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Discovery of a lunar air temperature tide over the ocean: a diagnostic of air-sea coupling

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The lunar semidiurnal (L_2) tide in the Earth's atmosphere is unique as a purely mechanically forced periodic signal and it has been detected in upper atmosphere winds and temperature and in surface barometric pressure. L_2 signals in surface air temperature, $L_2(T)$, have only been detected at a single land station (results published almost a century ago). We report observational determinations of $L_2(T)$ over the ocean by using data from 38 moored buoys across the tropical Pacific and Atlantic. In contrast to published speculation that $L_2(T)$ should be negligible over ocean, we find that the observed $L_2(T)$ is fairly close to that consistent with an adiabatic L_2 pressure variation. Any deviations from purely adiabatic behavior are a measure of diabatic effects on the surface air—expected to be dominated by damping processes, notably heat exchange with the ocean surface. With the aid of climate model simulations that include L_2 -tide-like variations, we demonstrate that our observations of $L_2(T)$ provide a unique diagnosis for the strength of air-sea coupling and a useful constraint on climate model formulations of this coupling.

npj Climate and Atmospheric Science (2018)1:25; doi:10.1038/s41612-018-0033-9

INTRODUCTION

The gravitational pull of the moon excites tidal variations in atmospheric circulation with L_2 frequency (period ~ 12.42 h). These variations are small—typically an order of magnitude less than solar diurnal and semidiurnal variations (which are forced primarily by the daily cycle of solar heating).^{1,2} However the lunar tide is singular as a purely mechanically (i.e., adiabatically) forced monochromatic (i.e., single frequency) variation and so observations of the L_2 tide can provide uniquely valuable diagnostics of atmospheric behavior that enhance understanding of the atmospheric circulation. A historically important example is provided by long-standing observations of L_2 variations in surface geomagnetic field that have informed our understanding of the ionospheric dynamo and upper atmospheric winds.³ Very recently it has been demonstrated that the L_2 circulation variations in the troposphere modulate tropical rainfall,⁴ and the measurements of the L_2 rainfall variation provide a valuable constraint on the representation of moist convective processes in climate models.⁵

In this study, we focus on measurements of the L_2 tide in surface air and show that the results provide important information on the thermal coupling of the atmosphere with the underlying surface. The L_2 variation in surface air pressure, $L_2(p)$, has been reliably determined at many stations where very long records of hourly measurements are available and its geographical and seasonal dependence have been established^{1,6–8} (Supplementary discussion for a brief review). The basic features of the observed $L_2(p)$ have been reproduced by simplified linear theory calculations of the response of the global atmosphere to the lunar gravitational forcing.^{2,9,10} The L_2 surface air temperature cycle, $L_2(T)$, is expected to have an amplitude ~ 0.01 K making it quite hard to detect among all the other atmospheric disturbances present. The only direct observational

report so far was obtained at a single location, Batavia (6°S , 106.5°E), by Chapman¹¹ using hourly data during 1866–1928 and this determination had large uncertainty due to sampling noise (Note that a recent study⁴ estimated the size of the troposphere-average $L_2(T)$ based on global reanalysis model data). In this study, we analyzed long data records at many ocean buoys to determine accurate values of $L_2(T)$ over the tropical ocean.

The L_2 forcing will excite a linear wave response in the atmosphere whose air temperature perturbations near the surface ($\hat{T}$: complex amplitude) are governed by

$$\hat{T} = \frac{i\omega}{i\omega + \beta} \hat{T}_{\text{ad}}, \quad (1)$$

where ω is the frequency ($=2\pi/12.42 \text{ h}^{-1}$), and $\hat{T}_{\text{ad}}$ is the adiabatic component defined as,

$$\hat{T}_{\text{ad}} \equiv \frac{R\bar{T}}{C_p\bar{p}} \hat{p}, \quad (2)$$

where $\hat{p}$ is the complex amplitude of $L_2(p)$ (see Methods section for details). Because there is no “external” thermal forcing for L_2 , the diabatic heating is expressed here as an effective Newtonian cooling with rate, β . Chapman¹² suggested that (over land at least) $L_2(T)$ should to first order be in adiabatic relationship with $L_2(p)$, with deviations from this purely adiabatic behavior serving as a diagnostic of damping of the temperature signal in the air near ground (which he suggested would be dominated by heat exchange with the underlying surface). Chapman's single-station determination¹¹ indeed showed a first order adiabatic relation between $L_2(T)$ and $L_2(p)$ but was not sufficiently precise to reliably determine the strength of any thermodynamic damping. Curiously Chapman¹² also suggested that air-sea coupling would be so strong that “over the oceans the lunar atmospheric tide will

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Received: 2 January 2018 Revised: 28 February 2018 Accepted: 6 March 2018

Published online: 24 October 2018

Fig. 1 **a** Semidiurnal harmonic of temperature observed at (0°N, 220°E) as functions of mean local solar time (MLST) and lunar phase (0 is new/full moon and 6 is half-moon). **b** As is **(a)** but for the anomaly from the mean S_2 variation (computed as the average over the lunar phase). The mean local lunar time (MLLT) is shown by magenta lines. **c**, **d** are as **(a–b)** but for the results averaged over the 38 buoy stations

appear to be isothermal at sea level". Our results will contradict this last assertion but will show that Chapman's idea of using determinations of $L_2(T)$ and $L_2(p)$ to infer damping rates can be useful if enough data are analyzed.

Note that for a propagating wave phenomenon such as the lunar tide, the thermodynamic "damping" effects, through either radiative transfer or sensible heat fluxes, will depend not just on the local T' but also on the T' in other layers of the atmosphere, and so the β in (1) may be complex^{13,14} (the real part of β must be positive, of course). In addition there is the possibility of an L_2 signal in latent heat release, although that is expected to be small in the (generally cloud/fog free¹⁵) air just above the tropical ocean.

RESULTS: ANALYSIS OF BUOY DATA

We have for the first time detected $L_2(T)$ over ocean by using data from dozens of moored buoy stations across the tropical Pacific and Atlantic. The observed 12-h harmonic determined from data during any day ($S_{2\text{-obs}}$) results from the superposition of true solar semidiurnal tide (S_2) and the L_2 that is dependent on the lunar phase (ν)², that is,

$$\begin{aligned} S_{2\text{-obs}}(\nu) &= S_2 + L_2(\nu) \\ &= s_2 \cos(\omega(t_{LT} - \phi_2)) + l_2 \cos(\omega(\tau - \lambda_2)) \\ &= s_2 \cos(\omega(t_{LT} - \phi_2)) + l_2 \cos(\omega(t_{LT} - \lambda_2) - \nu\pi/6) \end{aligned} \quad (3)$$

where t_{LT} is mean local solar time (MLST in h), s_2 and ϕ_2 are amplitude and phase of S_2 , τ is mean local lunar time (MLLT in h), l_2 and λ_2 are amplitude and phase of L_2 , and $\omega = 2\pi/12$ (h^{-1}). Here ν is an integer, increasing from 0 to 11 twice in each lunation (new moon to full moon and full moon to new moon). The relation, $t_{LT} = \tau + \nu + 12n$ (n : integer), is used for the transformation.² Figure 1 shows the $S_{2\text{-obs}}$ and the anomaly from the average over the lunar phase at the buoy at (0°N, 220°E) (panels a, b), as well as the results averaged over the 38 stations (panels c, d). Obviously the true S_2 , the component independent of ν , is dominant (~ 0.05 K; Fig. 1a, c); however, the anomaly from S_2 (~ 0.01 K) clearly shows the lunar phase progression (Fig. 1b, d), indicating a significant L_2 signal even for single-station data. The temperature maximum of L_2 is observed around at 10:00 MLLT regardless of ν (see Methods section for details about the procedure for determining the amplitude and phase of L_2). The effects of beating between the true S_2 and L_2 signals appears as a modulation of the amplitude for the observed S_2 from maximum (spring tide) around $\nu = 10$ to minimum (neap tide) around $\nu = 4$ (Fig. 1a, c).

Figure 2a shows the observed $L_2(T)$ (red arrows) and $L_2(p)$ (blue arrows) over the tropical Pacific and Atlantic, as well as the uniform-grid $L_2(p)$ product developed by Schindelegger and Dobszaw⁸ (SD16). The results are presented as harmonic dial vectors showing the amplitude and MLLT phase. $L_2(p)$ are converted to corresponding adiabatic temperature tide (referred to as $L_2(T_{\text{ad}})$) using Eq. (2). $L_2(T)$ has a coherent structure with the

phase at most stations being 10:00–12:00 MLLT. At the same time, a systematic geographical dependence is also obvious, particularly in the zonal direction; for example, both $L_2(T)$ and $L_2(T_{ad})$ are stronger at $\sim 180^\circ\text{E}$ and $\sim 220^\circ\text{E}$ compared to $\sim 200^\circ\text{E}$. Figure 2b shows the scatter plot of $L_2(T)$ and $L_2(T_{ad})$ at selected longitudes. It is seen that generally the amplitude of $L_2(T)$ is slightly smaller than that of $L_2(T_{ad})$, with the phase of $L_2(T)$ earlier than that of $L_2(T_{ad})$. This is consistent with a basic picture of an adiabatically forced mechanical wave slightly modified by thermal damping processes.

Figure 3 shows the $L_2(T)$ (from buoys) and $L_2(T_{ad})$ (from both buoys and the SD16 grid data), plotting all values. The mean amplitude and phase of $L_2(T)$ are $\sim 0.007\text{ K}$ and $\sim 11:20\text{ am/pm MLLT}$, respectively. It is clearly seen that the $L_2(T)$ amplitude is smaller than that of $L_2(T_{ad})$ and that $L_2(T)$ precedes $L_2(T_{ad})$ by $\sim 20\text{ min}$. The spread among the stations reflects the geophysical distribution of L_2 signals (Fig. 2b), as well as observational uncertainty.

The systematic difference between $L_2(T)$ and $L_2(T_{ad})$ indicates the existence of damping processes exerted on L_2 near the surface. The least-square estimation of β in Eq. (1) from buoy measurements (Methods section) shows that $\beta \sim 6 \times 10^{-5} \exp(i\pi/4)$, consisting of not only a real part (β_r) but also a significant imaginary part (β_i) (Fig. 3b). β_r^{-1} is $\sim 0.3\text{ day}$, which is much shorter than estimates of the purely radiative damping timescale

($\sim 2\text{ weeks}$) and also than the convective damping scale (1–10 days) in the tropical free troposphere.^{16,17}

DISCUSSION: NUMERICAL MODEL SIMULATIONS

Following Sakazaki et al.⁵ we have run a version of a realistic near-global climate simulation model with all diurnal variations of solar radiative heating of the troposphere or surface suppressed (the “Remote-F experiment”, Methods section). In Remote-F the daily cycle of stratospheric heating excites a solar semidiurnal (S_2) tide which propagates into the troposphere. For the atmosphere just above the tropical ocean surface this produces a situation very analogous to that of the L_2 tide – a zonal wave 2 westward-traveling pressure wave with period 12 h (vs 12.42 h for L_2) and no local diabatic effects except for those caused by the local longwave radiative damping and sensible heat exchange with the surface (plus any small latent heat release in the air directly above the tropical ocean surface). The same type of experiment was used to investigate the stratospheric role in diurnal cycles in precipitation and in surface pressure solar tides.^{5,18} In this configuration, the model was integrated for 12.5 months (15/11/1999–30/11/2000) and the results from the final 12 months of data were analyzed. Sakazaki et al.⁵ showed that the overall simulation including seasonal mean tropical rainfall is not strongly affected by the suppression of the tropospheric diurnal solar forcing. To

Fig. 2 **a** Horizontal distribution of harmonic dial vectors over the tropical Pacific and Atlantic ocean for (red) $L_2(T)$ from temperature measurements (38 stations), (blue) $L_2(T_{ad})$ from pressure measurements (7 stations), and (light-blue) $L_2(T_{ad})$ from grid dataset by SD16. **b–f** As **(a)** but for selected longitudes; only the end points of harmonic dial vectors plotted with the error bars

Fig. 3 **a** Harmonic dial of (red circles) $L_2(T)$ and (blue circles) $L_2(T_{ad})$ from buoy measurements (38 selected stations for $L_2(T)$; 7 stations for $L_2(T_{ad})$). The results from stations east of 160°E , where both $L_2(T)$ and $L_2(T_{ad})$ are determined (Fig. 2), are shown. The size of each circle denotes the number of days used for the analysis. Horizontal and vertical bars for each datum denote 95% confidence level with t -test (Methods section). Light-blue open circles are from the grid $L_2(T_{ad})$ data by SD16 for each location of $L_2(T)$ measurements. Solid and dashed arrows denote the average harmonic dial vectors for $L_2(T)$ from buoys and $L_2(T_{ad})$ from SD16. **b** Two arrows are adopted by those in **a**. Red curve starting from the averaged $L_2(T_{ad})$ (dashed black arrow) toward zero indicates the values that $L_2(T)$ should take if β , the coefficient of damping, is allowed to take a real, positive value (red circles show the $L_2(T)$ for selected β_r values)

Fig. 4 **a–d** As is Fig. 3a but for Remote-F model experiments with the C_H value relative to the standard values of (a) 50%, (b) 100%, (c) 150%, and (d) 200%. Red and blue dots show the results of $S_2(T)$ and $S_2(T_{ad})$; in the model, T is temperature output at 2 m and T_{ad} is calculated by using surface pressure. Only grid points over the tropical Pacific (10°S – 10°N ; 160° – 270°E) and Atlantic (10°S – 10°N ; 330° – 350°E) are used for analysis. See Fig. S6 for a different version in which red dots are plotted on top of the blue dots. **e–h** Frequency distribution of β ($= \beta_r + i\beta_i$) from the four model experiments for the tropical Pacific and Atlantic. Blue dots with the error bar shows the estimated β from L_2 observed by buoys

examine the effect of air-sea interaction processes, the Remote-F integration was repeated three more times with the standard surface heat exchange coefficients in the model (C_H) multiplied by factors (i) 0.5, (ii) 1.5, and (iii) 2.

Figure 4a–d shows the scatter plots of $S_2(T)$ and the $S_2(T_{ad})$ at the surface for the individual experiments with different C_H . Obviously the forcing for the S_2 tide in the model (stratospheric heating) differs from the gravitational forcing of the L_2 tide in the real world and so we focus only on the relationship between $S_2(T)$ and $S_2(T_{ad})$. All experiments show that the amplitude of $S_2(T)$ is slightly smaller than that of $S_2(T_{ad})$, while the phase of $S_2(T)$ is 10–20 min earlier than $S_2(T_{ad})$. These characteristics are quite consistent with our buoy observations of L_2 . At the same time, it is obvious that the difference between $S_2(T)$ and $S_2(T_{ad})$ increases, both in amplitude and phase, with increasing C_H . This suggests that the thermal interaction between air and sea is largely responsible for the damping process near the surface for a dynamical wave like L_2 tide.

Figure 4e–h shows the histograms of β ($= \beta_r + i\beta_i$), over the Pacific and Atlantic oceans, estimated for each experiment. Notably, not only the real part but also the imaginary part has a significant contribution in every experiment. Through all experiments, β may be reasonably approximated by, $\beta = B \exp(i\pi/4)$, where B is the amplitude; but B increases with increasing C_H , while the phase remains $\sim \pi/4$. The phase lag agrees well with our L_2 buoy measurements, while B best fits in the observed results in case of C_H is increased by 50–100% over the standard value used in this model in earlier climate simulations (including studies with

domains extending over tropical oceans^{19,20}) (Fig. 4g, h). The sensitivity of β to C_H found in our model shows that our L_2 results can be a practical constraint for the parameterization of near-surface-process in climate models.

Supplementary Information includes a discussion of the heat budget over the S_2 cycle in our Remote-F model experiments. Note, we can construct heat budgets only at the numerical model levels with the lowest located ~ 25 m above the surface (while the 2 m temperature is computed using similarity theory, Methods section). At altitudes < 500 m, the S_2 diabatic heating in Remote-F experiment comes largely from parameterized vertical diffusion processes (i.e., the vertical convergence of vertical sensible heat flux (τ): $-d\tau/dz$) (Figs. S3 and S4). At the lowest model level (~ 25 m) (and presumably also nearer the surface) the phase of $-d\tau/dz$ precedes $-T'$ by 1–2 h ($\sim \pi/4$ for L_2 and S_2), consistent with the estimated β . We further find that the phase of $-d\tau/dz$ shows an upward progression with altitude, indicating that it takes a finite time for the heat to be transferred upward (Fig. S4). By contrast, τ itself is found to be in phase with $-T'$ (Fig. S2) at the surface (τ_s), indicating that τ_s varies as a response to local T' near the surface. These findings suggest that the damping on dynamical wave near the surface is largely caused by the vertical convergence of upward-propagating sensible heat flux that was injected from the surface due to the air-sea interaction (proportional to $-T'$). A simple analytical model predicts that the phase lag between the diabatic heating and $-T'$ near the surface should be between 0 and $\pi/2$ (Fig. S5 and Supplementary Information).

We have observed the L_2 variation of surface air temperature at multiple locations, allowing us to now clearly determine these key features of L_2 at low-latitudes: (i) $L_2(T)$ is to first order a moon-synchronous signal with T' peaking around local lunar noon, (ii) the lunar temperature perturbations are roughly in adiabatic balance with the lunar pressure perturbations, but (iii) there is a clear indication of diabatic damping of the lunar temperature signal. We performed climate model experiments, including a close analogue of the L_2 tide, by suppressing the daily solar heating cycle in the troposphere; the model results are consistent with our L_2 observations when reasonable air-sea thermal coupling is included. This experiment is easy to perform and could be the focus for climate model intercomparisons to diagnose air-sea coupling.

METHODS

Derivation of Equations (1)–(2)

We apply the thermodynamic equation for $L_2(T)$ observed just above (3 m) the ocean surface. Vertical winds can be ignored, while any effects of mean horizontal advection are also expected to be small as the horizontal phase speed of the tide is much larger than the mean winds near the surface. The equation can be thus written as,

$$\frac{\partial T'}{\partial t} = \frac{R\bar{T}}{C_p\bar{p}} \frac{\partial p'}{\partial t} + \dot{Q} \rightarrow \frac{\partial T'}{\partial t} = \frac{R\bar{T}}{C_p\bar{p}} \frac{\partial p'}{\partial t} - \beta T', \quad (4)$$

where t is time; $\dot{Q}$ is L_2 diabatic heating rate per unit mass (K s^{-1}); T' and p' are $L_2(T)$ and $L_2(p)$, respectively; $\bar{T}$ and $\bar{p}$ are the mean state temperature and pressure, respectively; R is gas constant; C_p is specific heat of air. $\dot{Q}$ is then represented by an effective Newtonian cooling with the coefficient, β . When we assume a simple harmonic solution with frequency ω (i.e., $T' = \bar{T} \exp(i\omega t)$ and $p' = \bar{p} \exp(i\omega t)$), Equations (1) and (2) are obtained.

Analysis of Moored Buoy data

We analyze data from moored buoy arrays across the tropical Pacific and Atlantic oceans: the Tropical Atmosphere Ocean/Triangle Trans-Ocean Buoy Network (TAO/TRITON²¹) and the Prediction and Research Moored Array in the Atlantic (PIRATA²²), respectively. Since early 1990s, these arrayed buoys have been observing the air temperature and pressure at 3 m above sea level with a high-temporal resolution (10 min to 1 h for temperature; 1 h for pressure). The resolution of temperature measurements is 0.01 K with their accuracy being 0.2 K, while that of pressure measurements is 0.1 hPa with their accuracy being 0.01%. For temperature, we show the L_2 results obtained with only 10-min data (from ~1996 to September 2017) (temperature data from TRITON over the western Pacific (west to 160°E), whose resolution is 1-h, were not used for this study). For pressure, we show the results with hourly data up to September 2017 (there are no 10-min data for pressure). Figure S1 shows the data availability of 10-min temperature data for each buoy over the entire period up to September 2017. Although there are several data gaps, the measurements have continued over the last two decades.

Data are processed following the so-called Chapman-Miller (C-M) method^{2,23} except that we have employed an extensive quality control in advance to reduce the noise level as described below. First, we pick up days in which more than 60% are filled with valid data; for temperature (surface pressure) only the stations with the number of days meeting this criteria >4000 (>1500) are used for analysis (the criteria for pressure is weaker than that for temperature because the noise level is smaller). For each day (i.e., for 24-h time series), we applied a median-filter three times to eliminate the outliers such that any data outside of 3-sigma level from the median are considered as outliers (the median is calculated within each day with N -10 data, with N is the total number of data in the day for every iteration). Finally, the harmonic analysis is performed to extract the S_2 component. The coefficients for the S_2 cosine and sine curves are binned for each lunar phase (ν : integer from 0 to 11, where 0 (6) corresponds to the new/full moon (the half-moon)). Of all binned data for each ν , 50% around the median in each ν are used for further analysis.

The following calculation is based on the C-M method that will be briefly explained here. Let $A_2(\nu)$ and $B_2(\nu)$ be the sum of cosine and sine coefficients, respectively, over binned data for each ν . The cosine and sine

coefficients for lunar semidiurnal signals (l_a and l_b) are obtained as,

$$l_a = \frac{1}{K} \sum_{\nu=0}^{11} \left[A_2(\nu) \cos \frac{\nu\pi}{6} + B_2(\nu) \sin \frac{\nu\pi}{6} - A_2(\nu) \frac{N_A}{N} - B_2(\nu) \frac{N_B}{N} \right], \quad (5)$$

$$l_b = \frac{1}{K} \sum_{\nu=0}^{11} \left[B_2(\nu) \cos \frac{\nu\pi}{6} - A_2(\nu) \sin \frac{\nu\pi}{6} - B_2(\nu) \frac{N_A}{N} + A_2(\nu) \frac{N_B}{N} \right], \quad (6)$$

where

$$N_A = \sum_{\nu=0}^{11} N(\nu) \cos \frac{\nu\pi}{6}, \quad N_B = \sum_{\nu=0}^{11} N(\nu) \sin \frac{\nu\pi}{6}, \quad N = \sum_{\nu=0}^{11} N(\nu), \quad (7)$$

$$K = N \left\{ 1 - \left(\frac{N_A}{N} \right)^2 - \left(\frac{N_B}{N} \right)^2 \right\}. \quad (8)$$

Here, the third and fourth terms of Equations (5)–(6) are the correction terms taking into account any non-uniform distribution of days with respect to the lunar phase (ν). The amplitude and phase of L_2 (l_2 and λ_2 , respectively, in Eq. (3)) are thus derived as,

$$l_2 = \sqrt{l_a^2 + l_b^2}, \quad \lambda_2 = \frac{6}{\pi} \tan^{-1} l_b/l_a (\text{MLLT}). \quad (9)$$

On the other hand, the cosine and sine coefficients of S_2 (s_a and s_b) are obtained by

$$s_a = \sum_{\nu=0}^{11} \frac{A_2(\nu)}{N(\nu)}, \quad s_b = \sum_{\nu=0}^{11} \frac{B_2(\nu)}{N(\nu)} \quad (10)$$

The S_2 amplitude (s_2) and phase (ϕ_2) are thus

$$s_2 = \sqrt{s_a^2 + s_b^2}, \quad \phi_2 = \frac{6}{\pi} \tan^{-1} s_b/s_a (\text{MLST}) \quad (11)$$

The associated errors are estimated in the following procedure. We assume that for each day (i -th day), the observed S_2 component can be actually expressed as

$$\begin{aligned} S_{2-\text{obs}}^i(\nu) &= s_{2-\text{obs}}^i \cos \omega t + s_{2-\text{obs}}^i \sin \omega t \\ &= S_2 + L_2(\nu) + \epsilon_2^i \\ &= (s_a \cos \omega t + s_b \sin \omega t) + (l_a \cos(\omega t - \nu\pi/6) \\ &\quad + l_b \sin(\omega t - \nu\pi/6)) + (\epsilon_a^i \cos \omega t + \epsilon_b^i \sin \omega t) \end{aligned} \quad (12)$$

where the first term is the “true” S_2 independent of lunar phase; the second term is the L_2 ; the third term is the error. Now that we have obtained S_2 (Eq. 11) and L_2 (Eqs. 5–6), ϵ_a^i and ϵ_b^i are calculated for every day; then the errors on the harmonic dial are calculated as

$$E = \sqrt{\frac{1}{N} \sum_i (\epsilon_a^i{}^2 + \epsilon_b^i{}^2)} / \sqrt{N} \quad (13)$$

The error bars in Figs. 2 and 3 are defined as the 95% confidence levels for E with t -test (i.e., $\sim 2E$). For $L_2(T)$, we show the results at 38 stations with relatively small error bars (<0.0016 K; this limit is determined by trial-and-error).

Grid L_2 data from Schindelegger and Dobslaw (2016)

We use the global grid $L_2(p)$ dataset that was developed by Schindelegger and Dobslaw⁸ on the basis of 2315 selected ground-based measurements between 1900 and 2010 from land barometers and moored buoys (including TAO/TRITON and PIRATA used in this paper). It should be noted that they made a correction for the vertical displacement of buoy sensors due to oceanic L_2 tides. Although such correction terms are non-negligible,²⁴ we use data before the correction; these represent actual $L_2(p)$ determination in surface air (data were provided by M. Schindelegger, personal communication). For converting the $L_2(p)$ to $L_2(T_{\text{ad}})$ for this grid data, mean temperature and surface pressure (coefficients of Eq. 2) are derived from ERA-Interim reanalysis data for the period between 1998 and 2010.

Numerical simulations of L_2 -like variations near the surface

We use a comprehensive limited-area atmospheric model developed at the International Pacific Research Center (IPRC) of University of Hawai'i at Manoa,^{19,20,25} which we run in near-global mode. Recently this model was

used to study solar tides near the surface and their relationship with diurnal cycle in tropical rainfall.^{5,18}

We do not directly simulate L_2 signal (by adding the tidal potential to momentum equations²⁶), but reproduce L_2 -like signals near the surface by running the model with the same configuration of “Remote-forcing (Remote-F) experiment” developed by Sakazaki et al.⁵ That is, the diurnal variation of solar heating in the troposphere and surface is eliminated by replacing the shortwave heating and the shortwave flux at the surface with the daily-mean values on the same dates in a control run that was performed in advance. In this case, diurnal cycles near the surface are dominated by the S_2 dynamical tide that is remotely excited in the stratosphere and propagates into the troposphere. Since there is no external, diabatic diurnal forcing near the ground and the frequency of S_2 is close to that of L_2 , the S_2 signal near the surface in Remote-F is a close analogue of L_2 .

The model is run in near-global zonal strip mode with SSTs and lateral boundary conditions at 75°S and 75°N specified from observational reanalyses. The integration is from 11/15/1999 to 11/30/2000; the 1-year data from 12/1/1999 to 11/30/2000 are used for the actual analysis. In order to examine the effect of heat exchange between air and surface on the $L_2(T)$ – $L_2(T_{ad})$ relationship, the Remote-F experiment was repeated with surface heat exchange coefficient (C_H) changed by: (1) 50% decrease, (2) 50% increase, and (3) 100% increase.

Temperature at 2 m in altitude (T_{2m}) in the model is calculated based on the similarity theory of the surface layer. Diabatic heating due to the vertical mixing processes near the surface are calculated using the Mellor–Yamada–Nakanishi–Niino scheme.²⁷

Estimation of damping coefficient

By using the relationship between $L_2(T)$ and $L_2(T_{ad})$ for 38 buoy stations ($L_2(T)$ is from direct measurements while $L_2(T_{ad})$ is from SD16 data), the Newtonian cooling coefficient has been estimated. From Eq. (1), β is determined (β) with the least-squares-method so that the following quantity, L , takes the minimum.

$$L = \sum_n |(i\omega + \bar{\beta})\hat{T}^n - i\omega\hat{T}_{ad}^n|^2,$$

where ω is the wave-frequency ($= 2\pi/12.42$ h), $\hat{T}^n$ and $\hat{T}_{ad}^n$ are the complex amplitudes of $L_2(T)$ and $L_2(T_{ad})$ at n -th station (e.g., $L_2(T) = \hat{T}^n \exp(i\omega t)$). The uncertainty of the estimated β is defined as

$$E = \sqrt{\frac{1}{N_s} \sum_n |\beta^n - \bar{\beta}|^2 / \sqrt{N_s}},$$

where N_s is the number of stations (38) and β^n is determined for each station as,

$$\beta^n = i\omega \left(\frac{\hat{T}_{ad}^n}{\hat{T}^n} - 1 \right).$$

The 95% confidence level (with t -test) is defined as $2E$ (t -value for the degree of freedom of 38 is 2).

Data availability

Buoy data can be obtained through the web site of the Pacific Marine Environmental Laboratory: <https://www.pmel.noaa.gov/tao/drupal/disdel/>. Model results will be publically available at the Asia-Pacific Data-Research Center (APDRC) at IPRC (<http://apdrc.soest.hawaii.edu/>).

ACKNOWLEDGEMENTS

We thank Dr. Chunxi Zhang and Dr. Yuqing Wang for the use of their model and for their practical assistance, and Dr. Michael Schindelegger for providing the original L_2 grid dataset. We also thank Dr. Henryk Dobslaw and one anonymous reviewer for their instructive comments for improving the manuscript. T.S. was supported by the Japan Society for Promotion of Science. Figures 1–4, S1–4 and S6 were drawn with the GFD-DENNOU Library. The IPRC is partly supported by the Japan Agency for Marine–Earth Science and Technology (JAMSTEC).

AUTHOR CONTRIBUTIONS

T.S. and K.H. designed the analysis and numerical experiments. T.S. performed the data analysis and the numerical experiments. K.H. and T.S. wrote the paper.

ADDITIONAL INFORMATION

Supplementary information accompanies the paper on the *npj Climate and Atmospheric Science* website (<https://doi.org/10.1038/s41612-018-0033-9>).

Competing interests: The authors declare no competing interests.

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