PHOTOMETRIC VARIABILITY OF THE DISK-INTEGRATED THERMAL EMISSION OF THE EARTH

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ABSTRACT

Here we present an analysis of the global-integrated mid-infrared emission flux of the Earth based on data derived from satellite measurements. We have studied the photometric annual, seasonal, and rotational variability of the thermal emission of the Earth to determine which properties can be inferred from the point-like signal. We find that the analysis of the time series allows us to determine the 24 hr rotational period of the planet for most observing geometries, due to large warm and cold areas, identified with geographic features, which appear consecutively in the observer’s planetary view. However, the effects of global-scale meteorology can effectively mask the rotation for several days at a time. We also find that orbital time series exhibit a seasonal modulation, whose amplitude depends strongly on the latitude of the observer but weakly on its ecliptic longitude. As no systematic difference of brightness temperature is found between the dayside and the nightside, the phase variations of the Earth in the infrared range are negligible. Finally, we also conclude that the phase variation of a spatially unresolved Earth–Moon system is dominated by the lunar signal.

Key words: Earth – infrared: planetary systems – planets and satellites: individual – techniques: photometric

Online-only material: color figures

1. INTRODUCTION

Finding and characterizing habitable exoplanets is one of the key objectives of 21st century astrophysics. Transit spectroscopy already allows us to constrain some of the atmospheric properties of hot exoplanets such as composition (Tinetti et al. 2007), temperature structure (e.g., Knutson et al. 2008), or circulation (Snellen et al. 2010). The phase variations of the thermal emission or reflected light of hot exoplanets can be extracted from combined star+planet photometry, providing another powerful tool to characterize the climate of exoplanets. This has been achieved in transiting configurations (e.g., Knutson et al. 2007), including one terrestrial-mass planet (Batalha et al. 2011), and also for non-transiting hot Jupiters (e.g., Cowan et al. 2007). In theory, the thermal orbital light curves of non-transiting short-period rocky planets can also be measured with combined light photometry and used to detect atmospheric species (Selsis et al. 2011) or to constrain the radius, albedo, and inclination of the planet (Maurin et al. 2012). Transit spectroscopy and orbital light-curve measurements may be achievable with the James Webb Space Telescope or the Exoplanet Characterization Observatory (Tinetti et al. 2011) for a few habitable planets transiting M stars that can hopefully be found in the solar vicinity (Belu et al. 2011; Pallé et al. 2011). The characterization of an Earth analog around a G star by transit spectroscopy is much more challenging for several reasons: G stars are less numerous than M stars, the transit probability in the habitable zone is 10 times lower than it is for M stars, the orbital period (and thus the duration between two transits) is significantly longer for G stars, and the planet-to-star contrast ratio is less favorable for secondary eclipse spectroscopy. Direct detection thus seems to be necessary to study Earth analogs around G stars and several instrument concepts have been proposed (Traub et al. 2006; Danchi & Lopez 2007; Trauger & Traub 2007; Cash et al. 2009). Depending on the concept, it is either the light scattered in the visible or the infrared emission that can be detected. Spectroscopy and photometry can then be used to derive some of the planet properties. An important step toward these ambitious programs is to determine what level of characterization could be achieved when observing the Earth as a distant pale infrared dot.

In the optical range, broadband photometry can potentially give us information about the albedo and the cloud cover. The uneven distribution of oceans and continents enables the measurement of the 24 hr rotation period by the autocorrelation of the signal for several viewing inclination angles despite the presence of active weather (Pallé et al. 2004, 2008). Visible and near-infrared spectroscopical studies have been made by, e.g., Hearty et al. (2009), Cowan et al. (2011), Robinson et al. (2011), and Livengood et al. (2011) observing rotational and seasonal variations of the Earth spectrum and their influence on the detectability of the spectral signatures of habitability and life.

In this paper, we provide an integrated mid-infrared (4–50 μm) photometric time series model of the Earth, with 3 hr time resolution, constructed from over 22 years of available satellite data. From this geographically resolved data set, we derived the disk-integrated photometric signal of the Earth, seen as a point-source planet. The paper is organized as follows: In Section 2, we describe our model, the input data, and the assumptions that are considered. In Section 3, we describe the analysis of the time series and we discuss the retrieving of the rotation period, the variability of the signal, and the case of the Moon–Earth system. In Section 4, we summarize our conclusions.

2. METHODS

The Sun outputs most of its energy in the visible and near-infrared range of the electromagnetic spectrum which is absorbed by the Earth and re-emitted to space in the mid-infrared range. The equilibrium temperature of the planet can
be generally defined as

\[ T_{eq} = \left[ \frac{F_S(1 - A)}{4\sigma} \right]^{1/4} = \left[ \frac{L_S(1 - A)}{16\pi a^2\sigma} \right]^{1/4} \]

where \( F_S \) is the solar incident flux, \( A \) is the terrestrial albedo, \( \sigma \) is the Stefan–Boltzmann constant, \( L_S \) is the solar luminosity, and \( a \) is the Earth–Sun distance. For the Earth, this temperature is approximately \( T \sim 255 \) K and the emission reaches its maximum at 10 \( \mu \)m, although its spectral distribution is strongly affected by the broad absorption bands of atmospheric greenhouse gases.

The emission that a remote observer would detect, however, will be highly dependent on the season, the planetary region that is viewed, and the cloud distribution at that time. Here, we derive the variability and periodicities of this integrated flux as it would be seen from a remote perspective, using satellite data and a geometrical model to build the point-like signal received by the observer. The interest of this work lies in the facts that land masses can emit more infrared radiation than ocean areas, these latter have a bigger thermal inertia that make them more resistant to temperature changes, and the atmosphere of the Earth is optically thin enough in some windows of the thermal infrared to observe the influence of surface features on the emission at the top of the atmosphere (TOA).

2.1. Data

To construct the time series of the emitted infrared flux, we have used TOA all-sky upward long-wave (LW) flux integrated over the 4–50 \( \mu \)m wavelength interval. These data were obtained from the NASA Langley Research Center Atmospheric Sciences Data Center, they are part of the NASA/GEWEX SRB Project\(^5\) (Suttles et al. 1989; Gupta et al. 1992), and they cover a 22 year period from 1983 to 2005. The GEWEX LW algorithm used to create these data (Fu et al. 1997) uses a thermal infrared radiative transfer code with input parameters derived from the International Satellite Cloud Climatology Project\(^6\) (ISCCP; Rossow et al. 1996), the Goddard EOS Data Assimilation System level-4\(^7\) (GEOS-4), the Total Ozone Mapping Spectrometer\(^8\) (TOMS) archive, and the TIROS Operational Vertical Sounder\(^8\) (TOVS) data set. The data have a time resolution of 3 hr over the whole globe, and a spatial resolution of \( 1^{\circ} \times 1^{\circ} \) square cells in latitude and longitude. Typical maps of the outgoing mid-infrared radiation of the Earth, directly represented from the GEWEX data, are shown in Figures 1(a) and (b). The maps represent the average flux over the period 18:00–21:00 UT for 1987 July 1 and 2001 July 1, respectively. Some desert regions are noticeable, such as the Sahara, Arizona, or Atacama. Some cold and humid regions are remarkable too, such as Indonesian clouds or Antarctica.

2.2. Geometrical Model

In order to simulate the infrared observations of the Earth seen from any given direction, we have built a geometrical model of the emitted radiation of the Earth. This model calculates at any time the fraction of the planetary disk exposed in the direction of the remote observer and computes the integrated flux \( F_{\text{obs}} \) received by the observer at a distance \( d \):

\[ F_{\text{obs}}(t) = \frac{1}{d^2} \sum_{i} \left[ a_i \cdot \cos \theta_i \cdot \left( \frac{F_i(t)}{\pi} \right) \right] \]

where \( a_i \) is the area of each grid cell, \( \cos \theta_i \) is the cosine of the angle between the surface normal and the observer’s line of sight, \( F_i \) is the TOA flux given by GEWEX (with a resolution of 3 hr), and \( F_i/\pi \) is the specific intensity in the Lambertian approximation. We obtain a point-like source emission every 3 hr.

With the purpose of illustrating several possible planet viewing geometries, we have defined our observer position by equatorial coordinates at a radius distance of 10 pc. The equivalent of the declination angle on the planet surface gives us the latitude of the sub-observer point, which remains constant along the orbit. The position of the planet in the orbit and the position of the sub-observer’s point with time are derived from the ephemerides of the substellar point. The ephemeris data are taken from the JPL Horizons System\(^9\) (Giorgini et al. 1996).

In order to understand and interpret the variation of the Earth emission in relation with geographic features and climates, it is relevant to locate the sub-observer point by its geographic coordinates. Indeed, its latitude remains constant with time because precession and nutation are negligible during an orbit, and its longitude simply varies as \( \omega t \), \( \omega \) being the rotation rate of the Earth (with the exceptions of the polar cases where the planetary geometry does not change with time). Therefore, we can define the sub-observer point by its latitude and by its longitude at a given time. This time is set to January 1, 0:00 UT of the year considered. We consider four different initial longitudes for the sub-observer point: the meridian of the subsolar point, the morning terminator, the meridian of the antisolar point, and the evening terminator. We call these initial observing geometries conjunction (C), western quadrature (WQ), opposition (O), and eastern quadrature (EQ), respectively, although these terms should normally apply only to an observer located in the ecliptic plane (Figure 2).

It is important to note two facts: the first is that each observer sees the planet at a certain local hour during a whole rotation period and the second is that observers placed at the same latitude but at different longitudes see the same region of the planet at different local hours (if the (O) observer sees a given region of the planet during the winter midnight and the summer noon, the (C) observer sees the same region during the winter noon and the summer midnight). Furthermore, an observer over a polar latitude sees the same hemisphere of the planet along the time of observation. In need of a reference for time, we define its “local hour” as the Universal Time (UT). In this case, the seasonal variability and the diurnal variability (daily change in temperature in a certain region) produce the variation of the signal and not the changing of the planetary view (rotational variability), as we see in Section 3.2.

2.3. Limb Darkening

In order to convert the TOA flux into the disk-integrated flux received by a distant observer, we assume an isotropic (Lambertian) distribution of specific intensities at the TOA. In reality, specific intensities are not isotropic, producing limb-darkening

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\(^5\) http://www.gewex.org/
\(^6\) http://isccp.giss.nasa.gov/
\(^7\) http://daac.gsfc.nasa.gov/
\(^8\) http://jwockey.gsfc.nasa.gov/
\(^9\) http://ssd.jpl.nasa.gov/horizons
or -brightening depending on the thermal structure and composition of the atmosphere, as well as the wavelength. This approximation gives an exact result only if (1) the atmosphere is uniform (its structure and composition is the same at all latitudes and longitudes) and (2) the TOA flux has been obtained by integrating over all angles the actual distribution of specific intensities. If these conditions are fulfilled, then angular integration over the planetary disk is equivalent to the angular integration of the outgoing specific intensities done to compute the local flux at the TOA. Although condition (1) is obviously not fulfilled, the geographic variations of the atmospheric structure and the composition of Earth’s atmosphere are not steep and, as a consequence, significant departures are expected only for peculiar regions (for instance, the Sahara) when observed at the edge of the planetary disk. However, in the case where these regions do not remain all the time too far ($\gtrsim 70^\circ$) from the sub-observer point, this effect should not affect the amplitude or the periodicity of the photometric variations, but only affect their shape.

In order to evaluate the error caused (for a uniform atmosphere) when the outgoing flux is inferred from one single direction of propagation (two-stream approximation), we have used the Phoenix\textsuperscript{10} code (a one-dimensional line-by-line code with spherical geometry and full angular integration of specific intensities with a 0.5 precision), for composition and thermal profiles typical of the Earth. The version of Phoenix we have used was specifically developed for Earth-like atmospheres (Pailllet 2006). We find the maximum error on the disk-integrated spectrum to be about 5% for a monochromatic flux. At most wavelengths, and on the wavelength-integrated flux, the error is smaller than a few percent.

3. TIME SERIES ANALYSIS

Two examples of the annual time series of Earth’s outgoing mid-infrared radiation are plotted in Figure 3. The high

\textsuperscript{10} http://www.hs.uni-hamburg.de/EN/For/ThA/phoenix/index.html
frequency variability corresponds to the emitted mid-infrared flux due to Earth’s rotation super-imposed to the seasonal variation during the year. It is readily observable from the figure that the amplitude of the rotational variability is larger for an observer in the equatorial plane (black) and decreases toward that the amplitude of the rotational variability is larger for an observer in the equatorial plane (black) and decreases toward

### Table 1

| Viewing Angle | Orbital Amplitude | Rotational Amplitude |
|---------------|-------------------|----------------------|
|               | 1977              | 2001                 |
| 90° N (N. Pole)| 20                | 1                    |
| 45° N (Mid-lat)| 14                | 14                   |
| 0° (Equator)  | 6                 | 5(8)                 |
| 45° S (Mid-lat)| 7                 | 5(4)                 |
| 90° S (S. Pole)| 11                | 10                   |

**Notes.** Mean amplitude values of the orbital (seasonal) variability for the years of 1987 and 2001 and of the rotational variability over the months of 2001 January, April, July, and October for (O) and (C) (in parentheses) observers situated at different latitudes. Each value is calculated as the percentage of the average variation over the mean value. Data analysis for different years/months retrieves similar results.

### 3.1. Periodicities

The analysis of the cross-correlation of the time series with itself or autocorrelation, shown in Figure 4, is a technique that allows us to determine the rotation period of the planet and the lifetime of the cloud structures. The autocorrelation $A$ can be computed as

$$A(L) = \frac{\sum_{k=0}^{N-L-1} (F(k \Delta t) - \overline{F}) (F((k + L) \Delta t) - \overline{F})}{\sum_{k=0}^{N-1} (F(k \Delta t) - \overline{F})^2},$$

where $L$ is the time lag in number of 3 hr points, $N$ is the total number of points in the time series, and $\overline{F}$ is the mean flux. The autocorrelation is maximum when values are similar within a time lag distance and it is sampled according to the time resolution. This method was chosen before the Fourier transform to avoid the harmonics of the rotational period. The duration of the statistically significant peaks in the autocorrelated time series can give us an estimation of the lifetime of the cloud.
Figure 3. Time series of the mid-infrared emission flux for the two years of 1987 (top) and 2001 (middle and bottom). The sub-observer’s point is represented by the latitudes $90^\circ$ N, $60^\circ$ N, $45^\circ$ N, $30^\circ$ N, $0^\circ$, $30^\circ$ S, $45^\circ$ S, $60^\circ$ S, $90^\circ$ S, and $0^\circ$ longitude at the initial time (January 1, 0:00 UT) and the direction planes of opposition (O) (top and middle) and conjunction (C) (bottom).

(A color version of this figure is available in the online journal.)

Figure 4. Autocorrelation functions of the mid-infrared emission flux from Earth. 2001 January, April, July, and October for the latitudes $90^\circ$ N, $60^\circ$ N, $45^\circ$ N, $30^\circ$ N, $0^\circ$, $30^\circ$ S, $45^\circ$ S, $60^\circ$ S, and $90^\circ$ S.

(A color version of this figure is available in the online journal.)
structures, typically of around one week for Earth clouds (Pallé et al. 2008). In the outgoing mid-infrared radiation flux, a 24 hr rotation period is clearly shown, a value close to the true rotation period. This rotational signature has two origins. First, some large regions exhibit systematically high or low brightness temperatures. This is the case for Indonesian and Sahara areas, as the former is one of the most humid and cloudy regions on the planet, whereas the latter is warmer and drier than Earth’s average. Therefore, the two regions appear as fixed cold and hot features, respectively, even on averaged outgoing flux maps, as we can see in Figures 1(c) and (d) where the TOA emission is averaged over a month and in Figures 1(e) and (f) where it is averaged over a year. A smaller effect comes from the diurnal cycle (the change of brightness temperature between day and night in a region), which is negligible in most locations, because of humidity, clouds, or ocean thermal inertia, but noticeable in some dry continental areas as is shown in Figure 5. In fact, it is the diurnal cycle of the dry lands, which makes possible the detection of the rotation period for the case of the North-polar view during the more stable seasons (winter and summer), as it is shown in Section 3.2. Although it is always the same fraction of the planet (the whole Northern Hemisphere) in the field of view, the change in temperature along the day in the dry continental areas causes the rotational modulation as it is shown in Figure 4 (violet solid line). However, an observer looking at the South Pole does not detect a significant rotational variability, Figure 4 (violet dash-dotted line), not even for the austral summer when the effect of clouds would be minimized. In the Southern Hemisphere, the distribution of land is largely dominated by oceans whose high thermal inertia make diurnal temperature variability negligible.

3.2. Average Rotation Light Curves

Once the rotation period is identified, the observer can produce a typical rotation light curve by folding the time series obtained during weeks or months over the rotation period. This process averages out random cloud variability. It is clearly seen on the maps of Figures 1(c)–(f), where the clouds disappear for longer average times whereas the strongest features mentioned on the previous section prevail. Then the observer can plot this average rotation light curve as a function of an arbitrary longitude (in our case the longitude that we have chosen is the conventional geographic longitude for commodity). The shape of the light curve can then reveal brighter/fainter areas distributed in longitude. For instance, observers over a latitude of 30°N would note that the brightest and faintest point of the light curve occur when the longitude 0° (Sahara) or 135° (Indonesia) are, respectively, centered on their view. The results are represented in Figures 5 and 6. Figure 5 represents the rotational variability with local hour for several latitudes (in columns) and at the months of January (top row) and July (bottom row). Each chart corresponds to four observers placed at the same latitude referred and at the four initial longitudes O, C, WQ, EQ, previously defined. Each of these four planes corresponds to a certain local hour that changes along the orbital period. With this information, we can compare the temperature evolution for a given planetary region along the day (diurnal variability). The maxima show the largest variation temperature along the day whereas the minima hardly vary. For the cases where the observers are placed over the poles, the planetary view does not change with time so the local hour of the graph is taken just as reference of the observation time. For the North Pole case and for both summer and winter seasons,
the biggest variability, we have analyzed the emission of six re-
North Pole. For the identification of the region that produces
bands of high humidity and clouds that have lower brightness
ical Convergence Zone to these latitudes, which produces large
during summer. This is due to the migration of the Intertrop-
180
30
N–0
60
◦
N–30
◦
N, 60
◦
S, 30
◦
S, 45
◦
S, 60
◦
S, and 90
◦
S.
(A color version of this figure is available in the online journal.)

Figure 6. Rotation light curves of the mid-infrared radiation emitted from the Earth. For the months of 2001 January, April, July, and October, 0
 longitude and
itudes of 90
◦
N, 60
◦
N, 45
◦
N, 30
◦
N, 0
, 30
◦
S, 45
◦
S, 60
◦
S, and 90
◦
S.

In order to check the possible source of the diurnal variation
on the Northern Hemisphere, we have made a further analy-
of the emitted flux by geographic latitude bands. Figure 7
shows the mid-infrared emission of the Earth along one orbital
and different latitudes. It is
important to note the temperature evolution with time for each
gometry, implying a seasonal behavior. The equator shows a
warm stable temperature during the year, whereas the summers
of each hemisphere differ being hotter in the northern summer,
as it is shown in Figure 5. The maximum and minimum regions
do not change with the seasons, except for April and October
when the presence of the clouds can mask the signal.

In order to check the possible source of the diurnal variation
on the Northern Hemisphere, we have made a further analy-
showing the diurnal cycle effect previously mentioned. These hours coincide respectively to midnight and midday in the Sahara desert. The
greatest influence of the area is noticed at 30
◦
N latitude when
the Sahara is in the center of the planetary disk. In summer or winter the diurnal variability reaches the 2%, although the main contribution to the flux is that of the solar insolation
during the year. Figure 6 represents the rotational variability for observers placed at 0
 longitude and different latitudes. It is

3.3. Phase Variation

While Earth’s visible flux received by a remote observer is
modulated by the changing phase of the planet, the Earth does
not present significant phases when the integrated infrared flux is observed: the emission from the nightside contributes nearly
as much as the emission from the dayside (Selsis 2004). This
is shown in Figure 3 (middle and bottom), which represent
opposite observers and then opposite visible phases of the planet.
While observer (O) sees the winter midnight, the observer (C)
sees the winter noon which shows the same average temperature
but bigger rotational variation. Figure 5 shows this difference
between 12 hr and 0 hr; however, the seasonal variation dominates. Continental surfaces and the boundary layer above them (roughly the first 1500 m of the atmosphere) experience
a drop of temperature between the day and night. This diurnal
The diurnal cycle is insignificant above the ocean (due to the high thermal inertia of water), hence over ≈70% of the Earth’s surface. Day–night brightness temperature variations affect the outgoing thermal emission only in the 8–12 μm atmospheric window and above dry continents. This happens either over very cold regions, which in this case do not contribute much to the global emission, or over deserts, which represent a small fraction of the Earth’s surface. In addition, the diurnal cycle is much less pronounced above the boundary layer, at altitudes where most of the thermal emission is emitted to space. This is the reason why...
Figure 9. Time series for a North Polar observer (black) and the contribution of each latitude band to the signal. 90° N–60° N latitude band (blue), 60° N–30° N (green), and 30° N–0° (orange).

(A color version of this figure is available in the online journal.)

Figure 10. Moon–Earth mid-infrared emission light curves for one planetary orbit. 45° N (top row), Equator (second row), 45° S latitudes (bottom row), in columns the signals received by the observer’s placed in opposition, western quadrature, conjunction, and eastern quadrature at the initial time. The colors correspond to the Earth (black) and Earth–Moon system, with lowest (red) and highest (blue) inclination angles of the Moon’s orbit according to the observer’s geometry (the possible orbits are comprehended between the two).

(A color version of this figure is available in the online journal.)
phase-correlated variations of Earth’s brightness temperature are negligible compared with seasonal changes. Even for the observing latitudes in the 30°N–0° range, in which the Sahara diurnal cycle appears in the modulation, the winter midday is colder than the summer midnight as it is seen in Figure 5.

3.4. Moon–Earth System Light Curves

We have modeled the mid-infrared flux of the Moon. Due to a very low surface thermal inertia, the temperature map of the starlit hemisphere of the Moon can be calculated by assuming local radiative equilibrium at the surface (Lawson et al. 2007). When calculating the disk-integrated emission, the contribution of the dark side can be safely neglected due to the high temperature difference. Thus, the flux received depends only on the phase of the Moon as seen by the distant observer. The amplitude of the lunar phase variations depends on the elevation of the observer above the lunar orbit when we computed the Earth signal as a function of the latitude of the observer. To a given latitude of the observer, it corresponds to a range of possible elevations above the lunar orbit. In Figure 10, instead of calculating this elevation consistently with the chosen observer geometry, we bracket the orbital light curve of the unresolved Earth–Moon system with the two curves obtained by adding the lunar signal for the two extreme possible elevations. As pointed out by Selsis (2004) and Moskovitz et al. (2009), this presents phase variations dominated by the Moon. We note that the modulation from the satellite becomes negligible for a Moon-like satellite with 20% of the Moon radius, a ∼5% of the radius of the planet. The two main annual variations that modulate the IR emission from the point-like Earth–Moon system are due to the seasons of the Earth and the phases of the Moon. These modulations present a phase shift that depends on the observer geometry. For some geometries, these two modulations are coincidental. This happens if the maximum of the lunar phase corresponds to Earth’s annual emission maximum, which is for instance the case for an observer looking at northern latitudes who sees the Sahara at noon in July (see Figure 10, the first two panels of the left column). This particular observer will see only phase-correlated variations and may attribute this variability to a day–night temperature difference and conclude that the planet has less ocean coverage and a thinner atmosphere. With such particular geometry, seasonal variations could also be mistakenly attributed to the phases in the absence of a moon, unless the lunar origin of the modulation is identified using spectroscopy (Robinson 2011).

4. CONCLUSIONS

In this paper, we have constructed a 3 hr resolution model of the integrated mid-infrared emission of the Earth over 20 years in the direction of a remote observer randomly located. The seasonal modulation dominates the variation of the signal. As expected, it is larger for the polar views because the planetary obliquity causes a bigger annual insolation change for these latitudes. For equatorial views, the seasonal maximum occurs during the summer of the Northern Hemisphere, as the latter contains large continental masses whereas the Southern Hemisphere is dominated by the oceans.

The rotational variability is detectable because of the uneven distribution of oceans and continents with geographical longitude. The daily maximum of the mid-infrared flux is shown when dry large masses of land, such as the Sahara desert, are in the observer’s field of view. The daily minimum appears when cloudy humid regions such as the Indonesian area is visible, as iced big zones are confined to the poles. In the polar views, the distribution of land does not change with time but the diurnal temperature variation of large continental areas affects the signal, allowing the detection of the rotational period in the North Pole case. We find that the rotational variations have an amplitude of several percent, which is comparable to that of the seasonal variations for some latitudes. It is important to remark the strong influence of the weather patterns, humidity and clouds are sometimes able to mask the 24 hr rotation period of the signal for several days at a time. However, this effect can be solved by time folding.

It is important to point out that the Earth does not exhibit a significant modulation associated with phase variation (phase curve). This is because the integrated thermal emission does not generally probe the boundary layer (first kilometer of the atmosphere) where the diurnal cycle takes place. If unresolved, the Earth–Moon system would, however, present a phase variation of lunar origin. A satellite of the size of the Moon would introduce a strong phase variability that would completely dominate over the planet’s signal. This effect adds high complexity to its interpretation by photometry.

At the light of these results it seems that future infrared photometric observations of terrestrial planets can be useful in order to characterize their atmospheric and surface features. If the planet is not completely covered by clouds, as Venus is, the presence of strong surface inhomogeneities (continents) can be extracted from the daily variations. The seasonal cycle can also give estimates of the planet’s effective temperature, the variability in the obliquity of its orbit, and the distribution of land at larger scales. A further study with a Global Circulation Model combining Earth’s emitted flux is ongoing.

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