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Note

A new family of planets? "Ocean-Planets"

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Abstract

A new family of planets is considered which is in between the rocky terrestrial planets and the gaseous giants, "Ocean-Planets." We present the possible formation, composition and internal structure of these putative planets. We consider their oceans, as well as their possible Exobiology interest. These exoplanets should be detectable by Space missions such as Eddington, Kepler, and possibly COROT (launch scheduled in 2006). They have a density lower than that of rocky planets. Their rather large radius would make them attractive targets for exoplanet spectroscopic missions such as Darwin/TPF, all the more because a robust biosignature appears to exist.

Keywords: Extrasolar planets; Ices; Exobiology

1. Introduction

The extrasolar planetary systems discovered thus far show a surprising diversity of orbital parameters. Most of these systems do not resemble our own Solar System. There is presently no consensus as to why these planetary systems are so diverse, but it seems that migration due to interactions between planets and the protoplanetary disk is an important ingredient (Lin et al., 1996; Ward, 1997; Trilling et al., 1998). It seems reasonable to assume that planets resembling our Uranus and Neptune, or slightly less massive ones, may have formed in cold regions of a protoplanetary disk and migrated inward, possibly into the so-called "Habitable-Zone" where liquid water can be present at their surface. These planets would be extremely interesting as their large radius makes them rather easily detectable by transit missions (COROT, Eddington, Kepler) and analysable by Darwin/TPF. A planet with twice the Earth radius requires an integration time 16 times shorter than an Earth analogue, for the same distance and S/N conditions.

The interest of such planets is twofold, covering both Planetology and Exobiology. They would be a type of objects we do not have in the Solar System and would significantly extend the field of Planetology. The search for a form of life similar to that which has developed on Earth would open a new field in Exobiology because the conditions of the environment would be quite different from the terrestrial ones. The elements necessary to living bodies (P, S, Fe, Mg, Na, K, ...) could be brought to the surface by micrometeorites or found in the ocean as dissolved species. It could even tell us something about the emergence of life on Earth. For instance, if some form of life is discovered on an Ocean-Planet, it would indicate that it has occurred in the absence of Black Smokers because on these planets these structures are not expected, liquid water and silicates being separated by thousands of kilometres of ice.

Around stars with not too high a C/O ratio, planetesimals built in the cold regions of the protoplanetary disk contain a significant fraction of water ice. In our Solar System, this is the case for all the moons of the giant planets except Io. Uranus and Neptune can themselves be considered as "ice giants": their interior density is indeed very similar to that of compressed water ice (Podolak et al., 2000). However, Uranus and Neptune also contain about 1 to 4 Earth masses of hydrogen and helium in the form of an outer envelope. The case of the planets we consider in this Note is different because they contain much less hydrogen.

Of course, many parameters drive the structure and composition of planets. As planet formation is not well understood, and to limit the scope of this study, we choose to make the following assumptions:

(1) Only planets with masses in the range $1 < M/M_\oplus < 8$ are considered. The lower boundary is for the selection of planets that are easier to detect, and the upper one for objects that have not accreted a large amount of H_2 (Wuchterl et al., 2000). It is pointed out that the biggest objects of that type ($M \sim 6$–$8M_\oplus$) *are of special interest* because they are accessible to Transit detection and possibly to Radial Velocity measurements so that *their radius and mass could be determined simultaneously*.

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(2) These planets are supposed to form relatively far from the star (further than the snowline, e.g., $\sim$ 5–10 AU) and migrate inward (to $\sim$ 1 AU) on a timescale which is of the order of $\leqslant$ 1 Myr (e.g., Lin et al., 1996).
(3) We assume that the planets are mostly made of refractory material (metals and silicates) and ices. From what we know of the composition of the protosolar nebula, we envision respective fractions of $\sim$ 50% of metals and silicates, and $\sim$ 50% ices, by mass. Because of the relatively small masses and closeness to the star after the initial migration, the planets should contain only trace amounts of hydrogen and helium (Lammer et al., 2003). Most of the ices should be in the form of H_2O, but their exact composition is crucial and will be discussed further.

The questions that arise are:

(i) What should be the internal structure of these planets?
(ii) Under which conditions will a water ocean be present, and what will be its depth?
(iii) What are their atmospheric properties?

In this Note, we attempt to provide the first steps towards answering these questions.

2. Internal structure

2.1. An outline of the cooling history of a planetary water ball

The knowledge of the temperature of the planetary interior is essential for determining its structure. Depending on the T and P profiles, the interior may be solid, liquid or gaseous. For simplicity, we first examine the case of a giant ball made of pure water.

The energy acquired by the planet (mass M and radius R) is of the order of, e.g., $\sim (1/2)\varpi GM^2/R$, G being the gravitational constant, ϖ a factor accounting for the density distribution inside the planet ($\varpi = 3/5$ for a uniform density) and assuming that half of the energy was radiated directly to space (perfect gas approximation). This would correspond to a temperature $T_i \sim \mu\varpi GM/k_B R \sim 25{,}000$ to $75{,}000$ K for $M = 1$ to $6M_\oplus$ if all this energy was converted into heat and stored in the gaseous planet ($\mu \sim 18$ is the mass of a water molecule, k_B the Boltzmann constant, we used $R \sim 10{,}000$ to $20{,}000$ km). This estimate is very rough (for example, and among many simplifications, it neglects the fact that accretion did not entirely take place at the same time and that water is partially decomposed at these temperatures) but it shows that these planets could have had an early stage in which they were at least partially gaseous. We now show that this stage was necessarily short.

Considering first an isolated planet and assuming a gaseous atmosphere of pure water (with a grey opacity $\kappa \sim 0.1$ cm^2 g^{-1}, e.g., Auman, 1966). The level at which the atmospheric temperature T is equal to the effective temperature $T_{\rm eff}$ is for an optical depth ~ 1 (Eddington approximation) or pressure $P_0 \sim 2g/3\kappa$, g being the surface gravity, i.e., a few mbars. The assumption of an atmosphere of pure water vapour implies, in order for water to remain in vapour form at P_0, that $T_{\rm eff} \geqslant 270$ K. The planet can spend at most $\tau \sim E_g/(4\pi R^2 \sigma T_{\rm eff}^4) \sim 5$ to 20 Myr in this partially gaseous state (this timescale could be shorter because the effective temperature is likely higher at the beginning).

When the cooling of the atmosphere has led to a temperature lower than 270 K water begins to condense and falls down towards the planetary interior where it vaporises, absorbing the corresponding latent heat. This process provides an efficient cooling of the interior. Progressively, it cools down until condensed water cannot vaporise anymore and sinks towards the planetary centre as high-pressure ice. A thick ice shell forms. If the accretion rate is not high enough, the planet may never reach this gaseous state for water.

The case of a planet at a given distance from its star is discussed in Section 2.4. The basic cooling mechanism remains the same but the planet effective temperature is now governed by the radiative exchange with its surroundings. The migration time ($\leqslant$ 1 Myr) is shorter than the upper limit for the cooling time that we derived. Then, it is possible that an Ocean-Planet cools mainly when at $\sim$ 1 AU.

The presence of silicates/iron does not qualitatively alter this analysis: "rocks" are expected to rapidly differentiate from ices, settle towards central regions, and adjust thermally to the surrounding envelope by conduction.

Note that Io, Europa, and Ganymede are differentiated, as are all other major solid planets of the Solar System.

2.2. Composition of ices

We expect that initially, ices had a composition similar to that of comets, i.e., 90% H_2O, 5% NH_3, and 5% CO_2 by mass. Compared to the situation depicted in the pure ice ball case, this could cause major changes because the total amount of CO_2 and NH_3 is huge, several hundred times that present in Venus or the Earth (carbonates). If a significant part of these gases were in the atmosphere, it would have severe consequences: it would maintain the atmosphere into a hot state and prevent the formation of an ocean of liquid water. This is what happens in Uranus and Neptune: because they contain 1 to 4 Earth masses of hydrogen and helium (Podolak et al., 2000), even the small intrinsic heat flux maintains the atmosphere in a state such that the temperature at its bottom is larger than water's critical temperature (647 K). Consequently, the interiors of Uranus and Neptune are fluid (e.g., Cavazzoni et al., 1999).

However, four processes may limit the initial amount of CO_2 and NH_3 in the atmosphere:

(i) they are partially soluble in the liquid and solid. For example, at low pressure and $T \sim 300$ K, NH_4HCO_3 is soluble in water (12 g per 100 g water), as is $(NH_4)_2CO_3$ (100 g per 100 g water);
(ii) NH_3 and CO_2 are known to easily form hydrates/clathrates when the temperature is not too high (less than 280 K for CO_2) (Leliwa-Kopystynski et al., 2002; Sloan, 1998);
(iii) phase separation in the interior may sequester elements at deeper levels, in which case they would be unavailable to form a massive atmosphere.

The driving force for this sequestration is gravity. For instance, solid CO_2 is denser than solid H_2O, at least in the domain where experimental data are available ($P < 60$ GPa; Yoo et al., 1999; Hemley et al., 1987), e.g., by a factor 1.3 at 10 GPa and 1.2 at 50 GPa. If at high pressure H_2O and CO_2 separate into two phases the latter will sink down into the thick ice layer and *most of the* CO_2 *will be locked into the solid ice mantle*. How much carbon dioxide remains in the upper parts of the planet, including in the atmosphere, is an open question and conservatively, will be treated as a free parameter. However, it is pointed out that only a minute fraction of the CO_2 reservoir is in the atmosphere;

(iv) the evaporation processes discussed in Section 4.1 may erode some of the more volatile gases, especially in the early period of high X and EUV activity of the central star (Lammer et al., 2003).

2.3. Internal structure modelling

A model for planet interior is used that has been developed for the Earth interior and expanded to extrasolar terrestrial planets (Dubois, 2002). Quantities depend upon the radius, r (1D model). They are the planetary mass located between 0 and r, local density, gravity and pressure, $m(r)$, $\rho(r)$, $g(r)$, and $P(r)$, respectively.

Density is a function of material, pressure and temperature. The latter dependence is not major and will be neglected. High-pressure laboratory experiments have been performed that provide equations of state $\rho(P)$ for the different materials. The adopted relations are as follows:

– main metals in the centre of a telluric planet are Fe and Ni. The equation of state of iron by Anderson and Ahrens (1994), is used;
– main kinds of silicates in the Solar System are the Fe and Mg rich silicates. Their phase at high pressure is Perovskite ($P > 23$ MPa; Anderson, 1997) with the $\rho(P)$ relation as measured by Duffy and Arhens (1995);
– the phase diagram of H_2O is complex, with 10 ice phases (Durham and Stern, 2001). The ice $\rho(P)$ relation has been measured up to 128 GPa by Hemley et al. (1987) and we extrapolate it for higher pressures.

Inputs are: the total mass of the planet, the fractions of metals, silicates and ices. Outputs are the outer planetary radius, the outer radius of the different material shells, variation with radius of the different physical quantities.

For a given mass of the planetary components, a self-consistent solution of the density, gravity and pressure functions, $\rho(r)$, $g(r)$, and $P(r)$, is required that fulfils the mass conditions

$$M_i = 4\pi \int_{R_{i-1}}^{R_i} \rho_i(r) r^2\, dr, \tag{1}$$

where the index, i, corresponds to the metal, silicates and ice component, respectively, and R_i the outer radius of component (i); the gravity and pressure relations are

$$g(r) = Gm(r)/r^2, \tag{2}$$

$$P(r) = \int_r^{R_{\rm pl}} \rho(x) g(x)\, dx, \tag{3}$$

where $R_{\rm pl}$ is the planetary radius. *For a* $6M_\oplus$ *planet*, with the relative amounts of material as described in Section 1 and deduced from the Earth's values (Javoy, 1999), i.e., *m* 17% metals, 33% silicates, and 50% ices, the internal structure calculated is shown in Fig. 1. *The planetary radius is* $R_{\rm pl} = 2.0R_\oplus$, central pressure 1600 GPa and surface gravity $1.54g_\oplus$. The internal structure of a $6M_\oplus$ rocky planet and that of Earth are also shown for comparison.

2.4. Thermal structure

The interior structures may fall into three categories, depending on the stellar distance, orbital history and atmospheric composition:

(1) a hot gaseous planet where the temperature at $P = 22.1$ MPa (221 bars) is higher than the critical temperature of H_2O, 647 K, with a continuous transition from the outer atmospheric layers to the inner supercritical fluid (no surface);
(2) a planet with a liquid water surface;
(3) a planet with an icy surface. See also Stevenson (1999) for the case of isolated planets.

In the present paper we study case (2), that of an "Ocean-Planet," where an ocean is present with a surface temperature, T_S, in between the triple and critical temperature of water, 273 and 647 K. The Habitable Zone can be defined as distances to the star leading to case (2) and the fraction of case (3) where the surface ice can melt during the local summer.

Thereafter it is shown that the temperature profile $T(r)$, in the ice shell, follows the liquid/solid transition curve with pressure $T[P(r)]$ (Fig. 2 for low pressures).

The H_2O/silicate interface is derived to be at the temperature of the melting point of ice at the local pressure. For a $6M_\oplus$ Ocean-Planet, this pressure is 250 GPa (Fig. 1 caption) and the corresponding melting point of ice VII is 1150 K, as estimated from the Simon fusion equation in Mishima and Endo (1978). It must be noted that this equation predicts very well the later measurements by Fei et al. (1993).

It can be argued that the temperature at this interface is higher than the ice melting temperature and most of the water shell liquid. If so, a process analogous to that described in Section 2.1 would lead to the sinking of (high-pressure/high-density) ice and the fast build-up of an ice shell.

It can also be argued that the temperature at the H_2O/silicate interface is much lower than the melting temperature of ice VII and that internal heat is transferred by subsolidus convection in the ice layer. However, simple comparison with models built for the internal dynamics of the icy satellites of the giant planets in the Solar System shows that subsolidus convection in ice is not even sufficient to remove the heat produced by the decay of the radiogenic elements contained in the silicate shell (e.g., Grasset et al., 2000).

The only possibility is that heat is transferred by the melting of ice and migration of the melt towards the upper ocean. The amount of melt produced by the internal heating in the kind of planet described in this Note is on the order of 1 km thick layer per Myr at the interface. This is quite small compared to the size of the ice layer. Consequently, the temperature profile in the ice mantle follows the solid/liquid transition curve, i.e., from 1150 K at the silicate interface to the temperature at the bottom of the ocean.

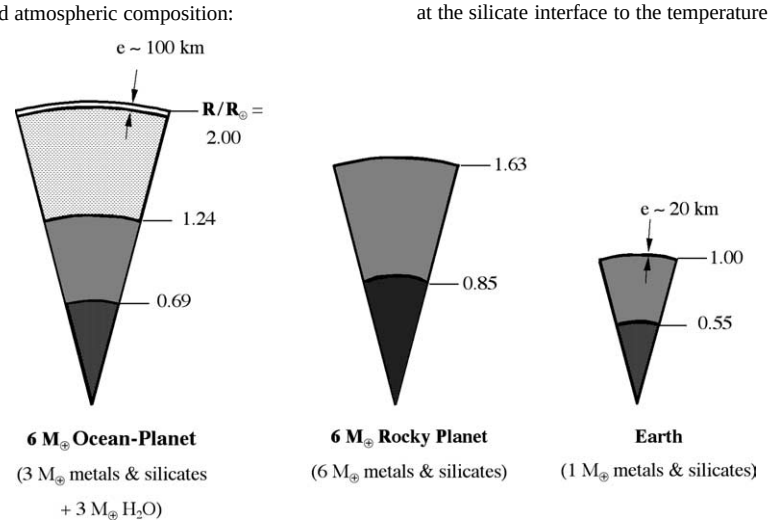

Fig. 1. From left to right: (1) calculated internal structure of a $6M_\oplus$ Ocean-Planet. Constituents are, from the centre to the outside, $1M_\oplus$ metals, $2M_\oplus$ silicates, and $3M_\oplus$ ice. The density (g cm^{-3}) at the centre, different interfaces and top is: 19.5, 15.6–8.2, 6.2–3.9, and 1.54; gravity ($g_\oplus$): 0, 2.1, 1.96, and 1.54; pressure (GPa): 1580, 735, 250, and ~ 1. The upper layer is a ~ 100 km thick ocean. The mean planetary density is 4.34 g cm^{-3}; (2) idem for a rocky planet with the same total mass ($2M_\oplus$ metals, $4M_\oplus$ silicates). Density is: 21.0, 15.5–8.1, and 4.1; gravity: 0, 2.72, and 2.24; pressure: 2200, 745, and 0. The mean planetary density is 7.74 g cm^{-3}; (3) for comparison, the structure of the Earth calculated with the same model is shown. It agrees fairly well with the actual one. Density is: 13, 9.5–5.2, 3.3; gravity: 0, 1.04, 1.00; pressure: 340, 130, 0. The mean planetary density is 5.57 g cm^{-3}.

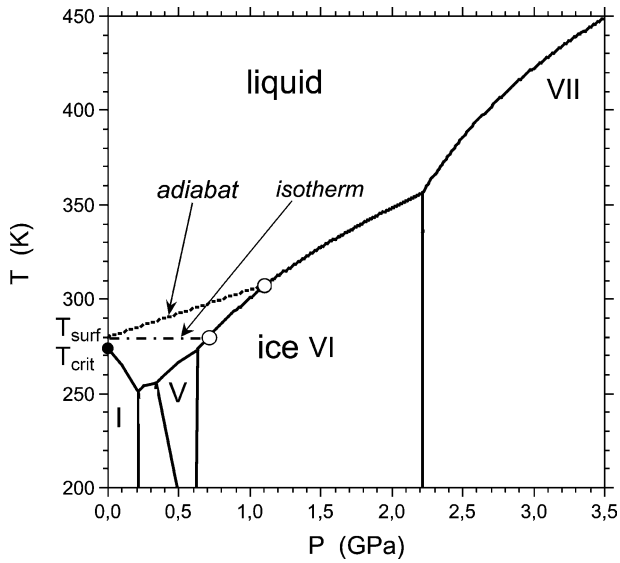

Fig. 2. Phase diagram for water (full lines). The dotted (dashed-dotted) line describes the temperature adiabatic (isothermal) profile in the ocean that results from a given surface value (here, 280 K = 7 °C). As explained in the text, the actual dependence is expected to be in between the adiabatic and the isothermal cases. In the adiabatic case, the ocean depth is 72 km, the ocean bottom temperature and pressure are 35 °C and 1.1 GPa. For a higher (lower) surface temperature, the ocean depth would be larger (smaller), i.e., $T_{\rm surf} = 30\,^\circ$C (0 °C) leads to a 133 km (60 km) ocean depth. A flatter T profile would lead to a shallower ocean. In the case of a isothermal profile, the depth is 40, 45, and 65 km for $T_{\rm surf} = 0$, 7, and 30 °C, respectively. The gas/solid or liquid transition curve is, at the scale of the figure, indistinguishable from the $P = 0$ vertical line. The filled circle is the triple point. The critical point ends that curve at ($T = 647$ K, $P = 22.1$ MPa). In the solid phase, the temperature follows that of the liquid/solid transition (Section 2.4).

The pressure in the silicate shell is larger than what is found in the Earth's interior. The behaviour of silicates in this domain is not known. By analogy with our planet, we assume that the silicates are composed of perovskite and magnesowustite and that the fossil heat plus that produced by the decay of long-lived radiogenic elements in the silicate shell, are transferred by subsolidus convection (e.g., Schubert et al., 2000). Without entering into the details of this convection one can say that the temperature profile within the silicate shell is made of a conductive temperature gradient in the cold thermal boundary at the top of the shell, an adiabatic temperature gradient in the well-mixed interior and a conductive temperature gradient at the hot thermal boundary layer at its bottom.

A picture of the temperature profile results. In the metal/silicate core, the temperature decreases from a high value in the centre (e.g., $\sim 10{,}000$ K for a $6M_\oplus$ Ocean-Planet by comparison between with the temperature at the centre of the Earth, 5000 K, and that for Jupiter, 20,000 K) to the ice fusion temperature at the silicate/ice boundary (e.g., 1150 K for that planet). In the ice shell, it follows the water liquid/solid transition to the ocean bottom temperature. The latter is discussed in Section 3.

3. Ocean

When such a planet is in the Habitable Zone of its star, its surface is liquid. A question arises, "how deep is the ocean?." As a first step, for the sake of simplicity, an ocean of pure water is considered.

The temperature is a function of depth, z, with $z = R_{\rm pl} - r$, and therefore pressure. The internal heat of the planet is transferred through the ocean. The *maximum* temperature gradient is adiabatic, otherwise convection in water would be triggered which is extremely efficient to transfer energy. We assume an adiabatic dependence for the temperature but it could be weaker, down to an isothermal one. One can note that for the Earth's ocean, the temperature at the sea floor is even lower than the temperature at the surface as a result of streams.

The adiabatic relation is:

$$dT/dP = \alpha T/(\rho_{\rm water} C_{\rm p}), \tag{4}$$

where α is the thermal expansion coefficient and $C_{\rm p}$ the heat capacity. For $P < 1$–2 GPa, values are $\alpha \approx 3 \times 10^{-4}$ K^{-1}, $\rho_{\rm water} \approx 1$ g cm^{-3}, $C_{\rm p} \approx 4$ J K^{-1} g^{-1}, and the temperature is given by:

$$T_{\rm adiabat}(z) \approx T_{\rm surface} + 20P \text{ (GPa)}. \tag{5}$$

The T $[P(z)]$ position in the phase diagram indicates whether water is liquid or solid (Fig. 2). The $P(z)$ relation depends upon the surface gravity g, the liquid water density at low pressure ρ_0 and its compressibility K. It reads $P(z) = -K\,{\rm Ln}(1 - g\rho_0 z/K)$. The bottom of the ocean is obtained when the liquid water temperature crosses the solidification line. The ocean depth, bottom temperature and pressure result (Fig. 2). For an adiabatic dependence of the temperature, a surface temperature $T_{\rm surf} = 7\,^\circ$C leads to an ocean depth of 72 km. For a higher (lower) $T_{\rm surf}$, the ocean depth would be larger (smaller).

If the T dependence is less steep than adiabatic, e.g., nearly isothermal, the ocean would be shallower.

4. Atmosphere

4.1. Composition

On Ocean-Planets, the expected primary volatiles are H_2O, NH_3, and CO_2. As mentioned, we assume, as an example, an initial composition of ice m 90% H_2O, 5% NH_3, and 5% CO_2. For carbonaceous compounds, this is an important difference with the atmosphere of bodies further away from their stars, as Titan in the Solar System, where carbon is mainly in species such as CH_4 because of the lower temperatures (Reynolds et al., 1987).

NH_3 is very sensitive to UV, especially for $200 < \lambda < 300$ nm where the opacity due to other species is negligible. The fraction of its initial reservoir that happens to be in the atmosphere will be photodissociated and converted into N_2 and H_2 in less than 2 Myr for a planet located at 1 AU from a G2V star. The produced H_2 is subject to hydrodynamical escape and can sweep away a fraction of N_2 because the atmosphere develops an exosphere with a high temperature $T_{\rm exo}$ governed by heating by extreme UV and stellar particles (Lammer et al., 2003). $T_{\rm exo}$ obtained with the XUV heating alone can be over 10,000 K which is much higher than the planet effective temperature $T_{\rm eff}$ (255 K for Earth).

The presence of these radiations has been observed in solar proxies by numerous satellites (ASCA, ROSAT, EUVE, FUSE, IUE). It indicates that a young GV star has continuous flare events producing a radiation environment several hundred times more intense than the solar environment today (Guinan and Ribas, 2002).

After the majority of the hydrogen is lost and the heavier atmospheric compounds begin to dominate in the upper atmosphere, a part of the nitrogen should be dragged away by a diffusion limited hydrogen flow (Zahnle et al., 1990). Because the amount of nitrogen left is not sure, the partial pressure of N_2 in the atmosphere is treated as a free parameter in the present paper. So is the CO_2 pressure.

4.2. Planet IR emission

The planet IR emission must be calculated in a self-consistent way as a function of the atmospheric composition and the stellar irradiation. A key ingredient is the temperature profile. It is to be remembered that an isothermal atmosphere would have no spectral feature, whatever its composition. As a preliminary calculation, for a given ground temperature, an atmosphere

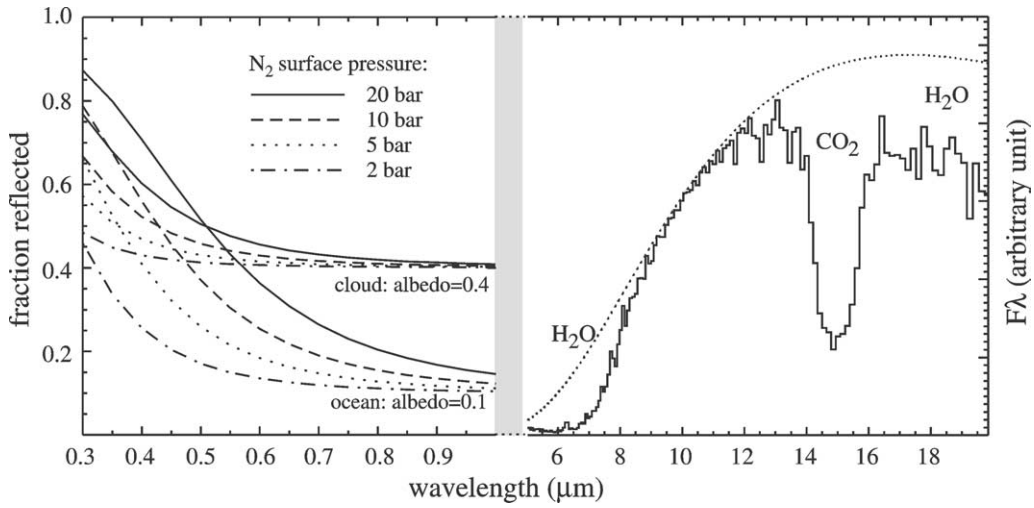

Fig. 3. Reflected and emitted light by a $6M_\oplus$ Ocean-Planet, versus wavelength. In the UV-visible, the fraction of the stellar flux reflected by the planet is shown for different values of the ground pressure of N_2, this gas being the main atmospheric component, and for situations with and without clouds. The light from the planet is the product of the stellar emission by this fraction. The former decreases with wavelength ($\propto \lambda^{-4}$) and becomes negligible at $\lambda > 3$ μm. In the IR, the emission is from the same planet with $T_{\text{surface}} = 320$ K $= 47\,°$C, $P_{N_2} = 5$ bars and an atmosphere saturated with water vapour. Its cold tropospheric trap ($T \sim -65\,°$C) is at 15 km. It contains CO_2 with a mixing ratio of 10^{-5}. As it is wet, and pretty warm at its surface, the planet is assumed to be cloudy with a full coverage of clouds at $h = 5$ km. Note the CO_2 absorption band at 15 μm and the strong H_2O ones at $\lambda < 8$ μm and $\lambda > 16$ μm.

saturated with water vapour is considered and an adiabatic decrease of T is calculated down to an arbitrary value of 180 K. This limitation is to prevent T from reaching unrealistic low values.

Figure 3 shows the interesting case of an Ocean-Planet with a 320 K (47 °C) surface temperature. This case of a rather hot planet with a super-humid atmosphere has no analogue in the Solar System. Both stellar reflected light and IR emission are shown.

4.3. Biosignatures?

Usually, O_2 or O_3, in association with H_2O and CO_2 are candidates biosignatures in a planetary atmosphere. To qualify them for an Ocean-Planet, one has to show that there are no abiotic processes that can produce a similar gas mixture.

Two cases of planetary atmospheres may result in a photochemical build-up of O_2 (Selsis et al., 2002):

(i) CO_2-rich atmospheres: $P_{CO_2} > 50$ mbar for dry atmospheres, $P_{CO_2} > 0.5$ bar for wet atmospheres, and
(ii) water-rich atmospheres loosing hydrogen to space after H_2O photolysis (for instance during a runaway greenhouse effect as described by Kasting, 1988).

Both cases require weak oxygen sinks (oxidation of rocks by soil weathering, and volcanic gases) to allow O_2 to reach significant levels.

Ocean-Planets may gather all these characteristics: a possible high CO_2 abundance, a water-rich upper atmosphere when the orbital distance is closer than the runaway greenhouse limit (Kasting et al., 1993) and low oxygen sinks in the absence of rocky surface and volcanism. Especially, due to their inexhaustible water reservoir, runaway greenhouse phases may last for most of their lifetime while, on Venus for instance, this phase is thought to be shorter than a few hundred million years (Kasting, 1988). Consequently, *the presence of* O_2 *is not a reliable biosignature on Ocean-Planets* because it can be the result of abiotic processes.

Can O_3 provide a better one? If O_2 is produced thanks to H_2O photolysis at high altitude, resulting hydrogenous radicals as $H^\bullet$, $OH^\bullet$, and $HO_2^\bullet$ react efficiently with O_3 and prevent the formation of a dense and detectable ozone layer. Producing abiotic O_2 from CO_2 photolysis implies a high CO_2 partial pressure that masks the O_3 band in the thermal infrared spectrum. Thus, as on terrestrial planets (Selsis et al., 2002), *the simultaneous detection of* O_3, H_2O, *and* CO_2 in the planetary mid-IR spectrum appears to be most reliable signature of an oxygen build-up of biological origin. This points out *the superiority of* O_3, *with respect to* O_2, *as a biosignature* (in presence of H_2O and CO_2).

5. Conclusion

We have shown that massive ice-rich planets possibly form in external regions of protoplanetary disks and migrate inward. Depending on their distance to the star and properties of their atmospheres, some of them can develop a surface water ocean and can be called "Ocean-Planets." Such oceans have a thickness of ∼ 100 km. For a given mass, they have a density significantly lower than rocky planets and therefore a larger radius, a favourable feature for their detection and study.

Mid-future space missions searching for planetary transits in the Habitable Zone (Eddington, Kepler), possibly coupled with Radial Velocity follow-up, should provide us with valuable information about their existence and properties. If they are as resistant, with respect to the evaporation and photolysis of their atmospheres as some models predict (Kuchner, 2003) COROT (launch scheduled in 2006) will detect the hottest ones. They have a density lower than that of rocky planets. Their rather large radius makes them attractive targets.

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