A new perspective on interiors of ice-rich planets: Ice-rock mixture rather than a layered structure

Allona Vazan,1,2* Re’em Sari,1 Ronit Kessel3
1 Racah Institute of Physics, The Hebrew University, Jerusalem 91904, Israel
2 Department of Natural Sciences, The Open University of Israel, 435701 Raanana, Israel
3 Institute for Earth Sciences, The Hebrew University, Jerusalem 91904, Israel

3 November 2020

ABSTRACT
Ice-rich planets formed exterior to the iceline and thus are expected to contain substantial amount of ice (volatiles). The high ice content leads to unique conditions in the interior, under which the structure of a planet may be affected by ice interaction with other metals. We use experimental data of ice-rock interaction at high pressure, and calculate detailed thermal evolution for possible interior configurations of ice-rich planets. We model the effect of migration inward on the ice-rich interior by including the influences of stellar flux and envelope mass loss. We find that rock and ice are expected to remain mixed, due to miscibility at high pressure, in most of the planet interior (>99% in mass) for a wide range of planetary masses. We also find that the deep interior of planetary twins that have migrated to different distances from the star are usually similar, if no mass loss occurs. Significant mass loss results in an interior structure of a mixed ice and rock ball, surrounded by a volatile atmosphere of less than 1% of the planet’s mass. In this case, the mass of the atmosphere of water / steam is limited by the ice-rock interaction. We conclude that when ice is abundant in planetary interiors the ice and rock tend to stay mixed for giga-years, and the interior structure differs from the simple layered structure that is usually assumed. This finding could have significant consequences on planets’ observed properties, and it should be considered in exoplanets characterisation.

Key words:

1 INTRODUCTION
Planets are formed by accretion of solids and gas from the protoplanetary disk (e.g., Pollack et al. 1996), and their composition depends on the formation location in the disk. Ice-rich planets were formed exterior to the iceline, where ice is as abundant as rock (Lodders 2003). Thus, they contain a significant fraction of ice (volatiles). Based on formation models, these planets didn’t reach the runaway gas accretion phase to become gas giant planets, but usually had conditions to accrete a fraction of gaseous envelopes (Mordasini et al. 2012; Lambrechts & Johansen 2014; Morbidelli et al. 2015). The interior structure of these planets is poorly understood. The interior of ice rich planets is usually modeled as a 4 layer object: iron core overlain by a rocky mantle, surrounded by an icy shell and in some cases covered with a gaseous (usually hydrogen and helium) envelope (e.g., Rogers & Seager 2010; Rogers et al. 2011; Dorn et al. 2017). While this structure is assumed for simplicity based on density consideration, it should be reexamined for two main reasons:

(1) The differentiation of the planet interior into distinct layers by gravity occurs only if the interior material is demixed (immiscible). Material mixing / demixing is determined by the thermodynamic properties of the components, namely, their tendency to interact as a function of pressure and temperature (Stevenson 2013). Laboratory experiments of ice and rock interaction (e.g., Kessel et al. 2015) show that ice and rock are miscible in each other at high pressure. Yet, no attempt was made to link these experimental results to thermal evolution of ice-rich planets.

(2) In the basic Core Accretion model accreted solids (metals) form a solid core, surrounded by a solar composition gaseous envelope (Pollack et al. 1996; Alibert et al. 2005). But detailed calculations of material and energy deposition during solid accretion indicate ablation of significant fraction of the solids in the gaseous envelope, resulting in a gradual metal distribution in the planet interior (Lozovsky et al. 2017; Brouwers et al. 2018; Bodenheimer et al. 2018; Helled & Stevenson 2017; Valletta & Helled 2019, 2020). Gradual composition distribution in the interior is consistent also with measurements of our outer solar system planets (Marley et al. 1995; Podolak et al. 2000; Leconte & Chabrier 2013; Vazan et al. 2016; Wahl et al. 2017; Vazan et al. 2018a; Debras & Chabrier 2018).
Thus, planets with gas envelopes may have gradual metal distribution in the interior.

Although ice-rich planets formed exterior to the ice-line, common planet-disk interaction theories predict planetary migration in the protoplanetary disk to be pervasive (Ward 1997; Papaloizou 2002; Bitsch et al. 2013; Mordasini 2018). Observational evidence of an evaporating or disrupted ice-rich planet around white dwarf (Veras & Fuller 2020) and of water in atmosphere of close-in Neptune (Kreidberg et al. 2020) indicate that ice-rich planets may get close-in and maintain their ice. When a planet migrates inward in the protoplanetary disk, its interior may change by the higher stellar flux over billions of years. The stellar flux reduces the planet luminosity and thus slows down the planet interior cooling. In addition, the higher stellar flux absorbed by the outer layers of the gas envelope can stimulate mass loss processes. Both effects can affect the thermal evolution of the interior.

Here, we explore possible interior configurations of ice-rich planets. We include experimental results of ice-rock interaction, possible initial structures, and migration effects. In section 2 we draw our model and the recent knowledge of interior formation, material interaction, and non-adiabatic thermal evolution. In section 3 we present resulting interior structures of planets with and without mass loss. We discuss implications and caveats in section 4, and summarise our conclusions in section 5.

2 MODEL

2.1 Initial structure

Exterior to the iceline rocks and ices are accreted together as solids. We assume that the solid building blocks outside the ice line are composed of 50% ice and 50% rock in mass (Lodders 2003). It is unclear whether the ice and rock stay mixed during the accretion process. Some models of solid accretion suggest evaporation of the ices in the envelope, while the rocks accreted to form the core (Podolak et al. 1988; Hori & Ikoma 2011; Venturini et al. 2016). However, later studies show that rocks also dissolve in the deep envelope when the core mass is smaller than a few Earth masses (Brouwers et al. 2018; Valletta & Helled 2019; Brouwers & Ormel 2020). Although ice vaporises at higher altitudes, when both ice and rock vaporises in the envelope they may tend to mix (Lock & Stewart 2017). In addition, large accreted solids that reach the deep interior without breaking (e.g., km size planetesimals), keep the ice to rock ratio in the deep interior similar to the ratio in the disk. Clearly, the exact ice and rock distribution in the interior requires further studies. Yet, formation studies hint that significant fractions of ice and rock are expected to be mixed in the planet interior at the end of the formation phase. For simplicity, and as a first order study, we assume here that all of the ice and rock are initially mixed in the interior.

A metal-rich planet has less gas than metals. Usually, if the gas to metal mass ratio exceeds unity the planet reaches a runaway gas accretion phase and becomes a gas giant (Pollack et al. 1996; Brouwers & Ormel 2020). The amount of accreted gas depends on parameters such as radiative opacity, accretion rate, and distance from the star (Lammer et al. 2014). Radius-mass relation of observed sub-Neptunes predict a gas fraction of up to 10% in mass (Lopez & Fortney 2014), consistent with theoretical studies (Ikoma & Hori 2012; Bitsch et al. 2015; Johansen & Lambrechts 2017). The ice-rich candidates in our solar system, Uranus and Neptune, have about 10%-20% of gas in mass (Helled et al. 2011; Nettelmann et al. 2013). Thus we take envelope of 10% hydrogen and helium (H,He) in solar ratio to be our standard initial case. As will be shown below, the exact fraction of gas doesn’t change the conclusions we make in this work about interior structure of ice-rich planets.

We consider two different initial structures, as representatives of two formation model approaches: gradual metal distribution and core-envelope structure. The gradual structure, motivated by new planet formation models (Valletta & Helled 2020; Ormel et al. 2020), has a deep pure metal (Z) region, followed by a gradual decrease in Z down to Z=0.1 in the outer envelope. For comparison we consider also the same composition in a simple core-envelope structure, with a pure Z core surrounded by a H,He envelope with uniform Z=0.1 enrichment. In Fig. 1 we show the initial density and metal distribution for 15 M\(_\oplus\) planets of these two models.

To summarise, our initial standard models are of gradual composition structure and of core-envelope structure. For clarity we focus our examples in this paper on 15 M\(_\oplus\) planets, initially composed of 90% metals (ice and rock in 1:1 ratio), and 10% H,He (solar ratio). For completeness we tested also models of smaller planetary masses - 10 M\(_\oplus\) and 5 M\(_\oplus\).

2.2 Ice-rock interaction

After formation, the long term ice and rock distribution depends on the tendency of ice and rock to mix or demix in different P-T regimes (Stevenson 2013, 2014). In planetary interiors separation (demix) leads to a differentiation into different composition layers by the planet’s gravity field, based on their densities. If, on the other hand, ice and rock are chemically mixed (miscible) then no differentiation into layers takes place. Laboratory data of ice-rock interaction in the context of Earth composition exists for many years.
The Earth-like planet would have mixed ice-rock in its interior. The Earth curves indicate that an ice-rich Earth-like planet would have mixed ice-rock in its interior.

To apply the ice-rock interaction we use data from laboratory experiments at high pressure by Kessel et al. (2015). In these experiments the authors studied the interaction of water ice and Earth-like rock at pressures of a few GPa. They determined the pressure-temperature of mixing and demixing of the two materials. Based on these experiments we derive the curves for ice-rock melting (wet solidus) and the ice-rock mixing (critical curve), as explained in appendix A below. In Fig. 2 we present the resulting map of mixing and demixing of ice and rock: grey shaded area signifies where ice and rock are completely mixed (miscible), brown shaded area is where ice and rock are partially mixed, and in the white area ice and rock are demixed (immiscible, separated). The water vapor curve of Alduchov & Eskridge (1996) appears in dashed black indicates the liquid-gas phase transition of pure water.

The boundaries between areas in Fig. 2 are an example for Earth-like rock and water in 1:1 mass ratio. The exact values and boundaries may vary with rock type (e.g., Kessel et al. 2005), and water fraction (see appendix A). Moreover, laboratory experiments have error bars on the measurements, and some inconsistencies between studies. However, those uncertainties - in the range of one GPa and of a hundred Kelvin - are quite small in the perspective of the entire pressure-temperature range of planetary interiors.

For reference we plotted in Fig. 2 pressure temperature profiles of current Earth (Pearson et al. 2003; Nomura et al. 2014), and possible structure of Jupiter (Vazan et al. 2018a). As is shown, the Earth and Jupiter are mainly in the complete ice-rock mixture regime. However, while Jupiter probably formed exterior the ice-line and thus is ice-poor, if the Earth were formed in the ice-rich region of the protosolar disk\(^1\), the ice would have been mixed with the rock in most of its interior, probably also preventing the differentiation of the iron core (see next paragraph). Moons, on the other hand, are expected to be differentiated. For example, the large moon of Jupiter - Ganymede - has a central pressure of about 10GPa and central temperature of about 1500K (Shibazaki et al. 2011), already below the miscibility regions. In this work we apply the data from Fig. 2 to thermal evolution of ice-rich planets, and examine their ice-rock mixing tendency.

**Iron-ice-rock interaction** The formation of Fe-dominated metallic cores is a consequence of insufficient O/Fe ratio (Trounnes et al. 2019). The fraction of oxygen in ice-rich planets is much larger than in terrestrial planets, and the iron is expected to be oxidised and therefore remain in the rock rather than segregate as metallic core alloys, while the leftover hydrogen dissolves in the rock or escapes by outgassing. As a result, the mixed interior structure scheme probably holds also for iron. Nevertheless, this estimate is based on limited knowledge. Experiments of ice-rock-iron mixtures in ice-rich conditions are lacking, and further studies are required to constrain its properties in planetary conditions.

\(^1\) The data in Fig. 2 cannot be applied to the real dry Earth, because dry and wet rock differ in their properties. The Earth curves indicate that an ice-rich Earth-like planet would have mixed ice-rock in its interior.

### 2.3 Thermal and structural evolution

Thermal evolution determines the planet temperature profile in time. The simplest approach for thermal evolution is of adiabatic planets (isentrope). While this approach can fit to some level a structure of 2-3 uniform shells, it cannot be used to model gradual composition distribution in planetary interiors, because it disregards the mutual effect of composition distribution on the heat transport (Ledoux 1947; Leconte & Chabrier 2012; Vazan et al. 2015; Müller et al. 2020). These effects can be significant also in interiors of ice-rich planets, as shown in Vazan & Helled (2020) for the interior of Uranus.

Our thermal evolution model is based on Vazan et al. (2018a) with modifications to metal-rich sub-Neptune planets, as described in Vazan et al. (2018b). The model allows for heat transport by convection, radiation and conduction, depends on the local conditions in time. The model encapsulates the mutual effect of composition distribution on the heat transport, and the non-adiabatic thermal evolution when required. In convective regions we calculate the change in structure by convective mixing.

We use equation of state (EoS) for a mixture of ice, rock, hydrogen and helium (improved version of Vazan et al. (2013)), based on the additive volume law. For simplicity, we first calculate an EoS for a mixture of ice and rock. Our rock and ice EoS includes a liquid-solid phase and a vapor phase, and thus can be used to model also steam atmospheres. We calculate the mixtures of ice, rock, hydrogen and helium in the evolving structure self-consistently during the thermal evolution. For radiative opacity we use the method of...
Valencia et al. (2013) to fit Freedman et al. (2008) tabular opacity for planetary atmospheres. The metal fraction in the opacity calculation is linked to the outer envelope metallicity in the model. Radioactive heating by long term radioactive elements in the rock, and heat from solid contraction are also included in the model, as described in Vazan et al. (2018b).

2.4 Migration and mass loss
The stellar flux is included as a boundary condition of the interior model. The optically thin outer atmosphere is assumed to be grey and plane parallel (Vazan et al. 2018b). We use a simplified model of linear migration, where we change the distance (stellar flux) from the formation location to a final location within 3 Myr. The migration timescale (Myr) is much shorter than the evolution timescale (Gyr), and therefore the exact migration path is negligible for the aims of this work.

As the planet migrates close enough to the star, the high stellar flux may cause mass loss. Mass loss from migrated-in planets can be via several mechanisms, among them photoevaporation (Owen & Wu 2013, 2017), hydrodynamic wind (Ginzburg et al. 2016; Gupta & Schlichting 2020), Jeans escape (Lammer et al. 2003; Zeng et al. 2019), and photodissociation (Howe et al. 2020). Current mass loss predictions vary with uncertain stellar properties and timescales (e.g., Mordasini 2020; King & Wheatley 2020). Metal-rich envelopes add more complication to the mass loss calculation by affecting the planet radius, atmospheric composition and mean molecular weight (Venturini et al. 2020b; Malsky & Rogers 2020).

As of today, theories for mass loss from metal-rich envelopes are lacking and fail to explain some of the observations (Kasper et al. 2020). Since in this work we focus on the effect of mass loss on the interior structure, and not on the exact mass loss process, we consider here 3 simple edge scenarios for close-in planets: no loss of mass, loss of hydrogen only, and loss of the entire gas fraction.

3 RESULTS
The results are shown at planetary ages of 10 Gyr. Earlier in the evolution temperatures are higher and therefore thermal differences between models are smaller. In addition, at higher temperature more mass is in the ice-rock mixed area, as shown in Fig. 2. Therefore, the models at the age of 10 Gyr are the lower bound for the effects we study here.

3.1 Interiors of different structure and location
In Fig. 3 we show two types of interior structures of 1.5 M\(\odot\) ice-rich giants, as evolved from the structures in Fig. 1: core-envelope (blue) and gradual Z distribution (red). The initial energy content and composition mass fraction of all cases are similar. The difference between the temperature profile of the core-envelope structure and the gradual Z structure on Gyr time is due to the composition distribution effect on the heat transport. The composition gradient suppresses convection and therefore the heat flux outwards is lower. As a result the deep interior of the gradual Z case is much hotter, while the outer layers cool more efficiently. We also find that the gradual structure maintains the composition gradient from formation, with minor redistribution by convective-mixing, mainly in the outermost convective zone. This result is anticipated for a steep composition gradient which are more difficult to mix (e.g. Vazan et al. 2016). Since metal-rich interiors must have steep metal gradients (shallow metal gradients require large gas fraction) we find that convective-mixing is insignificant in planets with metal-rich interior.

On top of the planet profiles in Fig. 3 we show the ice-rock mixing and demixing areas from Fig. 2. We find that the ice-rock mixed area covers >99% of the planet mass for all cases, where the metal-rich interior (marked in arrows) is clearly within the ice-rock miscibility region. We also plot in Fig. 3 the water-hydrogen miscibility curve of Ball et al. (2013). Above the water-hydrogen miscibility curve (green) the water is mixed with the hydrogen\(^2\). Consequently, ice and rock are well mixed in the deep interior and the inner envelope, and demixed in the outer envelope, where ice is miscible in hydrogen.

We also compare in Fig. 3 similar planets at different distances from the star. The planets have the same initial properties, one is located at 10AU (dashed) and one migrated to 0.1AU (solid) from the star. No mass loss is considered here, to isolate the effect of stellar irradiation on the cooling. As can be seen, even after 10 Gyr of evolution the deep interior is only slightly affected by the stellar irradiation, especially if the planet composition is gradually distributed. Although the outermost layers of the closer-in planet expand by the stellar flux, causing a significant radius increase, these layers contain little mass, and the deep interior is almost not affected. Unlike in gaseous planets, the radius (volume) of the metal-rich deep interior isn’t very sensitive to its energy content (Seager et al. 2007; Valencia et al. 2006; Rogers et al. 2011) and the structure remains similar. Thus, the deep interior of twin planets is similar for a close-in (0.01AU) and a further out (10AU) planets, if no mass loss takes place.

3.2 Interiors after mass loss
As the planet loses its lighter materials by the stellar flux, the mean molecular weight of the outermost layer increases. The heavier layer on top of a lighter layer drives inward mixing in the envelope. This process replenishes the outer layer with light materials. In the case of gradual composition distribution the mass loss process slowly flattens the interior composition gradient outside-in, allowing for more rapid cooling by large scale convection.

In the upper panel of Fig. 4 we show the pressure-temperature profile of initially 15 M\(\odot\) planet with gradual composition distribution that migrated to 0.05AU, under 3 different mass loss assumptions: no mass loss (cyan), mass loss of all hydrogen (blue), and mass loss of all hydrogen and helium (brown). We find that the completely demixed mass fraction of ice and rock is <0.01% for the cases shown in this figure. Namely, even without gaseous envelope most of the ice and rock in the interior are mixed. In the bottom panel we show the metal (ice+rock in 1:1 mass ratio) distribution in the interior. The metal distribution is affected by the loss of light materials and by the moderate convective-mixing.

We emphasis the interior structure for the 3 cases of mass loss and for the two interior structures in Fig. 5. As all the gas is lost (right sketch) the two interior structures become similar. The lower pressure in the outermost layers leads to ice-rock demixing. Consequently, the denser rock sinks to the mixed layer below, while

\(^2\) Under these conditions hydrogen is expected to be partially miscible also in molten rock (Kite et al. 2019).
the water form an atmosphere around the mixed interior. For high stellar flux the volatiles are in the form of vapor, building a steam atmosphere (Chambers 2017; Turbet et al. 2020).

Ice-rock tendency to demix on the surface (low pressure conditions) changes with temperature, and therefore is related to the distance from the parent star. In Fig. 6 we show the temperature-pressure profiles of planets without gas envelope, similar to the brown curve in Fig. 4, that have migrated to different distances from the star (solid colors). The level of ice-rock demixing on the planet’s surface depends on the irradiation by the parent star: ice-rock are mixed up to the surface at 0.02AU (red), ice and rock are partially mixed on the surface at 0.03AU (gold), steam (water vapor) atmosphere is formed at 0.1AU (blue), and partially condensed water atmosphere at 1AU (green). The extremely high irradiation flux at 0.02AU causes distinguishable increase in interior temperature by the slower interior cooling, in contrast to the further out interiors that we find to be similar. This increase may be an overestimation by our 1D model that cannot properly simulate day-night effects. Nonetheless, we limit our conclusion on similar interior properties for different distances from the star to be valid down to 0.03AU for our model parameters.

We suggest here, based on the nature of ice-rock interaction in planetary interiors, that the mass fraction of the water envelope is determined by the rock-ice tendency to demix, as a function of the planet pressure (planet mass and gas mass) and temperature (thermal evolution and stellar flux), and not solely by the total ice fraction in the planet.

3.3 Interiors of lower mass planets

We repeated the calculations shown in Fig. 4 for planets with lower initial masses and same composition as the 15 $M_{\oplus}$ model. In Fig. 7 we show the pressure-temperature profiles for initially 5,
10, 15 $M_\oplus$ planets after complete gas loss, at 0.05 AU. As expected, the mass fraction of the demixed ice and rock slightly increases with the decrease of planet mass. Lower mass planets have lower gravity (g) and lower temperature and therefore have larger fraction (mass fraction and absolute fraction) of separated ice. Nevertheless, for all cases ice and rock are completely miscible in more than 99% of the planet’s mass: 99.3% for 5 $M_\oplus$, 99.8% for 10 $M_\oplus$, and 99.9% for 15 $M_\oplus$. Planets with no mass loss (not shown here) have larger fraction of mixed (miscible) ice and rock, as the fraction increases with the envelope mass, due to increase in pressure in the metal-rich region. As ice-rich planets are predicted in the mass range of a few Earth masses and above (e.g., Venturini et al. 2020a), we suggest that ice and rock are miscible in each other for the mass range of ice-rich planets.

4 DISCUSSION

We find that interiors of close-in planets are similar in their properties to interiors of further out planets, down to 0.03 AU for our model parameters. As a result, information from observations of close-in planets may tell us on their further-out twins. For example, measurements of obliquity and Love numbers of close-in planets (e.g., Kellermann et al. 2018) might be useful to represent similar further-out planets. On the other direction, studying our solar system ice-giants can help our understanding of close-in ice-rich exoplanets\(^5\).

5 In this context, in Vazan & Helled (2020) we showed that Uranus models with gradual composition distribution and mixed ice and rock in 1:2 and 2:1 ratios can fit the measurements of Uranus.

The radius valley in the Kepler’s data (Fulton et al. 2017) may be explained by two populations of dry and ice-rich planets (Zeng et al. 2019; Venturini et al. 2020a). Here we suggest that the water/steam envelope of ice-rich planets is limited by the ice-rock interaction. Hence, even if the planet contains large fraction of ice, its water envelope might be less than 1% of the planet’s mass, because most of the ice in the interior is miscible in the rock. Observation of close-in planetary atmospheres in the coming years, by the ARIEL (Tinetti et al. 2018) and JWST (Greene et al. 2016) missions, will be able to shed some light on this topic by detecting the volatile abundance of planetary envelopes. Yet, the estimated mass of the envelope and its water content may not be representative of the bulk volatile content.

We show here that ice and rock tend to mix in planetary interior conditions, both for interiors with gradual composition distribution and for core-envelope structure. This finding emphasises the importance of material interaction in determining the interior structure of planets. In order to properly determine the planetary structure the interactions between the different species should be considered. Material distribution in the interior affect trends in observation data, such as the planetary radius-mass relation, and correlation between planetary mass and envelope metallicity (Thorngren et al. 2016; Welbanks et al. 2019; Chachan et al. 2019). Therefore, we suggest to include the ice-rock interaction we present here when interpreting exoplanet observation data.

4.1 Model caveats

Our findings are based on the knowledge that when ice-rock mixture reaches a supercritical fluid phase the ice and rock are completely miscible in each other. While this behaviour is well examined in lab experiments for pressure of a few GPa, it becomes unclear as...
we get deeper in the interior toward higher pressures. Molecule dissociation (Melosh 2007), and physical properties related to solid state phases at high pressure (e.g., Redmer et al. 2011; Musella et al. 2019, for ice and rock respectively) may affect our results and require further study. Moreover, additional thermo-chemical processes, such as hydrogen mixing with molten rock (Hirschmann et al. 2012; Chachan & Stevenson 2018; Kite et al. 2019), iron (Stevenson 1977), and other volatiles (Hori & Ikoma 2011; Kite et al. 2020) can influence our findings, toward more complicated interior structure.

Another important factor for the long term ice-rock distribution in the interior is their initial distribution. Here we assume accretion of mixed ice and rock. However, if from some reason the ice is being accreted after the rock, then no mixing is expected. The initial distribution is determined by the formation process. In particular, the size and accretion rate of solids and their volatility determine the deposition location and therefore the initial metal distribution. For example, icy pebbles evaporate exterior to rocky pebbles, while planetesimals of mixed ice and rock can get deeper as a mixture. Our simplified model doesn’t intend to model this, but to emphasise the importance and relevance of ice-rock interaction in the formation phase for the long term interior structure. Understanding the ice and rock initial distribution in the interior as an outcome of planet formation requires further study, and is essential in order to get more realistic models of ice-rich interiors.

Our interior evolution model is 1D model. Therefore, 2-3D effects such as orbital eccentricity of close-in Neptune planets (Correia et al. 2020) and day-night side differences are ignored. These simplifications can be justified, as the irradiation effect on the deep interior is found to be small for planets at distance beyond 0.03 AU. We also ignore for simplicity effects of tidal heating (Henning et al. 2009), which may heat the interior further and increase the ice-rock miscibility, and of self-rotation, which under certain conditions limits the mixing of elements (Lock & Stewart 2017).

4.2 Parameters’ uncertainty

We assume ice to rock ratio of 1:1 for formation exterior to the water iceline. However, water ice content exterior to the iceline can vary with stellar metallicity between 50% to 6% (Bitsch & Battistini 2020), and with short-lived radiogenic heating (Lichtenberg et al. 2019). Atmosphere-free objects in the outer solar system have an ice to rock ratio of 1:2 (Bierson & Nimmo 2019). This lower (than previous calculations, including this study. However, an increase in density is not expected to change qualitatively our findings. Opacity of metal-rich envelopes, although crucial for the thermal evolution, can be strongly affected by the metals micro-physics in time (Movshovitz & Podolak 2008) which is somewhat uncertain. Therefore, we tested several other values of higher and lower opacity. As expected, higher opacity slows the cooling and keeps the interior temperatures higher than is shown here, and lower opacity lead to slightly lower temperatures. In all the cases that we explored, the interior structure properties that we find here and the trends for migrated planets are similar.

The initial gas fraction is another uncertain parameter which varies with formation model, mass loss at disk dissipation, giant impacts, and mass loss by stellar flux. Nevertheless, higher or lower gas fraction doesn’t change our conclusions for the ice-rock interaction and the derived interior structure. Increasing the gas fraction of the envelope would increase pressure and temperature in the deep interior, and thus enlarge the mass fraction of metals in which ice and rock are mixed. Decreasing the gas mass fraction is modeled within the different mass loss cases of Fig. 4 and Fig. 7.

5 CONCLUSIONS

(i) Interior structure of an ice-rich planet differs from the simple layered structure of an ice layer on top of a rock layer. Ice and rock are well mixed (miscible) in most of the interior at any age, for a wide range of planetary masses.

(ii) Effect of migration on the deep interior thermal evolution is small. Deep interior structure and temperature of twin planets that migrated to different locations (beyond 0.03AU) are similar, if mass loss is insignificant.

(iii) Ice-rich planets with gradual metal distribution have hotter deep interiors, in comparison to the core-envelope structure. The gradual structure is stable along the thermal evolution.

(iv) Ice-rich planets with significant gas envelopes develop ice-rock demixing only in the outer gas envelope, where water is miscible in the hydrogen.

(v) Mass loss increases the metal enrichment of the envelope. For planets with gradual composition distribution the metal enrichment inwards flattens the composition gradient in time, resulting in large scale convection and fast cooling.

(vi) In the absence of hydrogen-helium envelope the ice and rock demix at the surface (at distances larger than 0.02AU). The rock is then dissolved in the deeper layers where ice and rock remain mixed, and the outer envelope is composed of ice (volatiles) in steam / condensed form.

(vii) The total ice / water content of a planet cannot be inferred from the atmospheric mass and abundance, as most of the ice is stored in the interior, mixed with the rock.

ACKNOWLEDGEMENTS

We thank Dave Stevenson and Morris Podolak for useful comments and discussions. We also thank Edwin Kite, Leslie Rogers and Tim Lichtenberg for discussion during the Exoplanet-3 virtual meeting. RS is partially supported by an ISF grant.

REFERENCES

Alduchov O. A., Eskridge R. E., 1996, Journal of Applied Meteorology, 35, 601
Alibert Y., Mordasini C., Benz W., Winisdoerffer C., 2005, A&A, 434, 343
Bali E., Audéat A., Keppler H., 2013, Nature, 495, 220
Bierson C. J., Nimmo F., 2019, Icarus, 326, 10
Interiors of ice-rich planets

Welbanks L., Madhusudhan N., Allard N. F., Hubeny I., Spiegelman F., Leininger T., 2019, ApJ, 887, L20
Zeng L., et al., 2019, Proceedings of the National Academy of Science, 116, 9723

Figure A1. Schematic picture of ice-rock interaction in a constant pressure as a function of the ice mass fraction ($X_{H_2O}$) and temperature (T). We use the temperature values of the dashed and dotted curves in each pressure to build the pressure temperature wet solidus and critical (miscibility) curves in Fig. 2, respectively.

APPENDIX A: ICE-ROCK INTERACTION CURVES

The ice-rock interaction is usually described in the temperature (T) - ice mass fraction ($X_{H_2O}$) space for a given pressure (Manning 2004). In Fig. A1, we show such a schematic picture of ice-rock interaction at a constant pressure. As the pressure increases, the two-fluid (melt + fluid) field shrinks (see Fig. 1 in Stalder et al. 2000). The dotted orange curve signifies the wet solidus temperature for the given pressure, and the dashed orange curve is our estimate for a temperature of complete miscibility. Below the wet solidus temperature, ice and rock are separated. Above the miscibility temperature, ice and molten rock are completely miscible (mixed). In between these two temperatures, the ice and rock are partially mixed in two phases: molten rock enriched with water, and water-rich fluid enriched with rock.

For each pressure, we take temperature values of these two orange curves, as found in laboratory experiments, to build the pressure temperature solidus and miscibility curves in Fig. 2. We use the experimental data by Kessel et al. (2015) for peridotitic mantle representing composition close to chondritic. We map the stable phases along the pressure and temperature range, and use the second critical point as found by Melekhova et al. (2007). We use extrapolation of Kessel et al. (2015) to low pressures by fit to the data from Schmidt & Poli (1998). It should be noted that the shape of the melt + fluid region (the half ellipse in Fig. A1) is not well constrained by the current experimental data. This is because most of the laboratory experiments are motivated by Earth’s studies, and therefore focus on low ice contents. The results for 50% ice that we use in this work are therefore uncertain, but temperature uncertainties are on the order of tens of Kelvins (see Kessel et al. 2015). More laboratory experiments are required in the high ice content regime, to study ice-rich interiors in more detail.