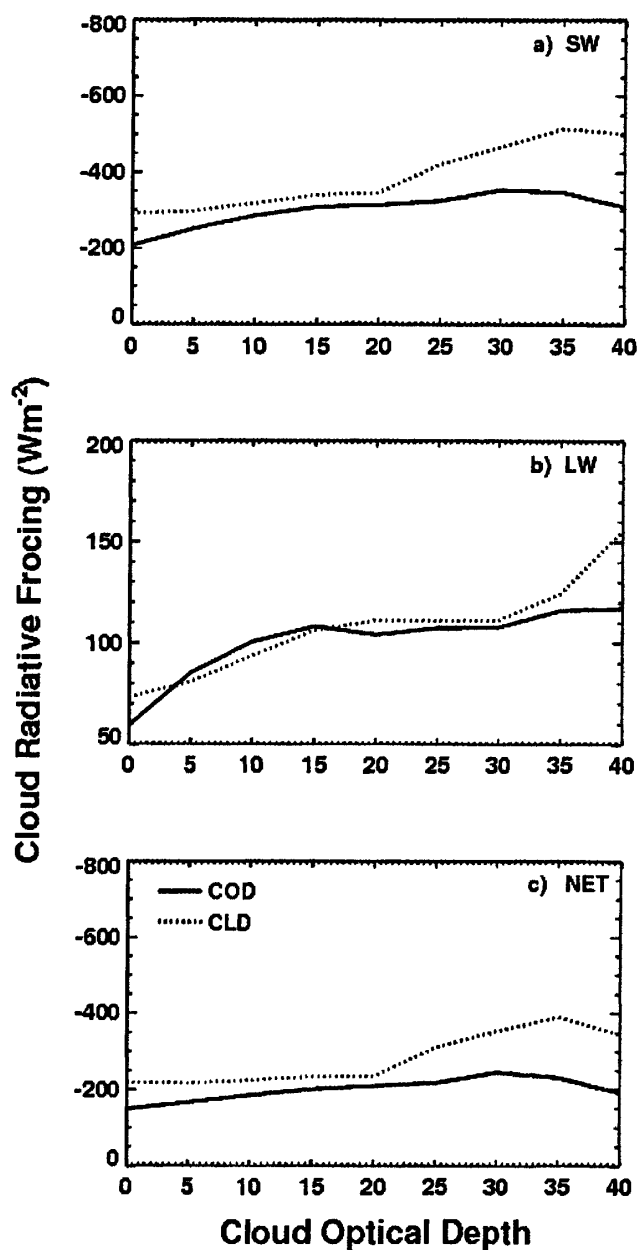


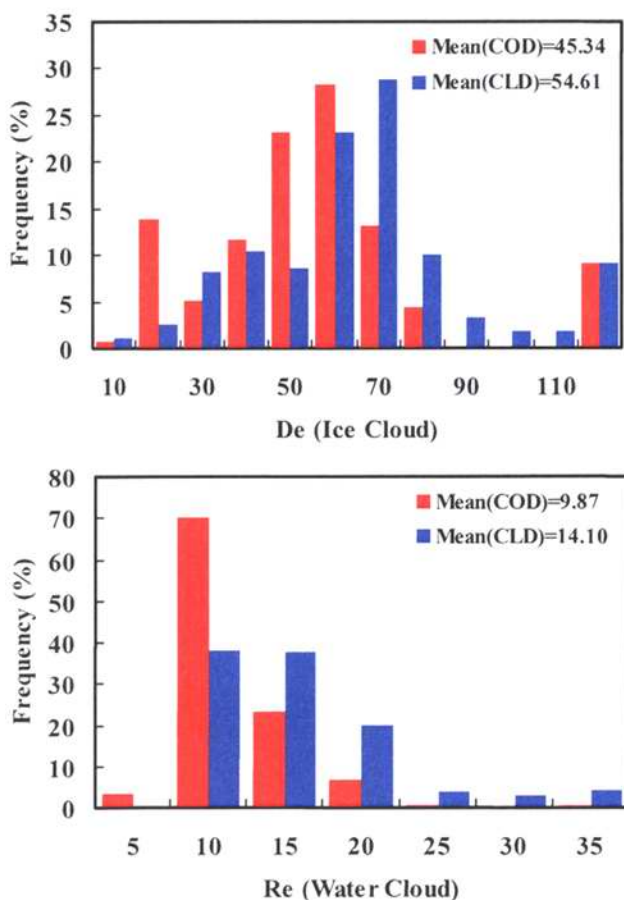
图 5.3 TOA 处 COD 和 CLD 区云短波、长波和净辐射强迫随云光学厚度变化的对比

辐射强迫值几乎没有差别。从这两幅图中我们看出，COD 与 CLD 区的云辐射强

迫差别随云光学厚度的变化大于随太阳高度角的变化。从上述的分析,结果清晰的表明当云层上部有沙尘存在时,沙尘气溶胶会显著影响云辐射强迫值。这种影响作用包括沙尘粒子的直接作用、间接作用以及半直接作用。

5.4 沙尘对云微物理特性的影响

图 5.4 是 COD 和 CLD 区冰云有效粒子直径 (De)、水云有效粒子半径 (Re) 和云光学厚度对比的频率谱分布。COD 区的 De 和 Re 的平均值分别小于 CLD 区值的 17.0%和 30.0%。COD 区 De 和 Re 的显著减少是因为在有沙尘的 COD 区,沙尘粒子增加了云滴的凝结核,从而抑制了云滴的生长,所以使得 COD 区的 De 和 Re 值相比较 CLD 区有了明显的减少。



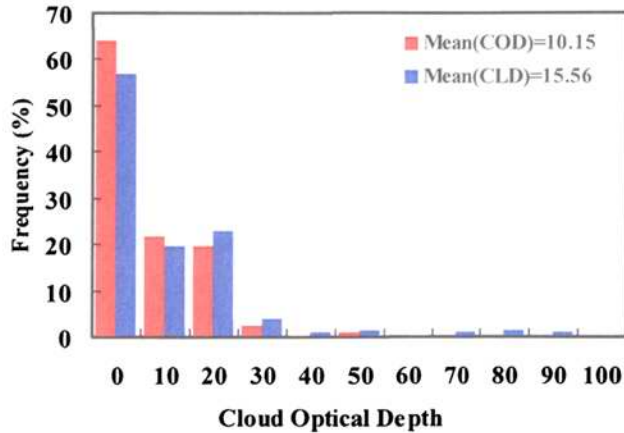


图 5.4 COD 和 CLD 区冰云有效直径、水云有效半径和云光学厚度比较频谱图

De 或是 Re 的减小会通过气溶胶第一和第二间接效应增大云的光学厚度和云的反照率 (Twomey, 1977; Albrech, 1989)。这里需要注意, 云光学厚度是与冰水路径/云水路径成正比的, 而与云滴粒子大小成反比。分析 MODIS 卫星数据结果表明, COD 区的平均云光学厚度为 10.15, CLD 区的平均值为 15.56, COD 区云光学厚度较 CLD 区减少了 34%。这种光学厚度的减少也许可以用气溶胶半直接效应来解释, 是由于沙尘的存在蒸发了低层云而造成了 COD 区云光学厚度的减小 (Huang, et al., 2006b)。虽然沙尘气溶胶辐射强迫对地表辐射通量和水循环的影响也许也是 COD 区云光学厚度减少的部分原因 (Jiang 和 Feingold, 2006), 但是沙尘气溶胶存在时的辐射强迫作用对云光学厚度的减少的相对贡献作用, 依然是将来研究探索领域的重要课题。以上结果已经证明, 沙尘气溶胶可以改变云的微物理和宏观物理特性, 从而影响云的辐射强迫。

为了进一步解释沙尘的作用, 我们利用 Fu-Liou 模式来进行数值模拟计算。在模式计算中, 沙尘层被放置在 4km 高度上, 从此高度到地面, 随着高度以 e 指数分布降低。大气垂直的温湿廓线, 已经各种气体浓度, 采用了 McClatchey 等人的中纬度夏季标准廓线 (McClatchey et al., 1971)。在模拟计算中, 云层高度为 3.5-5.5km, 下垫面地表类型为沙漠地表。

沙尘气溶胶的直接效应主要包括对入射太阳光的散射和吸收。因为沙尘粒子是一种吸收性气体, 它对大气层顶和地表的短波和长波辐射通量的影响, 表现出与其他气溶胶不同的特性。从图 5.5 的模式模拟结果可以看出, 随着沙尘气溶胶

光学厚度的增大,短波和净辐射强迫在大气层顶处逐渐减小,而在地表处则显著增大。这种差别表明,由于沙尘气溶胶的存在,减弱了云在大气层顶的冷却效应;但是增强了云在地表的冷却效应。而云冷却效应的增强/减弱事实上是由沙尘粒子的直接效应的冷却/增暖作用造成的。从图 5.5 中我们可以清楚的得到,当气溶胶光学厚度增大 0.1 时,根据不同的太阳高度角,在大气层顶处,短波瞬时云辐射强迫值的增幅范围是: $2.7 \text{ w/m}^2 \sim 15.7 \text{ w/m}^2$; 在地表,为 $-1.7 \text{ w/m}^2 \sim -29.7 \text{ w/m}^2$ 。当沙尘气溶胶存在于云层之上时,短波和净辐射增暖作用是因为沙尘气溶胶粒子的吸收性;同时无论沙尘粒子的吸收和散射都会阻止太阳能量到达地表,从而造成地表冷却作用的增强。对于长波来说,COD 区的云辐射强迫值一直是大于 CLD 区的,这主要是由于沙尘气溶胶的温室效应所造成的,但我们也发现长波的辐射强迫增加值很小,这是因为长波对沙尘粒子并不是相对的非常敏感。而 TOA 处沙尘的长波辐射作用小于地面的原因是在 TOA 处沙尘层上的云会完全掩盖沙尘的任何作用。

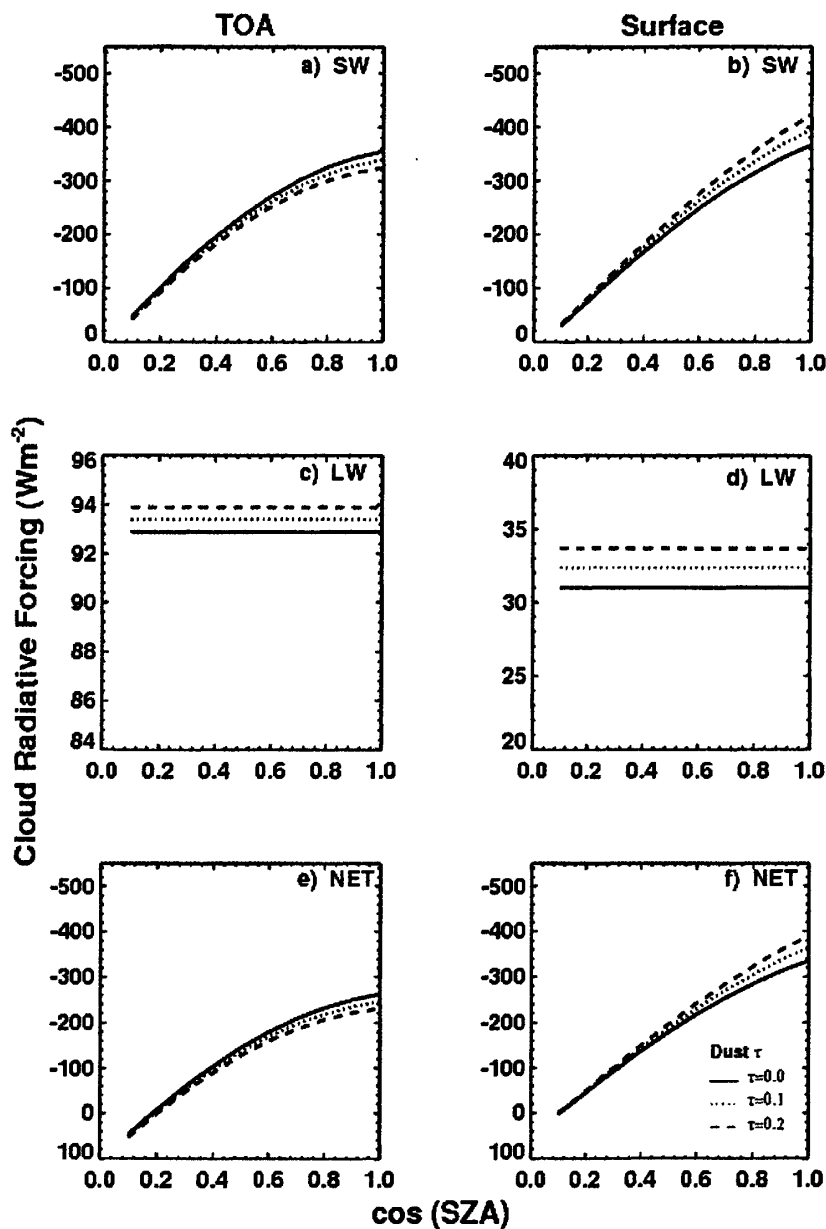


图 5.5 模式模拟有沙尘情况下云的瞬时辐射强迫值随太阳高度角的变化: a) TOA 处短波的辐射强迫; b) 地表处的短波辐射强迫; c) TOA 处的长波辐射强迫; d) 地表处的长波辐射强迫; e) TOA 处的净辐射强迫; f) 地表处的净辐射强迫。模式试验时云的光学厚度为 16, 沙尘类型为粗糙型, 下垫面为沙漠。

图 5.6 给出了云在 TOA 和地表处瞬时短波、长波和净辐射强迫随云光学厚度的变化。当云的光学厚度不断变化时, 沙尘在 TOA 处一直为增暖作用, 而在

地表则一直为冷却作用。以上模式模拟结果表明, 沙尘作为一种吸收性气溶胶, 其在 TOA 处的直接效应为增暖作用。

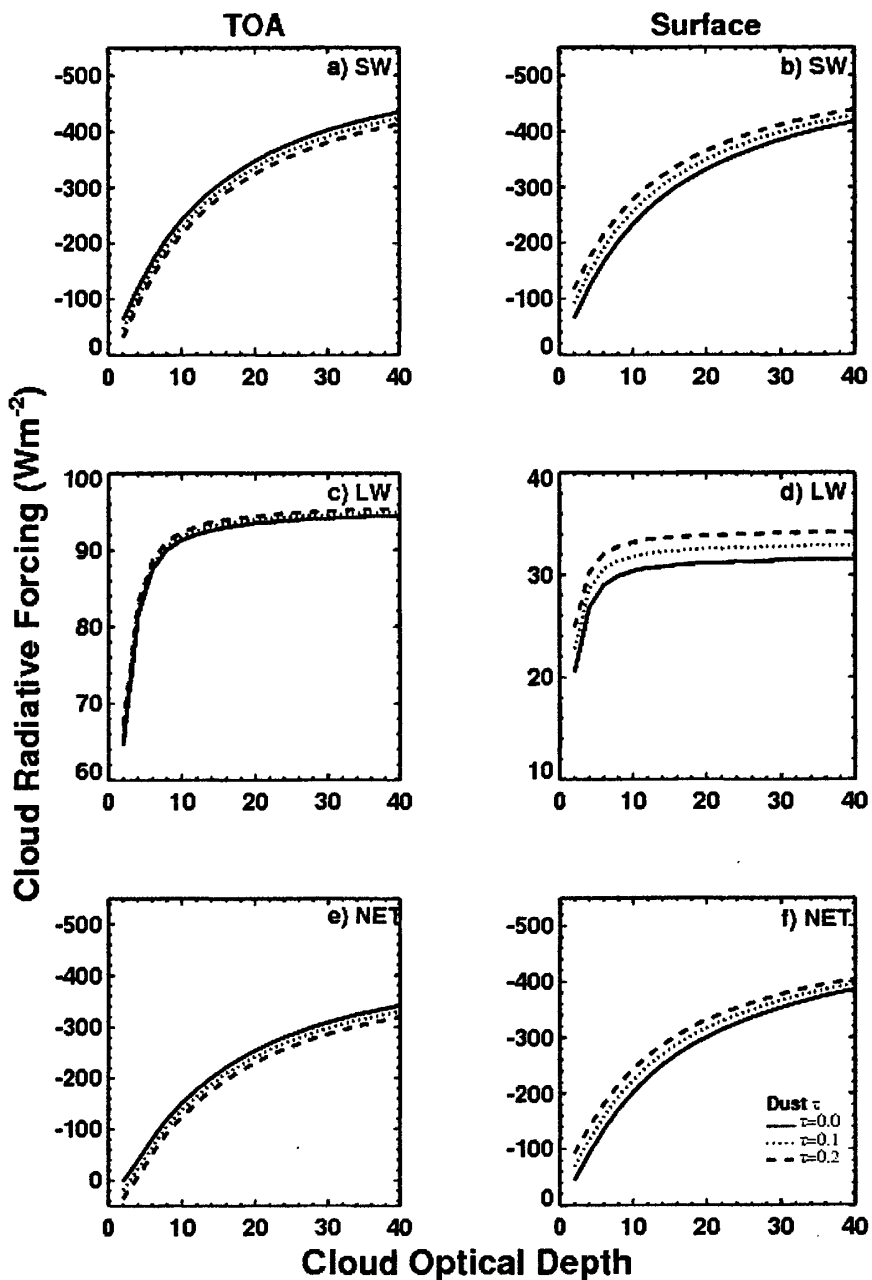


图 5.6 与图 5.5 一致, 不同的只是云辐射强迫随云光学厚度的变化

5.5 沙尘气溶胶直接辐射效应与间接效应和半直接辐射效应的区分

在 5.3 节中我们通过分析卫星观测数据，沙尘气溶胶在 TOA 处总的辐射效应为正值，即增暖作用。此处的总的辐射效应（也就是 COD 与 CLD 辐射强迫的差值）即包括了沙尘的直接辐射效应，也包括了通过改变云的微物理特性和宏观特性而作用的间接辐射效应和半直接辐射效应。但是由卫星观测数据分析得到的这三种影响效应是混合在一起的，我们很难仅通过卫星观测数据来将沙尘的直接效应和通过云产生的间接和半直接效应分离开来。为了探索解决这一问题，我们提出了一个简单的通过结合卫星观测数据和模式模拟结果来分离这两种效应的方法。

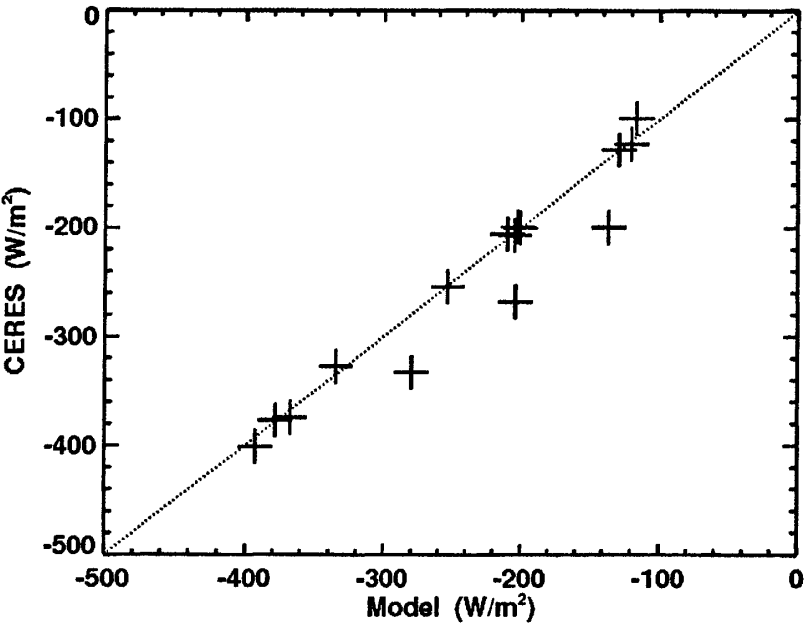


图 5.7 CLD 区 CERES 卫星观测的瞬时短波辐射通量与 Fu-Liou 模式模拟结果的对比散点图

通过图 5.7 的对比，我们可以看出 CERES 的卫星观测资料与 Fu-Liou 模式模拟结果显示出很好的一致性。这表明利用 Fu-Liou 辐射传输模式计算得到的 CLD 区 TOA 处的辐射通量是合理可信的。该结果表明通过计算分析卫星观测数

据和模式模拟结果的差来区分直接和间接/半直接效应的想法是一个合理、可行的方法。

估算间接和半直接效应的方法程序介绍如下：

1. 通过卫星反演资料得到 COD 区和 CLD 区云的微物理特性参数，如云光学厚度、水云有效粒子半径、冰云有效粒子直径以及云顶气压等。
2. 将第一步得到的 COD 和 CLD 区的云微物理特性参数，作为输入量输入 Fu-Liou 模式，分别计算两个区域的云辐射强迫。(此处要注意：尽管 COD 区有沙尘的存在，但我们在利用模式计算 COD 和 CLD 区域的云辐射强迫时，沙尘的光学厚度都设为 0。因为在 COD 区沙尘对辐射的影响已经通过改变云的微物理特性显现了出来)。
3. 沙尘气溶胶总的辐射强迫效应(包括直接、间接和半直接辐射效应)，可以通过下面的式子来计算：

$$ARF_{total} = F_{CERES}(COD) - F_{CERES}(CLD)$$

其中， $F_{CERES}(COD)$ 和 $F_{CERES}(CLD)$ 分别是由 CERES 卫星观测资料得到的 COD 和 CLD 区 TOA 处的辐射强迫值。

4. 沙尘通过改变的云的性质来影响辐射平衡的间接和半直接效应可以由下式来计算：

$$ARF_{indirect+semi} = F_{model}(COD) - F_{model}(CLD)$$

其中， $F_{model}(COD)$ 和 $F_{model}(CLD)$ 是由第二步得到的模式计算的 COD 与 CLD 区域的辐射强迫值

5. 通过第三步得到沙尘总的辐射强迫效应以及由第四步得到的间接和半直接效应后，我们可以很容易的得到沙尘的直接辐射强迫效应：

$$ARF_{direct} = ARF_{total} - ARF_{indirect+semi}$$

也就是说，沙尘的直接辐射强迫就是在晴空条件下沙尘总的辐射强迫值。

表 5.2 给出由挑选个例计算得到的 TOA 处，沙尘气溶胶的平均短波瞬时直接辐射强迫值以及间接和半直接综合辐射强迫值。沙尘气溶胶在 TOA 处的直接辐射强迫值为 22.7 Wm^{-2} ，占总辐射强迫效应的 21.6%。沙尘直接辐射强迫效应

的计算是由卫星观测数据得到的总的辐射强迫值与通过模式计算得到的间接和

Table 5.2 由挑选个例计算得到的 TOA 处, 沙尘气溶胶的平均短波瞬时直接辐射强迫值以及间接和半直接综合辐射强迫值及所占总辐射效应的百分比

	Mean SWRF Wm^{-2}	Standard Deviation	Mean Percentage (direct / total)	Standard Deviation
Direct effect	22.7	44.45	19.21%	9.80%
Indirect effect	82.2	41.15		

半直接综合辐射强迫值, 相减得到的。由此得到的 22.7 Wm^{-2} 与图 5.6 和 5.7 中模式计算结果的范围一直 (在图 5.6 和 5.7 中沙尘的直接辐射强迫值就是实线与虚线, 或是虚线与点线之间的差)。图 5.8 是计算得到的 16 个例子在 TOA 处直接辐射强迫值及间接和半直接综合辐射强迫值, 以及直接辐射强迫效应占总辐射强迫值的百分比。从表 5.2 和图 5.8 中, 我们可以非常清晰的看出, 沙尘的间接和半直接的综合辐射效应是沙尘辐射效应的主要作用。

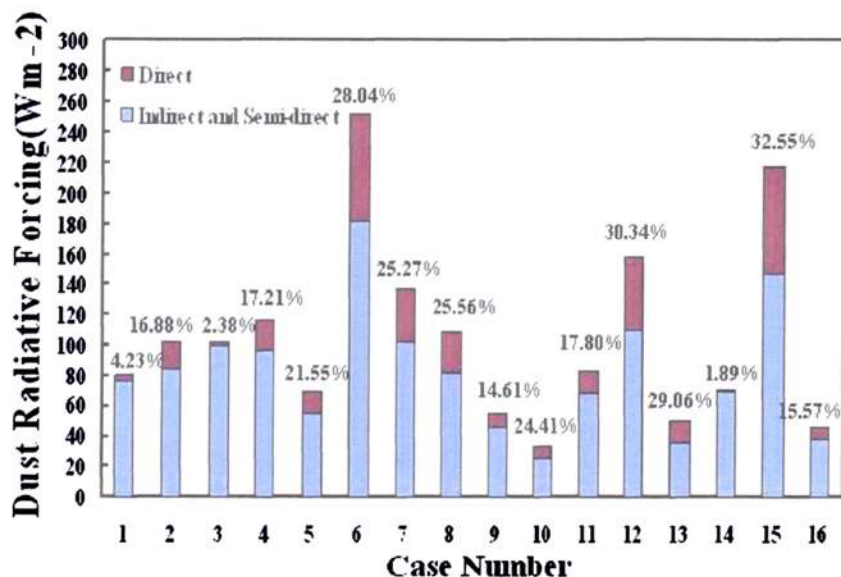


图 5.8 计算得到的 16 个例子在 TOA 处短波瞬时直接辐射强迫值及间接和半直接综合短波瞬时辐射强迫值

5.6 小结

东亚沙尘气溶胶通过直接辐射效应和改变云微物理和宏观特性的间接以及半直接辐射效应,对辐射能量平衡产生显著影响。以前的许多工作都是以研究晴空条件下的沙尘辐射强迫(即直接辐射强迫)为主。近年来有研究发现,沙尘气溶胶间接的通过改变云的特性对辐射平衡产生的影响,在气候变化研究中也扮演着重要角色(Sassen et al., 2003)。我们通过分析卫星数据发现,在大气层顶处有沙尘的 COD 区瞬时云辐射强迫绝对值明显小于无沙尘的 CLD 值。这表明,由于沙尘粒子的吸收性,当云上有沙尘存在时会对 TOA 处产生显著的增暖作用。我们通过对 MODIS 卫星资料的分析,已经可以清楚的证明,沙尘的存在会显著改变云的微物理特性,如:水云有效粒子半径、冰云有效粒子直径以及云的光学厚度。这些参数量在 COD 区的值明显小于 CLD 区的值。这也与以前的研究工作的结果一致(Huang et al., 2006a)。另外利用 Fu-Liou 辐射传输模式的计算,也证实了沙尘在 COD 区 TOA 处的增暖作用。

虽然我们可以通过卫星云微物理特征参数的卫星反演资料,清晰的分析证明得到沙尘粒子会对云的物理特性产生影响,由此对辐射能量平衡产生间接/半直接效应,但是仅仅利用卫星资料还是很难将沙尘的直接辐射强迫与间接/半直接辐射强迫作用区分开来。本章就是通过将卫星观测数据与模式模拟结果相结合,提出了一种简单的量化计算直接辐射强迫值和间接/半直接综合辐射强迫值的新方法。我们通过分析 2002-2006 四年挑选的 16 个例子,结合卫星观测和反演数据和 Fu-Liou 辐射传输模式模拟计算结果,发现沙尘间接/半直接综合辐射效应是沙尘气溶胶的主要辐射作用。16 个例子 TOA 处平均瞬时短波辐射强迫值为 82.2 Wm^{-2} , 占总的辐射强迫值的 78.4%。由卫星观测反演的辐射通量值和云微物理特性参数值计算得到的沙尘的直接辐射强迫值为 22.7 Wm^{-2} , 此结果在模式模拟计算的沙尘直接辐射强迫值的范围内。

虽然本章内容只是研究有云条件下东亚沙尘辐射作用的一个初步工作,但是我们提出了一个可以区分沙尘直接辐射效应和通过改变云产生的间接/半直接辐射效应的方法,对深入研究沙尘与云的相互关系有一定的帮助意义。在以后的工

作中,利用更多其他方法观测的数据资料会对量化计算间接和半直接辐射强迫值更有帮助。当然我们的方法还存在一定的不确定性,希望以后可以通过更多的例子做进一步的分析、验证。

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第六章 论文总结与未来工作展望

6.1 论文总结

气溶胶作为影响气候变化的众多因子之一,由于其在气候变化中作用的重要性和影响机制的复杂性,使得近年来对于气溶胶的研究受到高度重视,成为这个研究领域的热点。气溶胶不仅可以直接通过散射和吸收太阳辐射,改变大气层顶、地表以及大气内的辐射平衡;也可以作为凝结核,通过改变云量和云的微物理特性,从而间接的影响地气辐射能量平衡。我国西北地区是沙尘气溶胶的多发区,沙尘是大气气溶胶的主要种类之一,由于其对太阳辐射能量强烈的吸收作用,不仅会改变地表能量平衡,而且还会通过改变大气的辐射加热结构,加速大气低层水汽和云的蒸发,从而抑制和减弱局域降水,这也许对西北地区的干旱有加剧作用。所以定量计算研究西北半干旱地区气溶胶的辐射强迫作用,对了解气溶胶对这一区域气候变化和水分循环的作用,特别是对西北地区干旱的影响,有着重要意义。

本论文充分利用 Fu-Liou 辐射传输模式, CERES、MODIS、CALIPSO 等卫星遥感资料及地面观测资料,并借助于数值模式,首先从理论上对影响沙尘辐射强迫的各个参量做了敏感性测试,然后着重详细研究沙尘气溶胶的直接辐射强迫效应,精确定量计算了沙尘的直接辐射强迫值及其对大气辐射加热率的影响,进而初步研究了沙尘与云的相互作用对辐射能量平衡的影响,并在此基础上提出了一个将沙尘气溶胶直接辐射强迫效应与间接/半直接综合辐射强迫效应进行区分的方法。论文得到的主要结果如下:

(1) 各参数对沙尘辐射强迫计算影响的敏感性试验

利用模式计算气溶胶在 TOA 及地表的辐射强迫时,由于目前对沙尘气溶胶粒子的分布和物理、化学性质和地表性质等都还不是十分清楚,所以计算得到的气溶胶辐射强迫值有着很大的不确定性。为了定量了解这些关键模式输入量的不确定性对于计算沙尘直接辐射强迫值的影响,我们对各关键影响因子做了敏感性试验。气溶胶粒子与太阳辐射的相互作用主要由以下几个因子综合决定: 1. 气溶胶粒子光学性质(光学厚度、单次散射反照率、不

对称因子、埃斯屈朗指数)。2. 地表性质(地表反照率)。3. 大气廓线(水汽、臭氧), 所以我们对括号中所列的参数对辐射能量平衡的影响做了敏感性测试。结果表明, 在所有影响因子中, 沙尘粒子的单次散射反照率对辐射通量的影响最为显著, 其次沙尘的光学厚度以及下垫面的反照率都会对辐射通量产生较大的影响。所以在定量计算沙尘气溶胶的直接辐射强迫时, 沙尘粒子单次散射率和地表反照率的确定, 对结果的准确性至关重要, 但是目前这两个量却是所有影响因子中最难测量, 不确定性最大的两个因子。

(2) 沙尘气溶胶直接辐射强迫作用以及对大气辐射加热率影响的研究

2006 年 4 月发射的 CALIPSO 卫星上装载的激光雷达, 为我们提供了云和气溶胶的垂直分布廓线。这使我们得到一个前所未有的机会去细致了解云及气溶胶的辐射强迫及反馈相互作用。我们利用 CALIPSO 数据得到了塔克拉玛干地区的沙尘垂直分布廓线并结合 CERES 卫星资料, 计算了塔克拉玛干沙漠地区沙尘对大气层顶(TOA)和地表的辐射能量平衡的影响以及大气辐射加热率。开展的工作主要是: 首先挑选了 2007 年夏季塔克拉玛干沙漠地区一次强的沙尘过程, 利用 CALIPSO 卫星后向散射数据计算得到沙尘气溶胶的垂直分布廓线, 将其输入 Fu-Liou 辐射传输模式中计算 TOA 辐射值, 通过与 CERES 卫星观测值对比并调整模式中沙尘单次散射反照率性质来确定这一地区沙尘的光学特性。在确定沙尘的类型后, 进一步利用模式计算沙尘在 TOA 和地面的辐射强迫值以及大气辐射加热率。从结果可以得到 TOA, 地面和大气的日平均净辐射值分别为 44.41 , -41.89 , 和 86.31 Wm^{-2} 。由于沙尘的强吸收性使得很大一部分太阳辐射能量留在大气中, 从而改变大气的垂直结构。随着沙尘过程的加强发展, 沙尘层大气的加热率可以从 1K day^{-1} , 2K day^{-1} 曾加到 3K day^{-1} , 甚至沙尘最强时可以达到 5.5K day^{-1} 。这是首次量化的计算了在塔克拉玛干沙漠中强沙尘暴发生时, 沙尘层对大气的加热作用。

(3) 沙尘气溶胶与云相互作用的研究

气溶胶与云作为诸多影响气候变化因子中最不确定的两个因子, 其相互作用的研究一直是这个领域中研究的热点和难点。我们首先利用 CERES 卫星资料证实了沙尘气溶胶粒子对云微物理特性的影响, 并将资料分析结果与模

式模拟结果相结合,提出了一种新的研究沙尘气溶胶间接作用的方法。

(A) 沙尘气溶胶对云微物理特性的改变

开展的工作主要是利用四年的 CERES 卫星资料(2003—2006),分析比较了纯云区(CLD)和被沙尘污染过的云区(COD)云微物理特性的变化。结果表明因为沙尘粒子增加了云的凝结核从而使得COD区冰云的粒子直径和水云粒子半径比CLD 区分别减小17.0%和30.0%。这说沙尘粒子进入大气后会显著的改变云的微物理特性,从而进一步影响云的辐射强迫。然后我们分析了当云下有沙尘出现时,沙尘对云辐射强迫的影响。研究结果表明COD区,其短波、长波和净辐射强迫值比CLD区分别减少57.0%, 74.2% 和 46.3%。当云下有沙尘出现时,沙尘会明显的减弱云的冷却效应,云冷却效应的减弱可认为是由于沙尘增暖作用抵消的结果。为了进一步证明沙尘的增暖作用,我们利用Fu-Liou 模式对沙尘的直接作用进行数值模拟。结果表明随着沙尘光学厚度的增加,在TOA处云的短波及净辐射强迫值减小,而在地面则是增大的。这种差异说明沙尘在TOA上减弱了云的冷却效应而同时在地面增大了云的冷却效应。

(B) 提出一种新的研究沙尘气溶胶间接作用的方法

卫星资料分析的结果表明在 TOA 处沙尘气溶胶的辐射效应是增暖的,这其中包括直接效应以及由云微物理特性改变所引起的间接和半直接效应。但是我们无法只利用卫星资料将直接效应和其他两种效应分开。所以我们提出将卫星资料分析结果与模式模拟结果相结合,来达到这一目的。利用此种方法分析结果表明沙尘短波直接辐射强迫值为 22.7 Wm^{-2} , 间接与半直接辐射强迫和为 82.2 Wm^{-2} , 分别占总辐射效应的 21.6 %和 78.4%, 这说明间接和半直接效应是沙尘气溶胶的主要作用。

6.2 未来工作展望

由于缺少详细准确的西北地区气溶胶特性的数据库以及陆面反照率特征,我们在研究西北地区沙尘气溶胶辐射强迫作用时都为测站单点计算值,区域尺度上的辐射强迫作用的研究工作进展不大,存在着很大的不确定性。近些年来,为了解决这一问题,已经有许多工作用来扩充改进观测数据库,这包括:增加地基联

网观测站, 比如 Aerosol Robotic Network (AERONET); 改进和发展新的卫星探测器, 比如 MISR, CALIPOSE; 以及增加各区域的野外观测。目前在空间遥感探测方面, 美国 NASA 的 A-Train 计划已经拥有了 6 颗极轨卫星, 它提供了一个空前的机会来改进气溶胶光学参数认识; 在地基观测方面, 目前亚洲地区已经有 18 个 AERONET 联网观测点 (Che et al., 2005), 而且也进行了一系列的大型野外加强观测试验, 如亚洲气溶胶特性试验 (ACE-Asia) (Huebert et al., 2003), 亚洲大气环境气溶胶粒子试验 (APEX) (Nakajima, T., et al., 2003) 以及印度洋观测试验 (INDOEX) 等 (Ramanathan et al., 2001b)。虽然这些观测试验取得了很大的进展, 但是也有其局限性, 尤其是缺乏亚洲陆地内部的观测数据, 特别是中国西北内陆地区, 这样具有独特地理环境的气溶胶特性数据。2008 年兰州大学大气科学学院、中科院大气物理所、中国气象局、美国能源部和马里兰大学等单位联合举行的西北地区野外观测试验, 以兰州大学半干旱气候与环境观测站为中心, 另设景泰、张掖国家气候观测台两个移动观测点, 历时三个多月, 获得了我国西北地区气溶胶粒子特性的第一手资料。Ge 等人利用张掖站点的观测数据反演了当地气溶胶的光学特性, 结果表明西北地区气溶胶特别是沙尘粒子与其他地区的粒子的光学特性相比有明显差别, 具有强烈的吸收特性, $0.5\mu\text{m}$ 波段上粒子的单次散射反照率仅为 0.75, 将地面反演结果与多个卫星反演数据进行对比发现, MODIS 反演的这一区域的光学厚度值明显偏大 (Ge et al., 2010), 这说明 MODIS 传统的反演方法并不适合西北干旱半干旱地区亮地表条件下气溶胶的反演。所以如果利用一种气溶胶反演误差很大的卫星数据来计算区域尺度上辐射强迫, 是没有意义的。

在未来工作中我们想利用兰州大学半干旱气候变化教育部重点实验室观测站的地面长期观测资料以及野外加强观测试验数据, 反演西北地区粒子的光学特性, 并与各种卫星反演资料进行对比验证, 并借助于数值模式和适合西北地区气溶胶反演的卫星资料, 计算分析西北地区气溶胶的长期时空分布以及其在大气层顶和地表的辐射强迫, 深入研究气溶胶对西北地区区域能量平衡的作用以及对西北地区干旱的影响。

具体的实施想法如下:

[1] 利用地面观测数据反演气溶胶的光学特性

利用兰州大学半干旱气候变化教育部重点实验室观测平台（简称兰大气候站）长期观测以及野外加强试验观测的多滤波旋转遮光辐射仪（MFRSR）资料反演气溶胶光学厚度、粒子谱分布以及各散射特性参数，并与AERONET联网反演数据进行对比，得到西北地区长期可靠的气溶胶特性地基反演数据库。

[2] 利用地基反演数据验证评估卫星反演数据

将利用[1]得到的气溶胶光学厚度的地基反演数据，与MODIS卫星传统方法反演数据、MODIS卫星Deep Blue方法反演数据、MISR卫星反演数据和CALIPSO卫星反演数据进行验证对比分析，评估各种卫星反演方法在这一地区的可信度，并对反演误差相对较小的卫星资料进行订正，以期得到能够合理代表西北区域气溶胶光学厚度的卫星反演数据。

[3] 计算分析西北地区气溶胶长时间序列的时空分布

利用[2]得到整个区域气溶胶的合理卫星反演数据，计算分析整个西北地区气溶胶长时间序列的区域尺度上的时空分布。

[4] 西北地区区域尺度上辐射强迫的研究

利用[1], [2], [3]结果以及数值模式计算西北气溶胶在区域尺度上的辐射强迫。首先，为了提高数值模拟计算的精确度，验证输入参数的合理性，将站点地表辐射通量收支观测数据与模式模拟结果进行对比，进行辐射闭合试验。如果试验成功，则计算西北地区到达地表太阳辐射通量的变化以及区域尺度上气溶胶在大气层顶、地表和大气层中的短波、长波及净辐射强迫值。定量研究西北半干旱区域气溶胶对此区域能量平衡的影响以及对区域气候的作用。

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苏婧

2010 年 5 月 19 日

附录 A

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附录 B

在研期间论文发表情况

发表论文情况

1. Su, J., J. Huang, Q. Fu, P. Minnis, J. Ge, and J. Bi, 2008: Estimation of Asian dust aerosol effect on cloud radiation forcing using Fu-Liou radiative model and CERES measurements, *Atmos. Chem. Phys.* **8**, 2763-2771.
2. Huang J., Q. Fu, J. Su, Q. Tang, P. Minnis, Y. Hu, Y. Yi, and Q. Zhao, 2009: Taklimakan dust aerosol radiative heating derived from CALIPSO observations using the Fu-Liou radiation model with CERES constraints, *Atmos. Chem. Phys.*, **9**, 4011–4021.
3. 葛颢铭.刘玉芝.黄建平.苏婧.李积明.闭建荣, 2009: 利用 HITRAN 资料研究黑碳和沙尘气溶胶粒子的光学特性, *应用光学*, 30(2).
4. Fu Q., T. J. Thorsen, J. Su, J. Ge, and J. Huang, 2009: Test of Mie-based single-scattering properties of non-spherical dust aerosols in radiative flux calculations, *Journal of Quantitative Spectroscopy & Radiative Transfer*, doi:10.1016/j.jqsrt.2009.03.010
5. Ge J.M., J. Su, T.P. Ackerman, Q. Fu, J.P. Huang, and J.S. Shi, 2009: Dust Aerosol Optical Properties Retrieval and Radiative Forcing over Northwestern China during the 2008 China-US Joint Field Experiment. *J. Geophys. Res.*, (in press).

参加会议情况:

- [1] 苏婧, 2010 年 4 月 28 日. 2008 年中美联合观测期间西北地区沙尘气溶胶的光学与辐射强迫的研究. 半干旱气候变化教育部重点实验室 2009 年度学术年会, 兰州, 甘肃, 中国, 2010 年 4 月 27-28.
- [2] Ge J.M., J.Su, T.P. Ackerman, Q. Fu, J.P. Huang, J.S. Shi., Dec 14, 2009. Dust Aerosol Optical Properties Retrieval and Radiative Forcing over Northwestern China during the 2008 China-US Joint Field Experiment. 2009 AGU fall meeting, San Francisco, California, USA, December 14-18.
- [3] J.Su., Jul 4, 2009. 2009 Gordon Research Conferences (GRC): Radiation and Climate, New London, New Hampshire, USA, July 4- 10.

附录 C

已发表的部分论文

- [1] **Su, J.**, J. Huang, Q. Fu, P. Minnis, J. Ge, and J. Bi, 2008: Estimation of Asian dust aerosol effect on cloud radiation forcing using Fu-Liou radiative model and CERES measurements, *Atmos. Chem. Phys.* **8**, 2763-2771.
- [2] Huang J., Q. Fu, **J. Su**, Q. Tang, P. Minnis, Y. Hu, Y. Yi, and Q. Zhao, 2009: Taklimakan dust aerosol radiative heating derived from CALIPSO observations using the Fu-Liou radiation model with CERES constraints, *Atmos. Chem. Phys.*, **9**, 4011-4021.

Estimation of Asian dust aerosol effect on cloud radiation forcing using Fu-Liou radiative model and CERES measurements

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Abstract. The impact of Asian dust on cloud radiative forcing during 2003–2006 is studied by using the Clouds and Earth's Radiant Energy Budget Scanner (CERES) data and the Fu-Liou radiative transfer model. Analysis of satellite data shows that the dust aerosol significantly reduced the cloud cooling effect at TOA. In dust contaminated cloudy regions, the 4-year mean values of the instantaneous shortwave, longwave and net cloud radiative forcing are -138.9 , 69.1 , and -69.7 Wm^{-2} , which are 57.0 , 74.2 , and 46.3% , respectively, of the corresponding values in pristine cloudy regions. The satellite-retrieved cloud properties are significantly different in the dusty regions and can influence the radiative forcing indirectly. The contributions to the cloud radiation forcing by the dust direct, indirect and semi-direct effects are estimated using combined satellite observations and Fu-Liou model simulation. The 4-year mean value of combination of dust indirect and semi-direct shortwave radiative forcing (SWRF) is 82.2 Wm^{-2} , which is 78.4% of the total dust effect. The dust direct effect is only 22.7 Wm^{-2} , which is 21.6% of the total effect. Because both first and second indirect effects enhance cloud cooling, the aerosol-induced cloud warming is mainly the result of the semi-direct effect of dust.

albedo and suppress precipitation by modifying cloud microphysical properties, phenomena that are defined as the aerosol first and second indirect radiative effects, respectively (Twomey, 1977; Albrecht, 1989). Aerosols can also evaporate low-level clouds through so-called aerosol semi-direct effect (Grassl, 1975). Many observational and modeling studies have focused on the aerosol “direct” radiative effect, but recent research indicates that small changes in cloud microphysical properties (e.g., droplet size, phase) could induce large changes in cloud radiative forcing. Therefore, understanding and quantifying both the direct and indirect radiative effects are critically important in climate research (IPCC, 2007).

Dust storms that originate in Taklamakan Desert of China and the Gobi Desert of Mongolia occur frequently in recent years and become one of important aerosol types in East Asia (Huang et al., 2006a). The dust layers associated with these storms often travel thousands of kilometers at high altitudes moving from the continent to the open sea near Korea and Japan (Haywood et al., 1999; Higurashi and Nakajima, 2002; Takemura et al., 2002). Moreover, Asian dust particles, which could lead to both cooling by reflecting sunlight back to space and warming by absorbing solar and infrared radiation, make the radiative impact of Asian aerosols more complicated (Sokolik and Toon, 1999). Thus, it is imperative to examine the aerosol impact on radiative energy budget over East Asia where the aerosol sources are presumed to be much different from those in other regions.

Although there have been numerous studies about the effect of dust aerosols on the radiative energy budget in clear skies, few have investigated the radiative effects of Asian dust in cloudy conditions. Huang et al. (2006a) studied the impacts of dust storms on cloud properties and radiative forcing (RF) using data collected by the Moderate Resolution Imaging Spectroradiometer (MODIS) and Clouds and the Earth's Radiant Energy System (CERES) instruments on the Aqua and Terra satellites. They found, on average, ice

1 Introduction

It is well known that aerosols have a profound impact on the global and regional climate. Determining the radiative effects of aerosols is currently one of the most active research areas in climate study (IPCC, 2007). Aerosols can reflect and absorb the incoming solar radiation, which is called the aerosol direct radiative effect. They can increase cloud



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Table 1. Sixteen dust case images used in this study.

Case Image	Date	COD region (Lat. ° N, Lon. ° E)	CLD region (Lat. ° N, Lon. ° E)
1	2003/03/26	[44.0–45.0], [124.0–124.5]	[46.5–47.5], [123.0–124.5]
2	2003/04/09	[39.5–40.5], [78.7–79.7]	[40.5–41.5], [72.8–73.5]
3	2004/03/09	[43.8–44.8], [108.0–109.0]	[44.0–45.2], [114.8–115.9]
4	2004/03/27	[43.0–44.0], [114.5–115.5]	[46.1–47.1], [111.8–112.8]
5	2005/03/09	[44.9–45.0], [109.5–111.2]	[46.1–47.1], [113.0–114.0]
6	2005/04/06	[48.6–49.2], [118.1–118.9]	[49.3–50.3], [112.5–113.5]
7	2005/04/08	[41.5–41.7], [102.9–104.0]	[40.0–41.0], [104.5–105.5]
8	2005/04/09	[42.0–42.4], [115.3–115.8]	[41.5–42.5], [106.0–107.0]
9	2005/04/13	[45.1–45.3], [112.3–114.3]	[48.7–49.5], [115.8–117.1]
10	2005/04/20	[36.4–36.9], [116.5–117.4]	[36.5–37.5], [118.2–119.2]
11	2006/02/17	[44.0–44.7], [115.1–115.7]	[47.0–48.0], [114.0, 115.0]
12	2006/02/21	[43.8–44.5], [123.5–125.6]	[41.9–42.9], [123.2–124.2]
13	2006/04/17	[39.7–40.5], [104.9–105.6]	[42.4–43.0], [108.9–110.2]
14	2006/04/18	[41.4–42.1], [116.1–117.9]	[46.7–48.1], [120.1, 122.1]
15	2006/04/22	[42.4–42.8], [116.1–117.3]	[47.8–48.4], [116.2–117.3]
16	2006/04/28	[43.3–44.2], [104.8–105.8]	[50.0–51.3], [89.0–89.9]

cloud effective particle diameter, optical depth and ice water path of cirrus clouds under dust-polluted conditions are 11%, 32.8%, and 42% less, respectively, than those derived from ice clouds in dust-free atmospheric environments. Due to the altered cloud microphysics, the instantaneous net cloud RF changed from -161.6 W/m^2 for dust-free conditions to -118.6 W/m^2 for dust-contaminated clouds. However, the above-mentioned results were based on only 10 cases. In this study, we will extend the study of Huang et al. (2006a) by considering more cases, and further separate the dust direct effect from the aerosol indirect and semi-direct effects on cloud radiative forcing using a radiative transfer model along with satellite data analysis. By examining the influence of Asian dust aerosols on the radiative forcing both at the TOA and the surface, this study should lead to a better understanding of the interactions among dust aerosols, clouds and radiation.

2 Data and methodology

Four years (June 2003 to June 2006) of CERES Aqua Edition 1B Single Scanner Footprint (SSF; see Geier et al., 2001) data are used here. CERES SSF data sets combine CERES radiation measurements, MODIS cloud microphysical retrievals (Minnis et al., 2008¹) and ancillary meteorology fields to form a comprehensive, high-quality compilation of satellite-derived cloud, aerosol, and radiation budget infor-

mation for radiation and climate studies. There are about 140 parameters in the SSF data set. The current analysis uses the following SSF parameters including IWP, LWP, cloud top effective temperature (T_e), cloud effective droplet radius (R_e), ice crystal diameter (D_e) and optical depth, which were derived with the Visible-Infrared-Solar-infrared-Split-window Technique (VISST; see Minnis et al., 2004, 2008¹). Flux measurements from CERES are used to estimate the instantaneous RF. The CERES instruments measure broadband radiances at the top-of-the-atmosphere (TOA) in three spectral regions (0.2–5.0 μm ; 8–14 μm ; 5–100 μm) with a spatial resolution of about 20 km at nadir. These radiances are converted to TOA fluxes with angular distribution models selected according to the scene classification (Loeb et al., 2005).

To estimate the impact of Asian dust on both cloud radiative forcing and cloud microphysical properties, two regions were selected to represent the clouds in different environments in the data analysis. CLD (pure cloud) represents the area where clouds occurred in a dust-free atmosphere, while COD (cloud-over-dust) denotes the clouds are contaminated by dust. The CLD and COD regions are determined based on both MODIS images and surface meteorology station observations over China and Mongolia (Huang et al., 2006a). The surface stations report dust events in four categories: dust storm, wind-blown sand, floating dust, and no-dust. A cloudy region observed by the satellite is defined as CLD, if the surface observation is no-dust, or as COD, if the surface observation is dust storm, wind-blown sand, floating dust. Table 1 shows information for the 16 selected dust cases used in this study. These selected cases occurred during the 2003–2006 time period. Although the dust events often occur in dry regions, especially over desert region, most

¹Minnis, P., Sun-Mack, S., Young, D. F., Heck, P. W., Garber, D. P., Chen, Y., Spangenberg, D. A., Wielicki, B. A., and Geier, E. B.: Cloud property retrievals for CERES using TRMM VIRS and Terra and Aqua MODIS data, IEEE Trans. Geosci. Remote Sens., submitted, 2008.

dust events are caused by cold front system or cyclone system which can carry the cloud to dry region. For example, all dust cases studied in our paper except of the case 2 were associated with the Mongolian cyclone system. In the Mongolian cyclone system, the clouds formation often occur in warmer and moister air ahead of cold front and dust storms are generally at the edges of cold fronts, thereby the COD regions locate in the similar air condition of the ahead of cold front.

Cloud radiative forcing is a parameter that has been used extensively to study cloud-radiation interactions. Here, we use the outgoing short-wave (SW) and long-wave (LW) TOA fluxes to examine the cloud radiative forcing. The cloud radiative forcing is defined as the difference in the fluxes between the clear-sky and actual sky conditions (Charlock and Ramanathan, 1985; Hartmann et al., 1986; Ramanathan, 1987; Cess and Potter, 1987; Ramanathan et al., 1989).

The SW and LW fluxes are defined at the TOA and surface levels as

$$\begin{aligned} T_{SW} &= F_{SW}^{\text{down}} - F_{SW}^{\text{up}} \\ T_{LW} &= F_{LW}^{\text{down}} - F_{LW}^{\text{up}} \end{aligned} \quad (1)$$

and cloud radiative forcing is given as

$$\begin{aligned} C_{SW} &= T_{SW} - T_{SW}^{\text{CLR}} \\ C_{LW} &= T_{LW} - T_{LW}^{\text{CLR}} \\ C_{\text{NET}} &= C_{SW} + C_{LW} \end{aligned} \quad (2)$$

where F_{SW}^{down} (F_{LW}^{down}) and F_{SW}^{up} (F_{LW}^{up}) are downward (upward) SW and LW fluxes, respectively. The index "CLR" denotes clear-sky scenes: those without clouds and dust aerosols. Thus, the cloud radiative forcing derived in this study for COD conditions includes the radiative effects of both clouds and dust aerosols.

3 Radiation transfer model

The radiation transfer model originally developed by Fu and Liou (1992, 1993) and modified by Rose and Charlock (2002) is used to elucidate the observation results. It is a delta-four stream radiative transfer code with fifteen spectral bands from 0.175 to 4.0 μm in the SW and twelve LW spectral bands between 2850 and 0 cm^{-1} . The correlated k -distribution method is used to treat the non-gray gaseous absorption due to H_2O , CO_2 , O_3 , N_2O , and CH_4 (Fu and Liou, 1992). The surface albedo spectral dependencies are taken into account using a lookup table based on IGBP scene type. The radiation model allows for the input of multiple wavelength aerosol optical depth measurements while the aerosol wavelength-dependent single scatter albedo and asymmetry parameter are determined by choosing an assumed aerosol type or mixture of constituents.

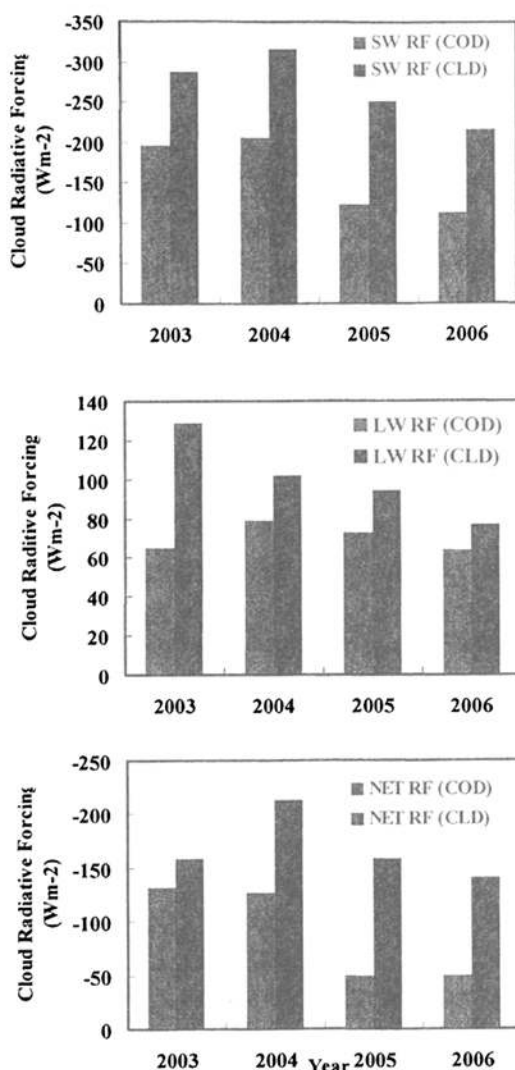


Fig. 1. Annual mean instantaneous TOA dusty (COD) and dust-free (CLD) cloud radiative forcing for (a) SW forcing, (b) LW forcing and (c) Net forcing.

4 Result analysis

Figure 1 shows the histograms of the annual mean instantaneous TOA CRF for dust-free cloud (CLD) and dusty cloud (COD) regions for 2003–2006. In each of these 4 years, the (absolute) values of SW, LW and Net forcing in COD region are all smaller than those derived from CLD areas. The 4-year average values of SW, LW and Net CRF are -138.9 , 69.1 , and -69.7 Wm^{-2} , respectively, for COD

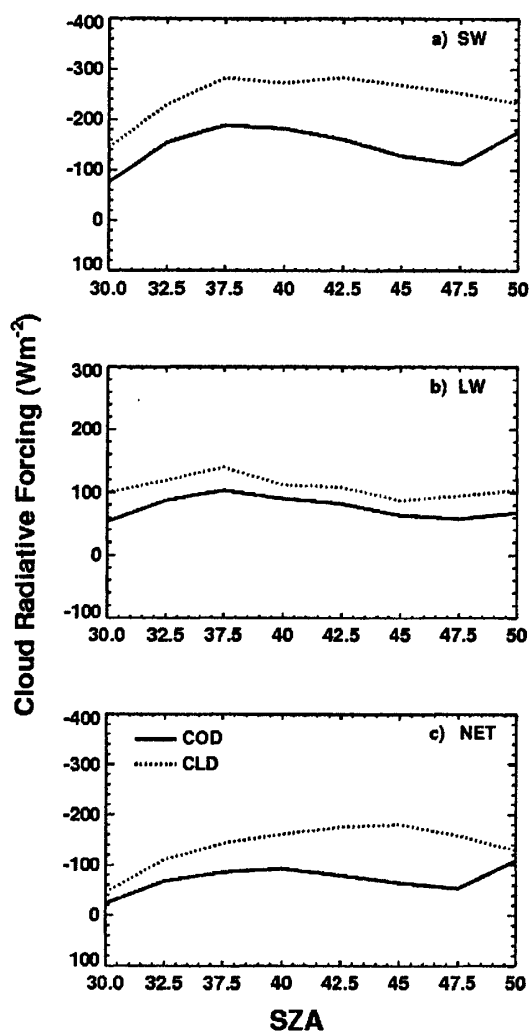


Fig. 2. Comparison of TOA dusty (COD) and dust-free (CLD) instantaneous cloud radiative forcing as a function of solar zenith angle for (a) SW forcing, (b) LW forcing, and (c) Net forcing.

regions. These values are only 57.0, 74.2, and 46.3% of those in CLD regions, respectively. Figure 1 indicates that the dust reduces the cloud modulation of the radiative energy budget including the cloud albedo (cooling) and greenhouse (warming) effects as well as the total cloud cooling effect. These results are consistent with Huang et al. (2006a) except that the net CRF in the COD regions is 40% lower. It may be due to stronger dust storms during 2005 and 2006 than during 2003 and 2004. Both previous studies (Huang et al., 2006a, b) suggest that the existence of dust under clouds significantly reduces the cooling by clouds. The reduced cooling due to

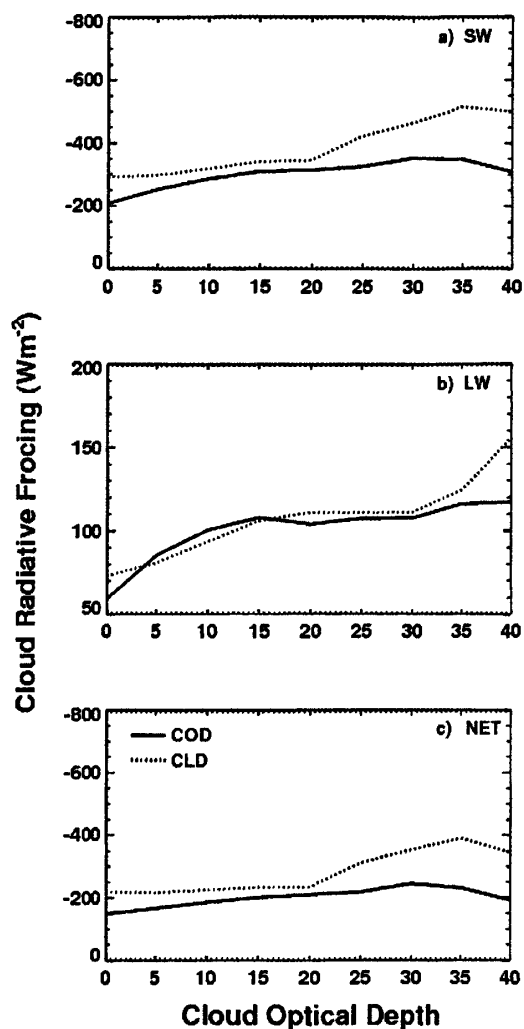


Fig. 3. Same as Fig. 2 but as a function of cloud optical depth.

dust in the COD regions can be considered as a warming effect of these aerosols. To confirm that the subtle differences in CRF are not caused by the clouds themselves, we compare the CRF in the CLD and COD regions as a function of solar zenith angle (see Fig. 2) and cloud properties, such as cloud top temperature, optical depth (see Fig. 3), particle size, and cloud water path.

Figures 2 and 3 compare the instantaneous SW, LW and Net CRF at TOA from the COD and CLD regions as functions of solar zenith and cloud optical depth, respectively. The absolute values of SW and Net CRF in Figs. 2a, c and 3a, c in COD regions are significantly smaller than those in CLD areas. However, for LW CRF, as shown in Figs. 2b and

3b, the differences between the COD and CLD CRF values are less distinct, especially in Fig. 3b where the cloud optical depth is smaller than 35. The CRF differences between the COD and CLD conditions depend more on the cloud optical depth than on the solar zenith angles (Figs. 2 and 3). The above analyses clearly show that cloud radiative forcing is influenced by aerosols when dust plumes exist under the cloud. These influences include direct, indirect and semi-direct effects of dust aerosols. Figure 4 shows the frequency distributions of ice cloud effective diameter (De), water cloud effective radius (Re) and cloud optical depth for the COD and CLD regions. The mean values of De and Re in the COD regions are 17.0% and 30.0%, respectively, less than those derived from CLD region. In the dust polluted area (COD), the dust may serve as extra condensation nuclei which restrain cloud particle growth and lead to the smaller values of De and Re compared to those in CLD regions.

The smaller De/Re associated with the aerosol first and second indirect effects (Twomey, 1977; Albrecht, 1989) would enhance the cloud optical depth and cloud solar albedo. (Note that the cloud optical depth is proportional to the IWP/LWP but inversely proportional to the effective size of cloud particles.) The average COD cloud optical depth, 10.1, is, however, 34% smaller than that, 15.6, for CLD areas. The decrease in the optical depth may be explained by the evaporation of low clouds due to the dust aerosol semi-direct effect that dominates the first and second indirect effects (Huang, et al., 2006b). Such a decrease may be also partly attributed to the aerosol radiative effects on surface fluxes and the hydrological cycle (Jiang and Feingold, 2006), although their relative contributions to the decrease of cloud optical depth remain as an important topic for future investigation. It is evident that the dust aerosol can change the cloud micro- and macro-physical properties and, thus, the CRF.

To further elucidate the dust effects, simulations were conducted using the Fu-Liou radiative transfer model. In the model calculations, the dust layer was placed below 4 km: the aerosol loading decreases exponentially with height and, at 4 km, it is $1/e$ of the surface value. The type of dust was selected as coarse mode. The single-scattering albedo and asymmetry factor of coarse mode dust are 0.7266 and 0.8613, respectively. Vertical profiles of temperature, pressure, humidity, and gaseous concentrations correspond to the mid-latitude summer (MLS) atmosphere of McClatchey et al. (1971). The cloud was placed between 3.5 and 5.5 km over a desert surface.

The direct effect of dust aerosols consists of scattering and absorption of incoming solar radiation. Because dust is an absorbing aerosol, it can exert different influences on the SW and LW fluxes at both the TOA and the surface. As dust optical depth (τ) increases (Fig. 5), the cooling resulting from the SW and Net CRFs decreases at the TOA, while it increases significantly at the surface. These differences imply that the dust aerosols decrease the cloud cooling effects at the TOA and increase the cloud cooling effects at surface. The re-

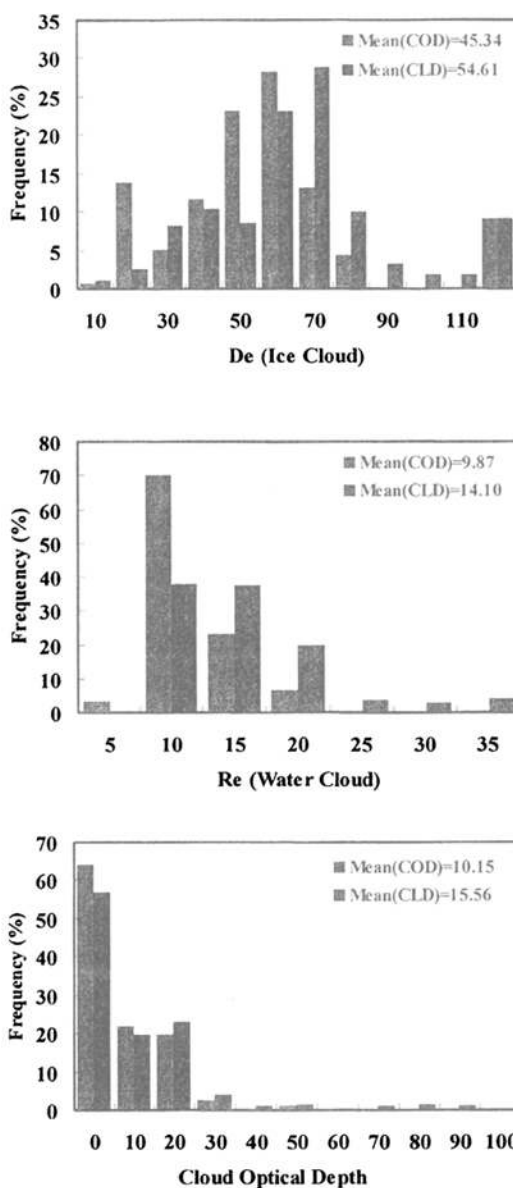


Fig. 4. Comparison of dusty and dust-free cloud properties for (a) ice cloud De , (b) water cloud Re and (c) cloud optical depth. The histogram intervals are $5 \mu m$ for Re , $20 \mu m$ for De and 10 for optical depth.

duced/enhanced cloud cooling effects are actually the direct warming/cooling effects due to dust aerosols. From Fig. 5, it is clear that when the aerosol optical depth increases by 0.1, the value of instantaneous SW CRF will change from 2.7 to

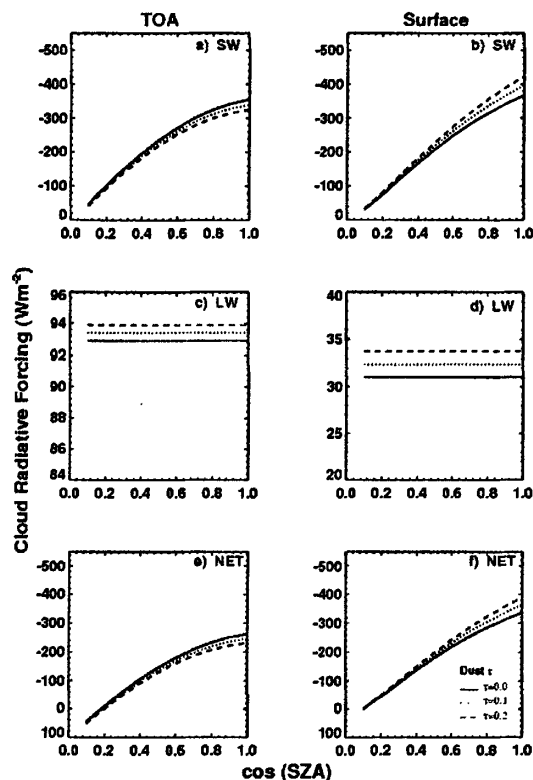


Fig. 5. Model simulated dusty instantaneous cloud radiative forcing as a function of solar zenith angle for (a) TOA SW forcing, (b) surface SW forcing, (c) TOA LW forcing, (d) surface LW forcing, (e) TOA net forcing, and (f) surface net forcing. The amount of aerosol loading is indicated on the plots, and cloud optical depth is 16. The aerosol type is Coarse Mode. The surface type is desert.

15.7 Wm^{-2} at TOA and -1.7 to -29.7 Wm^{-2} at surface, depending on the solar zenith angles. When dust aerosols occur under clouds, their SW and Net warming effects at the TOA are due to the dust absorption, while both the absorption and scattering by dust prevent solar energy from reaching the surface and cause a cooling there. The difference in LW CRF between dust and dust-free cloudy skies is always positive because of the greenhouse effect of dust aerosols. But it is small because the longwave flux is relatively insensitive to the presence of dust. The impact of these aerosols on LW radiation at the TOA is smaller than that at the surface because the clouds above the aerosol layer mask any dust effect.

Figure 6 shows the instantaneous SW, LW and Net cloud radiative forcings at the TOA and surface as functions of cloud optical depth over desert surface. When the cloud optical depth changes, the dust still has a warming effect at the TOA and cools the surface. These simulation results suggest

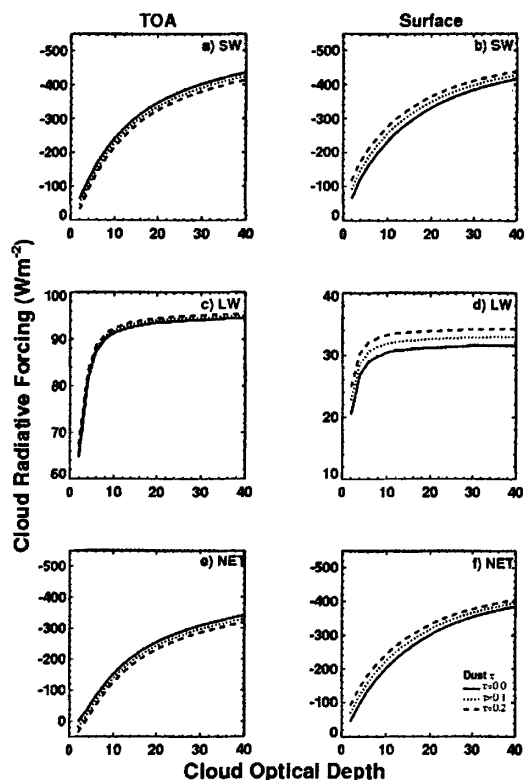


Fig. 6. Same as Fig. 5 but as a function of cloud optical depth.

that as an absorbing aerosol, dust has the direct effect of diminishing cloud cooling at the TOA, so, i.e., dust has a net warming direct effect at the TOA.

Satellite observations suggest that the total dust aerosol radiative forcing, which includes direct and, through cloud property changes, semi-direct and indirect effects (i.e., the difference in CRF between COD and CLD regions), is positive at TOA. However, it is difficult, using only satellite observations, to separate the dust aerosol direct effect from those caused by altered cloud properties. To estimate the separate effects, we propose a simple method that uses both satellite observations and the model calculations. Figure 7 compares the CERES-measured instantaneous SW CRF values with the Fu-Liou model simulation results for the CLD regions. The comparison shows good agreement between CERES observations and the Fu-Liou model simulation, indicating that the Fu-Liou model can reproduce the TOA radiation flux for dust-free cloud regions reasonably well. This result suggests that the contribution of dust aerosol direct and indirect/semi-direct effects can be estimated by using the differences between the satellite observations and the model simulations.

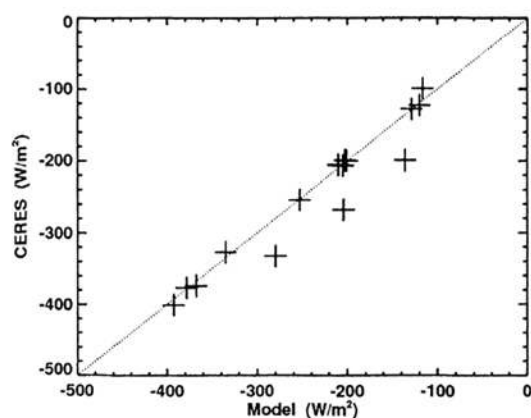


Fig. 7. Comparison of the instantaneous dust-free cloud CRF value (in CLD region) of CERES measurements with Fu-Liou model simulation.

The procedure for estimating the contribution of dust indirect and semi-direct effect is as follows:

1. Retrieve the dusty and dust-free cloud properties (i.e. cloud optical depth, R_e , D_e , cloud top pressure) through satellite measurements.
2. Simulate the dusty and dust-free cloud RF using satellite-retrieved cloud properties as input parameters. (Note that the dust optical depth was set to be zero in both simulations, although the cloud properties were changed by dust aerosols in the COD regions).
3. The total dust aerosol radiative forcing including direct and semi-direct and indirect effects is

$$ARF_{\text{total}} = F_{\text{CERES}}(\text{COD}) - F_{\text{CERES}}(\text{CLD})$$

where $F_{\text{CERES}}(\text{COD})$ and $F_{\text{CERES}}(\text{CLD})$ are the radiative energy budgets at the TOA from CERES observations under COD and CLD conditions, respectively.

4. The combination of dust aerosol indirect and semi-direct forcing related to cloud property changes is

$$ARF_{\text{indirect+semi}} = F_{\text{model}}(\text{COD}) - F_{\text{model}}(\text{CLD})$$

where F_{model} is the radiative energy budget at the TOA from the Fu-Liou model computed using satellite-retrieved cloud properties, as in steps 1 and 2.

5. The direct radiative forcing due to dust aerosols is

$$ARF_{\text{direct}} = ARF_{\text{total}} - ARF_{\text{indirect+semi}}$$

In other words, dust direct radiative forcing is the total dust radiative forcing in clear sky conditions.

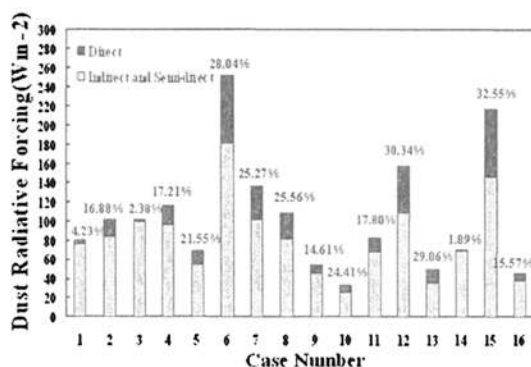


Fig. 8. The mean values of direct and combination of indirect and semi-direct instantaneous SWRF at TOA for 16 cases. The numbers represent the percentage of direct (direct/total).

The averaged dust aerosol direct and combination of indirect and semi-direct effects on the instantaneous SW RF at TOA for the selected cases are shown in Table 2. The averaged direct, and combined indirect and semi-direct instantaneous SW RF are 22.7 and 82.2 W m^{-2} , respectively, which correspond, respectively, to 21.6 and 78.4% of the total RF value. The dust aerosol direct effect of 22.7 W m^{-2} as derived from the satellite-measured radiation and cloud properties here is within the range of those from the model computations as shown in Figs. 5 and 6 (where the dust aerosol direct radiative forcing is the difference between the dashed/dotted line and solid line). The percentages of direct and combined indirect and semi-direct RF at TOA for the 16 selected cases are shown in Fig. 8. It is obvious from Table 2 and Fig. 8 that the combination of dust indirect and semi-direct effects is the primary contributor to the dusty cloud warming effect.

5 Conclusion and discussion

Asian dust aerosols have important influences on the radiative energy budget through both the direct effect and effects on the cloud microphysical properties. Most previous analyses of aerosol radiative forcing were limited to clear sky only (direct forcing). Recent studies have found that dust plays an important role in modulating climate through the indirect aerosol effect on cloud properties (Sassen et al., 2003). Our analyses of satellite observations show that the absolute values of instantaneous TOA CRF in COD regions are significantly smaller than those derived from CLD regions. This implies that when dust exists under a cloud it has a warming influence at TOA. It is clearly evident from analysis of satellite data that dust aerosols can change cloud microphysical properties, such as water droplet radius or effective ice particle diameter and cloud optical depth. These properties consistently yield smaller values when derived from COD

Table 2. Comparison of the averaged instantaneous RF effect at TOA.

	Mean SWRF Wm ⁻²	Standard Deviation	Mean Percentage (direct/total)	Standard Deviation
Direct effect	22.7	44.45	19.21%	9.80%
Indirect effect	82.2	41.15		

regions as compared to those derived over CLD regions. The results presented here are consistent with our previous study (Huang et al., 2006a). The warming effects of dust aerosols at TOA through dust aerosol direct effects in COD regions were also indicated by Fu-Liou model simulation.

Although the impact of the indirect/semi-direct effect of dust aerosols can be seen from the satellite-retrieved cloud microphysical properties, it is difficult to separate those effects on the radiative energy budget from their direct effect using satellite observations alone. This study has proposed a new approach by combining observation and simulation results to quantify the indirect and semi-direct effects of dust aerosols as well as their direct effect. Satellite observations along with the radiative transfer calculations show that the dust aerosol indirect and semi-direct effects dominate the dust aerosol total RF. The four-year (2002–2006) combined average indirect and semi-direct instantaneous shortwave radiative forcing (SWRF) is 82.2 Wm⁻², a value that accounts for 78.4% of the total RF value. The dust aerosol direct effect of 22.7 Wm⁻² as derived from satellite radiation and cloud observations here is within the range of those determined from the model simulations. Because both the first and second indirect effects cause cooling at TOA, the warming effect of dust aerosols in cloud-over-dust systems is due to the semi-direct effect of dust aerosols (Huang et al., 2006b). Although this paper is only a preliminary study of Asian dust effects in cloudy conditions, it sets the stage for more in-depth analyses by providing a method to separate the various types of aerosol effects. The use of other datasets in combination with the model and CERES data will further enhance the quantification of the different effects.

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Taklimakan dust aerosol radiative heating derived from CALIPSO observations using the Fu-Liou radiation model with CERES constraints

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Abstract. The dust aerosol radiative forcing and heating rate over the Taklimakan Desert in Northwestern China in July 2006 are estimated using the Fu-Liou radiative transfer model along with satellite observations. The vertical distributions of the dust aerosol extinction coefficient are derived from the CALIPSO (Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations) lidar measurements. The CERES (Cloud and the Earth's Energy Budget Scanner) measurements of reflected solar radiation are used to constrain the dust aerosol type in the radiative transfer model, which determines the dust aerosol single-scattering albedo and asymmetry factor as well as the aerosol optical properties' spectral dependencies. We find that the dust aerosols have a significant impact on the radiative energy budget over the Taklimakan desert. In the atmospheres containing light, moderate and heavy dust layers, the dust aerosols heat the atmosphere (daily mean) by up to 1, 2, and 3 K day⁻¹, respectively. The maximum daily mean radiative heating rate reaches 5.5 K day⁻¹ at 5 km on 29 July. The averaged daily mean net radiative effect of the dust are 44.4, -41.9, and 86.3 W m⁻², respectively, at the top of the atmosphere (TOA), surface, and in the atmosphere. Among these effects about two thirds of the warming effect at the TOA is related to the longwave radiation, while about 90% of the atmospheric warming is contributed by the solar radiation. At the surface, about one third of the dust solar radiative

cooling effect is compensated by its longwave warming effect. The large modifications of radiative energy budget by the dust aerosols over Taklimakan Desert should have important implications for the atmospheric circulation and regional climate, topics for future investigations.

1 Introduction

Aerosols influence the radiative energy budget directly by scattering and absorbing solar radiation (*direct effect*), and indirectly by altering cloud droplet size distribution and concentration (*indirect effect*) (Twomey, 1977; Albrecht, 1989). Absorbing aerosols, such as black carbon and mineral dust, could contribute to large diabatic heating in the atmosphere that often enhances cloud evaporation (*semi-indirect effect*) (Ackerman et al., 2000; Koren et al., 2004; Kruger and Graßl, 2004; Huang et al., 2006). The magnitude of the global mean radiative effect of dust aerosols is comparable to that of anthropogenic aerosols from sulphates and biomass combustion (Sokolik and Toon, 1999). Dust aerosols also have a significant greenhouse effect through longwave radiation which warms both the surface and atmosphere (e.g., Zhang and Christopher 2003). However, there are considerable uncertainties in estimating the radiative effects of dust aerosols. The net radiative effect at the top-of-atmosphere (TOA) could be either positive or negative, depending on several key variables, such as surface albedo, particle size, vertical distribution of the dust layer, dust optical depth, and



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the imaginary part of the refractive index (Liao and Seinfeld, 1998). Recently, Satheesh et al. (2006) studied the atmospheric warming due to dust aerosols over the Afro-Asian region. They found a reduction of solar radiation reaching at the surface with a lower atmospheric warming of 0.3 to 0.5 K day⁻¹.

The vertical distribution of dust aerosols is one of the critical parameters in the assessment of the dust radiative effect (Claquin et al., 1998; Zhu et al., 2007). A model study by Carlson and Benjamin (1980) showed that an elevated Saharan dust layer could change the atmospheric heating rate dramatically. Liao and Seinfeld (1998) claimed that clear sky long-wave radiative forcing and cloudy sky TOA shortwave (SW) radiative forcing of dust aerosols are very sensitive to the altitude of the dust and cloud layers. Meloni et al. (2005) also found that SW aerosol radiative forcing at the TOA has a significant dependence on aerosol vertical profiles.

The recently launched CALIPSO satellite provides a wealth of actively sensed vertical structures of aerosols over regional and global scales and provides an unprecedented opportunity to study the radiative effects of dust aerosols. Unlike the space-based passive remote sensing instruments, CALIPSO can observe aerosols over bright surfaces and beneath thin clouds as well as in clear sky conditions (Winker et al., 2006; Hu et al., 2006, 2007; Liu et al., 2004, 2008; Huang et al., 2007, 2008). One of the most distinct advantages of the CALIPSO lidar observations is that it provides a direct measure of the vertical structure of aerosols.

This paper estimates the dust aerosol radiative heating rate and the radiative effect during the dust events that occurred over the Taklimakan Desert in the summer of 2006. The Taklimakan Desert is a significant source of airborne dust that affects much of Eastern Asia, the Northern Pacific, and sometimes North America (references, Huang et al., 2008). The Fu-Liou radiation model (Fu and Liou, 1992, 1993) is used to compute the aerosol heating rates and radiative forcing. The vertical distributions of dust aerosol extinction coefficients used in the computations are derived from the CALIPSO lidar observations. We use the reflected solar radiation measured at the top of the atmosphere (TOA) from the Clouds and Earth's Radiant Energy System (CERES) (Wielicki et al., 1996) Single Satellite Footprint (SSF) to constrain the dust aerosol type employed in the radiation model. The combination of the radiation model with the CALIPSO and CERES observations should lead to a reliable estimate of the dust aerosol radiative effects.

The paper is organized as follows. The summer Taklimakan dust events and the dust aerosol extinction profiles from CALIPSO observations are discussed in Sects. 2 and 3, respectively. The radiation model and the CERES constraint of dust aerosol single scattering albedo and asymmetry factor are described in Sect. 4. The estimation of the Taklimakan dust aerosol radiative heating and forcing is presented in Sect. 5 and summary and conclusions are given in Sect. 6.

2 Summer Taklimakan dust events

The Taklimakan Desert is a desert in Central Asia, in the Xinjiang Uyghur Autonomous Region of China, which is known as one of the largest sandy deserts in the world. It covers an area of 270 000 km² of the Tarim Basin, which is 1000 km long and 400 km wide. The Taklimakan Desert is about 1 km above sea level, surrounded by mountains except for an opening at its northeast corner.

Most Asian dust studies have focused on the late winter and spring due to observed long-range dust transport (Iwasaka et al., 1983; Zhang et al., 1997; Murayama et al., 2001; Uno et al., 2001; Sun et al., 2001; Wang et al., 2005). Uno et al. (2008) reported the 3-dimensional structure of Asian dust outflow from a dust source region to the Northwestern Pacific Ocean. They found that the elevated dust was transported to the Pacific Ocean with the major dust layer maintaining a height between 2.5–4.0 km. There have been very few studies analyzing the specific signatures of summer and fall dust storms over the Taklimakan Desert. Recently CALIPSO lidar observations show that dust events occur throughout the year over the Taklimakan (Liu et al., 2008a) and that heavy dust storms are part of the summer weather (Huang et al., 2007). The impact of the Taklimakan dust storms on the radiative energy budget and the implication to the regional climate are open questions. In this study we will quantify the vertical structures of dust aerosol extinction coefficients and radiative effects during the dust event that occurred during the period of 26 to 31 July 2006 over the Taklimakan.

The start of the dust outbreak was associated with the intensive low pressure system over West Siberian with a trough extending from northwest toward southeast. This caused a large pressure gradient and strong northwesterly winds (>20 m/s) between 45° and 55° N, which resulted in the onset of this dust episode. During 26 July, a moderate wind and dust storm in the northern Taklimakan, accompanied by localized severe dust storms, developed and extended southward. Under the influence of this storm, a wind-blown sand and/or dust cloud persisted over northern Qinghai and Tibet through 31 July. The strong dust events mainly occurred between 27 and 29 July.

3 Vertical structures of dust aerosol extinction coefficient

CALIPSO lidar measurements are used to derive the vertical distribution of dust aerosol extinction coefficient. The CALIPSO lidar is designed to acquire vertical profiles of elastic backscatter at two wavelengths (532 and 1064 nm) from a near nadir-viewing geometry during both day and night phases of the Sun-synchronous orbit, which has a 13:30 LT equatorial crossing time. In addition to total backscatter at the two wavelengths, CALIPSO also provides

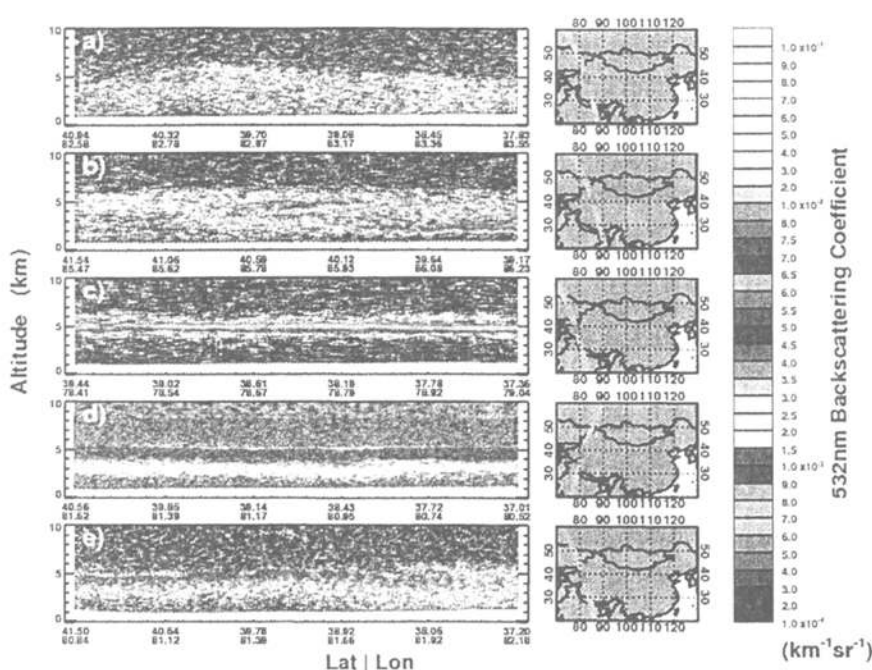


Fig. 1. The altitude-orbit cross-section of 532 nm total attenuated backscattering intensity (left panels) for the green-shaded portion of each track (right panels) over the Taklimakan Desert region (35° N–45° N) for (a) 24 July, (b) 26 July, (c) 29 July, (d) 30 July, and (e) 31 July 2006.

profiles of linear depolarization at 532 nm. The depolarization measurements enable the discrimination between ice and water clouds (Hu et al., 2007b), and the identification of non-spherical aerosol particles. The CALIPSO Level 1B data are used in this study and include a half orbit (day or night) of calibrated and geolocated single-shot lidar profiles with the highest vertical resolution. They include both 532 and 1064-nm attenuated backscatter and depolarization ratios at 532 nm. The product contains data from the nominal science mode measurement.

The dust aerosol optical depth for a given layer is retrieved in terms of backscatter from the CALIPSO lidar observations (Hu et al., 2006)] in the form,

$$\tau(z) = \frac{1}{2\eta} \ln(1 - 2\gamma'(z)S_{a,\text{eff}}) \quad (1)$$

where η is the layer-effective multiple scattering factor, which is 0.7 for this study (Omar et al., 2004). $S_{a,\text{eff}}$ is the product of η and the single-scattering lidar ratio (i.e., extinction-to-backscatter ratio), which has a value of 44 here (Omar et al., 2004), and $\gamma'(z)$ is the integrated attenuated backscattering coefficient β_a from the top to the base of the layer, which is defined as

$$\gamma'(z) = \int_{z_{\text{top}}}^{z_{\text{base}}} \beta_a(z) dz \quad (2)$$

The dust aerosol extinction coefficient, β , is thus obtained from

$$\beta(z) = \frac{\tau(z)}{\Delta z} \quad (3)$$

where $\Delta z = (z_{\text{top}} - z_{\text{base}})$ is the vertical resolution, which is 30 m below and 60 m above 8.2 km, respectively.

Figure 1 shows the CALIPSO orbit-altitude cross-section of the 532-nm total attenuated backscattering coefficient from 24–31 July 2006, where daytime data were used for 24, 26, 29, and 31 July and nighttime data were taken for 30 July. The CALIPSO data reveals that vertically extended dust layers are widespread throughout the Tarim Basin with peak lidar returns between 2.5 and 5.5 km above mean sea level (MSL). In general, the red-gray-white color scales used in CALIPSO data analyses, as shown in Fig. 1, indicate clouds and green-yellow-orange color features are aerosols. However, the heavy dust layers over the source regions are often misclassified as clouds in the current data products because the dust aerosol optical properties including the color ratio (defined as the ratio of 1064 to 532 nm attenuated backscatter) and backscattering intensity are similar to clouds during heavy dust loading episodes (optical depth $> 1 \sim 2$) (Liu, et al., 2008a)]. For example, the heavy dust layer over the Taklimakan Desert on 29 July is misclassified as cloud (gray in Fig. 1c). Figure 2 shows the volume depolarization

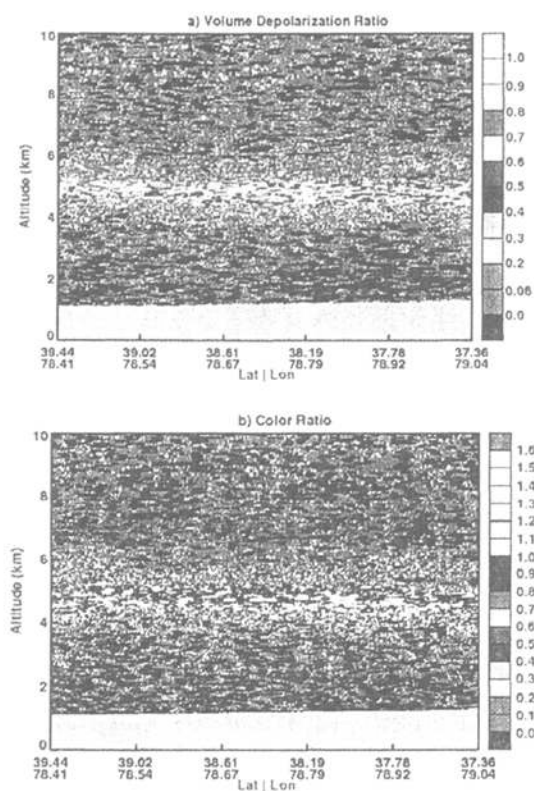


Fig. 2. The altitude-orbit cross-section of (a) volume depolarization ratio and (b) 1064-nm/532-nm backscatter color ratio for 29 July 2006 over the Taklimakan Desert region (35° N–45° N).

ratio (defined as the ratio of perpendicular-to-parallel components of received lidar signals at 532 nm) and the backscatter color ratio. The dust aerosols have a large depolarization ratio due to their nonsphericity (Fig. 2a), while they also have a large color ratio due to the relatively large size of the particles (Fig. 2b). On the other hand, the depolarization ratio is near zero for water clouds and other types of aerosols. Based on all this information from the CALIPSO measurements including the attenuated backscattering, depolarization ratio, and backscattering color ratio, we have identified the intensive backscattering layer in Fig. 1c as the dust layer. Independent observations from both *Aqua* MODIS (Fig. 3) and Cloudsat radar (figure not shown) confirm that the intensive dust layer occurred on 29 July.

The orbit-altitude cross sections (left) and orbit mean profiles (right) of the dust aerosol extinction coefficients are given in Fig. 4. For all cases, the dust layer can be distinctly identified. At the beginning of the dust episode (24 July), the extinction coefficient is $\sim 0.1 \text{ km}^{-1}$ from the surface to 5 km and then decreases with height above $\sim 5 \text{ km}$. On 26 July, the

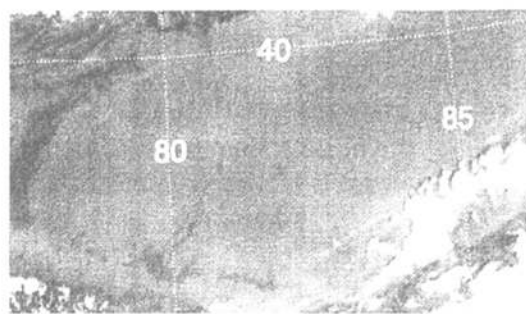


Fig. 3. The true color Aqua satellite image over Northwest China for 29 July 2006, in which channel $0.65 \mu\text{m}$, $0.56 \mu\text{m}$ and $0.47 \mu\text{m}$ are associated with red, green and blue colors, respectively.

elevated dust layer starts to develop as indicated by the enhanced dust aerosol extinction coefficients in Fig. 4b. From 29 to 31 July, the elevated dust layers are located at about 5 km, which leads to the transport of dust aerosols to the Tibetan plateau (Huang et al., 2007). For the heavy dust layer on 29 July (Fig. 4c), the mean dust extinction coefficient at 5 km is about 0.35 km^{-1} .

Figure 5 shows the column dust optical depth for those five days. The spatial variation of the optical depth is significant on 24 and 26 July, ranging from ~ 0.4 to 0.9 (Fig. 5a and b). On 29 and 30 July (Fig. 5c and d), the variation of dust optical depth along the CALIPSO track is small with a mean around 0.8 – 0.9 . On 31 July, the dust optical depth ranges from 0.6 to 0.8 (Fig. 5e).

4 Fu-Liou radiation model and dust aerosol type constrained with CERES measurements

We use the Fu-Liou radiation model along with the input of dust aerosol extinction coefficients from the CALIPSO observations to estimate the impact of dust aerosols on the radiative energy budget. This model was originally developed by Fu and Liou (1992, 1993) and subsequently modified by Rose and Charlock (2002) and Kato et al. (2005). It is a delta-four stream radiative transfer scheme with fifteen spectral bands from 0.175 to $4.0 \mu\text{m}$ in SW and twelve longwave (LW) spectral bands between 2850 and 0 cm^{-1} . The correlated k -distribution method is used to parameterize the non-gray gaseous absorption by H_2O , CO_2 , O_3 , N_2O and CH_4 (Fu and Liou, 1992) with the addition of CFCs and CO_2 in the window region (Kratz and Rose, 1999). The single-scattering properties of dust aerosols including normalized extinction coefficients, single-scattering albedo, and asymmetry factor are based on the four dust aerosol modes described by Hess et al. (1998). The single-scattering albedo (ω) and asymmetry factor (g) at $0.67 \mu\text{m}$ are given in Table 1 for these four dust aerosol types.

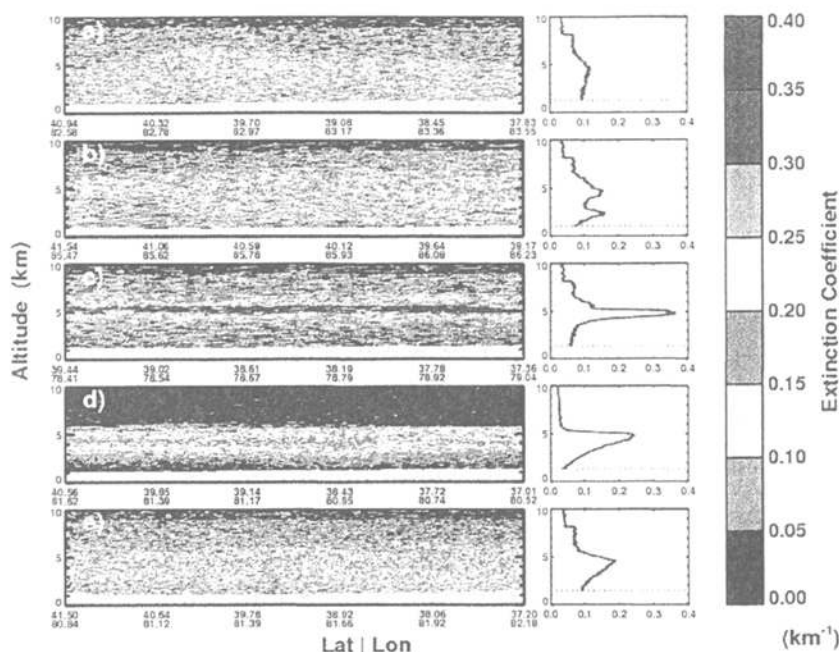


Fig. 4. The altitude-orbit cross-section of dust aerosol extinction coefficient (left panels) and orbit averaged vertical profile for (a) 24 July, (b) 26 July (c) 29 July, (d) 30 July, and (e) 31 July 2006.

Table 1. Single-scattering albedo (ω) and asymmetry factor (g) at $0.67 \mu\text{m}$ for 4 dust aerosol types used in Fu-Liou model.

	Nucleation mode	Accumulation mode	Transported mode	Coarse mode
Single-scattering albedo (ω)	0.9767	0.9203	0.89	0.7266
asymmetry factor (g)	0.6471	0.7143	0.7460	0.8613

For a given time and location, the pressure, temperature, and water vapor profiles are interpolated from the NCEP/NCAR reanalysis and ozone concentration is taken from the NCEP Stratospheric Monitoring Group Ozone Blended Analysis (SMOBA) product based on SBUV and TOVS observations. The surface albedo is based on MODIS observations as used by the CERES team (Minnis et al., 2008) with the spectral dependence prescribed for the desert scene type (T. Charlock, personal communication, 2007), which is validated with the CERES observations along the Fu-Liou radiation model. The mean value of the broadband surface albedo used in this study is 0.346 and the standard deviation is 0.005. Climatological concentrations are used for CO_2 , CH_4 , N_2O , and CFCs.

To evaluate the radiative heating and forcing due to dust aerosols, the aerosol optical properties such as extinction coefficient, single-scattering albedo and asymmetry factor are required. These properties depend upon dust characteris-

tics such as density, size distribution, and refractive index. The dust aerosol extinction coefficients at 532 nm are derived from CALIPSO measurements. In this study, we select the dust aerosol type that determines the single-scattering albedo and asymmetry factor and the spectral dependences of dust single-scattering properties by comparing the model-simulated reflected instantaneous solar radiative fluxes at the TOA with those observed by CERES. The CERES Aqua Edition 1B SSF data are used here (Wielicki et al., 1996). CERES SSF data sets combine CERES radiation measurements, MODIS cloud microphysical retrievals, and ancillary meteorology fields to form a comprehensive, high-quality compilation of satellite-derived cloud, aerosol, and radiation budget information for radiation and climate studies.

To optimize the dust aerosol single-scattering albedo and asymmetry factor used over the Taklimakan region, we compare the CERES TOA solar fluxes with Fu-Liou model simulations along the CALIPSO orbit using 4 different dust

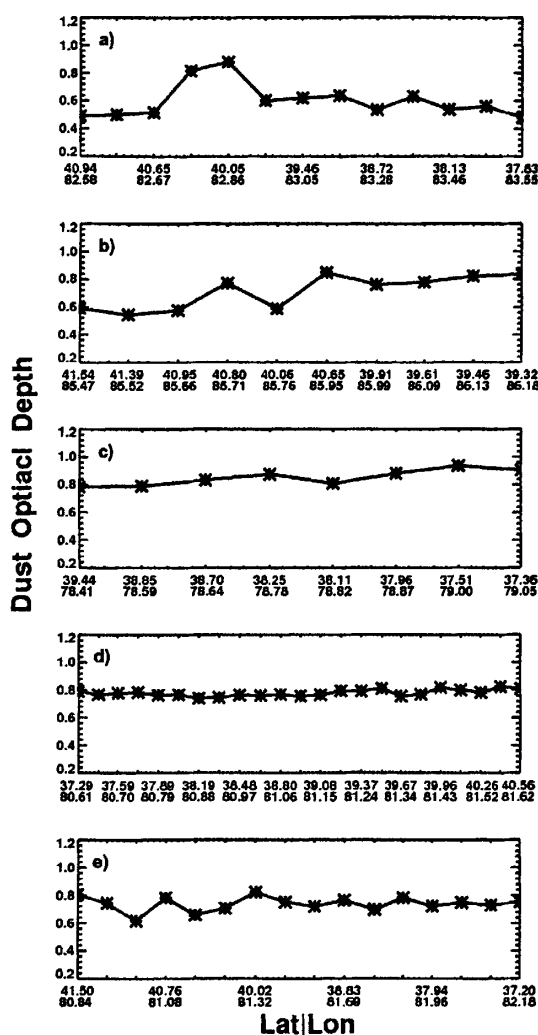


Fig. 5. Dust aerosol optical depth along CALIPSO track for (a) 24 July, (b) 26 July, (c) 29 July, (d) 30 July, and (e) 31 July 2006.

aerosol types from OPAC (Optical properties of aerosol and clouds: The software package OPAC) (Hess, et al., 1998). We found that the dust aerosol type that fits best is the transported mode. Figure 6 shows a comparison of the TOA reflected solar fluxes derived from the Fu-Liou model with CERES measurements along the CALIPSO orbit over the Taklimakan Desert region (35°N – 45°N) for the daytime cases (24, 26, 29, 31 July). The model-simulated TOA short-wave fluxes agree reasonably well with those from CERES. The averaged difference between the model simulations and CERES measurements is only 1.5 W m^{-2} . Thus, it is clear from this comparison that the radiative transfer model con-

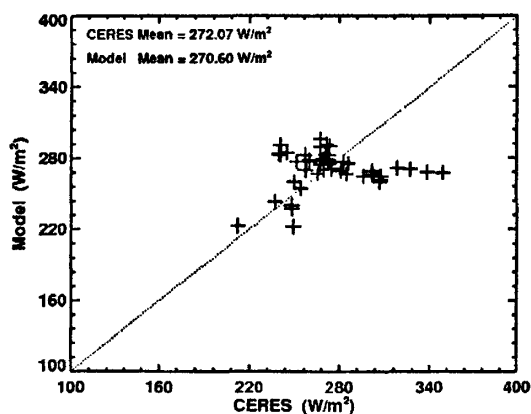


Fig. 6. Comparison of the modeled simultaneous TOA shortwave flux with CERES observations along the CALIPSO orbit over the Taklimakan Desert region (35°N – 45°N) on 24, 26, 29, and 31 July 2006.

strained with the CERES observations can be used to reliably determine the variation of dust aerosol radiative heating rates with the input of vertical distributions of dust aerosols from CALIPSO measurements.

We thus used the single-scattering albedo and asymmetry factor and their spectral dependences for the transported dust aerosol mode in all the simulations. The extinction coefficient is from

$$\beta(\lambda) = \frac{\beta(532\text{ nm})}{\beta_0(532)} \beta_0(\lambda), \quad (4)$$

where $\beta(532\text{ nm})$ is the dust extinction coefficient from CALIPSO at 532 nm and β_0 is the normalized extinction coefficients of the transported dust aerosol mode. Thus, our calculations allowed the extinction coefficient profiles to vary according to the CALIPSO observations.

5 Taklimakan dust aerosol radiative heating and forcing

Figures 7–9 show the impact of dust aerosols on the daily-mean (24 h average) SW, LW, and net heating rates, respectively. They are obtained as the differences between the simulated radiative heating rates with and without considering the observed dust aerosols. The dust has a significant effect on SW radiation. For relatively light (Fig. 7a), moderate (Fig. 7b and e) and heavy dust layers (Fig. 7c and d), dust aerosols heat the atmosphere via absorption of SW radiation by up to ~ 1 , 2, and 3 K day^{-1} , respectively. The maximum daily-mean solar radiative heating rate of 7 K day^{-1} is found at 5 km on 29 July. Figure 7 shows that the SW heating rates have a peak corresponding to the maximum dust aerosol extinction coefficient levels.

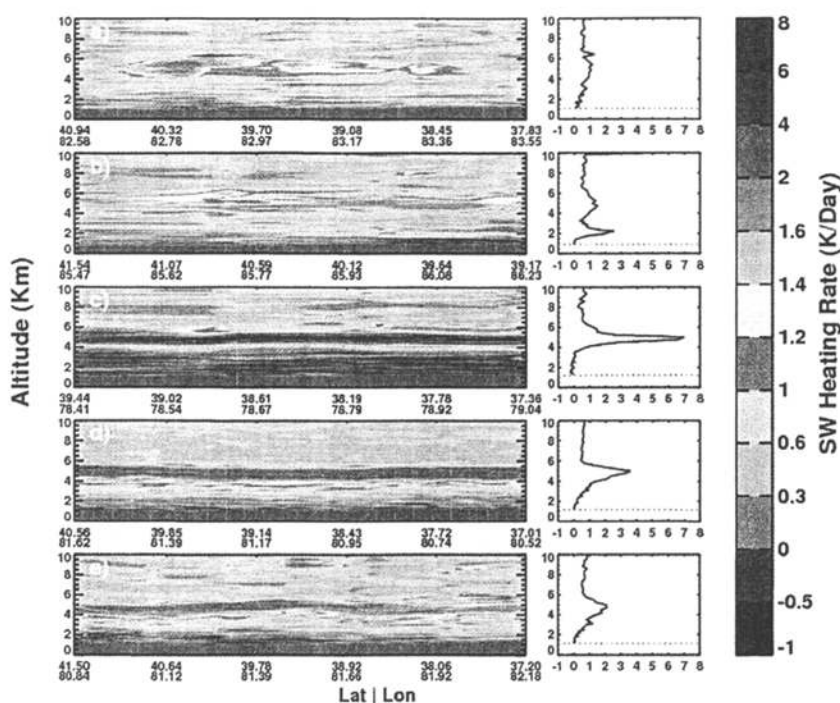


Fig. 7. The altitude-orbit cross-section of daily averaged shortwave heating rates due to dust aerosols (left panels) and the orbit-averaged vertical profile (right panels) for (a) 24 July, (b) 26 July (c) 29 July, (d) 30 July, and (e) 31 July 2006. Unit: K day^{-1} .

Table 2a. Daily mean radiative forcing of dust aerosols at TOA over the Taklimakan region. Unit: W m^{-2}

Date	24 July	26 July	29 July	30 July	31 July	Mean
SW	12.49	15.6	14.93	14.42	13.13	14.11
LW	28.64	28.86	32.73	27.64	33.65	30.30
Net	41.13	44.46	47.66	42.06	46.78	44.41

Table 2b. Same as Table 2a, but for surface. Unit: W m^{-2} .

Date	24 July	26 July	29 July	30 July	31 July	Mean
SW	-49.70	-67.0	-73.53	-67.27	-66.12	-64.72
LW	19.92	23.37	21.93	22.58	26.36	22.83
Net	-29.78	-43.63	-51.60	-44.69	-39.76	-41.89

Although dust aerosols appear to have less effect on LW radiative heating rates (Fig. 8), they do show a warming effect below the dust layers and cooling near the top of the layers. The maximum warming, which occurs near the surface, is typically about 0.5 K day^{-1} . The LW cooling ranges from

Table 2c. Same as Table 2a, but for in atmosphere. Unit: W m^{-2} .

Date	24 July	26 July	29 July	30 July	31 July	Mean
SW	62.19	82.6	88.46	81.69	79.25	78.83
LW	8.72	5.49	10.8	5.06	7.29	7.47
Net	70.91	88.09	99.26	86.75	86.54	86.31

near zero on 24 July to about -1.5 K day^{-1} on 29 July, which partly compensates the large solar radiative heating near the top of the dust layers. The net aerosol heating near the dust layer top is about 1 K day^{-1} on 24 July, $1.5\text{--}2.0 \text{ K day}^{-1}$ on 26 July, 5.5 K day^{-1} on 29 July, 3 K day^{-1} on 30 July, and 1.5 K day^{-1} on 31 July (Fig. 9). The net heating rates due to the dust aerosols are about $0.5\text{--}1.0 \text{ K day}^{-1}$ below and above the dust layers.

Table 2 shows the daily-mean dust aerosol radiation forcing at the TOA and surface and in the atmosphere over the Taklimakan region. Dust aerosols have a warming effect at the TOA in both SW and LW radiation, which are about 14 and 30 W m^{-2} , respectively. The net warming is 44.4 W m^{-2} . The positive solar radiative forcing is due to the elevated absorbing dust aerosol layer above the highly

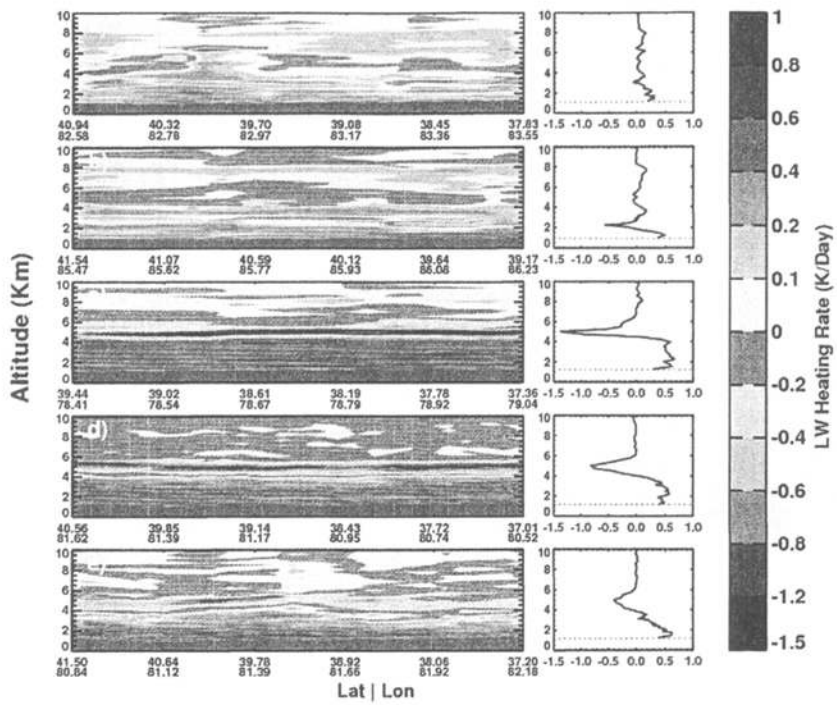


Fig. 8. Same as Fig. 7, but for longwave heating rates.

Table 3. Estimation of uncertainties of radiation forcing. Unit: W m^{-2} .

SSA $\pm 6\%$			Surface Albedo $\pm 10\%$			Lidar Ratio $\pm 20\%$			
	TOA	Surface	Atmos	TOA	Surface	Atmos	TOA	Surface	Atmos
SW	∓ 14.815	± 11.450	∓ 26.265	± 4.585	± 3.360	± 1.225	± 1.775	∓ 11.155	± 12.935
LW	∓ 0.360	∓ 0.370	± 0.010	0	0	0	± 4.820	± 3.480	± 1.335
Net	∓ 15.175	± 11.080	∓ 26.255	± 4.585	± 3.360	± 1.225	± 6.595	∓ 7.675	± 14.270

Table 4. Estimation of uncertainties of vertical averaged atmospheric heating rate. Unit: K/day

	SSA $\pm 6\%$	Surface Albedo $\pm 10\%$	Lidar Ratio $\pm 20\%$
SW	∓ 0.272	± 0.010	± 0.138
LW	∓ 0.004	0	± 0.019
Net	∓ 0.276	± 0.010	± 0.157

reflective desert surface. The dust aerosol radiative forcing in the atmosphere is also positive for both SW and LW radiation, which are 78.8 and 7.5 W m^{-2} , respectively, and the total warming is 86.3 W m^{-2} . The dust aerosols cool the

surface significantly by decreasing the incident SW radiation (-64.7 W m^{-2}) but warms it through the dust-emitted LW radiation (22.8 W m^{-2}). The net cooling at the surface is 41.9 W m^{-2} . Table 2 indicates that both SW and LW radiative forcing of dust aerosols play an important role in the radiative energy budget at the TOA and the surface.

6 Uncertainties in radiation forcing

Although this study attempts to minimize the errors by using reliable observations for the model input and to constrain the dust aerosol single-scattering properties, the estimated dust radiative forcing may have some unavoidable uncertainties. They are related to the uncertainties in the CALIPSO

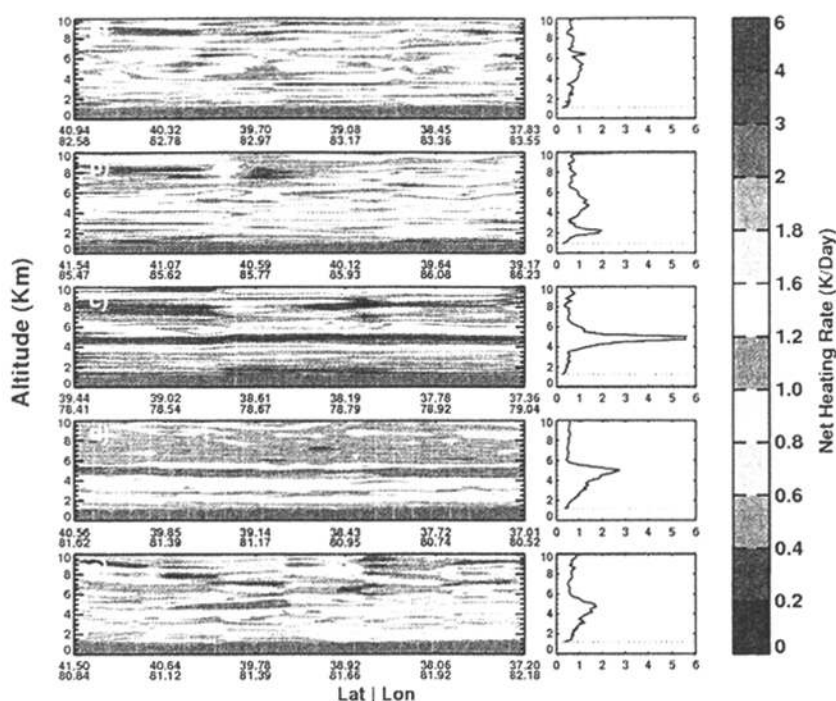


Fig. 9. Same as Fig. 7, but for net heating rates.

lidar ratio for the retrieval of dust aerosol extinction coefficient, surface albedo, and dust aerosol single scattering albedo (SSA). The range of uncertainties in the CALIPSO lidar ratio is about 20% (Winker et al., 2006). The sensitivity test showed that the lidar ratio uncertainty of $\pm 20\%$ can lead to uncertainties of about ± 6.8 , ± 7.6 , $\pm 14.4 \text{ Wm}^{-2}$ in the net dust radiative forcing at the TOA, in the atmosphere, and at the surface, respectively (see Table 3). The surface albedo is another possible source of error. If the surface albedo uncertainty is $\pm 10\%$, the dust SW radiation forcing at TOA and surface will be changed about ± 4.6 , $\pm 3.4 \text{ Wm}^{-2}$, respectively (see Table 3). Detailed discussion of uncertainties in SW radiative forcing with respect to surface albedo can be found in the studies of Claquin et al. (1998) and Liao and Seinfeld (1998). The major uncertainty in the estimated dust radiative forcing arises from the SSA. A ± 0.03 uncertainty in the AERONET SSA (Dubovik et al., 2002) can lead to a 12% uncertainty in the SW TOA forcing. Our sensitivity test revealed that the largest error is caused by the SSA uncertainty. The observation study of AERONET shows that maximum dust SW spectral weighted SSA may be 0.94 over Taklimakan, i.e., current value $0.89 \pm 6\%$ (Dubovik et al., 2002). Thus, the uncertainty range in the SSA is estimated to be $\pm 6\%$. A $\pm 6\%$ change in SSA (with a constraint that SSA should not be larger than one) can lead to about ± 15.2 ,

± 11.1 , $\pm 26.3 \text{ Wm}^{-2}$ uncertainties in the net dust radiative forcing at TOA, in the atmosphere, and at the surface, respectively (see Table 3). Although the aforementioned three parameters have large effects on the estimated dust radiative forcing, the pattern of dust heating rate is not affected. Only the magnitude of the dust heating rate varies with the changes of parameters. The SSA was found to have its largest effect in regulating the magnitude of the dust heating rate. If the SSA increases by 10%, the net vertically averaged atmospheric heating rate will change by roughly 0.27 K day^{-1} (see Table 4). The lidar ratio uncertainty of $\pm 20\%$ can lead to $\sim 0.16 \text{ K day}^{-1}$ uncertainty in the net vertical mean atmospheric heating rate (see Table 4), while 10% changes in the surface albedo can only cause 0.01 K day^{-1} changes in the net vertical mean atmospheric heating rate.

7 Discussion and conclusions

In this study, we investigated the impact of dust aerosols on the radiative energy budget over the Taklimakan Desert during dust episodes in late July 2006. The dust aerosol radiative forcing was estimated using the Fu-Liou model along with CALIPSO and CERES measurements. It was determined that dust aerosols warm the atmosphere over the Taklimakan, especially at the levels between 3 and 6 km where the

maximum aerosol extinction coefficients are found. In the dusty atmospheric layers, the dust typically heats the layer by up to $1\text{--}3\text{ K day}^{-1}$ depending on the dust concentration. The maximum daily mean net (shortwave + longwave) radiative heating rate reached 5.5 K day^{-1} at 5 km on 29 July. The averaged daily mean net radiative forcings of the dust aerosols, averaged over our case studies, were 44.4 , -41.9 , and 86.3 W m^{-2} , respectively, at the TOA, surface, and in the atmosphere. Among these forcings about two thirds of the warming effect at the TOA is due to absorption and emission of longwave radiation by the layer, while about 90% of atmospheric warming is due to absorption of the solar radiation. At the surface, about one third of the dust aerosol solar radiative cooling effect is compensated by the dust aerosol longwave warming effect. This study indicates that both shortwave and longwave radiative forcing of dust aerosols play an important role in the radiative energy budget at the TOA and the surface.

For the radiative effect of the dust aerosol, the most important factors for the shortwave forcing of dust aerosols at the TOA are the aerosol optical depth and the single scattering albedo while the longwave forcing is highly dependent on the vertical profile of the dust aerosols. Previous spectral, simultaneous remote and in situ observations suggest that the single scattering albedo of pure dust at a wavelength of $0.67\text{ }\mu\text{m}$ is predominantly in the range from 0.90 to 0.99, with a central global estimate of 0.96 (Forster, et al., 2007). This is in accordance with the bottom-up modeling of ω based on the hematite content in desert dust sources (Claquin et al., 1999; Shi et al., 2005). Analyses of ω from long-term AERONET sites influenced by Saharan dust suggest an average value of 0.95 at $0.67\text{ }\mu\text{m}$ (Dubovik et al., 2002), while unpolluted Asian dust during the Aeolian Dust Experiment on Climate (ADEC) had an average value of 0.93 at $0.67\text{ }\mu\text{m}$ (Mikami et al., 2006). These high ω values suggest that a significant positive shortwave radiative forcing by dust is unlikely (Forster et al., 2007). However, our results suggest that the single scattering albedo of Taklimakan dust aerosols is about 0.89 at $0.67\text{ }\mu\text{m}$ which is about 6% less than Saharan dust. The mean shortwave radiative forcing at TOA is as much as 14.1 W m^{-2} .

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