

# Role of ocean heat transport in climates of tidally locked exoplanets around M dwarf stars

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Edited by Robert E. Dickinson, The University of Texas at Austin, Austin, TX, and approved November 22, 2013 (received for review August 13, 2013)

**The distinctive feature of tidally locked exoplanets is the very uneven heating by stellar radiation between the dayside and nightside. Previous work has focused on the role of atmospheric heat transport in preventing atmospheric collapse on the nightside for terrestrial exoplanets in the habitable zone around M dwarfs. In the present paper, we carry out simulations with a fully coupled atmosphere–ocean general circulation model to investigate the role of ocean heat transport in climate states of tidally locked habitable exoplanets around M dwarfs. Our simulation results demonstrate that ocean heat transport substantially extends the area of open water along the equator, showing a lobster-like spatial pattern of open water, instead of an “eyeball.” For sufficiently high-level greenhouse gases or strong stellar radiation, ocean heat transport can even lead to complete deglaciation of the nightside. Our simulations also suggest that ocean heat transport likely narrows the width of M dwarfs’ habitable zone. This study provides a demonstration of the importance of exooceanography in determining climate states and habitability of exoplanets.**

planetary climate | exoplanet habitability | superEarth

**M**dwarf stars are the most common type of star in the Universe (1). The habitable zone (HZ) around M stars is close to such stars because of their weak luminosity (2). In consequence, habitable exoplanets orbiting M stars are likely to be tidally locked to their primary stars, so that one side of tidal-locking exoplanets permanently faces stars, and the other side remains dark. Previous studies have demonstrated the role of atmospheric heat transport in preventing atmospheric collapse on the nightside of terrestrial exoplanets located in the HZ of M stars (3–10). For a planet with an extensive ocean, its climate and habitability also involves ocean heat transport, which is known to be important in Earth’s climate (11). None of the existing studies has considered the role of ocean heat transport. Moreover, the climate also involves the spatial distribution of open water versus ice and the question of whether the planet becomes locked in a globally glaciated Snowball state. Simulation with a comprehensive Earth atmospheric general circulation model (AGCM) coupled to a slab ocean, without dynamic ocean heat transport, revealed an “eyeball” climate state, with a round area of open ocean centered at the substellar point and complete ice coverage on the nightside, even for very high CO<sub>2</sub> concentrations (4). In the presence of sea ice, ocean heat transport is likely to be especially important, because it is known from studies of the Snowball Earth phenomenon in Earth-like conditions that ocean heat transport is very effective in holding back the advance of the sea-ice margin (12–14). The distribution of sea ice on tidally locked exoplanets is only an issue for M stars, because planets with sufficiently dense atmospheres orbiting hotter stars in orbits close enough to yield tidal locking are likely to be too hot to permit ice and may even be too hot to retain water. The purpose of the present paper is to study how ocean heat transport and sea-ice processes influence the climate and habitability of tidal-locking exoplanets in the HZ around M stars by carrying out simulations with a fully coupled atmospheric–oceanic general circulation model (AOGCM).

## Methods

The AOGCM used here is the Community Climate System Model version 3 (CCSM3) (15), which was originally developed at the National Center for Atmospheric Research for studying Earth climate. The model includes a dynamic ocean with both wind-driven and deep-ocean circulations. The sea-ice component of CCSM3 consists of energy-conserving thermodynamics and elastic–viscous–plastic dynamics, and the model freezing point of sea water is set to  $-1.8^{\circ}\text{C}$ . The atmosphere component has 26 vertical levels from the surface to the model top of 2.6 hPa and horizontal resolution of  $3.75^{\circ}$  by  $3.75^{\circ}$  in latitude and longitude. The ocean component has 25 vertical levels, longitudinal resolution of  $3.6^{\circ}$ , and variable latitudinal resolutions of about  $0.9^{\circ}$  near the equator. Eddy parameterization used in the ocean model is the Gent–McWilliams parameterization (16).

The model is modified with planetary parameters the same as that of Gliese 581g (GI 581g) (17). The synchronous rotating period, radius, gravity, and incident stellar radiation are 36.7 Earth days, 1.5 times Earth radius,  $13.5\text{ m s}^{-2}$ , and  $866\text{ W m}^{-2}$ , respectively. Although GI 581g is not a confirmed exoplanet, its planetary parameters represent a hypothetical Super-Earth in the HZ of an M star. The stellar spectrum used here is an M star spectrum, with an effective temperature of 3,400 K. Unlike the solar spectrum, most of the M star luminosity is emitted in the near-infrared. Near-infrared radiation is much less reflected by ice and snow compared with visible light. Thus, ice–albedo feedback is weaker on exoplanets orbiting M stars than on Earth (18). Here, sea ice and snow albedos are set to 0.3 and 0.6, respectively. The substellar point is located at the equator and  $180^{\circ}$  in longitude, and both the eccentricity and obliquity are set to zero. The geothermal heat flux is also set to zero. We assume that the exoplanet is an aquaplanet with a uniform ocean depth of 4,000 m, which is the mean depth of Earth’s oceans.

The influence of CO<sub>2</sub> concentration on climate is probed with five different CO<sub>2</sub> levels: 3.6, 355, 10,000, 100,000, and 200,000 ppmv. It has been speculated that the atmosphere of aquaplanets could have high CO<sub>2</sub> levels because silicate weathering could be ceased by water coverage (19). On the other hand, it has been conjectured that CO<sub>2</sub> concentrations high enough to

## Significance

**Tidal-locking planets receive very uneven stellar heating because their one side permanently faces their stars and the other remains dark. While the dayside can be warm enough to sustain liquid water, the nightside could be so cold that any gases would condense out. Here, we perform simulations with a coupled atmosphere–ocean model to demonstrate the importance of exooceanography in determining the habitability of tidal-locking exoplanets around M dwarfs. We show that ocean heat transport can substantially extend the dayside habitable area and efficiently warm the nightside, so that atmosphere collapse does not occur. As greenhouse effect or stellar radiation is sufficiently strong, ocean heat transport can even cause global deglaciation. Ocean heat transport also likely narrows the width of M dwarfs’ habitable zone.**

Author contributions: Y.H. and J.Y. designed research; Y.H. and J.Y. performed research; Y.H. and J.Y. analyzed data; and Y.H. wrote the paper.

The authors declare no conflict of interest.

This article is a PNAS Direct Submission.

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This article contains supporting information online at [www.pnas.org/lookup/suppl/doi:10.1073/pnas.1315215111/-DCSupplemental](http://www.pnas.org/lookup/suppl/doi:10.1073/pnas.1315215111/-DCSupplemental).

maintain open water in the outer portions of the HZ might be incompatible with limitations imposed by sea-floor weathering (12). All other atmospheric compositions are kept the same as present-day Earth. The radiative transfer module of the model is approximately valid for atmospheres with CO<sub>2</sub> concentration lower than 200,000 ppmv and water-vapor column amount less than 1,200 kg m<sup>-2</sup>, although the simulated climate is very slightly colder than it should be (20, 21). For CO<sub>2</sub> levels above 200,000 ppmv, the effects of pressure broadening and collision-induced CO<sub>2</sub> absorption become significant (22, 23). In addition, we also examine climate states for exoplanets closer to both inner and outer edges of the HZ by performing simulations with a sequence of stellar radiation fluxes of 700, 866, 1,200, and 1,400 W m<sup>-2</sup> and with constant CO<sub>2</sub> concentration of 355 ppmv.

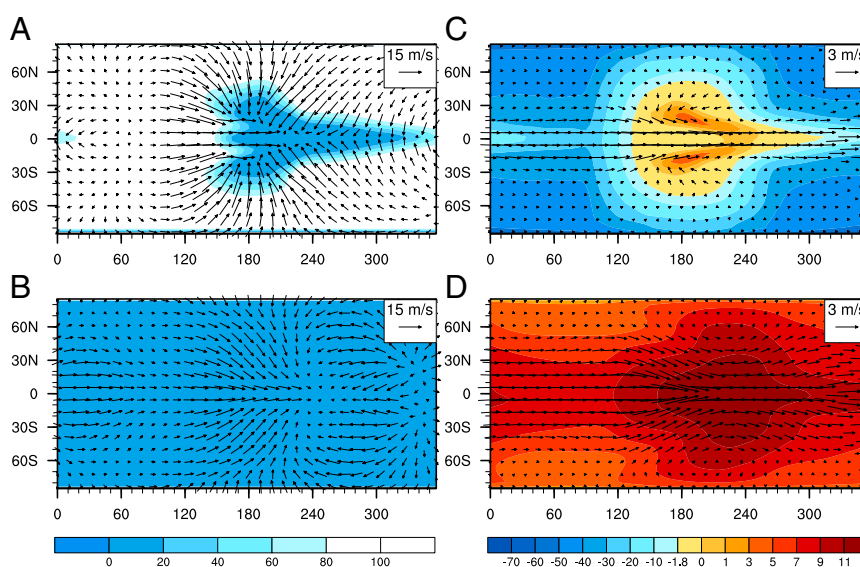
The model is initialized with present-day Earth atmosphere conditions and globally uniform vertical profiles of present-day Earth ocean temperature and salinity. All AOGCM simulations are run for about 2,200 Earth years, and the results shown here are based on averages over the last 100 y. To distinguish the role of ocean heat transport, we also carry out simulations with an AGCM coupled with a 50-m slab ocean. The AGCM used here is the atmospheric component of the AOGCM. All AGCM simulations are run for 80 Earth years. Results shown here are averaged over the last 5 y.

## Results

Fig. 1*A* shows AOGCM simulation results of sea-ice fraction and wind velocity at the lowest model layer for 355 ppmv of CO<sub>2</sub>. This level of CO<sub>2</sub> roughly equals the present-day CO<sub>2</sub> concentration in the Earth atmosphere. In the presence of a dynamic ocean, the open-ocean area (blue) is not like the round iris of an “eye” such as that in AGCM simulations coupled with a slab ocean (Fig. S1*A*; also figure 3 in ref. 4). Instead, the spatial pattern of the open-ocean region is more like a “lobster,” showing two “claws” symmetric to the equator and a long tail along the equator. The tail of open water extends eastward to the nightside. At the western side of the substellar point, sea ice is drifted eastward from the nightside toward the substellar point. The open-ocean region remains even for 3.6 ppmv of CO<sub>2</sub> and shows the similar lobster-like spatial pattern. For very high-level CO<sub>2</sub> (200,000 ppmv), sea ice is completely melted (Fig. 1*B*). By contrast, the nightside and a large part of the dayside remain frozen for the same level of CO<sub>2</sub> in the AGCM simulation (Fig. S1*B*), and the open-ocean region is only slightly expanded compared with that in Fig. S1*A*.

Spatial distributions of surface air temperatures ( $T_s$ ) (color shading) and surface ocean velocity (arrows) are illustrated in Fig. 1*C* and *D*. For 355 ppmv CO<sub>2</sub>, the highest  $T_s$  is not located at the substellar point but has two centers at each side of the equator with values of about 5–6 °C. This is because the eastward equatorial ocean current transports cold water from the dark side to the dayside. The lowest  $T_s$  is not located at either pole or the antistellar point but at subpolar regions of the nightside with values of about –60 °C. The lowest  $T_s$  is well above the condensation temperature of CO<sub>2</sub> (–78.5 °C at 1 bar of CO<sub>2</sub> partial pressure and much lower at lower partial pressures), so there is no risk of CO<sub>2</sub> being trapped in the cold regions. As CO<sub>2</sub> concentration is increased to 200,000 ppmv (Fig. 1*D*), the highest  $T_s$  on the dayside does not rise very much, only about 2 or 3 °C. However, nightside and polar surface temperatures increase drastically and are all above the freezing point of ocean water. Thus, temperature contrasts between the dayside and nightside and between the tropics and poles are largely reduced. Comparison between Fig. 1*D* and Fig. S1*D* reveals that as CO<sub>2</sub> concentration is sufficiently high, ocean heat transport is very efficient in warming the nightside and causing sea-ice melting there.

The flow pattern in Fig. 1 is the well-known Gill-type solution responding to a stationary equatorial heating source (24). The two cyclones at each side of the equator are the tropical Rossby-wave modes, and the eastward long tail along the equator is the tropical Kelvin wave. Similar flow patterns have been seen in atmospheric flows in previous simulation studies for tidally locked terrestrial exoplanets and hot Jupiters (e.g., 5–7, 10, 25, 26). Both the simulations for 355 and 200,000 ppmv CO<sub>2</sub> show a strong westerly equatorial ocean current on the order of a few meters per second. The equatorial westerly ocean current is mainly driven by westerly winds as demonstrated in ref. 27, although the tropical Kelvin and Rossby waves also contribute to the formation of the ocean current. This is different from the equatorial westerly atmospheric flow of tidal-locking exoplanets, which is driven by equatorward convergence of Rossby-wave momentum fluxes (10, 25, 26). The reason why the equatorial ocean current is so strong is because there are no meridional continental barriers and the wind-driven ocean current can be very strong (28, 29). The equatorial current here is just like the



**Fig. 1.** Spatial distributions of sea-ice fraction and surface air temperature. (Left) Sea-ice fraction (unit, %); (Right) surface air temperature (unit, °C); (Upper) 355 ppmv CO<sub>2</sub>; and (Lower) 200,000 ppmv CO<sub>2</sub>. In *A* and *B*, arrows indicate wind velocity at the lowest level of the atmospheric model (990 hPa), with a length scale of 15 m s<sup>-1</sup>. In *C* and *D*, arrows indicate ocean surface current velocity, with a length scale of 3 m s<sup>-1</sup>. Note that the color scale for surface air temperature is not linear. The substellar point is at the equator and 180° in longitude.

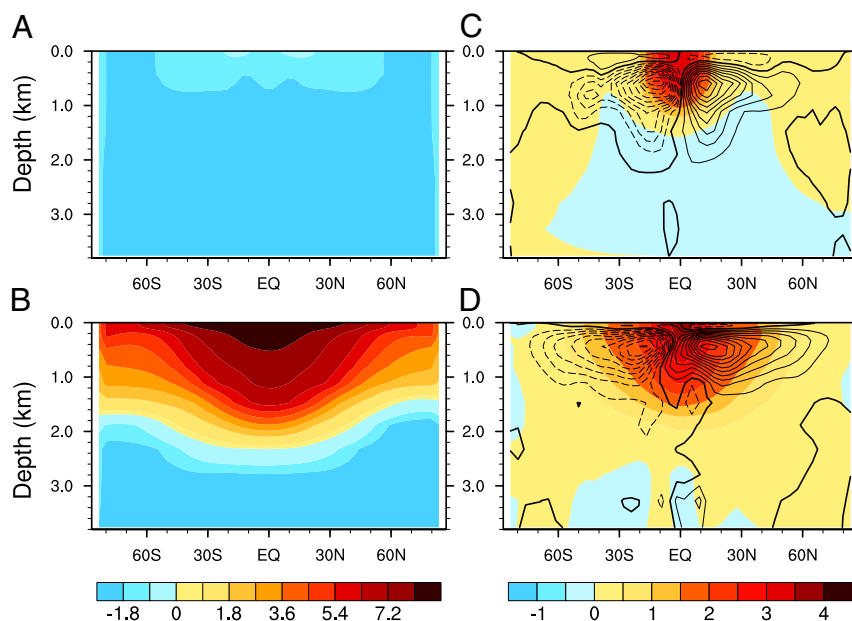
strong Antarctic Circumpolar Current in the Southern Ocean of Earth. Ocean currents, especially the equatorial jet stream, become stronger with increasing  $\text{CO}_2$  (Fig. 1D). This is because the open-ocean area becomes broader with increasing  $\text{CO}_2$ , so that wind stresses can force on the broader open-ocean surface and generate stronger zonal ocean currents.

Thermal structures and motions of deep-ocean layers also play important roles in ocean heat transports. Fig. 2 illustrates depth–latitude cross-sections of zonal-mean ocean potential temperatures and zonal velocities for 355 and 200,000 ppmv  $\text{CO}_2$ , respectively. For 355 ppmv  $\text{CO}_2$ , the relatively warm layer (higher than  $-1.8^\circ\text{C}$ ) is limited in the tropics and above 400 m in depth (Fig. 2A), whereas all other parts of the ocean show rather homogeneous temperature distribution. For 200,000 ppmv  $\text{CO}_2$ , the ocean temperatures increase greatly, especially in the tropics where the warm ocean layer, which has potential temperature higher than  $0^\circ\text{C}$ , extends to below 2,000 m near the equator (Fig. 2B). The highest temperature is up to  $8^\circ\text{C}$ , much higher than that of the 355-ppmv case. The increase in ocean temperatures is not only because of much stronger greenhouse effect but also the broader open-ocean area that absorbs more stellar radiation at much lower surface albedo than that of sea ice and snow. Zonal-mean zonal ocean velocities are shown in Fig. 2C and D. Both demonstrate equatorial jet streams. As  $\text{CO}_2$  increases from 355 to 200,000 ppmv, the equatorial jet stream becomes stronger and expands both downward and poleward. As shown below, the increases in both ocean temperatures and zonal velocities with increasing  $\text{CO}_2$  cause more ocean heat transport from the day-side to the night-side. Fig. 2B and D display a slight asymmetry between the two hemispheres, which is probably introduced by the time interval (100 Earth years) for averaging, which is not long enough.

Contours overlapped in Fig. 2C and D are the mean meridional mass streamfunction of ocean, which represents zonal-mean ocean meridional overturning circulations (MOC). In both plots, streamlines display two layers of circulation cells with opposite directions. The cells in the upper layer (above 300 m) are driven by wind stress. Westerly winds produce an equatorward

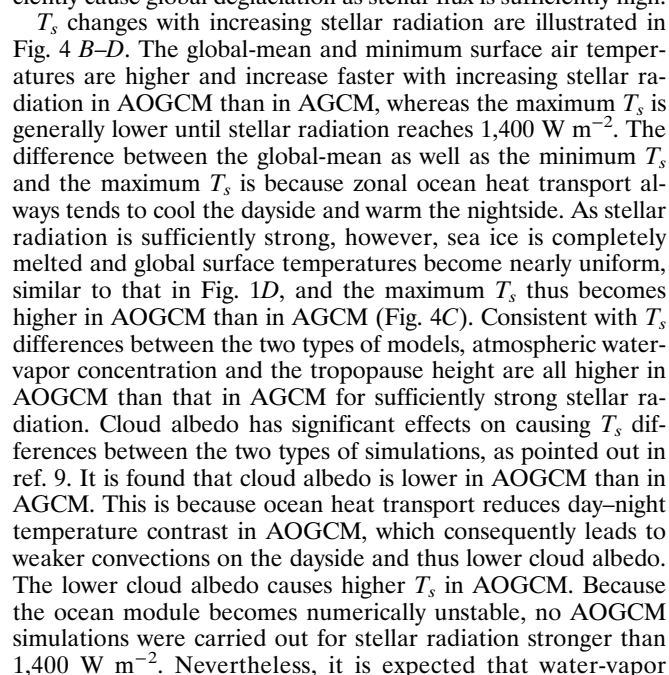
Ekman drift and therefore descending motion at the equator, and equatorward wind stress drives cold surface water from high latitudes toward the equator for compensation. Equatorial warm water at depth moves poleward and eventually upward at high latitudes in both hemispheres, forming enclosed ocean meridional circulations. The cells in the lower layer, with maximum values of about 1,000 Sv, are the so-called thermohaline circulations that are driven by meridional density contrast due to temperature and salinity differences. Sea-ice formation generates salty water in the cold ice-covered region, while sea-ice melting in the warm open-ocean region freshens ocean water there. Denser water at high latitudes descends and moves equatorward, while lighter water in the tropics rises and moves poleward, forming the thermohaline circulations. These meridional ocean circulations lead to exchanges between warm tropical water and cold high-latitude water and generate net poleward heat transport. In addition to the mean MOCs, eddy-induced MOCs also transport heat poleward, as shown in Fig. S2A and B. However, eddy-induced MOCs are about one order weaker than the mean MOCs. As  $\text{CO}_2$  is increased, both mean and eddy-induced MOCs become stronger and thus transport more heat from the tropics to high latitudes.

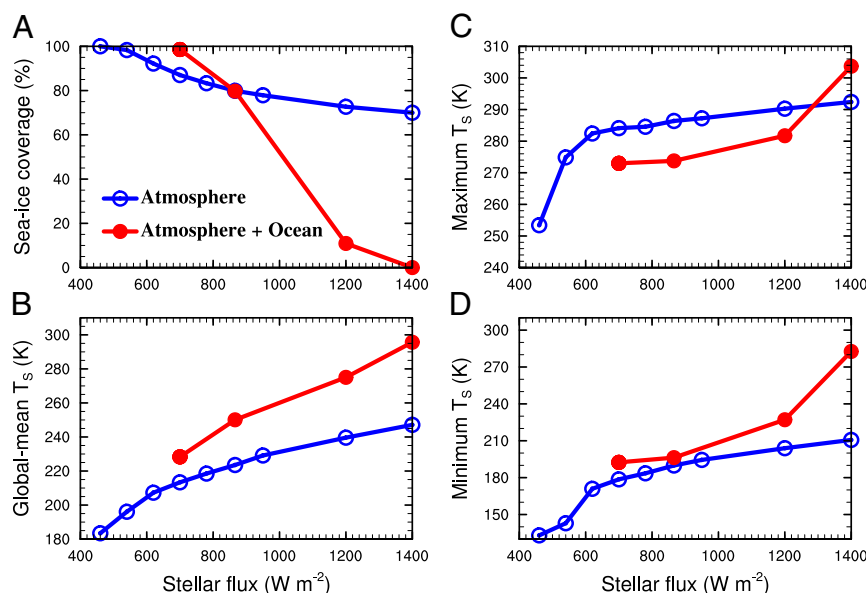
Atmospheric and ocean zonal heat transports from the dayside to the nightside are quantified and compared in Fig. 3. Both atmospheric and ocean zonal heat transports have maximum values at the equator and decrease with latitudes (Fig. 3A and C). This is consistent with the fact that both atmospheric and oceanic flows have maximum velocities at the equator. For 3.6 ppmv  $\text{CO}_2$ , the maximum zonal heat transport by the atmosphere is nearly twice greater than that by the ocean, i.e., 1.2 versus 0.7 PW. In general, atmospheric heat transport decreases with increasing  $\text{CO}_2$ . In contrast, ocean heat transport increases with increasing  $\text{CO}_2$ . For 200,000 ppmv  $\text{CO}_2$ , the maximum ocean heat transport is about twice greater than that by the atmosphere, i.e., 2.0 versus 0.9 PW. The decrease in atmospheric heat transport with increasing  $\text{CO}_2$  is because of largely reduced thermal contrast between the dayside and nightside (Fig. 1C and D). The increase in ocean heat transports with increasing  $\text{CO}_2$  is



**Fig. 2.** Depth–latitude cross-sections of zonal-mean ocean potential temperatures and zonal-mean zonal velocity. (Left) Ocean potential temperature (unit,  $^\circ\text{C}$ ); (Right) ocean zonal velocity (unit,  $\text{m s}^{-1}$ ); (Upper) 355 ppmv  $\text{CO}_2$ ; and (Lower) 200,000 ppmv  $\text{CO}_2$ . In C and D, yellow-red colors indicate westerly flows, blue colors indicate easterly flows, and contours are the mean meridional mass streamfunction. Solid contours indicate clockwise streamlines, and dashed contours are anticlockwise streamlines. Contour interval is 100 Sv.







**Fig. 4.** Comparison of sea-ice coverage and surface air temperatures as a function of stellar radiation fluxes between AOGCM (red) and AGCM (blue). (A) Sea-ice coverage; (B) global-mean  $T_s$ ; (C) the maximum  $T_s$ ; and (D) the minimum  $T_s$ . In both types of simulations, the  $\text{CO}_2$  concentration is fixed at 355 ppmv.

concentration and the tropopause height would be higher in AOGCM than in AGCM for stellar radiation stronger than  $1,400 \text{ W m}^{-2}$ . These results suggest that as a dynamic ocean is considered, an exoplanet would more readily enter the runaway greenhouse state if the exoplanet is close enough to the inner HZ edge and that ocean heat transport likely pushes the inner HZ edge outward.

## Discussion and Conclusions

We have demonstrated that ocean heat transports play critically important roles in determining the climate and habitability of tidally locked exoplanets in the HZ around M-type stars. In the presence of a dynamic ocean, the open-ocean area displays a lobster-like spatial pattern, instead of the “eyeball” state. For sufficiently high greenhouse-gas concentrations or strong stellar radiation, ocean heat transport is more efficient than atmospheric heat transport and can cause ice free on the nightside, greatly enlarging the habitable area of tidally locked Super-Earths (but no photosynthesis is possible on the nightside). Sea-ice thickness simulated by AOGCM here is much thinner than that simulated by models without a dynamic ocean. Furthermore, our simulations suggest that a dynamic ocean likely narrows the width of the HZ around M-type stars. The phase shift of surface temperatures between the hottest spots and the substellar point caused by ocean circulations may have observational consequences in both the infrared and visible phase curves of the system, which could be visible in future observational missions.

Exoplanets are likely to have a wide range of ocean depth or even have continents. Results in Fig. 2 suggest that ocean depth has important influences on ocean heat transports because the strong equatorial current and the warm ocean layer can both

extend downward to about 2 km for a strong greenhouse effect. Thus, both zonal and meridional ocean heat transports would largely decrease if ocean depth is shallower by one order. On the other hand, if ocean depth is thicker by one order, ocean heat transport would increase. Given that the ocean is separated by continents, especially by meridional continental barriers, continents would block ocean zonal heat transports from the dayside to the nightside. In such a case, the importance of ocean heat transport would be largely reduced, and the climate state would be more like the “eyeball” state although the open-ocean region is not a round area due to ocean heat transports over ocean-domain scales.

In addition to ocean heat transport, ice–albedo feedback and sea-ice dynamics also have contributions to climates of tidally locked exoplanets. Detailed analysis of these effects is beyond the scope of the present paper. We will report them in a separate paper. It is worth pointing out that it is difficult to make a full exploration on how ocean heat transport would alter the HZ width with current general circulation models (GCMs). Defining the outer and inner HZ boundaries involves high levels of  $\text{CO}_2$  and water vapor, respectively, and radiation transfer modules of current GCMs are not valid in resolving these problems.

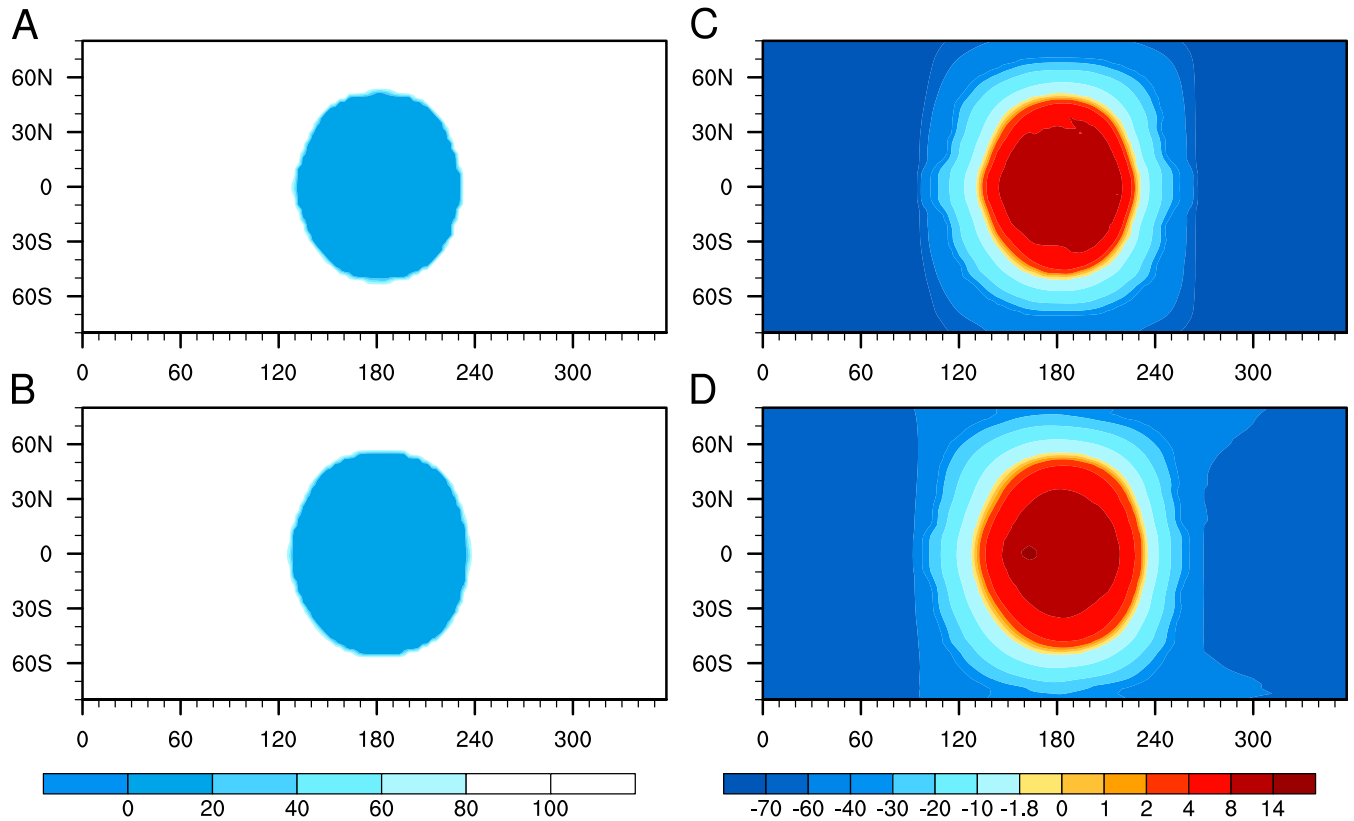
**ACKNOWLEDGMENTS.** We thank R. T. Pierrehumbert who sharpened our ideas and helped us greatly in revising early drafts of the paper. We also thank D. N. C. Lin for fruitful discussion on this subject. We are grateful to Y. Liu who helped us set up the model. This project was made possible through the support of a grant from the John Templeton Foundation. The funds from the John Templeton Foundation were awarded in a grant to the University of Chicago which also managed the program in conjunction with the National Astronomical Observatories, Chinese Academy of Sciences. This work is also supported by the National Natural Science Foundation of China (41025018) and by the National Basic Research Program of China (973 Program, 2010CB428606).

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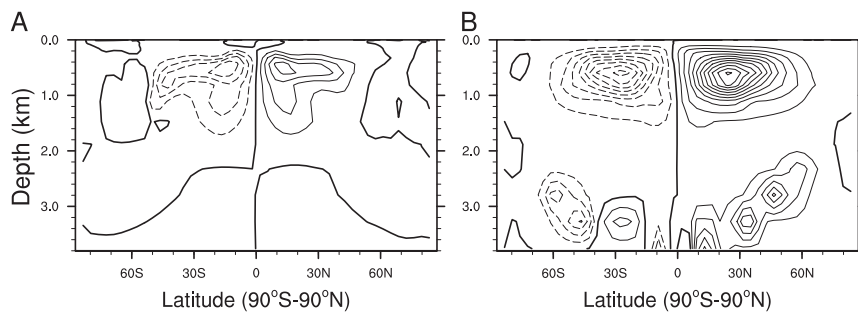
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# Supporting Information

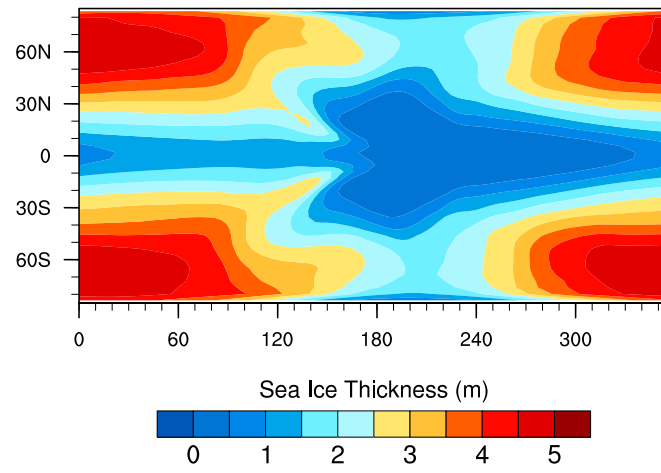
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**Fig. S1.** Sea-ice coverage and surface air temperature in AGCM simulations with a slab ocean. (Left) Sea-ice fraction (unit, %); (Right) surface air temperature (unit, °C); (Upper) 355 ppmv of CO<sub>2</sub>; and (Lower) 200,000 ppmv of CO<sub>2</sub>. For A and B, blue indicates open ocean, and white indicates sea ice. Note that the color scale is not linear in C and D.



**Fig. S2.** Mean meridional mass streamfunction of eddy-induced MOCs. Contour interval is 10 Sv. Shown are (A) 355 ppmv and (B) 200,000 ppmv of CO<sub>2</sub>. Solid contours indicate clockwise streamlines, and dashed contours indicate anticlockwise streamlines. The exoplanet studied here has a much slower rotating rate, 36.7 times slower than Earth's. Its gravity is 1.35 times greater than Earth's. Thus, the Rossby radius of deformation of its ocean is about 43 times larger than for Earth's oceans, according to the formula  $L_R = (gD)^{1/2}/f_0$  for a barotropic ocean as an example. Here,  $L_R$  is the Rossby radius of deformation,  $g$  is the gravity,  $D$  is the ocean depth, and  $f_0$  is the Coriolis parameter at the reference latitude.  $L_R$  is about 200 km for Earth's tropical ocean. Thus,  $L_R$  is about 8,500 km for the exoplanet studied here, which is much greater than the numerical resolution of the ocean model. Therefore, persistent ocean eddies are model-resolved, and eddies parameterized by the Gent-McWilliams parameterization are weak.



**Fig. S3.** Sea-ice thickness for 355 ppmv of CO<sub>2</sub>. Color interval is 0.5 m.