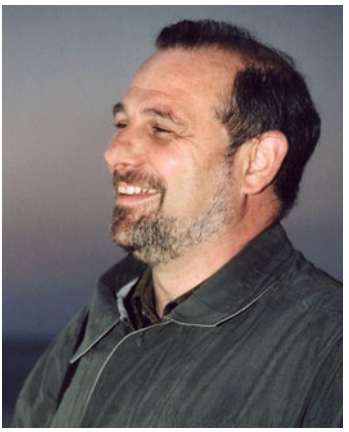
Biographies of the Authors



Lev V. Eppelbaum received M.Sc. from the Azerbaijan Oil Academy in 1982, and Ph.D. from the Geophysical Institute of Georgia in 1989. In 1982–1990 he worked as geophysicist, Researcher and Senior Researcher at the Institute of Geophysics in Baku (Azerbaijan). In 1991–1993 Eppelbaum completed postdoctoral studies in the Department of Geophysics and Planetary Sciences at Tel Aviv University; at present he occupies a position of Associate Professor at the same Department. He author of more than 300 publications including 5 books, 115 articles and 50 proceedings. His scientific interests cover potential and quasi-potential geophysical field analysis, integrated interpretation of geophysical fields, geodynamics and tectonics. Eppelbaum's research interests in geothermics include examination of temperature regime of boreholes, thermal interactions at a depth, near-surface temperature measurements, study of climate of the past and nonlinear analysis.

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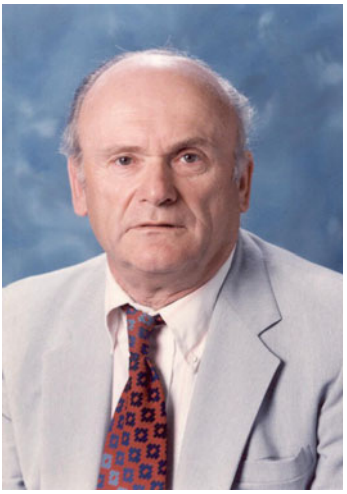

Izzy M. Kutasov holds Ph.D. in Physics from O. Schmidt Earth Physics Institute in Moscow. He was a Senior Lecturer at the School of Petroleum Engineering, University of New South Wales, Sydney, and a graduate faculty member at the Department of Petroleum Engineering and Geosciences, Louisiana Tech University. He worked for Shell Development Co., Houston, as a Senior Research Physicist. Now Dr. I.M. Kutasov is a consultant with BYG Consulting Co., Boston, USA. His research interests include the temperature regime of deep wells, transient pressure/flow analysis, well drilling in permafrost areas as well as deriving climate of the past from subsurface temperature measurements. He is author of more than 250 publications including four books and more than 100 articles. Dr. Kutasov presented his scientific achievements at more than 60 International Conferences. His developments in the thermal data analysis in oil and gas wells received a wide recognition in the world scientific and engineering community.

Arkady N. Pilchin has been involved in geoscientific research since he received his M.Sc. in Physics in 1973. He participated in a number of geophysical and geological research endeavors including geothermal investigations in the Caucasus, Caspian Sea and Middle Asia, developing physical-geological models of the regions, earthquake prognosis, methods of prognosis for overpressure, analysis of the thermodynamic conditions of the Earth's crust, and those of oil and gas fields. He authored a number of geothermal methods including determining heat absorbing and heat releasing strata within a sedimentary layer. After leaving the former USSR in 1990, he worked on geoscientific problems of different regions of the world such as the Eastern Mediterranean, Alps, Baltic Shield and Canada. He is the author of over 150 publications, including 2 books, 3 book chapters, several geothermal maps and 15 patents. His current scientific interests are related to the formation, evolution and thermodynamic conditions of: the magma-ocean, early Earth atmosphere, earliest lithosphere, water-ocean, ore deposits, and different generations of sulfur.

Appendix A
Computing Water Flow Geodynamics in Stratified Liquids (An Example from the Dead Sea Basin)

The Dead Sea (DS), with a total area of 1,049 km^2 is located on the border between Israel and Jordan. It is the lowest place on Earth (in 2013 it was about 427 m below the Mediterranean Sea level) with an arid climate (in the summer the temperature can reach 45 °C). Thermal and pressure parameters in this area are characterized by a strong inhomogeneity (Eppelbaum et al. 1996). The density of water in the vertical direction varies from 1,150 kg/m^3 at the DS surface up to 1,430 kg/m^3 at a depth of 310–330 m. This density gradient exceeds the normal water density gradient in the Earth's oceans and seas by more than 10^2. Thus, the DS Basin may be considered a distinctive native polygon of vertically stratified liquid (the role of rock stratification in the magma-ocean was shown in Sect. 6.1). The most significant source of this are temperature and density gradients based on the constant inflow of a mineral substance from great depths (apparently, from the upper mantle). To analyze the water flows geodynamics in the DS Basin, solutions for the motion of stratified liquid with different initial boundary values can be used. Given the very high current rate of change of the physical, chemical and geological features of this unique basin, the development of a geodynamic predictable model of the area is extremely important. As was shown earlier, density practically always has a reliable correlation with temperature.

A.1 Nature of the Problem

At present, the distinctive DS water is intensively utilized for extracting useful components contained in the liquid. This industrial activity, combined with the extremely arid climate has caused a significant drop in the sea level and the development of hundreds of sinkholes on the DS coast (Eppelbaum et al. 2008). For the DS, the conventional mathematical model describing sea behavior consists of a set of algebraic and a system of differential equations (e.g., Asmar and Ergenzinger 2002). The state of the sea is found by solving a system of ordinary differential equations formed by writing a system of mass balance equations for the water and eight major chemicals in the sea. The law of conservation is accounted for implicitly through evaporation calculations. However, this approach is unable to calculate all the peculiarities of the unique DS Basin. Interesting results were

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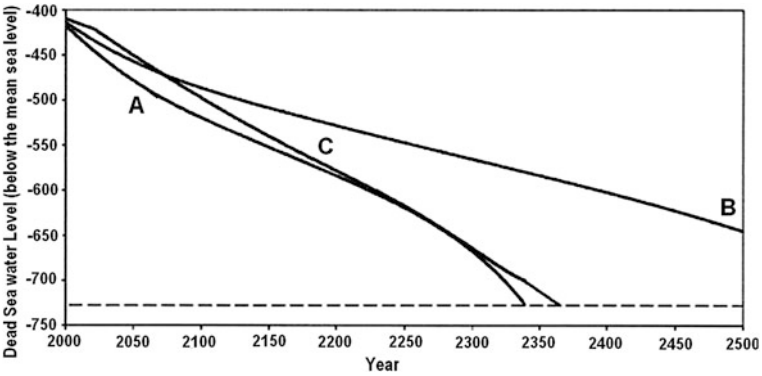

Fig. A.1 Predicted water levels in the Dead Sea up to the year 2500. **a** Scenario 1 shows a continuation of the current conditions. **b** Scenario 2 shows cessation of industrial extraction activity. **c** Scenario 3 shows simplified climate change (after Asmar and Erdenzinger 2004)

presented by Bubnov and Linden (1998) for a homogeneously rotating linear density-stratified fluid. However, this approach demands significant evaluation.

We suggested applying a more precise physical-mathematical model which has implications for forecasting environmental scenarios for the DS (three variants are presented in Fig. A.1) and in estimating the possible ecological impact of excavation of a channel between the Red and Dead Seas (the differences in chemical composition of water from these seas are shown in Table A.1).

A.2 Development of Mathematical-Physical Model

Let's consider the motion of stratified, rotating and compressible liquid (*SRCL*) in the Cartesian coordinates (x_1, x_2, x_3) which is rotating together with the *SRCL*. Let's assume that *SRCL* is rotating around axis Ox_3 and Coriolis vector $\mathbf{f} = (0, 0, f)$, where f is the double angle velocity of rotation. *SRCL* is stratified along the axis Ox_3; i.e., its density in undisturbed conditions depends on $x_3, \rho_0 = \rho_0(x_3)$. We assume below that $\rho_0(x_3) = Ae^{-2\beta x}, A > 0, \beta > 0$. Small *SRCL* motions under the influence of gravity acceleration without any external forces may be described as (Gabov and Sveshnikov 1986):

$$\begin{cases} \rho_0(x_3)\frac{\partial \mathbf{v}}{\partial t} + \rho_0(x_3)[\mathbf{f}, \mathbf{v}] + \nabla p + \mathbf{e}_3\rho_1 g = 0, \\ \frac{\partial \rho_1}{\partial t} + (\mathbf{e}_3 \cdot \mathbf{v})\rho_0'(x_3) + \rho_0(x_3)\mathrm{div}\ \mathbf{v} = 0, \\ \frac{\partial}{\partial t}\rho_1 = \frac{1}{c^2}\frac{\partial}{\partial t}p + \rho_0(x_3)\omega_0^2(x_3)\frac{(\mathbf{e}_3 \cdot \mathbf{v})}{g}, \end{cases} \tag{A.1}$$

where g is the gravity acceleration, $\mathbf{v} = (v_1, v_2, v_3)$ is the vector of *SRCL* particle motion, ρ_1 is the change in *SRCL* density caused by its motion, c is the acoustic velocity, p is the dynamic pressure and $\mathbf{e}_3$ is the ort of axis Ox_3 and $[\mathbf{f}, \mathbf{v}]$ denotes the vector product between two vectors. Value $\omega_0^2(x_3) \geq 0$ is the quadrate of

Table A.1 Some typical chemical analyses of the Dead Sea (*DS*) and the Red Sea (*RS*) (after Abu-Khader 2005; Ravizky and Nadav 2007, with supplements)

Element	*DS* concent-rations (mg/l)	*RS* concent-rations (mg/l)	Ratio, *DS*/*RS*
Chloride	224,000	23,000	9.74
Magnesium	44,000	1,500	29.33
Sodium	40,100	12,500	3.21
Calcium	17,650	500	35.3
Potassium	7,650	500	15.3
Bromide	5,300	70	75.71

Vyasyalay-Brent frequency. The condition $\omega_0^2(x_3) \geq 0$ indicates an absence of convective motion and stability of the density distribution $\rho_0(x_3)$ in the *SRCL*. It is known that physical parameters such as liquid density, acoustic velocity and pressure depend on temperature. Thus, the temperature is present in the formulas in an implicit form.

After several cumbersome transformations, Eq. (A.1) was reduced to the following type:

$$\frac{\partial^2}{\partial t^2}\left[\frac{\partial^2 u}{\partial t^2} - c^2\Delta_3 u + \left(f^2 + \beta^2 c^2\right)u\right] - \left[\omega_0^2 c^2 \Delta_2 u + f^2 c^2 \frac{\partial^2 u}{\partial x_3^2} - \beta^2 c^2 f^2 u\right] = 0, \tag{A.2}$$

where u is a scalar function on which unknown functions $\mathbf{v}$, ρ_1 and p depend.

A generalization of Eq. (A.2) was investigated in Yakubov (1989). Let's consider the following problem

$$\begin{cases} L(t, D_t, D_x)u(t,x) = \dfrac{\partial}{\partial t^2}\left[\dfrac{\partial^2 u(t,x)}{\partial t^2} - \displaystyle\sum_{i,j=1}^{3} \frac{\partial}{\partial x_i}\left(A_{ij}(t,x)\frac{\partial u(t,x)}{\partial x_j}\right)\right. \\ \qquad \left. + \displaystyle\sum_{i=1}^{3} B_i(t,x)\frac{\partial u(t,x)}{\partial x_i} + C(t,x)u(t,x)\right] + \displaystyle\sum_{i,j=1}^{3}\frac{\partial}{\partial x_i}\left(a_{ij}(t,x)\frac{\partial u(t,x)}{\partial x_j}\right) \\ \qquad + \displaystyle\sum_{i=1}^{3} b_i(t,x)\frac{\partial u(t,x)}{\partial x_i} + d(t,x)u(t,x) = f(t,x),\ (t,x) \in [0,T] \times \Omega, \end{cases} \tag{A.3}$$

where t is the time, A_{ij}, B_i, C, a_{ij}, b_i, d are sufficiently smooth functions (for an application, see Eq. (A.2), and depend on c, f, β and ω_0).

$$u(t,x)|_{x\in\partial\Omega} = 0, \quad t \in [0,T], \tag{A.4}$$

$$\frac{\partial^k u(t,x)}{\partial t^k}|_{t=0} = u_k(x), \quad k = 0,\ldots,3,\ x \in \Omega. \tag{A.5}$$

Equations (A.4) and (A.5) present the Dirichlet boundary condition and the Cauchy initial conditions, respectively. Thus, Eqs. (A.3)–(A.5) can be seen as a foundation for the development of a geodynamical model of *SRCL* in the DS Basin (Yakubov and Eppelbaum 2005).

A.3 Application of the Methodology

In the DS Basin (at the sea surface and at various depths) the following physical parameters were observed: temperature, gravity acceleration, water density, pressure, and acoustic velocity. Modern processing schemes can extend analytical continuations (downward, upward and sideward) of the physical observations to areas where we do not have these measurements. Thus, we can create a solvable system of differential equations and utilize this to create a precise physical-mathematical-geological model of the DS Basin. This model may be used not only for predictions, but also for the projected channel between the Red and Dead Seas, as well as to calculate the free electrochemical energy released from the unique DS Basin.

Appendix B
Water Production Using the Air-Cooling Method

The coastal regions in many regions of the world are characterized by their high moisture content—more than 15.0 g of water vapor per m^3 of air. This appendix is based on a phenomenon observed in the atmospheric boundary layer known as dew formation. At night, the boundary layer of air cools down considerably, which generally causes moisture saturation in the boundary layer and dew formation on the ground and plants. The ancient Greeks used this for water production (Zibold 1905). In more modern times, many investigations have been carried out to evaluate the quantitative potential of dew (Broza 1979; Kogan and Trahtman 2003; Nilsson 1996; Schemenauer and Cereceda 1992). However, all of these investigations have indicated that its potential is low.

To obtain fresh water by cooling humid air, we propose using a horizontal pipe. At the entrance to the pipe we will install a ventilator, which will introduce external humid air into the pipe. The outside wall of the pipe will be supplied with a special cooling system. A simplified scheme of this pipe is presented in Fig. B.1.

Let's consider the climatic conditions of Israel for illustrative purposes. According to Goldreich (2003) the average daily temperature, pressure and relative humidity in the coastal regions are $\bar{T} = 25\,^{\circ}\mathrm{C}$, $\bar{P} = 1{,}004\,\mathrm{mb}$ and $\bar{R}_H = 72\,\%$, respectively. Preliminary calculations using a relatively simplified thermodynamics method based on these average values show that if the air is cooled to 20 °C, 5.4 g of water can be obtained from 1.0 m^3 of air. If the radius of the pipe (cooling equipment) is 10 m and the velocity of the air inflow is 3.0 m/s, the air mass entering the pipe equals about 940 m^3/s. Therefore, the maximum amount of water obtainable from this system is 5.0 l/s; i.e., about 430 m^3 per day and 78,000 m^3 per summer period. These rough estimates indicate that the idea of producing considerable amounts of drinking water by air-cooling in the coastal regions of Israel is practical.

Let's make some preliminary calculations for conditions typical to Israel. Based on these data, one can compute the daily average specific saturating humidity by applying Teten's (1930) empirical formula:

$$\bar{Q}_S(\bar{P},\bar{T}) = \frac{3.8}{\bar{p}}\bar{E}\cdot\exp\left\{17.27\frac{\bar{T}-273.16}{\bar{T}-35.86}\right\}, \tag{B.1a}$$

where $\bar{T}$ is the absolute average temperature.

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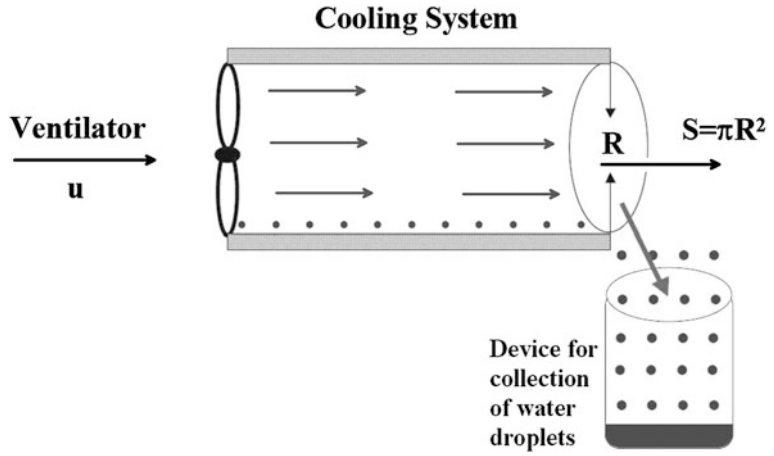

Fig. B.1 Schematic diagram of pipe for water production (after Tzivion and Eppelbaum 2004)

$$\bar{Q}_S(\bar{P} = 1{,}004\ \text{mb},\ \bar{T} = 298.16\ ^\circ\text{C}) = 19.63 \times 10^{-3}\ \text{g/g}. \tag{B.1b}$$

Thus, under these conditions, 1 kg of air contains 19.63 g of water vapor. Hence, we can calculate the daily average specific humidity in summer:

$$\bar{q}_v(\bar{P}, \bar{T}) = \bar{Q}_s(\bar{P}, \bar{T}) \cdot RH\,\% = 14.13\ \text{g/kg}.$$

If the air is cooled down to 20 °C, the specific moisture content will remain unchanged, and only $\bar{Q}_S$ will change, and its new value will be

$$\bar{Q}_S(\bar{p} = 1{,}004\ \text{mb},\ \bar{T} = 278.16) = 5.4\ \text{g/kg}. \tag{B.2}$$

Thus, we obtain the excessive water content—over saturation ΔS:

$$\Delta\bar{S} = \bar{q}_v - \bar{Q}_S = (14.13 - 5.4) = 8.73\ \text{g/kg}. \tag{B.3}$$

The average air density $\bar{\rho}$ for the above $\bar{p}$ and $\bar{T}$ values is roughly 1.2 kg/m^3. Hence, we obtain

$$\Delta\bar{S} = 10.5\ \text{g/m}^3.$$

Thus, 1 m^3 of air contains excess water of about 10.5 g.

Now we estimate the maximum amount of water that can be obtained from this excess water in humid air. During vapor condensation, the following values will change: $\bar{T}$, $\bar{q}_v$, $\bar{Q}_S(\bar{p}, \bar{T})$, $\Delta\bar{S}$.

We denote the new values of these functions as T^*, P^*, Q_S^* and ΔS^*. For simplification we assume that due to condensation the pressure remains unchanged and that after condensation the over-saturation becomes zero. Under such conditions, we calculate the maximum possible amount of condensed water

$\delta M_{cond}(\mathrm{g/kg})$. If δM_{cond} of the water is condensed, after condensation T^*, P^*, q_v^*, Q_S^* and ΔS^* take on new values:

$$P^* = \bar{P},\ T^* = \bar{T} + T',\quad q_v^* = \bar{q}_v - \delta M_{cond},\quad Q_S^* = Q_S^*(\bar{P}, T^*). \tag{B.4}$$

Here T' is the temperature disturbance due to condensation:

$$T' = \frac{L}{C_P}\delta M_{cond}, \tag{B.5}$$

where L is the specific heat condensation, and C_p is the specific heat capacity of the air at a constant pressure.

For purposes of simplification, we introduce the following notations:

$$A = 3.8/P,\quad b = 17.27,\quad T_0 = 273.16,\quad C = 35.86.$$

According to (B.2), we obtain:

$$Q_S^*(\bar{P}, T^*) = A \cdot \exp\left(b \cdot \frac{\bar{T} - T_0 + T'}{\bar{T} - C + T'}\right) \tag{B.6}$$

Given that $T' \ll \bar{T}$, we can rewrite (B.6) as follows:

$$Q_S^*(\bar{P}, T^*) = A \cdot \exp\left[b\frac{\bar{T} - T_0}{\bar{T} - C} + b\frac{T_0 - C}{(\bar{T} - C)^2}\right]T'. \tag{B.7}$$

Using (B.2) and (B.5), we can rewrite Q_S^* as follows:

$$Q_S^*(\bar{P}, T^*) = \bar{Q}_S(\bar{P}, \bar{T}) \cdot \exp\left[\frac{b(T_0 - C)}{(\bar{T} - C)^2}\frac{L}{C_P}\delta M_{cond}\right]. \tag{B.8}$$

Since the exponent is below unity, the final approximate value of Q_S^* is (Tzivion and Eppelbaum 2006):

$$Q_S^*(\bar{P}, T^*) \approx \bar{Q}_S(\bar{P}, \bar{T}) \cdot \exp\left[1 + \left(\frac{L}{C_P}b\right)\frac{T_0 - C}{(\bar{T} - C)^2}\delta M_{cond}\right]. \tag{B.9}$$

Differentiating (B.2) with respect to $\bar{T}$, we obtain:

$$\frac{d\bar{Q}_S(\bar{P}, \bar{T})}{dT} = \frac{b(T_0 - C)}{(\bar{T} - C)^2}\bar{Q}_S(\bar{P}, \bar{T}). \tag{B.10}$$

Substituting (B.10) into (B.9), we obtain:

$$Q_S^*(\bar{P}, T^*) = \bar{Q}_S(\bar{P}, \bar{T}) + \frac{L}{C_P} \cdot \delta M_{cond}\frac{d\bar{Q}_S(\bar{P}, \bar{T})}{d\bar{T}}. \tag{B.11}$$

Using (B.11) and taking into account that $\Delta S^* = q_v^* - Q_S^* = 0$ and $\bar{\Delta}\bar{S} = \bar{q}_v - \bar{Q}_S$, we obtain

$$\Delta\bar{S} - \left[1 + \frac{L}{C_P} \cdot \frac{d\bar{Q}_S(\bar{P},\bar{T})}{d\bar{T}}\right]\delta M_{cond} = 0.$$

Hence we finally derive a formula for determining the maximum possible amount of condensed water:

$$\delta M_{cond} = \frac{\Delta\bar{S}}{A(\bar{P},\bar{T})}, \tag{B.12}$$

where

$$A(\bar{P},\bar{T}) = \left[1 + \frac{L}{C_P} \cdot \frac{d\bar{Q}_S(\bar{P},\bar{T})}{d\bar{T}}\right]. \tag{B.13}$$

In the case of $\bar{Q}_S(\bar{P},\bar{T}) = 5.4\,\text{g/kg}$, $b = 17.27$, $\bar{T} = 278.16\,^{\circ}\text{C}$, $L = 600$ cal/g, $C_p = 0.24$ cal/g $\cdot$ grad we obtain:

$$A(\bar{p},\bar{T}) = 1.942. \tag{B.14}$$

Since $\Delta\bar{S} = 8.73\,\text{g/kg}$, we get that the maximum possible amount of condensed water will be about 4.5 g per 1 kg of air, i.e.

$$\delta M_{cond} \approx 4.5\,\text{g/kg} \quad \text{or} \quad \delta M_{cond} \approx 5.4\,\text{g/m}^3.$$

At $\delta M_{cond} \approx 4.5\,\text{g/kg}$ based on Eqs. (B.2)–(B.5) and (B.9) we get that new values of T^*, q_v^*, Q_S^* and $RH\,\%^*$ are equal to (Tzion and Eppelbaum 2006):

$$T^* = 289.4^{\circ}, \quad q_v{}^* = 9.63\,\text{g/kg}, \quad Q_S{}^* = 9.63\,\text{g/kg}, \quad RH\,\%^* = 100\,\%. \tag{B.15}$$

If we place a horizontal pipe with internal cooling in the vicinity of the sea coast perpendicularly to the latter, the amount of air entering the pipe will equal

$$M_{\text{air}} = U \cdot \rho_{\text{air}} \cdot S, \tag{B.16}$$

where M_{air} is the mass of the air entering the pipe with an area $S = \pi R_S$, U is the air velocity at the entrance to the pipe (m/s), ρ_{air} is the air density (kg/m^3).

If we assume that $R_S = 10$, $U = 3$ m/s, and $\rho = 1.2$ kg/m^3, then, according to (B.16), the mass of air entering the pipe in 1 s equals

$$M_{\text{air}} = 3.0(\text{m/s}) \cdot 1.2\left(\text{kg/m}^3\right) \cdot 3.14 \cdot 100\,\text{m}^2,$$
$$M_{\text{air}} \approx 1{,}130\,\text{kg/s} = 942\,\text{m}^3/\text{s},$$
$$M_{\text{air}} \approx 940\,\text{m}^3/\text{s}.$$

Since it is possible to obtain 5.4 g of water from 1 m^3 of air, the maximum amount of water obtainable from this system roughly amounts to 5,000 g of water in 1 s; i.e, about 18,000 kg of water per hour, or about 430 m^3 per day and 78,000 m^3 per summer period.

These rough estimates show that producing a considerable amount of water by air cooling is feasible. The obtained water can be used for drinking and other purposes. It should be noted that this methodology could be applied in Israel and in many regions of world at the sea coasts with high humidity.

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