Multi-Temporal MODIS Data Product for Carbon Cycles Research

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Abstract. A basic methodological premise in the driving of a carbon cycle model is that the findings obtained at the stand level can be applied at landscape, regional and global levels. So it is important to use spatially comprehensive data sets, especially satellite observations. Since the satellite signal received from an optical wavelength is sensitive to the characteristics of the terrestrial biosphere, remote sensing data can be used to retrieve many terrestrial parameters. In this paper we describe a satellite data set and derived products prepared for driving a carbon cycle model. It is derived from daily measurements by the MODIS onboard the TERRA and AQUA satellite. The data set was obtained through a procedure applied to remove atmospheric attenuation effects, cloud contamination and bidirectional reflectance effects to get land surface reflectance. The paper describes the correction procedures and the characteristics of the corrected data set, and briefly presents several derived products of biophysical parameters, including the normalized difference vegetation index (NDVI), the leaf area index (LAI) and the fraction of photosynthetically active radiation (FPAR).

Keywords: MODIS, carbon cycle, data set

1. INTRODUCTION

Scientific researches in the field of resources and environment are entering a new era characterized as globalization, systematization and quantification, in which methods, directions and interests of researches are changing greatly. The launch of EOS TERRA provided a new opportunity for the retrieval of quantitative information on terrestrial ecosystems. The Moderate Resolution Imaging Spectrometer (MODIS) onboard the TERRA satellite is able to acquire data in 36 spectral bands simultaneously with a best resolution of 250 m at nadir and a 1–2 day temporal resolution. With the successful launch of the AQUA satellite, the temporal resolution was increased to half a day, which made it the most comprehensive, advanced, up-to-date platform for remote sensing data. It is driving the quantitative remote sensing applications to a new era. The MODIS platform not only retains continuity and consistency with former NOAA, Landsat TM data, but also provides a new data source that includes atmosphere, land and ocean for integrated studies of scientific mechanism of the evolution of the
Three support teams have been established by National Aeronautics and Space Administration (NASA) to popularize the new satellite data source and support the MODIS Science Team. These teams include the MODIS Characterization Support Team (MCST), which is dedicated to the production of a high quality MODIS calibration product, satellite maintenance and data receiving; the MODIS Administrative Support Team (MAST), which is dedicated to the support of MODIS science team; and the MODIS Science Data Support Team (SDST), which is dedicated to the generation of high quality Level 2 through Level 4 MODIS science products for distribution (Justice et al., 2002). The MODIS Science Team is divided into four discipline groups: Atmosphere, Land, Ocean, and Calibration. The Team developed algorithm theoretical basis documents (ATBD) for MODIS products. Four large remote sensing data centers belonging to NASA process and distribute these data by SDST software. Such data cover many parameters required in terrestrial ecosystem research and are integrated in a uniform platform, which avoids the difficulties of obtaining data from several platforms that hampered in early research.

The MODIS data processing algorithm research and software development began in the Center for MODIS Scientific Data Processing, Institute of Geographical Sciences and Natural Resources Research, CAS in 2001. Retrieval software for many terrestrial parameters and MODIS preprocessing has been developed. This paper presents methods for the preprocessing of MODIS 1B data and the retrieving of terrestrial ecological parameters from preprocessed data.

2. METHODS AND DATA PRODUCTS

2.1 MODIS data

MODIS is a 36-channel visible to thermal-infrared sensor with an FOV of ±55°, a scene width of 2230 km, and a temporal resolution of 1–2 days. Its nadir resolution is 250 m (Channels 1 and 2), 500 m (Channels 3 to 7) and 1 km (other channels), respectively. Channels 1 to 7 are used for terrestrial ecological parameter retrieval, Channels 17 to 19 are used for precipitable water estimation, and Channels 1, 3 and 7 are used for aerosol inversion, which is useful for atmospheric correction of Channels 1 to 7. The data we receive are preprocessed in receiving stations to 1B level. Channels 1 to 19 and 26 give radiative DN values at the top of the atmosphere and related properties, which could be converted into reflectance or radiation at the top of the atmosphere. Channels 20 to 36 (excluding Channel 26) supply radiative DN values and descriptive information, which could be converted into radiation and brightness temperature. They also include MOD03 data which describe geometric information such as pixel location coordinates with a precision of 500 m, solar zenith, solar azimuth, satellite zenith, satellite azimuth, etc. All data are stored in HDF format.
2.2 Data preprocessing

2.2.1 Precipitable water estimation

Precipitable water can be retrieved by infrared or near-infrared channels during clear days. Precision depends greatly on the temperature and the temperature profile initially selected. The column content of water at the ocean surface can be retrieved by the radiation at 11 µm, where an atmospheric window is located. For land surface like forest, where the reflectance is lower, the split window technique is effective. However, when the reflectance of the land surface is high and its temperature is close to that of the atmospheric boundary layer, the radiation of the infrared channel is no longer sensitive to water vapor. Moreover, because of the spatial and temporal variability of the emissivity of the land surface, these methods may give rise to considerable errors when they are applied to land. However, there are strong absorption features at 0.90–0.94 µm, caused by water vapor, which could be used for precipitable water estimation (Kaufman and Gao, 1992). The reflectance of different underlying surfaces are not the same at the same wavelength, however, so it is impossible to obtain the transmittance of water vapor by the radiation from one channel only. The difference of solar radiation between absorption channels of water vapor and its non-absorption channels are required for precipitable water estimation. If the land surface reflectance does not change with wavelength, path radiation is merely a small part of direct solar reflectance, so the transmittance of water vapor in its absorption channels can be calculated by the reflectance ratio of an absorption channel to a non-absorption channel. For example,

\[ T_{0.94\mu m} = \frac{\rho_{TOA(0.94\mu m)}}{\rho_{TOA(0.865\mu m)}} \]

where \( T \) is the transmissivity of water vapor at given wavelength and \( \rho_{TOA} \) is the reflectance at the top of the atmosphere.

On the other hand, if the land surface reflectance changes linearly with wavelength, the transmittance of water vapor in its absorption channels can be calculated by the reflectance ratio of an absorption channel to the sum of two windows. For example,

\[ T_{0.94\mu m} = \frac{\rho_{TOA(0.94\mu m)}}{\left( c_1\rho_{TOA(0.865\mu m)} + c_2\rho_{TOA(0.865\mu m)} \right)} \]

where \( c_1 \) and \( c_2 \) are coefficients.

We now obtain the transmittance in absorption channels by the ratios mentioned above. A look-up table giving the transmittance in absorption channels and precipitable water can then be established using radiative transfer models, like MODTRAN and LOWTRAN. Finally, we can obtain precipitable water from the look-up table. If the table is not available, the following formula can be used for rough estimates:
\[ T = \exp(\alpha - \beta \sqrt{W}), \]  

where \( \alpha \) and \( \beta \) are coefficients in a look-up table, and \( W \) is the precipitable water content.

As to MODIS data, Channels 17 (0.905 \( \mu m \)), 18 (0.936 \( \mu m \)) and 19 (0.940 \( \mu m \)) are water vapor absorption channels, and Channels 2 (0.865 \( \mu m \)) and 5 (1.240 \( \mu m \)) are non-absorption channels. They can thus be used to calculate the 2-channel ratio and 3-channel ratio, respectively. The precipitable water in these three absorption channels can then be obtained by the look-up table method. Finally, the average water vapor is calculated as a weighted sum of water vapor in the three channels.

2.2.2 Aerosol optical thickness estimation

Several methods can be used for aerosol optical thickness estimation, using remotely sensed data directly. Among them, a method called the “dark object” method is relatively simple and is widely used. It allows the optical thickness and transmittance of aerosol to be obtained immediately by inversion of MODIS data in the visible and infrared region. The main idea is that the effect of aerosol on observed radiation usually decreases with wavelength as the power of \(-1\) to \(-2\), so its effect in the mid-infrared region is much less than in the visible region. After atmospheric correction and water vapor correction, the signals received by a sensor include two parts: reflectance and absorption of solar energy caused by aerosol, and reflectance caused by land surface. So both aerosol and underlying surfaces determine the signals if the surface reflectance is high, such as desert, while reflectance of aerosol is dominant over dark surfaces. Moreover, the land surface reflectance values are related to some extent at different wavelengths in the region of solar spectra, so the land surface reflectance of dark underlying surfaces in the red and blue (i.e. MODIS Channels 1 and 3) can be calculated as follows (Liang et al., 1997):

\[ \rho_{0.645\mu m} = 0.50\rho_{2.130\mu m} \]  
\[ \rho_{0.469\mu m} = 0.25\rho_{2.130\mu m}. \]

The reflectance of Channel 7, located in the near-infrared region, is rarely affected by aerosol, so it can be assumed to be equal to land surface reflectance after atmospheric correction and water vapor correction. The reflectance in the red and blue bands calculated according to (4) and (5) can be regarded as real reflectance of the land surface, that is, aerosol-corrected reflectance. The difference from reflectance at the top of the atmosphere acquired by satellite can be regarded as the effect of aerosol. We can obtain aerosol optical thickness by linear interpolation of the look-up table calculated by the radiative transfer model. The steps are as follows:
(1) Identification of dark objects

First choose a $5 \times 5$ search window in MODIS 1B image with a spatial resolution of 1 km. Then select dark objects according to priorities listed below in the window. If there are not enough dark objects, the criterion is revised according to a lower priority and the search is repeated. This is a loop process until the lowest priority is reached. If the number is still not sufficient, the search window will be extended and searched again until its size reaches a threshold. If this is the case, mark it as indeterminate. The priority standards are as follows:

- **First priority**: $0.01 \leq \rho_{2.130 \mu m} \leq 0.05$,
- **Second priority**: $\rho_{3.750 \mu m} \leq 0.025$,
- **Third priority**: $0.05 \leq \rho_{2.130 \mu m} \leq 0.10$,
- **Fourth priority**: $0.10 \leq \rho_{2.130 \mu m} \leq 0.15$.

When the underlying surfaces are cloud, water, ice or snow, the reflectance of which is very low in Channel 7 but high in the visible region, (4) and (5) are not fulfilled, so they should be eliminated.

(2) Correction of gas and water vapor

Water vapor, O$_3$ and CO$_2$ have absorption features affecting Channels 1 and 7, Channels 1 and 3, and Channel 7, respectively. So we should make some corrections prior to aerosol calculation as follows (Ouaidrari and Vermote, 1999):

\[
T_{g_{\text{H}_2\text{O}}} \left( U_{\text{H}_2\text{O}}, \theta_s, \theta_v \right) = \exp \left( - \exp \left( a + b \ln \left( m U_{\text{H}_2\text{O}} \right) + c \ln \left( m U_{\text{H}_2\text{O}} \right)^2 \right) \right),
\]

\[
T_{g_{\text{O}_3}} \left( U_{\text{O}_3}, \theta_s, \theta_v \right) = \exp \left( -amU_{\text{O}_3} \right),
\]

\[
T_{g_{\text{O}_2,\text{CO}_2,\text{NO}_2,\text{CH}_4}} \left( P, \theta_s, \theta_v \right) = \prod_{g_{\text{O}_2,\text{CO}_2,\text{NO}_2,\text{CH}_4}} \frac{1}{1 + a(\ln P / P_0)^b},
\]

where $T_g$ is the transmissivity, $\theta_s$ is the solar zenith, $\theta_v$ is the satellite zenith, $U_{\text{H}_2\text{O}}$ is the content of water vapor, $U_{\text{O}_3}$ is the content of O$_3$, $P$ is surface pressure, $P_0$ is standard pressure, $a$, $b$, $c$ are coefficients depending on wavelength, and $m$ is a parameter calculated as follows:

\[
m = \frac{1}{\cos(\theta_s)} + \frac{1}{\cos(\theta_v)}.
\]

(3) Calculation of land-surface reflectance in Channels 1 and 3

After obtaining transmittance of all channels by the methods mentioned above, we can calculate water vapor and O$_3$ corrected land-surface reflectance in
Channel 7 according to $\rho_i = \rho_{i,\text{TOA}} T_i$, where $T_i$ is the transmittance at channel $i$. We can then estimate the land surface reflectance of dark object in Channels 1 and 3 according to the correlation between Channels 3 and 7 according to formulae (4) and (5). Obtaining land surface reflectance in Channel 3 and atmospheric and water vapor corrected reflectance at the top of the atmosphere, we can calculate optical thickness and transmittance of aerosol by radiative transfer models like 6S, MODTRAN, etc. Here, a look-up table and linear interpolation are used to calculate aerosol optical thickness, so that radiative transfer models can be run operationally at pixel level, which could rapidly speed up the calculation.

2.2.3 Atmospheric correction

According to the radiative transfer equation, the relation between the satellite signals received and land-surface reflectance is expressed as follows:

$$L^m = L_0 + \frac{\rho F_d T}{\pi (1 - s \rho)}.$$  \hspace{1cm} (10)

where $\rho$ is land-surface reflectance, $L^m$ is signals received by sensor, $L_0$ is the directly reflected radiation from atmosphere towards sensor, $F_d$ is perpendicular radiance received on land, $s$ is the albedo of the atmosphere, and $T$ is the transmittance from sensor to land surface.

It is necessary to acquire $L_0$, $F_d$, $T$ and $s$ if we want to invert $\rho$ from $L^m$. But these values are determined by the optical conditions of the atmosphere and the wavelengths. If aerosol optical thickness, precipitable water and atmospheric pattern are given and the weather condition is assumed to be a clear day, we can obtain land surface reflectance at a certain wavelength. Unfortunately, even if we have obtained these parameters, it is still very difficult to solve the integral equation, which does not always have an accurate solution in the form of mathematical expressions. An approximate solution is a feasible and acceptable method for solving such equations, such as 2-flux approximation and 4-flux approximation, which is fast to calculate. The accuracy is not high enough, however, so the look-up table is often used to simulate some outputs according to certain input variables off-line. It is possible to obtain an accurate $\rho$ value by searching for outputs according to its inputs in the table and performing linear interpolation. We can see that there is a white aerosol in the center of Fig. 1, but this effect has been removed in Fig. 2.

2.3 Vegetation Index (VI)

The vegetation index is an empirical measure of vegetative cover. In the area of applications and research in satellite remote sensing, dozens of vegetation indices have been developed for different requirements, but only two of them are used in NASA standard products: EVI and NDVI. NDVI is the most widely used vegetation index:
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\[ NDVI = \frac{\rho_{0.865\mu m} - \rho_{0.645\mu m}}{\rho_{0.865\mu m} + \rho_{0.645\mu m}} \]  

where \( \rho_{0.859\mu m} \) and \( \rho_{0.645\mu m} \) are atmospheric corrected land-surface reflectance in Channels 2 and 1, respectively.

In addition, the Enhanced Vegetation Index (EVI) is also used in MODIS products, which corrects for some distortions in the reflected light by importing the reflectance in the blue channel:

Fig. 1. MODIS Band 2, 1, 4 composite image (before atmospheric correction).

Fig. 2. MODIS Band 2, 1, 4 composite image (after atmospheric correction).
EVI modifies NDVI by soil-adjusted coefficient $L$ and two coefficients $C_1$ and $C_2$, which correct the scatter of aerosol in red channel using the blue value. $\rho^*_{0.865 \mu m}$, $\rho^*_{0.645 \mu m}$ and $\rho^*_{0.469 \mu m}$ are reflectances at the top of the atmosphere without atmospheric correction.

Cloud distortion always exists in large-scale images. Vegetation index composition is a way to resolve this problem, which synthesizes satellite observed reflectance, VI, angular information and quality information from several days in an optimized way, so that a VI composite image with a certain time interval is produced. Figure 3 is a composite NDVI map of east and mid-Asia. The MODIS data are acquired from Beijing station and Wulumuqi (Xinjiang) station.

### 2.4 Leaf Area Index (LAI)

LAI is equal to half leaf area per unit of ground area, and is an indicator of various energy cycles, CO$_2$ cycles and material cycles in canopies. It is directly correlated with many ecological processes and parameters, such as evapotranspiration, soil water balance, photon interception in canopies, NPP, GPP, etc. Many researches proved the feasibility of LAI inversion by spectral measurements, although their accuracy might be variable. The algorithm used in our data set is based on classification of land covers (Chen et al., 2002):

Define,

$$SR = 1.27 \times \rho^*_{0.865 \mu m} / \rho^*_{0.645 \mu m}.$$  \hspace{1cm} (13)
For coniferous forest, \[ LAI = \frac{SR - B_c}{1.153}. \]  
(14)

Deciduous forest, \[ LAI = -4.15 \log\left(\frac{16 - SR}{16 - B_d}\right). \]  
(15)

Mixed forest, \[ LAI = -4.44 \log\left(\frac{14.5 - SR}{14.5 - B_m}\right). \]  
(16)

Fig. 4. LAI (24 September 2002).

Fig. 5. Instantaneous fPAR (24 September 2002).
Other type, \[ \text{LAI} = -1.6 \times \log \left[ \frac{14.5 - \text{SR}}{13.5} \right], \] (17)

where SR is a simple ratio for red and NIR; \( B_c, B_d, \) and \( B_m \) are background SR values for coniferous, deciduous and mixed forest, respectively. Figure 4 is the LAI map of north China according to the algorithm described above on 24 September, 2002.

2.5 fPAR

fPAR is the fraction of photosynthetically active radiation, which indicates the fraction of incoming solar shortwave radiation that is used for photosynthesis by vegetation. This is an important parameter, directly driving empirical photosynthesis models and simple process models. It changes with solar zenith \( (\theta_s) \), so the instantaneous fPAR (fPAR\text{Inst}) is calculated as follows (Cihlar et al., 2002):

\[ fPAR_{\text{Inst}} = (0.95 - 0.94 \times \exp(-0.4 \times \text{LAI} / \cos(\theta_s))) \times 100. \] (18)

Figure 5 is the instantaneous fPAR map of north China according to the algorithm described above on 24 September, 2002.

3. CONCLUSIONS

Satellite remote sensing is capable of obtaining a variety of real-time information about our earth. In this paper we describe methods for retrieving related parameters to drive carbon cycle models with MODIS data. The retrieval of ecological parameters from optical remotely sensed data is a complex operation, because the signals received by a sensor are affected by ground, atmosphere and the sensor itself. Usually, research on the inversion of ecological parameters focuses on the relations of spectral signals and vegetation parameters, which is critical for the regional scale, but is not adequate for larger-scale studies, for which data preprocessing is critical. Prior to MODIS, no satellite data can obtain both water vapor information and aerosol information simultaneously. MODIS gives us good opportunities for better data preprocessing, higher precision of reflectance inversion and ecological parameter estimation.

REFERENCES


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