SMALL-SCALE TURBULENCE AND VERTICAL MIXING

The vertical mixing of water masses by small-scale turbulence is one of the most important hydrophysical processes of the Black Sea. It regulates the physical end chemical conditions in the water column, e.g. the depth of oxygen penetration and the nutrient flows, thus controlling biological productivity.

The vertical mixing of deep and surface water and the rates of transfer of its physical and chemical features across a vertical density gradient is driven by several processes: thermal convection (in upper and bottom layers), small-scale turbulent diffusion, double diffusion (its thermal driven, so called, ”diffusive” regime), lateral thermohaline intrusion of alongslope descent dense Marmara Sea waters, and vertical components of currents.

The rate of turbulent exchange is controlled by the vertical density gradient ($d\rho/dz$). The greater this gradient, the more energy is needed for turbulent transfer and the slower it proceeds. Therefore the main parameter characterizing the potential speed of turbulent mixing, the coefficient of turbulent diffusion K_z, is an inverse function of the water stability which is proportional to the square of the Brunt-Väisälä (buoyancy) frequency N, calculated as $N^2 = -g/\rho(d\rho/dz)$, where g – gravitational acceleration. In the upper 200 m of the Black Sea N^2 ranges from $0.5\ 10^{-4}$ to $>6.0\ 10^{-3}\ s^{-2}$, then decreases to $\sim 10^{-6}\ s^{-2}$ at 1000 m and to zero beneath 1675-1780 m in the abyssal bottom layer of 300-400 m thickness (Fig 1).

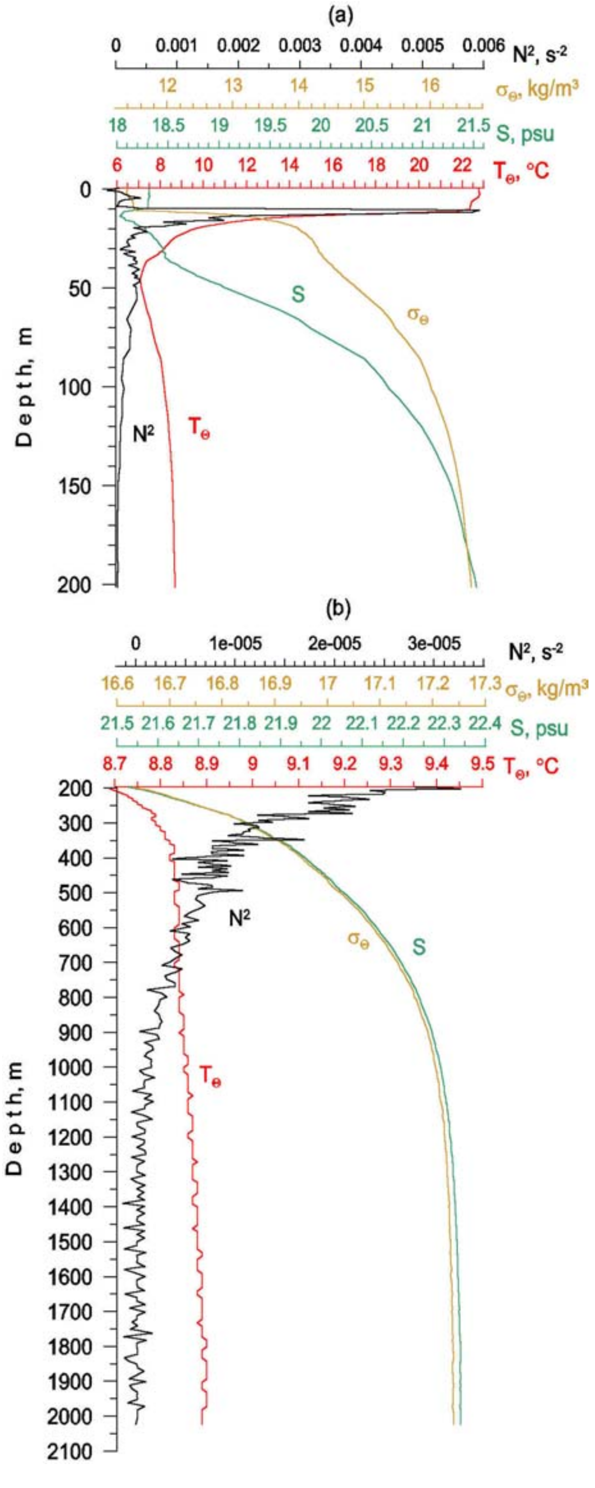

Fig. 1 – Vertical profiles of potential temperature (T_Θ), salinity (S), potential specific dencity (σ_Θ) and square of the Brunt-Väisälä frequency (N^2) in central area of the Black Sea in summer

A high dρ/dz in the Black Sea main pycnocline (at 50-150 m depth) drastically slows down vertical turbulent exchange and mixing (Fig 1). During the warm season (from April to October), an extremely steep seasonal thermocline (at 10-30 m depth), creates another obstacle to vertical mixing.

Thermal convection is a key mixing process in the Black Sea upper layer during autumn-winter period, when its water density increases due to cooling. The depth of autumn-winter convection vary in different parts of the Black Sea and depends on the year-to-year severity of winter. It proceeds down to 30-100 m.

Another place where thermal convection is the key mixing mechanism is the homogeneous bottom layer. The thermal convection is supported here by geothermal heat flow at $3\ 10^{-2}$ to $5\ 10^{-2}$ Wm^{-2}. The mantle is very thin under the Black Sea. It transmits a rather high Earth heat flow to the bottom layer of Black Sea abyssal. This geothermal heat flow is sufficient to support upward convection flow at the level of $0.3\text{-}0.7\ 10^{-6}$ m s^{-1}. It elevates near-bottom water from 2,100 m to 1,500 m depth in only 25-50 years.

This pool of water has properties, which very uniform in space and time: potential temperature, Θ=8.906±0.001°C, salinity, S=22.333±0.002 psu, specific potential density, σ_Θ=17.223±0.002 kg m^{-3}. The existense of a basin-wide layer with uniform depth of upper boundary and thermohaline properties syggest efficient homogenenization by convectively driven three-dimensional motions.

The top of this layer can identified by relatively sharp density step (thickness $\sim 10^1$ m) with local maximum N^2, which increased from background values of less than $1\text{-}2\ 10^{-5}\ s^{-2}$ to maximum value of about $8\ 10^{-5}\ s^{-2}$.

There have been a number of theoretical and experimental studies describing the processes that occur when a linearly stratified salt solution is heated from below (see Turner, 1973). Convective stirring first produces a layer at the bottom that is well mixed. This layer does not continue to grow indefinitely. Because heat diffuses faster than salt, a second layer eventually forms above the first followed by a third layer and so on. In time, many such layers should form with sharp interfaces separating turbulent convecting regions. Eventually the bottom layers coalesce to form a thicker bottom layer. Natural environments where this type of double diffusion are thought to occur in the Red Sea brines in that the 100 m thick well-mixed bottom convective layers have single sharp interfaces at their top. The Red Sea differs from the Black Sea in that it has two such convective layers.

In the gradient zones of seasonal thermocline and the main pycnocline a key role in mixing is clue to **small-scale turbulence**. The share of vertical water movements by currents in total vertical exchange was estimated between only 2-3% in halistatic areas and up to 10-12% in the core of the Black Sea Main Rim current (MRC).

In practice for the evaluation of vertical mixing the coefficient of efficient vertical mixing K_z^{eff} is used. This coefficient accounts for turbulent transfer plus the transfer by vertical upward or downward water flows. This coefficient can be calculated from the density gradient and current velocity. Its estimation in the Black Sea water based on the salt balance between the Black Sea and the Sea of Marmara. The salt transport by upper and lower Bosporus currents is balanced around 2 x 10^5 kg s^{-1}. The salt brought by the Mediterranean waters is spreading by intrusion along the isopicnal surface throughout its basin. Then, by turbulent exchange and reciprocal vertical flows this salt is gradually transferred to the surface. With the surface waters it is driven out from the Black Sea by the upper Bosporus current. So the ratio of total salt flow and the area of Black Sea at a given depth can be assumed to correspond to the vertical flow of salt per unit of area at this depth (F_z^S). Then the coefficient of vertical transfer K_z could be calculated as the ratio: $K_z^{eff} = F_z^S/(dS/dz)$ m^2 s^{-1}, where dS/dz is the vertical salinity gradient.

The calculations of K_z^{eff} from the salt balance by Bogdanova (1965) and later calculations for the water column for various seasons and regions (Boguslavski et al., 1989, Boguslavski and Ivaschenko,1989; 1990; Eremeev and Kushnir, 1996). showed a good coincidence In the upper mixed layer (UML) the intensity of turbulent exchange, characterized by K_z varies usually between ~ 10^{-5} to ~ 10^{-3} m^2 s^{-1} depending on season (Fig. 2). In summer the turbulent exchange in this layer is induced by wind waves. It proceeds to depths of 20 to 30 m during strong winds, when the values of K_z may attain in this layer up to 3 x 10^{-3} m^2 s^{-1}. In the seasonal thermocline layer summer K_z decreases to (2-5) x 10^{-5} m^2 s^{-1}. About the same low turbulence transfer exist around the year in the main pycnocline layer of the Black Sea central regions, while in the MRC zone it is usually (3-6) x 10^{-5} m^2 s^{-1} (Fig. 3).

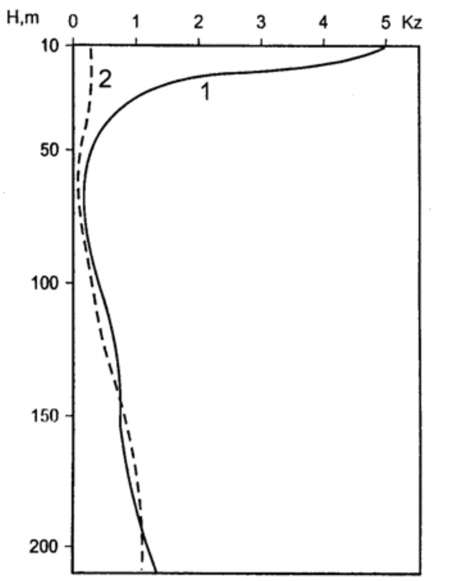

Fig. 2 - Seasonal variation of K_z^{eff} (10^{-4} m^2 s^{-1}) on vertical profiles: 1 — February; 2 — August. After Boguslavski et. al. (1989).

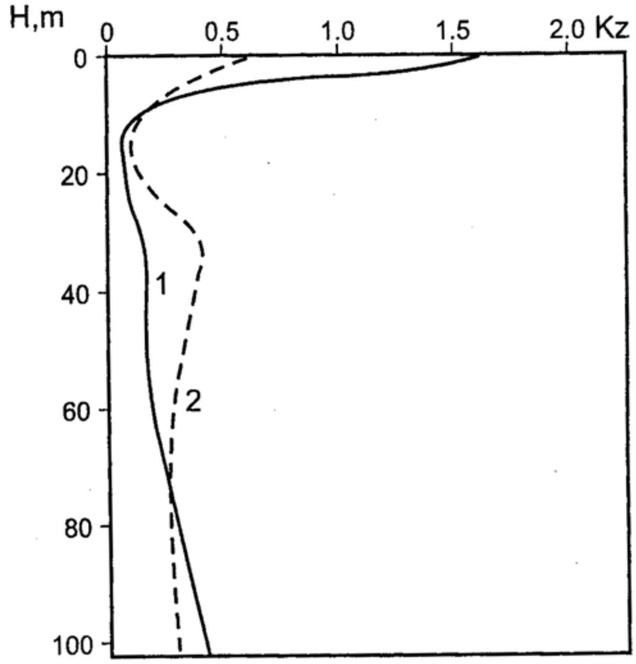

Fig. 3 - Profiles of K_z^{eff} (10^{-4} m^2 s^{-1}) in areas of central cyclonic gyres (1) and in coastal zone of convergence (2) in August. After Boguslavski and Ivaschenko (1990).

The turbulent exchange proceeds opposite the density gradient and thus needs energy. In the surface layer it is driven mainly by wind waves and vertical current shear. In the deeper layers it is supported by the energy of short-period internal waves and seiches due to their turnover and dissipation, by the hydrodynamical instability of currents and by the double diffusion in mesoscale and microscale layers with unstable stratification. The latter was confirmed by the complex character of transfer coefficient (K_z) profiles obtained by continuous recording (Fig. 4).

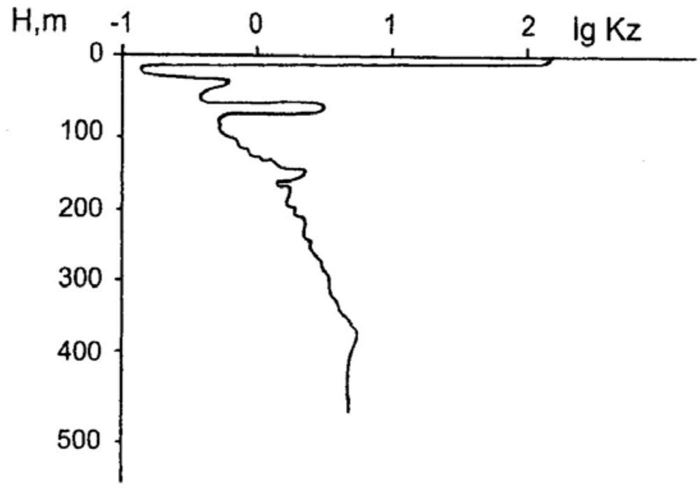

Fig. 4 - Vertical profile of K_z (10^{-4} m^2 s^{-1}) obtained by continuous recording in area to the south of Mean Rim Current, opposite the Crimea. After Eremeev and Kushnir (1996).

The calculation of the vertical turbulent exchange by internal wave below summer thermocline (Efremov, 1998) give values of coefficient from (1-9) x 10^{-6} m^2 s^{-1} in central areas of the Black Sea to (2-10) x 10^{-5} m^2 s^{-1} in the Rim current. The correspondent mean turbulent heat flux was 0.8 W m^{-2} and salt flux – 2.5 x 10^{-7} kg m^{-2} s^{-1}.

Another way of mixing deep and surface waters separated by the pycnocline is the **fine structure lateral thermohaline intrusion** of surface Black Sea waters entrained with Mediterranean waters when passing the Bosporus. When becoming heavier, these entrained waters sink and make inverse travel back to the Black Sea in the upper part of the lower Bosporus current. There they find their isopicnal surface and advect back under the chemocline (Fig. 5). This mechanism of overcoming chemocline stratification explains the relatively fast penetration of the Chernobyl Cesium (^{134}Cs and ^{137}Cs) into deep waters of the Black Sea (Buesseler et al., 1991). The above mentioned entrainment - the advectional mechanism of water mixing should be most effective during winter when cold surface waters participate in the formation of the entrained — transformed waters returning to the Black Sea under the pycnocline. In summer they are formed mostly from waters of the cold intermediate layer (Bodganova, 1965; Murray et al., 1991).

Lateral thermohaline intrusion of the Marmara Sea waters is most important below 100 m in the open part of the Black Sea (Fig. 5). These intrusions induce dissolved oxygen, hydrogen sulphid and nutrient anomalies in the water column.

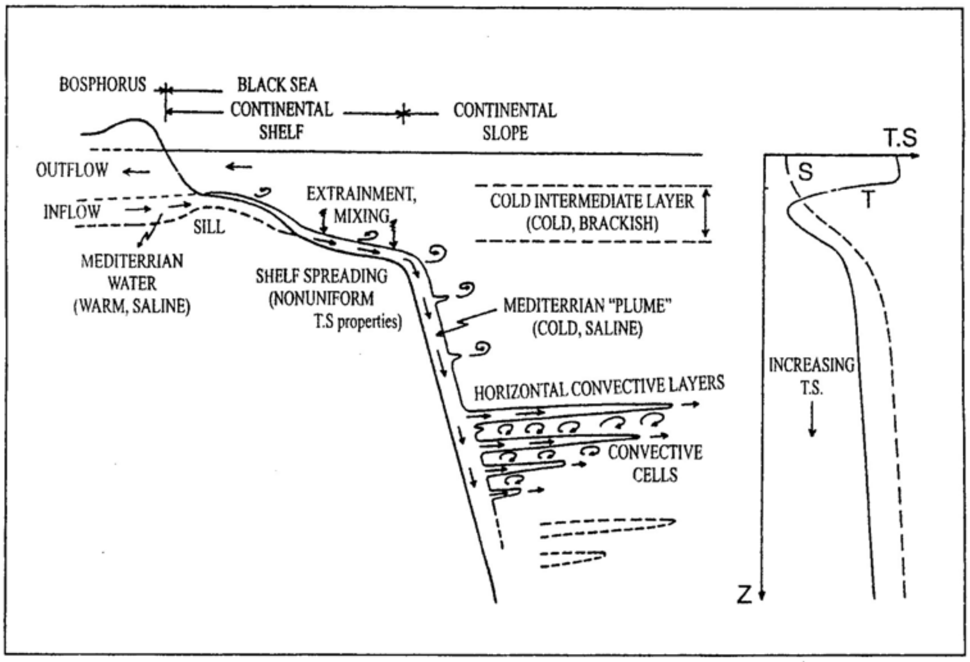

Fig. 5 Schematization of the boundary mixing processes driven by Mediterranean effluent issuing from the Bosphorus. After Ozsoy et al. (1992).

There are three distinct layers that may be distinguished in the Black Sea below the cold intermediate layer (CIL) and above the bottom homogeneous layer (BHL). The upper layer, above 500 m, is the layer where the Marmara Sea water downflow is a warm and salt water pattern. Within this layer, the lateral source of water, on average, supplies heat and salt into the pycnocline making up for the upward heat and salt flux maintained by eddy diffusion in the thermocline. Below 500 m, there is the intermediate quasiisothermal layer where injected water has the same thermohaline characteristics as ambient water, and the upward geothermal heat flux expires. It is possible that the vanishing gradient of potential temperature near depths of 500 m could be linked with this ultimate depth of penetration, and the double diffusive fluxes would likewise be expected to approach zero at this depth. This limit on the depth of efficient vertical mixing is consistent with other indicators of interior mixing, e.g. ^{14}C age distribution, showing smaller mean residence times of intermediate waters (depth < 500 m) compared to the more uniformly aged deep waters. The tritium penetration reaches similar depths.The geothermal heat is almost entirely spent in the third layer, below 650 m, where injected water is colder and saltier than water of the interior of the sea.

Simeonov et al.(1997) modelled the slope convection, heat and salt intrusions, which provide key mechanisms for internal mixing and water mass formation in the Black Sea. As they showed, this penetration depends both on the morphometric characteristics, and on the entrainment of the BSW by sinking MSW. Model simulated intrusions does not penetrate deeper than 500-600 m. This depth range was previously estimated from the observation data as the ultimate depth of penetration of the modified water (Ozsoy et al., 1993). The best agreement with the observations (thickness of the plume, bottom temperature and salinity, pathway of the plume) is simulated when the entrainment of CIW by the MSW is about 1:6.4. This entrainment rate correlates with other independent estimates.

The gradients of heat and salt in the deep Black Sea are both negative upward, and thus it is possible that vertical transport is controlled by **double diffusion**. Double diffusion occurs when the gradients of temperature and salinity have the same sign and thus have opposite effects on the density. The case such as in the Black Sea, where salinity is the stabilizing component, is called the diffusive regime, A master variable for double diffusion is the stability ratio, R_ρ = (Turner, 1973). Double diffusion comes into play when $1 < R_\rho < 70$ and is expected to become stronger as Rp approaches 1 (Turner, 1973). The value of R_ρ, in the deep Black Sea, progressively decreases with depth as the influence of temperature becomes greater and reaches a value of 2 just above the benthic bottom layer (Fig. 10b). A source of heat is necessary to drive this double diffusion, and it could be heat associated with Bosporus inflow and geothermal heat.

A vertical water exchange, provided by a turbulent conductivity coefficient and by hydrodinamical movement of water, induces active pumping of salt across the salinity gradient up to the layer of the main chemocline. This process creates an extremely steep and stable chemocline with a high density gradient in the upper water mass of the sea in conditions of fast geostrophic and synoptic motions. Because of the existence of rather fast currents at great depths, the turbulent exchange and vertical components of flows have enough energy to support a rather intensive vertical mixing in deep waters between 700 to 1,800 m. The geothermal energy also participates in this process supporting the rate of near-bottom convection at $0.5\text{-}0.7.10^{-6}$ m s^{-1}. This means that bottom waters may elevate from 1,500 m depth to the upper boundary of geothermal convection in 25-40 years. The existence of such mixing processes is evidenced first of all by the small gradients of salinity, density and temperature (Fig. 1).

Recent **microstrusture observations** over the Black Sea shelf and slope and in deep-sea regions reveal complex vertical structure and time variability of small-scale mixing.

(Lozovatsky, Fernando, 2002) estimated eddy diffusivities and variables characterizing turbulent mixing on the shallow (depth <30 m) shelf of the Black Sea near the Bulgarian coast in October 1989.

There are seversl prominent features of thermohaline structure and turbulence on the Black Sea shelf at the end of summer heating. These include:

1) an upper boundary layer (z<5 m) with a high turbulent kinetic venergy dissipation rate ($\varepsilon = 7.5\nu\langle du'/dz\rangle$), $\varepsilon > 10^{-8}$ W kg^{-1}, and diffusivity ($K_z = 0.2\, \varepsilon\, N^{-2}$), $K_z \sim (40\text{-}50) \times 10^{-5}\ m^2s^{-1}$,
2) an intermediate layer between the upper and lower boundary layers with patchy turbulence and $N^2 > 10^{-3}\ s^{-2}$, and $K_z \sim (2\text{-}15) \times 10^{-5}\ m^2s^{-1}$,
3) a strong narrow thermocline with mean temperature gradient dT/dz ~ 5°C m^{-1} and $N^2 > 10^{-2}\ s^{-2}$, $K_z \sim (0.2\text{-}0.3) \times 10^{-5}\ m^2s^{-1}$,separated the bottom boundary layer from the water above,
4) a bottom boundary layer with strong vertical shear of current velocity, and $N^2 < 5 \times 10^{-4}\ s^{-2}$, and $K_z \sim (2\text{-}3) \times 10^{-5}\ m^2s^{-1}$,

The above principal layers were embedded with transient phenomena resulting from the interaction between stratification, vertical shear of current and turbulence.

The relatively strong small-scale turbulence has not acted to destroy the stratification. Intermediate stratified region did not contain significant overturning activity.

A rough estimate of the weighted average vertical diffusivity over a unit area of the Black Sea shelf can be obtained as $<K_z> \sim 10^{-4}\ m^2 s^{-1}$. This estimate is close to Munk's estimate for the main oceanic pycnocline based on global energetics. It should be noted that the near-surface (0-3 m) layer contaminated by boat motion was not considered in this analysis. Therefore, the calculated value is probably an underestimation of the column-averaged vertical diffusivity under given background conditions.

Nevertheless, significant variation of K_z with depth shows that the column-averaged diffusivity $<K_z>$ is not an appropriate measure for the computation of vertical transports through the thermocline. The depth-dependent, layer-averaged diffusivities K_z rather than $<K_z>$ should be used for computation of vertical transports in Black Sea shelf waters.

Microstrusture observations over the Black Sea shelf and slope north of the Bosphorus in September 1994 (Ozsoy and Unluata, 1997) reveal a very small turbulent dapycnal diffusivity below the surface mixed layer. The turbulent dissipation rate, ε, peaked at 10^{-8} W kg^{-1} in the seasonal thermocline and fell below 10^{-8} W kg^{-1}in the CIL. Turbulent diapycnal diffusivities, $K_z = 0.2\ \varepsilon/N^2$, were mostly 10^{-6} to 10^{-5} $m^2\ s^{-1}$ in the CIL and less than 10^{-6} ($6.3x10^{-7}$) $m^2\ s^{-1}$ in the seasonal thermocline, where the mean stratification was $N^2 = 1.5 \times 10^{-4}$ s^{-2}. This is at the lower bound of observations in the open ocean, and only five times the molecular diffusivity of heat. This estimates, however, were collected during a few days over a very small area and under light winds (3 m/s). Consequently, it is probable that these results are indeed lower bounds in space and time.

During March 2003 Gregg and Yakushev (2005) observed microstructure profiles over the center of the Black Sea's Western Gyre after cold-air outbreak. Turbulent dissipation rates ε exceeded 10^{-7} W kg^{-1} throughout the mixed layer during the two windy periods and dropped below 10^{-8} W kg^{-1} in the remnant mixing layer when winds let up. They revealed intermittent turbulence in the pycnocline below the mixed layer witn overall ε below 10^{-8} W kg^{-1}. The accompanying mixing produced diapycnal idiffusivities K_z were only (1-4) x 10-6 $m^2\ s^{-1}$, and mean turbulent heat flux varied between 0.2 and 2.0 W m^{-2}. Consequently, turbulent fluxes were too weak to renew significantly either the lower 20% of the cold intermediate layer.

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