FORMATION OF HYDRATED SILICATES IN EDGEWORTH-KUIPER BELT OBJECTS. A. B. Makalkin, Institute of Earth Physics, RAS, Moscow, RF (e-mail: makalkin@uipe-ras.scgis.ru); Dorofeeva, V. A. Vernadsky Institute of Geochemisry, (RAS), Moscow, RF (e-mail: dorofeeva@geokhi.ru); V. V. Busarev, Sternberg State Astronomical Institute, Moscow University, RF; (e-mail: busarev@sai.msu.ru).

Introduction: Visible-range absorption bands at 600-750 nm are recently detected on two Edgeworth-Kuiper Belt (EKB) objects [1]. Most probably the spectral features may be attributed to hydrated silicates. In this paper we indicate possible conditions of the phyllosilicate formation in the EKB bodies.

The model and results: The EKB objects orbit the Sun outwards of the Neptune orbit, 30 AU to 50 AU, and are possibly the most primitive large solid bodies in the solar system. According to [2,3], the EKB objects formed in situ though some part of their material could come with projectiles from the formation zones of giant planets, mainly of Neptune and Uranus. Contemporary models of the solar nebula [4,5] yield very low temperatures and pressures of T = 15-30 K and $P = 10^{-9}$-10^{-10} bar at the radial distance (r) of 30-50 AU and the nebula age of about 10^6-10^7 yr, when the EKB bodies of sub-planetary size were formed. At these T-P conditions all volatiles excluding H_2, He and Ne were in the solid state (mostly ices and some organics), and the abundance of rocky (silicate) dust component was lower than that of ices in accordance with the solar ratios of elements [6].

Phyllosilicates may form in EKB bodies during aqueous alteration [7]. A liquid state of water requires a considerable elevation of temperature in the bodies' interior. Plausible mechanisms of heating were decay of radionuclides (short-lived ^{26}Al and long-lived ^{40}K, ^{235}U, ^{238}U and ^{232}Th) in the bodies and mutual collisions between the bodies. In this paper we consider the former mechanism only. As shown in model calculations [8], the long-lived radioisotopes were insufficient for total melting of ice fraction in the icy satellites of giant planets of radii up to 800 km, although partial melting was possible.

Was the concentration of captured ^{26}Al (half-life 7.2 x 10^5 yr) sufficient for melting of water ice in the EKB objects? If the time of EKB bodies' formation was substantially larger than the half-life of ^{26}Al, then independently of the isotope concentration it couldn't heat the EKB bodies with rather high efficiency.

The formation time of hundreds-km-sized EKB bodies was from about one million year [9] to several tens of million years [10]. Taking into account the model of cometary body formation by [9], accretion of bodies up to 100 km in radius at the EKB distances 35-50 AU within $\approx (1–1.5)\times10^6$ yr seems to be possible, though this time is near the lower limit of accretion timescales. In this consideration we suppose that formation of planetesimals at r of the EKB could begin several 10^5 yr after the collapse of the protosolar cloud.

If the accretion of bodies of radius R = 100 km had completed no later than a few ^{26}Al half-life times, the decay of this isotope provides enough heat to melt the water ice in the interiors of these bodies. To check this conclusion we adopt the mass fraction of silicates (silicates /(ices + dust)) in EKB parent bodies of 0.30 [11]. The rock component with chondritic (solar) abundances of refractory elements contains 1.3 wt. % of aluminum. We also adopt the ^{26}Al/^{27}Al ratio of 1×10^{-5} which is obtained from the "canonical" initial ^{26}Al/^{27}Al ratio of 5×10^{-5} and accretion time of a EKB body 1.6 Myr (after CAIs). This time possibly but not necessarily coincides with the age of the solar nebula (from the collapse stage).

The above figures, giving the ^{26}Al abundance, should be added with the decay energy of ^{26}Al = 3 MeV per atom and its decay constant $\lambda = 9.63\times10^{-7}$ year^{-1} to yield the heat production rate Q = 0.4 J kg^{-1} yr^{-1}. The time τ_m required to heat a large EKB body to the water-ice melting point and to melt the ice in its interiors can be estimated from the equation

$$\int_0^{\tau} Q\exp(-\lambda t)dt = \int_{T_0}^{T_m} c_p dT + L_f m_w\,, \qquad (1)$$

where T_0 = 30 K is the adopted value for the initial temperature of the body, T_m = 273 K is the melting temperature of water ice (a good approximation to at P< 25 MPa, characteristic for interiors of a EKB body of radius $R \le 300$ km), $L_f = 3.34\times10^5$ J kg^{-1} is the latent heat of fusion for H_2O, m_w = 0.38 is adopted for the H_2O mass fraction, c_p is the thermal capacity at constant pressure per unit mass for the body's material. With some overestimation of c_p at temperatures from 30 to 150 K we can take the temperature dependence of specific heat for all main components similar to that for water ice: c_{pw}=7.67 T J kg^{-1} K^{-1} [12]. In this approximation the values of thermal capacities (all in J kg^{-1}K^{-1}) is: $c_{p\ \mathrm{r}}$=3.1 T for rocks (mainly silicates), $c_{p\ \mathrm{CHON}}$ = 5.7 T for refractory organics, and $c_{p\ \mathrm{vol}}$ =10 T for volatile organics and gases (the approximation for gases is most crude, but this has little effect due to their low content). We use also mass fraction of CHON m_{CHON} = 0.22 and combined mass fraction of volatile organics and gases $m_{\mathrm{vol+g}}$ = 0.10. With these values we obtain the thermal capacity for the mixture $c_p \approx c_{p0}\,T$, where c_{p0} = 6.1 J kg^{-1}K^{-2}. After substitution of this value in Equation (1) and integration we have the majoring estimation for the time τ_m :

$$\tau_m = -\lambda^{-1}\ln\{1-\lambda[c_{p0}(T_i^2 - T_0^2)/2 + L_f m_w]/Q\} \approx 1.9\times10^6\,\mathrm{yr}\,. \qquad (2)$$

Thus the water ice in the bodies can be melted in less than 2 million years after the body formation and, hence, at the age of the solar nebula of 3.5 Myr. During this time only a surface layer of thickness $\Delta R \sim 10$ km could be frozen, as follows from the simple estimation:

$$\Delta R \sim \sqrt{\kappa\tau}\,, \qquad (3)$$

where κ is the thermal diffusivity related to the thermal conductivity k as $\kappa = k/(\rho c_p)$. The temperature dependence of κ for water ice I is $\kappa = \kappa_0 T^{-2}$, where $\kappa_0 \approx 9.1\times10^{-2}\,\mathrm{m^2K^2s^{-1}}$ [13]. However, the porosity of ices $p \approx 0.5$ decreases the thermal conductivity 5 to 50 times [14]. The porosity is the maximal at the surface and reduces to the low values at the bottom of the layer. Thus the reasonable estimate for the thermal diffusivity of the layer is $\kappa \sim 10^{-6}\,\mathrm{m}^2 s^{-1}$.

The consequences of water ice heating are much more important. First, huge amount of water ice evaporated at low pressures in the porous medium should recondense in the upper layers of the bodies, substantially reducing their porosity. As a result of insulation of the interiors from the outer space, the pressure below the upper layer of thickness ΔR would become higher than 1 bar and melting of water ice should occur when heated to $T > T_m \approx 270$ K. Probable admixture of volatile organics might slightly decrease this temperature. Thus, as follows from Equations (1) and (2), internal water ocean in the young EKB bodies forms at their age ≈ 1.9 Myr, that is after ≈ 3.5 Myr after CAIs and solar nebula formation.

Consider the evolution of the internal water ocean in a young EKB body of r = 100–300 km. The thermal convection in the ocean should be vigorous, if the Rayleigh number Ra is much higher than its critical value $\mathrm{Ra_{cr}} \sim 10^3$. We can estimate the value $\mathrm{Ra} = \alpha g d^3 \Delta T/(\kappa\nu)$, where $\alpha \approx 10^{-4}\,\mathrm{K^{-1}}$ is the volumetric thermal expansion coefficient of the mixture, dominated by liquid water, g is the gravitational acceleration ($g \sim 4\rho G R$), $d \sim$ 0.8–0.9 R is the convective layer thickness, ΔT is the temperature difference across the layer, $\kappa \sim 10^{-7}\,\mathrm{m^2\,s^{-1}}$ and $\nu \sim 10^{-5}\,\mathrm{m^2\,s^{-1}}$ are the thermal diffusivity and kinematic viscosity of the water–solids mixture. At d = 70 km and the very low value of ΔT = 1 K we nevertheless obtain a very high value Ra ≈ 10^{21}. The Nusselt number (Nu), which is the ratio of the total heat flow (including convective one) to the conductive flow is related to Ra by [15] $\mathrm{Nu} \approx 0.2\mathrm{Ra}^{1/3}$. With these data we can estimate the time of heat transport through the convective water ocean τ_c by relation (3) where ΔR and is substituted for $d \sim 0.8\ R$ and the molecular thermal diffusivity κ is substituted for the effective thermal diffusivity κ_e which accounts for convection, with $\kappa_e = \kappa \cdot \mathrm{Nu}$. At the above parameters we obtain $\tau_c \sim 10^3$ years. The time is very short relative to the thermal evolution time scale of the bodies $\tau_m \sim 10^6$ years. The latter is also the time scale for heat transport through the outer body's shell (thermal boundary layer) of thickness ΔR. Owing to the rapid radial heat transport through the water ocean its T is stabilized near the temperature of maximum water density ≈ 277 K (the adiabatic compression for hundreds-km-sized bodies is negligible) and probably never exceeds 280 K. After a lapse of time a continuing decrease of radiogenic heat production yields the freezing of the internal ocean beginning (as in a usual terrestrial ocean) from the upper layer.

The lifetime of the water ocean (till the beginning of its freezing) in the early EKB body of radius R can be estimated by comparing the heat flow F_1 underneath the solid shell (thermal boundary layer) of thickness ΔR and the heat flow F_2 in the shell. The flow F_1 is generated in the interiors being heated by the ^{26}Al decay and quickly transferred to the lithosphere. Thus we can write $F_1 \approx \tfrac{1}{3}\,\overline{\rho}\,(R-\Delta R) Q \exp(-\lambda t)$ (4)

on the assumption of ^{26}Al homogeneous distribution, where $\overline{\rho} = 1.4\times10^3\,\mathrm{kg\ m^{-3}}$ is the mean density of the body (calculated at the above fractions of components). Flow F_2 can be written as $F_2 = k\,\Delta T/\Delta R$, (5)

where k is the thermal conductivity of shell, ΔT = 273 –30 ≈ 240 K. We assume $k = 2\,\mathrm{W\,m^{-1}K^{-1}}$, taking into account the empirical relation for crystalline ice $k(T) = 567/T\ \mathrm{W\,m^{-1}K^{-1}}$ and compensating effect of increasing porosity from the base to the surface of the shell [16]. The freezing of the water ocean begins when the incoming flow from interiors F_1 becomes lower than the flow F_2 coming from the shell. By equating two flows from (4) and (5) we obtain the estimate of the lifetime of the ocean of liquid water as $\tau_l \approx$ 0.6–1.8 Myr for the bodies of radius R = 100–300 km respectively.

This lifetime is quite sufficient for silicates to form phyllosilicates by reaction with water.

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