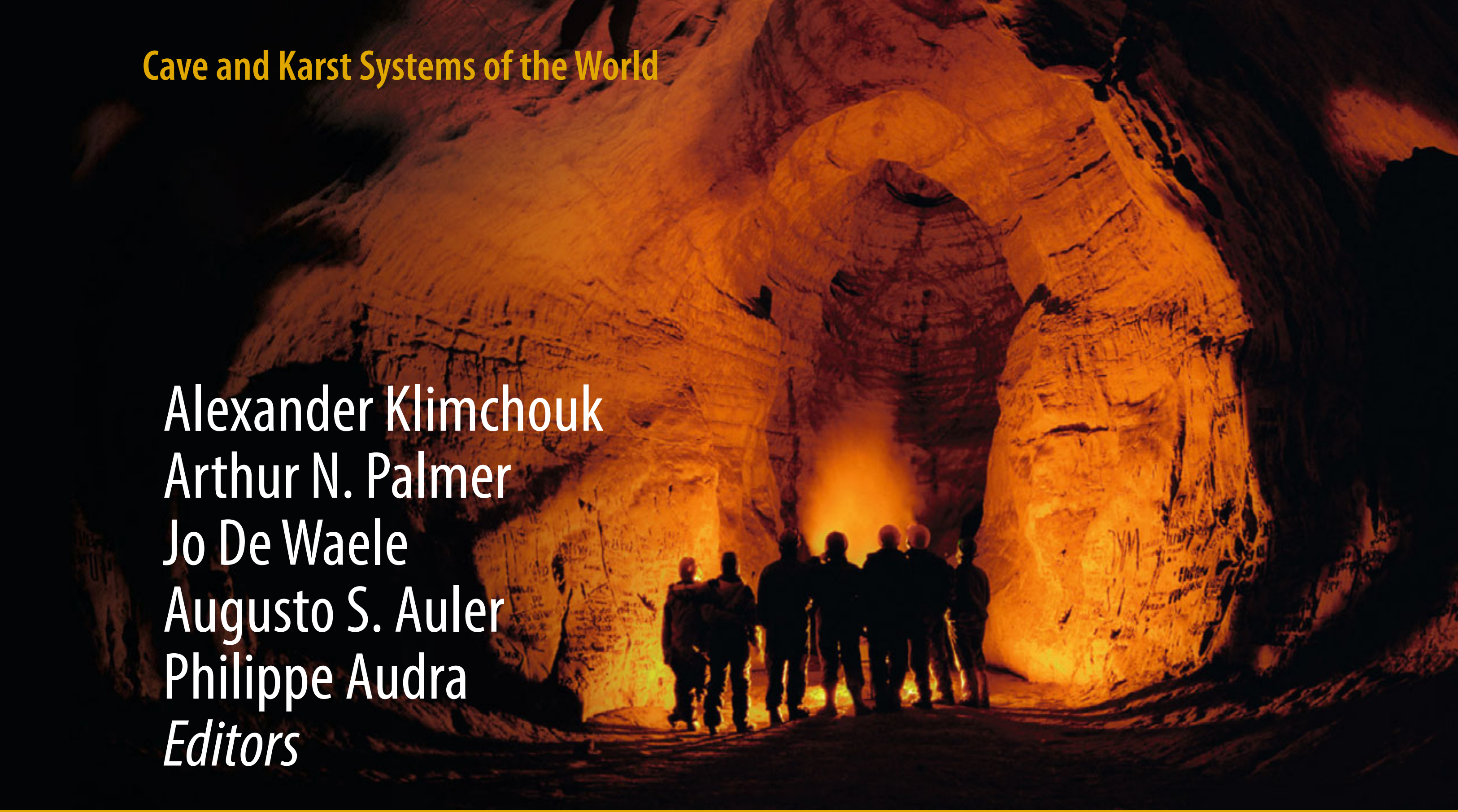

Cave and Karst Systems of the World

Alexander Klimchouk
Arthur N. Palmer
Jo De Waele
Augusto S. Auler
Philippe Audra
Editors

Hypogene Karst Regions and Caves of the World

Springer

Cave and Karst Systems of the World

Series editor

James W. LaMoreaux, Tuscaloosa, USA

More information about this series at http://www.springer.com/series/11987

Alexander Klimchouk
Arthur N. Palmer · Jo De Waele
Augusto S. Auler · Philippe Audra
Editors

Hypogene Karst Regions and Caves of the World

Springer

Editors
Alexander Klimchouk
Department of Hydrogeological Problems
Institute of Geological Sciences
Kiev
Ukraine

Arthur N. Palmer
Department of Earth and Atmospheric Sciences
State University of New York
Oneonta, NY
USA

Jo De Waele
Department of Biological, Geological, and Environmental Sciences
Universita di Bologna
Bologna
Italy

Augusto S. Auler
Instituto do Carste
Belo Horizonte, Minas Gerais
Brazil

Philippe Audra
University of Nice Sophia Antipolis
Sophia Antipolis Cedex
France

ISSN 2364-4591 ISSN 2364-4605 (electronic)
Cave and Karst Systems of the World
ISBN 978-3-319-53347-6 ISBN 978-3-319-53348-3 (eBook)
DOI 10.1007/978-3-319-53348-3

Library of Congress Control Number: 2017937725

Printed on acid-free paper

This Springer imprint is published by Springer Nature
The registered company is Springer International Publishing AG
The registered company address is: Gewerbestrasse 11, 6330 Cham, Switzerland

Preface

Ideas that karst can develop at depth without direct genetic relationship to the surface have a long history, but remained on the periphery of karstological thinking, not influencing the traditional paradigm of karst until the last 25 years. More attention to hypogene karst since 1990, and particularly the dramatic burst of studies in this field during the last decade, has changed our notion of hypogene karst from a curiosity to one of the fundamental categories of karst, at least of compatible importance with more familiar epigene karst.

Hypogene karst develops where the ascending flow of reactive fluids migrates across lithologies that are soluble in those particular fluids. Ascending fluid flow is ubiquitous in most of the upper crust in which deep-seated flow takes place. It is also a significant component of the circulation pattern in the uppermost zones of meteoric groundwater flow. A wide range of pressure–temperature conditions and the variety of compositions of fluids within the upper crust offer the potential for diverse dissolutional mechanisms to operate in a variety of lithologies. The depth limit for the development of dissolutional macro-porosity is difficult to establish, but available data suggest that it can form within depths of at least several kilometers. The time during which hypogene karstification may take place in deep-seated rocks is much greater than the common lifetimes for epigene karst systems in exposed formations. Hence, the potential for the development of hypogene karst is immense, not only in the continental domain but also in the oceanic domain.

The association of hypogene speleogenesis with ascending flow (leakage and discharge in confined fluid systems) was the major development that caused the recent burst in hypogene karst studies. It allowed (1) identifying the common genetic background and explaining the similarity of a large array of caves previously considered unrelated, which formed in different rock types by a variety of dissolutional mechanisms; and (2) interpreting hypogene karst in the context of regional hydrogeology and geodynamics. This triggered a dramatic expansion of regional and cave-specific studies and reinterpretation of many cases in light of the new conceptual framework.

Progress in regional studies has been reviewed in a series of conferences and their subsequent publications, such as "*Advances in Hypogene Karst Studies*" (edited by K. Stafford, L. Land, and G. Veni, 2009), "*Hypogene Speleogenesis and Karst Hydrogeology of Artesian Basins*" (edited by A. Klimchouk and D. Ford, 2009), "*Hypogene Cave Morphologies*" (edited by A. Klimchouk, I. Sasowsky, J. Mylroie, S.A. Engel, and A.S. Engel, 2014), and "*Origins, resources, and management of hypogene karst*" (edited by T. Chaves and P. Reehling, 2016).

This book was proposed as a next step in consolidating the growing wealth of regional data about hypogene karst. It is neither an inventory, nor a comprehensive coverage of all hypogene karst regions and caves of the world, but is rather a selection of regional and cave-specific case studies that represent a remarkable variability of relevant patterns and settings (geological, hydrogeological, tectonic, and geodynamic). In this way, it provides a solid reference for further generalizing and modeling studies of the topic, which may be the focus of a future collaborative volume on hypogene karst.

The book contains 61 chapters authored by 131 scholars from 25 nations and all continents. It starts with a chapter that reviews basic concepts about hypogene karst, speleogenesis, fluid dynamics, and hydrodynamic zoning of the upper crust, and outlines a pattern for classifying hypogene karst and its settings. Specific case studies are organized into four large geographic regions or continents. Although coverage is truly global, it is not uniform. Whereas 25 chapters are concerned with regions in Europe and 24 chapters deal with the North American regions, only 11 concern other parts of the world. This reflects the uneven distribution of research rather than scarcity of hypogene karst in underrepresented regions. On the basis of geological characteristics and fragmented reports of features scattered throughout petroleum and mining publications, it is evident that hypogene karst is widespread in many regions of Africa, Asia, Australia, and South America, although there are few focused studies. Moreover, even in Europe and North America, many areas have been recognized only recently to host hypogene karst, and its study is still ongoing. This means that next editions of volumes under this title will be needed.

Most contributions in this book deal with karst systems accessible to direct examination, which presently occur in the shallow subsurface. Most are relict systems, decoupled from their original genetic environments and brought into the shallow subsurface from considerable, sometimes large, depth. The advantages of direct observations and sampling, and of using methods developed in karst and cave science, make it possible to obtain unique information about patterns, processes, conditions, and controls of the origin of void-conduit systems at depth and about their hydraulic function. Many parameters of cave-forming environments can be reconstructed from mineralogical and geochemical footprints or inferred from other considerations, such as paleo-hydrogeological analysis. Studies of features that are analogues of deep-seated void-conduit systems are indispensable for interpreting data from drilling and geophysical surveys and for the development of conceptual models. Unfortunately, a wealth of information about the topic in mining areas and petroleum fields is difficult to obtain by karst scholars. Bridging the gap between karst science and industrial geology is a promising opportunity for further developments in hypogene karst studies.

Recognition of hypogene karst and the scale of its phenomena dramatically expand both the boundaries of karst and the significance of karst science far beyond the traditional, dominantly epigenic, karst paradigm. This has numerous scientific and practical implications. Hypogene karst studies hold a promise to help solve many problems in prospecting and exploration of deep petroleum, ore, and geothermal resources in soluble rocks. Proper reservoir characterization and modeling requires a skillful genetic interpretation of void-conduit systems and understanding of their hydraulic function. The role of hypogene karstification lies not only in enhancing reservoir properties but also in facilitating vertical fluid migration across heterogeneous strata. Therefore, it plays a dual role: It not only creates pathways for migration of hydrocarbons and metalliferous fluids to sites of deposition, but also contributes to the loss of the deposits by compromising the integrity of their seals. The latter aspect of hypogene karst also has important implications for the exploration of unconventional oil and gas resources and the sequestering of CO_2 and other troublesome fluids. The recognition of specific characteristics and functioning of hypogene karst is crucial for assessment and mitigation of environmental/engineering hazards, including sinkhole formation and groundwater flooding of mines.

The preparation of this book was initiated and coordinated by the Commission on Karst Hydrogeology and Speleogenesis of the International Union of Speleology, as a part of its ongoing HypoKarst project ("*Hypogene Karst & Speleogenesis: Nature, Processes, Mechanisms, Manifestations and Applications*"). The conference "DeepKarst 2016" held in Carlsbad, NM, organized and hosted by the National Cave and Karst Research Institute (NCKRI) has played an important role in the preparation of this book. Our special thanks to Margaret V. Palmer for her help in editing the chapters on North America.

The editors thank all contributing authors for their productive collaboration, as well as the many researchers and cave explorers who have documented caves and karst features in various parts of the world.

We hope that this book will stimulate further research into hypogene karst and caves around the globe, as well as interaction between karst scientists and industrial geologists.

Kiev, Ukraine	Alexander Klimchouk
Oneonta, USA	Arthur N. Palmer
Bologna, Italy	Jo De Waele
Belo Horizonte, Brazil	Augusto S. Auler
Sophia Antipolis Cedex, France	Philippe Audra

Contents

Contributors

Gennadiy Amelichev Crimean Federal University, Simferopol, Russia; Ukrainian Institute of Speleology and Karstology, Simferopol, Ukraine

Viacheslav Andreychouk Pope John Paul II State School of Higher Education, Biala Podlaska, Poland

Yemane Asmerom Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, NM, USA

Philippe Audra University of Nice Sophia Antipolis, Sophia Antipolis Cedex, France

Augusto S. Auler Instituto do Carste, Belo Horizonte, MG, Brazil

Pınar Avcı Department of Geological Engineering, Hydrogeological Engineering Section, Hacettepe University, Ankara, Turkey

Shirin Bahadorinia Karst Research Institute ZRC SAZU, Postojna, Slovenia

Fabrizio Balsamo NEXT—Natural and Experimental Tectonic Research Group, Department of Physics and Earth Sciences, University of Parma, Parma, Italy

Cinzia Banzato Department of Environment, Land and Infrastructure Engineering, Politecnico di Torino, Turin, Italy

Ingo Bauer Institut für Angewandte Geowissenschaften, TU-Darmstadt, Darmstadt, Germany

C. Serdar Bayarı Department of Geological Engineering, Hacettepe University, Ankara, Turkey

Pavel Bella State Nature Conservancy of the Slovak Republic, Slovak Caves Administration, Liptovský Mikuláš, Slovakia; Pedagogical Faculty, Department of Geography, Catholic University in Ružomberok, Ružomberok, Slovakia

Francisco Hilário R. Bezerra Geodynamics and Geophysics, Federal University of Rio Grande do Norte, Natal, Brazil

Jean-Yves Bigot Association Française de Karstologie (AFK), Montpellier, France

Kevin William Blackwood East Central University, Ada, OK, USA

Katalin Bolner-Takács Hungarian Speleological Society, Budapest, Hungary

Penelope Boston NASA Astrobiology Institute, Ames Research Ctr., Moffett Field, Mountain View, CA, USA

Van Brahana Department of Earth and Planetary Sciences, University of Arkansas, Fayetteville, AR, USA

Paul L. Broughton Calgary, AB, Canada

Paul A. Burger National Park Service, Anchorage, AK, USA

José María Calaforra Department of Biology and Geology, University of Almeria, Almería, Spain

Caroline L. Cazarin Centro de Pesquisa e Desenvolvimento Leopoldo Américo Miguez de Mello (Cenpes), Petrobras, Rio de Janeiro, Brazil

Matt Covington University of Arkansas, Fayetteville, AR, USA

Ilenia M. D'Angeli Italian Institute of Speleology, Department of Biological, Geological and Environmental Sciences, University of Bologna, Bologna, Italy

Donald G. Davis Denver, CO, USA

Nicole K. Davis Tucson, AZ, USA

Jo De Waele Italian Institute of Speleology, Department of Biological, Geological and Environmental Sciences, University of Bologna, Bologna, Italy

David D. Decker Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, NM, USA

Daniel H. Doctor U.S. Geological Survey, Reston, VA, USA

Virgil Drăguşin Emil Racoviță Institute of Speleology, Romanian Academy, Bucharest, Romania

Yuri Dublyansky Institute of Geology, University of Innsbruck, Innsbruck, Austria; Ukrainian Institute of Speleology and Karstology, Simferopol, Ukraine

Harvey R. DuChene Lake City, CO, USA

Renata Ennes-Silva Geodynamics and Geophysics, Federal University of Rio Grande do Norte, Natal, Brazil

Anita Erőss Department of Physical and Applied Geology, Eötvös Loránd University, Budapest, Hungary

Andrew R. Farrant British Geological Survey, Keyworth, Nottingham, UK

Jiří Filip Institute of Geology of the CAS v. v. i., Academy of Sciences of the Czech Republic, Prague, Czech Republic

Adriano Fiorucci Department of Environment, Land and Infrastructure Engineering, Politecnico di Torino, Turin, Italy

Lee J. Florea Indiana Geological Survey, Indiana University, Bloomington, IN, USA

Derek C. Ford Geography and Earth Sciences, McMaster University, Hamilton, ON, Canada

Joan J. Fornós Grup de Ciències de la Terra (Geologia i Paleontologia "Guillem Colom"), Universitat de les Illes Balears, Palma de Mallorca, Spain

Paolo Forti Italian Institute of Speleology, Department of Biological, Geological and Environmental Sciences, University of Bologna, Bologna, Italy

Amos Frumkin Cave Research Center, Institute of Earth Sciences, The Hebrew University of Jerusalem, Jerusalem, Israel

Ľudovít Gaál State Nature Conservancy of the Slovak Republic, Slovak Caves Administration, Liptovský Mikuláš, Slovakia

Sandro Galdenzi Jesi, Italy

Ermanno Galli Dipartimento di Scienze della Terra, University of Modena and Reggio Emilia, Modena, Italy

Marcus Gary Edwards Aquifer Authority, San Antonio, TX, USA

Fernando Gázquez Godwin Laboratory for Palaeoclimate Research, Department of Earth Sciences, University of Cambridge, Cambridge, UK; School of Earth and Environmental Sciences, University of St. Andrews, Fife, Scotland, UK

Angel Ginés Grup de Ciències de la Terra (Geologia i Paleontologia "Guillem Colom"), Universitat de les Illes Balears, Palma de Mallorca, Spain

Joaquín Ginés Grup de Ciències de la Terra (Geologia i Paleontologia "Guillem Colom"), Universitat de les Illes Balears, Palma de Mallorca, Spain

Stefan Glaser Krailling, Germany

Francesc Gràcia Grup de Ciències de la Terra (Geologia i Paleontologia "Guillem Colom"), Universitat de les Illes Balears, Palma de Mallorca, Spain

Michał Gradziński Institute of Geological Sciences, Jagiellonian University, Kraków, Poland

Jason D. Gulley School of Geosciences, University of South Florida, Tampa, FL, USA

İ. Noyan Güner Hydrogeological Research Unit, General Directorate of Mineral Research and Exploration, Ankara, Turkey

Tony Harrison Ivy Dene, Burtersett, Hawes, North Yorkshire, UK

Attila Hevesi Institute of Geography and Geoinformatics, University of Miskolc, Miskolc-Egyetemváros, Hungary

Carol A. Hill Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, NM, USA

Louise D. Hose Department of Geological Sciences and Engineering, University of Nevada, Reno, NV, USA

Jaroslav Hromas Cave Administration of the Czech Republic, Průhonice, Czech Republic

Daniel S. Jones BioTechnology Institute & Department of Earth Sciences, University of Minnesota, Minneapolis, MN, USA

O.I. Kadebskaya Mining Institute of the Ural Branch of the RAS, Perm, Russia

Stephan Kempe Institut für Angewandte Geowissenschaften, TU-Darmstadt, Darmstadt, Germany

Douglas W. Kirkland Carrollton, TX, USA

John Klenke Nye County Nuclear Waste Repository Project Office, Pahrump, NV, USA

Alexander Klimchouk Institute of Geological Sciences, National Academy of Sciences of Ukraine, Kiev, Ukraine; Ukrainian Institute of Speleology and Karstology, Simferopol, Ukraine

Sandor Kraus Hungarian Speleological Society, Budapest, Hungary

Boaz Langford Cave Research Center, Institute of Earth Sciences, The Hebrew University of Jerusalem, Jerusalem, Israel

Ladislav Lapčák University of Chemistry and Technology Prague, Prague 6, Czech Republic

Georgios Lazaridis Laboratory of Geology and Palaeontology, Department of Geology, Aristotle University of Thessaloniki, Thessaloniki, Greece

Szabolcs Leél-Őssy Department of Physical and Applied Geology, Eötvös Loránd University, Budapest, Hungary

José L. Llamusí Centre for Natural and Marine Environmental Studies, CENM-Naturaleza, Cartagena, Murcia, Spain

Frederick G. Luiszer Department of Geological Sciences, University of Colorado, Boulder, CO, USA

Vladimír Machovič University of Chemistry and Technology Prague, Prague 6, Czech Republic

Judit Mádl-Szőnyi Department of Physical and Applied Geology, Eötvös Loránd University, Budapest, Hungary

Giuliana Madonia Dipartimento di Scienze della Terra e del Mare, University of Palermo, Palermo, Italy

N.G. Maksimovich Institute of Natural Science of Perm State University, Perm, Russia

J. Martini Rue du Perbet, Saint Remèze, France

R. Mark Maslyn Littleton, CO, USA

Brian A. McMillan Tucson, AZ, USA

Marco Menichetti Department of Pure and Applied Sciences, University of Urbino, Urbino, Italy

Antoni Merino Grup de Ciències de la Terra (Geologia i Paleontologia "Guillem Colom"), Universitat de les Illes Balears, Palma de Mallorca, Spain

J. Michael Queen Carlsbad, NM, USA

Wasili Michajljow Foundation for Preservation and Studies of Caves, Bishkek, Kyrgyzstan

Joan Mylroie Department of Geosciences, Mississippi State University, Mississippi State, MS, USA

John Mylroie Department of Geosciences, Mississippi State University, Mississippi State, MS, USA

Angelo Naseddu Speleo Club Domusnovas, Domusnovas, Italy

Jean-Claude Nobécourt CRESPE, Vence, France

Rickard A. Olson Cave City, KY, USA

Bogdan P. Onac Karst Research Group, School of Geosciences, University of South Florida, Tampa, USA; Emil Racoviță Institute of Speleology, Romanian Academy, Cluj-Napoca, Romania

Wil Orndorff Virginia Natural Heritage Program, Christiansburg, VA, USA

R.A.L. Osborne Sydney School of Education and Social Work, The University of Sydney, Sydney, NSW, Australia

N. Nur Özyurt Department of Geological Engineering, Hydrogeological Engineering Section, Hacettepe University, Ankara, Turkey

Arthur N. Palmer Department of Earth and Atmospheric Sciences, State University of New York, Oneonta, NY, USA

Margaret V. Palmer Department of Earth and Atmospheric Sciences, State University of New York, Oneonta, NY, USA

Mario Parise Department of Earth and Geo-Environmental Sciences, University of Bari Aldo Moro, Bari, Italy

Emrah Pekkan Research Institute of Satellite and Space Sciences, Anadolu University, Eskisehir, Turkey

Lukas Plan Karst- und Höhlen-Arbeitsgruppe, Naturhistorisches Museum Wien, Vienna, Austria

Jason S. Polk Department of Geography and Geology, Western Kentucky University, Bowling Green, KY, USA

Erik Pollock Stable Isotope Lab, University of Arkansas, Fayetteville, AR, USA

Victor J. Polyak Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, NM, USA

Roi Porat Cave Research Center, Institute of Earth Sciences, The Hebrew University of Jerusalem, Jerusalem, Israel

Adriana Potra University of Arkansas, Fayetteville, AR, USA

Tomás Rodríguez-Estrella Department of Mining Engineering, Geology and Mapping, University of Cartagena, Cartagena, Spain

Andrés Ros Centre for Natural and Marine Environmental Studies, CENM-Naturaleza, Cartagena, Murcia, Spain

Laura Rosales-Lagarde Department of Physical and Life Sciences, Nevada State College, Henderson, NV, USA

Juan Sánchez Centre for Natural and Marine Environmental Studies, CENM-Naturaleza, Cartagena, Murcia, Spain

Geary M. Schindel Edwards Aquifer Authority, San Antonio, TX, USA

Tatiana A.R. Souza Instituto do Carste, Belo Horizonte, MG, Brazil

Michael Spilde Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, NM, USA

Christoph Spötl Institute of Geology, University of Innsbruck, Innsbruck, Austria

Kevin W. Stafford Department of Geology, Stephen F. Austin State University, Nacogdoches, TX, USA

Dominic Stratford Department of Archaeology, School of Geography, Archaeology and Environmental Studies, University of the Witwatersrand, Johannesburg, South Africa

Václav Suchý Praha 6, Czech Republic

Ivana Sýkorová Institute of Rock Structure and Mechanics v. v. i., Academy of Sciences of the Czech Republic, Prague 8, Czech Republic

Kinga Székely Institute for Nature Conservation, Budapest, Hungary

Penny M. Taylor Department of Geology, Mt. Holyoke College, South Hadley, MA, USA

Marjan Temovski Hertelendi Laboratory of Environmental Studies, Institute for Nuclear Research, Hungarian Academy of Sciences, Debrecen, Hungary

Rodney Tennyson Alpena, AR, USA

Levia A. Terrell Arcadis Engineering, East Syracuse, NY, USA

Jim Terry Toronto, AR, Canada

A. Koray Törk Cave and Karst Research Unit, General Directorate of Mineral Research and Exploration, Ankara, Turkey

Ádám Tóth Department of Physical and Applied Geology, Eötvös Loránd University, Budapest, Hungary

Andrzej Tyc Department of Geomorphology, University of Silesia, Sosnowiec, Poland

Elizaveta Tymokhina Crimean Federal University, Simferopol, Russia; Ukrainian Institute of Speleology and Karstology, Simferopol, Ukraine

Sam B. Upchurch SDII Global Corporation, Tampa, FL, USA

Hossein Karimi Vardanjani AbAtiPazhouh Consulting Engineers, Ahwaz, Iran

Marco Vattano Department of Earth and Marine Sciences, University of Palermo, Palermo, Italy

Bartolomeo Vigna Department of Environment, Land and Infrastructure Engineering, Politecnico di Torino, Turin, Italy

Jiří Zachariáš Faculty of Science, Charles University, Praha 2, Czech Republic

Types and Settings of Hypogene Karst

1

Alexander Klimchouk

Abstract

This chapter discusses the notion of hypogene karst, reviews its diversity and further develops the hydrogeological approach to classifying hypogene karst and its settings. Since an understanding of hypogene karst requires much deeper and broader hydrogeological and geodynamic context as compared to more familiar epigene karst, this chapter provides an overview of basic concepts about fluid dynamics and hydrodynamic zoning of the upper crust and about the influence of the mantle processes on crustal fluids. The relationships of hypogene karstification with metasomatism and other processes of fluid-induced transformations of rocks are examined. It is argued that the phenomena of the so-called ghost-rock karstification (commonly attributed to epigene settings) and cavernous decay (commonly attributed to external weathering) are manifestations of hypogene karstification and related alteration of rocks around conduits. Genetic categorization and discrimination of characteristic settings of hypogene karst are based on consideration of driving forces and conditions for fluid circulation and ascending flow in the upper crust in the context of tectonic/geodynamic positions and history of regions. Development and distribution of hypogene karst of the artesian type in gravitational flow systems of cratons are governed by the basin's configuration, topography and hydrostratigraphy. Hypogene karst of the endogenous type is governed by the geodynamic regimes and intimately related to cross-formational fluid-conducting systems. Hypogene karst is a significant component of fluid-induced lithogenesis and plays an important role in the porosity and permeability development in many sedimentary rocks and some metamorphic rocks.

Keywords

Hypogene karst • Deep hydrogeology • Geofluids • Karst and metasomatism • Hypogene karst types • Hypogene karst settings

1 Introduction

This chapter is aimed to clarify the notions of hypogene karst and speleogenesis, to review their diversity and outline approaches to reasonable classification of hypogene karst and settings of its development. The author believes that the problem of the origin of hypogene karst and of its further genetic subdivision should be approached from the hydrogeological perspective. Understanding of hypogene karst requires much deeper and broader hydrogeological and geodynamic context and commonly in much far-reaching retrospective than studies of more familiar epigene karst. Since hypogene karstification often occurs in greater depths than those commonly tackled in karst research and conventional hydrogeology, this chapter provides an overview of basic concepts about fluid dynamics and hydrodynamic

A. Klimchouk (✉)
Institute of Geological Sciences, National Academy of Sciences of Ukraine, Kiev, Ukraine
e-mail: klim@speleogenesis.info

A. Klimchouk et al. (eds.), *Hypogene Karst Regions and Caves of the World*,
Cave and Karst Systems of the World, DOI 10.1007/978-3-319-53348-3_1

zoning of the upper crust and about the influence of the mantle processes on crustal fluids.

The dissolution-dominated formation of macroscopic void-conduit systems at large depth is intimately related to other processes of fluid-induced transformations of rocks, particularly to metasomatism, which requires examining their relationships. This chapter will also demonstrate a genetic affinity of the phenomena of the so-called ghost-rock karstification and cavernous decay (commonly attributed to external weathering) and argue that both are manifestations of hypogene karstification.

Genetic categorization and distinguishing of characteristic settings of hypogene karst are derived from consideration of driving forces and conditions for fluid circulation and ascending flow in the upper crust in the context of tectonic/geodynamic positions and history of regions.

1.1 Genetic Types of Karst

The karst paradigm was changing during last decades, and the notion of karst has been defined from different perspectives (Klimchouk 2015). Karst was viewed as the totality of macroscopic dissolution phenomena in water-soluble rocks (a tradition traced back from E.-A. Martel), as terrain with distinctive hydrology and landforms (Ford and Williams 1989), as a geological environment (Huntoon 1995) or as a groundwater (fluid) flow system with specific properties (Worthington and Ford 2009; Klimchouk 2015). Some definitions equate karst to the process of dissolution, but reduction of a geological (hydrogeological) process to a chemical one is methodologically inappropriate and misleading. At the same time, karstification can be viewed as a complex process: of water–rock interaction (Zverev 1999), of metasomatic alteration of rocks (Ezhov et al. 1992), of hydrogeological mass transfer/mass transport (Klimchouk 2015), of destruction and obliteration of permeable soluble rocks (Sokolov 1962) or, more broadly, as a geological process (an interconnected set of processes) of transformation of soluble rocks by moving fluids with the dominant role of flow-dissolution feedback and respective self-organization of the groundwater flow system (Klimchouk 2015).

Regardless the approach, it was realized that most of the specific properties attributed to karst owe their origin to the development of organized dissolution porosity/permeability structures in soluble rocks (Ford and Williams 1989; Palmer 1991), i.e., karst is a function of speleogenesis (Klimchouk et al. 2000). Since speleogenesis is the primary mechanism of the formation of karst, genetic types of karst are to be distinguished based on types of speleogenesis (Klimchouk 2013b, 2015). Two fundamental types of speleogenesis, hypogene and epigene, are determined mainly by distinct hydrodynamic characteristics of the parent groundwater flow systems: (1) confined stratiform aquifer systems, or cross-formational fracture-vein systems, of varying depths and degrees of confinement, and (2) hydraulically open (unconfined), near-surface systems. They differentiate due to differences in hydraulic boundary conditions, geochemical and physical conditions of respective speleogenetic domains, hydrodynamic regimes of groundwater (fluid) flow and speleogenesis, and evolutionary trajectories of corresponding karst systems (Klimchouk 2015). Accordingly, two major genetic types of karst are distinguished within the upper part of the Earth's crust: hypogene and epigene. The respective karst systems differ in spatial distribution, the hydrogeological functioning, geomorphic expression, characteristics of void-conduit systems (their patterns, morphology, sediments, mineralogy, inhabitant biota, etc.) and related resources and hazards.

In contrast to epigene karst systems that develop in intimate interaction with the landscape and have both surface and underground components, hypogene karst evolves without direct genetic linkage with the surface, being originally represented exclusively by void-conduit systems at depth. When a hypogenically karstified rock formation is brought to the shallow subsurface by uplift and denudation, hypogene void-conduit systems are commonly relict (i.e., already decoupled from the original cave-forming environment). Although they can be intercepted by the denudation surface and become expressed in the landscape, the resultant geomorphic features are commonly destructive with respect to the hypogene karst system, i.e., they are not the inherent functional components of the latter (e.g., collapse features). In some cases, parts of hypogene karst systems can be overprinted by epigene karst systems, but the degree of their integration into the latter commonly remains limited due to the difference in driving forces, organization and functioning of respective parent flow systems. Since hypogene karst systems are initially comprised only of voids and conduits, the terms "hypogene karst (karstification)" and "hypogene speleogenesis" are frequently used interchangeably.

1.2 Definitions of Hypogene Speleogenesis

There are two different approaches on how to define hypogene speleogenesis. Palmer (2000) defined hypogenic caves as "*those formed by water in which the aggressiveness has been produced at depth beneath the surface, independent of surface or soil CO_2 or other near-surface acid sources.*" This approach, often termed "geochemical," emphasizes the place of the origin of the aggressiveness with respect to the surface. Another approach, known as "hydrogeological," is based on the acknowledgment that in any given geological environment the potential for dissolution and distribution of its effects, and hence localization, patterns, and morphology of the forming void-conduit systems, is determined largely by the regime, pattern and intensity of fluid flow, i.e., by hydrogeologic factors (Klimchouk 2007, 2015). With this approach, hypogene speleogenesis was defined as *"the formation of solution-enlarged permeability structures (void-conduit systems) by fluids that recharge the cavernous zone from below, driven by hydrostatic pressure or other sources of energy, independent of direct recharge from the overlying or immediately adjacent surface"* (Klimchouk 2007).

The above two approaches are not contradictory, but they determine slightly differing sets of caves and speleogenetic environments to be considered as hypogenic. The hydrogeological approach, though not specifying this directly, tacitly implies that the aggressiveness is produced at a depth below the cave-forming zone, whereas the rising flow is not necessarily implied by the geochemical definition. Based on the geochemical approach, Palmer (2007) places the artesian transverse cave development in evaporites into the realm of epigene speleogenesis, whereas cave development due to mixing at hydrochemical interfaces in unconfined aquifers, such as at the freshwater/saline water interface in coastal settings (Mylroye and Carew 1995), or the vadose/phreatic interface (Dreybrodt et al. 2009), is placed into the hypogenic category. Within the hydrogeological approach, the classifying of speleogenesis in these respective environments is the opposite.

The above-cited "hydrogeological" definition has been recently refined (Klimchouk 2015) by eliminating the non-specific and hence unnecessary requirement for upwelling fluids to be "*driven by hydrostatic pressure or other sources of energy*," and by adding more distinctness with regard to flow pattern and source of fluids. The primary criterion for distinguishing hypogene speleogenesis is that the cave-forming fluid comes from a hydrostratigraphically lower unit, although this unit is not necessarily the source aquifer. Cave development by flow uprising in the discharge segment of a phreatic conduit system (i.e., coming from below, relative to this segment), where flow is confined only by the conduit wall rock but recharge, through-flow and discharge occur within the same largely unconfined aquifer, should not be regarded as hypogene. With these reasons in mind, hypogene speleogenesis is redefined as *the formation of solution-enlarged permeability structures (void-conduit systems) by upwelling fluids that recharge the cavernous zone from hydrostratigraphically lower units, whereas fluids originate from distant, estranged (by low-permeability strata) or deep sources, independent of recharge from the overlying or immediately adjacent surface.*

The hydrogeological approach provides a theoretically and methodologically sound basis not only for defining and identifying hypogene speleogenesis but also for its further categorization and spatial and temporal prognosis in the context of global and regional hydrogeology and geodynamics.

1.3 Diversity of Hypogene Speleogenesis

Case studies included to this book demonstrate that hypogene speleogenesis operates in various geodynamic settings and geological/hydrogeological conditions, at varying depths (ranging from the shallow subsurface to several kilometers), and in rocks of different compositions (all kinds of carbonate rocks, evaporites, conglomerates, sandstones, quartzites, and even in igneous rocks) and ages (from Proterozoic to Pleistocene). Its distribution is not limited to continents—with the advent of new sensing technologies and expansion of offshore petroleum prospecting, evidence grow rapidly that hypogene karstification also occurs in the seafloor, especially in regions of active plate boundaries and hot spots (e.g., Betzler et al. 2011; Chen et al. 2015), although its proper identification is often hindered due to constraints of the traditional paradigm of karst (Michaud et al. 2005). Rising fluid flow migrates across lithologically diverse strata and aquifers, and upwelling is accompanied with changes in pressure and temperature. This causes disequilibrium conditions at different levels and supports diverse mineral reactions, including those resulting in the creation of macro-porosity. Hypogene karstification involves diverse dissolution mechanisms (Klimchouk 2012), operating either in combination or sequentially in time and space. The depth limit for hypogene speleogenesis is difficult to establish, but available data suggest that it operates within at least several kilometers.

In view of the broad variability of processes and conditions of hypogene speleogenesis and resultant void-conduit patterns, further subdivision into genetic types within this broad genetic category is needed, as well as a typification of settings of its development.

2 Hydrogeologic Context of Hypogene Speleogenesis

Whereas epigene karstification is associated mainly with meteoric flow systems in shallow hydraulically open unconfined settings, hypogene karst development commonly occurs at greater depths and is linked with ascending (discharge) components of flow systems of different nature and varying degrees of confinement. Thus, studying hypogene karst requires an understanding of much deeper and broader hydrogeological and geodynamic context, and commonly in much far-reaching retrospective.

The subject of "deep hydrogeology" (i.e., hydrogeology of those parts of basins and the metamorphic/crystalline basement that are below the uppermost interval routinely investigated for groundwater supply and geological engineering purposes—commonly below several hundred meters) is still controversial and ill-understood. Fundamental modern reviews performed from the essentially hydrogeological perspective are rare (e.g., Djunin 2000; Ague 2003; Djunin and Korzun 2005), although there are excellent reviews of geochemistry of deep aqueous fluids in sedimentary basins (e.g., Kharaka and Hanor 2004) and basement rocks (Frape et al. 2004; Bucher and Stober 2010). Deep fluid systems receive in-depth treatment in studies on the role of groundwater in geologic processes (e.g., Ingebritsen et al. 2006). Important progress has been made during last decades due to the intense quest for deep petroleum reserves and developments in the reservoir analysis (e.g., Kirkinskaya and Smekhov 1981; Cubitt et al. 2004; Shepherd 2009; Chilingar et al. 2005; Agar and Geiger 2014; Cathles and Adams 2005). Much of our understanding of deep basinal and crustal fluids has come from studies in sedimentology (diagenesis, basin's evolution; e.g., Kyser and Hiatt 2003; Galloway and Hobday 1996; Bjørlykke 1993; Warren 2006), petrology (metamorphism and metasomatism; e.g., Fyfe et al. 1978; Shmulovich et al. 1994; Ague 2003; Kissin 2009; Harlov and Austrheim 2013; Yardley 2013) and ore geology (e.g., Cathles and Adams 2005; Ingerbitsen and Appold 2012), particularly from geochemical/thermodynamic modeling and fluid inclusions. These studies, however, are more concerned with pervasive fluid fluxes. Channelized (focused) fluid flow remains much less studied, although researches of ore-forming systems provide some important insights on it (Ingerbitsen and Appold 2012). Evidence of massive fluid flow in the deep portions of the upper crust come from studies of metamorphic and metasomatic processes (e.g., Harlov and Austrheim 2013), particularly of fluid-dominated infilling of secondary porosity (including ore formation) and alteration of the original rock. High time-integrated fluid fluxes are inferred in many thermal aureoles and regional metamorphic belts (Rubenach 2013) and subduction zones (Bebout 2013; Schmidt and Poli 2014). Inhomogeneities in fluid distribution in both the lithospheric mantle and deep to middle crust are recorded by geophysical techniques (Kissin 2009; Unsworth and Rondenay 2013).

2.1 Origin of Aqueous Fluids

Free (i.e., mobile and available to flow) aqueous fluids in the crust are ultimately derived from atmospheric water, seawater, water that is physically and chemically bound in sediments and from devolatilization of the middle-lower crust and the mantle. In hydrogeology, the origin of aqueous fluids is commonly distinguished according to ways by which they penetrate into the lithosphere, or to processes by which they are released or generated within it (Fig. 1).

Meteoric water is the water derived from atmospheric precipitation and infiltrated from the surface. *Connate water* is the water (commonly seawater) trapped in the pores of a rock at the time of their formation. Waters of these two origins dominate in the uppermost portion of the crust.

Metamorphic (thermogenic) fluids originate as the result of devolatilisation reactions in sedimentary or magmatic rocks in course of metamorphism, which occurs around magma chambers or at large depth. *Magmatic (juvenile) fluids* are released during crystallization or solidification of magma within the crust (also called *volcanic fluids*) or originate from the mantle and core degassing (called by some *trans-magmatic fluids*; Pinneker 1983).

Meteoric and connate waters are considered exogenic (meteoric waters have also been termed epigenetic; Crossey et al. 2006), whereas metamorphic and magmatic origins of fluids can be treated as endogenic (e.g., Pinneker 1983; Crossey et al. 2006). Where rising to and across the sedimentary cover, metamorphic and trans-magmatic (mantle input) fluids are inevitably mixed in proportions which are often difficult to determine, so that referring to them as to endogenic fluids is convenient (Crossey et al. 2006). The term "basinal waters (fluids)" is often used for deep fluids that have originated in a basin, i.e., expelled connate water and metamorphic water.

Aqueous fluids in varying physical states constantly circulate in the crust, either directly or via the processes of the geologic cycle (Fig. 1), and may retain in the subsurface continuously during several successive hydrogeological cycles (i.e., periods starting with subsidence and marine sedimentation, continuing through uplift, denudation and exposure and closing with the new marine transgression; Kartsev et al. 1969). This leads to ubiquitous mixing of fluids of different origins. Meteoric waters can be found in the relatively pure (unmixed) form in uppermost parts of the crust, well flushed by the meteoric regime (particularly in recharge segments of groundwater flow systems), and connate waters may overwhelmingly dominate in young subsiding basins. Upon their generation, metamorphic fluids immediately mix with preexisting pore fluids, as well as with magmatic fluids, if any. Endogenic fluids (metamorphic and magmatic) mix with formation waters during upflow across sedimentary successions. The term "formation waters (fluids)" is commonly used to underscore genetic uncertainty of waters in deep parts of basins, especially mature ones, which are almost universally of mixed origins (Kharaka and Hanor 2004).

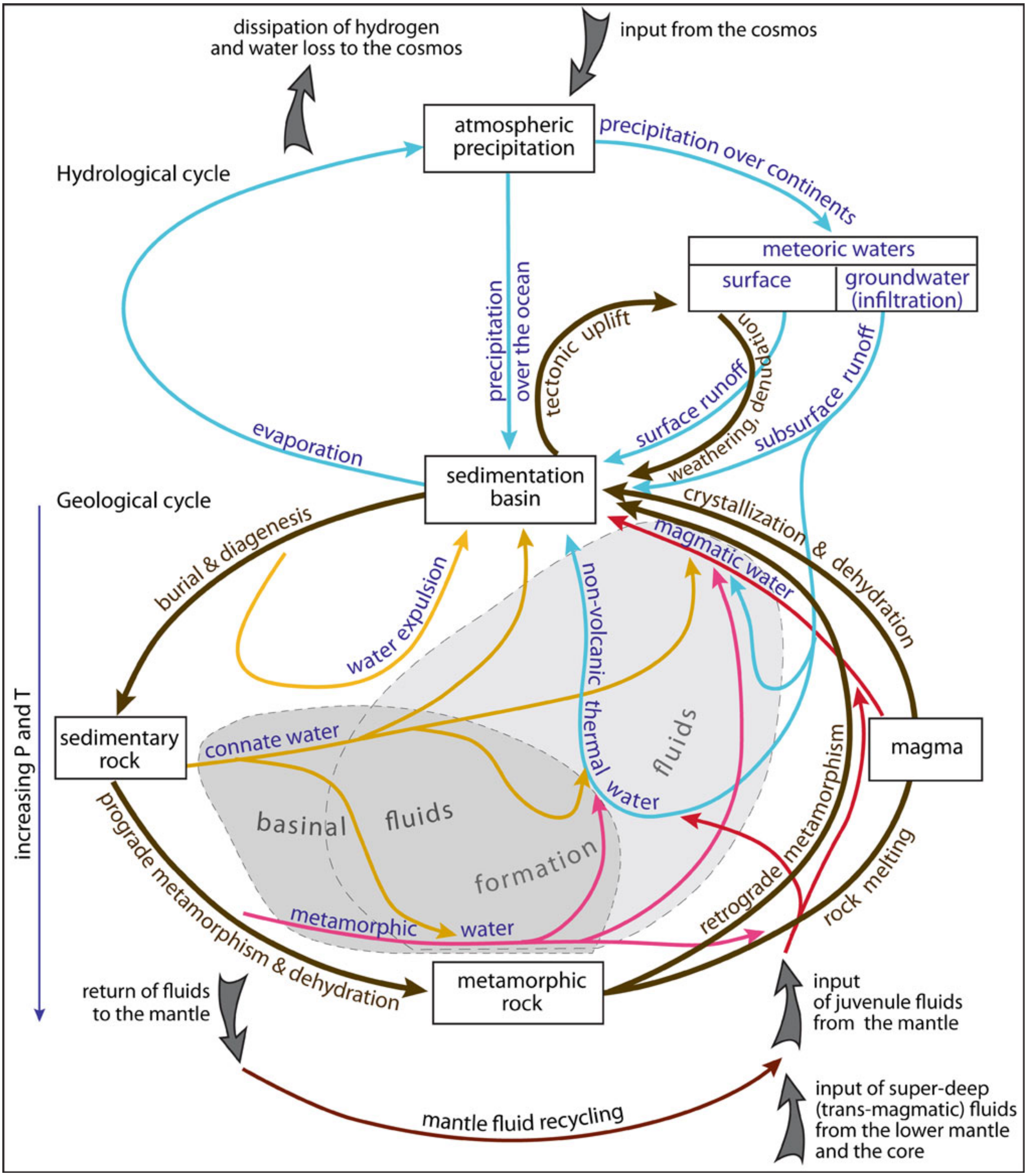

Fig. 1 Relations between the hydrologic and geological water cycles and genetic paths of fluids in the crust. *Thicker dark brown arrows* denote the petrological cycle. *Thinner arrows* of various colors denote genetic and mixing paths of fluids in the crust. The overall configuration of the cycles and their compartments follow Pinneker (1980), but the nomenclature and relationships of fluids are severely modified

2.2 Flow Regimes

Different fluid flow regimes occur in the upper crust, determined by variations in the nature and magnitude of pressure and the degree of hydrodynamic confinement (Kissin 1985; Bjørlykke 1993; Galloway and Hobday 1996; Shvartzev 1996; Deming 2001; Hiscock and Bense 2014). Importantly, the notion of a flow regime (the source and realization of fluid motion) should not be confused with the origin of the fluid itself.

2.2.1 Compactional Regime

In young subsiding basins, the dominant flow drive in progressively buried strata is the pressure head generated by compaction due to increasing lithostatic loading, which requires an expulsion of pore fluids (connate and meteoric) from the sediments. The compactional regime is also termed expulsion, elision and exfiltration regime. Flow is directed generally upward and outward, on the regional scale—from areas of greatest subsidence to basin margins. Varying susceptibility of sediments to compaction and pressure solution, as well as to chemical water–rock interactions, leads to increasing heterogeneity in porosity and permeability. Where low-permeability confining formations are present above compacting sediments, significant overpressure can develop. When tectonism is imposed, the compactional regime can be also generated by tectonic compression in the vicinity of collision and uplift areas. Compaction-driven fluid flow is complex because it is inseparable from rock deformation and because the hydraulic properties that limit fluid flow through the rock matrix, such as permeability and porosity, are dynamic (Connolly and Podladchikov 2013).

The compactional regime in mature sedimentary basins is unlikely to drive considerable fluid fluxes due to its transitional nature (Deming 1994; Djunin 2000; Djunin and Korzun 2005) and the loss of most of the connate water during earlier stages of the basin development. It is considered of limited importance for hypogene speleogenesis (Klimchouk 2013b). However, it can be important during the early stages of burial (during eogenesis and the transition to meso-genesis, during and immediately after progressive burial), especially in sequences containing alternating sediments with contrasting susceptibility to compaction, from which water is expelled at drastically different rates. It can be particularly important in creating cross-formational permeability structures in upper, still poorly indurated successions due to the focused expulsion of overpressured fluids from the underlying strata. The fluid-focusing role of such early structures can be retained or renewed during the later stages of diagenesis. The abundance of such structures, well expressed as seafloor pockmarks and expulsion chimneys beneath them, is shown by numerous studies during the last decades (e.g., Hovland et al. 2002; Hovland 2003; Cartwright and Santamarina 2015).

2.2.2 Thermobaric Regime

With still deeper burial and further rise of temperature and lithostatic loading, the thermobaric regime develops, in which the fluid pressures are caused mainly by the thermal expansion of water and/or by the release of water by mineral dehydration in a low-permeability environment. This regime underlies the compactional regime and commonly mingles with it. Thermobaric flow is commonly highly overpressured and directed generally upward. The thermobaric regime can be initiated by the deep endogenous regime, with which it mingles at greater depth.

2.2.3 Deep Endogenous Regime

Aqueous fluids originating from metamorphism and devolatilization in the lower crust, as well as magmatic/juvenile fluids released from the mantle and the core (superdeep fluids; Letnikov 1992, 2001; Lukin 2014, 2015), can intrude up into the upper crust and sedimentary cover, introducing anomalously high energy. These intrusions contribute to the thermobaric regime in the deepest portions of sedimentary basins and cause strong anomalies in parameters of thermal, baric, hydrochemical, geochemical and permeability fields.

Upwelling of endogenous fluids through the upper crust is a manifestation of the global Earth degassing process. Letnikov (1992) emphasized the distinction between two branches of the Earth degassing:

1. Monotonously diminishing in time, whole-planet degassing of the upper part of the lithosphere, resulted in its depletion in fluids, and in lowering of the fluid front to depth. The energy potential of the deep lithospheric fluid systems declined since the Archean through the Cenozoic. The character of the distribution of these systems changed in time from continuous in the Archean, through clustered in the Proterozoic and linear in the Phanerozoic, to discrete in the Cenozoic.
2. On the background of the general lithospheric degassing—a pulsed degassing of the Earth liquid core via mantle plumes. The plume-related degassing occurs periodically and impacts certain areas varying in size, during varying periods of time.

Deep-rooted mantle plumes (superplumes) have been the subjects of numerous studies (e.g., Larson 1991; Letnikov, 1992; Condie 2001; Ernst and Buchan 2001; Maruyama et al. 2007; Yuen et al. 2007) and vigorous debates. Recent work of French and Romanowicz (2015) provided robust evidence, based on an advanced seismic tomography technique and a global whole-mantle shear-wave velocity model, for large, vertically continuous, low-velocity columns in the lower mantle beneath many prominent hot spots. The imaged trans-mantle conduits, rooted at the mantle-core boundary, are broader (typically 600–800 km) than commonly invoked thermal plume tails, which suggests that they are long-lived and have a thermochemical origin (French and Romanowicz 2015).

Superplumes are thought to be composed mainly of reduced gasses, primarily by hydrogen, and they possess tremendous energy potential when departing from the mantle-core boundary (pressure >1300 kbar, temperature >4000 °C, hydrogen enthalpy 1200–1000 kJ g^{-1}; Letnikov and Dorogokupets 2001). In the course of upwelling through the mantle, their thermal energy is renewed by exothermic reactions of hydrogen and reduced gasses with oxygen-bearing mantle minerals, so that superplumes are capable of reaching the crust without major loss of thermal energy (Letnikov 1992). Additionally, superplumes may assimilate water from the top lower mantle, where its large reserves are now inferred from petrologic modeling, laboratory experiments and seismic data processing (Murakami et al. 2002; Schmandt et al. 2014). When reaching the core, superplumes give rise to its tectonic activation, magmatism, metamorphism and the origin of various fluid systems, among other geologic consequences (Larson 1991; Condie 2001). The area of invasion of superplumes into the lithosphere can be as large as of many thousands of km^2. The evolution of superplumes is accompanied by their structural differentiation into derivative plumes varying in sizes, thermobaric regimes, geodynamic and fluid-dynamic activity, that determines variations in their lithogeodynamic impact on particular basins (Lukin 2014, 2015). Breaking through the brittle/ductile transition zone, plume-derived fluid systems determine the deep endogenous regime in the upper crust, with fluids propagating to and across the sedimentary cover via magmatic columns or deep-rooted faults. Invasion of deep endogenous fluids gives rise to various transformations of sedimentary rocks, including diffuse and infiltration (focused) metasomatism (Korzhinskiy 1957; Zharikov et al. 2007), and fracture and dissolutional porosity creation and infilling (Ezhov et al. 1992; Klimchouk 2012).

Kropotkin (1986) introduced a useful notion of "degassing tubes" (chimneys)—sub-vertical contours across the sedimentary succession that embrace manifestations of deeply derived gasses, zones of abnormally high pressures and geochemical and thermal anomalies related to invasion of deep gas–vapor fluids (including hydrocarbons) and their vertical migration. Lukin (2015) expanded the concept of degassing tubes to include multifarious alterations of rocks due to endogenous metasomatism induced by rising fluids of the deep origin. Kropotkin (1986) distinguished two branches of degassing: *"hot degassing,"* where fluids migrate through the crust along with magma, or across high-temperature zones in the vicinity of intrusions, and "*cold degassing*," where rising reduced fluids do not encounter magmatic bodies and other strongly heated rock masses.

The deep endogenous regime has great potential to support endogenous hypogene speleogenesis in various rocks, including siliciclastic, in deep parts of basins affected by endogenic fluids.

2.2.4 Meteoric Regime

The meteoric regime (syn. gravitational, infiltration, topography-driven regime) evolves with the beginning of a continental exposure and occupies the uppermost parts of the crust. Infiltrated meteoric waters, driven by gravitational head imposed by differences in the topography of the water table (and ultimately of the land surface), increasingly flush out the formation waters from basins. The meteoric groundwater circulation can penetrate to depths of over 3 km affecting deeply buried strata in geologically and topographically favorable conditions, especially in tilted and faulted basins adjacent to mountainous areas, and in intermountain basins. Localized freshwater discharges at depths of several hundred meters below the sea level are documented in many coastal regions, and meteoric groundwater flow was found by ocean drilling far offshore. During prolonged exposure periods, the meteoric regime progressively substitutes the compactional regime in the uppermost crust, although the compactional and other endogenous regimes may still predominate in deep environments. Patterns and dynamics of groundwater flow driven by this regime are discussed in the context of the upper hydrogeologic story in Sect. 2.4.1.

2.2.5 Other Flow Drives

Other significant drivers for flow in the deep parts of basins include convection induced by thermal or salinity gradients (variable density flow) and seismic pumping driven by dilation and contraction in seismically active areas. The convection regime is favored where pronounced density inversion with depth (due to heating and sometimes decreasing salinity) is combined with high vertical permeability across large thicknesses (permeable relatively homogenous successions, steeply dipping aquifers, or extensively faulted or fractured sections), allowing large-scale vertical fluid movement due to density differences (Galloway and Hobday 1996). Convection circulation can be pronounced in the vicinity of magmatic intrusions, in thick sequences containing evaporites, and in coastal regions. The potential of density-driven convection within the compactional or thermobaric regimes is controversial (Bjorlykke et al. 1988; Hanor 1987; Phillips 1991). Seismic pumping is defined as fluid flow driven by repeated episodes of dilation and contraction in seismically active areas, particularly near and within faults zones (Sibson et al. 1975). Fluids are driven toward newly opened fractures and dilated pore space.

2.2.6 The Nature, Evolution and Interaction of the Regimes

Since waters involved in the meteoric regime are mainly of the meteoric origin, and since the meteoric regime is intimately controlled by the surface topography and other external factors such as climate, it can be considered as exogenous. Other forced flow regimes are characterized by abnormally high pressures and generally upward flow. Since the drives for flow in these regimes are created by internal processes, these regimes can be termed endogenous.

The fluid flow regimes are transitional in geological timescales, and their domains, boundaries and relative importance, as well as properties of fluids, change during basinal and post-basinal evolution. Changing relationships between the regimes reflects stages (geostatic, transitional, hydrostatic and endogenous; Pinneker 1977) in the hydrogeologic evolution of basins. The compactional regime dominates young actively subsiding basins, and still deeper burial leads to the increasing development of the thermobaric regime in the deepest portions (the geostatic stage). Respectively, the fluid flux under these regimes attains the maximum. With the termination of subsidence and progression of uplift and continental exposure, these regimes dissipate and respective fluxes cease, being increasingly displaced by the expanding meteoric regime (the transitional stage). Invasion of meteoric waters starts from the basin's margins, but with the growth of the topographic relief, the meteoric regime embraces the upper story in the internal regions. It further expands in depth and becomes dominating in mature basins (the hydrostatic stage). In collision belts, the compactional regime can be reactivated and affect adjacent basins due to the loading imposed by thrust sheet stacking. The thermobaric regime can be maintained at depth, and the deep endogenous regime can expand upward in geodynamically active areas, where mantle plumes and asthenosphere upwellings contribute to abnormal P-T conditions and fluid flow. The expansion of the endogenous regime (the endogenous stage) may occur at any stage of the basin evolution.

Since mixing of fluids contrasting in physicochemical properties often results in bursts of the aggressiveness, zones of interaction between different regimes are potentially favorable for hypogene speleogenesis. Hypogene speleogenesis is commonly a part of mixed flow systems, where rising flows of the endogenous regimes interact with the meteoric waters. The geometry of interaction zones is controlled by respective fluid potentials and geological heterogeneities and is time-dependent. The boundaries can be blurred, but they are often more distinct where they coincide with low-permeability strata of a regional extent. Importantly, the domain of the meteoric flow, perched on ubiquitously ascending fluids of the endogenous regimes, is often pierced by rising cross-formational injections and plumes of deep fluids driven by the endogenous regimes, guided by major sub-vertical tectonic disruptions (Klimchouk 2012). Thermobaric and geochemical anomalies induced by such intrusions are often traced vertically across several aquifers in stratified shallower aquifer/aquitard systems.

2.3 Endogenic Fluids in the Deep Crustal Settings and Their Connection to the Upper Crust

Fluids in the middle/deep crust originate from in situ chemical reactions driven by the heating of rock masses (see Sects. 2.1 and 2.2.3) and come from degassing of the asthenosphere and the core. Fluid flux from crustal devolatilization is a major contributor to planetary volatile cycling and principal geologic processes (Ague 2003).

Within the huge range of P–T conditions in the middle/deep crust and mantle, aqueous fluids are highly variable in composition (generally represented by H_2O, different nonpolar gasses like CO_2 and CH_4, and different dissolved metal halides like NaCl or $CaCl_2$) and exhibit specific physicochemical properties (Liebscher 2010). The cited work is an excellent recent review of phase relations in one- and multi-component fluid systems at high pressure and temperature. The supercritical fluids, combining comparably low viscosity with high solute contents, are very effective metasomatising agents (Liebscher 2010; Yardley 2013). Although solubilities of many rock-forming minerals in brines vary in a complex manner with changes in P-T conditions, pH and fluid composition, they are commonly much higher in the middle/deep crustal settings than at ambient conditions (Newton and Manning 2002, 2005; Caciagli and Manning 2003; Dolejs and Manning 2010), suggesting a possibility of enhanced dissolution of various rocks. Evaluation of dissolution potential and processes (including metasomatic ones) in high P-T conditions from the perspective of karstification is still an open research area. When a multi-component endogenous gas–vapor fluid rises through permeable zones, down along the pressure and temperature gradients, complex processes of phase transformations and geochemical evolution occur, with the formation at different levels of alkaline solutions and strong acids, such as HCL and H_2SO_4 (Malyshev 2011), which causes the vertical zoning of specific environments aggressive to different lithologies (Klimchouk 2013b).

The origin and behavior of fluids in the deep crustal settings are a matter of much controversy. Metamorphic fluid fluxes are likely controlled by dewatering rates of metamorphic piles undergoing devolatilization (Yardley and Bodnar 2014). Estimates of fluxes of metamorphic fluids by different methods for different regions vary in six orders of magnitude (Ague 2003). The highest fluxes are inferred to occur in subduction zones where dehydration of subducting oceanic lithosphere occurs (Breeding and Ague 2002; Schmidt and Poli 2014) and in the continental collision and mid-crustal shear zones. Aque (2003) estimated that regional devolatilization fluxes for zones of continental collision with the dominant pervasive flow are less than $\approx 10^4$ m^2 m^{-2}, averaging around 500 m^3 m^{-2}, but focusing of flow by structural features such as ductile shear zones may provide fluxes in excess of 10^5 m^3 m^{-2}. The total flux from the deep crust in active mountainous belts is estimated to be in excess of $\approx 10^{17}$ kg Myr^{-1} (Ague 2003). Fluxes of volcanic fluids can be high around and above magma bodies crystallizing in the crust (Oppenheimer et al. 2014), especially those intruded into shallow crustal settings (Yardley and Bodnar 2014). The input of external H_2O-bearing fluids, such as those derived from dilatancy pumping (Sibson et al. 1975), dehydrating schists (Aque 2003; Schmidt and Poli 2014) or mantle and core sources (see Sect. 2.2.3), can contribute significantly to the total flux.

Highly overpressured fluids inevitably escape upward to the upper crust where they invade hydrothermal systems and mix with formation fluids. The “single pass” flow model is widely accepted for fluid dynamics in the middle/lower crust (Fig. 2; Ague 2003; Yardley and Bodnar 2014). Single pass fluid flow can be pervasive or channelized into fractures and faults or along more permeable layers. A possibility of the “multi-pass flow” recirculating by convection is commonly questioned due to generally low-permeability environment and problematic passing of the brittle/ductile transition (BDT), although studies in some active mountain belts (e.g., in Pakistan and New Zealand) suggest that shallow fluids can penetrate close to this boundary (Templeton et al. 1998; Poage et al. 2000). In the upper crust, multi-pass systems driven by convection are common in the vicinity of thermal anomalies (geothermal circulation). A study by Wing and Ferry (2002) suggests that the metamorphic fluid flow can locally be very complex and include upward, downward, up-temperature and down-temperature components, illustrating that the common current notions of fluid dynamics in the middle/deep crust are highly generalized and incomplete.

From the perspective of hypogene karstification, one of the key issues is assessing the potential for upflow of deep fluids to the upper crust where they may contribute to the endogenous regime and cause anomalous conditions in the sedimentary cover (Kissin 1967; Ezhov et al. 1992; Klimchouk 2012). Of particular importance here is the permeability distribution.

Ingerbitzen and Manning (2010) and Ingerbitzen and Appold (2012) pointed out that metamorphic petrologists have long recognized that permeability of the middle/deep crust is a dynamic parameter responding to tectonic stress, dewatering and fluid production, and other geochemical reactions. The above-cited works comment that this view is in stark contrast to the traditional hydrogeologic concept of permeability as a relatively static material property that exerts control on fluid flow. It is worth noting, however, that this view is certainly compatible with the understanding of permeability in karst science, where its dynamic self-adjustment is seen as one of the defining attribute of karst (Worthington and Ford 2009; Klimchouk 2015).

A large-scale crustal permeability estimate based on geochemical data and heat and mass transport models for rocks that underwent substantial fluid flow (Manning and Ingerbitzen 1999) yielded a permeability-depth curve log $k \approx -14 - 3.2 \log z$ (where permeability k is in m^2 and depth z is in km; Fig. 3a), but the data below about 12.5 km are actually fitted just as well by a constant permeability of $10^{-18.3}$ m^2 (Fig. 3b; Ingebritsen and Manning 2003). In other publication (Ingebritsen and Manning 2010), data have been compiled that represent local-to-regional-scale, transient, permeability-generation events that entail permeabilities much higher than these mean k–z relations would suggest. These data yield a curve log $k \approx -11.5 - 3.2 \log z$. Both data sets suggest a high variance and strong depth dependence of permeability at crustal depths of less than 10–15 km, with less variance and essentially no depth dependence below this level, which supports a general distinction between the hydrodynamics of a brittle upper crust and a ductile lower crust that is dominated by devolatilization reactions and internally derived fluids (Ingebritsen and Manning 2010). Ingebritsen and Manning (2010) noted that even the disturbed-crust values may underestimate the maximum transient permeabilities.

The change in the curves in Fig. 3b to vertical at depth of about 12 km corresponds to the brittle/ductile transition. It is broadly accepted that this transition occurs, depending on the density of the rock column and the geodynamic regime, at depths varying from 7 to 15 km (even at lesser depths in some cases), and that the maximum compaction of rocks and vanishing of open porosity in the base of the brittle domain causes steep decrease in permeability in this interval, thus creating a global hydrologic seal (Mukhin 1965; Ivanov 1966, 1970; Ezhov and Vdovin 1970; Etheridge et al. 1983; Ezhov and Lysenin 1986, 1988; Bailey 1990; Fournier 1991; Ivanov and Ivanov 1993). This interval (the “buffer” sub-zone of the meso-zone in Fig. 4c and d, according to Ezhov and Lysenin 1986, 1990) separates the upper domain of predominantly hydrostatic fluid pressures from the underlying domain of irregular very high and (still deeper) nearly lithostatic pressures. At deeper levels, where the ductile regime predominates, rocks are weak and the near-lithostatic fluid pressures encountered during

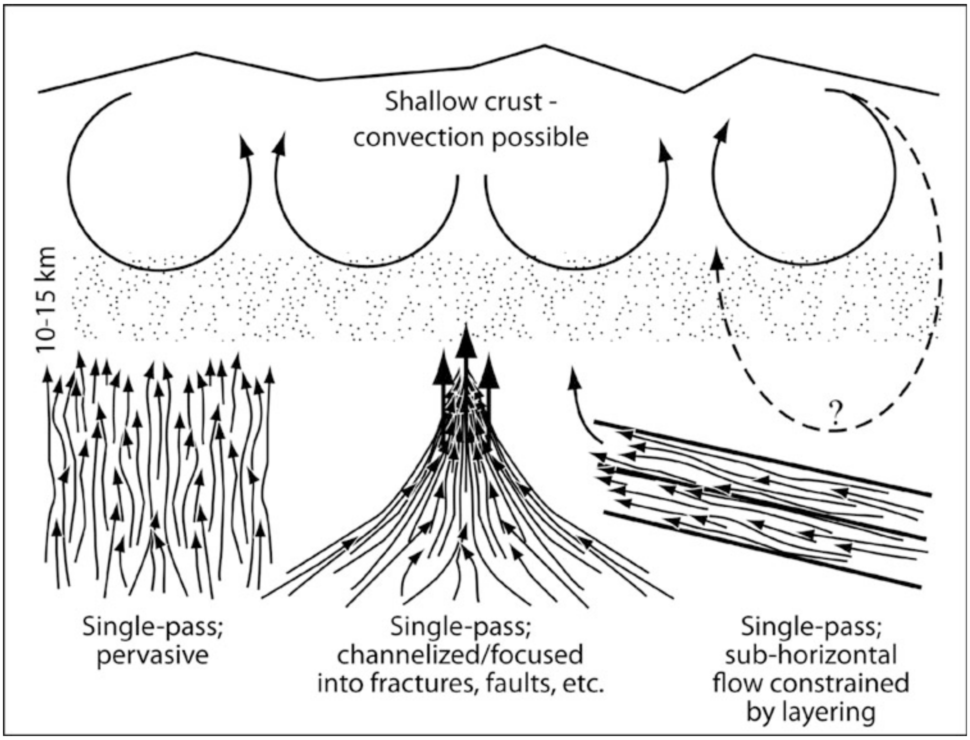

Fig. 2 Schematic cross section through the crust illustrating some possible modes of metamorphic fluid flow (reprinted from Treatise on Geochemistry, vol. 3, Ague JJ, Fluid Flow in the Deep Crust, pp. 203–247, 2003, with permission from Elsevier)

metamorphism result in relatively high permeabilities. The BDT (the "buffer" sub-zone) serves as a planetary-scale regulator of devolatilization of the deep crust and the mantle degassing, restricting the transfer of deep fluids to the upper crust and the hydrosphere (Ezhov and Lysenin 1986, 1988; Ivanov and Ivanov 1993). It is generally thought that large deep faults originate above the BDT but not cross through it, and many researchers believe that fluid transfer through this boundary occurs mainly by magmatic intrusions or diffusion.

Ezhov and Lysenin (1988) inferred that breakthroughs of deep fluids through the buffer sub-zone occur in a localized and pulsed manner, where and when the fluid pressures below build up high enough to exceed tensile strength and cause hydrofracturing. In this way, highly overpressured fluids periodically invade into the brittle upper crust and propagate upward, creating zones of baric, thermal and geochemical anomalies in the above-lying zone of hydrostatic pressures. Similar views were developed by Connolly (1997, 2010) and Connolly and Podladchikov (2013) into an elaborate and elegant model suggesting that metamorphic devolatilization reactions generate deformation-propagated fluid flow in form of pulses of fluid that travel upward as porosity waves, leaving trails of interconnected pore space in their wake. These trails act as preferential pathways for subsequent fluid flow. The models of periodic breakthroughs of deep endogenous fluids into the upper crust are corroborated by mounting evidence (from studies in many geological disciplines, particularly in petroleum geology, ore geology, geochemistry, sedimentology and non-volcanic hydrothermalism) of episodically high fluid pressures, fluid flow occurring in pulses or in multiple phases, and often at extraordinary rates, and of multifarious anomalies of fluid properties and composition in deep parts of the upper crust.

That deep fluids can traverse the BDT, and upflow to the upper crust in an advective regime without magma melts is strongly supported by numerous unambiguous identifications of mantle-derived volatiles, such as noble gasses and CO_2, in shallow aquifers in various regions devoid of recent volcanic activity. Kulongoski et al. (2005) showed a strong correlation between deep crust (^4He) and mantle-derived (^3He) helium isotopes in the eastern Morongo Basin, California, and association of sources for these components with faults. They dismiss a basin-wide diffusive flux of the mantle helium as the dominant process to transfer mantle volatiles to the shallow crust. They suggest that an advective flow regime drives He transport through the crust and that episodic fracturing in the Eastern California Shear Zone regulates mantle- and deep crust-derived fluxes of helium. Kennedy and van Soest (2007) show that mantle volatiles (He) leak over a wide area in the Basin and Range Province, southwestern USA, with $^3He/^4He$ ratios much higher than expected for regions with little magmatic activity. This indicates the enhanced flow of mantle fluids through the lower/middle crust and the BDT. These authors suggest that the shear force twisting the regional strain generates vertical faults that link the brittle upper crust with the ductile lower crust. Strain localization induced by an increasing dextral shear component superimposed on the extensional stress

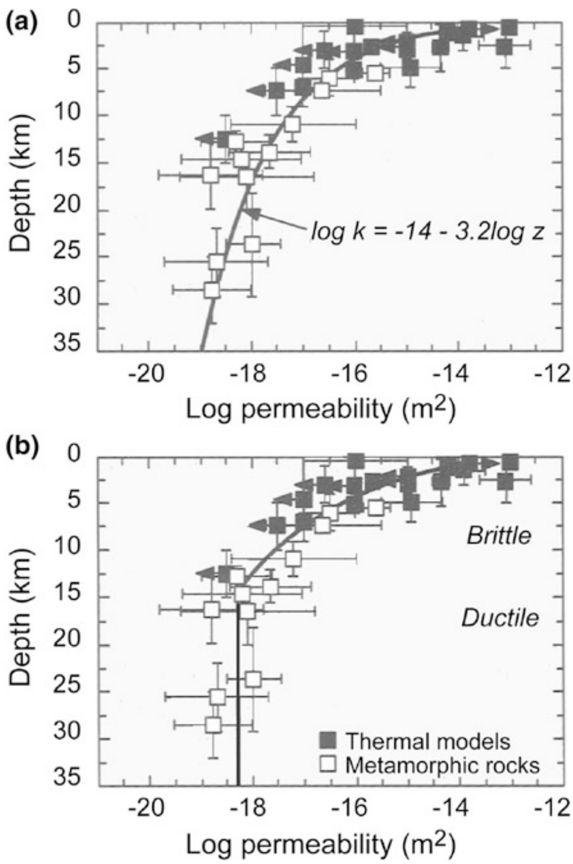

Fig. 3 Estimates of permeability based on hydrothermal modeling and the progress of metamorphic reactions showing **a** log fit to data and **b** data below 12.5-km depth fitted with a constant value of $10^{-18.3}$ m^2 (reprinted from Journal of Geochemical Exploration, issue 78–79, Ingebritsen SE and Manning CE, Implications of crustal permeability for fluid movement between terrestrial fluid reservoirs, pp. 1–6, 2003, with permission from Elsevier)

field must mechanically couple the brittle and ductile crustal zones, generating vertically oriented downward fault splays that extend through the ductile crust and into the mantle. These splays would act as conduits for fluid flow (Kennedy and van Soest 2007).

Newell et al. (2005) and Crossey et al. (2009, 2016) also provide ample evidence of heterogeneous mantle degassing and inputs from deep crustal sources throughout the western USA. Mantle-derived volatiles, including noble gasses and CO_2, are contained in travertine-depositing thermal and nonthermal springs in a variety of locations and tectonic settings throughout the entire region, such that many aquifer systems are influenced by mixing of deeply sourced and circulated waters. Karlstrom et al. (2013) determined that 27% of helium in springs in the Colorado Rocky Mountains is mantle derived, and 76 ± 20% of CO_2 come from deep (endogenic) sources. The endogenic CO_2 component in springs yields an integrated annual flux of deeply derived CO_2 to the groundwater system of $\sim 1.4 \times 10^9$ mol/yr (Crossey et al. 2009). Newell et al. (2005), Karlstrom et al. (2013) and Crossey et al. (2016) show that variations in mantle helium signals correlate best with low seismic-velocity (fluid-rich) domains in the mantle and lateral contrasts in mantle velocity.

Karlstrom et al. (2013) and Crossey et al. (2016) summarized features of the distribution of mantle helium signals in the western USA. The maximum $^3He/^4He$ values are associated with the active volcanic center in Yellowstone, reinforcing the models for its plume origin and emphasizing the role of magmatic intrusions in transferring mantle fluids to the crust. Very high values (approaching the values characteristic of mid-ocean ridges) are also found above the Cascadia subduction zone and San Andreas transform, indicating that plate boundaries provide zones for flux of asthenosphere-derived volatiles. High but heterogeneously distributed $^3He/^4He$ values elsewhere in the western USA, even in areas away from active volcanic regions and plate margins, unambiguously indicate that in regions of extending continental lithosphere mantle-derived volatiles can indeed transit through the brittle/ductile transition and rise to mix with aquifers in the shallow crust along deeply penetrating extensional fault networks (Crossey et al. 2016). There is a growing body of evidence for asthenosphere-derived volatiles, particularly CO_2, in the upper crust from many other regions of Cenozoic lithospheric extension, based on the integration of $\delta^{13}C$ data with noble gas isotopic tracers. Such regions include (just to name a few) many regions of China (Xu et al. 1995), the Pannonian and the Vienna basins (Sherwood-Lollar et al. 1997), the Bohemian massif (Weinlich et al. 1998), the Great Artesian Basin of Australia (Crossey et al. 2011; Italiano et al. 2013) and the Central Anatolia of Turkey (Bayari et al. Chap. 27). It is not surprising that those regions host remarkable examples of hypogene karst.

2.4 Hydrodynamic Zoning of the Upper Crust in Continents

Fluid properties and regularities of fluid flow dynamics systematically change with depth. There are several approaches to generalizing these changes within various schemes of vertical hydrodynamic zoning. Zones are distinguished based on (Fig. 4c) generalized hydrostratigraphy (hydrogeologic stories; Vsevolozhskiy 1983), hierarchy of gravitational flow systems (Tóth 1963, 2009), circulation intensity (Ignatovich 1950), dominating flow regimes (Shestopalov 2014; Vsevolozhskiy and Kireeva 2014) and distribution of fluid pressures (Mukhin 1965; Ezhov and Lysenin 1990). Although categories and terminology of these schemes are sometimes difficult to reconcile (Fig. 4),