Calculation of the global land surface energy, water and CO₂ fluxes with an off-line version of SiB2

C. Zhang,¹ D. A. Dazlich,¹ D. A. Randall,¹ P. J. Sellers,² and A. S. Denning¹

Abstract. Global land surface fluxes of energy and CO₂ have been simulated using an off-line version of a biosphere-atmosphere model, SiB2, forced with analyzed or observed atmospheric boundary layer mean potential temperature, water vapor mixing ratio, and wind, surface downward solar and thermal radiation, and precipitation. The off-line model is called SiBDRV. Soil and vegetation boundary conditions were specified from satellite data and other sources. The European Centre for Medium-Range Weather Forecasts (ECMWF) data assimilation system products were used to derive the atmospheric and radiative forcings. Precipitation was based on station observations. The SiBDRV results were compared with corresponding simulation results produced by the Colorado State University general circulation model (CSU GCM), with the ECMWF assimilation system output and with observations. Differences between the surface energy budget components and the surface climatology produced by SiBDRV and the ECMWF assimilation system are due to differences in the land surface parameterizations between the two models. SiBDRV produced lower surface latent heat fluxes and larger sensible heat fluxes than the ECMWF data assimilation, partly due to large canopy resistant term explicitly formulated by SiB2 and possible precipitation differences between the SiBDRV precipitation forcing and the ECMWF data. Differences between the SiBDRV and the CSU GCM results are due to the different climates associated with the ECMWF assimilation system output, which is strongly constrained by assimilated observations, and by the CSU GCM, which is run in pure simulation mode. More specifically, the major reasons for the surface energy and CO₂ budget differences between the SiBDRV and the GCM are greater incoming solar radiation in the GCM and differences in the precipitation patterns. The simulated global annual carbon uptake by the terrestrial biosphere is similar in SiBDRV and the GCM. The annual gross primary productions of SiBDRV (116 Gt) and the GCM (113 Gt) agree well with other studies, using either ecological process models or empirical regression models. SiBDRV takes up 10 and 5% more CO₂ than the GCM in January and July, respectively. The seasonally varying land surface CO₂ fluxes estimated by the SiBDRV and the GCM both compare reasonably well with the results of other calculations.

1. Introduction

Land surface processes affect climate mainly through the surface-atmosphere exchanges of energy, momentum, and CO₂ across the atmospheric boundary layer. The land surface energy and CO₂ budgets are still uncertain, however, due to the complexity of the land surface processes and their interactions with the climate system. A comparison of 19 atmospheric general circulation models (GCMs) concluded that variations of the surface-energy fluxes in response to prescribed 4 K sea surface temperature changes varied greatly from model to model, mainly due to differences in the simulated hydrological cycles and the parameterizations of longwave radiation exchange, clouds, and boundary layer processes [Randall et al., 1992]. These differences can arise in part from differences in the parameterization of land surface processes. The 19 models used in the study mentioned above incorporated a wide range of parameterizations, from simple bucket models to relatively sophisticated vegetation parameterizations.

The biosphere-atmosphere transfer scheme (BATS) [Dickinson, 1984] and the simple biosphere model (SiB) [Sellers et al., 1986] were developed to explicitly represent the effects of terrestrial vegetation and soil properties on surface-atmosphere fluxes. A new version of SiB (SiB2), described by Sellers et al. [1996a], includes more realistic canopy photosynthesis and stomatal control processes; vegetation transpiration is linked to carbon assimilation, and the rates of CO₂ exchange and water flux between the land surface and the atmosphere are calculated simultaneously. Shao et al. [1994] reviewed more land surface models involving intercomparison studies among the models.

Sato et al. [1989] published results from simulations using the first version of SiB (SiB1), as incorporated into the Center for Ocean-Land- and Atmosphere version of the U.S. National Meteorological Center’s GCM. They showed that the surface fluxes of energy, momentum, heat, and moisture were in reasonable agreement with the available field observations. Recently, Randall et al. [1996] reported the results of simulations using SiB2 coupled to the Colorado State University
eral circulation model (CSU GCM) coupled with exactly the same version of SiB2 and with the surface fluxes provided as part of the original ECMWF assimilation system output. The results are also compared with other available data sets. Figure 1 summarizes this strategy.

The differences between the land surface energy budget components and surface climateology produced by SiBDRV and the ECMWF assimilation system can be attributed to the differences between the land surface used in the two models. On the other hand, discrepancies between the SiBDRV results and the GCM simulation may be related to climatic differences experienced by the land surface.

This is a new attempt to reasonably estimate the land surface fluxes at the global scale, using a comprehensive land surface model (SiB2) forced by observations. Although some comprehensive field measurements for the surface energy and turbulent fluxes are now available at several locations, such as FIFE, HAPEX, Boreal Forest, surface energy fluxes at continental scale are not available at the present time. The surface fluxes also vary spatially. Thus a simulation such as SiBDRV can provide a practical way to obtain surface energy, water, and CO2 fluxes information at the global scale and for a period of long time.

In section 2 we briefly describe SiB2. The boundary and atmospheric forcing data sets are discussed in section 3. Section 4 describes the experiments. In section 5 we analyze and discuss the results. A summary is given in section 6.

2. A Brief Description of SiB2

SiB2 includes one canopy layer. There are three soil layers: a surface soil layer, a root zone, and a deep soil layer. A canopy photosynthesis submodel [Collatz et al., 1990, 1991, 1992; Sellers et al., 1992] has been incorporated, with a prognostic stomatal conductance and an explicit calculation of the photosynthetic CO2 flux between the atmosphere and the land surface. A new snow parameterization has also been included.

SiB2 has 10 prognostic variables: the vegetation canopy temperature ($T_c$), surface soil temperature ($T_s$), and deep soil temperature ($T_d$); water intercepted on the canopy ($M_{c,w}$) and on the surface soil layer (puddles, denoted by $M_{p,w}$); snow intercepted on the canopy ($M_{c,s}$) and on the surface soil layer ($M_{s,s}$); and surface soil wetness ($W_s$), root-zone soil wetness ($W_{s,z}$), and deep soil wetness ($W_s$).

Each of the three temperatures changes as a function of the surface energy fluxes. Following Sellers et al. [1996a], the sensible heat, moisture, and CO2 fluxes are formulated in terms of their potential differences divided by aerodynamic resistances. Three components of evapotranspiration are included in SiB2: direct evaporation from intercepted water on the canopy and ground, transpiration of water vapor regulated by the stomatal conductance of the canopy, and evaporation from the surface soil layer. Temperatures for dry and wet surfaces are assumed to be the same.

The land surface albedo is determined by applying a two-stream radiative transfer model to the canopy layer [Sellers et al., 1986]. Surface downward solar and thermal radiation are each partitioned into direct and diffuse visible (wavelength less than 0.72 μm), direct and diffuse near infrared (0.72–4.0 μm), and thermal infrared (wavelength greater than 4.0 μm) bands.

Photosynthesis is the source of biomass production over land and has been identified as one of the major CO2 sinks for the atmosphere [e.g., Schlesinger, 1991]. As described by Sellers et

![Diagram illustrating the relationships among the simulations and data used in this study. SiBDRV is driven by a combination of European Centre for Medium-Range Weather Forecasts (ECMWF) analyses, ECMWF assimilation system output, and precipitation observations. The results produced by SiBDRV are compared with the ECMWF surface fluxes, with observations, and with simulations produced by the Colorado State University general circulation model (CSU GCM), which have the exactly the same SiB2 land surface model.](image-url)
Table 1. List of Biomes As Defined and Used in SiB2

<table>
<thead>
<tr>
<th>Biome Number</th>
<th>Biome Name</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Broadleaf evergreen trees</td>
</tr>
<tr>
<td>2</td>
<td>Tall vegetation: broadleaf deciduous trees</td>
</tr>
<tr>
<td>3</td>
<td>Tall vegetation: broadleaf and needleleaf trees</td>
</tr>
<tr>
<td>4</td>
<td>Tall vegetation: needleleaf-deciduous trees</td>
</tr>
<tr>
<td>5</td>
<td>Short vegetation/C4 grassland</td>
</tr>
<tr>
<td>6</td>
<td>Short vegetation: broadleaf shrubs with bare soil</td>
</tr>
<tr>
<td>7</td>
<td>Short vegetation: dwarf trees/shrubs</td>
</tr>
<tr>
<td>8</td>
<td>Agriculture/C3 grassland</td>
</tr>
</tbody>
</table>

al. [1992, 1996a], canopy photosynthesis is included in SiB2. The canopy carbon assimilation rate is regulated in three ways: rubisco limitation, light limitation, and transport limitation. The plant takes in CO2 and releases water vapor through the leaf stomata. The stomatal conductance is expressed as a function of the net assimilation rate and the CO2 concentration at the leaf surface, as discussed by Collatz et al. [1991 and 1992] and Sellers et al. [1992, 1996a]. Canopy transpiration of water vapor is thus directly related to canopy carbon assimilation through the canopy conductance. Transpiration in turn may feedback on the canopy conductance by affecting the canopy’s environment.

The monthly CO2 flux between the terrestrial biosphere and the atmosphere is estimated diagnostically as the difference between the monthly soil respiration $R_{soil}$ and the net canopy assimilation rate, $A_n$. Following Randall et al. [1996], the CO2 flux is estimated as

$$F_{CO2} = R_{soil} - A_n,$$

where $R_{soil}$ is calculated using

$$R_{soil} = \sum_{\text{lyear}} \frac{R_n}{(\Delta t)},$$

$\bar{R}_n$ is the monthly mean soil respiration scale, which is a function of the layer-averaged soil temperature, root-zone soil moisture, and soil texture [Randall et al., 1996]. The annual gross primary production (GPP) of the terrestrial biosphere is expressed as

$$GPP = \sum_{\text{lyear}} A_n(\Delta t),$$

In equations (1)–(3) we assume that the annual GPP is totally released by soil respiration, so that the annual mean CO2 flux is zero at each location.

Nine vegetation types (biomes) are defined in SiB2, as listed in Table 1. The vegetation type for each grid cell is used to interpret satellite data to determine the values of parameters used in SiB2. These include the morphological properties of the canopy structure, the soil layer depths, canopy heights and leaf optical parameters, as well as physiological properties relating to canopy assimilation. These are generally not time-invariant quantities. Properties describing seasonal biosphere changes are derived for each biome based on the NDVI observations obtained through satellite remote sensing, including the fraction of incident photosynthetically active radiation (PAR) absorbed by the green canopy (FPAR), the canopy greenness, the leaf area index (LAI), and aerodynamic quantities, as shown by Randall et al. [1996] and Sellers et al. [1996b].

3. Forcing Data

The off-line SiB2 simulations need atmospheric forcing data, namely, the potential temperature, water vapor mixing ratio, and horizontal wind components in the atmospheric boundary layer; precipitation rate; and the downward solar and longwave radiation at the surface. The frequency of the input forcing data must be at least hourly in order to adequately resolve the diurnal cycle of the atmospheric physical processes and surface-atmosphere fluxes.

Table 2 lists the data sets and sources used for generation of the atmospheric forcing data. The data are all for the year 1987. They include (1) the daily total precipitation data observed through the global observation station network [Shea et al., 1994]; (2) the monthly mean precipitation diurnal cycles as simulated with CSU GCM [Randall et al., 1991]; and (3) six-hourly surface analyses and other assimilation system output obtained from the European Centre for Medium-Range Weather Forecasts (ECMWF).

The ECMWF assimilation system products were generated using operational cycle number 28, as described by Blondin [1988, 1991]. These products include surface temperature,
moisture, winds, and surface energy fluxes on a six-hourly basis. Three soil layers were considered but without explicit vegetation. The two prognostic temperatures and two soil wetnesses for the top two layers were based on derivations of Budyko [1956] and Deardorff [1978]. The deepest layer temperature and soil wetness were assigned fixed “climatological” values. The surface sensible and latent heat fluxes were parameterized, using the bulk formulas of Louis [1979] and Louis et al. [1982]. These products were produced as the ECMWF model was run in an operational data assimilation mode, in which observed winds, temperatures, and humidities, and other data were inserted every 6 hours. In effect the model interpolates the observations in space and time, providing a dynamically consistent representation of the state of the atmosphere. It is important to recognize, however, that the results produced in this way are not pure observations because the formulation of the model does affect the results. Nevertheless, the ECMWF assimilation system products for the winds, temperature, and humidity are strongly constrained by the assimilated observations where the data density is high enough.

As discussed below, we also use certain other ECMWF assimilation system products, including for example the surface solar radiation. This is most assuredly not an observation, although the values obtained have been influenced by observations, e.g., of humidity. The surface sensible and latent heat fluxes produced by the ECMWF assimilation system are also used here, not as input to SiBDRV but rather for comparison with results produced by SiBDRV. While these ECMWF surface sensible and latent heat flux products are not observations, they have been influenced by the assimilated observations of the winds, temperature, and humidity. We refer to these and similar ECMWF model-generated fields as “ECMWF assimilation system output.”

3.1. Atmospheric Boundary Layer Data

The surface transfer parameterization used in SiBDRV (and the CSU GCM), which is based on the work of Deardorff [1972], accepts as input “mixed layer” values of the potential temperature ($\Theta_m$), water vapor mixing ratio ($q_m$), and the horizontal wind components $u_m$ and $v_m$. During the day these values are characteristic of the well-mixed layer that is typically observed to fill the upper portion of the atmospheric boundary layer. At night these values are interpreted as being representative of the upper portion of the typically stable and relatively shallow nocturnal boundary layer.

To generate suitable values for use with SiBDRV, the ECMWF 2-m surface temperature ($T_{sfc}$), 2-m dew point temperature ($T_{dew}$), and 10-m surface winds ($u_{10}$ and $v_{10}$) were extrapolated up to a height of 50 m above the land surface, which we assume is a representative depth of the lower part of the mixed layer in the sense defined above. Surface similarity flux-profile relations [Paulson, 1970; Businger et al., 1971] were used to generate these values based on the surface turbulent fluxes of momentum, heat and moisture, and surface pressure from the ECMWF assimilation system together with the SiB2 surface roughness length and displacement height. We are aware that surface similarity theory may not work well for very rough and heterogeneous surfaces [Garrett, 1977]. The error becomes small, however, above twice the canopy height. The six-hourly mixed layer values were further interpolated to obtain hourly values, with the aid of a cubic spline method [Burden and Faires, 1985].

Figure 2 compares the 2-meter air temperature from the ECMWF assimilation system, $T_{sfc}$, with the climatology of Legates and Willmott [1990a]. The temperatures from the ECMWF assimilation system are generally cooler than those of Legates and Willmott, with global mean differences of 2.2 and 2.0 K in winter and summer, respectively. In regions such as eastern Siberia, the northern boreal forests, East Asia, and the East and West Coasts of North America, there are larger temperature differences (more than 4 K). This is consistent with the discussion of Blondin [1988], who showed that the version of the ECMWF model used to produce the assimilation products employed here overestimated surface evaporation, and so produced excessively cool surface temperatures, particularly in the northern hemisphere.

The surface transfer parameterization of Deardorff [1972] requires as input the depth of the atmospheric boundary layer. For the SiBDRV simulations this depth was set to 50 mbar. In the GCM and SiBDRV simulations the surface-boundary transfer coefficients are calculated diagnostically as functions of the bulk Richardson number.

3.2. Surface Radiation Forcing

SiBDRV uses the SiB2 surface albedo prescription. Nevertheless, the land surface albedos produced by the ECMWF assimilation system are needed for the calculation of the downward surface shortwave radiation flux, which was not available in the ECMWF assimilation system output. The albedos were estimated using

$$\alpha_{sfc} = \left(1 - \frac{a \cdot M_g}{Z_m}\right) \cdot \alpha_{soil} + \left(\frac{a \cdot M_g}{Z_m}\right) \cdot \alpha_{snow}$$

(4)

for forested regions, and

$$\alpha_{sfc} = (1 - A_{snow}) \cdot \alpha_{soil} + A_{snow} \cdot \alpha_{snow}$$

(5)

for bare soil or short grass cover. In (4), $\alpha_{soil}$, $\alpha_{snow}$, and $M_g$ are the snow-free background soil albedo, the snow albedo, and the snow water mass per unit area, respectively, and $Z_m$ is
the effective snow-masking depth. Our methods to determine $Z_m$ and $\alpha_{\text{snow}}$ are discussed in the Appendix.

Estimates of the downward surface solar radiation fluxes were first derived from the snow-free land surface albedo $\alpha_{\text{soil}}$ by combining the ECMWF assimilation system output and the net surface solar radiation, using Goudriaan's [1977] method to decompose the total solar radiation into visible ($S^v$) and near-infrared ($S^r$) components. An iteration was used to compute the land surface albedo, given the snow-free surface albedo, total surface net solar radiation, snow, and cloud cover.

Figure 3 shows the estimated mean surface albedo $\alpha_{\text{df}}$ for December, January, February (DJF), and June, July, and August (JJA), as well as the differences between the ERBE (Earth Radiation Budget Experiment) [Barkstrom, 1984] clear-sky albedo and the estimated surface albedo. The estimated ECMWF land surface albedo is in good agreement with the ERBE clear-sky albedo, although, of course, they are not expected to be the same because the ERBE clear-sky albedo includes the effects of Rayleigh scattering by the gaseous atmosphere. The seasonal global mean differences are 3.8 and 2.8% for winter and summer, respectively. There are larger differences at higher latitudes and over snow/ice covered land, in part because ERBE is generally unable to identify clouds over snow and ice surfaces.

The SiBDRV-simulated land surface albedos are compared with the ECMWF values, and with the GCM (Figure 3). There are large surface albedo differences between ECMWF and SiBDRV. In the northern winter the surface albedo used in the ECMWF assimilation system is about 5 to 25% higher than the SiBDRV albedo over the northern hemisphere but is lower than the SiBDRV albedo in the southern hemisphere. During the northern summer, ECMWF has higher land surface albedos in the northern boreal forest regions and near the Arctic Circle and lower surface albedos in the midlatitudes and over much of the southern hemisphere. This is due to differences in the treatment of snow-free soil/vegetation albedo between the ECMWF assimilation system and the SiB2. The snow-free soil albedos from the ECMWF system are set as constants for each land point, without consideration of seasonal vegetation changes, while SiB2 does consider the seasonal vegetation changes. For the Sahara desert, the ECMWF assimilation system used a higher surface albedo than SiB2; SiB2 prescribes the desert albedos following the ERBE clear-sky albedo [ Sellers et al., 1996b]. The SiBDRV surface albedos for snow-free surfaces agree well with the values obtained in the GCM simulation for the summer months (Figure 3); on this basis we believe that the radiation components have been adequately approximated by the empirical formula.

The surface downward solar radiation was calculated from the surface albedo and surface net solar radiation flux. The hourly surface downward solar radiation flux, $\text{SWD}(t)$, was interpolated from the six-hourly data using

$$\text{SWD}(t) = \text{SDT} \cdot \Phi(t),$$

where $\text{SDT}$ is total accumulated solar energy during the day and $\Phi(t)$ is a weighting function, given by

$$\Phi(t) = \sum_{i} [1 - C(t)] \cdot \sin \left( \frac{\pi t}{DL} \right).$$

In (7), $C(t)$ is the total cloud cover (fraction), $DL$ is daylight length (hours), and $SS$ and $SR$ are the times of sunset and sunrise, respectively.

The surface downward thermal radiation flux (LWN) was computed from the net surface longwave radiation flux (LWN) and the ground surface temperature ($T_g$), using $LWN = LWN + e \sigma T_g^4$. We have assumed that the surface emissivity $\varepsilon$ is 1.0 for all types of land surface. A cubic spline method was used to interpolate the six-hourly downward radiation data into an hourly data set.

### 3.3. Precipitation

The observed daily total precipitation data set for 1987 [Shea et al., 1994] was used to generate the hourly precipitation data, aided by the GCM-simulated monthly mean diurnal cycle. The interpolation was done using

$$P_f(t) = P_d \chi(t),$$

$$\chi(t) = P_{\text{del}} \left( \sum_{i=0}^{23} P_{\text{del}}(t) \right),$$

where $P_d$ is daily total precipitation, $P_{\text{del}}$ is the monthly mean diurnal cycle (first and second diurnal harmonics only) of the precipitation, as simulated by the CSU GCM [Randall et al., 1991], and $P_f(t)$ is the hourly interpolated precipitation rate. Negative values of $P_{\text{del}}$ were set to zero.

Figure 4 shows a comparison of the precipitation rate used as input to SiBDRV, as described above, and the precipitation
climatology of Legates and Willmott [1990b], for December, January, and February (DJF) and June, July, and August (JJA). The SiBDRV precipitation is in fair agreement with the precipitation climatology, although the climatological global means over land are about 0.37 and 0.49 mm d\(^{-1}\) greater in winter and summer, respectively. The largest differences are in the tropics, particularly in India and Southeast Asia, where the SiBDRV values are generally less than the climatology. It should be noted that we are comparing precipitation from a particular year with a multiyear climatology; the differences may thus be due to interannual variations. The SiBDRV precipitation was not corrected for rain gauge errors, as was done by Legates and Willmott [1990b], and we expect generally to underestimate the precipitation without such corrections.

Finally, the total precipitation rate was arbitrarily divided equally into convective and stratiform precipitation. This partitioning is needed for the surface runoff and infiltration computations in SiB2 [Sellers et al., 1996a].

### 3.4. Coupling of SiB2 and Forcing Data

The surface downward solar and longwave radiation fluxes, precipitation rate, and the atmospheric boundary layer's wind, temperature, and mixing ratio are provided hourly to SiB2. As the model runs, the forcing variables are interpolated linearly for each integration step of 6 min.

### 4. Experimental Design

The six-hourly ECMWF assimilation system output provided to us started from 0000 GMT on January 1, 1987, and ended at 2400 GMT on December 31, 1987. The data were at 2.5 \(\times\) 2.5° grid resolution and were regridded to 4 \(\times\) 5°, which is the resolution used in this study.

Exactly the same land surface parameterization (SiB2) was used in SiBDRV and in the GCM simulation. The biophysical and soil parameters of the land surface boundary conditions were based upon the 1987 NDVI data and global biome distributions. Thus the input atmospheric forcings and the land surface parameters are consistent with each other.

SiBDRV was initialized using a state from an earlier GCM simulation with SiB2. SiBDRV was then integrated for five simulated years, with repeating 1987 conditions. The first four simulated years were used to allow SiBDRV to equilibrate to the prescribed forcing. The fifth and last simulated year is analyzed here. Although the GCM was run for 10 years (from 1978 to 1988), only the data from 1987 are compared with the SiBDRV results.

### 5. Results

#### 5.1. Surface Air Temperature and Water Vapor Mixing Ratio

Figure 5 shows the DJF and JJA seasonal mean surface air temperatures \((T_s)\) as simulated by SiBDRV, as well as comparisons of SiBDRV results with the 2-meter surface air temperatures \((T_{2m})\) from the ECMWF assimilation system output.

![Surface Air Temperature (K)](image)

Figure 5. SiBDRV-simulated surface air temperature (top panel) and comparisons of SiBDRV with the 2-m ECMWF surface air temperature (second panel) and with the GCM simulated surface air temperature (third panel), for DJF and JJA. A comparison of the GCM results with the surface air temperature climatology of Legates and Willmott [1990a] for DJF and JJA are shown in the bottom panel. The contour interval is 5.0 K in the top panel and 1.5 K in other panels.
and with the GCM-simulated surface air temperatures. The surface air temperature and humidity in both SiBDRV and the GCM simulations are representative of the air in the canopy and are diagnostically computed from the corresponding surface and mixed layer values.

The $T_a$ obtained with SiBDRV agrees well with $T_{2m}$ from the ECMWF assimilation system output over much of the global continents, for both DJF and JJA. The SiBDRV climate is slightly warmer than the climate produced by the ECMWF assimilation system in the midlatitude deserts for JJA and in the deserts in the southern hemisphere for DJF.

The GCM simulation gives much warmer $T_a$ than does SiBDRV over much of the global continents, especially over the northern hemisphere in DJF. The seasonal global mean $T_a$ differences are 3.0 and 1.3 K in DJF and JJA, respectively. The larger $T_a$ differences in DJF are due to less snow cover in the GCM simulation (discussed later), where temperature differences are about 10 K warmer. In JJA, the $T_a$ differences are more likely affected by the surface solar radiation which is related to cloud effects, as discussed later.

Figure 5 also compares surface air temperatures of the GCM with the observations in DJF and JJA. The global mean temperatures of the GCM are much closer to the observations, with global mean temperature differences of 0.6 and $-0.3$ K in DJF and JJA, respectively. However, the GCM is warmer than observed over the northern hemisphere in DJF and cooler than observed over much of the northern hemisphere in JJA. Figure 4 shows that the GCM generally produces more precipitation in the regions that the model makes cooler than observed.

Figure 6 shows the seasonal mean surface air water vapor mixing ratio ($q_a$) as simulated by SiBDRV, and a comparison with the 2-m mixing ratio from the ECMWF assimilation system ($q_{2m}$), and with the simulated surface water vapor mixing ratio in the GCM. The seasonally varying global mean values are close among the simulations and the ECMWF assimilation system output, although the mixing ratio differences show slightly moister surface air in SiBDRV than in the GCM or the ECMWF assimilation system output. More specifically, SiBDRV has moister surface air in the southern hemisphere in DJF and is generally moister in the subtropical regions in the northern hemisphere in JJA. The $q_a$ differences between the GCM and the SiBDRV simulations are mainly located in the southern hemisphere and the sub-tropics in DJF and in the Amazon, North America, India, Southeast Asia, and desert regions during JJA, where precipitation differences (Figure 4) are very well correlated with the $q_a$ differences.

5.2. Solar and Terrestrial Radiation

Figure 7 shows the surface net solar radiation fluxes produced by SiBDRV, and comparisons with ECMWF, with the CSU GCM, and with the retrievals of Li and Garand [1994] for winter and summer. The observed all-sky surface net shortwave radiation [Li and Garand, 1994] was derived from the ERBE clear-sky solar radiation at the top of the atmosphere and the ECMWF total precipitable water data, using an empirical radiative transfer parameterization.

In the seasonal global mean, SiBDRV absorbs 6.4 and 7.9 W m$^{-2}$ less solar radiation on land than does ECMWF, for DJF and JJA, respectively. The mean differences between the GCM and the SiBDRV, however, show that the GCM absorbs 18.7 and 8.1 W m$^{-2}$ more net shortwave radiation flux on the land surface in DJF and JJA, respectively. The SiBDRV incident global surface shortwave radiation is about 4.0 W m$^{-2}$ (3.4%) and 9.0 W m$^{-2}$ (4.6%) more than the Moats et al. [1994] all-sky surface shortwave radiation for DJF and JJA, respectively.

SiBDRV tends to absorb less surface net shortwave radiation than ECMWF in South America and Australia but more in the northern hemisphere land regions in winter. During summer, the middle- and high-latitude regions of the northern hemisphere and much of the southern hemisphere land absorb less surface net shortwave radiation in SiBDRV than the ECMWF assimilation system, but for the Sahara desert and at high latitudes of the northern hemisphere, SiBDRV absorbs more solar radiation than ECMWF.

Snow cover differences, shown in Figure 8, produce large albedo differences between the ECMWF assimilation system output and the SiBDRV simulation in winter (Figure 3). Over the Eurasian boreal forests in high latitudes, there is more snow in SiBDRV than in the ECMWF assimilation system output, but nevertheless SiBDRV gives a lower surface albedo, because the ECMWF assimilation system includes less of a vegetation masking effect and uses a higher albedo for the snow-free soil. The snow cover produced by the ECMWF assimilation system was based on values calculated from surface air temperature and precipitation, except for parts of Canada and Europe where snow cover was observed at stations, area averaged, and interpolated [European Center for Medium-Range Weather Forecasts (ECMWF), 1987]. The maximum snow depth was limited to 140 cm in Canada and Europe. Unfortunately, the precipitation produced by the ECMWF assimilation system was not available for this study.

Comparisons between the SiBDRV and the GCM (Figure 8)
show that in DJF the continents absorb more shortwave radiation in the GCM than in the SiBDRV, except on the east coasts of South America, South Africa, and the Tibetan plateau. During the northern summer, SiBDRV absorbs more shortwave radiation in western North America, the west coast of South America, eastern Asian continent, and the southern Sahara region. The snow cover differences, as shown in Figure 8, only partially explain the higher surface albedo and the smaller surface net solar radiation produced by SiBDRV in the northern hemisphere in winter. There is little surface albedo difference between the SiBDRV and the GCM (Figure 3) in the summer, except at very high latitudes, where SiBDRV shows a higher surface albedo because of persistent snow cover.

The relatively large amount of absorbed surface solar radiation in the GCM, shown in Figure 8, is due to excessive downward surface solar radiation. The clouds used for SiBDRV (from the ECMWF assimilation system output) were compared with the total cloud cover as simulated by the GCM (not shown). We found there was more cloud cover in the GCM than in the ECMWF assimilation system output. The seasonal global mean cloud differences were large: 24 and 30% in DJF and JJA, respectively. This seems in conflict with the fact that there is generally more surface net solar radiation flux over much of the land area in the GCM (Figure 8). The explanation is that the GCM clouds are optically thinner than the clouds produced by the ECMWF assimilation system. The global mean total cloud covers produced by the ECMWF assimilation system are 53 and 44% over land for DJF and JJA, respectively. The cloud cover produced by the ECMWF assimilation system agrees much better with the International Satellite Cloud Climatology Project (ISCCP) data. On the other hand, the GCM had 29 and 25% more total cloud cover than shown by the ISCCP data set, for DJF and JJA, respectively.

Excessive surface net solar radiation in the GCM tends to increase the surface temperature and the turbulent fluxes of sensible and latent heat, as well as the longwave radiative cooling of the surface. The surface net longwave radiation is also affected by cloud cover, atmospheric water vapor and temperature, through the downward component of the longwave radiation. Figure 9 shows the surface net longwave radiation in SiBDRV (negative contours show surface cooling), and a comparison with the ECMWF assimilation system output, and with the GCM simulation. The GCM generally produces more surface longwave cooling than SiBDRV in the northern hemisphere in both DJF and JJA, because of the warmer land surface in the GCM. On the West Coast of North America and over the tropical and southern hemisphere continents, there is more surface longwave cooling in SiBDRV, because there is less cloud and a relative warmer and drier land surface in those regions.

There are differences in precipitation between the SiBDRV and the GCM (Figure 4). Of course, precipitation is correlated with cloud cover to some extent, so that cloudy regions have less incoming surface solar radiation and a cool and moist land
surface. A more recent version of the GCM includes a new stratiform cloudiness parameterization [Fowler et al., 1995] and the GCM-simulated precipitation distributions are in much better agreement with observations.

The surface net longwave radiation fluxes agree well between the ECMWF assimilation system and the SiBDRV, as shown in Figure 9, although SiBDRV produces slightly more surface net longwave cooling than the ECMWF assimilation system in the southern hemisphere during DJF, and in the midlatitude desert regions of the northern hemisphere during JJA. The land surface temperature affects the upward surface longwave radiation. In the ECMWF assimilation system the soil heat capacity is set to a constant $2.4 \times 10^6$ J m$^{-2}$ K$^{-1}$. In SiB2, however, the soil heat capacity is calculated as a function of soil moisture such as discussed by Sellers et al. [1996a] and is affected by precipitation and snow accumulation/melting. Over deserts and other dry regions, SiBDRV tends to have warmer ground and more surface longwave cooling than the ECMWF assimilation system. In far eastern Siberia, SiBDRV produces less surface cooling due to more snow on the ground in JJA (Figure 8).

5.3. Sensible and Latent Heat Fluxes

Figure 10 shows the land surface sensible heat flux simulated by SiBDRV and comparisons with the GCM simulation and the ECMWF assimilation system output. The surface sensible heat flux differences between SiBDRV and the ECMWF assimilation system output (Figure 10) depend very much on location. In winter, SiBDRV produces less sensible heat flux over much of the globe but more in North America and Asia. In summer, SiBDRV tends to produce a larger sensible heat flux over land, especially over North America, North Africa, and at high latitudes of Eurasia. More sensible heat flux is produced in the GCM than SiBDRV (Figure 10), over much of the globe in winter, except in the Amazon and southern Africa. In summer the GCM produces a much larger sensible heat flux than SiBDRV in central North America, Amazon, North Africa, and India.

Figure 11 shows that the ECMWF data assimilation system generally produces stronger surface moisture fluxes than SiBDRV in winter, with a global mean difference of 9.1 W m$^{-2}$. In summer there are greater differences between SiBDRV and the GCM. The ECMWF data assimilation system gives more latent heat flux than SiBDRV over much of North America, the Amazon region, tropical Africa, and Southeast Asia. The GCM generally produces much more surface latent heat flux than SiBDRV over much of the globe. The seasonal global mean differences of the surface moisture fluxes are 11.9 and 21.3 W m$^{-2}$ for DJF and JJA, respectively.

The partitioning between the surface sensible and latent heat fluxes is largely determined by the ground wetness and vegetation distributions. In the ECMWF model the surface sensible and latent heat fluxes are estimated by the bulk transfer method of Louis [1979] and Louis et al. [1982]. The latent heat flux was corrected by an efficiency function which depends on soil wetness and the temperature of the surface soil layer. According to Blondin [1986] the ECMWF assimilation system tends to overestimate surface evaporation and simulates excessively dry surface soil and excessively cold surface temperatures, especially in the northern hemisphere summer, when compared with surface meteorological observations.

Total surface evaporation in SiB2 is the sum of direct evaporation from interception, soil surface evaporation, and tran-

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**Figure 9.** Net surface net longwave radiation of SiBDRV (top) and comparisons with the ECMWF assimilation system output (middle) and with the GCM simulations (bottom), for DJF and JJA. The contour interval is 10 W m$^{-2}$ in the top panel and 5 W m$^{-2}$ in the bottom two panels.

**Figure 10.** Surface sensible heat flux of SiBDRV (top) and comparisons with the ECMWF assimilation system output (middle) and with the GCM simulations (bottom), for DJF and JJA. The contour interval is 10 W m$^{-2}$ in all panels.
Figure 11. Surface latent heat flux of SiBDRV (top) and comparisons with the ECMWF assimilation system output (middle) and with the GCM simulations (bottom), for DJF and JJA. The contour interval is 10 W m$^{-2}$ in all panels.

respiration from vegetation, part of the evaporation is regulated by vegetation stomata, which in turn is affected by the root-zone moisture supply and the canopy temperature. The precipitation (Figure 4) is well correlated with the surface moisture flux, indicating that more precipitation generally results in a larger surface latent heat flux.

In the SiBDRV simulation the maximum possible moisture flux from the land surface to the atmosphere is practically determined by the prescribed atmospheric temperature and humidity and the prescribed downward surface shortwave and longwave radiation fluxes. The partitioning between the surface sensible and latent heat fluxes is regulated by the responses of the vegetation to the prescribed precipitation and the ground wetness.

In the GCM simulation, on the other hand, the surface moisture fluxes, the atmospheric temperature and mixing ratio, precipitation, and the soil moisture all interactively feedback on each other. It is not clear which are the forcings and which are the responses. If we assume that the surface turbulent fluxes of heat and moisture are responses to the atmospheric and radiation forcings in the same way as in SiBDRV, then we may interpret the differences of sensible and latent heat fluxes between the GCM and the SiBDRV.

5.4. Carbon Assimilation

Photosynthesis is the only significant mechanism for carbon uptake by the terrestrial biosphere, and it is one of the major sinks for atmospheric CO$_2$. In Figure 12 we present the simulation results for the canopy net carbon assimilation rate in units of $\mu$mol carbon m$^{-2}$ s$^{-1}$ from SiBDRV, and the GCM, and their differences, for DJF and JJA. The net assimilation rate, as represented in SiB2, is total canopy photosynthesis minus canopy leaf respiration [Collatz et al., 1991; Sellers et al., 1992]. Most of the net carbon assimilation occurs in the tropics, with rates between 6 and 12 $\mu$mol carbon m$^{-2}$ s$^{-1}$ throughout the year. The temperate and boreal forests have strong carbon assimilation rates during their growing seasons and show strong seasonal variations. The carbon assimilation rates of the northern temperate and boreal forests are largest during summer and near zero during the winter. Tropical C4 grassland appears to have strong seasonal assimilation changes as well, following the dry and rainy seasons. The maximum assimilation rates for tropical grassland are in DJF, as expected, because that is the wet season for those regions. There is much less seasonal variation in the tropical evergreen forests.

The global mean carbon assimilation rate shows strong seasonal variations too. The seasonal global mean assimilation rate in summer is about twice as large as in winter. This is mainly due to there being more land in the northern hemisphere. SiBDRV and the GCM have similar seasonal variations of the global mean carbon assimilation. The winter global mean assimilation rates are about 56 and 55% of the summer values, for SiBDRV and the GCM, respectively.

Although the seasonality of carbon uptake by vegetation is similar in SiBDRV and the GCM, the relative differences between SiBDRV and the GCM show 10 and 5% more global mean carbon assimilation in SiBDRV than in the GCM, for January and July, respectively. This indicates that the GCM climate generally exerts greater stress on the terrestrial biosphere than the SiBDRV input climate. The major differences are located in the tropical grass land regions in DJF (Figure 12), where there is more precipitation in SiBDRV than in the
GCM (Figure 4). In summer, much of the net carbon assimilation difference occurs in North America, Southeast Asia and India, and South America because of the precipitation differences, which causes the GCM to impose more soil moisture stress on the vegetation in those regions. The distributions of the carbon uptake differences are significant for the land surface carbon budget and are due to the different climates of SiBDRV and the GCM. The GCM-simulated feedbacks between the terrestrial biosphere and the climate, discussed by Randall et al. [1996], may account for the net carbon assimilation differences.

The global total net assimilation rates simulated by SiBDRV and the GCM are shown in Figure 13. The terrestrial biosphere carbon budget includes carbon assimilation by vegetation and respiration by plant leaves, stems and roots, as well as soil respiration. Total carbon uptake by the canopy minus leaf respiration is defined as the gross primary production (GPP) of ecosystem. The GPP minus roots and stem respiration is the net primary production (NPP) of the terrestrial biosphere ecosystem. The global total net carbon assimilation shown in Figure 13 is thus comparable to the GPP if we consider that stem photosynthesis and leaf respiration are very small in comparison with total canopy photosynthesis [Schlesinger, 1991]. Estimations of annual mean global GPP over land are 113.4 and 116.6 Gt carbon per year from the GCM and SiBDRV simulations, respectively. It has been estimated that approximately 50 to 60% of the gross carbon assimilation is used for plant respiration [Schlesinger, 1991]. Thus the NPP estimates of SiBDRV and the GCM are about 52 and 51 (of 45% GPPs) Gt carbon per year, respectively, which are less than the global mean NPP estimation of 63 Gt carbon per year made by Leith [1975] who correlated NPP measurements with local temperature and precipitation. More recent estimates of NPP by Melillo et al. [1993] and Potter et al. [1993], however, gave 53 and 48 Gt carbon per year respectively, based on comprehensive ecosystem models driven by climate observations and GCM output. Our results illustrate that estimates of global NPP are sensitive to the climate forcing; we see a less estimate of the GPP in the GCM than in SiBDRV, with about 10 and 5% differences of SiBDRV in winter and summer, respectively (Figure 13). The ratio of NPP to GPP varies among biomes and changes with forest age, so that the 45% NPP to GPP ratio is an estimation of the rule of thumb. The global mean GPP values simulated in the GCM and SiBDRV are reasonably close to other calculations and agree with the total NPP values of 45–65 Gt carbon per year suggested by Schlesinger [1991, p. 120].

SiBDRV and the GCM both have maximum global GPP in July and minimum GPP in January (Figure 13) as expected. The SiBDRV GPPs are larger than the GCM estimates for most months, except in April and May. The greatest differences are in winter, indicating that the SiBDRV climate is more hospitable to vegetation than the GCM climate.

Figure 14 shows the distributions of CO₂ fluxes according to Fung et al. [1987], and the corresponding results from SiBDRV and the GCM, for DJF and JJA. The CO₂ fluxes for SiBDRV and the GCM are diagnostically calculated under the assumption that annual total carbon uptake by vegetation is balanced by total soil respiration throughout the year, as discussed in section 2.3. In winter the mean CO₂ fluxes between the atmosphere and the terrestrial biosphere are close to zero over much of the land area in high latitudes because of photosynthesis shut-down, frozen soil, and snow accumulation. There are CO₂ releases from soil to atmosphere in western Europe for all three simulations, with a slightly higher CO₂ release over a larger area in the GCM, but less CO₂ release in SiBDRV. Much warmer ground temperatures and lower snow cover (Figure 8) over that region in the GCM, relative to SiBDRV, favors more CO₂ release from the soil in the GCM. During DJF, much of the CO₂ uptake by the terrestrial biosphere is in the southern hemisphere, where photosynthesis

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**Figure 13.** Monthly variations of the gross primary production (GPP) simulated in SiBDRV (dotted-dashed line) and in the GCM (solid line). The annual mean GPP is shown for SiBDRV (dotted line) and the GCM (dashed line). The units are Gt C per year.

**Figure 14.** Global distributions of the seasonal mean CO₂ fluxes simulated by Fung et al. [1987], in SiBDRV, and in the GCM. The contour interval is 0.3 μmol carbon m⁻² s⁻¹. The dashed lines indicate carbon uptake by terrestrial biosphere, and the solid lines indicate a net carbon flux into the atmosphere.
assimilates more carbon than soil respiration releases. For all of the simulations the strongest CO₂ fluxes are found over the C4 grassland areas of the southern hemisphere. The greatest carbon assimilation occurs here in the GCM and SiBDRV simulations, because for C4 plants, carbon assimilation is more efficient at higher temperatures, compared to C3 plants [Collett et al., 1991, 1992]. Both the GCM and the SiBDRV show negative carbon fluxes (from atmosphere to biosphere) in India and Southeast Asia, where Fung’s calculation shows positive CO₂ release (into the atmosphere) instead. There are differences in the estimates of soil respiration and carbon uptake of the terrestrial biosphere between Fung’s method and SiB2. Fung et al. [1987] calculated the carbon uptake rate by using the NPP values of vegetation classification weighted by a function of monthly NDVI and the soil respiration rate by the same NPP values weighted by empirical functions depending on climatological surface air temperature. In western North America and the Amazon region the GCM and SiBDRV show significant differences because of their different climates, and especially precipitation (Figure 4).

During JJA, major CO₂ sinks are located in the northern boreal and temperate forests for all three simulations (Figure 14), and CO₂ releases occur over the grasslands in the southern hemisphere. SiBDRV and Fung show net carbon uptake in Southeast Asia, which is contrary to the GCM results. SiBDRV shows CO₂ release to the atmosphere in western North America, where the summer is dry and warm. Both Fung and the GCM, however, show strong carbon sinks in that region. SiBDRV experiences much less precipitation in western North America (Figure 4) and generally a drier and warmer climate, which often is closer to observations, so that there is much less carbon uptake in SiBDRV than in the GCM (Figure 14).

Figure 15 shows monthly variations of the zonally averaged total CO₂ fluxes between the atmosphere and the terrestrial biosphere, as simulated in SiBDRV and the GCM, and also the CO₂ flux of Fung et al. [1987]. Generally, there is good agreement among the three simulations. We see stronger seasonal variations in northern middle and high altitudes as the result of the strong seasonality of the biosphere and strong climate variations during the year. There are weaker seasonal variations in the tropics, but both SiBDRV and the GCM produces stronger seasonal variations than Fung’s estimate. In the tropics the carbon assimilation and soil respiration are affected mainly by precipitation, in the wet and dry seasons. In middle and higher latitudes both SiBDRV and the GCM start net carbon uptake earlier than the estimate of Fung et al. [1987], between April and May. This earlier CO₂ uptake is supported by the observed decrease of atmospheric CO₂ concentration in the high latitudes in April and May, as discussed by Randall et al. [1996].

There are no global CO₂ flux measurements for independent comparison. However, the simulation of atmospheric CO₂ concentration with atmospheric tracer transport models, provided with various surface CO₂ sources and sinks, gives an indirect but independent measure of the CO₂ flux associated with the terrestrial biosphere. Denning [1994] simulated the atmospheric CO₂ budget, using the GCM and various prescribed surface CO₂ flux maps. He found that in middle and high latitudes, Fung’s land surface CO₂ flux estimate tends to produce too much atmospheric CO₂ in the spring, indicating a too weak CO₂ uptake by the terrestrial biosphere. On the other hand, the SiB2-simulated CO₂ fluxes compared well with observations at the same latitudes.

Although the terrestrial biospheres in SiBDRV and the GCM are represented by the same model, the two coupled systems respond differently to their different climates. There are differences between the soil respiration rates simulated by SiBDRV and the GCM in the middle and high latitudes, where the GCM shows a stronger net carbon release into the atmosphere than SiBDRV between January and March. During spring the GCM starts to produce net carbon uptake at 60°N latitude in April, which is about one month earlier than estimated by Fung. SiBDRV is close to Fung’s estimate at that latitude, showing stronger soil respiration in spring. Net carbon uptake begins about 0.5 month earlier than in Fung’s results, however.

The major seasonal changes of the CO₂ flux are similar in the GCM, Fung’s computations, and SiBDRV, with strong CO₂ uptake in the growing season and soil respiration in spring and fall. The soil respiration continues into the winter season in both the SiBDRV and the GCM, but there is much less soil respiration in Fung’s estimate for the same season. This may be due to the fact that the GCM and SiBDRV use the layer-averaged soil temperature to calculate soil respiration, instead of the climatological air temperature used by Fung. In the tropics the GCM and SiBDRV simulate much stronger seasonal changes of CO₂ flux than are found in Fung’s estimate. The differences of the tropical CO₂ fluxes are particularly large.
in Africa, where the GCM and SiBDRV show more net assimilation in summer (Figure 14). Among the three simulations, SiBDRV shows the strongest seasonal variations in the tropics.

6. Conclusions
The global land surface energy, water, and CO₂ fluxes have been simulated using an off-line land surface process model based on SiB2, using ECMWF data assimilation products to prescribe the atmospheric boundary layer's mean potential temperature, mixing ratio, and wind; ECMWF assimilation system output to prescribe the downward surface solar and thermal radiation; and various observational data sources to prescribe the precipitation and soil and vegetation boundary conditions. The SiBDRV results have been compared with the interactive climate simulations performed with the CSU GCM, with the ECMWF assimilation system output, and with available observations.

The differences between the land surface energy budget components and the surface climatology produced by SiB2 and the ECMWF assimilation system are due to the differences between the land surface parameterizations used in the two models. The surface net radiation produced by the ECMWF data assimilation system was redistributed and controlled by the soil and vegetation model in SiB2, which produced less surface latent heat flux. This is because the large canopy resistant term explicitly formulated by SiB2 and the possible precipitation differences between the SiBDRV precipitation forcing and the ECMWF data. On the other hand, discrepancies between the SiBDRV results and the GCM simulation are related to climate differences, especially differences in incoming solar radiation, precipitation, and snow cover.

The surface flux differences between the ECMWF assimilation system output and the SiBDRV may be affected by inconsistencies in the precipitation forcings seen by SiB2 and the ECMWF land surface parameterization. We are unable to evaluate this effect, however, because the ECMWF precipitation data are not available to us.

The seasonal mean surface-air temperature simulated by SiBDRV is generally cooler than that simulated by the GCM, particularly for DJF. There is more surface net solar radiation in the GCM simulation than in SiBDRV. The GCM tends to have more surface longwave cooling, less surface sensible heat flux, and more latent heat flux. The distributions of the differences in surface sensible and latent heat fluxes are correlated with the precipitation differences.

SiBDRV and the ECMWF assimilation system output are naturally more similar in surface air temperature and water vapor mixing ratio, because the input data for SiB2 are derived mainly from the ECMWF assimilation system output. The differences between the ECMWF and the SiB2 land surface parameterizations tend to produce different surface energy fluxes, however. The differences between the surface net solar radiation produced by SiBDRV and the ECMWF data assimilation system are due to differences in the land surface albedo, especially due to snow cover effects, vegetation masking, and the snow-free soil albedo. Generally, SiBDRV tends to produce less surface net solar radiation than the ECMWF data assimilation system. There is slightly more surface longwave cooling in the SiBDRV output. SiBDRV tends to reduce the surface latent heat flux and produces more surface sensible heat flux, compared with the ECMWF assimilation system output, particularly in the northern hemisphere summer.

The CO₂ budget of the global land surface has been estimated using SiBDRV and compared with the results of the GCM simulation and with the estimates of Fung et al. [1987]. Compared with the GCM, SiBDRV produces about 10 and 5% more global carbon uptake in January and July, respectively, because of climate differences; evidently, the climate produced by the ECMWF data assimilation system is more favorable for plant growth than the GCM's climate. The annual gross primary productions agree well between the SiBDRV and the GCM. Both estimates are well within the range of other studies, using either ecological process models or empirical regression models.

The seasonal CO₂ exchanges between the land surface and the atmosphere were diagnostically estimated for both the SiBDRV and the GCM and agree reasonably well with Fung's calculations. Monthly variations of the zonal total CO₂ flux show that carbon uptake in the boreal and temperate forests in SiBDRV and the GCM occurs about a half month to one month earlier than in Fung's results. There is more soil respiration in the winter months in the GCM, because of the warmer ground. Soil respiration and carbon uptake agree well between SiBDRV and Fung's results for the transition seasons.

Appendix

Equation (4) is similar to the formulation of Barker et al. [1994]. For snow-free conditions the land surface albedo was set equal to the ECMWF assimilation system output background soil albedo; that is, \( \alpha_{\text{soil}} = \alpha_{\text{soil}} \). The values of the snow water mass per unit area are also taken from the ECMWF assimilation system output. We used a constant ratio of 5 to convert snow water mass into snow depth in meters. In (4), \( Z_m \) is defined as the effective snow-masking depth of the vegetation canopy, which corrects for the shading effect of vegetation on snow on the ground and is written as

\[
Z_m = Z_1 \frac{LAI}{LAI_{\text{max}}}.
\]

Here LAI is the leaf area index. We use \( LAI_{\text{max}} = 1.5 \). The ratio \( LAI/LAI_{\text{max}} \) corrects for the canopy leaf shading on ground snow. The SiB2 canopy height \( Z_1 \) and seasonal changes of LAI are used in (6). The snow area fraction, according to Sellers et al. [1996a], is given by

\[
A_s = 13.2, \ldots, W_{\text{snow}}, \quad 0.0 \leq A_s \leq 1.0,
\]

which is an empirical relationship based on remote sensing data obtained with the scanning multichannel microwave radiometer (SMMR) sensor on Nimbus 7 [Chang et al., 1987, 1990].

The snow surface albedo, \( \alpha_{\text{snow}} \), was estimated as the sum of visible snow albedo, \( \alpha_{\text{snow}^v} \), and near-infrared snow albedo, \( \alpha_{\text{snow}^n} \), weighted by the downward surface visible (\( S^v \)) and near-infrared (\( S^n \)) radiation fluxes; that is,

\[
\alpha_{\text{snow}} = \frac{\alpha_{\text{snow}^v} \cdot (S^v + 0.5S^n)}{S^v + S^n},
\]

where \( \alpha_{\text{snow}^v} \) and \( \alpha_{\text{snow}^n} \), and the temperature-dependent snowmelt factor \( F_{\text{melt}} \) are

\[
\alpha_{\text{snow}^v} = 0.8F_{\text{melt}},
\]

\[
\alpha_{\text{snow}^n} = 0.4F_{\text{melt}}.
\]
\[ F_{\text{min}} = 1 - 0.04 \left( T_g - T_f \right), \quad 0.6 \leq F_{\text{min}} \leq 1.0. \] (15)

In (15), \( T_g \) is the ground surface temperature and \( T_f = 273.16 \text{ K} \).

Acknowledgments. We thank the Data Services at European Centre for Medium-Range Weather Forecasts (ECMWF) for providing the surface analysis data set and the Data Support Section of National Center for Atmospheric Research (NCAR) for providing the precipitation data. We are also grateful to Inez Fung of the University of Victoria for providing the CO₂ flux data and to Zhanqing Li of the Canadian Center for Remote Sensing, who provided the SRB data. This work has been supported by NASA’s Earth Observing System (EOS) program, through the Sellers-Mooney Interdisciplinary Science Project under NASA contract NASS-31730. Computing resources were provided by the NASA Center for Computational Science (NCSS) of the Goddard Space Flight Center.

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(Received September 7, 1995; revised April 16, 1996; accepted April 16, 1996.)