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Karst in the Earth's Crust: its distribution and principal types

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Some problems in theoretical karst science ('karstology') are considered. An attempt is made to match the fundamentals of karstology with recent ideas on the structure of the lithosphere and the vertical zonation of the hydrosphere. Boundary conditions of karstogenesis and karst zoning are discussed. The boundaries and the structure of the karstosphere, as well as the place of karst among other geological processes are defined.
The book will be of interest for karstologists, hydrogeologists, geologists and geographers.

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Preface

Over the past few decades, primarily due to advances in deep and super-deep drilling, abundant new information on the deep-seated parts of the Earth’s crust has became available. This enhanced knowledge about the workings of the deep hydrosphere has led to the need to reconsider some established views, e.g.: - the possibility of the development of karst at great depths; - the vertical zoning of karst; - the possibility of karstification of “insoluble” rocks (i.e. of some rock types that are essentially insoluble in the near-surface conditions on Earth); - the sources of aggressive waters; etc. Understanding of the distribution of karst in the Earth’s crust, as well as of the factors controlling karstification at great depth, acquires increased importance as deep-seated karst reservoirs are being targeted for hydrocarbon and geothermal exploration.

A separate problem is the lack of consensus among researchers about the nature of some known processes of karst development, e.g.: - karstification by thermal waters and ore-bearing solutions; - karst related to sulfuric-acid bearing waters; - karst in silicate and aluminosilicate rocks, etc. Despite their differences, these types of karst process are controlled by a limited number of parameters of the lithosphere and the hydrosphere such as the temperature of waters and rocks, their composition and physical states. The changes of karst processes with depth are causally related to the changes in these parameters, which provides the basis for their joint consideration. In this publication we attempted to develop a comprehensive picture of relationships between the various types of karst recognised today (exokarst, endokarst, hydrothermal karst, hypogene karst, silicate karst, etc.) and to provide the spatial and physical “coordinates” of karst in the Earth’s crust.

The authors' contributions to this book were defined, in large part, by their respective fields of professional expertise. Yuri Ezhov and Gennady Lysenin were involved in hydrogeological studies of deep-seated formations associated with hydrocarbon exploration; Yuri Dublyansky studied hydrothermal karst caves and ore deposits; and Viacheslav Andreychouk's main research topic was the conditions and specifics of karst development in different settings and lithologic media. The work is theoretical in its nature and, of course, the conclusions reached by the authors cannot be deemed "final". Rather, the authors consider this work to be a journey into some poorly charted frontier regions of karst enquiry, and as an invitation for discussion with others who are concerned with them.

The first version of this book was published in 1992 in Russian (publication of the Institute of Geology and Geophysics, Siberian Branch of the Russian Academy of Sciences, Novosibirsk, Russia). The book was dedicated to the memory of Y.A. Ezhov, who passed away in 1991.

The current, second edition of the book is slightly modified and updated. The authors express their gratitude to Derek Ford for thorough substantive and editorial review.

Chapter 1. Non-traditional types of karst

Defining the non-traditional types of karst

Before addressing the issue of the distribution of karst in the Earth's crust and describing its various types, we need to define the subject of our study. There are several tens of definitions of karst in the scientific literature (see Timofeev et al. 1991 for a terminology review). In the most general terms, karst may be defined as *a process of interaction between soluble rocks and different waters, as a result of which characteristic features develop on the Earth's surface and underground*. Most definitions emphasize the close connections between the process and the resulting features (cause and consequence), the dominant role of dissolution, and the peculiarities of the resulting morphologies (the predominance of cavities and depressions). Many definitions contain qualifiers addressing the more specific characteristics of karst, such as the composition of the host rock, types of waters involved, nature of their circulation, impact of karst development on landscape, etc.

In this work we adopt the definition of Andreychouk (1991), who viewed karst as a *system of processes and phenomena developing and occurring underground and at the Earth's surface as a result of the interaction (dissolution, transport, and deposition of matter) of natural waters with, rocks that are soluble in the given situation.* This definition is broad enough to encompass all the processes and features discussed in this paper.

Several types of what may be termed "**non-traditional karst**" will be discussed below in more detail. They are silicate karst, hydrothermal karst, endokarst, and ore karst. The reasons why these types are considered non-traditional include their comparative rarity and limited distribution and the fact that, generally, they are insufficiently studied. They are related, genetically or paragenetically, to “foreign” processes, which can conceal their karstic nature; and they are commonly developed in rocks that are considered non-karstifiable.

We emphasize that expression “non-traditional karst” is but a working term which bears no genetic or classifying connotations.

Silicate karst

Silicate karst is *karst developing in rock composed, partly or entirely, of silicate minerals*. Many researchers have reported morphological forms and features similar to karstic ones in silicate rocks. However, the paucity of factual data and, more importantly, the lack of theoretical models, hinder the resolution of the issue whether karst processes can be effective in such rocks.

Examples of surficial and subterranean dissolution-related karst-like (or truly karstic?) features in silicate (metamorphic and igneous) and siliceous rocks (sedimentary and metamorphic), such as small depressions, rillenkarren, solution-widened fissures, cavities and caves are known well in the literature (e.g., Mainguet, 1972; Jennings, 1983; Pouyllau, 1985; Busche & Erbe, 1987; Young, 1988). Perhaps the best-known examples occur in tropical areas of South America (Guiana and Brazilian highlands), central and South Africa and in Australia (e.g., Martini, 1987; Galan & Logarde, 1988; Thomas 1994), where the karst-like features are developed in different varieties of quartzite. Peculiar surficial and subterranean karst forms were reported from siliceous sedimentary rocks, opoka (a porous, flinty and calcareous sedimentary rock), and glauconitic sandstone in the Trans-Urals (Russia) by Andreychouk (2000).

Many researchers studying geomorphology in tropical regions (Ollier, 1965, Twidale, 1982) and some others (Jakucs, 1977) emphasize the important role of dissolution in the weathering of granites, which leads to development of karst-like landforms (e.g., rillenkarren, pinnacles similar to karst remnants) and types of relief (polygonal hilly relief similar to cockpit karst). The similarity in surface morphology of granite and limestone terrains was noted more than 50 years ago (Bulla, 1954), and some researchers identified such peculiar forms in granites as karst (Rasmusson, 1959). Tentatively corrosion forms were also reported from other silicate rocks, such as basalt and dunite (Loffler, 1978).

The solubility of silicate rocks is low in the conditions generally prevailing at the Earth's surface (but not at great depth!); accordingly, karst is a slow process in such rocks. In order to develop "classical" karst forms (e.g., caves), silicate rocks therefore must remain subject to karstification for geologically long periods of time. Recognizing this, Maksimovich (1975) introduced the term **bradikarst** (from Greek *bradus* – slow, weak) for karst in ferruginous quartzite, and in silicate rocks in general. According to relative speed of development, Maksimovich subdivided karst onto three major categories: - **tachikarst** (from Greek *tachus* – fast) – very fast karst (in halite and sulfate rocks); - **karst** (mostly in carbonate rocks); and - **bradikarst** – very slow karst (in silicate rocks).

It is widely accepted that the controlling role in silicate karst solution in tropical settings is played by organic matter (guano, humus, animal remnants, etc.) which acidify the infiltrating waters. However, during dissolution of silicate minerals, alkalinity may have even greater importance (**Fig. 1**). Strongly alkaline waters (pH>10) are very aggressive toward silicate rocks (Mitsiuk, 1974). Very high pH values are characteristic deep in the Earth's crust. It can be conjectured that at least some of the large

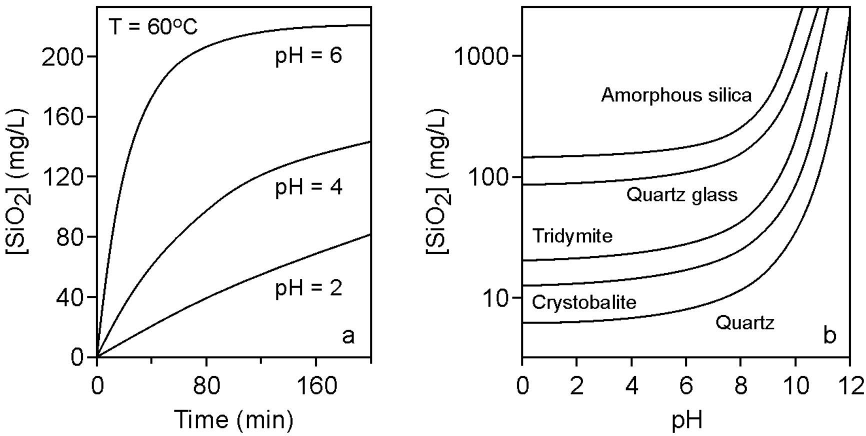

Fig. 1. Solubility of SiO_2 in water in different conditions (according to Mitsiuk, 1974): a – the dissolution rate of amorphous SiO_2 at T = 60°C and varying pH; b – solubility of different varieties of silica with change of pH in standard conditions.

caves in quartzite (e.g. in the Guiana Highlands) might have been originally formed at significant depth and filled with mineral deposits (e.g., calcite) which later was removed by dissolution in the near-surface conditions.

Sedimentary silicate rocks composed predominantly by detrital remnants of microorganisms (radiolarian, etc.) do not require high-alkalinity solutions to be dissolved effectively. The solubility of amorphous silica, which constitutes the bulk of such rocks, is up to an order of magnitude greater than the solubility of quartz, and may reach tens and hundreds of mg/L (Pulina & Andreychouk, 2000) (**Fig. 1**).

In some cases, cavities in silicate rocks develop by dissolution in thermal waters. Such cavities were reported from quartzite, quartz sandstone, jasperoid, skarn (Dublyansky, 1990), as well as from quartz veins (Dublyansky, 1990; Andreychouk & Lavrov, 1992).

Hydrothermal karst

The concept that some caves could have been formed by ascending thermal waters and gases was first introduced as early as in the mid-19th century (Noggerath, 1845; Desnoyers, 1845; as reported in Shaw, 1992 and Bosák, 2000). Somewhat later, studying Pb-Zn deposits in carbonate rocks, Pošepny suggested that these ores were emplaced in dissolution cavities and that the latter owe their existence to the same solutions from which, at later stages, the ores were deposited (Pošepny, 1893). A little later, in his capital “Treatise on Metamorphism”, Van-Hise provided explanations for how and why hydrothermal solutions move advectively through the rocks, and what causes their aggressiveness (Van-Hise, 1904). He conjectured that most hydrothermal solutions originate as common meteoric waters that become heated during their circulation deep in the Earth’s crust (this was a visionary insight amply corroborated several decades later, with the advent of stable isotopes in hydrogeology!).

These early works laid a foundation for the concept of **hydrothermal karst**, which at first attracted little attention from either ore geologists or karst researchers. The first detailed works elucidating the role of hydrothermal karst activities in the localization of ores at Pb-Zn deposits were published only in the 1920s (Locke, 1926, Grigoriev, 1928). In the 1930-40s, karst researchers started to pay more attention to the subject. The concept has become firmly established only in the second half of the 20th century.

The most comprehensive summary on hydrothermal karst from the perspective of karst science was published by Maksimovich (1969). The current understanding of the role of hydrothermal karst in ore systems was summarized by Kutyrev et al. (1989). An attempt at the synthesis of these two lines of research, ore-geological and karstologic was undertaken by one of the authors of this paper (Dublyansky, 1990).

A number of different terms was used to define karst developed by the action of thermal, commonly mineralized, waters, e.g., **ore-thermal** (Makarenko, 1947), **thermomineral** (Michal, 1929-1930, Kunsky, 1957), **endogenic** (Dublyansky & Kropachev, 1981; Ivanov, 1983), **hypogenic** (Kutyrev & Liakhnitsky, 1982), and **hot** (Kutyrev et al., 1989). These terms were later superseded by presently universally accepted term **hydro-thermal karst**. In the broadest sense, hydrothermal karst is defined as the *process of dissolution and possible subsequent infilling of cavities in the rock by the action of thermal water* (Dublyansky, 1990, 2005). Kutyrev et al. (1989) defined hydrothermal karst as the *result of dissolution, metasomatic replacement, transport of the dissolved material by flowing water and gases of deep-seated origin, which enter the massif from below or laterally*. The latter authors stressed the important role of hydrothermal karst in the emplacement of ore deposits.

Besides carbonate rocks hydrothermal karst is known to occur in sulfate and silicate rocks, rock salt, as well as in sulfide ore bodies. Presently active hydrothermal karst is most common in areas of alpine orogeny, as well as in zones of modern volcanism. Development of hydrothermal karst involves the movement of significant amounts of heated waters; therefore it is commonly found an association with deep-seated faults, as well as in zones of thinning of the Earth's crust. Ford (1988) estimated that no less than 10% of all studied karst caves underwent a stage of hydrothermal development.

Hypogene karst

The concept of the **hypogene speleogenesis (hypogene karst)** has been developed in recent years (Ford & Williams, 1989; Worthington &

Ford (1995), Palmer, 1991, 2000, Klimchouk & Ford, 2000; Ford, 2006, Klimchouk, 2000, 2007). Hypogene karst is defined as *the formation of caves by water that recharges the soluble formation from below, driven by hydrostatic pressure or other sources of energy, independent of the recharge from the overlying or immediately adjacent surface* (Ford, 2000). Hypogene karst thus is one component of dual system of karst, the other component being hypergene (or epigenic) karst[1]. Consistent with the logic of the definition above, the hypergene karst develops under action of waters which are recharged from overlying or immediately adjacent surfaces. It is apparent that the definitions above rely on a loosely defined quantifier (overlying or immediately adjacent surface).

The concept of hypogene speleogensis was discussed in depth by Klimchouk (2007), who put forth the notion that the specific hydrogeological settings which lead to hypogene speleogenesis transcend the particularities of the physico-chemcial mechanisms which create the aggressiveness of water toward rocks (e.g., the presence of sulfuric acid or elevated temperatures). The concept of hypogene karst development thus epitomizes the hydrogeological approach to karst studies.

Endokarst

Two similar, but rather loosely defined terms, **endokarst** and **endogenic** karst, can be found in geologic and karst publications. For instance, Băcăuanu et al. (1974) use the term ‘endokarst’ to describe any underground karst features. The term ‘endogenic karst’ can refer to: - karst

[1] We note that the term **epigenic** karst is now strongly entrenched in literature as an antonym for hypogene karst. Newerthelless, the term hypergene (or hypergenic) appears to be more appropriate to us (see also Klimchouk, 2007, p. 6). The pair of terms: hypergene-hypogene (processes, minerals, deposits) is in wide use in geological literature to discriminate between the deep-seated and surficial phenomena. In contrast, the term epigenesis is commonly used in geological literature to define a geologic change in mineral content of rock after the rock was formed (e.g., epigenetic mineralization = secondary mineralization). The latter term has thus more temporal rather than spatial connotation.

formed by endogenic processes (Mechrsprächiges…, 1973); karst developing due to flow of hydrothermal waters (Ivanov, 1983); - karst related to circulation of postmagmatic solutions (Dublyansky & Kropachev, 1981). Kutyrev et al. (1989) use ‘endokarst’ as a synonym of hydrothermal karst.

In this paper the term **endokarst** is assigned a very different and strict meaning. The new definition exploits the fundamental differences between the karst process in the upper part of the Earth's crust where fluid pressures do not exceed hydrostatic values (hydrostatic pressure; HSP), and karstification in the lower parts of the crust, where ascending aggressive fluids have suprahydrostatic pressures (SHSP) (Ezhov et al., 1988). Endokarst is defined as *karst developing in the lower part of the karstosphere*[2]*, where the internal (telluric) energy is dominant and the rock mass is under lithostatic pressure which exceeds its tensile strength*. Movement of water in this zone of the Earth's crust occurs via buoyancy, the discharge occurs in a pulsed mode through hydrofracturing of a buffer interval (see below), and recharge occurs by influx of new volumes of fluid from the deeper parts of the Earth's crust and upper mantle. Under these conditions a cavity can only exist if it is filled with fluid having supra-hydrostatic pressure.

Ore karst

Ores emplaced in karst cavities were recognized as a specific genetic type of ore deposits in the later 19th Century by Pošepny (1871). At the beginning of the 20th Century, a karstic origin for ore-bearing cavities was recognized in many major ore deposits, such as Raibl and Iglesias (Pb and Zn) in Europe, Tyuya-Muyun (U, Ra, and V) in Asia, Tsumeb (Pb, Zn, Cu,

[2] The term ‘karstosphere’ is discussed in detail in Chapter 3.

Ge, Ga, Cd, and As) in Africa, and the numerous deposits of the Mississippi valley-type (Pb and Zn) in America. The term **ore karst** was introduced into the geological literature by Ivensen (1937) who, however, failed to provide a formal definition. The term was used both to define karst cavities of any origin which contain ores (Kreiter, 1941) and (predominantly) underground cavities in carbonate rocks developed due to dissolution by waters whose aggressiveness is related to sulfuric acid released during oxidation of sulfide ores (Kreiter, 1941; Nikolaev, 1962). For the latter situation Sokolov (1960) used the term **sulfuric acid karst**. Yet another use of the term 'ore karst' was to define the dissolution cavities developed within sulfide ore bodies (Aprodov, 1962; Sokolov, 1962). Later, this lithologic type of karst was termed **sulfide karst** (Gariainov, 1980). In recent years, to define karst hosting ore mineralization, the term **ore-bearing karst** has been used. Liakhnitsky (1987) defines the latter as *localization of ore material in previously formed cavities (and systems) of karstic and hydrothermal karstic origin through processes of inflow and infiltration of cold surficial and thermal waters*.

We suggest that the two terms, **ore karst** and **ore-bearing** karst must be differentiated. The former refers to the *process* of accumulation of ores and associated gangue minerals, residua, etc. during karstification (e.g., accumulation of residual-infiltration and infiltration ores; Tsikin, 1985); in other words, cavities and their infilling are syngenetic. The latter refers to the physical *phenomenon* of ores being found in karst cavities. Processes, leading to these occurrences might be of different nature, and not necessarily karstic.

Chapter 2. Karst in the Earth's crust

Hydrodynamic zoning of the Earth's crust

A scheme for the planetary-scale hydrodynamic zonation of the Earth's crust was developed by Russian scientists in the 1960s-80s (Ivanov, 1966; 1970-a;-b; Ezhov & Vdovin, 1970; Ezhov & Lysenin, 1986; 1988). The scheme was based on the synthesis of large amounts of data from observations in deep boreholes, from seismic studies carried out in different regions on all continents and, on experimental results of rock deformation at high PT parameters as well. In this scheme, shown in **Figure 2**, the underground hydrosphere is divided into three planetary-scale hydrodynamic (baric) zones: I – zone of hydrostatic pressures; II – zone of transitional pressures; and III – zone of lithostatic pressures. The three zones have developed in response to the geologic evolution of the crust. Each of them has its particular baric, temperature, hydrochemical, and petrophysical properties and performs specific hydrogeological functions.

Zone I (hydrostatic zone, HSZ) comprises the upper part of the underground hydrosphere. It contains free waters flowing under simple gravity as well as artesian and other confined waters with elevated pressures. The pressure in this zone derives primarily from the hydrostatic pressure of groundwater entering the system from surface recharge areas. Free waters typically form descending or lateral flows. Under certain tectonic and lithologic circumstances, water under pressure may also develop ascending flows (flows directed upward along inclined permeable beds during recharge from compacting rocks and from lower zones with suprahydrostatic pressure; discharge in the overlying deposits along faults, in areas of weakening of aquitards, etc.).

Based on the flow dynamics, the HSZ zone can be subdivided into three hydraulically connected sub-zones): I-A – an uppermost sub-zone of active circulation; I-B – a sub-zone of impeded circulation; and I-C – a sub-zone of strongly impeded circulation. With depth, the porosity and fracture density of rocks decreases due to their compaction and the temperature of waters increases. The TDS (total dissolved solids) content in waters also increases as their chemical character changes from dilute sodium- and, in regions of predominantly carbonaceous rocks, calcium-bicarbonate (such waters effectively dissolve carbonate and sulfate rocks, and rock salt) in sub-zone I-A through sodium-sulfate and magnesium-chloride to calcium-chloride in sub-zone I-B. Concentrated calcium-chloride brines dominate the I-C sub-zone; the aggressiveness of such brines with respect to most types of the bedrock is very low.

The pressure of waters in the aquifers, and in the HSZ as a whole, is close to hydrostatic and, as a rule, does not exceed it by more than 10-15 %. The pore pressure in dense aquitard and aquiclude rocks (primarily, clays) exceeds the hydrostatic pressure by 15-20 %. Aquicludes constitute hydrodynamic barriers impeding the ascending movement of fluids and commonly serving as caps for oil and natural gas deposits.

We note that pore pressures exceeding the pressure in the aquifer should not be called “anomalous” (Anomalously High Aquifer Pressure; AHAP). Geologically short-lived, truly anomalous high aquifer pressures can develop in some parts of zone I (most commonly in sub-zone 1-B), where they are often associated with thermal, gas, and other anomalies. This phenomenon is usually explained by localized breakthroughs of high-pressure fluids from underlying **zone of transitional pressures** (meso-zone; MZ). Depending on the specific lithologic and geotectonic setting, thickness of HSZ ranges from 0.5-0.6 km up to 6-7 km and more.

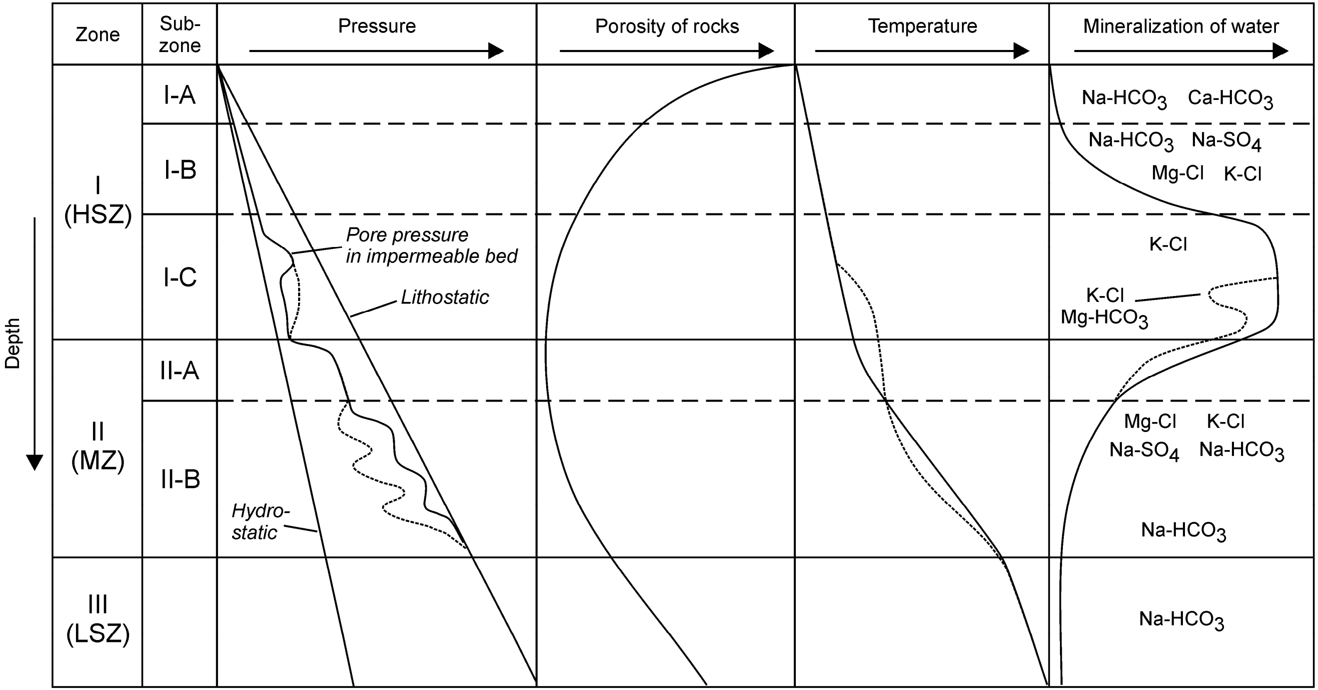

Fig. 2. Scheme of the vertical hydrodynamic zoning of the Earth's crust showing changes in some important parameters (after Ezhov & Lysenin, 1986; 1988): Solid line – parameters before the impulse breakthrough of fluids through buffer sub-zone II A; dotted line – parameters after the breakthrough. Definitions of zones and sub-zones are given in text.

The HSZ is thinner under areas of alpine orogeny and thicker under old platforms and folded mountain structures (Ezhov & Vdovin, 1970; Ezhov & Lysenin, 1986).

Zone II (meso-zone, MZ) is transitional between zone of the hydrostatic pressures (I; HSZ) described above and the underlying zone of the lithostatic pressures (III; LSZ). Across this zone, the tremendous suprahydrostatic pressures (SHSP) generated in zone III may be transferred and attenuated. The pressure in the MZ increases with depth from nearly-hydrostatic to nearly-lithostatic values. The increase is non-linear and commonly abrupt. Most occurrences of SHSP recorded in deep water-bearing intervals in boreholes (pressure gradients of 1.2 to 2.3 MPa/100 m) are associated with this zone.

Based on the fluid dynamics and some specific effects, the MZ can be subdivided into two sub-zones (**Fig. 2**): II-A – a buffer sub-zone or the sub-zone of maximum compaction of rocks; and II-B – a sub-zone of variable suprahydrostatic pressures or of reduced rock density. Sub-zone II-A plays the role of a planetary-scale regulator of defluidisation of the deep-seated parts of the Earth's crust. When pressure of fluids beneath this zone reaches values sufficient for natural hydro-fracturing, the fluids break through the buffer interval into the shallower zones. These impulse breakthroughs create baric, hydrochemical, thermal, gas and other anomalies in the HSZ (Ezhov & Lysenin, 1987; 1988).

The inevitability of the formation of an interval of maximum compaction of rocks was demonstrated on theoretical and computational grounds by Ivanov (1966; 1970-a;-b). He demonstrated that within the Earth's crust, both on continents and under oceans, the closure of open pores and fractures occurs at given depths. As a result, a virtually impermeable zone develops, corresponding to sub-zone II-A in our scheme. The closure depth depends on the strength of the water-filled rocks. The sealing properties of this interval may be further enhanced due to precipitation of secondary minerals.

This meso-zone can develop in rocks of any composition; it is common, however, that either its upper boundary or the entire buffer sub-zone develops in regional clayey or chemogenic strata, because in such rocks the hydrodynamic barrier confining the SHSP could be created more readily than in other, more competent, rocks.

In sub-zone II-B both the porosity and the fracture density of rocks increase due to the corrosive action of aggressive fluids ascending from below, and due to hydro-fracturing. This "loosening" of rocks does not encompass the whole volume of the sub-zone but is restricted to the areas of preferential migration of super-dense fluids. The highest SHSP gradients are found in the least permeable rocks. With depth, the confining properties of aquicludes deteriorate due to the increasing tendency for hydro-fracturing to occur. The gradients between the aquifer and the pore pressures gradually diminish.

The temperature in the meso-zone typically increases with depth faster than in HSZ. In addition, an inversion of the chemical zoning of water occurs, associated with pulsed influxes of deep-seated (mantle, meta-morphic) waters. With increasing depth, the chemical types (facies) of water change from calcium-chloride through magnesium-chloride and sodium-sulfate to sodium-bicarbonate ones. The total mineralization of waters decreases. CO_2 and H_2S prevail among the dissolved gases (Ezhov, 1978; 1981; Ezhov & Lysenin, 1985; 1986). During a breakthrough of deep-seated fluids through buffer sub-zone II-A, the hydrochemical inversion can locally be transferred into the HSZ.

An important feature is that the pressure in the meso-zone changes with time due to the periodic breakthrough of pulses of deep-seated fluids: it reaches the highest values between breakthroughs and drastically decreases during a breakthrough phase (see **Fig. 2**). The dynamics of the periodic discharge of fluids across the buffer interval was recently documented by fluid inclusions studies in Southeastern Brazil (Faleiros et al., 2007).

The thickness of the meso-zone depends on the regional lithologic and structural properties, as well as on the properties of fluids: it ranges between 900-1500 m and several kilometers. It is presumed that the thickness of meso-zone is inversely proportional to the density of the rocks in which it develops. Under seas and oceans, due to additional pressure from the water column, the upper surface of the meso-zone is closer to the sea floor than it is to the land surface on the continents. At sufficiently great depths of water (several kilometers) sub-zone II-A could occur at or near the sea floor itself. In such circumstances, any deep-seated thermal fluids breaking through the buffer zone will discharge in and mix with the sea water. Their dissolved mineral loads (including ores) are precipitated on the sea floor or in unconsolidated sediments there due to the fall of pressure and temperature (Ivanov, 1970-b).

Zone III (lithostatic zone, LSZ) comprises the lowest parts of the Earth's crust that have been studied (to a depth >12-15 km). The pressure in this zone is controlled by lithostatic load and by endogenic sources. General pressure gradients change from 2.1 to 2.6 MPa/100 m, depending on the mean density of water-bearing rocks. The lithostatic zone hosts the ascending fluxes of thermal liquid-vapor fluids, of endogenic and metamorphic origin, that constitute the major source of recharge of the meso-zone.

A characteristic feature of the LSZ is the decreased density of rocks due to permanent natural hydro-fracturing. The data from very deep drilling suggest that hydro-fractured rocks are concentrated in certain horizons and zones along the migration pathways of the deep-seated fluids. The degree of porosity and fracturing in such zones is significantly higher than in the meso-zone, and may reach the values characteristic of the hydrostatic zone (Ivanov, 1970-a). Due to hydro-fracturing, "traditional" aquitard and aquiclude rocks lose their confining properties in the lithostatic zone. Salts, due to their high plasticity and fluidity, are squeezed up into the upper parts of the geological cross-section.

Temperature further increases with depth in the lithostatic zone, but at a slower rate than in the hydrostatic zone and meso-zone (Ivanov, 1966; 1970-b; Ezhov, 1981). The content of dissolved gases increases to extremely high values, mostly at the expense of acidic components.

Waters in the LSZ belong to sodium-bicarbonate type and have relatively low mineralization. They are enriched in specific deep-seated components, such as CO_2, He, F, As, etc., some of which have characteristic isotopic properties (Ezhov, 1978; Ezhov & Lysenin 1985). Compared to their counterparts from near-surface settings, the fluids in the LSZ have anomalously high aggressiveness with respect to many sedimentary, metamorphic and igneous rocks. Yakutceni (1984) suggests that at high PT water acquires properties of strong acids. The predominantly ascending character of the fluxes of water ensures removal of the dissolved matter from the lithostatic zone.

The most prominent example of the high energy contained in the lithostatic zone is the mud volcanism reported in areas of alpine orogeny. This energy can also be expressed as hydrovolcanism (Ivanchuk, 1974) and hydrothermal activity. The LSZ is the major source of sodium-bicarbonate thermal waters discharging at the Earth's surface. They are sometimes enriched in Si.

The vertical hydrodynamic zonation of the Crust described above is a cumulative result of the processes of sedimentation, rock transformation, and tectonics. In turn, physical properties of the zones affect different geologic processes. For example, Ivanov (1970-b) argues that hydro-thermal mineralization could be controlled not by the cooling of fluids or by their migration into local zones of tectonic stress but, rather, by the abrupt pressure drop associated with the passage of the ore-bearing fluid through the buffer interval. The presence of rocks with lowered density and containing fluids under suprahydrostatic pressures plays prominent role in the development of faults, particularly nappes and overthrust folding. Of course, such hydrodynamic zoning must play an important role in the distribution of karst processes in different parts of the Earth's crust.

The vertical zoning of karst

The scheme introduced above implies that two fundamentally different settings will exist in the subterranean hydrosphere: one of hydrostatic pressures and one of suprahydrostatic pressures. Accordingly, the karsto-sphere (see Chapter 3 for discussion) should be subdivided into two storeys: the exokarstic (corresponding to zone I) and the endokarstic, encompassing zones II and III (Ezhov et al, 1988; Ezhov & Lysenin, 1991).

The development of karst in the upper, exokarst storey is primarily governed by external factors, such as the infiltration and inflow of meteoric water, the expulsion of formation waters under the pressure of the overlying rocks, by topographic relief, climate, etc. In the lower, endokarst storey not only the carbonates but also the silicate and aluminosilicate rocks, traditionally considered only weakly or very weakly soluble, can be subject to intense dissolution by fluxes of ascending deep-seated fluids (**Fig. 2**). For example, one of the most abundant and stable, resistant forms of silica under crustal conditions, quartz, at T = 300 to 350℃ and P = 200 to 250 MPa acquires an aqueous solubility similar to that of gypsum and anhydrite in typical near-surface conditions (up to 0.1 mas.%; Fyfe et al. 1981; Yakutseni, 1984).

The vertical zonation of karst proposed here is compared in **Fig. 3** with the most commonly used schemes of other authors, as well as with the scheme of the vertical hydrodynamic zoning. The characteristic features of the sub-zone of active water circulation I-A include: - infiltration and inflow of meteoric recharge karst waters; - relatively high velocities of all flow, descending, lateral, and ascending; - local drainage into erosion entrenchments, large lakes, and the sea. The intensity of karstification is high, leading to the development of abundant and diverse surficial and subterranean features, including large subaerial and subaqueous cavities. The intensity generally decreases with depth.

In the two lower sub-zones, I-B and I-C, of the HSZ (= exokarstic storey), the drainage is of more regional extent, broadly directed towards different parts of the ocean basins. In addition, the discharge may be upwards along large disjunctive structures. The dynamics of karst water flow, controlled by the hydrostatic head from the areas of infiltration recharge, is low and diminishes with depth. As in the I-A sub-zone, descending, lateral, and ascending flows exist but their velocities are low. Karstic waters commonly discharge through large subaqueous (submarine) springs or through subaerial springs near the shores of seas and oceans. The intensity of karstification is low, mostly resulting in development of pores, vugs, small cavities, and solution-widened fractures.

Larger cavities form, rather as the exception, in the areas of enhanced permeability of rocks associated with disjunctive geologic structures. The major agent of dissolution is CO_2 produced by oxidation and biochemical processes (e.g., oxidation of hydrocarbons and disseminated organic matter), as well as that introduced during breakthroughs of low-salinity, gas-charged fluids from the deep-seated zones of the Earth's crust.

In the endokarstic storey, the planetary-scale MZ and LSZ zones are responsible for the global drainage of the deep parts of the Earth's crust through the ascending movement of deep-seated fluids and their periodic pulses of discharge into the HSZ via hydro-fractures. Little is known at present about karst in this storey. Most of information about karstification of rocks in the suprahydrostatic zone can be gleaned from the results of deep drilling. Studies of drill cores, geophysical borehole testing, and analyses of the technological drilling information suggest that karst at such depths is mostly represented by solution pores, vugs, and solution-enlarged fractures. In many sedimentary basins of the world the cavernosity of Paleozoic and Mesozoic carbonate rocks in the SHSZ may reach 12 to 28 %, and their permeability $n \cdot 10^{-14}$ to $n \cdot 10^{-12}$ m^2 (the maximum value reported is $8 \cdot 10^{-12}$ m^2; Yakutseni, 1984; Maksimov et al., 1984).

<table>
<tr><th colspan="4">Hydrodynamic zoning of Eath's crust</th><th colspan="5">Zoning of karst</th></tr>
<tr><th>Zone</th><th>Sub-zone</th><th>Flow direction</th><th>Drainage</th><th>Karsto-sphere storey</th><th>Karst intensity</th><th>Primary karstified rocks</th><th>Hydrodynamic zone of karst water</th><th>Zone of karstosphere</th></tr>
<tr><td rowspan="3">I
(HSZ)</td><td>I-A</td><td></td><td>Local</td><td rowspan="3">Exokarstic</td><td>High</td><td rowspan="3">Halite,
sulfate,
carbonate</td><td>Aeration
Seasonal fluctuation
Saturation</td><td>Exokarstic</td></tr>
<tr><td>I-B</td><td rowspan="2"></td><td rowspan="2">Regional</td><td rowspan="2">Low</td><td>Deep circulation</td><td>Mesokarstic</td></tr>
<tr><td>I-C</td><td rowspan="4"></td><td>Hypokarstic</td></tr>
<tr><td rowspan="2">II
(MZ)</td><td>II-A</td><td></td><td>Trans-
migration</td><td rowspan="3">Endokarstic</td><td>Cementation</td><td></td><td rowspan="3"></td></tr>
<tr><td>II-B</td><td></td><td rowspan="2">Global</td><td rowspan="2">Variable,
up to high</td><td rowspan="2">Sulfate,
carbonate,
silicate,
Al-silicate</td></tr>
<tr><td>III
(LSZ)</td><td></td><td></td></tr>
<tr><td colspan="7">Ezhov, Lisenin, and Andreichuk (1988)</td><td>Sokolov (1962)</td><td>Maximovich (1979)</td></tr>
</table>

Fig. 3. Vertical hydrodynamic zoning of the Earth's crust compared with schemes of hydrodynamic zoning for the karstosphere. See the text for definitions of the hydrodynamic zones and sub-zones.

Large cavities have not been reported from the endokarstic storey. This, however, might be the result of methodological and technical difficult-ties of detecting and identifying them. For instance, the yield of drill-hole core from karstified intervals typically decreases due to the destruction of the bedrock by the drilling itself. Large hydraulic depressions created during the testing of such intervals leads to the closure (collapse) of any open spaces because the rock cannot withstand the lithostatic load. The latter point serves to emphasize that any cavities formed within the endokarstic storey may exist only if filled with fluids under suprahydrostatic pressures. The drop of the fluid pressure leads to the destruction of such cavities.

Dissolution forms can also be preserved if they are subsequently filled with minerals precipitating from solutions. For example, some large karst features reported in silicate rocks (e.g., large caves in quartzite in Venezuela and Gabon) could be attributed to dissolution at high temperatures and pressures in the conditions prevailing in the endokarstic storey, followed by their infilling with secondary minerals (usually calcite), their subsequent shift into the upper exokarstic zone in the course of general crustal uplift and denudation and, finally, dissolution and removal of the infilling material by aggressive infiltration waters (Ezhov et al., 1988). Smaller forms of endokarst (solution pores, caverns, stylolite seams) are relatively common in silicate rocks.

Karst developing in the deep parts of the Earth's crust may become highly important in studies of hydrocarbon reservoirs. It is well known that the capacity of pore reservoirs in oil and gas deposits typically diminishes with depth, where the fluids increasingly accumulate in fractured reservoirs. In order to store the same amount of hydrocarbons as in pore reservoirs, fractured reservoirs must be much thicker. The thickness of hydrocarbon reservoirs in the conditions of SHSP (i.e., the upper part of the endokarstic storey) is limited because the pressure required for hydro-fracturing in the buffer interval is quickly reached (Lysenin & Ezhov, 1986; 1987). In such conditions, karstic reservoirs hold high potential for creating economic deposits of hydrocarbons. Relatively thin karstified intervals may host substantially greater amounts of hydrocarbons in comparison with equivalent pore or fracture reservoirs. An important point here is that, in conditions prevailing in the endokarstic storey, karstic reservoirs may also develop in silicate and aluminosilicate rocks, in addition to the carbonates.

The zonation of hydrothermal karst

It is empirically established that the minimum temperatures in the upper part of the endokarstic storey range between 80 and 100℃. Karst developing there thus conforms to any definition of hydrothermal karst. Hydrothermal karstification is commonly thought to be an azonal process, meaning that its development transcends the vertical hydrological zoning of the Crust (Tsikin, 1981; 1985). This apparent lack of zoning may be related to a number of reasons, including (i) occurrence of fossil hydrothermal karst which, indeed, might not be related to the contemporary hydrological zoning; (ii) the fact that hydrothermal karst zoning generally is at much larger scales, comparable to those of the hydrosphere, and (iii) because such "normal" hydrosphere-scale zonation can be overprinted by anomalous hydrothermal karst. Here we introduce the working term '**anomalous hydrothermal karst**' to describe *the hydrothermal process developing in zones where the steady-state thermal field of the hydrosphere is disturbed*.

This typically happens in spatially restricted zones of enhanced permeability which allow the concentrated (focused) ascent of water; such water carries the heat from the deep-seated levels up toward the Earth's surface (**Fig. 4**).

According to the medium in which the karstification takes place, hydrothermal karst may be subdivided into **subaerial** and **subaqueous zones**. The possibility for hydrothermal karst to develop in subaerial conditions was proposed by Pávai-Vajana (1931), Müller (1974), and Szunyogh (1982). The major agent of karstification here is the condensation of water vapor forming above the surface of the underground thermal lakes.

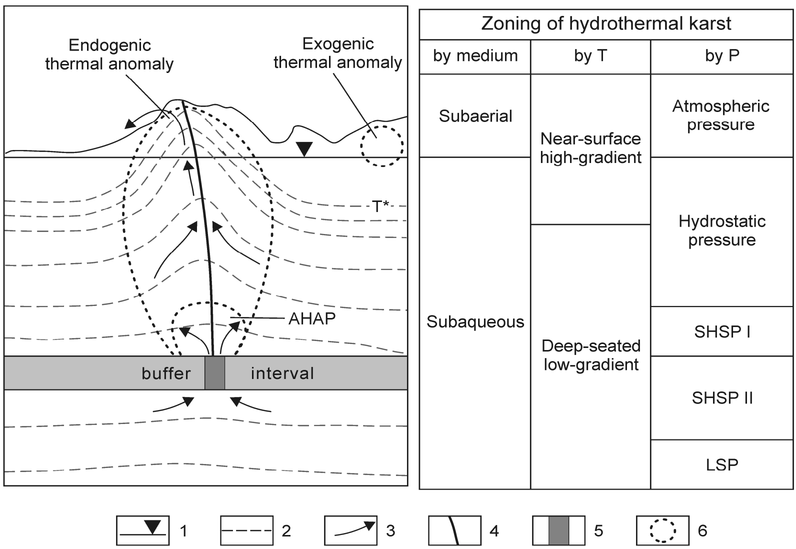

Fig. 4. Zoning of hydrothermal karst in the Earth’s crust: 1 – water table; 2 - isotherms; 3 – flow direction; 4 – axis of enhanced-permeability zone; 5 – hydrofracturing of the buffer interval; 6 – areas of "anomalous" hydrothermal karst and area of the anomalously high aquifer pressure (AHAP). T* - conventional minimum temperature of thermal water; SHSP I – suprahydrostatic pressure; SHSP II suprahydrostatic pressure exceeding the tensile strength of rock; LSP – lithostatic pressure.

Based on temperatures, hydrothermal karst may be subdivided into two broad zones (Dublyansky, 1997; 2000). The **near-surface high-gradient zone** exists where thermal waters occur close to the Earth's surface, creating elevated temperature gradients there. The anomalous hydrothermal karst defined above is restricted to this zone. In it, hypergene factors may also play an important role in the karst development. The **deep-seated zone** is located deeper beneath the Earth's surface. It is characterized by significantly smaller thermal gradients and is little affected by hypergene factors.

The following areas can be defined as **azonal** hydrothermal karst:

1. **Areas of hypogene thermal anomalies**. These are areas where waters carrying heat acquired in deep-seated settings ascend toward the Earth's surface as more or less concentrated, and rapid, flows. Most occurrences of modern, active hydrothermal karst fall into this category.

2. **Areas of local hypergene anomalies**. An example of such anomalies is the development of karst near sulfide ore bodies. Although waters taking part in this process are "normal" meteoric waters, they acquire enhanced aggressiveness and elevated temperatures through exothermic oxidation reactions. A characteristic feature of such karst is the descending character of water flow, which is generally not typical of hydrothermal karst.

In some situations hypergene waters can be significantly heated at or very close to the topographic surface. An interesting example was reported by Scherbakov (1960). Studying uranium-vanadium mineralization in Southern Fergana, he described zones of dissolution of clayey and carbon-rich schists: "*The process of dissolution was so intensive that, in places, only delicate quartz films remained, whereas the schist was entirely removed. … Taking into account the fact that in summer the entire rock* *surface, facing south, is heated to at least 70℃ and that the zone of constant temperatures of 15-17℃ lies at a depth no less than 3-4 m, we get, in the dry climate of Southern Fergana and in presence of ascending solutions, peculiar agglomerations of mineral bodies forming before our eyes and corresponding to the temperatures associated with hydrothermal processes*." (Scherbakov, 1960).

Possibly there could be other hypergenic mechanisms of heating. Anthropogenic activities might contribute significantly (e.g., by the discharge of heated industrial waters into karstified formations, the mining of sulfide ores, etc.).

The following zones may be defined on the basis of differing fluid pressures: - zone of atmospheric pressure; - zone of hydrostatic pressure; - zone of suprahydrostatic pressure not exceeding the strength of rocks; - zone of variable suprahydrostatic pressures exceeding the strength of rocks; and - zone of lithostatic pressure (see **Fig. 1**).

The zone of atmospheric pressure corresponds to zone of subaerial hydrothermal karst. The zone of hydrostatic pressure encompasses the upper 3-5 km of the Earth's crust. Hydrothermal karst associated with the hydrothermal ore deposits is mostly found in this zone. The suprahydrostatic pressure zones encompass the upper part of the exokarstic and all of the endokarst storey of the karstosphere.

Boundary conditions for karst genesis

Conditions suitable for karstification are not ubiquitous throughout the Earth's crust. They exist only in its upper part, where rocks may be dissolved by fluids in liquid or liquid+vapor states. In **Figure 5** the potential for karstification is plotted against temperature and pressure, the parameters which, together with the state of water and rock, define the aggressiveness of the former and the solubility of the latter. The state, in turn, defines the character of water flow and the mechanisms (processes) responsible for the permeability of rocks. **Figure 5** defines the principal fields in which the conditions of water and rocks differ drastically. For simplicity, the geothermal gradient is taken to be constant, and numerical values of depth and temperature are not assigned because, due to the variability of properties of the lithosphere they also vary across a broad range. Generally speaking, the scheme presented in **Fig. 5** encompasses the upper part of the Earth's crust (i.e., depths traditionally dealt with in karst studies) as well as zones of suprahydrostatic pressures up to depths of ca. 12-15 km.

The area including the field of cold karst and part of the hydrothermal karst field, bounded on the right by the liquid-vapor boundary and by the buffer interval at the bottom is the main subject area of hydrogeology. The waters here are virtually always in a liquid state, and their movement is described by well developed laws of hydrodynamics. Ground water flows in various modes – from viscous to laminar to turbulent. The permeability of rocks is defined primarily, by their porosity and fractures, the latter playing the leading role in karst development.

Discriminating between the cold (meteoric) karst and hydrothermal karst fields is somewhat ambiguous. By definition, hydrothermal karst is created by thermal waters. The term "thermal waters", however, is itself ambiguous. There are numerous criteria for defining thermal waters, which are used in different applications. In hydrogeological and karst studies the two methods most commonly applied are: (1) the method of fixed temperatures (i.e., some particular numerical value of the temperature is formally assigned the role of a boundary between cold (= ambient) and thermal waters), and (2) the method of climatic gradation; e.g., in the

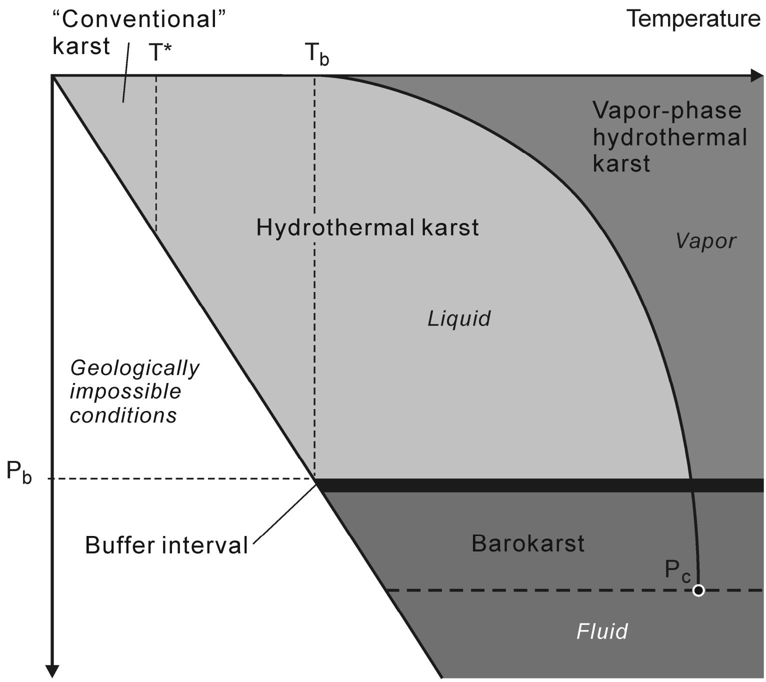

Fig. 5. Boundary condition of karst in the Earth’s crust: P_c – critical point for water; T* - conventional minimum temperature of thermal water; T_b – temperature at buffer interval (empirically determined; 80-100℃); P_b – pressure equal to the tensile strength of rocks.

European hydrogeological school a given water is considered thermal if its temperature exceeds the mean annual temperature of the area by ca. 4℃ (Schoeller, 1962).

Both approaches have their advantages and drawbacks. At the moment, however, we can simply accept that starting from some "boundary" temperature T^* the waters will be considered thermal. The upper boundary of the zone of hydrothermal karst will therefore be defined by either the isothermal surface, T^*, or the topographic surface in places where thermal waters discharge at the surface as ‘hot’ or ‘warm’ springs. According to regional geothermal studies, the upper surface of thermal waters (here taken as $T^* = 20℃$) lies at depths of less than 20 to 50 m in active volcanic areas, 500 m on platforms, and as deep as 1000 to 1500 m in areas of development of permafrost (Frolov, 1976).

Karst developing to the right of the two-phase equilibrium curve, the vapor-phase type of hydrothermal karst, is poorly studied. One example is the Larderello field in Tuscany (Shirone, 1954; Scherbakov, 1964), where karst is developed by the action of the complex vapor-gas mixture, H_2O-CO_2-B_2O_3-H_2S, the temperatures reach 200℃, and pressures reach 3 MPa. It can be assumed that the major role in karst dissolution is played by the condensate, not by the overheated vapor *per se.* The term "hydrothermal karst" remains therefore applicable. The mechanisms of this type of karstification are poorly understood. The movement of vapor-gas mix in the rocks is controlled by pore- and fracture permeability and can be described in terms of gas dynamics. Specific mechanisms of dissolution, transport and deposition of the matter are yet to be studied.

The impermeable buffer interval and all rock beneath it constitute the endokarstic storey of the karstosphere, where the lithostatic pressure exceeds the strength of rocks. There is presently no comprehensive theory describing fluid-rock interaction in these conditions. Only fragmentary data based on general physical principles and the limited empirical data obtained by deep drilling are available. The state of water within the endokarstic storey can be described as either liquid, or gas, or fluid. The lithostatic pressure is completely transmitted to water in the rocks. Any cavities in such an environment can only exist if they are filled with over-pressurized fluid. Flow is caused by the density difference between the fluid and the surrounding rock, as well as by pressure differences in different parts of the system. The overall direction of fluid movement is upward. It is discharged into the overlying exokarst storey by means of natural hydro-fracturing of the buffer interval. The minimum temperature beneath the buffer interval ranges between 80 and 100℃. The wa ters carry 2 to 3 times less dissolved matter compared to the lower part of the exokarstic storey, and are enriched in HCO_3 and CO_2.

Elucidation of the mechanisms and dynamics of the development of cavities within the endokarst storey remains a challenge. The major uncertainty here stems from the scarcity of data regarding the dynamics of fluids, as well as from the lack of quantitative models of fluid-rock interaction in such high thermal and baric conditions.

The upper thermal (and, respectively, the lower spatial) limits of karst development cannot be defined with certainty. At high temperatures and pressures the boundary between liquid and vapor states in water is "blurred". In this case we have to deal with a state conventionally called "fluid", which may be envisioned as a poly-component mixture of volatiles with the major components being water in a super-critical state plus the products of its thermal and electrolytic dissociation (Osnovy gidrogeologii, 1980-1983).

Chapter 3. The karstosphere

The concept of the karstosphere

The term ‘karstosphere’ was introduced by Maruashvili (1970), who defined it as a *spatially discontinuous layer comprised of the global areas of karstifiable rocks*. He suggested that karst processes are restricted to sedimentary rocks; the karstosphere, thus, is confined to the sedimentary crust and encompasses only some of its formations. It has the two storey architecture described above. The karstosphere is also defined as the *assemblage of those parts of the sedimentary cover composed of the readily soluble rocks that experience intensive chemical action of waters and which possess the complex of characteristics known as karst* (Maruashvili & Tintilozov, 1981; 1982). Maximovich (1979) broadly defined the karstosphere as *part of the lithosphere which serves as the arena for karst*. He extended the concept to include metamorphic and magmatic rocks (quartzite, marble, carbonatite, etc.). Andreychouk (1986), emphasized the dynamic character of the karstosphere and defined it as a *relatively continuous layer of the Earth’s crust, a geosystem characterized by the turnover of soluble rocks which are being formed primarily in the oceans and are being destroyed on the continents*.

It is now apparent that under the appropriate physical and chemical conditions, virtually any rock could become karstifiable. The definition of ‘karstosphere’ thus needs to be expanded. Here we define **karstosphere** as the *discontinuous cover encompassing the upper part of the Earth’s crust (to a depth of 12-15 km), within which karst processes have been active in the past, or are active at present, or could become active in the future*. Because water is the intrinsic “agent” for karst processes, the lower limit of the karstosphere approximately corresponds to the lower boundary for the existence of water in the liquid state (**Fig. 5**). Because the thermal and baric conditions in the Earth's crust are variable, the depth to the lower boundary of the karstosphere will also vary spatially. The karstosphere is an evolutionary phenomenon in the history of the lithosphere, intimately associated with the establishment of the subterranean hydrosphere.

The character of karst, the intensity of its development within the Earth's crust and on its surface, vary in response to the heterogeneous composition of the crust, topography, landscape-climatic conditions, as well as human activity. Despite some seemingly unique occurrences, the conditions needed for the development of karst anywhere have many common characteristics, such as lithology, mountainous or flat terrain, natural landscape zoning, the degree to which karstifiable rocks are covered by non-karstifiable ones, etc. These and other characteristics are used in the definition of types of karst and karstified terrains (**typology** of karst). An alternative way to reflecting the differences found in karst is **subdivision into regions**, i.e. territories within which specific types of karst are predominant. Although karst type and karst region are quite distinct notions, they both represent ways of generalizing concepts of the genetic specifics of karst within given territories or parts of the Earth's crust. Together they characterize the **karstogenetic settings**, a concept, however, that remains loosely defined.

The structure of the karstosphere, karstogenetic settings, and karstogenetic situations

The hydrodynamic zoning of the Earth's crust is planetary in scale. The exokarst and endokarst storeys defined within this zoning represent the highest-order components of the structure of the karstosphere. The differences between karst developing within these two storeys stem from the difference in energy sources (exogenic energy *vs.* endogenic energy), as well as from qualitative changes of the thermal, baric, hydrodynamic, and hydrogeochemical conditions occurring within crust with increasing depth. The conditions within the exokarstic and endokarstic storeys are heterogeneous, of course, yet certain characteristic properties can be defined (Ezhov et al., 1988).

The exokarstic storey represents the upper part of the karstosphere between the topographic surface and a depth of 0.5 to 7.0 km. The storey is characterized by the presence of meteoric infiltration and expelled (elision) waters having predominantly hydrostatic pressures and moderate (<80-100℃) temperatures. The karst waters discharge into river valleys, lakes and seas. The intensity of karst development decreases with depth within this storey.

The endokarstic storey encompasses the lower part of the karstosphere, at depths below 0.5 to 7.0 km. Karst processes act at pressures exceeding the hydrostatic and at elevated temperatures (>80-100℃), and involve metamorphic and igneous fluids acting upon the carbonate, sulfate, chloride, silicate, and aluminosilicate rocks. The discharge occurs in a pulsed manner through hydrofracturing of the buffer sub-zone of maximum lithostatic compaction of rocks that separates the endo- and exokarstic storeys. Although cavities developing within the endokarstic storey are typically small, the overall void capacity of the rocks here is increased relative to the lower part of the exokarst storey. Endokarstic reservoirs in this storey could accumulate hydrocarbons.

The zone of maximum lithostatic compaction of rocks serves as the boundary between the storeys of the karstosphere (Ezhov & Lysenin, 1987, 1990). This zone is a thin interval within the Earth's crust in which the rocks are under maximum compaction, their structural framework is significantly deformed, the pore- and fracture structure is destroyed, and their fluid holding capacity is drastically diminished. Cementation processes dominate over material removal processes. This narrow zone can be considered to be a distinctive transitional karstogenetic setting.

The Earth's crust is heterogeneous both vertically and laterally. The first-order subdivision is into continental and oceanic crust types, which possess different vertical and lateral construction, types of tectonic structures, patterns of tectonic movements, heat flows, etc.

Combining this model of global structure and the vertical hydrodynamic (baric) zoning, the Earth's crust can be subdivided into two storeys with four major components, defined here as the **principal karstogenetic settings**: (1) exokarstic continental; (2) exokarstic oceanic; (3) endokarstic continental; and (4) endokarstic oceanic (**Fig. 6**). The interfaces between these four settings, i.e., the buffer interval (sub-zone II-A) and the boundary between continents and oceans (the sea shore in its broad sense on the exokarstic storey, and the coupling between continental and oceanic crust in the endokarstic storey) represent distinctive settings in their own right and are defined as **transitional karstogenetic settings** (see **Fig. 6**).

To summarize, the **karstogenetic setting** is defined here as the complex of conditions and factors of karst development characteristic of the first-order blocks of the karstosphere: continental and oceanic exo- and endokarstosphere (principal settings) and their interfaces (transitional settings). The taxon "karstogenetic setting" is thus intermediate between taxons "karstosphere storey" and "karstogenetic situation".

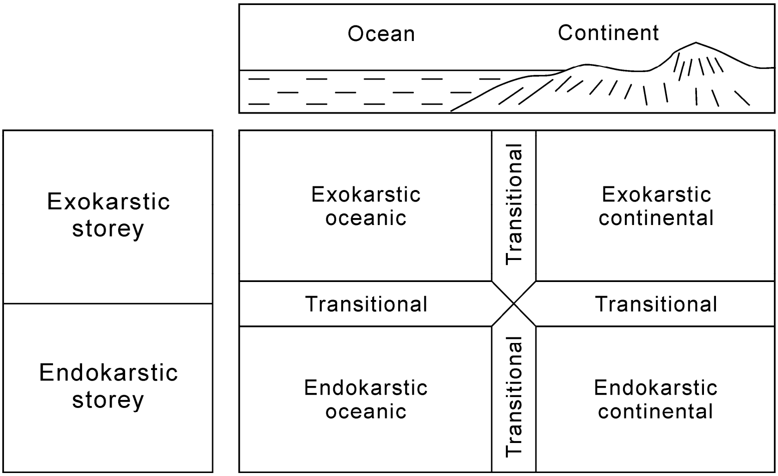

Fig. 6. Principal and transitional karstogenetic settings within the karstosphere.

Both principal and transitional zones are in turn heterogeneous in terms of the conditions and factors controlling the development of karst locally and regionally within them. Their diversity in the continental exokarstic setting has been well studied. It is in this setting that the numerous types of karst have been defined on the basis of the different combinations of conditions and factors (e.g., mountainous karst, platform karst, river karst, nival karst, watershed karst, permafrost karst, barren karst, etc.). The elucidation of the fine structure of this component of the karstosphere is beyond the scope of this work; we note only that all taxons of karst defined in the literature fit into one of the karstogenetic settings introduced above and correspond to specific **karstogenetic situations.**

Our empirical knowledge of the conditions of karst development in the other karstogenetic settings, principal and (particularly). transitional, is very limited. Relatively loose terms characterizing karst in the oceanic and the transitional (land-ocean) exokarstic settings can be found in the literature (e.g., sea karst, littoral karst, submarine karst, shoreline karst, karst of coral reefs, etc.). Information on karst in oceanic, continental and transitional settings within the endokarstic storey is insufficient to define any specific karstogenetic situations there.

Maximovich (1979) subdivided the karstosphere into three principal zones (storeys) that are different: exokarstosphere, mesokarstosphere and hypokarstosphere (see **Fig. 3**). The **exokarstosphere** is the uppermost storey, in which karst processes occur in the open mass transfer system and are associated with intense removal of material, development of surficial and subterranean forms, and carbonate sedimentation on the surface and in caves. The **mesokarstosphere** is the intermediate storey, corresponding to the geochemical zone of catagenesis and hydrodynamic zone of the impeded water circulation. The **hypokarstosphere** is the deepest storey, corresponding to the geochemical zone of catagenesis and the hydrodynamic zone of strongly impeded water circulation. Maximovich' zones correspond to the sub-zones of the hydrostatic zone (HSZ) proposed in this work (see **Fig. 3**).

Chapter 4. Non-traditional types of karst: specifics and relationships

Building upon the ideas regarding karst development in the Earth's crust discussed in the previous chapters, here we evaluate the genetic, spatial and logical relationships of karst in the broad sense and its non-traditional components.

Karst in the broad sense

In previous chapters we have argued that rocks of virtually any origin or lithology could become soluble under given conditions in the crust. If the two other essential conditions of karst development are met (i.e. the rock is permeable and there is fluid moving through it), the rock is subject to karstification, and its part of the Earth's crust becomes part of the karsto-sphere.

In our opinion, a genetic classification should be based on the sources of energy of the processes. None of the terms discussed above (i.e., hydrothermal karst, endokarst, etc.) appears to be appropriate as presently defined. The new scheme proposed is presented in **Table 1**. We use the most general and independent physical parameters, temperature and the pressure, to define the particularities of karstification within the different genetic types.

We introduce the term '**heterogeneous karst'** as a genetic category defining the development and subsequent infilling of cavities in rocks subject the joint forces of endogenic and exogenic energy; it is the product of the mixing of the endogenic and exogenic karst processes in space and time. Dissolution processes occurring in zones of horizontal flow of thermal waters, which commonly also include some waters from exogene sources

(Zaitsev, 1940; Kutyrev et al., 1989), and currently active near-surface hydrothermal karst fall into this category.

As is apparent in **Table 1**, the trio of terms: exogenic karst, heterogeneous karst, and endogenic karst encompasses all of karst (karst *sensu lato*). Below we use this scheme to discuss the relationships between the "non-traditional" types of karst described in Chapter 1 and karst as a whole.

Silicate karst

Silicate karst is specifically limited to the silicate rocks. In genetic terms, cavities in silicate rocks can be exokarstic, endokarstic, or heterogeneous in origin. In order for silicate karst to develop, besides the four principal factors, several additional geological conditions must be met, e.g., a long time available for karstification, low overall denudation rate competing with dissolution, etc. Favorable combinations of these conditions do not occur often.

Migration of silica is a relatively common phenomenon in hydrothermal karst because quartz is a typical gangue mineral deposited from thermal solutions. One of the controlling factors for the development of endogenic silicate karst is the concentration of flow to achieve dissolution. In the absence of focused flow paths, volumetric decompaction of rocks occurs instead of cave development. Hydrothermal karstification in silicate rocks result in some peculiar forms, as is apparent from the rare examples published in the literature (e.g., Dublyansky, 1990).

Table 1. Storeys of the karstosphere, genetic types of karst and the defining factors

Storeys of the karstosphere	Genesis (by energy)	Karst type (by temperature)	Karst type (by temperature and pressure)
Exokarstic (predominance of exogenic factors)	Exogenic karst	Ambient-temperature karst (ATK)	ATK at AP (subaerial) ATK at HP ("conventional" karst) ATK at SHSP (expelled formation waters, catagenesis, AHFP)
		Hydrothermal karst (HTK) of local thermal anomalies	HTK at AP (subaerial) HTK at HSP (local exogenic thermo-anomalies)
	Heterogeneous karst	Hydrothermal karst (HTK)	HTK at AP (subaerial) HTK at HSP (mixing zones) HTK at SHSP (expelled formation waters, catagenesis)
Endokarstic (predominance of endogenic factors)	Endogenic karst	Hydrothermal karst (HTK)	HTK at HSP HTK at SHSP (expelled formation waters, catagenesis) HTK at SHSP exceeding tensile strength of rocks (barokarst)

Notes. AP – atmospheric pressure; HSP – hydrostatic pressure; SHSP – suprahydrostatic pressure; AHFP – abnormally high formation pressure. The term *barokarst* corresponds to *endokarst* as defined by Ezhov, Lysenin, & Andreychouk (1988).

Hydrothermal karst and endokarst

Different researchers assign different meaning to the terms *endokarst* and *hydrothermal karst*. They are used as genetic categories, as spatial categories, and to define specific settings for karstification. It is apparent from **Table 1**, however, that ’hydrothermal karst’ should not carry any genetic connotation. The process is polygenetic; it may be related to endogenic and exogenic sources of energy, and can develop in both the endo- and exokarstic storeys.

The specific of hydrothermal karst is the involvement of thermal waters, whereas **specific for endokarst** (as defined by Ezhov, Lysenin, & Anderychouk, 1988) is karst solution/precipitation under pressures which exceed the tensile strength of the rocks. *Endokarst* can be substituted for *barokarst* in discussions pertaining to the origin or specific mechanisms of the processes. The term *endokarst* indicates the link between this type of karst and the endokarst storey with its particular energy sources.

Hypogene karst

According to the currently accepted definition of **hypogene karst**, its **specific** is the independence of recharge from the immediately overlying or adjoining areas of the open surface. In this context, the factors causing aggressiveness of water are not relevant. The term *hypogene karst*, therefore, applies equally to cave systems developed by hypogenic transverse speleogenesis (as discussed in Klimchouk, 2007) in ambient-temperature waters in artesian basins, and to caves developed at sites of localized discharge of high-temperature fluids. Logically, *endokarst* also falls into the broad category of hypogene karst.

Ore karst and ore-bearing karst

The **specific of ore-bearing karst** is the localization of ores and gangue minerals in karst cavities. Neither the origin of the cavities nor the lithology (chemistry) of rocks in which these cavities are developed is taken into account in this definition. Ore-bearing karst can result from: (1) Spatial superposition of two genetically independent processes, karstification followed by ore deposition; (2) Development of karstification and ore deposition as two consecutive (sometimes alternating) syngenetic stages of the hydrothermal processes; (3) Development of a single karst-ore process (residual; residual-infiltration; and infiltration groups of karst mineral/ore deposits; Tsikin, 1985). In the first case the term defines the phenomenon only, whereas in the second and third, both the phenomenon and the causative processes are included. These last two cases thus are better defined by term **ore karst**, whose **specific** is the *genetic unity (syngenetic character) of the processes of karst development and ore deposition*.

In terms of their origin, the ore-bearing karst cavities could be exogenic, endogenic, or heterogeneous. Therefore, all genetic classifications developed above for karst in the broad sense are applicable to ore karst. Despite its different origins, ore karst is known only within the exokarst storey at present. There is no information on the occurrence of ore bodies in karst cavities developing in the endokarstic storey.

Chapter 5. The position of karst among other geological processes

Introduction of the terms **hydrothermal karst**, **barokarst**, and **endokarstic setting** expands the meaning of the concepts of **karst** and the **karstosphere**. The latter now defines a *global zone in the Earth's crust within which certain specific alterations of matter occurs*. This requires definition of the logical relationships between karst (in the broad sense) and other processes of rock alteration such as weathering, metamorphism, etc.). The fields of development of these processes in the Earth's interior are shown in **Fig. 7**.

Regional *metamorphism* requires temperatures exceeding 300℃ (Turner & Verhoogen, 1951). *Weathering*, *diagenesis* and *catagenesis* proceed at lower temperatures. Melting of rocks (*anatexis*, *palingenesis*) begins at temperatures ranging from 500-600℃ (granitoid rocks rich in volatiles, particularly water) to 900-1000℃ (basic rocks with low volatile content). Further increase of temperature leads to *magmatism*. Water exists in the liquid state and rocks in the solid state primarily above the zone of the anatexis and palingenesis. Pospelov (1973) defined *metasomatism* as a "through" (overarching) process that operates, as a component, in all of the other processes of rock transformation.

Karst and metasomatism

Considering karst (*sensu lato*) can develop within the zones of weathering, diagenesis, catagenesis and metamorphism, it is tempting to parallel it with metasomatism. Definitions of the latter, however, vary. Traditionally, metasomatism is defined as a *type of metamorphism in the process of which the composition of the rock changes*. Processes similar to karst are excluded from this definition: "…*processes accompanied by … development of voids in the rock and their subsequent infilling do not belong to the class of metasomatic processes*" (Korzhinsky, 1953, p. 332).

The reason of this "rejection" of cavity formation lies in the traditional concept of metasomatism as that of the "point" and instantaneous act of dissolution-precipitation (Lindgren's metasomatism). The ambiguity of this concept has been demonstrated by Pospelov (1973). In the physico-chemical system of metasomatism he distinguished three zones: frontal, transitional and condensational: in case of simple dissolution these zones can be defined as zones of dissolution, transport, and deposition. To clarify the question, Pospelov proposed the concept of **extended metasomatism** in which operation of these zones is "stretched" in time or space or both. Extended metasomatism includes karst processes: "… *significant development of destructive processes may lead to the development of large caverns, karst-like cavities, and collapses of rock fragments* …" (Pospelov, 1973, p. 106).

To define one of the structural types of metasomatism, Pospelov introduced the term **intersomatism**: "Dissolution during metasomatism need not necessarily result in the development only of micro-pores. In the course of the extended form of metasomatism it may lead to the development of macro-pores or even chambers… As the dimensions of cavities increase, the processes of metasomatism grade into introsomatism and later – intersomatism … therefore, it appears reasonable to consider the processes of intersomatism, to be closely related to the volume-disequilibrium developments in metasomatism, locally resulting in early development of cavities as part of the process" (Pospelov, 1973, pp. 42-43). The quotations above demonstrate that the notion of karst may be encompassed entirely by the concept of intersomatosis. The karst process,

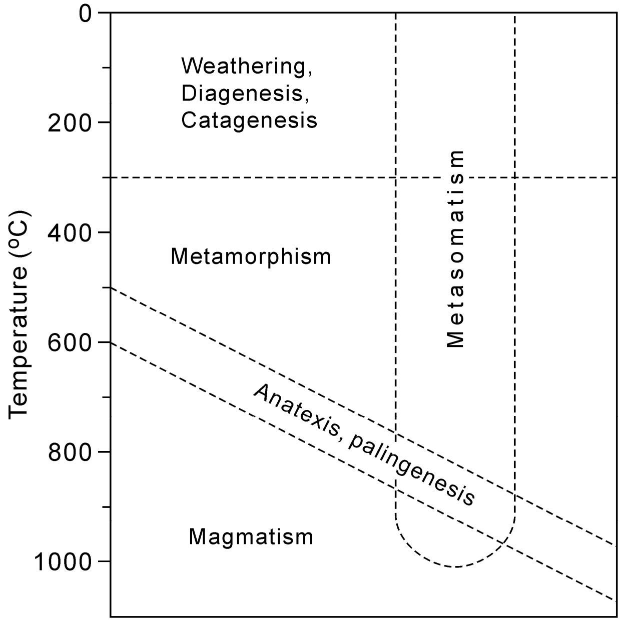

Fig. 7. Major processes of rock transformation in the Earth's crust (according to Eliseev, 1959; Turner & Verhoogen, 1951; and Pospelov, 1973).

thus, can be viewed as a component of the more general process of metasomatism as defined by Pospelov. It should be noted, however, that although deep-seated karst processes appear to be perfectly compatible with this conception of metasomatism, in hypergenic, near-surface karstification other processes that are not metasomatic may play quantitatively important roles; consider underground river erosion or biological activity for example.

Karst as a metasomatic process: characteristic features

Below we elaborate on the main features which define karst within the broad conception of metasomatism.

Formation of cavities (speleogenesis)

Initial dissolutonal development of cavities. This feature is common for karst and intersomatosis. In the earliest stages, cave development is mostly controlled by dissolution. Subsequently, the enlargement of caves can be assisted by other geomorphic processes such as collapse, mechanical erosion, etc. Apparently, karst is the only metasomatic process in which erosion may play a role.

Localization of cavities. Tsikin (1981) proposed the attribute of *localization* as a feature specific to karst. The succession of cave forms, with decreasing degree of localization, can be defined as e.g.: individual cavity – cave – cave system – karstic aquifer. It appears that the degree of localization generally decreases with depth. This is related to the increasing isotropy of the medium and the parameter fields (temperature, pressure, flow velocities, etc.). Because of that, development of individual large caves within the endokarst storey is possible only as an exception. Karst developing near the Earth's surface commonly displays the highest degrees of localization (e.g., large individual caves).

Dimensions of cavities. This topic is closely related to the preceding one. Karstic reservoirs have characteristic dimensions of cavities ranging between millimeters and tens of centimeters. These dimensions are within the range of macro-pores characteristic of metasomatism. These dimensions also define the lower limit for karst flow conductors; this concept is justified by the fact that the type of the function defining the movement of waters in such conductors remains unchangeable (Ford, 1988). The dimensions of individual karst caves and cave systems can surpass these values by large margins. Well-known examples are the largest subterranean cave chambers, Sarawak, Belize, Verna, and

Carlsbad Cavern (all in excess of 1 million m^3) as well as hundreds of known cave chambers of 100,000 to 500,000 m^3 (Ford, 1988).

The Involvement of allochthonous processes. This feature is most common in hypergene karst. Because such cavities are “open” systems, they are impacted by a plethora of the processes occurring on the Earth’s surface. Among processes not common in metasomatic systems are the actions of: - rapidly flowing water (mechanical erosion); - below-zero temperatures; - living organisms (including actions of man = technogenic impact) and others.

Infilling of cavities (speleolithogenesis)

Incomplete infilling (volumetric disequilibrium). This is not unique, but a very characteristic feature of karst. It is more typical in exogenic karst compared to endogenic karst. In exogenic karst during the stage of speleogenesis the dissolved matter is deposited outside the cave. With rare exceptions, the processes of destruction (dissolution) and transport dominate over the processes of deposition of matter, so that the volumes of cavities become far greater than the volumes of cave deposits. Reversal of this relationship is typically caused by superimposed, non-karstic processes. A characteristic example is the infilling of caves with ores during weathering of ore-rich rocks (Samama, 1986).

The important role of mechanical transport and sedimentation. This stems from the characteristic of cavity localization discussed earlier. Development of individual fluid channels leads to concentration of flow and, as a consequence, to the increase in the kinetic energy of the latter. In favorable conditions, channel flow can transport and deposit large amounts of solid matter. The presence of suspended particles in the flow increases its erosional impact.

The important role of allochthonous deposits. Because exogenic karst commonly opens onto the Earth's surface, many "exotic" deposits are commonly found in caves, e.g., cryogenic, aeolian, organogenic, and anthropogenic (technogenic) deposits.

Chapter 6. The cavernosity of rocks at great depths

Over the history of karstology, the lower boundary of karstification has been shifted progressively deeper as new facts regarding the development of karst became known. The paradigm of 1930s and 40s was that the depth of karst processes is limited by the level of river valleys cutting through a karst region. The level of the local hydrographic network was considered to be the “base level of karst” or “base of corrosion”. However, at the end of 1940s, new schemes of karst water zoning were put forth, which considered not only the local base level of erosion but also sea level (works of Z.A.Makeev,1948; F.A. Makarenko, 1947; D.S. Sokolov, 1947,1951, and 1962). Later, the entire sedimentary cover was identified as the arena of karst processes (Maruashvili, 1970), and the areas of igneous and metamorphic complexes in the Earth’s crust were included in the karstosphere (Maximovich, 1979). Most schemes of vertical zoning of karst waters encompass only the zone of hydrostatic pressures, which means that infiltration is assigned a leading role in recharge of karst waters. Although the classification of Maximovich (1979) introduced expelled formation waters as an agent of karstification in the lower part of the hyperkarstic zone, this is part of the zone of strongly impeded ground water circulation and, therefore, the processes there are controlled by exogenic energy (see **Fig. 3**).

Meanwhile, already in the 1960s, karst features in carbonate rocks containing fluids under suprahydrostatic pressure were identified, and novel schemes of vertical zoning of the subterranean hydrosphere that took into account the pressures of fluids were published by oil geologists (K.A. Anikeev, 1963-1964; Y.V. Mukhin, 1965).

Karst cavities more than 2 m in size in Devonian and Carboniferous limestone and dolomite were encountered at depths of 0.5 to 1.5 km on the western slope the Kizelovski basin and the eastern slope of the North-Urals oil and gas basin. These cavities commonly hosted secondary deposits (terrigenous, and chemogenic carbonate and sulfate). Drops of drilling tools by 3 to 4 m and intense losses of drilling mud were encountered at depths of several km when drilling carbonate strata underlying rock salt in the Peri-Caspian depression (Proshliakov et al., 1987). Abundant cavities and karstified zones were identified at a depth of ca. 3 km in limestone and dolomite of the pre-Jurassic basement in Western Siberia (Nizhne-tabaganskaya, Kalinovskaya, Gerasimovskaya, Maloichskaya and other zones; Zhuravlev, 1988).

Smaller karst forms (pores, caverns, solution-enlarged fractures) were detected in drill-hole cores retrieved from great depths in many petroliferous basins of the world. For example, in Southern Mangishlak, the Triassic limestone and dolomite of the middle volcanogenic-carbonate unit are intensely corroded at a depth of 4 km and below. Their effective cavernosity ranges between 7.4 and 21.5 % (mean 13 %; Timurziev, 1984). Enhanced porosity and cavernosity was reported for Middle Carboniferous and Early Permian limestones at depths of 4 to 5 km within many zones of the Peri-Caspian depression (e.g., Kenkijakskaya and Karagachskaya zones; effective porosity 10-14 %; permeability $6 \cdot 10^{-13}$ m^2; Maximov et al., 1984). A general review of the data from North America published by Maximov et al. (1984) reveals that Mesozoic and Paleozoic carbonate rocks may exhibit enhanced secondary porosity, up to 15-20 %, at depths up to 5-10 km.

Yakutseni (1984) reported that the effective capacity of carbonate reservoirs increases with depth (from 2 to 9 km) due to fracturation and karstification, rising by 2 to 9 % in Cambrian and Ordovician rocks, 13 to 18% in Devonian rocks, and 5 to 11% in Mesozoic rocks. Minsky (1979) argued that high-capacity reservoirs in carbonate rocks can develop at depths exceeding 7 km. Summarizing their extensive studies, Proshliakov et al. (1987) concluded that the dissolution of limestone and dolomite plays a leading role in the development of the filtration and accumulation properties of carbonate reservoirs at depths of 7-8 km.

The data presented above unequivocally indicate that karst can develop throughout the entire section of the Earth's crust that is accessible by drilling. The distribution of karst voids at different depths is irregular. Relatively large cavities, with dimensions exceeding meters, only occur in the top few kilometers. Smaller pores, vugs, and enlarged fractures prevail deeper. It is to be noted that this spatial heterogeneity can be exaggerated, at least partly, by the technical difficulties of identifying large cavities at great depths. Nevertheless, the available data from many regions of the world strongly suggest that the cavernosity of carbonate rocks within the 4 to 5 km depth interval is increased relative to that of the overlying units. An important point is that the deep-seated occurrences of karst in many regions (e.g., Southern Mangishlak, Peri-Caspian depression, petroliferous basins of the USA, etc.) are associated with suprahydrostatic pressures in the fluids. This means that waters of non-meteoric origin could have taken part in their development.

Various hypotheses have been put forth to explain these deep-seated karst features. Conventionally, they can be grouped into the *infiltrogenic*, *paleokarstic*, and *endokarstic* concepts.

The earliest, **infiltrogenic,** concept is based on the idea that the decrease of karstification with depth in carbonate rocks is an effect of the infiltration waters that penetrate deep into the Earth's interior. The paradigm of karst development formulated in early 1960s envisions the progressive slow-down of circulation and decrease in carbonate aggressiveness of water with depth, resulting in the overall decrease of the intensity of karstification (Sokolov, 1962). Karst cavities discovered by drilling at depths up to 1 km were explained by the very long-term impact on carbonate rocks of waters which moved exceedingly slowly along tectonically "weakened" zones of enhanced permeability. The aggressiveness of water is postulated to be a result of various geochemical processes which generate carbon dioxide.

The depth of penetration of infiltration waters into the Earth's crust is finite, so that below some given depth the infiltrogenic concept is no longer applicable. Obviously, this concept cannot be valid at depths where pressure is suprahydrostatic, because these zones represent impermeable barriers for gravity-driven infiltration waters.

The **paleokarstic concept** presumes that presence of karst features at any depth is explained by the burial of carbonate strata that were karstified at an earlier time in near-surface continental or island settings. It is believed that open karst cavities may be preserved even where pressure increases to lithostatic values, due to the high rigidity of much limestone and dolomite. Paleokarstic epochs can be inferred from hiatuses in carbonate sedimentation, by erosional unconformities and the intensely karstified intervals associated with them. Many tens of paleokarstic epochs have been identified in different regions of the continents, during which one or another carbonate formation experienced intense karstification by meteoric waters. For example, four major hiatuses were documented in drill logs for the carbonate rocks near the northern margin of the Peri-Caspian depression (pre-Bashkirian, pre-Moskovian, pre-Permian, and pre-Kungurian; Proshliakov et al., 1987). Each of these hiatuses is associated with a maximum interval of open porosity in the limestone and dolomite. It is to be noted that the first two hiatuses occur at a depth of 4 to 5 km.

Intense karstification of limestone and dolomite comprising the Triassic carbonate-volcanogenic complex at a depth exceeding 4 km in the Southern Mangishlak is interpreted as the product of the pre-Jurassic continental erosional epoch. However, enhanced cavernosity of the rocks of this complex was also noted in several zones within the central part of the Southern Mangishlak depression, at places where this complex is conformably overlain by upper-Triassic volcanogenic-terrigenous deposits (Rabinovich et al., 1985).

Limestone and dolomite of the basement of the West Siberian plate are intensely karstified at depths of 3 to 4 km. Karstification there is attributed to post-Hercynian (post-Variscan) peneplanation during the Triassic and, in places, middle- to late-Jurassic periods. Numerous caves, potholes, shafts, and smaller karst features were formed in Paleozoic carbonate strata, filled with secondary deposits, and subsequently buried under thick Meso-Cenozoic deposits (Zhuravlev, 1988).

An interesting hypothesis explaining the origin of deep-seated open karstic cavities was put forth by Turyshev (1965). On the basis of laboratory and field experimental studies he concluded that one possible means of creating the open, deep-seated cavities found in limestone and dolomite within the western slope of the Urals and the pre-Urals was the dissolution and removal of ancient gypsum infilling the cavities. According to drilling results, deep cavities in carbonate rocks within these regions are commonly filled with gypsum and anhydrite. Turyshev (1965) argues that groundwater loses its carbonate aggressiveness at shallower depth compared to its sulfate aggressiveness. As the water moves deeper into the Earth's crust, the latter can even increase due to a variety of geochemical processes that may remove the sulfate and calcium ions from solution (e.g., cationic exchange, crystallization, etc.). As a result, the ancient gypsum and anhydrite filling paleokarstic cavities can be removed preferentially by infiltrating waters.

The hypothesis of Turyshev is important in that it provides a viable chemical mechanism for karstification at any depth. Hydrodynamics (i.e., the impossibility of meteoric waters infiltrating into zones of suprahydrostatic pressures), however, imposes serious constraints on its applicability.

Recent advances in studies of hydrothermal activity and hydrocarbon potential at great depths have provided factual material for development of the **endokarstic concept,** which postulates dissolution by ascending f fluids of metamorphogenic and igneous origins (collectively defined as *endogenic*).

We believe that deep-seated karstification must be considered in the context of the vertical hydrodynamic zoning of the Earth’s crusts discussed above (Chapter 2). In the upper, exokarst storey, solution cavities can develop and exist for a (geologically) long time. Their existence does not depend on the pressure of fluid filling cavities; in fact, they can be preserved even when filled with air in the aeration zone. In contrast, in the lower, endokarst storey, cavities must be filled with fluid under suprahydrostatic pressure sufficient to prevent their collapse under the lithostatic load.

The character of fluid circulation and the evolution of cavernosity within the two storeys are illustrated in **Fig. 2**. The slowing of circulation and loss of water aggressiveness result in a general trend of decreasing intensity of karstification with depth in the upper, exokarst storey. Local and regional drainage of karst waters is controlled by local erosional entrenchment and by sea level. In the lower, endokarst storey, despite the prevalence of smaller cavities (pores, vugs, and enlarged fractures), the overall potential karst capacity of the carbonate rocks may be greater than in the lower parts of the exokarst storey that are characterized by regional groundwater drainage. There is variable development of karst in time and space in deep-seated rocks subject to the impact of upwelling fluids (global drainage). A similar structure for the subterranean hydrosphere, based on fluid dynamics, was proposed as early as the mid-1960s by V.F. Derpholz (1979), who introduced the concepts of circular and radial hydrodynamics.

The transitional zone separating the two storeys is characterized by complex dynamics in which physical and chemical conditions alternate between those favoring dissolution of rocks and those conducive for deposition of minerals. This area encompasses the upper part of the buffer interval (II-A in **Fig. 2**) and the lower part of the zone of sluggish circulation (I-C). Karst development in this zone is governed by pulses of highly pressurized fluids with low mineralization but saturated with CO_2 that break through the buffer interval, as well as by the variation of thermal and baric parameters (Ezhov & Lysenin, 1988; Ezhov et al., 1988).

A number of features promote karstification within the endokarst storey: - low mineralization (5-10 g/L and less) and the sodium-bicarbonate character of the upwelling fluids; - high contents of CO_2 (60-70 vol.% and greater); and - suprahydrostatic pressures which, prior to breakthrough, can reach lithostatic values.

Proshliakov et al. (1987) reported that in the upper 5 km of the Earth’s crust the pH of groundwater decreases from 8-9 to 5-6 (based on data from different regions of the former USSR). These authors argued that the decreasing pH provides one potential mechanism for preserving, and even enhancing, the reservoir properties of the carbonate rocks at great depths.

This hypothesis, however, has not been thoroughly tested.

pH measurements of fluids extracted from zones of suprahydrostatic pressures in a number of other petroliferous basins have found the fluids to be weakly alkaline to alkaline (pH≥7-8). For instance, in the Timan-Pechora basin the pH decreases from surface to a depth of ca. 3 km (from 7-8 to 4-5). At greater depth, however, the trend reverses (**Fig. 8**, **II**). Judging from the distribution of aquifer pressures and from calculations, the upper hydrodynamic storey in this basin has a thickness of 3 to 5 km. In the Volga-Ural basin, the pH tends to decrease to a depth of ca. 3 km, although the data set shows strong dispersion; **Fig. 8**, **I**. This trend is characteristic of the upper exokarstic storey. The South-Caspian depression is an area of alpine orogeny. There, a slight kink on the pH-depth graph can be observed at a depth of 1-2 km (**Fig. 8**, **III**) and is interpreted as the boundary between the exokarstic and endokarstic storeys. The data discussed here therefore do not support a universally acidic composition for fluids in the endokarstic storey but, equally, a predominance of alkaline conditions cannot be demonstrated unequivocally. In addition, it may be noted that numerous determinations of pH of hydrothermal fluids in areas of active volcanism yield both acidic and alkaline values.

Within most petroliferous regions the depth to the top of the endokarstic storey ranges between 3 and 5 km; it tend to be shallower in areas affected by the alpine orogeny. Its location is inferred from the appearance in drill cores of intervals of dense rock in which the open porosity is strongly reduced; beneath such intervals, suprahydrostatic pressures are typically encountered.

Let us evaluate the respective roles which endokarst and paleokarst could play in the development of deep-seated high-capacity reservoirs in carbonate rocks. Paleokarstic cavities, i.e., cavities formed in previous

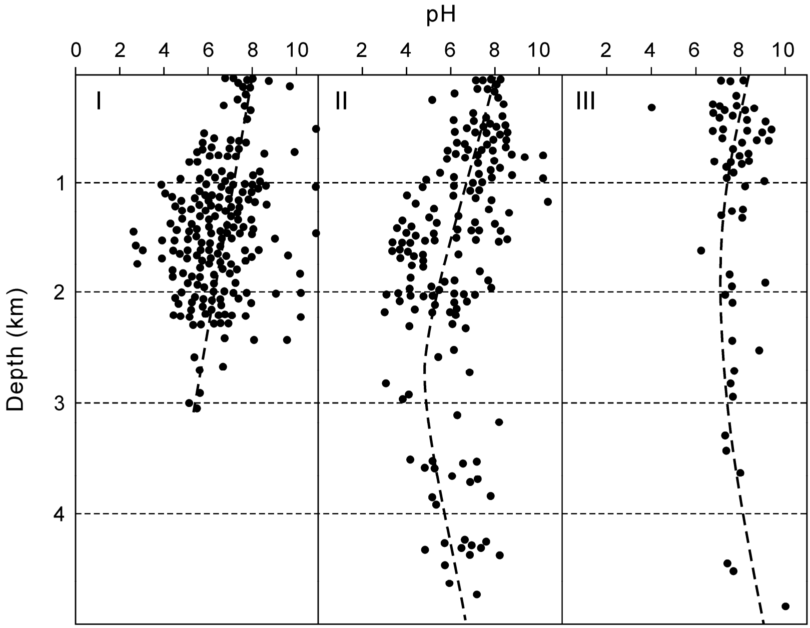

Fig. 8. The pH of ground waters in three petroliferous basins: (I) the Volga-Urals basin; (II) the Timan-Pechora basin; and (III) the South-Caspian basin. Compiled by Y. Ezhov from unpublished reposrts.

continental epochs in the exokarst environment, may be moved down into the endokarstic storey during subsequent tectonic subsidence of the region. They can be preserved only if they are filled with some secondary deposits. In the course of subsidence, the rocks and caves in them must pass through the buffer interval. At that time the rock experiences pressures which exceed their tensile strength, which means that any open voids in them will be closed, and any fluids contained therein will be expelled upward. Only if the caves are filled with solid material, can they "survive" this squashing effect of lithostatic pressure.

Having crossed the buffer interval, a cave filled with soluble mineral material (e.g. gypsum) could become open again due to dissolution and removal of the infilling by upward flowing, low-salinity fluids. The mechanism proposed here is analogous to that postulated by Turyshev (1965), the difference being that the Turyshev model applies to exokarstic settings, whereas in the present model it applies to the endokarst (barokarst).

The infilling, protecting the karst void from destruction during its passage through the buffer interval may also be of minerals which are only weakly soluble in the conditions prevailing in the exokarstic storey (e.g., silicates and aluminosilicates). In the conditions of the endokarstic storey, solubility of such minerals is increased, and becomes comparable to that of gypsum or anhydrite in near-surface conditions.

At moderate pressures calcite is more soluble in acidic solutions, whereas silica dissolves more readily in alkaline conditions. In the endokarstic storey, where high values of pH are common (see **Fig. 8**), the solubility of silica-rich rocks may exceed the solubility of the carbonate rocks; hence, silicate karstification may contribute to the development of high-capacity reservoirs.

We conclude therefore that paleokarstic cavities within the endokarst storey could acquire reservoir properties only if they are “rejuvenated” by the dissolution processes operating in the endokarst. The high-capacity reservoirs known in this storey, thus, owe their existence to the distinct attributes we have described in the endokarst.

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