An "A-Train" Strategy for Quantifying Direct Climate Forcing by Anthropogenic Aerosols

BY THEODORE L. ANDERSON, ROBERT J. CHARLSON, NICOLAS BELLOUIN, OLIVIER BOUCHER, MIAN CHIN, SUNDAR A. CHRISTOPHER, JIM HAYWOOD, YORAM J. KAUFMAN, STEFAN KINNE, JOHN A. OGREN, LORRAINE A. REMER, TOSHIHIKO TAKEMURA, DIDIER TANRÉ, OMAR TORRES, CHARLES R. TREPTE, BRUCE A. WIELICKI, DAVID M. WINKER, AND HONGBIN YU

New-generation, formation-flying satellites offer tremendous benefits to aerosol science but require, as well, rigorous planning for coordinated research.

CONCEPTUAL FRAMEWORK. Understanding how anthropogenic aerosols impact the Earth's climate system is a daunting challenge. Quantifying these impacts will require a multifaceted, global observing system and a capacity for integrating diverse data (Seinfeld et al. 1996; Heintzenberg et al. 1996; Charlson 2001). Over the past decades, as evidence of the importance of aerosol impacts has accumulated, major components of the required observing system have been developed. A program for coordinating and integrating these observations, dubbed the PARAGON (see appendix A for acronym definitions) initiative, has recently been proposed in *BAMS* (Diner et al. 2004a). Implementing this program will require advance planning and worldwide scientific cooperation. Such coordination is critical if the opportunity that is afforded by new and enhanced satellite sensors (see "A-Train contribution" section) is to be realized.

In pursuit of these goals, we offer here an integration strategy for the specific problem of quantifying DCF by anthropogenic aerosols. We note that the quantification of "direct" forcing (i.e., the modification of radiative fluxes by aerosol particles themselves)

AFFILIATIONS: ANDERSON AND CHARLSON—University of Washington, Seattle, Washington; BELLOUIN, BOUCHER, AND HAYWOOD—Met Office, Exeter, Devon, United Kingdom; CHIN, KAUFMAN, AND REMER—NASA Goddard Space Flight Center, Greenbelt, Maryland; CHRISTOPHER—University of Alabama in Huntsville, Huntsville, Alabama; KINNE—Max Planck Institute for Meteorology, Hamburg, Germany; OGREN—NOAA/CMDL, Boulder, Colorado; TAKEMURA—Kyushu University, Fukuoka, Kyushu, Japan; TANRÉ—University of Lille, Lille, France; TORRES—JCTE University of Maryland, Baltimore County, Baltimore, Maryland; TREPTE, WIELICKI, AND WINKER—NASA Langley Research Center, Hampton, Virginia; YU—University of Maryland at Baltimore County, Baltimore, and NASA Goddard Space Flight Center, Greenbelt, Maryland

CORRESPONDING AUTHOR: Theodore L. Anderson, Dept. of Atmospheric Sciences, Room 408 ATG, University of Washington, Seattle, WA 98195-1640
E-mail: tadand@u.washington.edu
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is, in many ways, a prerequisite to tackling the more complex problem of "indirect" forcing (i.e., aerosol modification of cloud radiative properties). This is because quantifying DCF requires knowledge of the global distribution of anthropogenic aerosols.

Conceptually, the problem is framed as a need for complete global mapping of four parameters. The choice of these parameters, as described in this section, is based on a blend of historical, practical, and theoretical considerations, with an emphasis on practicality with respect to satellite mapping. The satellite focus is on the "A-Train," a constellation of six spacecraft that will fly in formation from about 2005 to 2008. This strategy is offered as an initial framework—subject to improvement over time—for scientists around the world to participate in the A-Train opportunity.

Relevant data for constraining DCF exist across a wide range of time and space scales. Consider the scale differences among the following: local in situ measurements, vertically resolved remote measurements, column-integral remote measurements, and grid-box averages from chemical transport models. The successful integration of these data is far from trivial. It will require i) a clear conceptual framework that links aerosol properties to climatic effects and ii) the identification of common metrics within this framework to permit both the comparison of disparate data and the diagnosis of discrepancies.

In developing a conceptual framework, we begin by recognizing that the concept of climate forcing, that is, an exogenous perturbation of the Earth's energy budget, has played a key role in climate change research. Over the past decade or more (see, e.g., Houghton et al. 1990, 1996, 2001), the equation

$$\Delta T = \lambda \Delta F \qquad (1)$$

has provided the basic conceptual framework whereby forced changes in global mean surface temperature ΔT are analyzed in terms of two fundamental aspects: climate forcing ΔF, and climate sensitivity λ (key symbols are defined in appendix A).

Quantification of DCF can similarly be divided into two fundamental aspects, namely, the amount of aerosol that is present in the atmosphere as a result of human activities and the efficiency of that aerosol at perturbing the net flux of solar radiation. Because it is the most widely available and the best-validated aerosol product that can be retrieved from satellites, optical depth δ is the most practical measure of aerosol amount for global assessments.[1] Given this, it is logical to define a radiative efficiency parameter E with units of watts per square meter per unit optical depth. Here, E is the transfer function required to convert globally distributed empirical knowledge of δ into estimates of forcing:

$$\Delta F = \delta E(\delta). \qquad (2)$$

Note that E depends on δ due to multiple scattering effects when δ is large. For the purpose of this paper, we define ΔF and E with respect to TOA forcing for solar radiation.[2]

As shown previously, E depends on a variety of aerosol and geophysical parameters.[3] For a highly absorbing aerosol, E can switch from a cooling influence to a warming influence. Positive forcing (warming) can also occur for weakly absorbing aerosols over highly reflective surfaces, such as snow and clouds.

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[1] Modeling studies have sometimes formulated aerosol climate radiative efficiency with respect to the column-integrated component mass concentration (Boucher and Anderson 1995; Hobbs et al. 1997; Houghton et al. 2001, their Table 6.4). Because this quantity is difficult to measure with in situ techniques and estimate by remote techniques, a global observational assessment would be impractical.

[2] We note that BOA forcing by anthropogenic aerosols, while not directly affecting the Earth's energy balance, is now recognized to have important effects on regional climate and the hydrological cycle (Ramanathan et al. 2001; Kaufman et al. 2002). While it is not the focus of this article, the quantification of BOA aerosol forcing would be a natural by-product of the efforts described herein.

[3] These dependencies are illustrated by the following equation, adapted from Sheridan and Ogren (1999).

$$E(\delta \ll 1) = -F_0 \Phi T_{at}^2 (1 - A_c) \omega \beta_{up} \left[(1 - R_s)^2 - \left(\frac{2R_s}{\beta_{up}} \right) \left(\frac{1}{\omega} - 1 \right) \right]. \qquad (F1)$$

Equation (F1) approximates the TOA, clear-sky forcing by a thin aerosol layer averaged over a partially cloudy region; R_0 is the mean TOA solar irradiance (i.e., 342.5 W m^{-2} in the global average), Φ is the mean daytime value of the secant of the solar zenith angle (to account for slant-path optical depth; e.g., 2 in the global average), T_{at} is the atmospheric transmission above the aerosol, A_c is fractional cloud cover, ω and β_{up} are the layer mean aerosol single scattering albedo and upscatter fraction, respectively, and R_s is the surface reflectivity.

Because the atmospheric aerosol consists of natural and anthropogenic components, one must distinguish climate forcing (i.e., DCF) from the DRE of the total aerosol. Importantly, anthropogenic aerosol is mostly found in the fine-mode portion of the aerosol size distribution.[4] Because of this, satellite observations can offer a powerful constraint on estimates of aerosol anthropogenic fraction by providing information on the fine-mode fraction of optical depth f_f, that is, the fraction of δ caused by particles smaller than about 1-μm diameter. Thus, the observation-based approach seeks information on the global distribution of fine-mode aerosol δ_f, where

$$\delta_f = \delta f_f. \tag{3a}$$

Next, this approach represents the anthropogenic aerosol δ_a as the anthropogenic fraction of the fine mode

$$\delta_a = \delta_f f_{af}. \tag{3b}$$

We can now rewrite Eq. (2) in terms of climate forcing,

$$\text{DCF} = \Delta F_a = \delta_a E_a = \delta f_f f_{af} E_a, \tag{4}$$

where E_a is the radiative efficiency of the anthropogenic aerosol.

CURRENT KNOWLEDGE. The uncertainty in DCF has variously been estimated as spanning a factor of 2–3 (Ramaswamy et al. 2001, their Table 6.11) or a factor of 5–10 (Penner et al. 2001, their Table 5.11). Either way, the uncertainty is unacceptably large for constraining climate simulations and climate change projections. Dramatic improvement is possible during the A-Train satellite era, but will require a coordinated strategy in which advanced satellite observations are effectively combined with systematic suborbital measurements to test and refine the calculations of CT/RTMs. The parameters laid out in Eqs. (2)–(4) are intended to help achieve this. Here we examine the current state of knowledge with respect to these parameters.

Table 1 summarizes a comparison of δ_i (component optical depth), δ, and δ_f among 14 CT/RTMs (Kinne et al. 2003, 2005). Table 2 summarizes the current knowledge of E over the oceans by comparing results from a number of recent studies. Figure 1 illustrates the global distribution of δ_f and δ_a. Figure 1 can be used to compare i) satellite (Fig. 1a) versus model (Fig. 1b) estimates of δ_f, ii) model estimates of δ_f (Fig. 1b) versus δ_a (Fig. 1c), and iii) model estimates of δ_a in two different seasons (Fig. 1c versus Fig. 1d).

Table 1 indicates that δ is much better constrained than are component optical depth or δ_f. For δ, the range among the models is a factor of 1.4,

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[4] Human augmentation of the global dust burden constitutes a forcing that is ignored by our method; however, this is currently thought to be small (Tegen et al. 2004).

TABLE 1. Chemical transport model intercomparison for component optical depth. Results taken from Kinne et al. (2005). These are the updated results from the AeroCom project (Kinne et al. 2003), which presently involves 18 modeling groups. Here, all models providing sufficient information to determine the selected parameters are used. There are 14 models for median and range, and 7 models for evolution. The analyzed parameter is global mean aerosol optical depth at 550 nm. Median values have been multiplied by 1000 for clarity. Component abbreviations follow: SU = sulfate, BC = black carbon, OC = organic carbon, DU = mineral dust, SS = sea salt.

	SU	BC	OC	DU	SS	Total	Fine[a]	f_f
Median	34	3.6	19	29	33	133	60	0.49
Range[b]	3.4	4.2	2.4	4.1	3.2	1.4	2.5	2.9
Evolution[c]	67%	125%	27%	150%	46%	21%	65%	30%

[a] Fine mode is defined here as the sum of components SU, BC, and OC.

[b] Range: Ratio of the second-highest value to the second-lowest value. This definition is intended to reduce the effect of possible outlier models.

[c] Evolution: For seven of the models, the latest output can be compared to output from the same model as reported in Kinne et al. (2003). The changes represent model evolution (revised emission rates, extinction efficiencies, transport or removal parameterizations, etc.), because the effects of year-to-year variations in meteorology on global mean parameters are negligible. Here we quantify model evolution as the root-mean-square change divided by the median value for each parameter, expressed as a percentage.

TABLE 2. Shown are δ, DRE, and E over the ocean from various studies. DRE is defined as the top-of-atmosphere, clear-sky,* 24-h-averaged effect on solar radiative flux due to the total aerosol. Here E is considered with respect to optical depth at 550 nm, and has a modest dependence on δ, such that values derived at different optical thicknesses are not expected to be equal, even if all intensive aerosol properties were the same. Definitions: sample fraction: fraction of area viewed by the satellite that provided valid data contributing to the global or regional average; δ: aerosol optical depth at 550 nm; DRE and E: direct aerosol radiative effect and radiative efficiency per unit δ, respectively, for solar radiation, clear-sky conditions only, and 24-h averaging.

a. Estimates based on models

Study	Notes	Region	Sample fraction	δ	DRE (W m^{-2})	E (W m^{-2} δ^{-1})
Aerosols over the global oceans						
Takemura et al. (2002)	1	Global oceans	100%	0.089	−1.9	−22
Yu et al. (2004)	2	Global oceans	100%	0.106	−2.9	−27

b. Estimates based on satellite observations

Study	Notes	Region	Sample fraction	δ	DRE (W m^{-2})	E (W m^{-2} δ^{-1})
Aerosols over the global oceans						
Loeb and Kato (2002)	3	Tropical oceans	N/a	0.12	−4.6	−38
Chou et al. (2002)	4	Global oceans	N/a	0.104	−5.4	−52
Christopher and Zhang (2002)	5	Global oceans	2.2%	0.15	−5.3	−35
Bellouin et al. (2003)	6	Global oceans	N/a	0.125	−5.1	−41
Yu et al. (2004)	7	Global oceans	N/a	0.128	−4.6	−36
		Std dev/mean		13%	8%	17%
Regional aerosols over the ocean, annual averages						
Christopher and Zhang (2002)	5	Australia outflow	18.3%	0.172	−6.7	−39
		East Asia outflow	5.0%	0.143	−6.2	−43
		North Africa outflow	2.4%	0.114	−6.3	−55
		North America outflow	2.2%	0.067	−4.0	−60
		South Africa outflow	7.4%	0.139	−6.3	−46
		Remote ocean	2.5%	0.039	−0.9	−22
		Std dev/mean		45%	45%	30%
Regional aerosols over the ocean for specific seasons						
Satheesh and Ramanathan (2000)	8	India outflow (Jan–Feb–Mar)				−30
Haywood et al. (2003a)	9	North Africa outflow (25 Sep)		1.480	−64.5	−24
Markowicz et al. (2003)	10	Asian outflow (Apr)		0.391	−10.8	−28
Li et al. (2004)	11	North Africa outflow (Jun–Jul–Aug)		0.36	−12.6	−35
		North Africa outflow (Nov–Dec–Jan)		0.16	−4.2	−26
		Std dev/mean				15%

* For CTMs, "clear sky" values refer to averages over the entire oceans with all clouds removed. For satellite observational studies, clear-sky values refer to the averages over just the cloud-free portions of the ocean where the satellite retrieval could be performed.

[1] SPRINTARS chemical transport model with five aerosol components: Takemura et al. (2002) present the sum of solar and terrestrial forcings; here, we report the forcing for solar wavelengths only from the study.

[2] GOCART radiative transfer model with five aerosol components.

[3] Solar flux from CERES on the TRMM satellite and optical depth at 630 nm from VIRS on TRMM: Optical depth and radiative efficiency are converted here to 550-nm wavelength by assuming an Ångström exponent of 1.0.
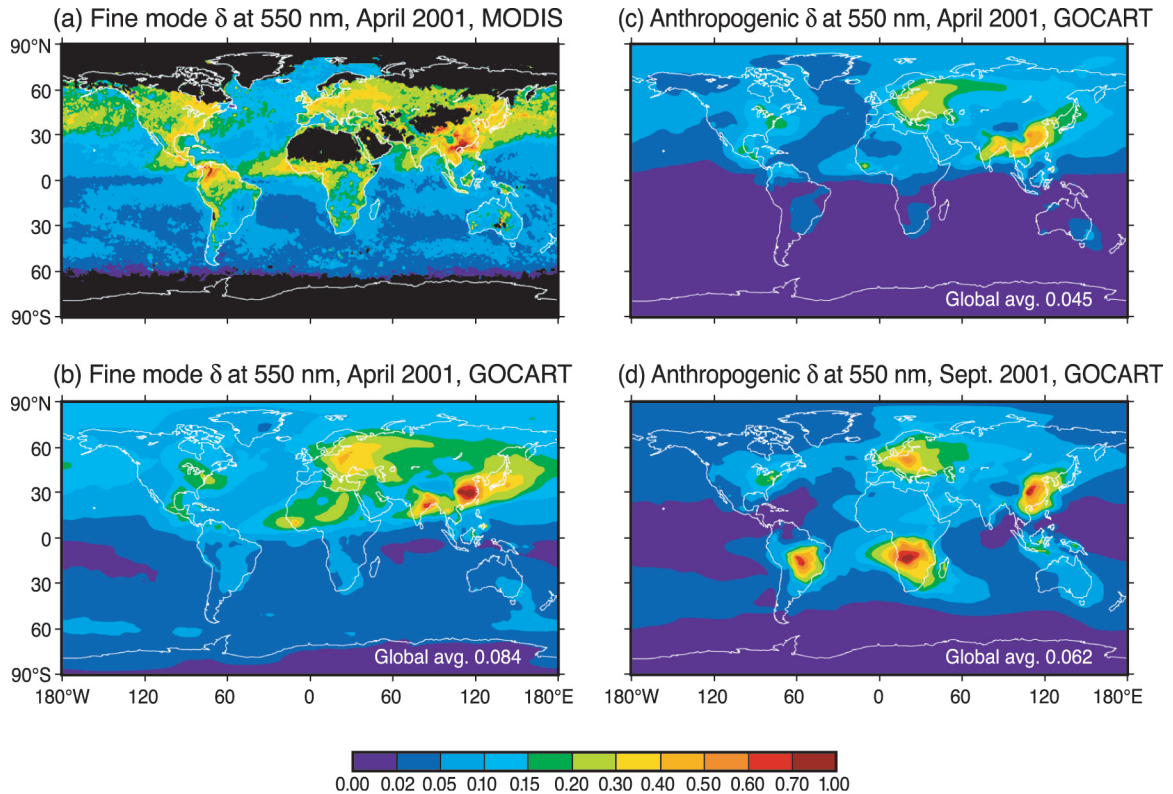
FIG. 1. Satellite and model data for fine-mode aerosol and anthropogenic aerosol. Satellite data are the level 3, version 4 MODIS product with 1° × 1° resolution. Model data are from the GOCART chemical transport model with 2° × 2.5° resolution. Fine-mode optical depth from (a) MODIS and (b) GOCART for Apr 2001. Anthropogenic aerosol optical depth from GOCART for (c) Apr and (d) Sep 2001.

[4] Radiative transfer model calculation based on aerosol optical depth and optical properties retrieved from the SeaWIFS satellite.

[5] Solar flux from CERES/*Terra* satellite and aerosol optical depth from MODIS/*Terra* satellite: Christopher and Zhang (2002) report instantaneous forcings that are converted here to diurnal averages by dividing by 2. This is the globally averaged conversion factor as determined by radiative transfer calculations, but it will vary somewhat by latitude, causing error in the values reported here for the regional aerosols.

[6] Analysis of POLDER-1 satellite data: In the paper only forcings were reported, not optical depths, but the latter were calculated as part of the study and are provided here.

[7] Aerosol optical depth data from MODIS/*Terra* satellite assimilated into GOCART with five aerosol components.

[8] Solar flux from CERES/TRMM satellite and aerosol optical depth from AERONET station at Kaashidhoo Climate Observatory, Maldives, India Ocean: Radiative efficiency is computed as the slope of the regression. Data are from the winter monsoon period (Jan–Mar 1998 and 1999), and capture continental outflow from Indian and Southeast Asia.

[9] A single, heavy dust event on 25 Sep 2000: A radiative transfer model, constrained by satellite flux retrievals (CERES/*Terra*), airborne radiometric and in situ measurements, and surface radiometers, is used to calculate aerosol optical depth and top-of-atmosphere flux. Haywood et al. (2003a) report instantaneous values. Here, additional radiative transfer calculations have been performed to convert the results of 24-h mean forcing. The conversion factor was 1/3.6.

[10] Solar flux from CERES/*Terra* satellite and aerosol optical depth from a shipboard sunphotometer: The average is from over six nearly cloud-free days in the Sea of Japan during April 2001. The radiative transfer model is used to convert instantaneous measurements to a 24-h average. The conversion factor was approximately 1/2.0.

[11] Regional study using solar flux from CERES/*Terra* and aerosol optical depth from MODIS: Instantaneous radiative efficiency is determined by regression and converted to a 24-h average using a radiative transfer model.

and model evolution (over a time period of about 1 yr) has resulted in an rms change of 21%. In contrast, the range for component optical depths are factors of 2–4, and model evolution has resulted in changes greater than 50% for three of the five components. This situation likely reflects the fact that observational constraints on global mean δ (largely from satellites) are currently much stronger than constraints on the global abundance of individual aerosol components (largely from land-based, in situ, chemical measurements). This implies, in turn, that the development and validation of satellite methods for detecting aerosol type is critical to advancing current knowledge.

Fine-mode optical depth δ_f can be retrieved from satellites, but it is not clear whether this retrieval has impacted CT/RTMs. The range among models for this parameter is a factor of 2.5, and recent model evolution has resulted in an rms change of 65% (Table 1). These values are similar to the uncertainties in component optical depth. An important factor here is that the accuracy of the satellite retrieval of δ_f is not presently known, and, for the MODIS product at least (cf. Fig. 1a and 1b), there are indications of strong discontinuities at land–ocean boundaries, which may be artifacts.

The anthropogenic fraction of fine-mode optical depth f_{af} has yet to be carefully investigated, even though i) it can readily be calculated by modern CTMs and ii) a great deal of chemical data exist that could potentially be brought to bear upon its assessment. Comparing Figs. 1b and 1c, we see that f_{af} for April 2001 was 0.54 (i.e., 0.045/0.084), according to the GOCART model.

Many studies have investigated aerosol radiative efficiency E.[5] However, it is not possible in all cases to compare these values because of differences (and sometimes ambiguities) in definition. Table 2 provides several current examples in which the quantities have all been adjusted to correspond to TOA shortwave diurnal mean forcing during clear-sky conditions over the ocean, normalized with respect to δ at 550 nm. For the studies that report E over the global oceans, values range over about a factor of 2. Because all of these studies were attempting to derive the same quantity, this range reflects the methodological uncertainty. Recent studies have begun to

investigate this. For example, Zhang et al. (2005) show a 20% range in satellite-retrieved E over the ocean, depending on the selection of the angular distribution model, and Loeb and Manalo-Smith (2005) show a 24% effect on derived E associated with changing from one cloud-clearing scheme to another.

Discrepancies among the regional studies shown in Table 2 reflect a combination of methodological uncertainty and true variability. The regional variability of E appears to be substantial, according to the results of Christopher and Zhang (2002) shown here in Table 2b (where the method used is constant across regions). The causes of this variability cannot readily be assessed because the required ancillary data are not available. Note that the satellite estimates of E use the ratio of DRE (derived from broadband flux retrievals) to δ (retrieved from narrowband radiances), but provide little diagnostic information on the factors controlling E [see Eq. (F1) in footnote 3]. This highlights an important role for in situ measurements: assessing the column mean values of these controlling factors in selected locations would greatly aid in interpreting the satellite retrieval of E.

We can summarize this section by considering the current uncertainty in clear-sky DCF. Based on Eq. (4) this could be estimated as the propagated sum of the uncertainties associated with the global mean values of each of the four parameters (i.e., δ, f_f, f_{af}, E_a), plus the uncertainties associated with their spatiotemporal covariations. The space–time covariations have neither been assessed nor has there been a global-scale assessment of f_{af}. Therefore, uncertainties associated with these aspects of the problem are unknown. Uncertainty in the global mean value of δ_f (the product of δ and f_f) is 2.5X as indicated by the range among current models shown in Table 1. (Additional uncertainty could arise from errors in the methods or assumptions that are common to all the models.) The global mean value of E_a has not been assessed, but its uncertainty is unlikely to be smaller than the 1.5X uncertainty indicated by the range among studies of E over the global oceans (Table 2a). Propagating these parameter uncertainties implies a 3X uncertainty in their product. Thus, the lower limit on uncertainty in DCF is about 3X and the upper limit is not known. Although incomplete, this analysis is consistent with the uncertainty estimates from IPCC (Ramaswamy et al. 2001, their Table 6.11; Penner et al. 2001, their Table 5.11). Given the rapidly evolving measurement and modeling capabilities, the research community would appear to be poised to improve this situation dramatically.

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[5] Most previous studies refer to this quantity as "forcing efficiency." However, we recommend reserving the word "forcing" for externally imposed changes. The term "radiative efficiency" refers to either DRE or DCF.

QUANTIFICATION STRATEGY. Despite considerable effort, the reduction of uncertainty in the forward calculations[6] of aerosol climate forcings has proven, to date, to be an elusive goal. In large part, this reflects a rapidly expanding list of forcing mechanisms—the discovery of new ways that anthropogenic aerosols may be perturbing the climate system. A more basic dilemma, however, is the lack of rigorous, empirical knowledge of the global distribution of climatically relevant aerosol properties. Progress on this front is relatively straightforward and is the key to improved quantification of DCF. The "Conceptual framework" section identified a proposed set of relevant parameters. This section, along with Table 3, outlines an orderly process for improving knowledge of these parameters and applying that knowledge to the scientific questions surrounding DCF. Because no single research group, agency, or country could implement this strategy by itself, we offer it as a proposed framework for cooperative research.

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[6] Forward calculations are based on aerosol science and can be distinguished from inverse calculations in which aerosol forcings are derived from the temperature record (Anderson et al. 2003).

TABLE 3. The four-step approach to the quantification of clear-sky DCF is shown.

Step 1: Test and refine knowledge of aerosol optical depth δ

Goal: Global map of $\delta(x, y, t)$

Task i: Compare model–satellite, satellite–satellite, and satellite–suborbital to establish uncertainties

Task ii: Assimilate satellite data for refined global map

Step 2: Test and refine knowledge of radiative effect* DRE

Goal: Global mean, clear-sky DRE and its regional/temporal variations

Required information: Global map of $E(x, y, t)$ to be combined with map of $\delta(x, y, t)$ from step 1

Task 0: Intercompare radiative transfer codes

Task i: Invoke a CTM-based map of $E(x, y, t)$

Task ii: Assess $E(x, y, t)$ empirically from satellite retrievals and suborbital validation experiments

Task iii: Assimilate satellite data for refined global map of $E(x, y, t)$

Step 3: Test and refine knowledge of climate forcing** DCF

Goal: Global mean, clear-sky DCF and its regional/temporal variations

Required information: Global maps of $f_f(x, y, t)$, $f_{af}(x, y, t)$, and $E_a(x, y, t)$ to combine with global map of $\delta(x, y, t)$ from step 1

Task i: Invoke CTM-based maps of $f_f(x, y, t)$, $f_{af}(x, y, t)$, and $E_a(x, y, t)$

Task ii: Assess fine-mode fraction $f_f(x, y, t)$ empirically from satellite retrievals and suborbital validation experiments

Task iii: Assess anthropogenic fraction of the fine mode f_{af} empirically in selected regions from in situ chemical measurement programs and process studies

Task iv: Assess E_a empirically from satellite retrievals and suborbital validation experiments in regions dominated by fine-mode anthropogenic aerosols

Task v: Assimilate satellite data for refined global maps of $f_f(x, y, t)$, $f_{af}(x, y, t)$, and $E_a(x, y, t)$

Step 4: Attribute climate forcing to aerosol components

Goal: Quantify climate forcing due to specific anthropogenic sources

Required information: Chemical composition and source of the aerosol that is causing DCF

Task i: Take CTM-based estimate using assimilation of results from steps 1, 2, or 3

Task ii: Assess chemical composition empirically as part of an in situ satellite underflight program

Task iii: Assimilate underflight information for a refined global estimate

*Radiative effect: Portion of the energy budget with an atmospheric constituent (W m^{-2}).

**Climate forcing: Imposed change in energy balance by human activity or by some specific forcing agent (W m^{-2}).

Clear sky. Table 3 summarizes our proposed stepwise approach to the global quantification of clear-sky DCF. The guiding philosophy is that, at each point in the development and refinement of the research effort, existing knowledge should be consolidated into an overall framework. Over time, this knowledge will become increasingly grounded in observations. At each stage, however, it is important to assess uncertainties, assess where we stand with respect to the key scientific questions, and assess what information needs are most critical for further progress. The various steps can, and should, be undertaken in parallel. For example, initial results from step 1 will immediately enable task *i* of the remaining three steps to be accomplished. Thereafter, progress in all four steps can proceed in an iterative fashion (see information online at www.atmos.washington. edu/~cheeka/DCF/DCF.html for a detailed description of each step).

Strategic challenges. Here we highlight the key challenges that must be confronted in implementing the proposed DCF strategy. This section is intended to give the reader a feeling for the complexity of the task, the need for explicit coordination, and the need, in some cases, for departing from common research practice.

CLOUDS. The variety and ubiquity of clouds presents an enormous challenge to observational quantification of clear-sky DCF. Inadequate cloud-clearing causes cloud contamination or an upward bias in the satellite view of the clear-sky aerosol. Excessive cloud clearing causes downward bias if thick aerosols are mistakenly identified as cloud. A general problem is that advanced satellite retrievals often require large, cloud-free areas. This is illustrated by the combined CERES–MODIS retrieval of clear-sky radiative efficiency (Christopher and Zhang 2002). When the high-resolution, cloud-clearing criteria from MODIS were applied to the CERES pixels (30 km at nadir), only 2.2% of those pixels over the ocean were available for the combined analysis (Table 2b).

GLOBAL COVERAGE. The fractional coverage (spatial and temporal) for most satellite products is surprisingly limited. For example, i) sun-synchronous polar-orbiting satellites measure at only one time of day, ii) retrievals of δ for passive sensors are restricted to regions with appropriate cloud and surface characteristics, iii) retrievals of δ for spaceborne lidar are restricted to essentially a zero swath width, iv) retrievals of more advanced products (e.g., E and f_f) are generally possible only for high values of δ, and v) advanced retrievals that involve more than one sensor are only possible in regions where these sensors overlap and where the data quality requirements of all sensors are met.

NEED FOR MODELS. The limited coverage offered by even the most "global" of aerosol observational methods (i.e., satellites) has two implications. i) The calculation of global forcings (DRE and DCF) inherently requires a model-generated global map of aerosol and geophysical characteristics. Satellite observations do not replace model-generated maps; rather, they serve to refine and constrain them. ii) All satellite data must be assessed with respect to whether it is representative. Assessing whether satellite observations constitute a biased or a representative sample requires that they be put in the context of a complete global map of aerosol and geophysical characteristics. To provide this context, a global aerosol modeling capability should be incorporated into the observational strategy.

SATELLITE VALIDATION. Satellite retrieval products cannot be considered reliable until they have been independently validated. The validation strategy should achieve the following: i) test whether the parameter can, in fact, be retrieved to within the theoretically estimated uncertainty, ii) provide a new, empirically based estimate of retrieval uncertainty, and iii) establish the limits of retrieval validity in terms of the required signal magnitude and required viewing conditions (underlying surface reflectivity, glint angle, absence of thin or scattered clouds, etc.). Validation of δ is relatively mature. A worldwide network of intercalibrated sun photometers has been established for this purpose (Holben et al. 1998, 2001) and numerous intensive campaigns have provided additional high-quality tests (Livingston et al. 2003; Schmid et al. 2003; Redemann et al. 2005). In contrast, there are no long-term programs or intensive campaigns designed to test the satellite retrievals of E or f_f. Therefore, a major challenge to the observational quantification of DCF is to design and implement a program of suborbital measurements capable of validating these quantities (see the "Systematic, suborbital underflights" section).

INTEGRATION. Going beyond validation, the application of satellite data to DCF requires integration with the knowledge gained from suborbital approaches. Such integration will be difficult unless a subset of suborbital measurements is carefully coordinated

with the satellite observations. Existing measurement programs, designed for other purposes, are unlikely to suffice. For example, traditional, ground-based, 24-h-averaged measurements of aerosol chemistry are unlikely to be useful for interpreting instantaneous satellite measurements that apply either to the entire column (as in passive retrievals) or to a zero-width swath that rarely passes close to the surface station (as in active lidar retrievals).

Cloudy sky. In general, the direct radiative effect of aerosols when clouds are present in an atmospheric column is close to negligible (Haywood and Shine 1995; Haywood et al. 1997). An important exception, however, is the case of absorbing aerosol located above cloud, which could potentially exert a large warming influence (Haywood et al. 1997; Jacobson 2002; Takemura et al. 2002; Keil and Haywood 2003). Unfortunately, because clouds are both highly reflective and extremely variable, passive satellite methods are unable to measure aerosols in the presence of clouds. CALIOP, unlike any previous satellite sensor, will be able to assess the frequency and amount of aerosol above boundary layer clouds. However, the climatic effect of this aerosol depends almost entirely on its single scattering albedo ω, which CALIOP cannot readily determine. Thus, a high-leverage use of suborbital measurements will be to determine ω for a representative subset of the aerosol-above-cloud cases detected by CALIOP.

A-TRAIN CONTRIBUTION. In essence, quantifying DCF is a problem of global mapping of key aerosol and geophysical parameters. The A-Train (Stephens et al. 2002) is a constellation of six polar-orbiting satellites that offers an unprecedented opportunity to improve empirical knowledge of these parameters and their covariations. For aerosol sensing, it features improved versions of instruments with a long heritage (e.g. MODIS, POLDER, TOMS), as well as a lidar instrument for vertical aerosol profiling (CALIOP). Spaceborne lidar was the principal instrumental recommendation of a National Academy report on research needs for the improved quantification of aerosol forcing (Seinfeld et al. 1996).

Appendix B lays out the capabilities of key A-Train sensors with respect to DCF. The real power of the A-Train mission comes with the synergies among these sensors. These involve both overlapping and complementary capabilities in terms of retrieved quantities, sensitivity, resolution, and coverage. An integrated strategy for the A-Train would begin with intercomparisons of common retrieval products

and proceed to the development of joint retrieval products. Satellite intercomparisons (e.g. Abdou et al. 2005) provide a robust way to test for unanticipated retrieval error. Such tests are likely to be far more comprehensive (although less definitive) than suborbital validation experiments. Potentially important joint retrievals are i) retrievals of E using aerosol radiative forcing from CERES combined with δ from MODIS, OMI, or POLDER, ii) retrievals of ω using OMI measurements combined with aerosol-layer-height data from CALIOP, iii) vertical profiles of fine- and coarse-mode extinction by constraining CALIOP lidar inversions with δ and f_f data from MODIS and/or POLDER, and iv) developing global maps of component optical depth in terms of a simple, three-component scheme (dust, sea salt, and fine mode), which could involve all of the sensors. (Both the intercomparison tests and joint retrievals are described in more detail online at www.atmos. washington.edu/~cheekca/DCF/DCF.html.)

SYSTEMATIC, SUBORBITAL UNDER-FLIGHTS. The PARAGON initiative recognizes that a "program of sustained airborne measurements is a key ingredient of an integrated observing system" for aerosols, particularly with regard to the validation and demonstration of coherence between model predictions and remote sensing observations (Diner et al. 2004b). An online version of this paper provides a detailed design for a program of A-Train underflights in the eastern United States to support the quantification of DCF. This is intended as a paradigm for other regions around the globe (see Fig. 1d), where anthropogenic pollution is likely to have a large effect on radiative balance. The basic approach is given here.

Coupling to surface measurement programs. An enormous amount of surface-based aerosol data is routinely acquired via long-term observing programs (Kahn et al. 2004). Especially relevant to quantification of DCF are in situ measurements of aerosol optics and chemistry (e.g. Delene and Ogren 2002), sun-photometer measurements of δ (e.g. Holben et al. 1998, 2001), and lidar measurements of aerosol vertical distribution (e.g. Welton et al. 2001; Bösenberg et al. 2003).

To make optimal use of these programs, two additional facts must be considered. First, the CALIOP ground track will rarely pass close enough to any given station to permit a direct comparison to the surface measurements (see Fig. 2). Second, aerosol microphysical and optical properties that are

retrieved from remote observations at the surface (rather than directly measured) are themselves in need of validation (Haywood et al. 2003b; Kahn et al. 2004). Thus, airborne measurements can serve two complementary objectives: i) to validate the inversion-based remote retrievals from surface stations and ii) to validate the satellite retrievals whenever flights along the orbital track can be arranged. In this sense, flight programs can provide a bridge between the continuous data at these fixed stations and the global, but intermittent, data from the A-Train.

Key objectives. Flight logistics should be designed around the primary goal of acquiring a statistically significant number of samples that are collocated with the zero-width-swath CALIOP sensor. For each region, this sample set should include the full spectrum of measurement conditions in terms of aerosol type, concentration, vertical structure, surface reflectivity, meteorological regime, and season. This would seem to dictate a routine flight program, sustained over a large portion of the CALIOP 3-yr lifetime. The choice of measurements should be guided by the need to establish the ability of the A-Train sensors to retrieve the common metrics defined herein and to connect those retrieved products to aerosol composition, including aerosol condensed water content.

In situ measurements are needed for three fundamental purposes: i) zero-order validation of coherence between retrieved and in situ versions of extinction, fine-mode fraction of extinction, and fine-mode fraction of optical depth (We must prove that the in situ, satellite, and surface remote platforms are measuring the same phenomenon in order to know that these datasets are indeed integratable.); ii) observational data on variables that are required for the calculation of radiative efficiency E [Eq. (F1) footnote 3], but are difficult or impossible to measure remotely (These include, for example, total and submicron single scattering albedo, total and submicron backscatter fraction, and the portion of total and submicron scattering that is due to aerosol hydration at ambient relative humidity. In addition, measurements of ω will be critical for validating satellite retrievals of this quantity.); and iii) basic aerosol chemical measurements for connecting
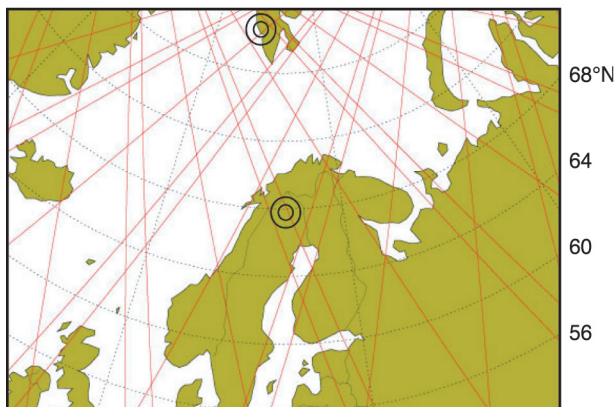
(a) 16 days, Eastern U.S.
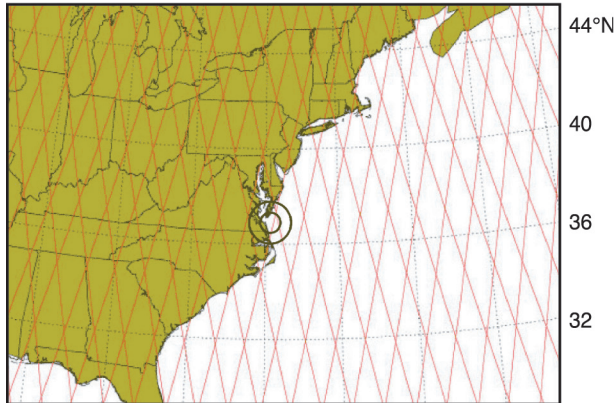
(b) 16 days, Central Europe

(c) 3 days, Arctic

FIG. 2. Approximate ground tracks of the CALIPSO lidar. Circles indicate 50- and 100-km-radius regions around the central point. Sixteen-day repeat tracks are shown for (a) the eastern United States with circles centered on the CERES Ocean Validation Experiment (COVE) station at the Chesapeake Bay Lighthouse (36.8°N) and (b) central Europe with circles centered on Leipzig, Germany (51.5°N). The frequency of orbital tracks passing within these distances from a fixed ground station varies with latitude. This is highlighted by (c) 3 days of ground tracks over the Arctic with circles centered on Kiruna, Sweden (67.8°N), and Ny Ålesund (Spitzbergen), Norway (78.9°N). While most anthropogenic aerosol forcing is thought to exist in the Tropics and midlatitudes, statistically significant amounts of correlative suborbital samples can be acquired most readily at high latitudes.

retrieval products to CTMs (especially important for the assimilation/inversion method of deriving aerosol source strengths; in addition, chemical measurements will provide guidance in deriving the f_{af}).

Remote measurements, specifically, airborne sun photometry and shortwave flux radiometers, would greatly enhance the flight program by i) determining ambient extinction by an independent method (differential sun photometry) to check the sampling efficiency and the hydration correction involved with the in situ measurements, ii) providing a well-calibrated measurement of E for the lower troposphere based on optical depth and shortwave flux divergence, and iii) measuring the spatial variability of optical depth, especially in the vicinity of the fixed surface stations and of the CALIOP orbital track.

Timeline. The *Aqua, Aura,* and PARASOL satellites are in orbit and CALIPSO is scheduled for launch in the fall of 2005. The POLDER and CALIOP design lifetimes run through 2007 and 2008, respectively. Therefore, the ideal time period for acquiring correlative, suborbital measurements will be 2006–07, with continuing value through 2008.

CONCLUSIONS. We have presented a conceptual framework for integrating observations and models with respect to the problem of DCF by anthropogenic aerosols. This framework is based on four fundamental parameters, which we propose as common metrics through which the integration could logically take place, namely, the midvisible optical depth δ, radiative efficiency per unit optical depth E, fine-mode fraction of of optical depth f_f, and anthropogenic fraction of the fine mode f_{at}. The choice of these parameters is based on a blend of historical, practical, and theoretical considerations, with an emphasis on practicality with respect to satellite mapping. For example, f_f provides a practical approach to estimating the anthropogenic fraction of optical depth.

Current global knowledge of these parameters (Tables 1, 2, and Fig. 1) leads us to estimate that uncertainty in clear-sky DCF is at least a factor of 3, but could be dramatically reduced using new satellite measurements and a coordinated research strategy. A four-step plan for shifting the current model-based estimates to an increasingly empirical basis is outlined. The plan emphasizes the need for a carefully designed, systematic program of suborbital, airborne measurements to provide a means of validating the advanced aerosol retrievals and to provide the ancillary data that are required for interpreting the satellite products.

These recommendations are offered as an initial framework—subject to improvement over time—to facilitate coordination among scientists interested in DCF with respect to the wealth of observations afforded by the A-Train constellation of satellites (for greater detail see www.atmos.washington.edu/~cheeka/DCF/DCF.html). The critical time frame for the correlative measurements is from the present through 2008.

Implementing the proposed strategy will require a high level of scientific cooperation among research groups, agencies, and nations. In addition, it will require departing in significant ways from current plans and common practice. For example, satellite studies of the global aerosol rarely report sample coverage or estimate sampling bias. Suborbital research programs have yet to incorporate in a significant way the goals of i) in situ validation of the inversion-based retrievals from remote measurements at surface stations, ii) establishing the ability of satellite sensors to retrieve f_f and E, or iii) routine airborne sampling collocated with a zero-width-swath spaceborne lidar. These, and other recommendations described herein, are the logical consequences of adopting the PARAGON philosophy, wherein the need for the integration of knowledge is built into the observing and analysis strategy.

APPENDIX A: DEFINITIONS
a. Acronyms

ADEOS	Advanced Earth Observing Satellite
AeroCom	Aerosol model and observation intercomparison project (described in Kinne et al. 2005)
A-Train	Constellation of six satellites (described in "A-Train contribution" section and appendix B
BOA	Bottom of atmosphere
CALIPSO	Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations (scheduled to launch in the fall of 2005)

CALIOP	Cloud and Aerosol Lidar with Orthogonal Polarization (CALIPSO lidar)
CERES	Clouds and the Earth's Radiant Energy System, deployed on the Terra (December 1999–present) and Aqua (May 2002–present) satellites
CTM	Chemical Transport Model
CT/RTM	Chemical Transport Radiative Transfer Model
DCF	Direct climate forcing (by anthropogenic aerosol)
DRE	Direct radiative effect (of the total aerosol)
GOCART	Goddard Chemistry Aerosol Radiation and Transport
IPCC	Intergovernmental Panel on Climate Chnage
MODIS	Moderate Resolution Imaging Spectroradiometer, deployed on the Terra (December 1999–present) and Aqua (May 2002–present) satellites
OMI	Ozone Monitoring Instrument (successor to TOMS, launched on Aura in July 2004)
PARASOL	Polarization and Anisotropy of Reflectances for Atmospheric Sciences coupled with Observations from a Lidar (launched 18 December 2004)
PARAGON	Progressive Aerosol Retrieval and Assimilation Global Observing Network
POLDER	Polarization and Directionality of Earth's Reflectances, to be flown on PARASOL
RMS	Root-mean-square
RIM	Radiative Transfer Model
SeaWIFS	Sea-viewing Wide Field-of-view Sensor
SPRINTARS	Spectral Radiation-Transport Model for Aerosol Species
TOA	TOP of atmosphere
TOMS	Total Ozone Mapping Spectrometer
TRMM	Tropical Rainfall Measuring Mission
UV	Ultraviolet
VIRS	Visible Infrared Scanner

b. Key symbols

δ	Aerosol optical depth (at 550 nm)
δ_a	Optical depth of the anthropogenic aerosol
δ_f	Optical depth of the fine-mode aerosol
E	Radiative efficiency (W m^{-1} per unit δ at 550 nm)
E_a	Radiative efficiency of the anthropogenic aerosol
f_{af}	Anthropogenic fraction of δ_f
f_f	Fine-mode fraction of δ
λ	Climate sensitivity (in Eq. 1) or Wavelength (elsewhere)
ω	Single scattering albedo or ratio of scattering to extinction

APPENDIX B: KEY A-TRAIN SENSORS FOR QUANITIFICATION OF DCF.

MODIS on Aqua. Multiwavelength radiances are measured over a 2,300-km-wide swath with 1-km nadir resolution or better. Aerosol retrievals of δ and f_f are based on six wavelengths over the oceans and two wavelengths over land. Additional relevant products include land surface typing, which is useful for determining surface reflectivity, and numerous cloud properties, including fractional coverage. *Aqua* was launched on 4 May 2002.

CERES on Aqua. Broadband solar and terrestrial radiances are measured from limb to limb with 20-km nadir resolution. Twin instruments are used to optimize angular sampling over the hemisphere of scattered and emitted radiation. The retrieval of TOA solar flux under clear-sky conditions can be compared to the expected flux for an aerosol-free atmosphere. The difference constitutes an estimate of DRE (Haywood et al. 1999; Loeb and Kato 2002; Christopher and Zhang 2002). When combined with optical depth data, this method provides an estimate of E (Table 2).

OMI on Aura. This is a hyperspectral UV and visible imaging radiometer with 13 km × 24 km nadir resolution over a swath of 2600 km. Aerosol retrieval products are δ and ω in the near-UV (Torres et al. 1998, 2002, 2005). If the data on ω can be extended spectrally to the visible, it will provide a critical constraint on radiative efficiency E [see Eq. (F1), footnote

3]. Because the retrieval is sensitive to the height of the aerosol layer, a combined OMI–CALIOP retrieval has great potential. *Aura* was launched on 8 July 2004 for a 5-yr mission.

POLDER on PARASOL. This is a nine-channel wide-field-of-view imaging polarimeter. Its nadir resolution is 6 km and its swath is 1400 km. Along-track view angles are ±51°, which provide multiangle viewing of each point on the surface over a time period of ~300 s. Previous versions were flown on ADEOS-1 (November 1996–June 1997) and ADEOS-2 (December 2002–October 2003), and were used to estimate clear-sky DRE over the oceans (Bellouin et al. 2003) and the global distribution of fine-mode aerosol (Tanré et al. 2001). It will be used to retrieve δ and δ_f over the ocean and δ_f over land (Herman et al. 2005). PARASOL was launched on 18 December 2004 for a 2-yr mission.

CALIOP on CALIPSO. Measurement of laser backscatter at two wavelengths (532 and 1064 nm), with polarization sensitivity at 532 nm, will provide data on aerosol vertical structure and properties. The ability to measure above low clouds and beneath thin cirrus will allow the first global-scale constraints on cloudy-sky aerosol forcing. CALIOP provides data at nadir only (see Fig. 2 for an example of orbital track spacing), which imposes special logistical requirements for collocated suborbital correlative measurements. While CALIOP can detect boundary layer clouds at its maximum resolution (30 m vertical and 300 m along track), aerosol retrievals require considerable averaging and will have typical resolutions of about 120 m vertically and 10 km along track. Aerosol retrieval products are δ, and vertical profiles of aerosol backscatter, extinction, and depolarization. CALIPSO is scheduled for launch in the fall of 2005 for a 3-yr mission.

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