Climate Change and Groundwater

Edited by
W. Dragoni and B. S. Sukhija

Geological Society
Special Publication 288

Climate Change and Groundwater

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GEOLOGICAL SOCIETY SPECIAL PUBLICATION NO. 288

Climate Change and Groundwater

EDITED BY

W. DRAGONI
University of Perugia, Italy

and

B. S. SUKHIJA
National Geophysical Research Institute, Hyderabad, India

2008
Published by
The Geological Society
London

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Published by The Geological Society from:
The Geological Society Publishing House, Unit 7, Brassmill Enterprise Centre, Brassmill Lane, Bath BA1 3JN, UK

(*Orders*: Tel. +44 (0)1225 445046, Fax +44 (0)1225 442836)
Online bookshop: www.geolsoc.org.uk/bookshop

British Library Cataloguing in Publication Data

A catalogue record for this book is available from the British Library.

ISBN 978-1-86239-235-9

Typeset by Techset Composition Ltd, Salisbury, UK

Printed by Antony Rowe, Chippenham, UK

Distributors

North America
For trade and institutional orders:
The Geological Society, c/o AIDC, 82 Winter Sport Lane, Williston, VT 05495, USA
Orders: Tel +1 800-972-9892
Fax +1 802-864-7626
E-mail: gsl.orders@aidcvt.com

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Orders: Tel +1 918-584-2555
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India
Affiliated East-West Press Private Ltd, Marketing Division, G-1/16 Ansari Road, Darya Ganj, New Delhi 110 002, India
Orders: Tel +91 11 2327-9113/2326-4180
Fax +91 11 2326-0538
E-mail: affiliat@vsnl.com

Preface

The concept of 'Climate Change' (basically implying global warming) has been with us for two to three decades. Not a day passes without a new story on climate change, and often the evidence and the hypotheses offered are uncertain and conflicting. This suggests that there is still much to learn about the causes, the magnitude and the persistence of the processes that are going on; however, in spite of all the uncertainties, the reality of global warming is no longer questioned.

Global warming will impact the hydrological cycle, and there is evidence that this is already occurring, with many countries facing more frequent droughts than in the past, whilst rainfall and flooding have intensified in some areas. Climate change will directly affect groundwater recharge, timing of recharge events, storage in aquifers, the quality of groundwater and the freshwater/seawater interface. These impacts are of paramount importance for many reasons, not least that groundwater is, and will be, the main solution to water scarcity during droughts and in permanently arid or semi-arid areas. It is imperative, therefore, to establish rational management and conservation plans.

With this in view, the IAH Working Group on Groundwater and Climate Change organized a special session on 'Impact of Climate on Groundwater Resources' during the XXXII International Geological Congress (held in August 2004, in Florence, Italy). Those research workers, expert in climate change and groundwater, who were unable to participate in the conference have also contributed to this volume. It contains thirteen papers covering a variety of topics related to change in climate and groundwater resources.

This volume provides only a glimpse of some of the more important aspects associated with the general theme of *Climate Change and Groundwater*. The book will, however, stimulate interest and research into some of the topics covered which hopefully will help towards developing tools for coping with the environmental problems anticipated in the coming decades.

The editors wish to thank Angharad Hills, Nick Robins, Sally Oberst and Alison Tucker, of the editorial staff of the Geological Society, for their cooperative efforts and patience.

WALTER DRAGONI
BALBIR S. SUKHIJA

Acknowledgements

The editors are grateful to the following people for their assistance with their reviewing of the papers submitted to this special publication.

Dr Andreas Angelakis, National Foundation for Agricultural Research, Iraklio, Greece
Prof Andrew Binley, Lancaster University, UK
Dr Costanza Cambi, Perugia University, Italy
Prof Mebus A. Geyh, Rübeland 12 - OT Bannetze 29308 Winsen/Aller Germany
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Prof Giovanni Maria Zuppi, 'Ca' Foscari' University, Venice, Italy

Contents

Climate change and groundwater: a short review

W. DRAGONI[1] & B. S. SUKHIJA[2]

[1]*Dipartimento di Scienze della Terra, Università di Perugia, Piazza Università 1, 06001 Perugia, Italy (e-mail: dragoni@unipg.it)*

[2]*National Geophysical Research Institute, Hyderabad-500 007, India (e-mail: bssngri@rediffmail.com)*

Abstract: There is a general consensus that climate change is an ongoing phenomenon. This will inevitably bring about numerous environmental problems, including alterations to the hydrological cycle, which is already heavily influenced by anthropogenic activity. The available climate scenarios indicate areas where rainfall may increase or diminish, but the final outcome with respect to man and environment will, generally, be detrimental. Groundwater will be vital to alleviate some of the worst drought situations. The paper analyses the main methods for studying the relationships between climate change and groundwater, and presents the main areas in which hydrogeological research should focus in order to mitigate the likely impacts.

This article has two aims. The first is to present a summary of the current knowledge of the relationships between climatic variations and water resources, with emphasis on groundwater. The second aim is to review the main issues that groundwater specialists will have to face and study in order to minimize the impact of climatic variation and to protect groundwater resources.

The climate has changed in the past, is changing presently and will change in the future. The scale of the fluctuations varies from hundreds of millions of years to decades or less (for example Huggett 1991; Goudie 1994; Issar 2003; Lamy *et al.* 2006; Yang 2006). The present climatic trend (i.e. a warming trend), which is no longer a hypothesis but a planet-wide observation, may correspond to a natural warming phase, probably at the scale of a few hundreds years, which began in the nineteenth century; the warming is being accelerated and increased because of the anthropogenic release of greenhouse gases from fossil fuels burnt during the last two centuries. The main concern raised by global warming is that climatic variations alter the water cycle; indeed, in many cases, the data show that the hydrological cycle is already being impacted (Dragoni 1998; Buffoni *et al.* 2002; Labat *et al.* 2004; Huntington 2006; IPCC 2007).

Today there is a very large consensus, supported by an impressive set of observations and analyses, that anthropogenic activity is the main factor causing the present global warming (Trenberth *et al.* 2006; Kerr & Balter 2007). However, the Intergovernmental Panel on Climate Change (IPCC), does not provide total certainty to this view, but only indicates a probability greater than 90% (IPCC 2007) and a few recent papers raise some doubts about the driving role of greenhouse gases (de Jager & Usoskin 2006; Stanhill 2007; Svensmark 2007). Indeed, a heated dispute is going on, as there is a minority of scientists who claim that the main reason for the present climatic behaviour is natural (sun variability being the most probable) and that, very likely, the future warming will be moderate (Essex & McKitrick 2003; Landscheidt 2003; Santer *et al.* 2004; Michaels 2005; Singer & Avery 2006; IDAG 2005; Shaviv 2005; Scafetta & West 2006; Zastawny 2006; Lockwood & Fröhlich 2007). This issue is critical, because the worst possibilities considered by the IPCC indicate that the temperature will rise by several degrees and the warm phase will last for centuries, with dramatic consequences beyond those that can reasonably be defined at present. In any case, today, there is an unanimous consensus on the forecast that the warming will persist for decades, no matter what action is taken (Michaels 2005; Singer & Avery 2006; Trenberth *et al.* 2006; IPCC 2007). As the warming process continues, it will bring about numerous environmental problems, among which the most severe will relate to water resources (Loáiciga 1996, 2000; Milly *et al.* 2005; Holman 2006; IPCC 2007).

The magnitude of future consequences can be inferred from the dramatic effects caused by the natural and 'moderate' climatic changes that occurred during the last millennium, during which millions of deaths all over the world were caused directly by the alternation of droughts and short cool–warm periods (Lambe 1977; Goudie 1994; Dragoni 1998; Brown 2001; Fagan 2001; Davis 2002). The development (and in some cases the

From: DRAGONI, W. & SUKHIJA, B. S. (eds) *Climate Change and Groundwater*. Geological Society, London, Special Publications, **288**, 1–12.
DOI: 10.1144/SP288.1 0305-8719/08/$15.00

disappearance) of many civilizations was determined by natural and 'moderate' climatic change (Stewart 2005; Brooks 2006; Cremaschi *et al.* 2006; Kumar *et al.* 2006; Issar & Zohar 2007). Clearly a comprehensive knowledge of climate variations in space and time is vital in order for human society to adapt and survive. The key issues in the study of climate change (Oldfield 2005) are:

(i) what will be the amplitude and rate of global climate change over the next century and beyond;
(ii) how will the global mean climate be expressed in terms of extreme droughts and floods, sea level changes, groundwater recharge, soil degradation, deforestation, loss of biodiversity, and changes in ecosystem functioning, especially in view of the human-induced greenhouse effect; and
(iii) how do the complex changes involved affect the key issues of vulnerability and sustainability of water resources for the human population in general and groundwater in particular.

The importance of the relationship between groundwater and climatic change cannot be overstated. The global volume of groundwater is estimated at between 13% and 30% of the total volume of fresh water of the hydrosphere (Jones 1997; Babklin & Klige 2004) and groundwater provides 15% of the water used annually (Shiklomanov 2004*b*), the remainder being from surface water. Aquifers mitigate droughts as they have a high storage capacity and are less sensitive to climate change than surface water bodies. Surface water baseflow is, of course, groundwater discharging from store.

General observations regarding scenarios, future climate and water resources

The IPCC prepared five reports, the latest of which, in a preliminary version, was released in January 2007. The conclusions of this report most relevant to water resources and groundwater are (IPCC 2007):

- Projected warming in the twenty-first century shows geographical patterns similar to those observed over the last few decades. Warming is expected to be greatest over land and at the highest northern latitudes, and least over the Southern Oceans and parts of the North Atlantic Ocean;
- Snow cover is projected to contract. Widespread increases in thaw depth are projected over most permafrost regions;
- The more optimistic globally averaged rises in sea level at the end of the twenty-first century are between 0.18–0.38 m, but an extreme scenario gives a rise up to 0.59 m;
- It is very likely that hot extremes, heat waves and heavy precipitation events will continue to become more frequent; and
- Increases in the amount of precipitation are very likely at high latitudes, whereas decreases are likely in most subtropical land regions.

The IPCC scenarios (global and regional) are based on the results from Global Circulation Models (GCMs), traditionally considered by the IPCC to be the most reliable tools for obtaining indications regarding the future climate (Troen 1993; Kattenberg *et al.* 1996; IPCC 2007). Uncertainties conspire to make the model output, a rough approximation at best, of what could happen under various assumptions of greenhouse gases emissions (Covey 2003; Friedlingstein *et al.* 2003; Bender *et al.* 2006; Hegerl *et al.* 2006; Schmidt *et al.* 2004; Masson-Delmotte *et al.* 2006; van Ulden & van Oldenborgh 2006; Zhang *et al.* 2006; IPCC 2007; Schneider 2007). Future scenario outputs may even be contradictory (Rosenberg *et al.* 1999; Gagnon & Gough 2005; Stephenson *et al.* 2006; IPCC 2007; Kripalani *et al.* 2007; Li *et al.* 2007) and the results are averages over vast areas (IPCC 2007; Jacob *et al.* 2007; Ruosteenoja *et al.* 2007). By using Regional Circulation Models (RCMs), nestled within a GCM, one can arrive at averaged results (in terms of rainfall and temperature) for areas as small as 600 to 2500 km^2, but the results often depend more on the choice of the initial GCM than on the choice of the emission scenarios (Hay *et al.* 2006; Graham *et al.* 2007; Ruosteenoja *et al.* 2007; Olesen *et al.* 2007). This is unsatisfactory for defining the impact of climatic variations on water resources, and for planning intervention strategies for mitigating the likely impacts.

The inadequacy of the GCMs suggests that other approaches, although empirical, should be used together with the GCMs. The 'analogue approach' gives information that is more specific than that given by the GCMs by reconstructing past climates (i.e. temperature and precipitation) in a given area. These reconstructions can be used to construct future scenarios by analogy. The analogue approach assumes that, if a given average temperature variation corresponded to a given variation in rainfall or in water resources in the past, a similar temperature variation in the future will cause similar effects. Thus it is accepted that this occurs regardless of the causes of the variations in temperature, which may also be of different types, such as variations in the solar constant or the concentration of greenhouse gases in the atmosphere. This assumption can only be maintained if the temperature variations being compared are similar and if they occur during similar atmospheric boundary conditions, i.e. during time periods in which the planet has cryosphere, oceans and land masses in similar conditions (Wigley *et al.* 1986; Dragoni 1998). Similarity can be accepted only if the 'palaeoanalogue periods' go back no more than a few millennia or, on a more detailed scale, if the palaeoanalogues consist of multi-annual instrumental series, such as those of the warm years at the beginning and the end of the twentieth century. Of course the future scenarios based on such palaeoanalogues are not quantitatively sound, and cannot be extrapolated confidently into the future beyond a few decades. Thus, despite the progress made by GCMs and by information obtained by the analogue approach, it must be recognized that a definition of scenario given about twenty years ago is still valid: 'scenarios are not meant to be predictions of future climate; rather they are meant to be internally consistent pictures of a plausible future climate, a basis for other workers to evaluate the possible impacts of climatic change on Man and society' (Wigley *et al.* 1986).

Data and information on past climatic and hydrological conditions are important for verifying whether a GCM or RCM is potentially reliable: only a model that provides good results for the present and/or for past climate can be reliably used for constructing future scenarios (Bell *et al.* 2003; Karl & Trenberth 2003; Dearing 2006; Sloan 2006).

Confidence is increased if one or more models and the analogue approach independently indicate a similar future scenario in terms of temperature and rainfall. Nevertheless, the actual intensity and spatial and time variability of rainfall and temperature for a given scenario and a given region still remain uncertain. The same degree of uncertainty is retained when translating rainfall and temperature to evapotranspiration, runoff and aquifer recharge, whatever procedure is adopted (Strzepek & Yates 1997; Di Matteo & Dragoni 2006).

Another consequence of the uncertainties intrinsic to the climatic scenarios is that the impact of the conditions provided by the scenarios on hydrogeological systems are tentatively simulated for different, and more or less arbitrary values for the climatic factors (Rosenberg *et al.* 1999; Loáiciga *et al.* 2000; Taeuea *et al.* 2000; Nijssen *et al.* 2001; Yusoff *et al.* 2002; Allen *et al.* 2004; Gagnon & Gough 2005; Jha *et al.* 2006; Vicuna & Dracup 2007; Olesen *et al.* 2007). The results provided by this approach show the most probable direction of change, the sensitivity to different factors which regulate hydrological systems, and point to the processes which will be modified most by future climate variation.

Main tools to study the relations between groundwater and climate variations

General considerations

A good knowledge of the geology and hydrogeology of the study system is an essential prerequisite to investigating the impact of climate change. Ideally, the study of groundwater resources should be based on a reliable, continuous and dense database of hydrometeorological data and soil moisture, covering a long time interval. These data should be coupled with a large amount of spatially distributed quantitative information such as hydraulic conductivity and porosity. However, adequate data are rarely available and work is often qualitative. Thus a high quality network of data collection needs to be established. The networks are essential not only to evaluate the present situation but also to follow the evolution of the processes and, therefore, to validate the knowledge gained through time. Natural experimental systems, not directly influenced by anthropic activities (in terms of variations in the use of the land or irrigation) may be used to isolate, at least locally, the effects of climate variation from those derived by Man's presence. Management of water resources has brought, even on a regional or continental basis, significant variations (Vörösmarty *et al.* 2004; Shiklomanov 2004*a*).

Remote sensing provides a convenient way of assessing the spatial and temporal variations in water fluxes in different components of the water cycle. Attempts to estimate small scale temporal changes in the gravity field due to redistribution of water are in progress. The GRACE (Gravity Recovery And Climate Experiment) satellite is aimed at observing the gravity field to a high accuracy (*c.* 1 cm in terms of geoid height and a spatial resolution of 200–300 km). Any redistribution of water masses in different parts of water cycle may result in time variation of gravity field.

Numerical hydrological and hydrogeological models, in spite of being a simplified representation of reality, are invaluable tools for describing and understanding hydrogeological processes. The model serves both the understanding the system and prediction, once the model is validated and calibrated using historical data. Models should ensure internal consistency, compatibility with uncertainties, compatibility with constraints of data and robust performance (Oldfield 2005). In order to assess the effect of climate on a particular groundwater system, dedicated monitoring of various parameters is required. There are many modelling codes for flow in both the unsaturated zone and the saturated zone and various water quantity and quality issues can be accounted for depending upon the boundary conditions, subsurface properties and process representation. The models can differ in scale and in detail depending upon specific process (e.g. evapotranspiration or groundwater mass transport).

Isotope methods

During the last four decades, isotope methods have been developed which are based on proxy records of climate change in the past in different media, and such studies are termed as palaeohydrology/palaeoclimatic studies (Parrish 2001; Mazor 2003). The different media/materials through which climate change can be studied are:

(i) oxygen and carbon isotope composition of benthic and planktonic foraminifera (Bar-Matthews *et al.* 2003);
(ii) hydrogen and oxygen isotopic ratios of organic matter (Sauer *et al.* 2001; Shu *et al.* 2005; Webb & Longstaffe 2006);
(iii) ice cores (Thompson *et al.* 1998; Hondoh 2000; Thompson *et al.* 2000); and
(iv) carbon and oxygen composition of carbonates, cave deposits, lake, groundwater (Li *et al.* 1989; Bar-Matthews *et al.* 1998; Frumkin *et al.* 1999; Niggemann *et al.* 2003; Sasowsky & Mylroie; 2004; Pentecost 2005; Parker *et al.* 2006).

Most media are basically formed through precipitation which in turn originates or is modified by climatic processes, and these can be reconstructed on the basis of their isotopic composition, which in many cases provides some information about rainfall, aquifer recharge and the water table position at the time of deposition (Cremaschi & Di Lernia 1999; Nelson *et al*. 2000; Drake *et al*. 2004; Garnett *et al.* 2004; Mariani *et al*. 2007).

The vadose zone of an unsaturated aquifer contains information about climate change in the range of tens of years to thousand of years. This information is contained within the solute (chloride) and the ^{3}H, ^{2}H, ^{18}O profiles. In case of homogenous soils (without cracks and fissures) infiltrating water moves through the unsaturated zone by displacement (Zimmermann *et al.* 1967). Estimates of recharge using thermonuclear tritium peaks of 1963–64 have been made (Anderson & Sevel 1974; Sukhija & Shah 1976; Allison & Hughes 1978). Chloride profiles have also been used to measure mean recharge and residence time of water in vadose zone (Allison & Hughes 1978; Edmunds & Walton 1980; Sukhija *et al.* 1988; Lo Russo *et al.* 2003).

The residence time of a tracer in the vadose zone determines the number of years for which the climate information may be available. Cook *et al.* (1992) estimated the persistence time of isotopes before they are smoothed by diffusion and dispersion to vary from several years to 10 000 years. Climate fluctuations over 5–10 years are discernible in chloride profiles in regions like Cyprus (Edmunds & Walton 1980; Edmunds *et al.* 1982, 1988); however, two chloride profiles in Senegal (Edmunds *et al.* 1992) have a length of record of 30–50 years and exceptionally 475 years (at Longa). Attempts are in progress (Lal 2000) to study palaeoclimatic recharge during the last thousand years using ^{32}Si.

Impacts and adaptation: some topics on which to focus research

Impact of climate change on groundwater recharge and discharge

Any variation in the regime and quantity of precipitation, together with variations in temperature and evapotranspiration, affects groundwater recharge. In general, groundwater recharge will increase in areas where precipitation is increased and vice versa. Groundwater recharge will increase also in areas where permafrost thaws (Potter 2002; Kitabata *et al.* 2006). Most of the consequences of changes in recharge will be detrimental. There is general agreement that many areas of currently high precipitation are expected to experience precipitation increases, whereas many of the areas at present with low precipitation and high evaporation, now suffering water scarcity, are expected to have rain decreases (IPCC 2007; Issar & Zohar 2007). As conditions change rapidly, the existing infrastructure network will have to be reshaped rapidly (Potter 2002; Semadeni-Davies *et al.* 2007).

The groundwater recharge is the residual flux of water added to the saturated zone resulting from the evaporative, transpirative and runoff losses of the precipitation. It can take place by diffuse infiltration, a preferential pathway, and through surface streams and lakes. Thus groundwater recharge is a sensitive function of the climatic factors, local geology, topography and land use. Generally, measured groundwater recharge is a site specific quantity and this complicates the problem of defining its regional impact. The broad scenarios given by the GCMs should be considered only as a very preliminary basis for investigations to be carried out to understand the effects of the changing climate on groundwater. Any research on the variation of recharge has to be based on data and investigation specific to the hydrogeological system under consideration. These investigations must be based on detailed knowledge of the geological structures, and may involve the use of complex models which consider multi-component interaction, hydrological-atmospheric processes, hydrological boundary conditions and identification of model sensitivity to parameter uncertainty. In one way or another, the results given by the models must have some form of calibration, and this inevitably leads to using the known present or past conditions of the systems under investigation.

Generally, variations in aquifer recharge not only change the aquifer yield or discharge, but also modify the groundwater flow network, e.g. gaining streams may suddenly become losing streams, groundwater divides may move position.

The effects of recharge variation on groundwater flow has been considered by various authors, sometimes not within the context of seasonal variations in recharge. Meyboom (1967) has shown on a seasonal basis how the decrease or absence of recharge changes the flow relationships between recharge areas and shallow, contiguous lakes. An important paper by Winter (1999) shows how climatic conditions affect the direction of groundwater flow and the relationship between superficial hydraulic bodies and subterranean waters, at different scales. Cambi & Dragoni (2000) studied the effects of climatic variations on the Bagnara spring, located in the Umbria-Marche Apennines (Italy), in an area where both the analogue approach and the most recent GCMs forecast a decrease in rainfall and, therefore, of groundwater recharge (De Felice & Dragoni 1994; IPCC 2007). The Bagnara spring is located on the west slope of an asymmetric anticline (Mount Pennino Anticline), the core of which comprises permeable limestone and is bounded by a low permeability marl formation. The spring is located at the interface between the marl formation and the limestone. The recharge over the anticline core partly feeds the spring, and partly feeds a deeper, regional, flow (Fig. 1).

A numerical model was built and calibrated in different stages in order to quantify the possible effects of climatic variations on the Bagnara spring system. The simulations confirmed that if there is a decrease of recharge, the area feeding the spring shrinks, while the area feeding the lower regional flow system increases. This implies that any decrease in annual recharge will produce a larger decrease in the spring yield and to a smaller percent decrease of the regional flow. It may, therefore, be necessary to develop new techniques for capturing some of the water feeding the regional flow, before it reaches the polluted alluvial plains or deep evaporitic formations where salinization occurs (Cambi & Dragoni 2000).

In the areas already suffering from water scarcity and in those where the rainfall will decrease and the climate will get drier, it is crucial to try to increase recharge artificially. The techniques for the reuse and recharge of the aquifers by means of low-quality reclaimed water play a crucial role.

Groundwater discharge is another key element in the water cycle which includes loss of water from the aquifers to surface water, to the atmosphere and abstraction for human needs. The influence of the climatic changes on the discharge can be assessed by measuring, both spatially and temporally, the base flow to the rivers, lakes, wetlands and oceans and by studying the role of vegetation in transpiration.

Climatic change and fresh water discharges to the oceans. During the last decades it has been realized that the contribution of subterranean discharge of groundwater to the oceans is large, perhaps as

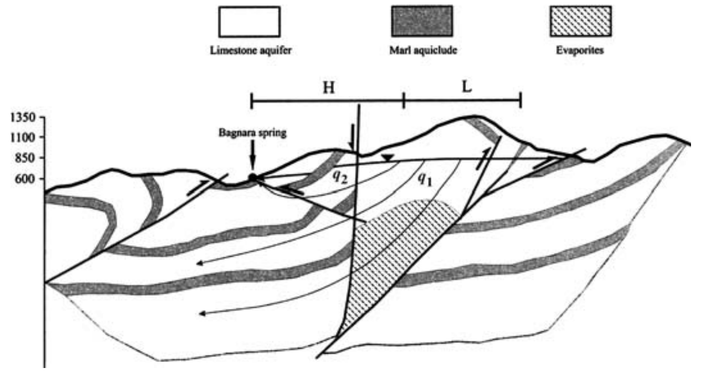

Fig. 1. Hydrogeological diagram of the Mount Pennino anticline and the Bagnara spring. Thick arrows indicate faults. Thin arrows indicate the flow path, towards the regional flow (q_1) and the spring (q_2). H, recharge area of the spring (flow towards the higher boundary); L, recharge area of the regional flow (flow towards the lower boundary). If the recharge decreases, H decreases and L increases, while the value q_2/q_1 decreases. Elevation is in metres above mean sea level, horizontal scale is equal to vertical scale.

much as 12 000 km^3/year (Speidel & Agnew 1988); a more recent review paper, presenting the methods for quantifying submarine groundwater discharge, indicates that the process is essentially ubiquitous in coastal areas (Burnett *et al.* 2006). All this water is lost to the sea, often of acceptable quality, and research to improve the measurement and recovery of it should be strongly enhanced.

Groundwater and reforestation. As trees are CO_2 sinks, reforestation coupled with new tree plantation has been considered key to maintaining control over CO_2 in the atmosphere (Schellnhuber *et al.* 2006; IPCC 2007). However, in forested areas, groundwater recharge is generally lower than in non-forested areas, and thus the water table and groundwater storage are generally lower (Scanlon *et al.* 2006). Recent research, based on computer simulations suggests that, despite carbon dioxide absorption, reforestation in high latitudes would help warming, because the tree would decrease albedo and increase evapotranspiration; conversely in the tropics the trees would have an overall cooling effect (Bala *et al.* 2007). These findings support the idea that tropical arid and semi-arid areas should, whenever possible, be reforested and afforested, as, according to some preliminary estimates, the sequestration of CO_2 could be in the range of 30–50% of industrial emissions (Issar 2006, personal communication). The water to carry on such activities should be provided by surface runoff, sewage from urban centres and, most important, by low quality, fossil groundwater, which is largely present in the arid and semi-arid areas of the world.

Climatic change, groundwater and landslides. The variations in the level of groundwater inevitably entail a variety of geomorphological and engineering effects. Slope stability is strongly influenced by the water pressure in pores and fractures and, therefore, by groundwater. In areas with increasing groundwater recharge, there will be increased slope instability. An inverse evolution would be expected in those areas in which recharge decreases. The relationships between all the factors which determine the stability of a slope, such as its geological set up, rock resistance, morphological situation, groundwater recharge, neutral pressure distribution and the quality of the climatic scenarios, are complex and spatially variable so that it is difficult to draw general conclusions, and research on the subject is relatively scarce (Dehn & Buma 1999; Dikau & Schrott 1999; Dehn *et al.* 2000; Malet *et al.* 2005; Dixon & Brook 2007). In an area around Bonn (Germany) a process-based, spatial and temporal model for groundwater variations and slope stability indicates that the most unstable conditions occurred during the transition from the more humid Little Ice Age to a dryer, recent climate (Schmidt & Dikau 2004). In the Italian Dolomites there appears to be a close correlation between landslides and climatic variations, to the extent that many of the identified and dated landslides can be considered as indicators of climatic change (Corsini *et al.* 2004).

Climatic change, groundwater rebound and sinkholes. In those areas in which mining activity was intense and where working mines required dewatering, a progressive groundwater rebound results when mining activities and pumping cease. Groundwater rebound, which may last for many years and involve very large areas, can cause problems of engineering stability and pollution (Banks & Banks 2001; Razowska 2001; Burke *et al.* 2005). Conversely, a fall in groundwater level may cause collapse of cavities with a roof close to the surface, and the formation of sinkholes. This risk is widespread in karst areas (Ford & Williams 1989; Waltham *et al.* 2004). The problems of areas with groundwater rebound and sinkholes should be tackled whilst bearing in mind future recharge variations as indicated by climatic scenarios.

Water scarcity and traditional techniques for water resource management. The problems of water shortage in arid or semi-arid areas have been tackled using traditional techniques in some areas (Pandey 2000; Radhakrishna 2004). These include water harvesting and use of qanats. Rainwater harvesting can be used to recharge groundwater via recharge ponds (Sukhija *et al.* 1997). Qanats are drainage tunnels commonly found in the arid and semi-arid areas of Europe, North Africa and Asia (Castellani & Dragoni 1997; Issar & Zohar 2007). The issue of the traditional techniques to face the present and future aridity is an important one, as these techniques allowed the survival of human populations in difficult areas and during changing climates in the past. However, these techniques are incompatible with intensive agriculture and a high population density. Moreover, qanats drain aquifers permanently, even during periods in which water is not needed. It is probable that in specific areas and with the use of present knowledge, these simple and low cost techniques can be revived and applied with much improved results.

Climatic change and groundwater quality. The climate is expected not only to affect input (recharge) and output (discharge), but also to influence the quality of the groundwater. For example, water recharged during an arid period may have a higher concentration of salts and hence higher TDS, while during a wet period the converse may occur (Sukhija *et al.* 1998). However, to appreciate such changes long-term monitoring of rainfall and groundwater quality is required. It is also possible to link the occurrence of certain ions in groundwater to particular water–rock processes that occurred during specific past climatic periods.

Final observations

Neither the impacts of climatic change on water resources nor the possibilities offered by groundwater to mitigate drought are new. What is new is the global dimension of the environmental change and its permanency. Compared to the climatic changes of the past, the present change is taking place in a world where many vast areas are densely populated, with a high water demand. Even if the climate were not to change, a water crisis will still occur. Increased average temperature during the last few decades overshadows the impact of anthropogenic activity and its impact on water resources which are larger than those caused by any recent past climate change (Vörösmarty *et al.* 2004; Bouwer *et al.* 2006). Indeed in many areas the lack of water reflects a decrease in rainfall (i.e. climatic change), but often the underlying reason is an increase in consumption (Falkenmark & Lannerstad 2005).

In order to overcome the present and future water and environmental problems it is necessary to try to predetermine the problems through focused research, based on a good set of meteorological and hydrological data, which at present are far from satisfactory. The protection and restoration of ecosystems that provide critical water resources, such as those protecting recharge areas, wetlands and mountain forests is critical. There is a need to reduce the gap between the water supply and demand with more efficient irrigation systems, training of farmers, recycling of waste water, water conservation through public awareness and groundwater legislation for better groundwater management. Another key point is international co-operation, both in research and in the rational distribution of water resources.

These actions are widely agreed by national and international bodies and by the agencies that work for the environment and water resources (Vrba & Verhagen 2006). Some of these actions, such as transboundary management of resources, are complex, and can only be implemented slowly. Installation of monitoring networks is less difficult provided the will is there, but data needs may never be satisfied (Vörösmarty & Sahagian 2000; Shiklomanov 2004*b*; Di Matteo & Dragoni 2006).

The authors, while they assume full responsibility for any errors and imprecisions, thank Gian Maria Zuppi warmly for his painstaking review of the manuscript.

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Some aspects of groundwater regime in Bulgaria with respect to climate variability

A. D. BENDEREV[1], T. V. OREHOVA[1] & E. K. BOJILOVA[2]

[1]*Geological Institute, Bulgarian Academy of Sciences, Acad. G. Bonchev Str., bl. 24, 1113 Sofia, Bulgaria (e-mail: orehova@geology.bas.bg)*

[2]*National Institute of Meteorology and Hydrology, Bulgarian Academy of Sciences, Tzarigradsko chaussee 66, 1784 Sofia, Bulgaria*

Abstract: Groundwater regime in Bulgaria is influenced by climate variability. The impact is evident especially for karst water. A time series analysis of spring discharge for selected karst basins was performed. The impact of the 1982–1994 drought period on groundwater regime was detected. For springs that drain open and mountainous karst, the impact of climate variability is similar to that on surface waters. In fact, the difference in degree of influence of the drought period is related to the specific geological structure of the karst massifs and recharge conditions. Furthermore, the porous waters are characterized by a weaker reaction to such an effect. In general, groundwater use during the 1982–1994 drought period was impacted by climate variability due to limited resource availability.

Climate variability is of importance for the territory of Bulgaria, and particularly its effect on groundwater. This influence depends on the general characteristics of the groundwater parameters as well as the specific natural conditions in the country. Long-term observations of groundwater allow deviation of the groundwater regime to be assessed with respect to climate influence.

For the aims of this research, data from the National Hydrogeological Network (NHGN) are processed. The NHGN of Bulgaria has been in operation since the late 1950s. Our study concerns a long-lasting drought period (1982–1994) in the territory of the country. Discharge data and water levels in wells on a monthly and annual basis are used for the analysis in accordance with the needs of water resource planning.

Studies similar to ours have been developed to estimate droughts in the whole of Europe using monthly and annual temporal scales (Alvarez Rodriguez & Estrela Monreal 2000). Applying the clustering technique, ten groups were summarized for the 1901–1996 period. According to Alvarez Rodriguez & Estrela Monreal (2000), north Bulgaria belongs to the Middle and Lower Danube basin, and south Bulgaria to the Eastern Mediterranean basin.

Various studies have concluded that in the last few decades the drought situation in many European regions has become more severe, due to an increase in frequency, duration or intensity of low flows (Bernhard & Doll 2001; Demuth & Stahl 2001; DVWK 1998). A further increase driven by global and climate change impacts is expected, in particular for southern European areas (Bernhard & Doll 2001; Watson *et al.* 1997).

Drought is a phenomenon that is not constrained by international boundaries and can therefore afflict many countries simultaneously. As low flows and droughts commonly cover large areas for long time periods, it has been suggested that these events should be studied within a regional context (Demuth & Stahl 2001; Tallaksen 2000).

Overview

Bulgaria is located in southeastern Europe. The country is bounded by the Danube River to the north, the Black Sea to the east, Greece and Turkey to the south and Serbia and Macedonia to the west. The territory of Bulgaria, despite its small area of around 111 000 km^2, is characterized by a large variety of relief forms. The altitude in the country ranges from 0 m (near the Black Sea) up to 2925 m (Moussala peak in the Rila mountains). The average elevation of the country is around 470 m above mean sea level (a.m.s.l). Around 31% of Bulgaria is represented by plains (up to 200 m a.m.s.l.) and around 41% by hilly territories (up to 600 m a.m.s.l.). Semi-mountainous areas (up to 1000 m) represent 15%, and mountainous territories 13%. The higher mountains are located in the southwestern part of Bulgaria. An important relief feature is the Balkan mountain chain, which crosses the country from west to east and divides the country into two parts: northern and southern. This chain is of primary significance and presents a local climatic

From: DRAGONI, W. & SUKHIJA, B. S. (eds) *Climate Change and Groundwater*. Geological Society, London, Special Publications, **288**, 13–24.
DOI: 10.1144/SP288.2 0305-8719/08/$15.00

boundary. The Balkan mountain chain is the main water divide of discharge flow in two opposite directions: to the Danube river and to the Aegean Sea. Furthermore, in the eastern part of the country the river runoff is directly into the Black Sea.

Bulgaria is characterized by a rather complicated geological structure. The outcropping rocks are of different chemical and lithological content, genesis and age. There is a large variety of sedimentary, magmatic and metamorphic rocks, and also volcanic-sedimentary rocks from Precambrian to Quaternary age. In parts of the rocks the porous, fractured and karst groundwater is formed.

Climatic characteristics and hydrological zones

The climate of Bulgaria is influenced by intensive atmospheric circulation from the Atlantic Ocean coast, continental Europe, the Aegean Sea and the Black Sea. The transfer of air masses is determined by the location of mountain chains in the territory of the country. In this respect, we should mention the Balkan range (Stara Planina mountain). The Balkan range is a barrier stopping the cold continental air masses and their propagation to the south. The Rila–Rhodopes massif, which is situated in the central southern part of the country, is a barrier to the Mediterranean influence. Important to climate variability are horizontal air movements such as vertical air drifting due to altitude. A mountain climate occurs at high altitude.

The specific orographic characteristics of the country are reasons for climatic differences and definition of the main drainage basins. Three main drainage basins or so-called hydrological zones are distinguished (Fig. 1).

Danube hydrological zone

The zone occupies north Bulgaria without its eastern part (number 1 in Fig. 1). The southern border of the basin is the Balkan range. This zone includes the drainage basin of the Iskar River going through the Balkan range. The upper recharge zone of the Iskar River is located in the Rila mountains. The hydrological zone is characterized by a temperate climate to a (European) continental climate. The Danube zone covers around 45% of the territory of Bulgaria.

Black Sea zone

This zone (number 2 in Fig. 1) is characterized by a temperate climate, and a climate with Mediterranean influence south of the Balkan range. The Black Sea zone covers around 13.8% of Bulgaria.

Aegean Sea zone

This zone (numbers 3 and 4 in Fig. 1) occupies south Bulgaria. It is possible to divide it into Western Aegean and Eastern Aegean basins. The zone is characterized by a transition from temperate climate to Mediterranean (continental–Mediterranean)

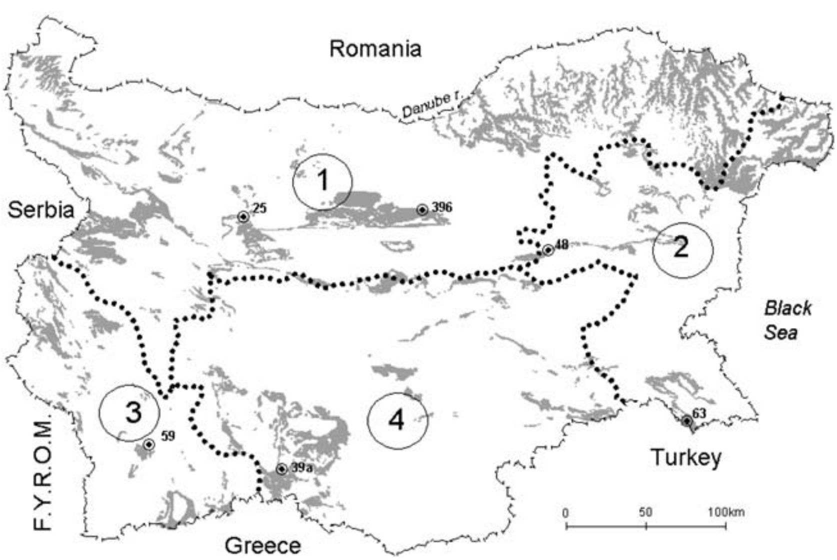

Fig. 1. Hydrological zones in Bulgaria: 1, Danube; 2, Black Sea; 3, Western Aegean; 4, Eastern Aegean. Dots indicate location and numbering of studied springs.

climate. The southern parts of the zone and river valleys are influenced by hot air masses moving from the south. The Aegean Sea zone covers the remaining 41.2% of the territory of the country.

Groundwater in Bulgaria – an overview

The structural–tectonic conditions are of primary importance for the occurrence of groundwater. This overview reflects a large variety of hydrogeological settings. They have influence on the groundwater regime, and on its vulnerability to climate variability. In the territory of Bulgaria three main hydrogeological units are distinguished (Antonov & Danchev 1980).

Low-Danubian artesian region

This region covers the north part of Bulgaria, from the Balkan mountains up to the Danube River. It is a typical platform structure, stratified, with very well defined aquifers separated by aquitards. The groundwater is accumulated in Neogene sands and friable sandstones, Palaeogene limestone and Mesozoic carbonate formations. In the southern part of the artesian basin at the foot of the Balkan range, local karst basins are widely distributed due to folding and fault processes. Alluvial aquifers are formed along the rivers and are superimposed over earlier formations.

Intermediate hydrogeological region

The higher parts of the Balkan and the areas south of it up to the Rila–Rhodopes massif are included within this region. It is characterized by a complicated block structure. Different rocks (magmatic, sedimentary and volcanic-sedimentary) bear fissured groundwater (including thermal waters) that is related to weathering zones and large tectonic dislocations. In local outcrops of carbonate rocks, detached karst basins are formed. Typical of this region is the presence of considerable depressions, filled in with unconsolidated Neogene and Quaternary deposits in which porous groundwater with large resource is accumulated.

Rila–Rhodopes massif region

This region coincides with the Rila–Rhodopes mountain massif. Here in the western part, magmatic rocks of different ages are presented, as well as Precambrian highly metamorphosed rocks with important weathering zones. Thus, fissured water is widespread in this region. In the zone with a large occurrence of Proterozoic marbles, important karst basins are formed, having typical mountainous character. In the eastern part of the region the predominant distribution has Cenozoic volcanic-sedimentary groundwater with fissured, layered water bodies and partly Precambrian metamorphic rocks bearing fissured groundwater.

General characteristics of karst basins in Bulgaria

Due to the complex geological structure and variable relief in Bulgaria, numerous karst basins are formed. According to Boyadjiev (1964) there are more than 170 karst basins. Each of them is characterized according to its location, recharge and drainage conditions, with inherent regime and different vulnerability to human impacts. In Figure 1 the main outcropping karst collectors in the country are presented in a light grey colour. The dots in the figure locate the karst springs that will be discussed.

The following classification for the main karst basins in Bulgaria is proposed.

Karst basins related to well-defined aquifers

These are related to carbonate formations of the Moesian platform. The important part of the formations is saturated, and aquifers are mainly unconfined. The confined aquifers are developed where these formations are covered. Due to the large volume of accumulated groundwater, the springs draining them are characterized by relatively stable discharge. The most important karst springs in Bulgaria, the Devnia springs, are of this type.

Karst basins in the lower parts of the country related to block carbonate bodies

The major portion of these is below the local erosion base. This type of karst basin is distributed mainly in the intermediate hydrogeological structure zone. The recharge is mainly due to precipitation, and the groundwater is karst-fissured and fissure-karstic. The karst basin in Palaeogene limestone near the town of Chirpan, and some others in the Zemen passage, belong to this type. The regime of karst springs is influenced by relatively low transmissivity of the medium and low volume of accumulated groundwater in the saturated zone. Due to the smooth relief, the soil layer is the most important factor that governs its regime.

Large karst basins with open karst and relatively smooth relief

The recharge is mainly due to rainfall. In most cases these karst basins are in the thick vadose (unsaturated) zone and insignificant phreatic zone. Variations of the discharge are related to areal recharge. Several karst basins formed in shallow synclines in Fore-Balkan (southern part of the Moesian platform) belong to this type (Gabare, Krushuna and Musina springs).

Mountain karst basins with exceptional rainfall recharge

These are formed in the Balkan and Rila–Rhodopes massif zones. They are characterized by a large difference in altitude between recharge and discharge zones and large diversity. The spring outlets appear in the lowest part of the karst basins in contact with low-permeability rock formations. Most of them are perennial, with high variation in discharge. The characteristic feature for such springs is the presence of a local saturated zone near the spring issue (for example Bistretz spring).

Mountain karst basins with river recharge

These karst basins do not differ from the previous type from the geological or geomorphological point of view. The main difference is the available permanent river recharge that leads to their specific regime. Iskretz, Glava Panega and Matnitza springs are of this type.

High mountain karst basins with important snowmelt recharge

These are karst basins with areal recharge distributed at high altitude above 2000 m in which snow cover is present up to early summer and in some parts all through the year. These conditions are of primary importance to the regime of karst springs. A typical karst basin of this type is the Razlog karst basin located in the region of north Pirin.

National Hydrogeological Network in Bulgaria

The National Hydrogeological Network (NHGN) in Bulgaria was found in 1958–1961. Nowadays the National Institute of Meteorology and Hydrology is responsible for data collection, processing, maintenance of stations and archives, supplying the state institutions and consumers with data from NHGN, and dissemination of information. Data received from NHGN includes time series for groundwater levels in observational wells and discharge for springs (Orehova & Roussev 2004).

The frequency of discharge measurements using a current meter is mainly once per month. For selected karstic springs, daily data are obtained using rating curves. At some springs the water level is recorded by limnigraph (water-level recorder), at other stations, water level is measured every day by observers. For the majority of the springs the measurements of discharge are made once or twice a month without daily observations of water level.

Water level in observational wells is measured usually once a month. Water level analogous recorders are available only for limited parts of the stations.

Many karstic basins in different parts of Bulgaria give rich material for analyses. Various karstic springs are characterized by a specific regime that represents the dynamics of the recharge and particular structure of the basin. Observational wells in kettles and into alluvial deposits along the rivers are subject to investigations of groundwater regime.

The groundwater quality is determined four times per year for selected springs and pumping stations. Basic chemical components and nitrates are defined.

General characteristics of chosen karst springs

For the purposes of this study, some typical karstic springs were selected. The springs are chosen to represent the main hydrological zones of Bulgaria. The location of the springs is represented in Figure 1. The numbering of the chosen springs is in accordance with the system used in NHGN.

Spring no. 25, Glava Panega

This spring is one of the largest in Bulgaria. It drains the main part of the so-called Zlatna Panega karst basin (according to Antonov & Danchev 1980). The karst basin is built from Upper Jurassic limestone and is part of a large anticline that is highly dislocated. The drainage basin of the spring is situated in the northern part of the Stara Planina mountain and is characterized by mountain relief. An important role in the recharge of the spring is played by the Vit River, which loses an important part of its runoff 6.5 km ESE from the spring outlet. Other main sources of spring recharge are rainfall and snowmelt in the drainage basin. The recharge from snowmelt depends not only on snow distribution within the drainage basin of the spring, but also on the altitude. The delayed snowmelt process is related to large differences in the altitudes of the basin, ranging over 700 m. The water penetration is relatively fast due to numerous karst forms. A large number of vertical and steeply inclined caves is registered in the region. The location of the spring issue is determined by the contact between limestone and Lower Cretaceous marbles, which form a barrier for the karst water. The spring is concentrated, vauclusian, rising from a steeply inclined cavern 50 m deep.

Spring no. 396, Musina

This spring drains a part of so-called Lovetch-Tarnovski karst basin (Antonov & Danchev 1980) built from Lower Cretaceous limestone. The limestone forms a shallow syncline with fully outcropping carbonate rocks. The drainage basin is a low plateau with altitude 200–300 m a.m.s.l. The basin is located in the Fore-Balkan. On the surface many dolines are observed that contribute to the fast penetration to depth of water from rainfall and snowmelt. In this region, precipitation is the main source for groundwater recharge. Geological and geomorphological settings determine the spring location. The karst water flows out from a horizontal cave. The entrance of the cave is situated in the northern part of the outcropping limestone, in the bottom of the rock cliff.

Spring no. 48, Kotel

This spring is located in the lower part of the Eastern Stara Planina mountain. The karst basin is a well-drained syncline, and the karstified rocks are Senonian limestone. The karst is of mountain type with numerous surface and underground karst forms. The vadose zone is of considerable thickness and predetermines the existence of some of the deepest caves in Bulgaria (the maximum amplitude between the zones of recharge and drainage is almost 1000 m). The recharge of karst waters is due to rainfall. The total precipitation in the region is above the average value of the country. Areas with permanent and temporal water loss are registered within the karst basin. The karst spring is situated in the eastern part of the basin and develops several outlets that form a lake. There is hydraulic connection of the water in the lake with karst water in the neighbouring cave.

Spring no. 63, Malko Tarnovo

This spring occurs in the southern limb of the syncline. Karstified rocks are Triassic marbles, marble limestone and dolomites. The recharge of groundwater is mainly due to rainfall over outcropping carbonate rocks. The relief of the region as a whole is favourable for predominantly surface runoff. In this part of the country, the snow cover is not durable.

Spring no. 59, Jazo–Razlog

This spring drains a typical alpine karst basin, the so-called Razlog karst basin (Antonov & Danchev 1980). The basin is formed in a high horst-anticline with elevation of the highlands between 2500 and 2925 m a.s.l. The spring outlet is located at an altitude of 950 m a.m.s.l. The Proterozoic marbles are subject to the process of karstification, and are fissured and weathered. Due to the typical mountain climate, precipitation is abundant and predominantly of snow type; snow cover exists until late summer. This is a barrier spring, as the location of the spring outlet is predetermined by tectonic disturbances in the east. The water flows out through a coarse proluvial formation built mainly from boulders and marble pebbles, with minor participation of sand and clay. Another spring issues in the neighbourhood, draining the higher part of the phreatic zone in the same karst basin.

Spring no. 39a, Beden

This is a barrier spring. The spring outlet is located in the northern part of the outcropping Proterozoic marbles that are highly karstified. These marbles form the largest karst basin in the Rhodopes mountain: Nastan-Trigrad karst basin (Antonov & Danchev 1980). The basin is faulted into blocks and as a result several separate karst zones are formed. The recharge is mainly due to precipitation and loss of surface water including river runoff. The karst is of mountain type with large development of surface and underground karst forms.

Analysis method

The empirical and theoretical distributions of the annual spring discharges were studied. The method for analytical approximation and multi-criterial optimization of the empirical distributions proposed by Gerassimov (1988) was used. The main ideas of the method are as follow: (i) approximation of the empirical distribution functions by regression analysis of appropriately transformed coordinates; (ii) transformation of the empirical cumulative frequencies to normal distribution quantiles, and taking logarithms of the random variables; and (iii) choice of the most appropriate approximation. Table 1 presents water discharges for

Table 1. *Spring discharges (in m^3/s) for chosen probabilities of exceedance*

No.	95%	90%	50%	10%	5%
25	2.286	2.540	3.653	5.210	5.755
396	0.137	0.173	0.361	0.701	0.840
48	0.287	0.324	0.487	0.711	0.789
63	0.195	0.214	0.290	0.382	0.411
59	0.690	0.759	1.046	1.415	1.538
39a	0.434	0.492	0.736	1.052	1.158

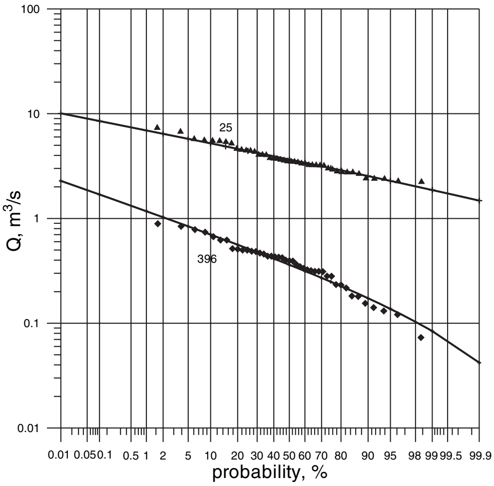

Fig. 2. Probability distribution curves for springs 25 and 396 located in the Danube basin.

selected probabilities of exceedance. In Figures 2 and 3 the probability distributions for some springs are given.

River runoff variability during the 1982–1994 drought period

A recent study (Gerassimov *et al.* 2001, 2004*a*, 2004*b*) concerned water resources in Bulgaria during the 1982–1994 drought period. An investigation was made of precipitation (*P*, mm) and river discharge (runoff depth, *h*, mm) for two periods: one with a duration of 13 years (the drought period) and the other with a duration of 106 years (1890–1995). In these studies the river discharge is calculated and used in total depth of runoff. The authors concluded that the drought period had a strong negative impact on river runoff in the country.

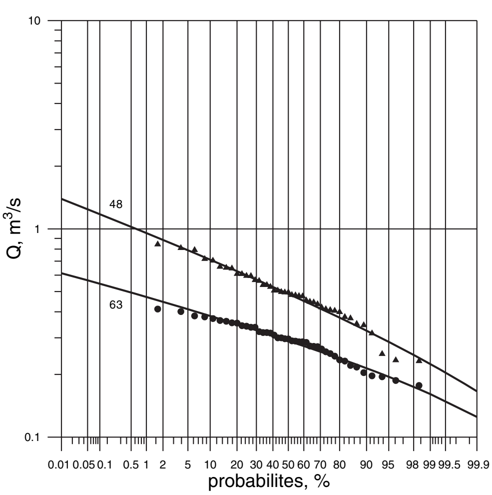

Fig. 3. Probability distribution curves for springs 48 and 63 located in the Black Sea basin.

The statistical descriptors of data such as the (arithmetic) mean ($\bar{X}$) and standard deviation (σ) of discharge and precipitation, with comparison to the long period of 106 years, are presented in Table 2 according to Gerassimov *et al.* (2004*a*). From Table 2 it is possible to conclude that during the drought period the average values and standard deviations are considerably lower compared to those of the long series. The reductions of these values are most significant for the Danube basin, followed by the Aegean Sea and Black Sea basin.

Groundwater regime variability during the 1982–1994 drought period

The orographical, climatic and hydrological conditions influence groundwater occurrence, recharge intensity and regime. Our previous studies concerning the drought period in Bulgaria confirmed the sensitivity of karst springs and shallow groundwater to climate variability (Orehova & Bojilova 2001; Gerassimov *et al.* 2004*a*). For these studies the examples of springs and observational wells from the NHGN were used.

The impact of the 1982–1994 drought period on groundwater was observed all over the territory of the country. The consequences of drought for the springs were comparable: 20–30% reduction of discharge. The results are similar for the springs draining karstic massifs of different geological ages: Precambrian marbles, Triassic dolomite and limestone, Upper Jurassic and Cretaceous limestone.

Table 2. *Statistical structure of discharge (*h*, mm) and precipitation (*P*, mm) of the drought period 1982–1994*

Drainage basin		$\bar{X}_{13}$ (mm)	$\bar{X}_{13}/\bar{X}_{106}$ (mm)	σ_{13} (mm)	σ_{13}/σ_{106}
Danube	*h*	98.3	0.638	36.1	0.642
	P	633.6	0.865	82.2	0.791
Black Sea	*h*	112	0.733	41.9	0.710
	P	598.7	0.93	68.1	0.600
Aegean	*h*	182.6	0.719	52.7	0.791
	P	658.7	0.873	67.2	0.626
Total for Bulgaria	*h*	138.2	0.695	41.7	0.747
	P	640.2	0.877	66.5	0.639

From Gerassimov *et al.* (2004*a*).

Table 3. *Deviations of average values (ε, %) for spring discharge and hydrological zones (river runoff) in relation to their 37-year values*

Object (spring or hydrological zone)	1960–1996*	1960–1981	1982–1994	1985–1994
Spring 25, Zlatna Panega	–	13.5	−21.7	−22.2
Spring 396, Musina	–	16.8	−28.1	−45.0
Danube hydrological zone	−6.0	20.5	−32.3	−39.0
Spring 48, Kotel	–	12.5	−20.5	−24.9
Spring 63, Malko Tarnovo	–	9.9	−15.9	−20.0
Black Sea hydrological zone	4.9	14.8	−29.9	−38.9
Spring 59, Jazo–Razlog	–	14.7	−22.9	−25.9
Spring 39a, Beden	–	14.6	−19.3	−30.0
Aegean hydrological zone	−3.9	14.6	−25.3	−34.0
Total river runoff	−3.9	17.0	−27.7	−35.8

*In relation to the period 1890–1995 (106 years).

The effect of the 1982–1994 drought period on the groundwater regime was studied (Orehova & Bojilova 2001). Representative springs and wells with long observation periods from the three main hydrological zones were chosen. Most of the observational stations refer to shallow groundwater with recharge from rainfall and snowmelt. A limited number of stations represents deep aquifers. In this study the influence of the drought period on shallow porous and karstic aquifers was investigated. These aquifers are potentially most vulnerable to droughts.

The processed data give evidence for a major drop of recharge to the aquifers in Bulgaria during the 1982–1994 drought period and also for a decrease of groundwater resources. The time series of discharge for springs and groundwater level for observational wells give important information on the groundwater regime in Bulgaria. Some of them have 40-year long observational periods and reflect the influence of climate variability on groundwater.

Assessments for karst springs

For quantitative comparative estimations, the periods investigated were the 37-year longer period 1960–1996, and three shorter periods of 22 years (1960–1981), 13 years (1982–1994) and 10 years (1985–1994). The percentage deviations of the shorter periods in comparison with the longer period are calculated. To make an assessment, deviations for the same periods for the river runoff in the three main hydrological zones are obtained. In Table 3 some results for the chosen karstic springs are given.

The chronological structure of the investigated periods in reduced variables $\psi = (X - \bar{X})/\sigma_x$ with reference to the values of mean and standard deviation ($\bar{X}$, σ_x) is presented for the three hydrological zones and selected springs in Figures. 4–6. From Table 3 and Figures 4–6 we draw the following conclusions.

- The shift of the groundwater regime in Bulgaria during the drought period is similar to that for

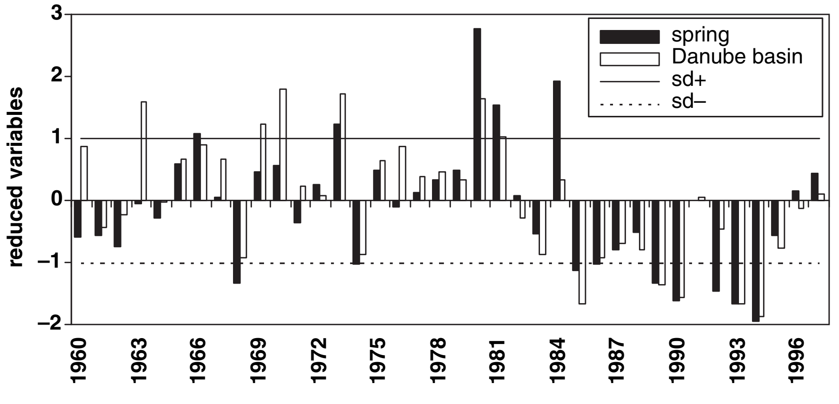

Fig. 4. Chronological graphs for spring 396 near Musina village and Danube hydrological zone.

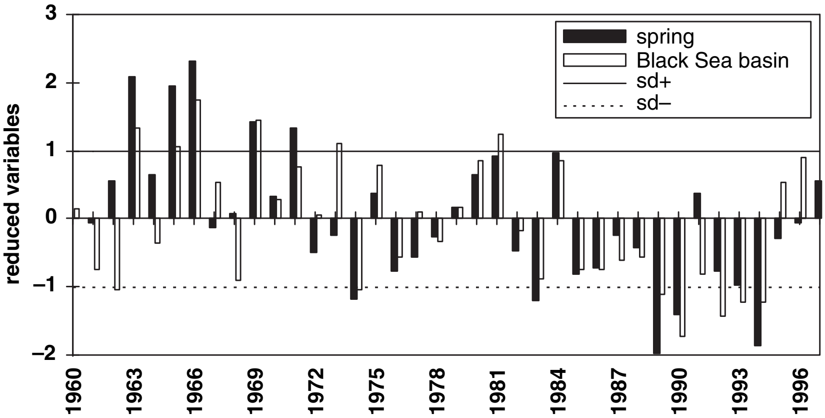

Fig. 5. Chronological graphs for spring 48 near Kotel town and Black Sea hydrological zone.

river runoff; the corresponding deviations are in the same range.

- The drought period 1982–1994 and especially the shorter component 1985–1994 are characterized by deep depression of the groundwater.
- The chronological structure of spring discharge and groundwater levels during the drought period resembles that of river runoff.

During the 1982–1994 drought period the recharge to the groundwater diminished. Orehova & Benderev (2004) applied two techniques to quantify the multi-annual recharge of Kotel spring basin. The reduction of recharge during the drought period is also assessed (around 70 mm). Such estimations require detailed knowledge of relief features, geological and hydrogeological settings, and hydrological data on springs draining the karst massif.

Despite the long observation period, no natural regularities can be settled now concerning the regime of springs in different hydrological and climatic zones. This inference is most likely related to highly varied conditions of selected karst basins.

From the two springs of the Danube zone, spring no. 396, Musina, is the most vulnerable to climate variability. We suppose that it is related to preponderant recharge due to precipitation. The precipitation in its drainage basin, having no possibility to generate surface runoff, percolates through the karst forms. This is the cause of the high vulnerability of karst waters in this region to drought. Permanent river recharge for the Zlatna Panega karst basin from the Vit River is a reason for the relatively high resistance to drought of the spring Glava Panega.

The two springs from the Black Sea basin show a similar reaction to drought. In both cases karst formations perform a retention function. This function is most visible in the region of Strandja mountain (Malko Tarnovo town), where the level of karstification is lower.

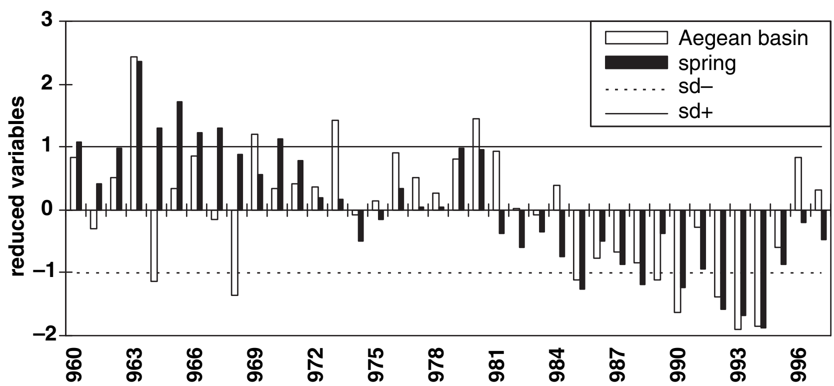

Fig. 6. Chronological graphs for spring 59 Jazo, and Aegean hydrological zone.

Table 4. *Absolute and relative declines of the groundwater levels for the drought period*

River basin	Observational well no.	H_{av} (m) 1960–2000	σ (m)	H_{av} (m) 1982–1994	ΔH (m)	$\Delta H/\sigma$
Tundja	271	2.09	0.351	2.38	−0.29	−0.827
Tundja	273	7.02	0.968	7.72	−0.70	−0.723
Maritza	526	7.86	0.329	8.08	−0.22	−0.699
Maritza	287a	2.92	0.655	3.43	−0.51	−0.779

H_{av}, multiannual average water level; σ, standard deviation of H_{av} (1960–2000).

The two springs from the Aegean Sea basin drain karst basins in mountain regions. In the case of Razlog karst basin, the flow distribution is stable during the whole year due to considerable snow cover and an extended period of snowmelt.

Assessments for observational wells

To investigate groundwater variability in shallow porous aquifers, observational wells in proluvial–alluvial formations were selected. Two regions from the central part of the country are presented. In Table 4 the examples of observational wells in Kazanlak kettle (Figure 7) and Upper Thracian valley are given. The numbering of the chosen wells is in accordance with the system used in NHGN. The wells are located in the Eastern Aegean basin under similar climatic conditions. The Kazanlak kettle occupies the inside graben orientated in a west–east direction and is bounded to the north by Stara Planina mountain and to the south by Sredna Gora mountain. The kettle is filled with proluvial–alluvial deposits in which an aquifer is formed. The groundwater level in the terrace is hydraulically connected to the Tundja River.

The Upper Tracian valley is the largest superimposed depression in Bulgaria. The valley is filled in with aquiferous Pliocene and Quaternary deposits of different genesis. The recharge is mainly due to river water, surface runoff and rainfall water. The Maritza River is the main collector in the valley. The Upper Tracian valley is characterized by more complicated conditions than the Kazanlak kettle.

For the chosen observational wells in proluvial–alluvial deposits the average drawdown of water level is in the range 0.2–0.7 m for the period

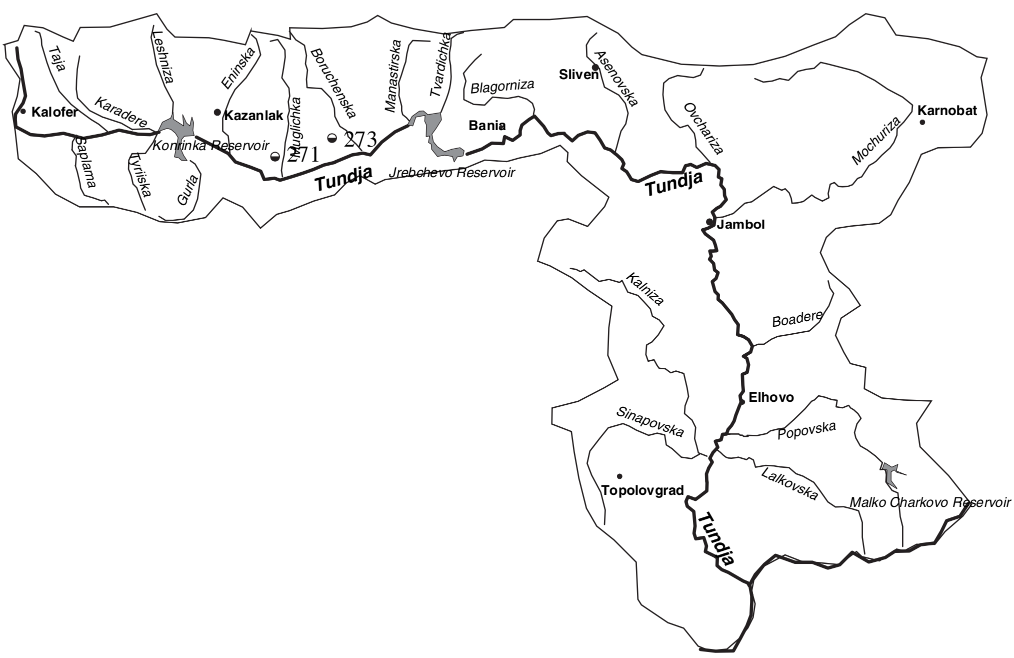

Fig. 7. Tundja River basin.

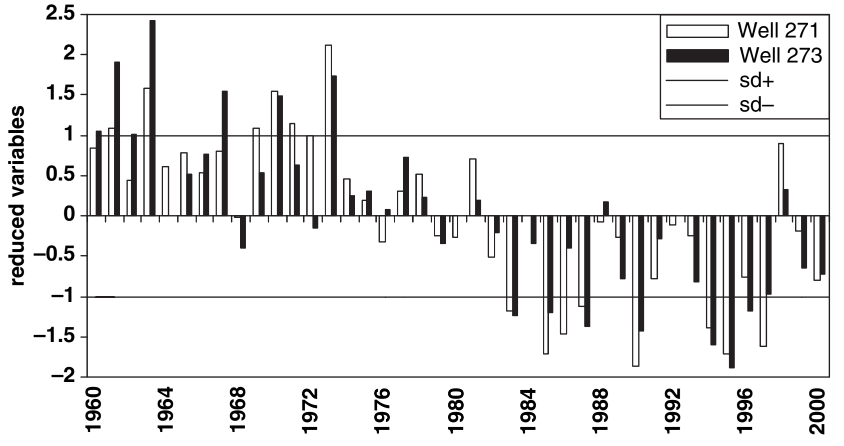

Fig. 8. Groundwater level variability for observational wells from the Kazanlak kettle, Tundja river basin.

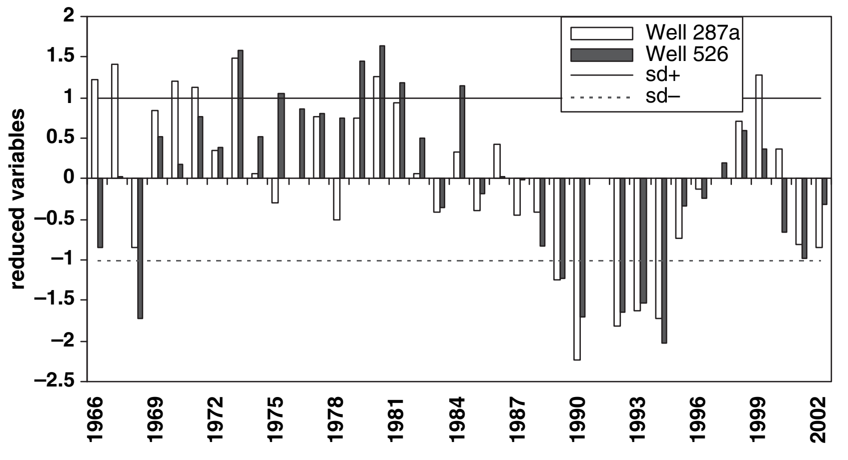

Fig. 9. Groundwater level variability for observational wells from the Upper Thracian valley, Maritza river basin.

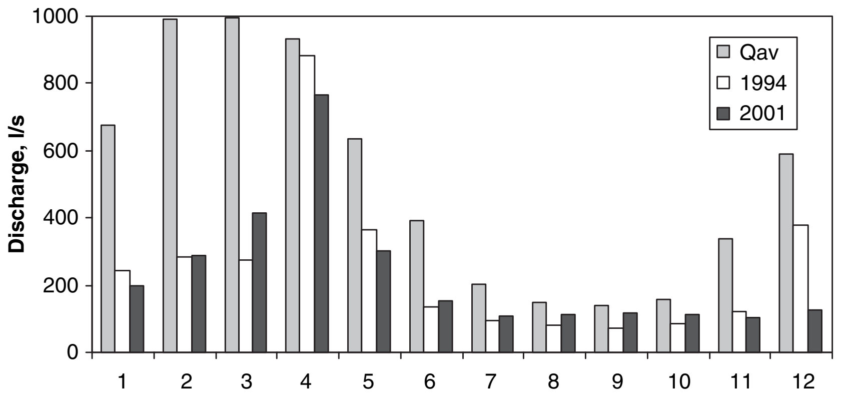

Fig. 10. Monthly distribution of discharges for two dry years (1994, 2001). The reference period for average monthly data (Qav) is 1961–1990.

1982–1994 (see Table 4). These values depend not only on reduction of recharge to the aquifer, but are subject to the influence of variable hydrogeological setting and parameters as well. In the last column of Table 4, relative deviations are presented in a comparable range. This feature can be seen in Figures 8 and 9, where the chronological structures of water level for selected observational wells are presented in reduced variables.

Mild winters and groundwater

Mild winters were analysed by Andreeva *et al.* (2003). Mild winters in Bulgaria, according to the classification given in the paper, are characterized by temperatures above normal, and precipitation about normal or below. Droughts during the cold period for the years had a strong negative influence on groundwater. Examples of monthly distribution of discharge for two dry years are presented in Figure 10 for hydrogeological station no. 48 (Kotlenski springs). The reference period for average monthly data (Qav) is 1961–1990.

Concluding remarks

Climate variability affected the groundwater regime in Bulgaria. Karst and shallow porous aquifers showed their vulnerability to long-lasting droughts. It seems that groundwater in karst regions was most affected by the reduction of recharge during such periods due to direct connection between surface and groundwater in many carbonate terrains.

The processed data give evidence for a major drop of recharge to the aquifers in Bulgaria during the 1982–1994 drought period and also for decrease of groundwater resources. In situations of water shortage the interest in available groundwater resources is enhanced for domestic and industrial water supply. It must be taken into consideration that groundwater, similarly to surface water, is vulnerable to droughts.

Our research does not concern deep aquifers, therefore it is not possible to establish a reliable conclusion for them. Owing to the limited number of observational stations and diversity of the hydrogeological setting, the effect of regionalization has not been studied.

Groundwater in Bulgaria is a widely used resource that is unfortunately vulnerable to drought. It is necessary to perform good management for sustainable water use. The aim of this type of research is to introduce the problems of the negative impact of drought periods to society. It is necessary to investigate the formation of groundwater resources and to study their vulnerability to droughts. The rational utilization of water resources should take into consideration surface and groundwater together.

Drought analysis is a major issue for drought policies and implementation of strategies for drought awareness. Work on drought analysis may be useful for water and social economic technicians, institutions or policy-makers because it gives the basis for studying impacts of droughts.

The authors acknowledge that the part concerning groundwater investigations was taken from Orehova & Bojilova (2001).

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Effects of climate change on groundwater resources in Campania (southern Italy)

D. DUCCI[1] & G. TRANFAGLIA[2]

[1]*Department of Geotechnical Engineering – "Federico II" University, Piazzale Tecchio, 80-80125 Naples, Italy (e-mail: daniela@unina.it)*

[2]*Department of Inland and Marine Waters Protection – National Agency for Environmental Protection – APAT, Via Curtatone, 3-00185 Rome, Italy*

Abstract: In order to estimate the influence of global climate change upon the hydrological regime, variations in the water budget prompted by precipitation and temperature changes were evaluated in the region of Campania (southern Italy). In many parts of the region, precipitation distribution in the last 20 years shows a marked reduction. During the same period, Campania also experienced a regional temperature increase of about 0.3°C. Water budgets, calculated in a geographical information system environment for the region's hydrogeological structures, show a mean decrease of 30% of average infiltration within the present climate scenario. The structures most affected are carbonate aquifers, with the flow of springs being significantly reduced (about 70 m^3/s). The most severely affected zones are the mountainous areas in the southern and northern parts of Campania.

Direct human-induced effects on the use of groundwater resources consist in the depletion of the groundwater table and the reduction in spring discharge, while indirect effects lie in contamination and in the reduction of surface water flow. A further effect on aquifer recharge and hence water levels in aquifers and springs is brought about by climate change. Indeed, climate change impacts on groundwater have been identified as a critical problem for groundwater managers. All in all, groundwater resource management in western Europe has to cope with increasing human pressure, hence increasing water demand, and, in almost all regions, a decrease in groundwater resources due to climate change.

In 2004 groundwater resources accounted for 86.4% of Italy's drinking water requirements and as much as 99.7% of such requirements in Campania. Groundwater management thus has to take due account of groundwater budgets and their possible changes so as to achieve a sustainable equilibrium between annual recharge and extraction.

At present, academics are widely involved in the task of ensuring the sustainable use of groundwater resources. For example the 2004 UNESCO–GRAPHIC project (Groundwater Resources Assessment under the Pressures of Humanity and Climate Change) deals with groundwater resources assessment and forecasting under the various pressures of human activities and climate changes.

Evaluation of the changes in groundwater recharge rates requires determination of a groundwater budget to establish the role of each climatic variable (P and T) in groundwater recharge. In fact, the basic groundwater balance equation – applicable when there are no interactions with other aquifers or surface waters (rivers, lakes, seas, etc.) and in the absence of exploitation and artificial recharge – generally defines groundwater recharge as the difference between precipitation and the sum of evapotranspiration and surface runoff. If the climate gauge network is dense, it is possible, working in a geographical information system (GIS) environment, to evaluate the effect of each climatic variable on recharge in space. Moreover, if the climate data series are long, the influence of each climatic variable on the recharge in time may be estimated. Both such evaluations have been performed in this paper to estimate in a simple fashion the role in recharge of decreased precipitation and of increased evapotranspiration due to warmer temperatures, and to highlight the negative balance between recharge and discharge in Campania, as testified by the recorded decrease in groundwater levels and spring discharge.

Lithological and hydrogeological setting

The region of Campania in the southern part of the Italian peninsula has an area of about 13 500 km^2, and a coastline along the Tyrrhenian Sea. The landscape is dominated inland by the Apennine mountain ranges reaching altitudes of 1000–2000 m, accounting for 32% of the land area. Coastal plains account for a further 18%, while the rest of

From: DRAGONI, W. & SUKHIJA, B. S. (eds) *Climate Change and Groundwater*. Geological Society, London, Special Publications, **288**, 25–38.
DOI: 10.1144/SP288.3 0305-8719/08/$15.00

the region consists of low-altitude hills and valleys. Geologically speaking, Campania consists of three coastal plains (of the rivers Garigliano, Volturno and Sele) filled by alluvial and pyroclastic deposits, of the Roccamonfina and Vesuvius volcanoes, pyroclastic hills in the Phlegrean Fields, and carbonate Mesozoic mountains surrounded by impervious arenaceous–clayey–marly sediments (Fig. 1) (Budetta *et al.* 1994; Celico *et al.* 2004*a*).

In the alluvial–pyroclastic plains, permeability is medium–high, chiefly depending on the sediment grain size. In the Volturno river plain, the main aquifer, confined or semiconfined, is located in the hydrogeological units (with sand and gravel grain size) located below the Campanian ignimbrite tuff. The aquifer is also recharged from the Mesozoic limestone mountains. Shallow aquifers have also been recognized in the coastal sector, along the course of the Volturno river, and in the southeastern sector.

Nitrate, manganese and iron are the most common groundwater contaminants in the area. Iron and manganese occur naturally, but high nitrate levels indicate severe contamination from human activities (application of nitrogen fertilizers and wastewater disposal).

In the southern part of the region, the Sele river plain contains a multilayered coastal plain aquifer (thickness >200 m) covered by 50 m of silty–clayey deposits. In the hills of the Phlegrean Fields and the volcanoes of Roccamonfina and Vesuvius the piezometric surface indicates a radial groundwater flow chiefly towards the aquifers of the adjacent plains. The Vesuvius (and the old Somma volcano) aquifer consists of lavas, scoria and pyroclastics: the groundwater body is prevalently unconfined and very deep below the soil surface. The main aquifers are found in the carbonate massifs, which have a very high permeability due to a well developed karstic network. The groundwater bodies reach the springs with copious

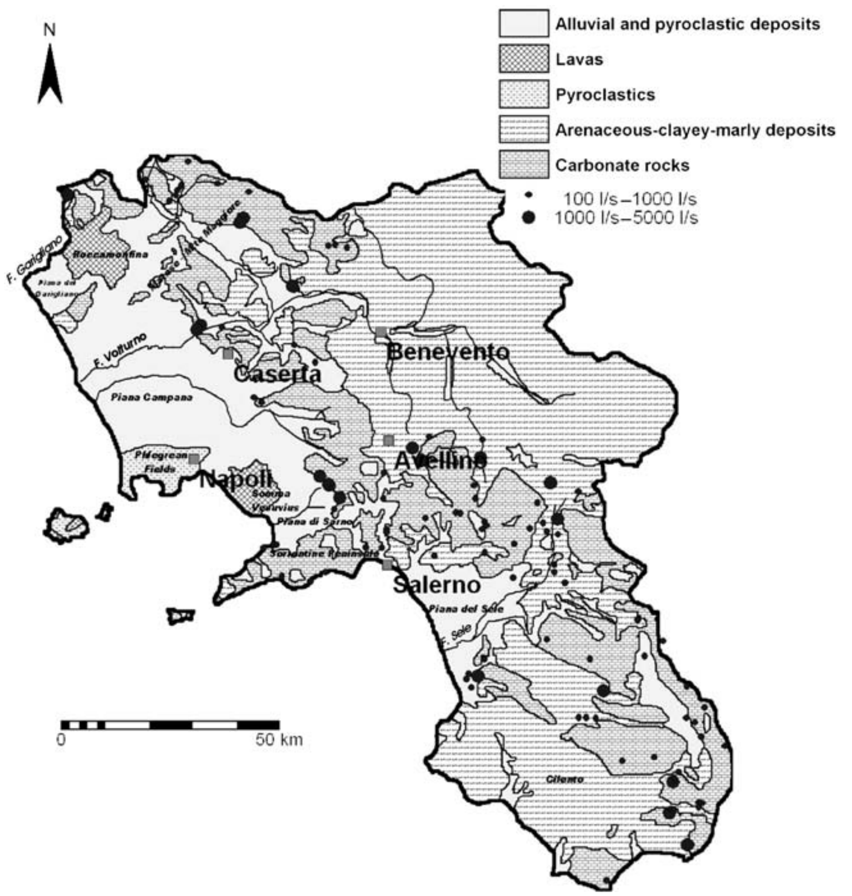

Fig. 1. Lithological map of Campania.

discharge (in some cases more than 1 m^3/s), located at the foot of the mountains, and feed the adjacent plains. The total spring discharge of Campania is about 70 m^3/s.

The regional climate gauge network

The regional rain gauge network has changed greatly in recent years as the mechanical rain gauge stations were replaced by digital stations funded after the tragic landslide event which took place in May 1998 in the area around the town of Sarno (>150 fatalities). The new digital rain gauge network became fully operational in the year 2000. The old network, with inadequate density for forecasting and warning in the event of landslides and floods, is very effective for analysing monthly and yearly data, as it represents a very large, continuous data set. Indeed, it comprises more than 250 stations distributed evenly throughout the area, 110 of which have almost complete data from 1951 and about 40 of which have rainfall data from 1920. The average elevation is approximately 350 m, exceeding 1000 m in only a few locations. By contrast, the region's temperature gauge stations are considerably less numerous, with very few having reliable, continuous data.

In this study a series of yearly and monthly climatic data up to the year 1999 (more than 100 years in some cases), were examined for 110 stations (Fig. 2). Finally, rainfall and temperature data were analysed over the time period 1951–1999. The reason for choosing this time period was mainly the availability of representative climatic data for the Campania area.

Climatic characteristics of Campania

The region of Campania has a Mediterranean climate, affected by the Azores, Siberian and South African anticyclones and the Aleutian and Icelandic lows, with hot, dry summers and

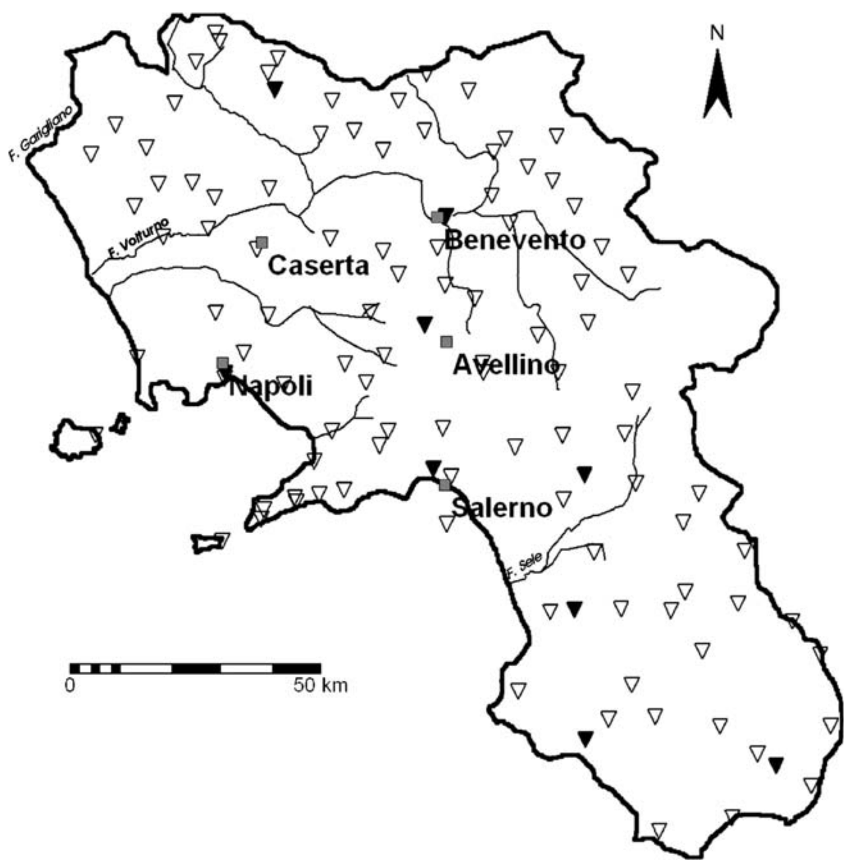

Fig. 2. Climate gauge stations examined in this study. Triangles represent rain gauge stations. Filled triangles represent the temperature and rain gauge stations.

moderately cool rainy winters. Mean annual temperatures are in the range of about 10°C in the mountainous interior, 18°C in the coastal areas, and 15.5°C in the plains surrounded by the carbonate massifs. In Campania the correlation between temperature and elevation is extremely high (generally >0.9), with a gradient of about -0.5°C to -0.7°C each 100 m.

The Italian rainfall regime consists of four different types: (1) Alpine continental; (2) Alpine sublittoral; (3) Apennine sublittoral; and (4) Marine. The rainfall regime in Campania is Apennine sublittoral, with a maximum in autumn/winter. Precipitation is influenced mainly by the mountain chains, in terms of elevation (often 1500–2000 m a.s.l.), location of ridges (barrier effect) and proximity to the Tyrrhenian Sea.

The lowest mean annual rainfall, about 700 mm, occurs in the eastern part of the region, on the other side of the Apennine watershed; the highest, about 1800 mm, occurs in the central part of the Apennine ridge.

Climate change analysis in Campania

Climatic phenomena are often the product of two or more, simple, interacting non-linear processes. As a result, chaotic processes in the atmosphere are extremely sensitive to small disturbances. Small variations in atmospheric turbulence can result in very different outcomes and then it becomes impossible either to measure the system accurately or to predict its future state (Bryant 1997). The actual atmospheric phenomena taking place over an area are the final stage of a number of different processes occurring on different scales, therefore the estimation of the areal distribution of meteorological parameters from point observations has been, and probably will remain, one of the most difficult issues within geophysics.

The occurrence of intense flooding causing landslides in autumn and winter in Campania depends on small cyclonic areas, the dynamics of which follow the genesis of tropical cyclones (hurricanes), but show a low level of energy (Tranfaglia & Braca 2004). Such meteorological systems, together with convective systems and orographic rainfall, can be intensified by the higher contribution of heat at the sea surface and often cause sudden flooding in coastal regions and in mountain regions exposed to sea winds.

A major challenge to climate researchers is to determine the degree of predictability associated with these and other events. The analysis of historical series of rainfall recorded in four large Italian watersheds shows that more than 50% of their interannual variance is significantly explained by the 22-year harmonics. It is proposed by Mazzarella *et al.* (2003) that the 22-year solar magnetic activity is able to influence the zonal circulation over the Mediterranean basin and the relative rainfall.

Mazzarella & Tranfaglia (2000) investigated the capability of the historical rain gauge network belonging to the Naples Hydrographic Service to measure the annual rainfall in Campania. They found that the value of the fractal dimension D was equal to 1.84, with a confidence level higher than 99%, within a scaling region enclosed between 8 km and 64 km with a dimensional deficit equal to 0.16 (2–1.84). A 50% lower limit value of the scaling region (4 km) represents the optimal value of the network resolution.

Analysis by deterministic and statistical methods

With a view to verifying the homogeneity of the historical series of annual rainfall data from the stations in Campania, a classic analysis of the double-mass curve was performed. The cumulative values of rainfall depths measured in each station were compared with the corresponding values of rainfall measured at the nearest stations.

For example, analysis between the stations of Naples University and Naples Hydrographic Service and between the stations of Naples Capodimonte and Naples University, besides evident homogeneity of the historical series, also shows a difference between rainfall values in the order of 5%. From the same analysis it becomes evident that between the stations of Naples Capodimonte and Naples Hydrographic Service there is significant homogeneity (constant slope of the double-mass curve), but also a substantial coincidence between the historical series (the curve of the double-mass is fitted by a straight line with a 45° slope).

On the basis of such considerations, a 182-year historical series of precipitation (from 1821 up to 2003) for the city of Naples was constructed, supplementing the data lacking in the Naples Capodimonte series with those of the Naples Hydrographic Service series (Fig. 3).

In order to verify the influence of extreme events on short-term trends (Burroughs 2001), regression analysis was performed based on the least-squares method. A scatter plot of annual precipitation versus number of rainy days was examined. A best-fit curve was sought to verify whether the data have a significantly linear trend. A good correlation between annual precipitation and number of rainy days was found ($R^2 = 0.78$, Fig. 4). It is important to establish a threshold for the significance of the correlation coefficient

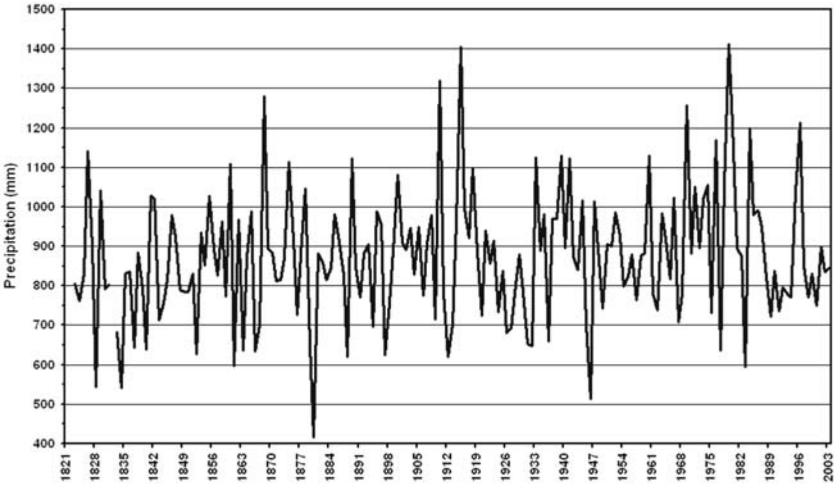

Fig. 3. Annual rainfall (mm/a) measured at Naples Capodimonte from 1821 to 2003.

(Swan & Sandilands 1995). A statistical test was carried out for all available data. The significance of the correlation coefficient was estimated using a t-statistic. The value of r can be tested against a hypothesis H_0 ($\rho = 0$). The data are taken without bias from a population which is normally distributed with respect to both variables. It was found that t equals 12.86 and critical $t = 1.67$, with $\alpha = 0.05$ and degrees of freedom $= n - 2 = 47$. As the calculated t exceeds the critical t, the null hypothesis is rejected and it can be stated that a significant correlation exists between annual rainfall and number of rainy days in Campania. The observation of the number of rainy days and rainfall data in the last decade shows that the hypothesis proposed by some authors (Alpert *et al.* 2002; Kitoh 2003) of a Mediterranean-wide decrease in the number of rainy days in respect to the rainfall rate in recent years, does not seem confirmed by the data from Campania region where there has been merely a slight levelling of the slope compared with the curve in Figure 4, and there has been no inversion in the trend.

In order to show the values of individual years that do not follow the signal (Bryant 1997), a curve was also sought to best-fit mean annual time series of precipitation in Campania. Confidence limits were fitted to the trend line to highlight the extreme values.

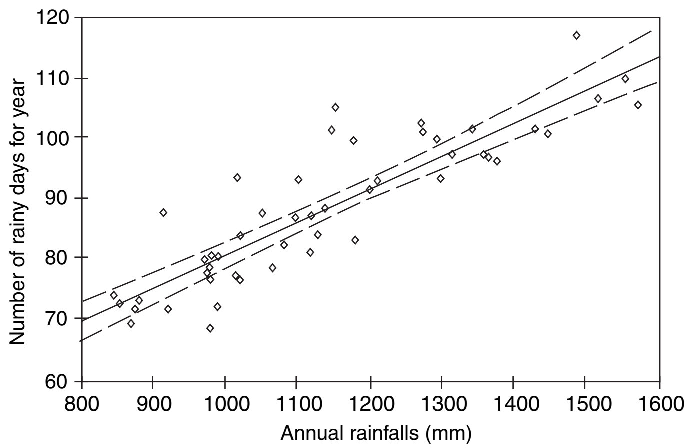

Fig. 4. Mean annual rainfall versus mean rainy days in Campania (estimated for 49 rain gauge stations from 1951 to 1999) with confidence bands for mean.

Impact of climate changes on groundwater resources

Spatial variation in mean annual rainfall

Input data included mean monthly and annual precipitation and temperature data from 110 rain gauge stations. Figure 2 shows the distribution of the rain gauge stations in the area. The correlation between precipitation and elevation is extremely low (0.1), as expected in such a large area. Indeed, it has been demonstrated (Guida *et al.* 1980) that the correlation $P-h$ is quite high only for small homogeneous areas (subzones), but is very low for a whole morphologically complex region (e.g. the Cilento area or Campania). The correlation between the annual rainfall at each station is extremely low (0.1), and between the annual rainfall of each station the standard deviation varies between 240 and 510 mm/a. The Pearson correlation values between the stations show the similarity of stations within a climatic zone rather than within the same mean elevation, according to the considerations previously expounded.

In this study to evaluate the variability of annual rainfall data, the rainfall rate was calculated for each year by interpolating the rain gauge data, using kriging interpolation techniques. The maps were constructed in a GIS environment, allowing the digital rainfall model (DRM) to be defined for each year (pixel: 100 m $\times$ 100 m).

Rainfall variability was observed by comparing the 49 yearly DRM raster maps (1951–1999) instead of operating as usual with point yearly data (related to each rain gauge station). The rainfall maps constructed by this method highlight the differences between areas. Moreover, the applied technique allows the problem of data paucity to

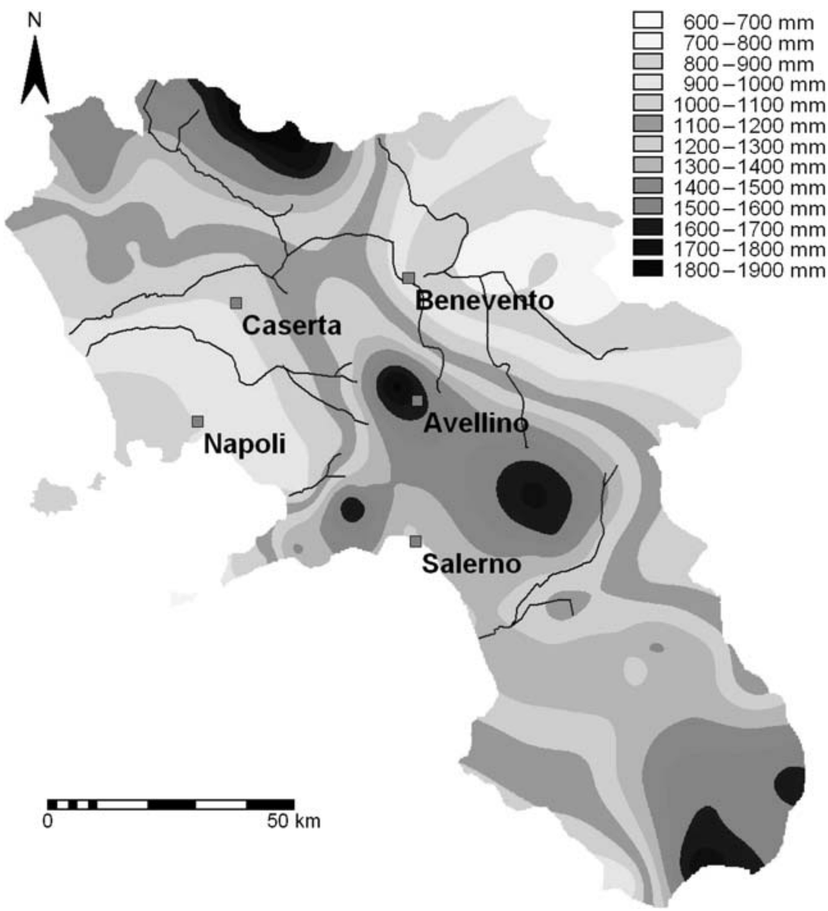

Fig. 5. Mean annual rainfall map (mm/a) for the years 1951–1980.

be overcome and minimizes the effect of occasionally rainy or dry years on the general rainfall pattern (Ducci 1999).

Statistical comparison between the 49 DRM highlighted the differences between the period 1951–1980 and the drier period 1981–1999. Figures 5 and 6 show the mean annual rainfall maps for the years 1951–1980 and 1981–1999, respectively. The rainfall maps were obtained by the average of the annual DRM raster maps for 30 years and 19 years, respectively. The rainfall maps constructed according to this method differ markedly in some sectors from the average rainfall maps obtained by interpolation of mean annual data.

For the whole region, the annual volume of rainfall is 16 000 Mm^3 for the period 1951 to 1980, and 13 500 Mm^3 for the period 1981–1999, with a decrease of about 15% (Fig. 7).

Variation in the evapotranspiration rate

The selection of the method to compute the evapotranspiration was based on data availability at regional scale. Detailed methods such as the Penman–Monteith equation, recommended by the Food and Agriculture Organization of the United Nations, were not feasible to apply due to lack of data (especially crop–soil data).

Therefore, annual evapotranspiration (Er, mm) was computed by using Turc's empirical formula (Turc 1961), as several studies have confirmed the reliability of this formula for central and southern Italy (Bono 1993; De Felice *et al.* 1993; Dragoni & Valigi 1991, 1993):

$$Er = \frac{P}{\sqrt{0.9 + P^2/L^2}}$$

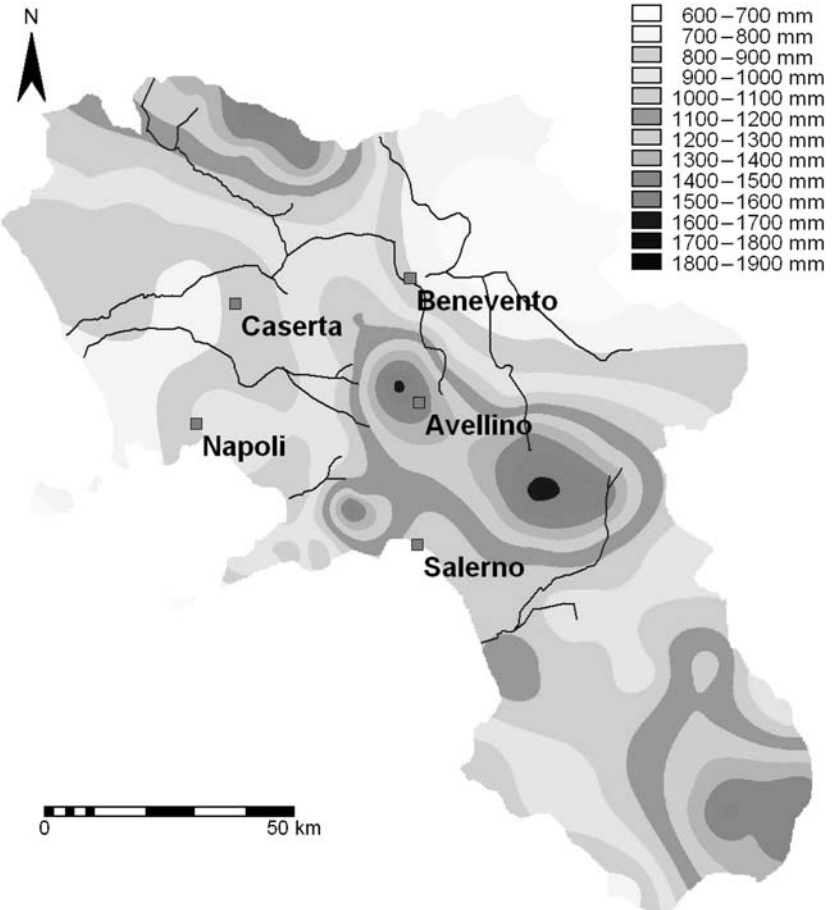

Fig. 6. Mean annual rainfull map (mm/a) for the years 1981–1999.

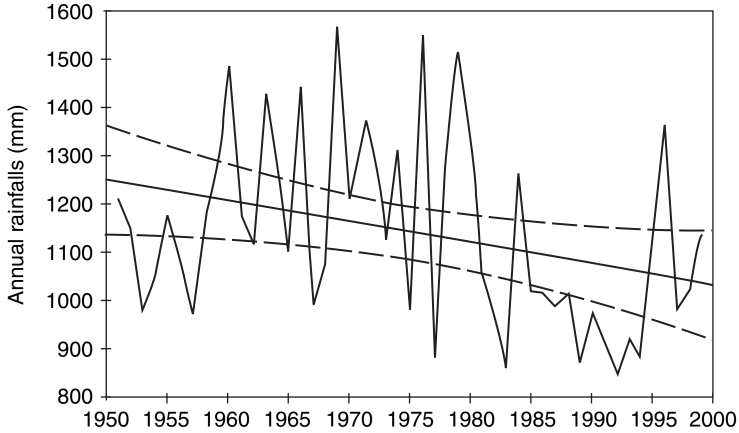

Fig. 7. Mean annual rainfall in Campania (1951–1999) with confidence bands for mean.

where: P = yearly rainfall (mm/a); $L = 300 + 25\ T + 0.05\ T^3$ (the capacity of the atmosphere to evaporate water); T = mean annual temperature (°C).

In the whole region the number of stations with reliable and continuous data is limited (about ten). Fortunately, the strong correlation between temperature and elevation allows the values of T to be combined with the digital elevation model (DEM) by linear regression. Hence the DEM is the best spatial descriptor for this climatic variable. The temperature maps were constructed in a GIS environment, for each year, from 1951 to 1999, on the basis of the DEM. The strong correlation between altitude and temperatures (0.9 in 62% and 0.8 in 93% of the years) emerges in Table 1,

Table 1. *Regression equation for each year between elevation and temperature and correlation factor R^2*

Year	Direction coefficient	Intercept	R^2	Year	Direction coefficient	Intercept	R^2
1951	−0.0068	18.020	0.9488	1975	−0.0068	17.090	0.8031
1952	−0.0070	18.393	0.9881	1976	−0.0070	16.717	0.8026
1953	−0.0086	18.256	0.9803	1977	−0.0065	17.448	0.8634
1954	−0.0085	17.625	0.9841	1978	−0.0067	16.591	0.8636
1955	−0.0075	17.858	0.9767	1979	−0.0075	17.717	0.9717
1956	−0.0079	16.936	0.9718	1980	−0.0077	17.168	0.9253
1957	−0.0069	17.847	0.9779	1981	−0.0074	17.694	0.9446
1958	−0.0070	18.013	0.9360	1982	−0.0068	17.806	0.8352
1959	−0.0083	18.414	0.8287	1983	−0.0061	16.807	0.7621
1960	−0.0079	18.538	0.9670	1984	−0.0066	16.747	0.8208
1961	−0.0099	19.095	0.9188	1985	−0.0069	17.918	0.8012
1962	−0.0065	18.568	0.8919	1986	−0.0073	17.794	0.7610
1963	−0.0076	18.371	0.9678	1987	−0.0069	17.884	0.6847
1964	−0.0095	19.048	0.9801	1988	−0.0068	18.093	0.6701
1965	−0.0070	18.270	0.9340	1989	−0.0075	17.681	0.9920
1966	−0.0086	18.734	0.8751	1990	−0.0074	18.835	0.9440
1967	−0.0081	18.592	0.8529	1992	−0.0077	19.097	0.9941
1968	−0.0064	18.047	0.8788	1993	−0.0079	19.218	0.9859
1969	−0.0077	17.926	0.9848	1994	−0.0084	20.301	0.9968
1970	−0.0074	18.036	0.9910	1995	−0.0082	18.136	0.8821
1971	−0.0075	17.681	0.9920	1996	−0.0085	18.357	0.8746
1972	−0.0072	17.475	0.9366	1997	−0.0083	18.907	0.9417
1973	−0.0070	17.289	0.8094	1998	−0.0073	17.604	0.9896
1974	−0.0067	16.826	0.8608	1999	−0.0074	18.849	0.9445

The year 1991 is not evaluated due to paucity of data.

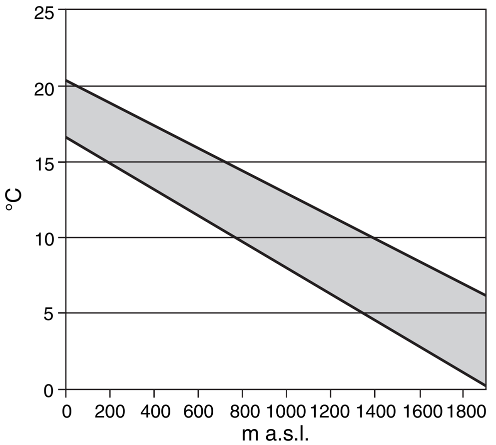

Fig. 8. Linear regression between elevation and temperature for the years 1951–1999.

where the regression equation for each year and the corresponding R^2 are reported. Figure 8 summarizes the trends of the linear regression with a medium gradient of about -0.7°C each 100 m.

The temperature variability was observed by comparing the 48 yearly raster maps of the temperature (1951–1999, excluding the year 1991 with only two stations working). The temperature maps highlight the differences between areas according to elevation: the higher increases in temperature are in mountainous areas and the lower in the coastal plains. Statistical comparison between the 48 maps allowed its to detect the differences between the period 1951–1980 and the warmer period 1981–1999. Figure 9a and b show the mean annual temperature maps for the years 1951–1980 and 1981–1999, respectively. Evaluating the real evapotranspiration by the Turc method, in the last 20 years the percentage of evaporated water is 6% greater than in the 30 years before, increasing from 52% to 58% of the precipitation volume, due to the increase in temperature.

The map of the variation in total flow (runoff + recharge) obtained from the difference between precipitation and evapotranspiration is very significant (Fig. 10). Moreover the areas with a decrease of more than 200 mm are mountainous areas where permeable rocks (limestone, lavas) crop out. The role of lithology in the infiltration decrease is outlined below.

Variation in infiltration

Due to the difference in the position of the groundwater divide and the watershed divide, it is almost impossible to evaluate runoff directly, using data from river gauge stations. Recharge is estimated as a percentage of the surface and subsurface runoff as a function of the permeability of the outcropping rocks. The recharge coefficient c_r ranges from 10–20% in impervious rocks to 100% in limestones with strong karstification. The lithological map of Campania (Fig. 1) was

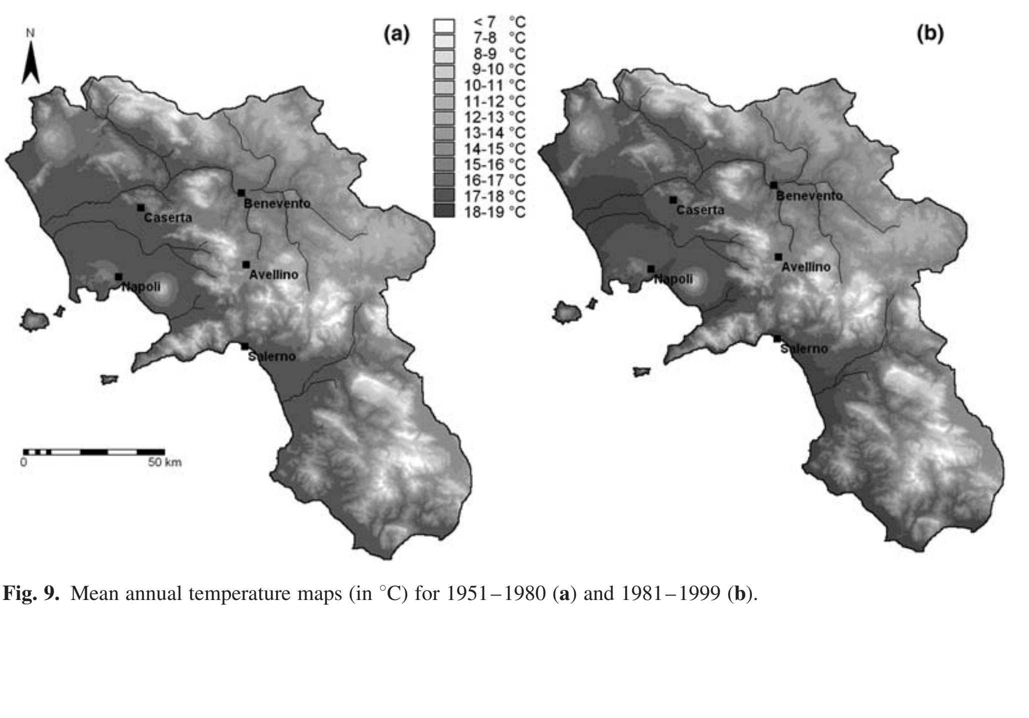

Fig. 9. Mean annual temperature maps (in °C) for 1951–1980 (**a**) and 1981–1999 (**b**).

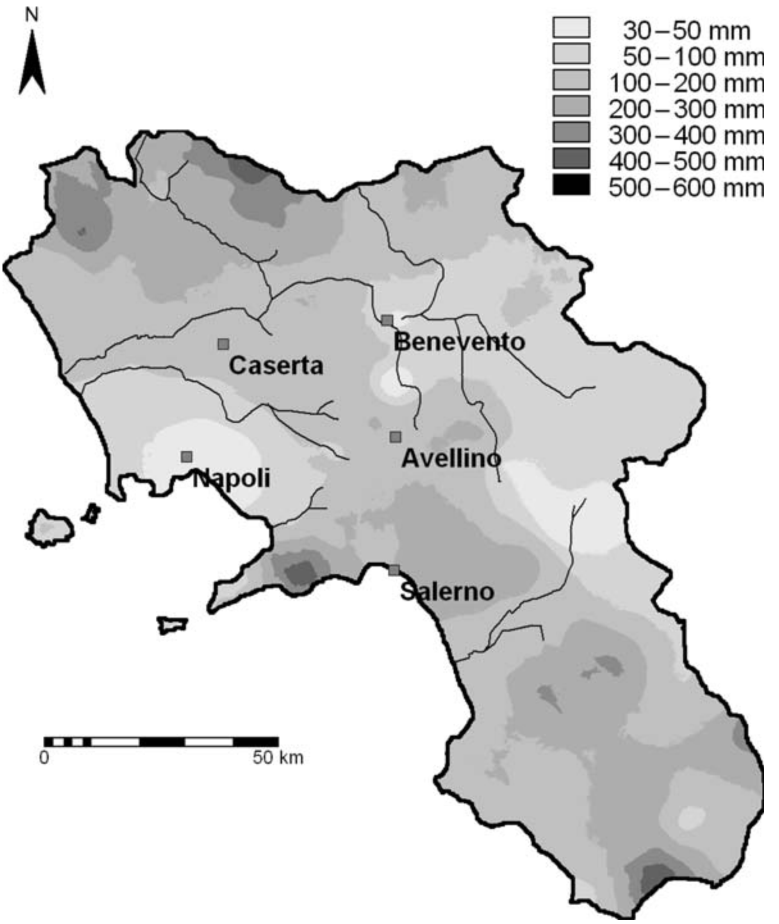

Fig. 10. Difference in mean annual flow (runoff + recharge; mm/a) between 1951–1980 and 1981–1999.

used to evaluate the recharge coefficient on the basis of the permeability of the mapped units (Table 2). The coefficients are multiplied (pixel by pixel) by the difference between precipitation and evapotranspiration to compute the value of the groundwater recharge.

Table 2. *Recharge coefficients of the main lithologic units cropping out in Campania*

Lithologic unit	Recharge coefficient (%)
Alluvial and pyroclastic deposits	60
Lavas	80
Pyroclastics	40
Arenaceous–clayey–marly deposits	20
Carbonate rocks (prev. Limestone)	90

In the last 20 years the volume of infiltration has been 30% lower than in the previous 30 years, due to the decrease in rainfall and to the increase, in percentage terms, in evapotranspiration resulting from the temperature rise of about 0.3°C. This is in agreement with previous studies carried out in central Italy (Cambi & Dragoni 2001) and in a basin in Campania (Ducci *et al.* 2000).

The recharge maps (Fig. 11a and b) highlight the differences between areas. The areas most severely affected by the recharge reduction, where the decrease sometimes exceeds 50%, are the Matese carbonate area (in the northern part of the region), the area around Avellino, the Sorrentine Peninsula carbonate area (located west of Salerno) and the Cilento area (mountainous coastal area in the southern part of the region). The least affected area is the zone surrounding the town of Naples. In terms of differences, and hence the amount of

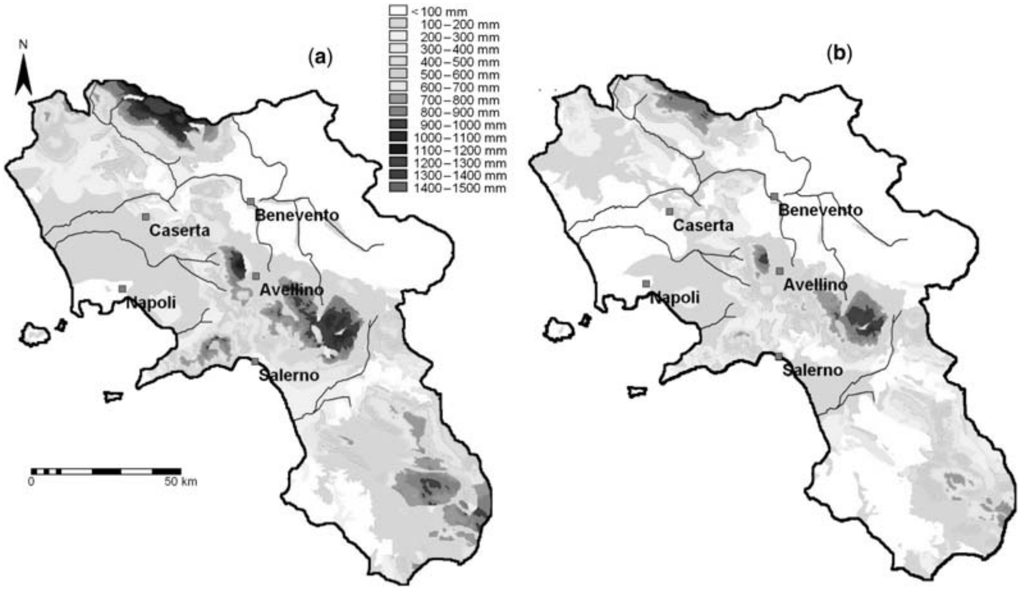

Fig. 11. Mean annual recharge maps (in mm/a) for 1951–1980 (**a**) and 1981–1999 (**b**).

recharge, the zones most affected by the reduction are often in the Apennine chain, as these carbonate areas are the 'freshwater reservoirs' of Campania (Celico *et al.* 2004*a*). In these areas there are few rain gauge stations, but this lack of data has relatively little effect in determining recharge since the recharge reduction was verified on stations at high altitude with very long, continuous data series (such as the Montevergine rain gauge station, at 1287 m a.s.l.). In order to highlight this aspect, the *recharge* values (mm/a) were multiplied by the recharge area to obtain the total amount of infiltration. The difference (Fig. 12) is 1163 Mm^3 (37 m^3/s).

Variation in spring discharge and piezometric surfaces

The impact of precipitation on groundwater levels is more or less delayed and attenuated by porous media (Kresic 1997). To define the normal hydrodynamic behaviour of an aquifer it is necessary to develop some stochastic models of the input–output type. Even the simplest stochastic model provides considerable information on the aquifer's structure and on the connections between hydrologic variables.

In Campania there are several long series of piezometric and discharge data. For instance, the Acerra Capomazzo well (in the coastal alluvial Volturno river plain), which is a low-depth (about 15 m) well in fine–medium grained pyroclastic deposits (shallow aquifer), has groundwater level measurements from 1926 until 2000 every three days (Italian Hydrographic Survey 1948–1999). As regards groundwater discharge, the springs of the Matese mountains in the northern part of Campania and the springs of Mt Terminio, with more than 1 m^3/s of discharge, have extensive records on a monthly basis.

In Campania in the last 10–20 years a sharp reduction in the discharge of many important springs at the foot of carbonate aquifers has been recorded, sometimes up to 50% of the previous discharge (Celico *et al.* 2004*b*). Moreover, in the coastal plains, bordered by carbonate massifs and hydraulically connected to them, during the last 15 years the fall in the piezometric level of the aquifers has been, at some points, as much as 6 m.

Such decreases should not be solely attributed to the recharge decrease, but also clearly depend on human activities and consequently on the increase in exploitation. Indeed, the increase in groundwater tapping may depend on a series of factors, such as the increase in population, the industrialization of the area, and changes in land use, but also on the increase in water demand due to the drier climate.

Hence in this research, piezometric and discharge data, although extensively available, were not used. Indeed, the link between recharge decrease and exploitation increase makes it impossible to establish a clear relationship between the

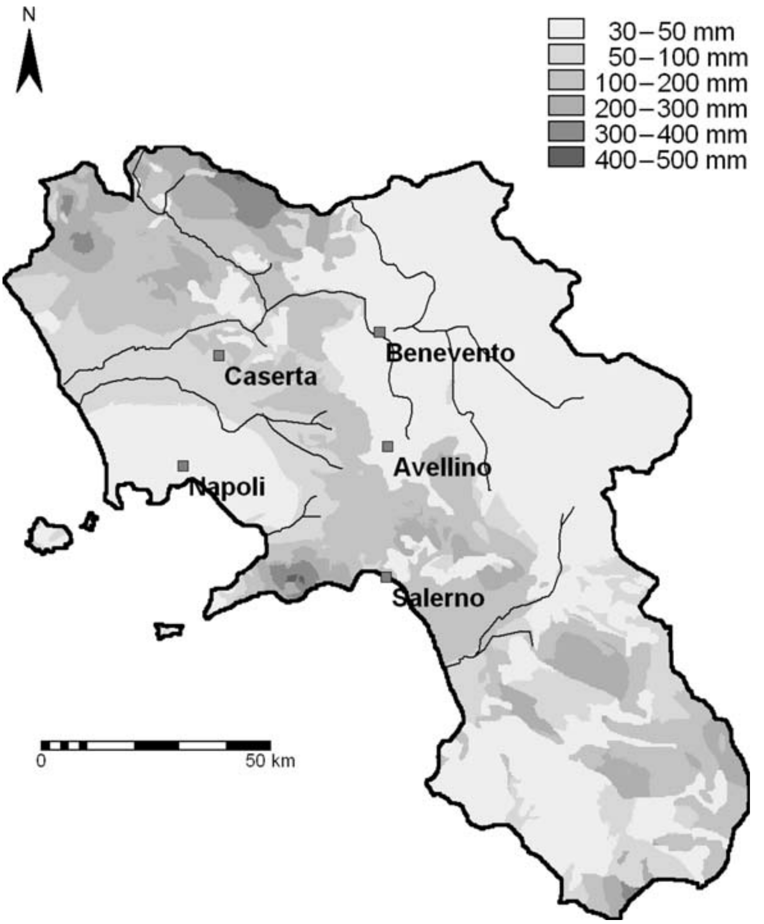

Fig. 12. Difference in mean annual recharge (in mm/a) between the years 1951–1980 and 1981–1999.

variations in piezometric and discharge data and the changes in precipitation/recharge.

Discussion and results

The analysis performed herein stressed the following points.

In the last 20 years precipitation in Campania has decreased by about 15%, although the reduction is not uniformly distributed throughout the region. The most severely affected areas are those at higher elevations. The substantial rainfall reduction is also evident from the trends at stations with rainfall data records from 1920.

The number of rainy days has decreased roughly in the same way as the rate of precipitation. Consequently, rainfall intensity is unchanged in time.

The analysis of temperatures demonstrates a clear increase, of an average of 0.3°C (from 0.2°C in coastal flat areas to 0.5°C in mountainous areas). These trends are in agreement with the results of the UK Hadley Centre global climate model, running on monthly climatic data from Mediterranean countries (i.e. HadCM2 in De Wrachien *et al.* 2002). In the southern Mediterranean, this model indicates a rainfall decrease of about 10–15% and a temperature increase of 1.5–2.5°C by the year 2050.

Spring discharge depletions and negative fluctuations of the piezometric surface are recorded in the area. These are important indicators of groundwater resource reduction, but they are also affected by the increase in water resource exploitation.

Concluding remarks

In Campania in the last 20 years, the precipitation distribution shows reduction in rainfall, and a mean regional temperature increase (0.2–0.5°C). The rise in temperature and decrease in precipitation have had a sequence of direct effects on the hydrological cycle, with particular regard to the evapotranspiration rate, soil moisture, surface runoff, and finally groundwater recharge.

In recalculating the water budgets in a GIS environment for the hydrogeological structures of the region with these climatic parameters, the average infiltration decreases by up to 30%. The average infiltration rate measured at the end of the 1980s was about 3850 Mm^3/a, while the average for the last 20 years is 2700 Mm^3/a, with a difference of about 37 m^3/s.

If this trend continues, it can be expected that in the next 50 years groundwater resources will decrease by about 70% and groundwater management in Campania will need to be reviewed. The water demand is currently about 1000 Mm^3/a. By 2050, under the previously described groundwater scenario, between 6 million (population of Campania) and 9 million people (combined populations of Campania and Puglia, now supplied by groundwater from Campania) will face a crisis.

The forecasts for the decrease in groundwater resources, albeit evaluated empirically (in future it would be wiser to model the hydrological budget in a more rigorous way), give an idea about the future scenario. It is essential to ascertain future groundwater resource availability since knowledge of water resources is a precondition for effective water management, in terms of both a reduction in water consumption and the search for alternative water supplies.

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Climate change, drought and groundwater availability in southern Italy

M. POLEMIO & D. CASARANO

CNR – IRPI, Department of Bari, Via Amendola 122 I, 70126 Bari, Italy
(e-mail: m.polemio@ba.irpi.cnr.it)

Abstract: Data for the period 1821 to 2003 from 126 rain gauges, 41 temperature gauges, eight river discharge gauges and 239 wells, located in southern Italy, have been analysed to characterize the effect of recent climate change on availability of water resources, focusing on groundwater resources. Regular data are available from 1921 to 2001. Many analysis methods are used: principal component analysis, to divide the study area into homogenous portions; trend analysis, considering the Mann–Kendall, Student-*t* and Craddock tests, autocorrelation and cross-correlation analyses, and seasonal, annual and moving-average variables, applying the spatial analysis to each variable with a geographical information system approach.

A widespread decreasing trend of annual rainfall is observed over 97% of the whole area. The decreasing trend of rainfall worsens or decreases as mean annual rainfall increases; the spatial mean of trend ranges from −0.8 mm/a in Apulia to −2.9 mm/a in Calabria. The decrease in rainfall is notable after 1980: the recent droughts of 1988–1992 and 1999–2001 appear to be exceptional. On a seasonal basis, the decreasing trend is concentrated in winter; a slight positive trend is observed in summer, the arid season in which the increase is useless as it is transformed into actual evapotranspiration. The temperature trend is not significant and homogeneous everywhere if the temperature increase seems to prevail, especially from about 1980. Net rainfall, calculated as a function of monthly rainfall and temperature, shows a huge and generalized negative trend.

The trend of groundwater availability is so negative everywhere that the situation can be termed dramatic for water users, due not only to the natural drop in recharge but also to the increase of discharge by wells to compensate the non-availability of surface water tapped by dams, as a direct effect of droughts.

The significant meteorological factors behind droughts include not only rainfall, temperature and evapotranspiration, but also atmospheric circulation patterns (EEA 2001).

Considering Europe and particularly the Mediterranean area, low pressure generally settles over Iceland, and high pressure over the Azores involving the Mediterranean area; this feature occurs quite normally in the summer, very often from March to October. A change in position and/or duration and intensity of anticyclones leads to rainfall/temperature anomalies. A common feature of Mediterranean droughts is the persistence of high-pressure systems. In the Mediterranean area, drought seems to be teleconnected to La Niña, an anomalously high cooling of the equatorial Pacific Ocean; out of 14 La Niña events, occurring between 1865 and 1990, 13 were associated with droughts in the Mediterranean area (Conte & Colacino 1994).

The Mediterranean Oscillation, as is the case for the North Atlantic Oscillation, is defined as the normalized pressure difference between two stations, in the case of the Mediterranean Oscillation Algiers and Cairo; the considered variable is the 500 hPa surface height. Piervitali & Colacino (2002) show the Mediterranean Oscillation is anticorrelated to rainfall in Italy in the period 1951–1995.

The spectral analysis of Italian temperature and rainfall series shows short- (two years), medium- (eight years) and long-term (14–26 years) oscillations caused mainly by solar cycles, according to Nanni & Lo Vecchio (1997), and also by the Atlantic ocean–atmosphere oscillation, according to Brunetti *et al.* (2000).

In the Mediterranean countries, drought is often the result of a sequence of dry years (EEA 2001). Considering the central-western Mediterranean basin, Piervitali & Colacino (2002) detect a trend towards a drop in rainfall equal to −3.2 mm/a from 1951 to 1995 while Piervitali *et al.* (1997) describe an increase in temperature of 1°C during the period from 1860 to 1995, higher than recognized on a global scale.

If Italy is considered, this pattern is confirmed but a distinction can be drawn between northern and central-southern Italy (Brunetti *et al.* 2004). The Italian climate has become warmer and drier, especially in the south, with an increase of both heavy precipitation events and long dry spells, as

From: DRAGONI, W. & SUKHIJA, B. S. (eds) *Climate Change and Groundwater*. Geological Society, London, Special Publications, **288**, 39–51.
DOI: 10.1144/SP288.4 0305-8719/08/$15.00

revealed by a study of more than a century of measurements up to 2000. Brunetti *et al.* (2004) show the temperature trend is positive for every season in the south and for autumn and winter in the north; in the former case the annual temperature trend is equal to 0.7°C/100 years and the winter trend is equal to the maximum, as shown also by Nanni & Lo Vecchio (1997) on the basis of a study of data from 1866 to 1975. The southern annual rainfall trend is −1 mm/a, about double that of the northern trend, and these trends, seem more or less null only for the winter season.

Other authors determine the annual rainfall trend equal to −2.2 mm/a in Italy, considering data from 1951 to 1996 (Brunetti *et al.* 2001) and equal to −4 mm/a in the same areas of southern Italy (Brunetti 2002).

Temperature and precipitation trends seem anticorrelated in Italy, as observed by Brunetti *et al.* (2000), using seasonal temperature and rainfall data for 1866 to 1995, and by Cambi *et al.* (2000), using also proxy data of lakes during the last 3000 years; Cambi *et al.* have evaluated the linear gradient ranging from −130 to −40 mm/°C, using regularly monitored data.

Some droughts can be man-made via mismanagement of the resources. Physical and human driving factors include the storage of catchments and aquifers and socio-economic factors controlling water demand. In southern Europe water consumption climbed from 7.1 km^3/year in 1900 to values 15 times higher in 1995, and further increases are expected in the years ahead (16.5 times higher in 2010; Shiklomanov 1999).

With regard to Italy, Brunetti *et al.* (2004) observed the drought worsening from 1980 onwards, studying data from 1950; in the drought period of 1988–1990 a deficit of 43% in Italian total rainfall was recorded (EEA 2001). Cambi *et al.* (2000) observed the effect of droughts on two mountain springs uninfluenced by human activity, evaluating the annual trend of spring discharge equal to 7% (1942–1991) and to 19% of mean discharge (1974–1995).

Future scenarios could be worse: by 2050, annual rainfall is expected to increase in northern Europe and decrease elsewhere in Europe. Temperature and potential evaporation will rise everywhere, with a huge impact on the driest regions of southern and eastern Europe (EEA 2001). Broadly stated, it seems that simultaneous drops in rainfall, increasing evapotranspiration and water demand are occurring in this period in southern Europe, contributing to groundwater resource depletion, decreasing piezometric levels and affecting several related environmental issues, such as seawater intrusion, contamination of land and water, and desertification.

All studies previously cited are characterized by low gauge density or by considering relatively small areas. This study focuses on the whole of southern Italy, involving the Apulia, Basilicata, Calabria and Campania regions (Fig. 1), where the mean annual precipitation (MAP) is equal to 901 mm.

This study is based on rainfall and air temperature (hereafter referred to as simply 'temperature') monthly data from 126 gauges, river discharge data from 8 gauges, and piezometric data of 239 wells. The main purposes are to determine the existence of a trend of these variables, to determine the role of drought periods, and to describe the contribution of variation of the water cycle on the availability of groundwater.

Climatic data and spatial approach

Rainfall and temperature monthly time series of the Italian hydrological service (Servizio Idrografico e Mareografico Nazionale, SIMN), have been considered (SIMN 1916–2000). A total of 126 rainfall gauges were finally selected from among 817 gauges (SIMN 1976) (Fig. 1). The gauges were selected to obtain a sufficient gauge density and spatial continuity mainly of rainfall and secondly of temperature, covering the maximum monitoring period with the minimum of data gaps; 41 of the selected gauges were also temperature gauges. The unpublished data are available courtesy of the Naples, Bari and Catanzaro SIMN departments. Data before 1915 were collected by Eredia (1918). The time series can be considered almost complete only from 1921 to 2001, the so- called main study period (MSP). Climatic time series utilized by previous authors (Polemio & Casarano 2004; Polemio *et al.* 2004*a*) have been enlarged, reducing gap percentages, and have been improved afterwards by homogeneity evaluation.

Residual gaps of time series are filled using multiple regressions based on a selection of the best correlated data series of the nearest gauges. The multiple regressions were performed on the normalized deviation of the considered value from the annual mean of the same time series, considering the best-correlated time series ($r > 0.7$) of the nearest gauges (up to six).

The spatial analysis of each variable is carried out by interpolation of point values (gauges) in a Geographical Information System (GIS) environment, operating with a 1-km spaced grid. The mean annual precipitation (MAP) plot, as in any case of variables related to rainfall, has been calculated in the MSP for each cell by weighting data from the 12 nearest gauges, with weights proportional to the inverse square of the distances (Fig. 1).

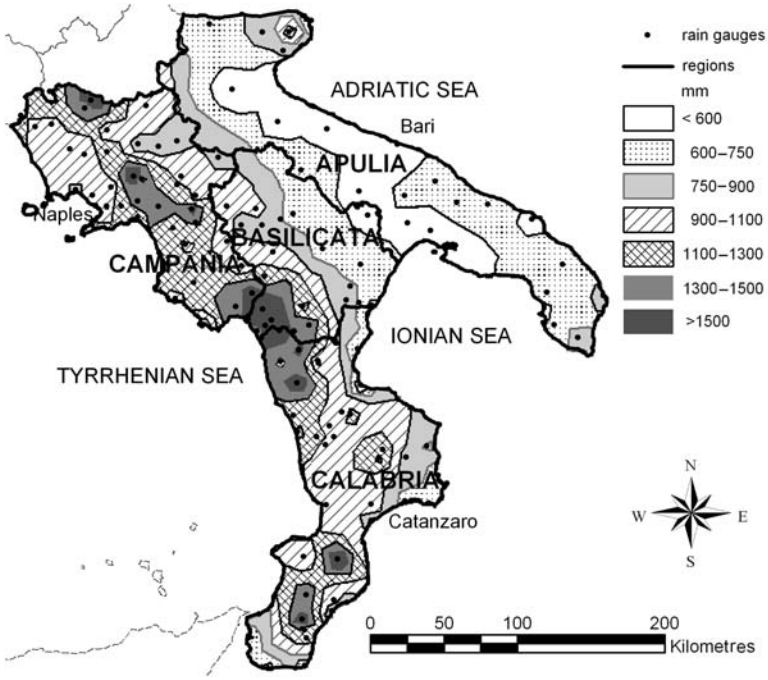

Fig. 1. Studied area, selected rain gauges and mean annual precipitation of main study period.

The climatic homogeneity of the study area has been evaluated on the basis of principal component analysis applied to the deviation of annual rainfall from the average normalized by the respective standard deviation; each time series or gauge is considered an individual while each year is a character or component.

Two main homogeneous climatic areas (HCAs) can be recognized, located along the Tyrrhenian and the Ionian–Adriatic coasts respectively (Fig. 2). The former can be divided into two sub-areas, Campania and the Tyrrhenian Basilicata–Calabria. The latter includes the Ionian Basilicata–Calabria and all of Apulia. Transitional sub-areas between the Tyrrhenian and Ionian–Adriatic influenced areas can be considered sub-areas of the SW portion of Calabria and Inner Basilicata.

The principal component analysis result is coherent with the spatial rainfall distribution in southern Italy (Fig. 1), mainly influenced by humid air circulation and by its arrival along the Italian coasts across the Mediterranean sea, moved by winds of the third and fourth quadrants and secondly by altitude and by the Apennine range which is located along the Tyrrhenian coast in Calabria and Basilicata and inland in Campania.

The statistical and spatial calculations (Table 1) have been performed considering both administrative regions (Fig. 1), where local governments are entirely in charge of water cycle management, and homogeneous climatic areas (Fig. 2). The results are quite similar; they are described here preferring the administrative zonation in order to enhance the message to water cycle managers.

Changes of climate, droughts and net rainfall

Time and spatial variability of annual rainfall

The annual rainfall trend is determined as the Angular Coefficient (AC) of the least-square line for each rainfall time series in the MSP. The increasing trends or positive values of AC are typical only of 12 of the whole 126 series; the maximum observed slope is about 2.5 mm/a. Decreasing trends are observed for 114 series (90%); the minimum is about −9 mm/a. If a 5% significance level is considered for correlation coefficients, 60 negative trends are found versus only two positive trends.

There are 17 time series available before 1921 (three before 1829) and they are located in Campania and Apulia (Polemio & Casarano 2004). If the start of the study period is moved back with respect to the MSP, in Campania a downward trend in rainfall

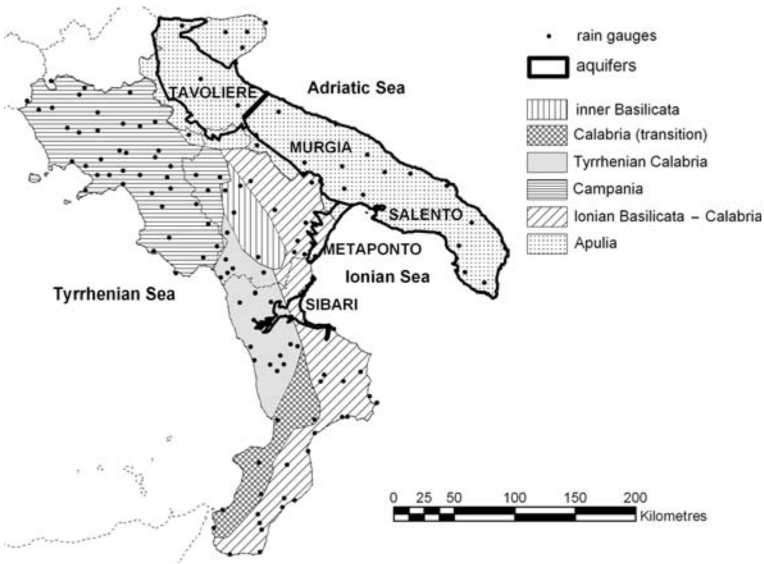

Fig. 2. Homogeneous climatic areas as classified by principal component analysis, rain gauges and aquifers.

is not evident while in Apulia a slight but almost continuous downward trend in rainfall is quite evident, even in the nineteenth century.

The results of the MSP are consistent with previous studies if average spatial values are considered (Brunetti 2002; Cambi *et al.* 2000; Piervitali & Colacino 2002). The higher density of gauges used in this study implies that the trend range is wider and the determination of extreme values is more accurate.

The spatial analysis of AC shows that 96.8% of the study area is affected by a negative trend (Fig. 3). Considering MAP and AC values as cell attributes, the spatial average of AC or the trend values for MAP class highlight the fact that the rainfall trend worsens or decreases as the MAP increases (Table 2). This figure is extremely worrying in the context of water management because high MAP areas are wide Apennine portions of the drainage basins of the artificial lakes which guarantee a relevant percentage of water supplies.

The reliability of detected rainfall trends has been evaluated by the Mann–Kendall test (Mann 1945; Kendall 1975). The Mann–Kendall variable S is:

$$S = \sum_{i=2}^{k} \sum_{j=1}^{i-1} \operatorname{sgn}(z_i - z_j)$$

where z_i is the rainfall of the i-th year of the considered gauge z, and k is the number of data or dur-

Table 1. *Regions or HCA and rainfall*

	MAPR (mm)	TR (mm/a)	PVR (mm)	PVR/MAPR (%)
Region				
Apulia	644	−0.80	−65	−10.1
Basilicata	893	−1.81	−145	−15.9
Calabria	1043	−2.87	−230	−22.0
Campania	1118	−2.44	−196	−17.5
HCA				
Apulia	650	−0.82	−66	−10.2
Inner Basilicata	986	−1.66	−133	−13.5
Tyrrhrnian Calabria	1262	−2.90	−232	−18.4
Calabria (transition)	1054	−3.12	−250	−23.7
Ionian B.-C.	865	−2.33	−186	−21.5
Campania	1118	−2.39	−191	−17.1

For each row the mean is determined considering the MSP: MAPR, MAP of a region; TR, precipitation trend; PVR, precipitation variation due to the trend and to the duration of MSP.

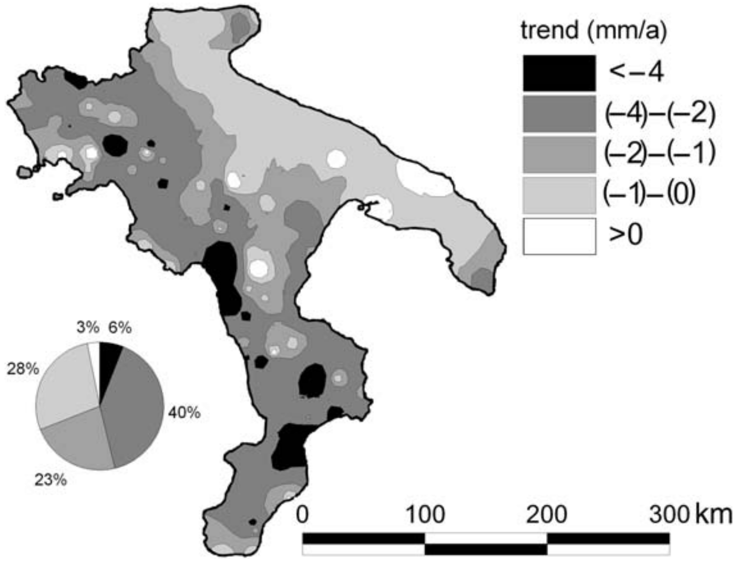

Fig. 3. Trend of annual rainfall as angular coefficient of regression line and pie chart of rainfall trend classes and area of interest in the main study period.

ation of the time series. S is distributed with null mean and variance σ^2 function only of k. This statistic, normalized to the respective standard deviation, highlights a negative trend with regard to 98% of the area, with the Mann–Kendall variable lower than average for more than one standard deviation over 75% of the area, and more than two standard deviations over 39%. It is clear that there is a relevant and generalized downward rainfall trend in the MSP.

The Mann–Kendall test has been improved taking into consideration three more detailed approaches (Douglas *et al.* 2000; Hirsch *et al.* 1982; Polemio *et al.* 2004*a*): pre-whitened series, trend estimator and spatial correlation calculation.

Pre-whitened series may be necessary since the Mann–Kendall test formulates the hypothesis that the data of a time series are independent and not autocorrelated. If the time series is autocorrelated a false trend could be detected. Each time series was then pre-whitened, obtaining a new series with null autocorrelation typical of a white process, and subjected to the Mann–Kendall test again. Widespread negative values of the Mann–Kendall variable and downward rainfall trends were substantially confirmed.

An extension of the Mann–Kendall test also allows an alternative and independent estimate of the trend. It is defined, for each rainfall time series, as the median of the slopes $d_{ij} = (z_i - z_j)/(i - j)$ of any combination with i and j equal to 1, 2, ..., k and $i \neq j$. This estimator, being based on a median, is more 'robust' with respect to extreme or anomalous values in the time series. The results are substantially similar to these obtained with the standard approach. The spatial correlation of the m rainfall time series allows the number of 'equivalent' independent rain time series to be estimated. It is useful to estimate the expected variance of a regional average for the Mann–Kendall variable, and then the significance of the trend over a wide area.

If S_i is the Mann–Kendall variable value for the ith rain gauge and all the time series have the same length, then the variance is equal for all the S_i, and a regional average $\bar{S}$ for the Mann–Kendall variable can be calculated. If the m time series are not correlated with each other, then the variance of $\bar{S}$ would

Table 2. *Spatial average of angular coefficient (SAAC) of rainfall straight line trend for MAP class areas*

	MAP class (mm)						
	<600	600–750	750–900	900–1100	1100–1300	1300–1500	>1500
SAAC (mm/a)	−0.64	−1.00	−1.89	−2.38	−2.64	−3.01	−4.74

Table 3. *Distribution of the Mann–Kendall variable for the 41 temperature time series*

	>0	$>\sigma$	$>2\sigma$	<0	$<-\sigma$	$<-2\sigma$
Original series	21 (51.2%)	16 (39.0%)	11 (26.8%)	20 (48.8%)	9 (21.9%)	7 (17.1%)
Pre-whitened series	24 (58.5%)	11 (26.8%)	2 (4.9%)	17 (41.5%)	8 (19.5%)	3 (7.3%)

simply be σ^2/m. Since time series are actually correlated, with ρ_{ij} correlation coefficient between the rain gauges i and j, it is possible to define an 'equivalent' gauge number m_{eq}:

$$m_{eq} = \frac{m^2}{m + 2\sum_{k=1}^{m-1}\sum_{l=1}^{m-k}\rho_{k,k+l}}$$

The variance of the regional average $\bar{S}$ will then be $\sigma_s^2 = \sigma^2 m_{eq}$.

The selected rain gauges have been divided into three groups on the basis of HCAs and administrative boundaries: Campania (41 gauges), Apulia (28 gauges) and Calabria–Basilicata areas (56 gauges). If the original (not pre-whitened) series are considered, m_{eq} is 1.95 for the Campania group, 1.85 for the Apulian group and 2.54 for the Calabria–Basilicata group. The regional average $\bar{S}$ is, respectively, $-2.81\ \sigma_s$, $-1.17\ \sigma_s$ and $-3.59\ \sigma_s$. For the pre-whitened series, m_{eq} is 1.84 for Campania, 1.85 for Apulia and 2.47 for Calabria–Basilicata, and $\bar{S}$ is, respectively, $-2.26\ \sigma_s$, $-1.18\ \sigma_s$ and $-3.21\ \sigma_s$. It can be assessed that the negative trend is only slightly attenuated if the autocorrelation is considered, and thus the substantial statistical relevance of trend results is confirmed.

The existence of relevant rainfall variations in periods shorter than MSP can be better highlighted with the moving average analysis: the deviation from MSP average of moving averages of decreasing duration (5, 3 and 2 years) is considered (Fig. 4).

The 5-year duration allowed the evaluation of significant deviations from the average over long periods. Dry periods were recorded in Apulia for 1942–1950, 1988–1992 and 1997–2001; the last two periods were almost the driest periods in Basilicata, Calabria and Campania.

The 5-year average deviation has been continuously negative since 1978 in Basilicata and Calabria and since 1983 in Campania, whereas the negative deviation in Apulia was observed from 1980 to 1995. The analysis of 3-year and 2-year moving averages shows the drought duration of 2 or 3 years is longest from 1980 as from this year the minimum rainfall has been reached and has been exceeded one or more times in each considered region, particularly with the latest drought of 1999–2001. In Apulia and Campania, some dry periods of the late 1920s and 1940s were as dry as the latest droughts.

The decade's average analysis also highlights a persistent succession of low-rainfall years and drought periods from about 1980; the results are consistent with those of other authors, obtained using different time series and lower data density (Brunetti *et al.* 2004). It can be hypothesized that the observed downward trend in rainfall is strongly influenced by low rainfall observed after 1980. On a

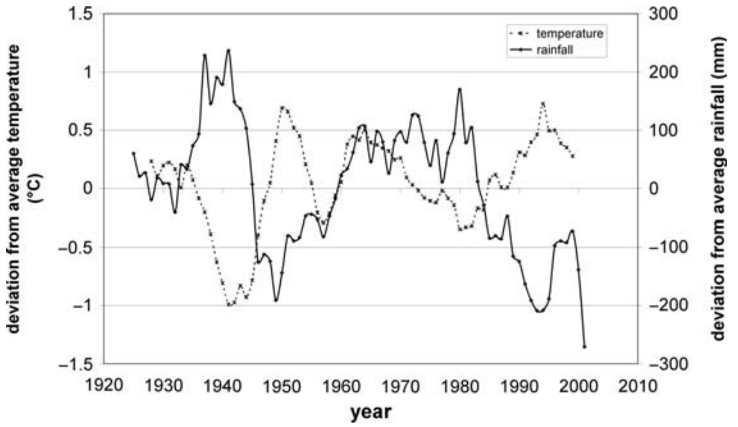

Fig. 4. Regional moving average of 5-year annual temperature and rainfall, expressed as deviations from mean values.

HCA basis, the 1981–2001 average is lower than the 1921–1980 average of 10% in Apulia, 14% in Basilicata–Calabria and 16% in Campania.

The Student *t*-test is used to assess whether each time series of 1921–1980 and 1981–2001 can be considered part of the same population. The 1981–2001 average is lower than the 1921–1980 average for 98% of the time series. The 5% and 1% significance level is, respectively, found for 75% and for 53% of the time series: the rainfall of the latest 20 years can be considered anomalously low.

Monthly data have been utilized to characterize the rainfall seasonal trend. The most important contribution to the annual negative trend is due to the winter (the months from December to February, which is the rainiest season) rainfall trend (Fig. 5). The precipitation deficit of the last 20 years is mostly due to a reduced contribution of winter rainfall. Spring (March–May) and autumn (September–November) also show negative trends, although much less evident. A positive trend appears for summer, the arid season: the effect is null in terms of water resources availability.

Temperature and net rainfall trend

Monthly temperature series are available from 1924 onwards. To fill the residual monthly gaps the time series were grouped into HCA. A different approach, based on gap filling and homogenization, was necessary for 15 time series of gauges located in Campania due to the abundance of very anomalous values, in particular for the last 20 years (the homogeneity evaluation is based on the Craddock (1979) test).

The linear trend analysis shows the temperature trend is not as homogeneous as the rainfall trend both in the whole study area and in each HCA (Tables 3 and 4). A prevailing increasing trend is observed in Campania but is weak in Apulia and is substantially absent in the remaining area.

To apply the Mann–Kendall test, pre-whitening of the temperature series is necessary due to their significant autocorrelation. The Mann–Kendall *S* distribution is close to Gaussian; there is a slight prevalence of increasing trends in the Campania region where an increase of temperature starting from about 1980 is observed, as shown in Figure 4. However, this is not enough to assess a significant and generalized temperature trend over the whole area in MSP, since this behaviour is not so evident elsewhere. These results are thus quite different from those by other authors (Brunetti *et al.* 2004; Cambi *et al.* 2000), probably due to the strong difference of data set length and spatial density of examined gauges.

The real or actual evapotranspiration E_r was calculated using Turc's formula (Turc 1954) with the correction suggested by Castany (1968), using temperature and rainfall monthly data. In this way an approximate but simple evaluation of the annual variation of actual evapotranspiration can be obtained.

The average annual net rainfall (ANR) of a time series ranges from 52 to 1565 mm for the whole group of 41 available time series in the period 1924–2001. The AC of net rainfall (ACNR) is strongly negative everywhere. The absolute value of ACNR is directly correlated to MAP: it increases or gets worse as MAP increases. ACNR ranges from −0.4 to −4.3 mm/a, grouping the time series by

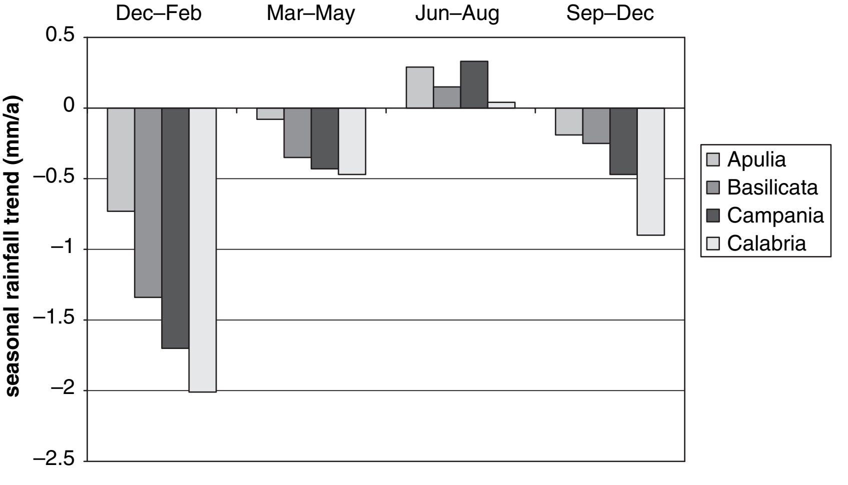

Fig. 5. Spatial average of seasonal rainfall trends of the main study period.

Table 4. *Number of time series as percentage of the total for each HCA and classes of angular coefficient of temperature trend (ACT, °C/100 years)*

HCA	ACT < −1	−1 < ACT < −0.5	+0.5 < ACT < +1	ACT > +1
Apulia	7	7	20	7
Calabria–Basilicata (all)	17	25	8	8
Campania	7	0	13	27

MAP (Table 5). In the whole period, the reduction of net rainfall can be roughly assessed as from 27% to 33% of ANR: this percentage is everywhere higher than that calculated for actual rainfall.

This dramatic situation is due to different phenomena. First of all, the downward rainfall trend is more relevant during winter, when generally net rainfall reaches maximum levels and actual evapotranspiration minimum levels. The entire winter decrease in actual rainfall becomes a decrease in net rainfall. The summer is arid everywhere and the actual evapotranspiration is less than the potential one due to low rainfall. The increase in summer rainfall is completely 'burned' by actual evapotranspiration. The recent trend towards a rise of annual temperature and the regime variation amplify the effect of rainfall variations.

Groundwater availability and role of climate change

Main characteristics of selected aquifers

The effects of recent climate variations on groundwater availability are evaluated considering five wide hydrogeological structures (HSs). In each HS the shallow or outcropping aquifer is considered; three are porous, two are constituted by carbonate rocks, all are coastal aquifers.

The Apulian Tableland HS, Tavoliere HS, consists of a shallow and large porous aquifer within a conglomerate sandy–silty succession, less than 60 m deep, with a clayey impermeable bottom (Polemio *et al.* 1999). It is deep enough to allow seawater intrusion only in the vicinity of the coast. Groundwater is phreatic inland or far from the coast, in the recharge area, whereas it is confined in the remaining part of the aquifer; maximum piezometric levels reach 300 m a.s.l.

Table 5. *Net rainfall trend and MAP classes considering data from 1924 to 2001*

MAP Class	ANR (mm)	NRT (mm/a)	NRV/ANR (%)
<600 mm	85.5	−0.39	−33.1
600–900 mm	227.9	−0.89	−27.1
900–1300 mm	464.8	−1.99	−32.2
>1300 mm	967.9	−4.30	−32.3

ANR, average net rainfall; NRT, net rainfall trend; NRV, net rainfall variation from 1924 to 2001 and ANR ratio.

Except for the Tavoliere, the Apulian region is characterized by the absence of rivers and the non-availability of surface water resources due to its karstic nature. Considerable groundwater resources are located in large and deep carbonate coastal aquifers, as in the case of the Gargano (not considered in this study due to the low data availability), the Murgia and the Salentine Peninsula (Salento) HSs. The Murgia and Salento areas show some common features (Cotecchia *et al.* 2005). They consist of large and deep carbonate aquifers, constituted mainly of limestone and dolomite rocks. Carbonate rocks are affected by karstic and fracturing phenomena, which occur well below sea level, whereas intruded seawater underlies fresh groundwater owing to a difference in density. Confined groundwater is more widespread inland; groundwater is phreatic everywhere along a narrow coastline strip. The maximum piezometric head is about 200 m a.s.l in the Murgia area and 5 m a.s.l in the Salento area (Spizzico & Tadolini 1997).

Five rivers cross the Metaponto plain, located along 40 km of Ionian coast. Marine terraced deposits, mainly sands, conglomerates and silts, crop out in the upland sectors of the Metaponto plain, while alluvial, transitional, marine and coastal deposits crop out in the coastal plain and along the rivers (Polemio *et al.* 2003). Two main types of porous aquifers can be distinguished in the Metaponto plain. The first one encloses the aquifers of the marine terraces and the alluvial river valley deposits. The marine terrace aquifers display medium to high hydraulic conductivity; the river valleys regularly break their spatial extent. The aquifers of the river valleys display low to medium hydraulic conductivity and they do not generally permit an accumulation of significant groundwater resources. The second type of aquifer includes one of the coastal plain deposits and has a medium hydraulic conductivity. This aquifer is the most exploited one for practical purposes due to its extension (about 40 km wide), thickness, continuity across the plain and also because its outcropping surface is more affected by economic growth and increasing water demand.

The groundwater of the coastal plain aquifer flows in a multilayered aquifer; it is mainly phreatic, otherwise it is confined due to an upper, almost impervious and outcropping stratum.

The Sibari plain is located in NE Calabria and covers the final sector of the Crati river. The Sibari plain is bordered by the carbonate relief of the Pollino Massif to the north and by the intrusive and metamorphic rocks of the Sila Massif to the south; it is composed of sedimentary lithotypes, varying from sand to marl and clay and including gravel locally.

The Sibari plain houses multilayered aquifers, the recharge of which is partially ensured by groundwater flowing from massifs and by leakage of rivers. For this study the shallow sandy aquifer alone has been considered (Polemio & Petrucci 2003).

Hydrological data

Piezometric data (Table 6) and river discharge measurements (five time series ranging from 1930 to 1992 for the Tavoliere and three series ranging from 1929 to 1971 for the Metaponto plain; SIMN 1916–2000) are considered together with already analysed rainfall and temperature data. The continuous or regular monthly piezometric data are derived from gauges managed by SIMN (1916–2000) or by the Irrigation Development Agency (Regione Puglia 1983). Occasional and recent data were collected on-site by the IRPI hydrogeological staff (Polemio & Dragone 2003, 2004; Polemio *et al.* 2003, 2004*b*, 2004*c*; Polemio & Petrucci 2003) and by other sources (Regione Puglia 2002) for the Tavoliere aquifer; Lopez *et al.* (2003) and CASMEZ (1987) for the Sibari aquifer.

Data from 58 wells or piezometric gauges are available for the three Apulian HSs – the Tavoliere, the Murgia and the Salento (Polemio & Dragone 2004). The piezometric data sets regarding the Tavoliere are available for a minimum of 17 years and for a maximum of 55 years, covering a continuous period between 1929 and 1994 (Polemio *et al.* 1999). Continuous data are available from 1973 to 1978 for the Murgia and the Salento. Furthermore, sporadic recent data were collected in Apulia for the periods 1995–1997 and 2001–2003.

Piezometric time series of monthly data are available in the Metaponto plain for 60 wells in two periods, 1927–1940 and 1951–1984 (Polemio & Dragone 2003). Occasional but high density data are available in the whole plain for 1953 and 1990 and in a selected study area for each season of 2002 (Polemio *et al.* 2004*a*).

Data from 121 wells in the Sibari plain were considered, for which discontinuous piezometric data are available from 1932 to 2002. Data are concentrated in the 1930s, the 1950s and the 1970s. The surveying was managed by different institutions in these different periods: the location and the identification of wells are not detailed enough to permit linking of the series. The oldest data were regularly collected from 1932 to 1940 in 27 wells; this data set has been used as a reference for spatial analyses. In June 2002 a high density piezometric survey was carried out. Due to the shortness of regular piezometric time series, the analysis for the Sibari plain is limited to the spatial analysis of piezometric surface modifications.

Data analysis

Piezometric data are explored by typical approaches of time series analysis such as autocorrelation, cross-correlation and trend analysis tools, and of spatial analysis, using kriging to obtain grid data of piezometric surfaces to compare using simple arithmetic operations and volume determinations.

The piezometric value recorded in any month is strongly dependent upon the values of the previous months, the link being significant, diminishing as the time lag increases. The duration and the

Table 6. *Piezometric data availability for each hydrogeological structure (HS) and straight line trend (AC)*

HS	Well number	Data range	Minimum AC (m/a)	Trend more probable at 2002 or 2003
Tavoliere	11	1929–2002	−0.408	High decrease
Murgia	30	1965–2003	−0.240	High decrease
Salento	17	1965–2003	−0.060	Decrease
Metaponto	60*	1927–2002	−0.236†	Decrease
Sibari	121	1932–2002	‡	Decrease

*The number of wells available for occasional years is higher and variable.
†In the periods 1927–1940 and 1951–1984.
‡Determination not available due the characteristics of data set.

intensity of this dependence, called the memory effect, is a function of specific yield, saturated thickness, hydraulic conductivity and extent of the aquifer. High values of these parameters are typical of aquifers of high quality to tap groundwater; in these cases the autocorrelation decreases slowly as the lag increases.

The cross-correlation for each piezometric series is determined by comparing it pairwise with data from a hydrogeologically significant series of rainfall, temperature and, where available, river discharge, the data of which are utilized step by step with increasing lag from 1 to 12 months. The cross-correlation coefficient expresses a measure of effect of the latter variable, rainfall, temperature or river discharge, on the variability of the former variable, the piezometric height or level.

The spatial analysis is utilized to complete the trend analysis of piezometric data when sporadic but high density data are available.

The three considered porous aquifers are subject to similar hydrological conditions: the range of mean annual rainfall and temperature is, respectively, 440–600 mm and 15.9–16.9°C. The Tavoliere area is a little cooler and drier than the other two areas. With regard to the monitored river discharges, the whole range of mean annual values is 1.0–20.0 m^3/s.

In the Tavoliere aquifer the autocorrelation coefficient always decreases slightly and quite linearly. There is a very high autocorrelation in 56% and 22% of wells, decreasing the coefficient respectively from 1 to about 0.8 and to 0.5, increasing the lag up to 12 months. The autocorrelation of remaining wells is insignificant after six lags.

The piezometric height is cross-correlated with river discharge, temperature (in this case this is an anticorrelation) and rainfall in decreasing order of maximum absolute value of coefficient. The river bottom is generally higher than the piezometric height in the monitored locations.

The maximum of the cross-correlation coefficient (MCC, as absolute value if the coefficient is negative, as in the case of temperature) and of maximum significant lag (MSL, the maximum lag for which there is statistical significance of cross-correlation) are, respectively, 0.5 and 2 months for discharge, 0.4 and 3 months for temperature, and 0.3 and 5 months for rainfall. MSL seems well correlated to depth to water, and thus appears to be a useful parameter to evaluate the time necessary to transfer a surface water impulse to groundwater.

The fact that even temperature variations are significant, more widespread than rainfall, has already been observed in similar hydrogeological conditions (Polemio *et al.* 1999). This can be explained by considering the nature of the climate, which is semi-arid everywhere for the selected aquifers. In this type of climate, the temperature is significant because of two separate and cyclic phenomena. The first is a natural phenomenon, real evapotranspiration, which 'regulates' the availability of net rainfall for infiltration from autumn to spring. The second is anthropogenic and is mainly linked to groundwater discharge from spring to autumn, due to high temperatures and potential evapotranspiration: the farms use more groundwater to offset the water deficit. In this way, temperature variation more than rainfall explains piezometric variations over the whole hydrologic year.

The trend of river discharge has been characterized by five gauges: the trend is clearly decreasing, especially since 1980, as in the case of rainfall in the whole Apulian region. It should be recognized that this variable is also influenced by human activity due to depletion by dams or by diffuse withdrawals from rivers for farm use, which strongly increased by 1980.

The piezometric trend everywhere is decreasing (Table 6). The continuous piezometric lowering has transformed many confined wells into phreatic wells; after that, the shallow groundwater of the Tavoliere is completely depleted in places. In terms of straight line trend, the trend everywhere is strongly negative, constituting a severe problem for groundwater discharge by wells (Table 6). The trend is confirmed by the spatial analysis of sporadic 2002 data: the spatial mean of piezometric decrease is 7.93 m over about 15 years.

In the Metaponto plain, the autocorrelation coefficient is quite linear from the maximum, a bit lower than 1, to the minimum, equal to not less than 0.3, increasing the lag up to MSL, equal to 6 months (after that the piezometric values are independent). The memory effect is high everywhere, but generally higher where groundwater is confined.

The MCC is generally less than 0.5 while MSL is 3 months, with some exceptions up to 4 months in the case of rainfall. As in the case of the Tavoliere, the MCC is low and lower than the absolute value of temperature MCC which is, in this case, also greater than the river discharge MCC. The temperature MCC is generally less than 0.7 while MSL is 2 months, with some exceptions of up to 4 months; everywhere, the coefficient is negative. The river impulse in terms of piezometric variations is extremely quick and important: MSL is generally 1 month and MCC less than about 0.6.

The trend of the piezometric series has been defined both on a synoptic scale, for the whole plain, and on a detailed scale, in the selected study area. With regard to the synoptic scale, the trend is described by AC while for the detailed scale it is the result of a comparison of the piezometric surfaces of different years.

The piezometric minimum occurred generally between 1952 and 1954 (up to 1984), when the exploitation of the aquifers was very high, after the end of land reclamation works (Polemio & Dragone 2003).

The time series analysis shows a widespread negative trend for the period 1927–1940 with a piezometric drop, on average, equal to 0.05 m/a. In the same period of time, the rainfall trend is slightly positive locally. This figure can be explained, as the groundwater represented the only irrigation resource during these years. Conversely, a positive trend is observed from 1951 to 1984 (75% of the total available time series), even if rainfall and river discharge trends are not positive. This figure can be explained considering the fact that many dams were built in this period; the dams started to supply more irrigation water than ever before. This allowed a reduction of groundwater tapping and also created a sort of artificial recharge, due to over-irrigation.

A new trend variation, a widespread downward trend, started during the 1980s; this trend remains unchanged. During 1988–1991 a heavy drought hit the area. The effect was the depletion of artificial lakes and the massive re-utilization of groundwater. The piezometric effects of drought in 1990 were relevant, particularly as compared to the situation in 1953, defined almost as a minimum until 1984. Negative piezometric variations are generally also prevalent in the case of the selected study area, where a more detailed analysis was carried out during 2002 (Polemio *et al.* 2003). For this area, the 2002 spatial mean of piezometric height is 1.12 m less than that of 1953, while that of 1990 is 0.34 m higher.

The maps of the piezometric variations in the Sibari plain highlight a decreasing trend which started around the 1950s, assuming as reference piezometric surface that of the 1930s. The plain as a whole is characterized by piezometric heights in the 1950s which were higher than in the 1930s (positive piezometric variation). Major decreases of piezometric height (negative variations) appear inland during the 1970s while two separate positive variation areas remain, involving main rivers and portions of the coast. Both a lowering of the piezometric negative variations, in term of absolute values, and a narrowing of the positive variation area are observed in 2002, the positive variation area becoming quite similar to a strip along the coast. The spatial mean of piezometric variation of 2002 is the lowest or the worst observed; it is equal to a lowering of 4.42 m with respect to the 1930s.

In the case of the considered Apulian carbonate hydrogeological structures, the autocorrelation piezometric coefficients consistently show a progressively declining trend, starting from one to the statistically significant minimum, everywhere not less than 0.5, increasing the lag to 4 months for the Murgia and 5–6 months for the Salento area. The consistent memory effect of Apulian groundwater is a characteristic feature which is of great importance during droughts or dry spells. The Salento has shown very strong and long-lasting memory effects, which is only further proof of the good hydrogeological characteristics of these aquifers.

There is a cross-correlation between the piezometric and climatic variables of an acceptable significance level for a time lag up to 4 months. The effects of rainfall are perceptible up to a maximum period of 2–3 months, whereas the best correlation with temperature is felt with a time lag of 4 months. The temperature variations are more significant than rainfall in some portions of the Murgia and the Salento.

The calculated piezometric trend, generally speaking, is downward, since there is a widespread tendency, albeit in some cases a very slow one, towards a piezometric drop. The lowest piezometric decrements have been observed in the Salento area, which has an AC in the range of -0.060 to -0.012 m/a; worse AC values are typical of the Murgia HS (Table 6). In the Murgia, as in each HS, the AC approaches zero the closer one gets to the coastal areas, as would be expected.

During 2002, a widespread and dramatic drought period ended. On the basis of the available data set, the most likely piezometric trend, ending in the second half of 2002, was very negative and serious in terms of the sustainability of the groundwater demand, over the entire area covered by porous aquifers, as in the case of the Tavoliere, the Metaponto plain and the Sibari plain (Table 6). This situation is confirmed by sporadic data of 2003 in the case of the Murgia and the Salento, notwithstanding the effect of more than a year of abundant rainfall.

Conclusions

A widespread decreasing trend of annual rainfall is observed over 97% of the whole area from 1921 to 2001. The spatial average of trend value and MAP highlight the fact that the rainfall trend worsens or decreases as the MAP increases. This phenomenon is extremely worrying because high MAP areas are wide Apennine portions of the widest drainage basins of the artificial lakes which guarantee a relevant percentage of water supplies. The spatial mean of trend ranges from -0.8 mm/a in Apulia to -2.91 mm/a in Calabria.

The downward trend is mainly the effect of a succession of low-rain years; this succession is anomalous from about 1980, in terms of frequency and intensity of annual rainfall less than MAP; in this context, the droughts of 1988–1992 and 2000–2001, the worst since 1921, appear to be more important.

On a seasonal basis, the downward trend is concentrated in winter: the precipitation deficit of the last 20 years is mostly due to a reduced contribution of winter rainfall.

A Mann–Kendall test does not show a significant prevalence of negative or positive temperature trends. Although in some stations the highest temperatures have been recorded in the last ten years and a slight increase seems to prevail, especially from about 1980 onwards, this is not enough to determine a significant and generalized increasing temperature trend for the whole area.

The annual mean of net rainfall ranges from 52 to 1565 mm. The trend of net rainfall is everywhere strongly negative; in the whole study period the reduction of net rainfall can be roughly assessed from 27 to 33% of the annual mean.

The selected aquifers show high hydrogeological characteristics as confirmed by the consistent memory effect, which is not shorter than 4 months. This characteristic is of great importance during droughts or dry spells.

The cross-correlation with piezometric level shows that the variability of groundwater availability can be explained in terms of rainfall, temperature and, where it exists, river discharge variability. The significant lag or duration of this influence generally decreases from the first variable, rainfall, to the last, river discharge, and respectively from 3–5 to 1–2 months. The intensity of this influence, in terms of maximum cross-correlation coefficient, is generally due to temperature, river discharge and rainfall in decreasing order.

The piezometric trend is on the decline everywhere, so widespread as to determine serious effects in terms of groundwater discharge sustainability. The worst trend in each aquifer or structure ranges from 0.06 to 0.41 m/a notwithstanding the limiting effect of sea level boundary condition. Detailed spatial studies show an average decrease of 7.93 m over the last 15 years in the Tavoliere, 1.1 m over 50 years in the Metaponto plain and 4.4 m over 70 years in the Sibari plain.

During 2002 the latest widespread and dramatic drought ended. On the basis of the data set available, the most likely piezometric trend, ending in the second half of 2002, was a very serious one over the entire area covered by porous aquifers, as in the case of the Tavoliere, the Metaponto plain and the Sibari plain. This situation is confirmed by sporadic data of 2003 in the case of the Murgia and the Salento, notwithstanding the effect of more than a year of abundant rainfall.

The whole piezometric downward trend appears to be due to the overlapping effects of natural recharge and of increasing well discharge. The effects of the latter phenomenon appear to be influenced by the progressive availability of surface water resources tapped by dams. The increasing role of these water resources to cover the water demand does not prevent a return to pumping water from aquifers during the recent and unusual droughts.

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Climatological cycles in groundwater levels in a detritic aquifer

J. A. LUQUE-ESPINAR[1], M. CHICA-OLMO[2] & E. PARDO-IGÚZQUIZA[2]

[1]*Instituto Geológico y Minero de España (IGME), Urbanización Alcázar del Genil 4, 18006 Granada, Spain (e-mail: ja.luque@igme.es)*

[2]*CEAMA/Department of Geodynamics, University of Granada, Avenida Fuente Nueva s/n, 18074 Granada, Spain (e-mail: mchica@ugr.es)*

Abstract: The spectral analysis of waterhead measurements on a network of piezometers has allowed the analysis of the spatio-temporal variability of data in the Vega de Granada aquifer (southern Spain). This analysis has detected several long-term cycles in the piezometric time series for the period 1966–2001. A complete analysis, using different spectral methods of those time series, has detected four different cycles: a decadal period (11 years), a 3.2-year period, an annual period and a semi-annual period. The annual cycle is ubiquitous, as a reflection of the annual hydrological cycle. The decadal cycle can be traced to hydrological variations involving a climatological regime induced by sunspot activity (11-year cycle). The piezometers located close to streamflows reaching the aquifer or along the main river, in areas where the main river recharges the aquifer, present a statistically significant decadal cycle. The explanation for the presence and spatial distribution of the 3.2-year cycle is similar to the decadal cycle, though it is identified in fewer piezometers.

In a general sense, spatial and temporal changes in groundwater levels in an aquifer are conditioned by both anthropogenic and natural factors. The former are related, mainly, to the effect of pumping, artificial recharge, etc. The latter have a number of causes, usually involving recharge by rain infiltration or superficial runoff, although other factors may be considered, such as changes due to atmospheric pressure or, in the case of coastal aquifers, the evolution of tides (Tison 1956; Nilsson 1968; Carr 1971; Fetter 1994). Frequently, series of the waterhead observed in control points over a long time show cyclic behaviour ranging from short-term (i.e. hours, days) to long-term variations (i.e. semester, annual or decadal cycles). However, our ability to study such cycles is affected, firstly, by the real presence of such cycles in the hydrodynamic behaviour of the aquifer system, and secondly, if present, with the availability of a record of measurements long enough to allow the cycles to be detected statistically. This aspect concerning experimental data series is important because long-term cycles in short records may have the appearance of a trend, which is obviously spurious.

Spectral analysis is a powerful statistical technique widely used for analysing the presence and statistical significance of cycles in time series. Given N experimental data of waterhead recorded in a piezometer at an equally spaced time sampling interval Δ, the range of frequencies that can be investigated goes from the Nyquist frequency $1/2\Delta$ to the Rayleigh frequency $1/N\Delta$. Additionally, frequencies close to the Nyquist one will be affected by aliasing if the sampling interval Δ is not short enough to represent the cycle with the highest frequency affecting the waterhead evolution. Also, frequencies close to the Rayleigh one are likely to be affected by red noise (i.e. noise in the low frequencies). The origin of red noise may be the presence of a trend, long-period cycles or random components. Although spectral techniques can be used to ameliorate the previous inconveniences, the effective range of frequencies that can be investigated with confidence will be an interval shorter than the one defined by the Nyquist and Rayleigh frequencies. In our case study, for example, with the sampling interval Δ being a month and N ranging from 200 to 350, so that the length of the time series is from 20 to 30 years, the identifiable cycles will range from 6 months to 120 months (10 years), i.e. the cycles occur at least twice in the time series. Such are the limitations imposed by the sampling interval and the length of the records.

A careful spectral analysis determined the statistical significance of spectral peaks in the power spectrum of the waterhead time series for 53 piezometers irregularly distributed throughout the Vega de Granada alluvial aquifer (SE Spain).

Hydrogeology of the Vega de Granada aquifer

The Vega de Granada is an important Mediterranean aquifer located in an alluvial plain surrounded

From: DRAGONI, W. & SUKHIJA, B. S. (eds) *Climate Change and Groundwater*. Geological Society, London, Special Publications, **288**, 53–62.
DOI: 10.1144/SP288.5 0305-8719/08/$15.00

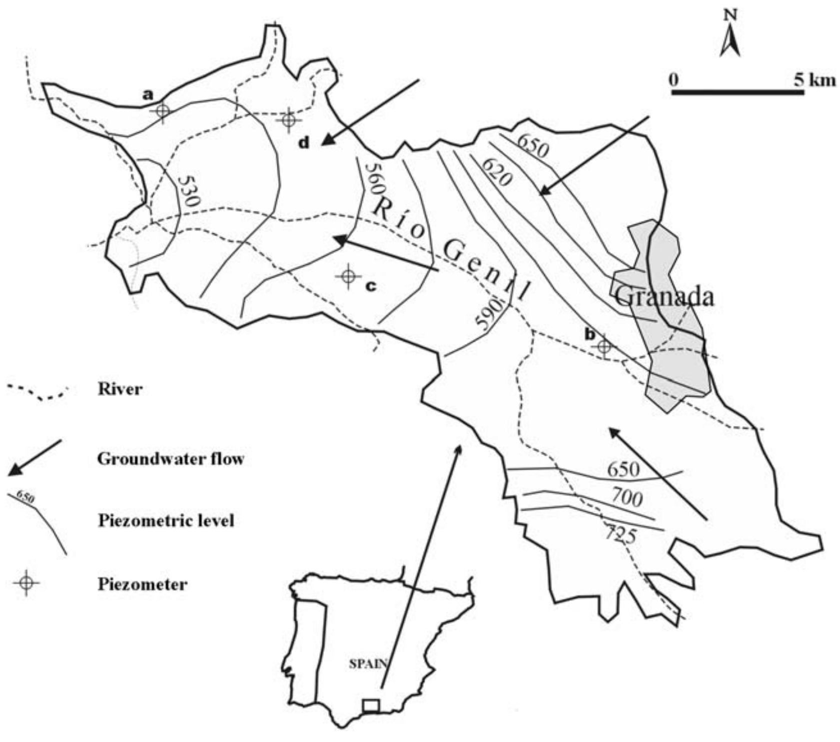

Figure 1. Geographical location on the Vega de Granada aquifer in Southern Spain with the location of piezometers represented in Figure 3a–d.

by mountains in southern Spain, within the province of Granada (Fig. 1). The aquifer presents a multilayer structure, with the superposition of sedimentary materials showing a varied range of permeabilities. The surface of the aquifer has a smooth topography between the altitudes of 530 and 760 m above sea level and it is the receptor of a drainage basin of 2900 km^2. The annual mean rainfall over the aquifer is 450 mm, although it can reach 1000 mm above some points of the drainage basin such as the Sierra Nevada range. The aquifer has a surface of around 200 km^2. Its Quaternary alluvial sediments reach a thickness of 250 m in the middle and diminish towards the northern and southern borders to 50 m. The sediments are mainly gravel, sands, silts and clay, with frequent spatial changes of lithofacies, resulting in a multilayer unconfined aquifer. There is an increase of the silt and clay fraction from the central axis of the aquifer (defined by the Genil river) towards the northern and southern ends (PNUD/FAO 1972). Figure 2 shows a cross-section along the main river, where lithofacies and permeabilities were interpreted from drilling data. As seen, the changes in permeability in the aquifer are considerable, which may explain the spatial variations in the spectra calculated; however, the role of the distance to sources and to the aquifer borders would appear to be more important (Luque-Espinar 2001).

The mean transmissivity of the aquifer is around 4000 m^2/day with a wide range of variation from 40 000 m^2/day in some central sectors, to 100 m^2/day at the border when clay materials are more frequent. The mean effective porosity is estimated at 6%, with most values ranging from 1 to 10% (PNUD/FAO 1972). The mean groundwater flow direction is from east to west, with the steepest gradients in the NE and eastern sectors (Fig. 1).

The main inputs into the aquifer come from the infiltration of superficial runoff and return from irrigation water, plus infiltration of rainfall water. An important part of the superficial runoff comes from snow melting in the Sierra Nevada range (with altitudes up to 3000 m). Total runoff water is estimated at 400 hm^3/year, and the renewable aquifer resources are estimated at 180 to 230 hm^3/year (PNUD/FAO 1972; ITGE 1989). The output from the aquifer is mainly natural drainage towards rivers, while well pumping represents around 30 hm^3/year (ITGE 1989).

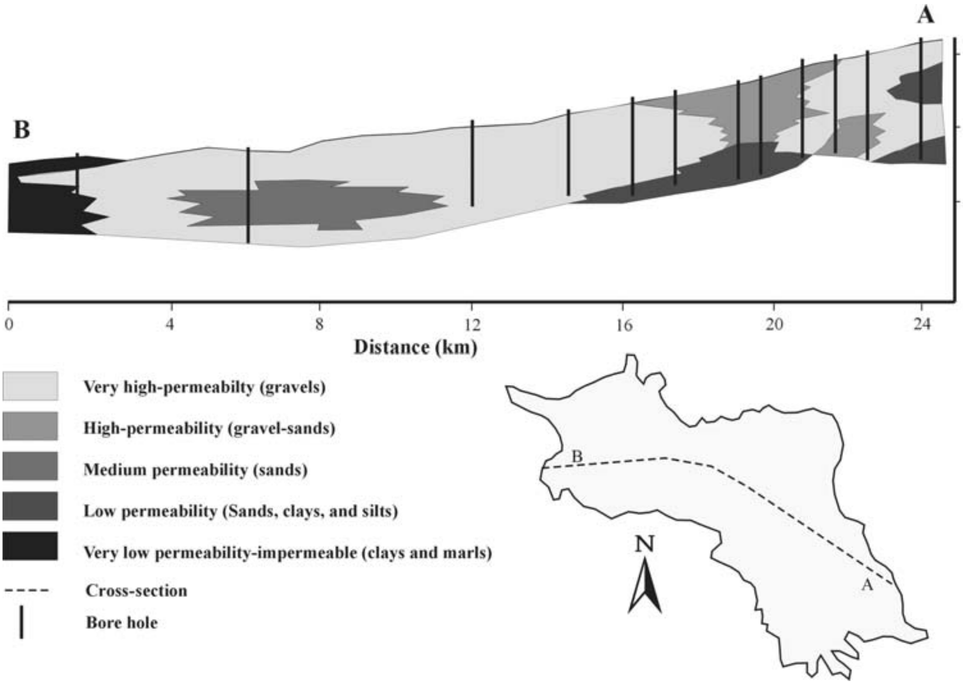

Fig. 2. Cross-section of the vertical permeability distribution from identification of borehole data along the profile A–B.

Waterhead data

The original data used in this study come from the control network of the aquifer, carried out by the Instituto Geológico y Minero de España (IGME) and Confederación Hidrográfica del Guadalquivir (Ministry of Environment). The piezometric series expand up to 35 years with the oldest data from 1968. The number of piezometric series used in this work was 53, scattered all over the aquifer. Observations were made on a monthly basis for most parts, and sometimes every two months. Thus, the series average nearly 150 measurements, which is sufficient for studies of this nature. The data are measured almost monthly, but not on the same day of the month. The sampling with a constant step of one month does not introduce any considerable distortion of the spectral content. The database was checked for errors and, afterwards, the time series were sampled with a constant step of one month using linear interpolation. Figure 3 shows the time series at four piezometers, located in different sectors of the aquifer. These series can be considered representative of the water level variations (Luque-Espinar 2001). Piezometer A (Fig. 3) is situated at the western end, in a sector of the aquifer where fine-grained sediments predominate. The piezometric time series presents a cyclical aspect of long periods with minor piezometric variations; no monthly fluctuations are seen. Piezometer B (Fig. 3) is near the eastern border, in the main recharge area. In this area, gravel-size sediments are abundant, and permeability values are very high. Evolution over time shows long cycles with monthly fluctuations. Piezometer C (Fig. 3) is located in the central part of the aquifer, near the main river, which contains water all year round. The high permeability values are due to the abundance of gravel and sand. The series is characterized by distinctive annual cycles within the long-range variations. Finally, piezometer D (Fig. 3) is located between A and C, in the aquifer's discharge area. The sand-sized detritic fraction predominates, with intercalating gravel and lime parcels. Monthly oscillations are very pronounced, while long-term ones are scarcely observed.

Spectral analysis

The statistical technique used to find cyclic components in a time series is known as spectral analysis (Jenkins & Watts 1968; Yevjevich 1972; Bras & Rodríguez-Iturbe 1985). The signal component represents the structured part of the time series, made up of a small number of embedded

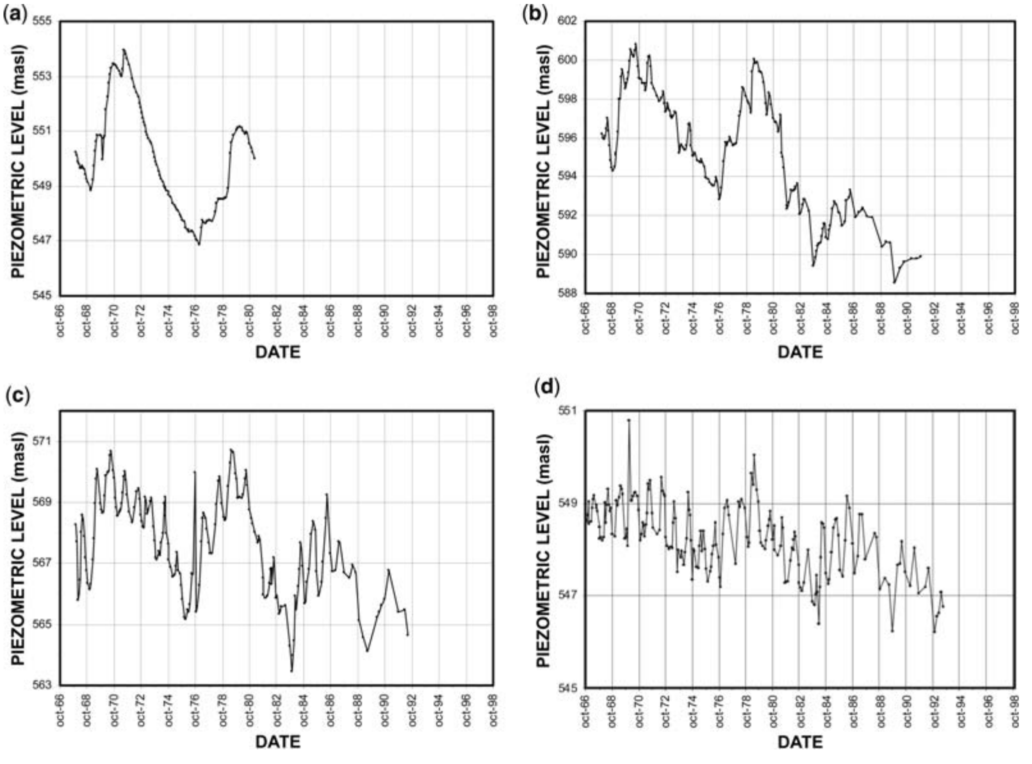

Fig. 3. Representative time fluctuations of piezometric levels in the aquifer showing an increase of the irregular data (**a**–**d**). Levels are in metres above sea level (masl).

periodicities or cycles repeated over a long time. The noise is a random component; it may be white noise, but more often will be red noise. A time series can be represented by a finite number of measurements. In the present case, a piezometric time series is represented by a succession of water-head data measured at regular or irregular time intervals. When a cycling component is added to another cyclic component of a longer period than the length of the time series, it will give an apparent trend. This, together with possible real trends and other factors, gives rise to noise in the low frequencies, known as red noise.

Harmonic analysis is another name used to denote the estimation of cyclic components in the time series. The time series is supposed to be a linear combination of sinusoidal functions of known periods but of unknown amplitude and phase. The modulus of the amplitude is related to the variance of the time series, explained by the oscillation at each frequency. The representation of the square of the modulus versus frequency is known as the power spectrum. There are a number of methods that can be used to infer the power spectrum: the periodogram (Papoulis 1984), the Blackman–Tukey approach (Blackman & Tukey 1958), maximum entropy (Burg 1972), and Thomson multitaper (Thomson 1982), among others. Each methodology has advantages and disadvantages, for which reason a good strategy is to use various methods and compare the results. This was done with the time series of the piezometric head; nevertheless we can affirm that the indirect method of Blackman–Tukey is a robust approach which gives acceptable results with our data sets. As an example of this comparative analysis, Figure 4 shows the results obtained with the methods of (a) maximum entropy; (b) Thomson multitaper; and (c) Blackman–Tukey. The results obtained were similar, yet we consider the Blackman–Tukey approach to be more robust and therefore more adequate for the analysis of time series.

The power spectrum is calculated from the covariance function by:

$$\hat{S}(\omega) = \frac{1}{\pi}\left\{\lambda(0)\hat{C}(0) + \sum_{k=1}^{M}\lambda(k)\hat{C}(k)\cos(\omega k)\right\} \tag{1}$$

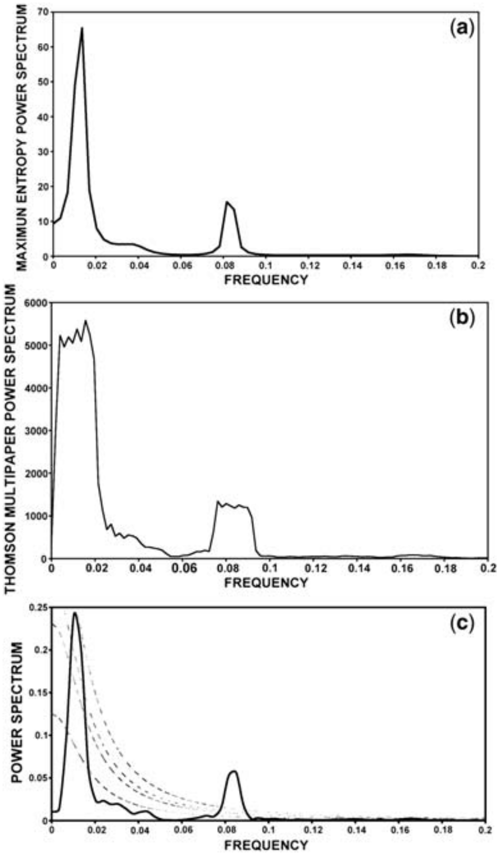

Fig. 4. Power spectra of piezometric data represented in Figure 3c using (**a**) maximum entropy (**b**) Thomson multitaper method and (**c**) Blackman–Tukey methods. In (c) dashed lines represent, starting from the bottom, the underlying power spectrum and the 90%, 95% and 99% confidence levels.

where $\hat{S}(\omega)$ is estimated power spectrum for frequency ω, $\hat{C}(k)$ is estimated covariance function for the kth lag and $\lambda(k)$ is weighting function, known as a lag-window, which is used to give less weight to the covariance estimates as the lag increases. For large lags, the estimated covariance function is less reliable. The lag-window used was the Tukey window:

$$\lambda(k) = \frac{1}{2}\left\{1 + \cos\left(\frac{\pi k}{M}\right)\right\} \qquad 0 \leq k \leq M \quad (2)$$

where M is maximum number of lags for the covariance function used in the spectral estimation. The maximum number of lags is $N-1$, with N being the number of experimental data; however, with large values of M a great number of peaks will be seen in the estimated power spectrum, most representing spurious cycles. On the other hand, if M is very small, significant cycles would not be seen in the estimated power spectrum. For this reason we used a number of $M = N/2$ in order to resolve peaks, and a value of $M = N/4$ in order to find out which are the most significant peaks.

In addition to using a small value for N, confidence levels were estimated for the inferred power spectrum. Our approach consists of fitting a background power spectrum with no cyclic component, but rather a smooth continuous spectrum, which is done by fitting the spectrum of an autoregressive process of order one, i.e. AR(1). The parameter of this process is estimated from the experimental data. Then, we take into account the known result for the one-side confidence band of power spectrum estimator:

$$\mathrm{P}\left(\upsilon\frac{\hat{S}(\omega)}{S(\omega)} < \chi^2_{\upsilon,\alpha}\right) = 1 - \alpha \qquad (3)$$

where P is probability operator, $\hat{S}(\omega)$ is power spectrum estimate for frequency ω, $S(\omega)$ is underlying power spectrum for frequency ω, υ is number of degrees of freedom – for the Blackman–Tukey estimate with a Tukey lag-window, the number of degrees of freedom is $2.67N/M$, $\chi^2_{\upsilon,\alpha}$ is the α quantile of a chi-squared distribution with υ degrees of freedom and α is significance level. For this study, we established confidence levels of 90%, 95% and 99%.

Results

The harmonic analysis of the time series of waterhead variations on the network of piezometers detects the presence of four distinctive periodicities: a decadal cycle, a cycle of 3.2 years, an annual cycle and a semi-annual cycle. The decadal cycle is related to the climatic 11-year cycle, in turn related to sunspot activity in the context of the North Atlantic Oscillation (NAO). The 3.2-year cycles could be related to the climatic cycle of the NAO (it has been recognized that the NAO has an influence on climate on a cyclic basis known as the quasibiennial oscillation, which ranges between two and four years). The annual cycle is related to the hydrological annual cycle, and similarly the half-year cycle is related to two precipitation seasons in a single year. Table 1 gives the percentages presented by the cycles according to the established confidence intervals.

The 11-year cycle is apparent in most of the piezometric series (85%), marking one of the

Table 1. *Percentage of cycles according to confidence intervals*

Cycle	Level of significance (%)					
	>99	99–95	95–90	<90	D	ND
11-year	24	23	10	28	85	15
3.2-year	4	6	9	17	36	64
Annual	68	11	8	13	100	0
Semi-annual	2	4	2	13	21	79

ND, Not distinguishable; D, distinguishable.

main features of the temporal behaviour of the aquifer piezometric level, together with the annual cycle. The 3.2-year cycle, present in one-third of the series, also has a sparse distribution throughout the aquifer. The annual cycle appears in all the piezometric series, as expected. Finally, the semi-annual cycle is weakly represented, in just 21% of the series. This might be related to a precipitation cycle (in this area there are two precipitation seasons in a single year), which would have a minor influence on the recharging of the aquifer.

Figure 5 shows the typical power spectra found in the harmonic analysis. Not all cycles are equal in their presence and intensity in the power spectra of the different piezometers. As shown in Table 1, while decadal and annual cycles are detected in most of the piezometers, the 3.2-year and half-year cycles are detected only in a set of them. According to our probabilistic estimation, there is also a difference in the statistical significance of every cycle from one piezometer to the next and, consequently, there is a spatial variability in their importance in different areas of the aquifer.

The procedure used to assess this spatial variability was to give a code to every cycle at each piezometer according to its statistical significance in the estimated power spectrum of each piezometer. The categorical code used was:

- × the cycle is not distinguishable in the power spectrum;
- ∘ the cycle is distinguishable but is not statistically significant at the 90% level;
- ○ the cycle is statistically significant at the 90% level but not at the 95% level;
- ◯ the cycle is statistically significant at the 95% level but not at the 99% level;
- ⬭ the cycle is statistically significant at the 99% level.

The previous categorization, although arbitrary, allows us to highlight differences in the behaviour of cycles when the categories are represented on maps. Figure 6a–d present the spatial distribution of the statistical significance of peaks for the decadal, 3.2-year, annual and semi-annual cycles, respectively.

Figure 6a presents the spatial distribution of the statistical significance of the decadal cycle. The most significant presence of this cycle (category 4) can be seen in the borders of the aquifer and in relation to rivers and other watercourses that represent input of water from an important drainage network. We believe this is because climatic variations of around 11 years, related to the sunspot cycle, may only be seen in the variability of rainfall when integrated in an area such as the drainage basin. This implies a variability in the amount of water that enters into the aquifer from surface drainage networks; it would be detectable near the mouth of those tributary channels, but its effect is damped in the interior of the aquifer or farther away from the surface watercourse. The lesser significance of the 11-year cycle in the central area can be attributed to the small variation of the water level in this zone, where the signs of this cycle would be reduced. In the eastern sector, in contrast, the fluctuations in the piezometric level are pronounced, and decreases or increases accumulate, forming waves over an approximately 11-year period, well reflected in the spectral analysis. Moreover, for this cycle we found that the piezometric series were highly parallel to graphs of the NAO index (Hurrell 1995) in that the more positive the NAO index, the lower the piezometric levels. In this sense, it is generally accepted that the NAO marks climatic behaviour at these latitudes (Hurrel 1995; Qian *et al.* 2000; Rodrigo *et al.* 2000), though some authors relate this cycle to sunspot activity (Eddy 1976; Reid 1993).

Most researchers point to the NAO as the cause of the three-year cycle (Pozo-Vázquez *et al.* 2000), although some studies conclude that the behaviour of the rainfall in this region could be influenced by the El Niño South Oscillation (ENSO) (Rodo *et al.* 1997). Thus, the statistically significant presence of this cycle would be produced in a set of piezometers located in sectors where the vertical permeability in the proximities of surface currents

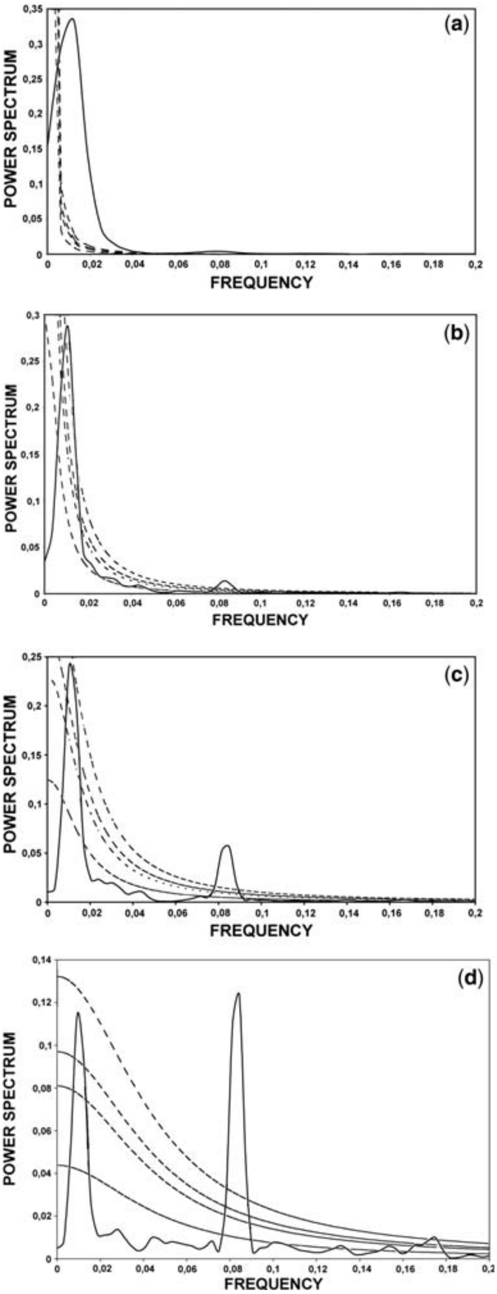

Fig. 5. Power spectra piezometric levels represented in Figure 3(a–d) from Blackman–Tukey.

is high. There is also a preferential location at points near the aquifer borders, or by torrential watersheds and more permeable detritic stretches. The NAO cycle is a subtle one that may be difficult to detect because its influence may be small in comparison with the decadal and annual cycles. We believe this cycle is exhibited in the form of minor pulses on the piezometric levels that will essentially coincide with stormy episodes. The lack of statistical significance or the absence of this cycle from other piezometers of the aquifer could be attributed mainly to a lower permeability of the sector where the piezometer is located, and to the recharge being more localized.

Figure 6c reflects how the annual cycle is highly significant (category 4) in most of the piezometers. This ubiquitous presence is logical because it is related to the yearly hydrological cycle, which affects the whole aquifer. From Figure 6d it may be seen how the semi-annual cycle is detected only at some piezometers. While there is no simple explanation for its spatial distribution, it might be related to some other climatic cycle (in this area there are two precipitation seasons in a single year).

Conclusions

Most studies of the temporal variability of piezometric levels that appear in the literature refer to short-term cyclic variations, with periods ranging from hours to days or weeks, such as the variations deriving from tidal waves that may be observed in coastal aquifers. Some series presenting long-term fluctuations have been studied qualitatively, and then compared with rainfall data, with which there is not necessarily a parallel (Hanson & Dettinger 2005).

The most relevant aspects of the present study are the long piezometric series involved and the spectral methodological focus, as there are few bibliographic precedents. Detecting longer cycles requires a long series of records, preferably on a network of monitoring points: there may be a spatial variability in the hydrodynamic parameters of the aquifer affecting the preservation of cycles in some places, while the cycles are unobservable at other places. There were four different cycles detected in this study (not all being detected at all piezometers): decadal, 3.2-year, annual and semi-annual. The decadal cycle is perhaps the most interesting one. It is the cycle with the longest range that is related to sunspot activity, yet it is not expected in waterhead evolution. It is seen more clearly at the piezometers in close proximity to drainage channels at the border of the aquifer or along the main river entering the aquifer. This is the case because climatic variations of rainfall in relation to sunspot activity are amplified by the effect of the drainage basin, which will produce a runoff series well correlated with the climatology. Thus, it has the same effect on piezometers close to where those drainage channels enter the aquifer. The presence of the three-year cycle would have a similar general explanation, yet because the rainfall regime and its spatial distribution are different, variations in the piezometric level would be seen only in the more permeable sectors of the aquifer borders. In this way, the relationship between the surface drainage network and the aquifer recharge would be established as a reflection of the climate. The existence of cycles in the series can therefore be traced to climatic variations and the hydrodynamic characteristics of the aquifer itself, as the signal received by the piezometric levels is modulated by the functioning of the aquifer. Local or distorting phenomena that may be caused by stormy episodes are not manifest in the piezometric levels, which would clearly reflect only the predominant cycles of the regional climate.

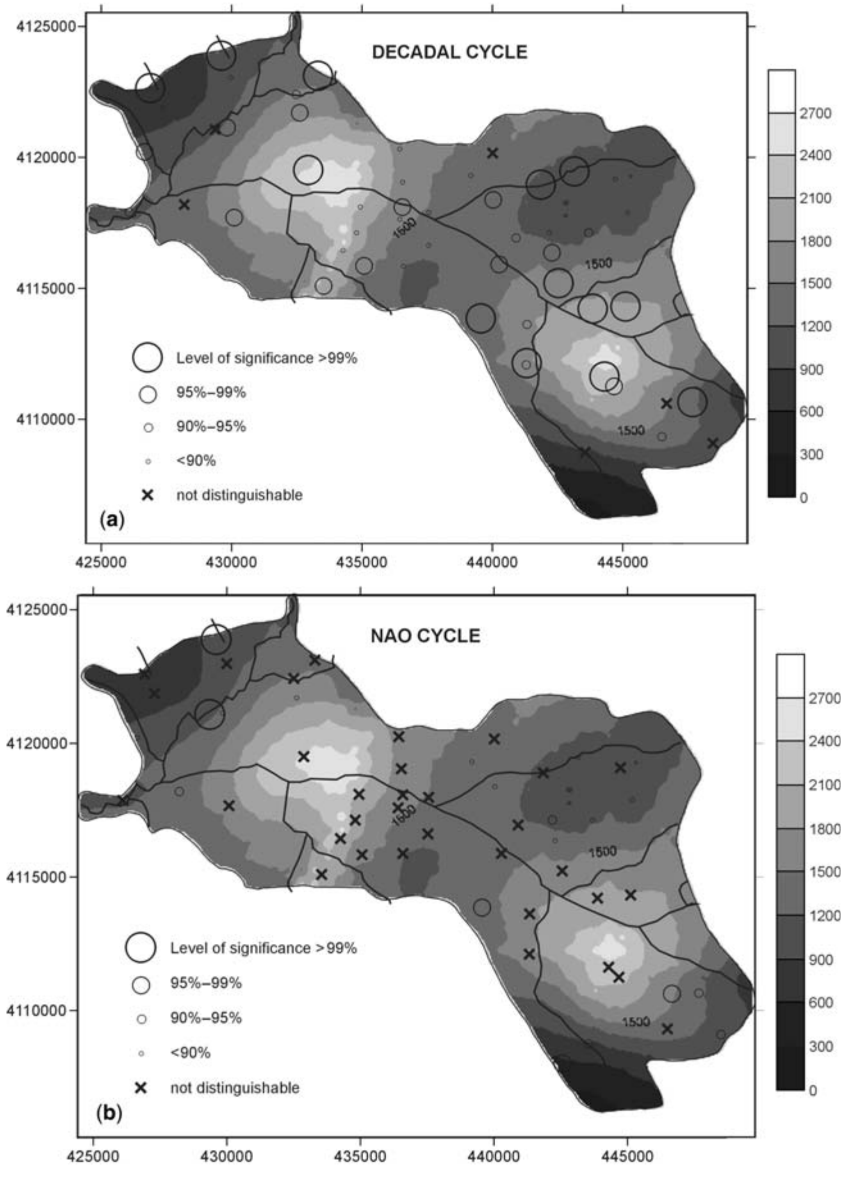

Fig. 6. Mapping of statistical significance of the cycles, represented over the transmissivity (m^2/day) kriged map: (**a**) decadal, (**b**) NAO, (**c**) annual and (**d**) semi-annual cycles. The axes are expressed in UTM coordinates.

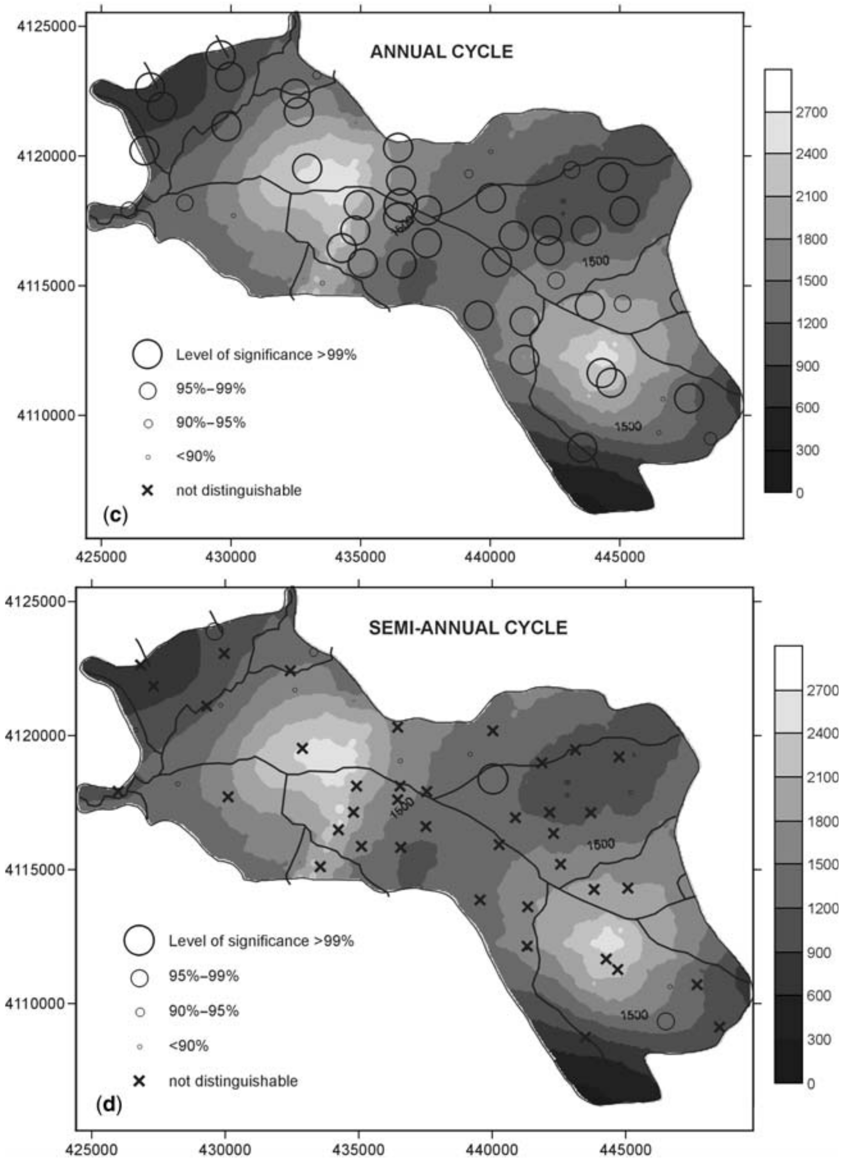

Fig. 6. (*Continued*).

In conclusion, we underline the importance of long-term monitoring of piezometric levels. Processing this information with the methodology put forth here constitutes, in our opinion, a sound reference for the study of climatic changes.

We are grateful for the financial support given by the Spanish MEC (Project BTE2002-00152) and Junta de Andalucía (Group RNM122). The third author is a Ramón and Cajal Grant Holder from the Ministry of Education and Science (MEC).

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Yperia Krini spring (central Greece): inferences on climatic changes from its 2000 years of history

D. RAPTI-CAPUTO[1] & B. HELLY[2]

[1]*University of Ferrara, Department of Earth Sciences, via Saragat 1, I-44100 Ferrara, Italy (e-mail: cpr@unife.it)*

[2]*Institut Fernand-Courby (UMR 5649 du CNRS), Maison de l'Orient, 7 rue Raulin, F-69007 Lyon, France (e-mail: Bruno.Helly@mom.fr)*

Abstract: The development of past civilizations and the foundation of towns have always been strictly linked to the availability of water. In this paper, we analyse more than 2000 years of evolution of Yperia Krini spring in Thessaly (Greece), by investigating possible variations in terms of water discharge. In particular, the integrated analysis of geological, hydrological, hydrogeological and historical data relative to the spring, called by Sophocles a 'gift of God and source of life', allowed us to understand the role played by both climatic variations and anthropogenic activities on the behaviour and the characteristics of the local underground water resources within the Thessalian plain.

The Yperia Krini spring is located in the middle of the town of Velestino (southern Thessaly, Greece), at an altitude of 105 m; it generates a small lake with an area of about 0.15–0.2 hectares (Fig. 1). The lake is bounded all around by a wall, while some artificial channels control the water outflow. The modern town of Velestino, which is built along the northeastern slope of Mount Malouka, roughly corresponds to the well known archaeological site of Pherae. The ancient city of Pherae commanded a fertile district near the southern verge of the Plain of Pelasgiotis, with important routes and easy access to the sea. Human settlement in this sector is know from the Neolithic and Bronze Ages. As a Greek city, Pherae took a prominent place in mythology as the home of Admetus. In historic times, Pherae was one of the most important Thessalian towns, flourishing especially in the Classic and Hellenistic periods (Intzesiloglou 1994).

In the history of Pherae, the spring called Yperia Krini has always been mentioned as the representative monument of the old town as it is for the modern one. It is evident that the abundant water of the spring has around the interest of the inhabitants, travellers and the poets of antiquity. According to a citation of Sophocles (825 N; see Pearson 1917), the Yperia fountain is a 'gift of God and a source of life' (*'Yperia Krini' nama theofilestaton*). Indeed, there is no doubt that the Ancients clearly acknowledged that the Yperia spring was a fundamental element of the environment they lived in and that the agricultural richness of the region was a direct consequence.

Although the quality of life of the town is no longer so closely related to the amount of water discharge occurring at the spring, any negative variation had always worried local people. Indeed, since 1989, the spring has suffered a significant loss in water discharge, which fell from 200 l/s to only 30 l/s. In order to understand the causes of this phenomenon and to define the working conditions of the spring, a specific hydrogeological study was carried out. Following the geological principle that 'the present is the key of the past', we tentatively investigated possible quantitative variations that have occurred at the spring over time.

Geological and structural framework

The town of Velestino is located at the base of the eastern slope of the Central Hills, also referred to as Revenia (Sivignon 1974), which represent one of the major structural highs of Thessaly (Caputo 1990) bounded by the Larissa Basin to the east, and the Karditsa Basin to the west. This hilly area is mainly characterized by Pliocene to Upper Pleistocene weakly deformed subhorizontal sediments. In the broader study area, these deposits unconformably cover different lithologies of the Subpelagonian Zone, among which are rocks of the ophiolitic suite, limestones and different terrigenous materials. A low grade metamorphism generally affects all these rocks.

The tectonic setting of the area is complex because it is characterized by some overthrusting units with a southward vergence, as inferred from the regional dips towards NNW (Fig. 2). In particular, within the study area, the principal tectonic unit

From: DRAGONI, W. & SUKHIJA, B. S. (eds) *Climate Change and Groundwater*. Geological Society, London, Special Publications, **288**, 63–77.
DOI: 10.1144/SP288.6 0305-8719/08/$15.00

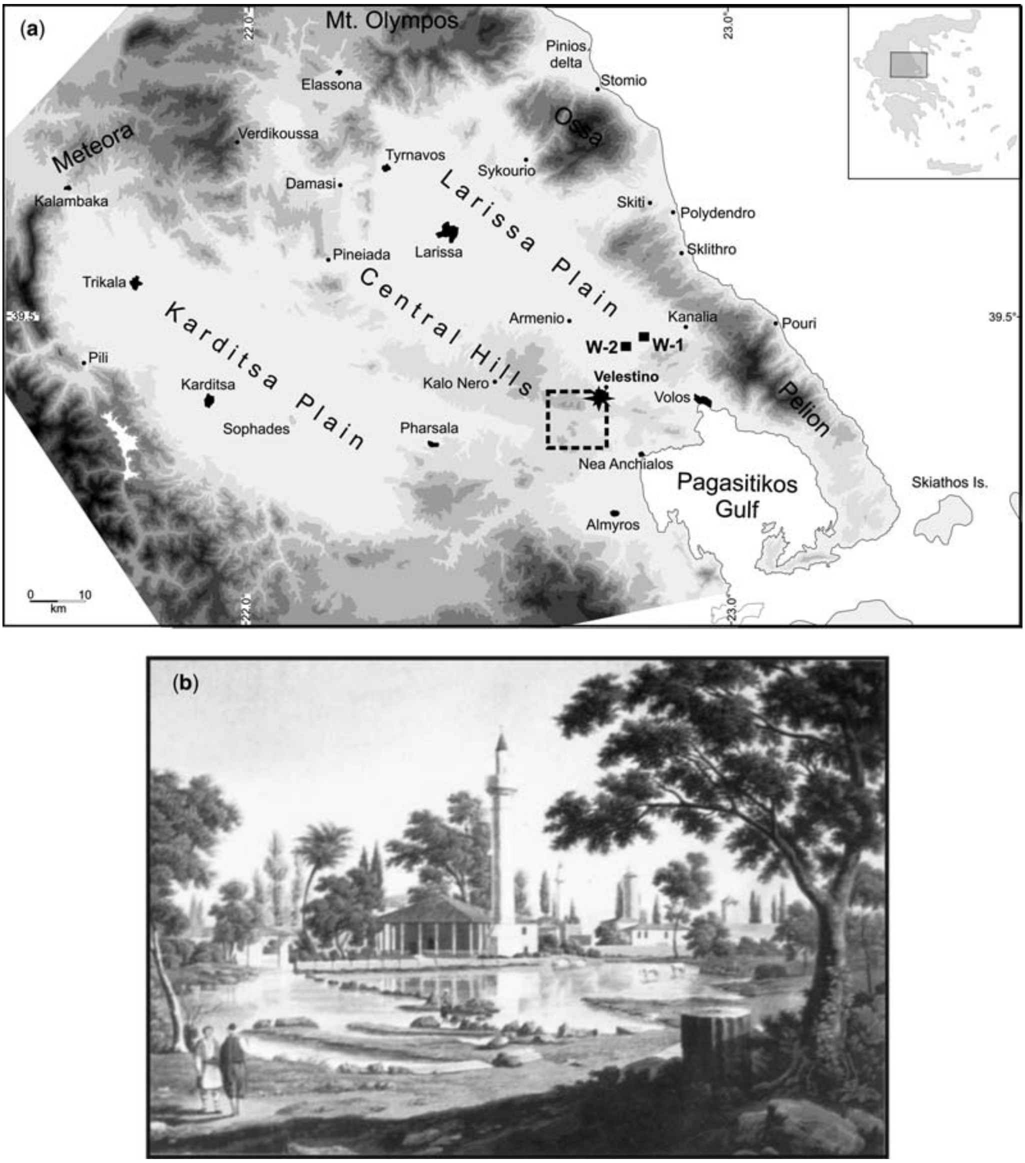

Fig. 1. (**a**) Location map of the Yperia Krini spring (star) and of wells W-1 and W-2 (squares). The box indicates the location of Figure 2a. (**b**) A representation of the spring in the eighteenth century.

consists, from base to top, of mafic rocks, limestones with rudists, a volcanoclastic sequence grading upwards into siliciclastic deposits.

Due to differential erosion, the Cretaceous limestones generate an almost continuous ridge from Mount Malouka, near Velestino, to the Chalkodonio Massif. The thickness of this carbonate stratigraphic unit is highly variable, being more than 200 m at Mount Malouka, but decreasing westwards to only a few metres (Fig. 3). Further to the west, the huge carbonate body, which generates the Chalkodonio Massif, represents a different and higher tectonic unit. It is noteworthy to say that, in this area, the two carbonate units are in direct (i.e. tectonic) contact and this observation will be a key point in the discussion about the hydrogeological system connected to the Yperia Krini spring.

During Pliocene times, when most of Thessaly was a lacustrine area, these carbonate massifs represented small islands that were never completely

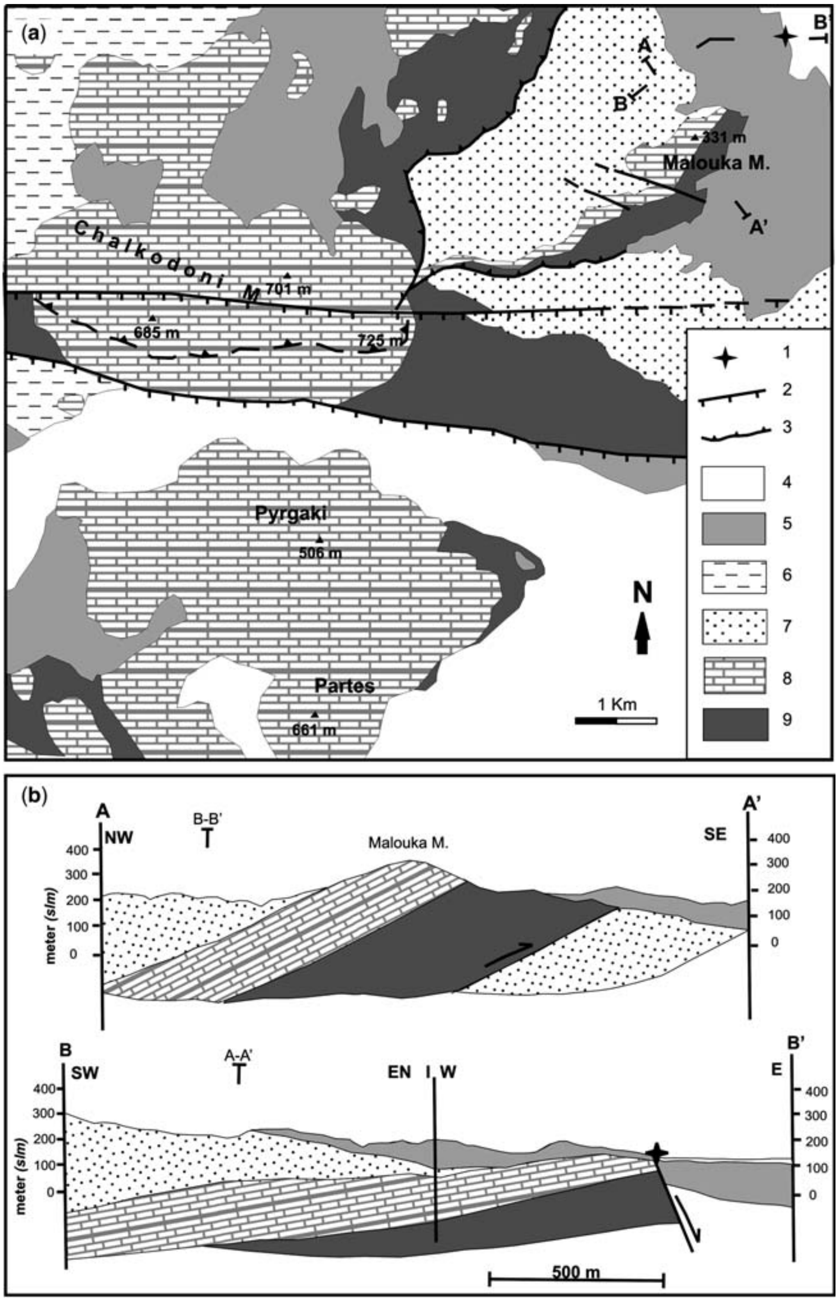

Fig. 2. (**a**) Simplified geological map of the investigated area (see Fig. 1a for location). Key: 1, Yperia Krini spring; 2, normal faults; 3, thrusts; 4, Holocene alluvial deposits; 5, Upper Pleistocene alluvial deposits (Red Beds); 6, Pliocene fluvio-lacustrine deposits; 7, Cretaceous terrigenous sediments; 8, Cretaceous limestones; 9, ophiolitic rocks. (**b**) Geological sections across the area (locations shown in (a)).

submerged nor covered by younger sediments as occurred all around there. The onlap geometry of the lacustrine deposits can be clearly observed near the carbonate massif of Kalo Nero, representing a very similar geological setting (Caputo 1990). From Late Pliocene and during the Quaternary, two major deformational events affected the region and caused the evacuation of

Fig. 3. The morphological ridge of Mount Malouka consisting of limestones with rudists characterized by highly variable thickness and representing an important 'bottleneck' within the hydrogeological system (Lm, Cretaceous terrigenous sediments; Fy, Cretaceous limestones; Oph, ophiolitic rocks).

waters from the lake, the consequent growth of a hydrographic pattern and the uplift of the Central Hills. This latter phenomenon caused the deep entrenchment of the former sediments and the deposition of thick but localized alluvial bodies along the borders of the structural highs. The town of Velestino is located along such a morphological escarpment and the Upper Pleistocene Red Beds are largely represented (Fig. 1). The continuous tectonism affecting the region caused further uplift of the area and the consequent entrenching of the youngest sediments, as can be observed immediately south of the town of Velestino (Fig. 2).

Hydrogeological framework

According to surficial observations, and due to their subaerial position for several million years, all the carbonate rocks cropping out in the area are deeply and intensely karstified (Fig. 4). Although the spring emerges from the Upper Pleistocene Red Beds, the source volume of the spring consists mainly of rocks belonging to the Mesozoic bedrock. Due to the complex geological and tectonic setting of the area uphill of Yperia Krini and its surroundings, the hydrogeological system is a typical karstic one. As a consequence, the basin and mechanism that feed the spring cannot easily be defined. In order to constrain this problem and to better define the aquifer system that feeds the spring, the hydraulic parameters and the hydrochemical characteristics have been investigated.

Climatic data

In order to define the replenishment conditions of the aquifer, we used monthly values of precipitation and mean temperature obtained from the pluviometric station of Sotirio, for the period 1971–1995 (Y.E.B. Larissa). This station is located about 15 km north of Velestino, at an altitude of 47 m a.s.l.

From the distribution of the temperature of the air, averaged monthly and annually (Fig. 5a) for the years 1973–1995, the mean annual value for the whole period is 14.5°C with variations between 12.8°C (1974 and 1975) and 17°C (1995). By averaging the temperatures of each month (Fig. 5b), the coolest month is January, with a mean temperature of 4.8°C, while the warmest month is July with a mean of 26.1°C.

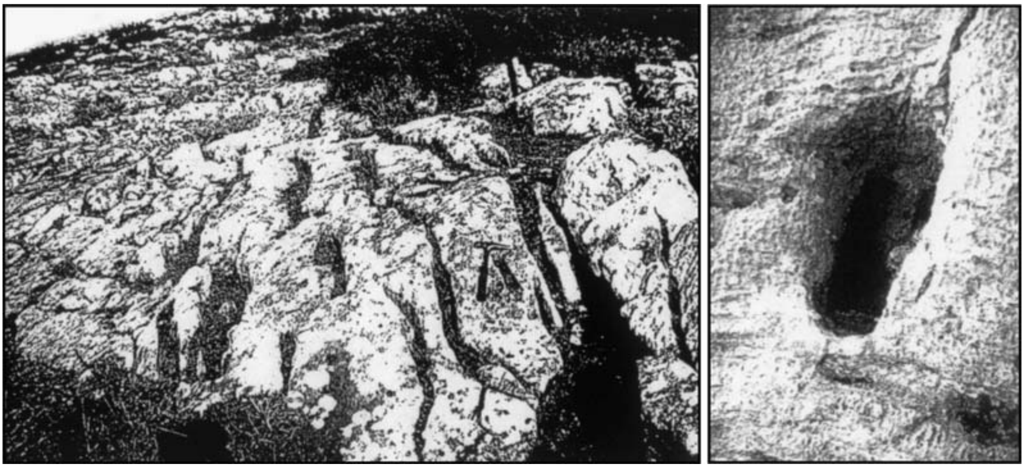

Fig. 4. Karstic features within the carbonate rocks of Mount Malouka confirming the high infiltration coefficient.

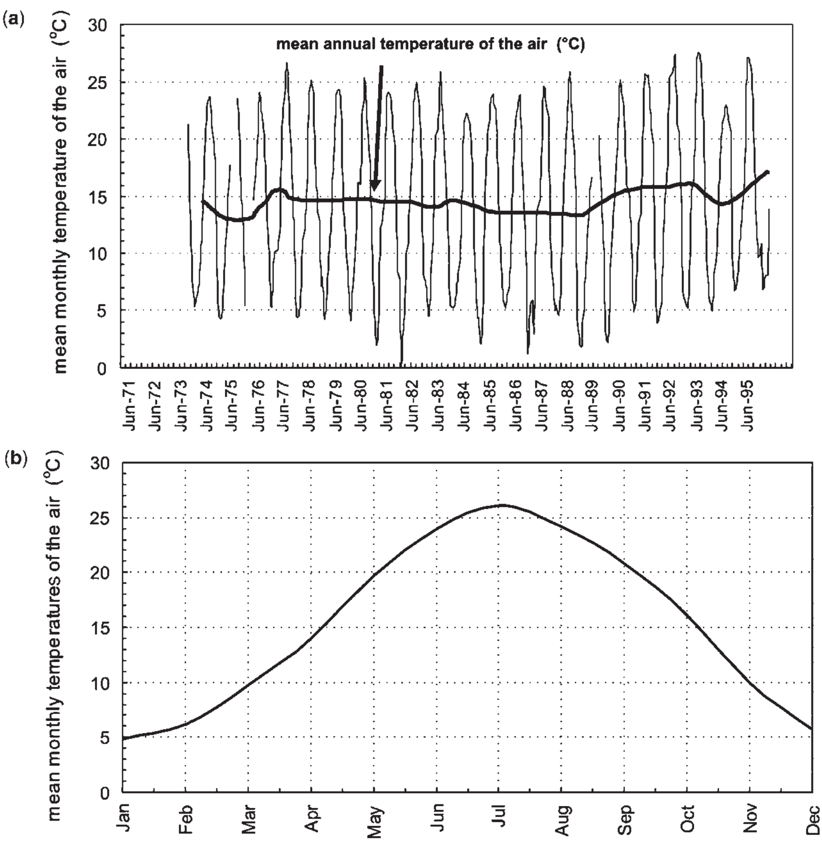

Fig. 5. Graphs representing (**a**) mean monthly and mean annual temperatures of the air and (**b**) mean monthly temperatures recorded at the pluviometric station of Sotirio for the period 1973–1995.

Similarly, from the distribution of the monthly (Fig. 6a) and annual (Fig. 6b) precipitations for the years 1971–1995, the mean annual precipitation is 420 mm with annual variation between 726 mm (1982) and 220 mm (1977). During the dry subperiod 1984–1990, the mean annual precipitation was 325 mm, representing a reduction of 22.5% with respect to the mean annual value of the whole period.

The maximum, mean and minimum values of the monthly averaged precipitation for the same period are represented in Figure 6c. The driest month of the year is August, with mean rainfall of 12.6 mm, while the wettest month is November, with 72.3 mm. The comparison of Figure 5b with Figure 6c indicates that the dry season is from June to September. During the other months, the recharge of the humidity of the soil and the consequent replenishment of the aquifer occur.

Water discharge of the spring

In order to define the hydraulic behaviour of the aquifer, monthly measurements of water discharge of the spring for the period 1973–1995 (Y.E.B. Larissa) were used. Both annual and pluriannual variations of water discharge are clearly evident in Figure 7. The mean annual discharge of the spring is 241 l/s (Fig. 8), showing three maxima during 1973, 1983 and 1987, with 335, 320 and 317 l/s, respectively, and a minimum value of 75 l/s in 1994, which corresponds to only 30% of the mean annual discharge for the whole period.

However, if we consider the subperiod 1973–1988, the minimum annual value is never lower than 230 l/s, while during the period 1989–1994 there is a continuous decrease of the mean annual discharge of the spring, with a strong reduction from 227.5 l/s to only 75.2 l/s, which also corresponds to a strong reduction of water stored in the reservoir.

By comparing the mean annual values of water discharge with those of precipitation (Fig. 8) for the period 1973–1995, we can observe that during the years 1973–1988, both positive and negative variations of the rain always induce comparable variations of the spring water discharge, though with some hysteresis. In contrast, from 1988 to 1994, while the distribution of precipitation is more or less uniform, though lower than average, the annual discharge is abated as much as the 86.5%, that is from nearly 220 l/s to 30 l/s during the summer of 1994. However, the observed rainfall reduction does not justify such a large contraction of the water discharge (Coutagne 1968). Possible explanations will be discussed later.

It is well known that the shape of a hydrogram is mainly related to the area extent of the karstic aquifer and to its hydraulic parameters (e.g. Maillet 1905; Mangin 1975; Soulios 1991; Padilla *et al.* 1994; Bögli 1994). In order to estimate (i) the volume of the groundwater stored within the aquifer and (ii) the volume of water emerging from the spring, the Maillet (1905) method is applied:

$$Q_t = Q_0 e^{-\alpha t} \tag{1}$$

where Q_t is the water discharge of the spring (m^3/s) at the time t from the beginning of the dry period, and Q_0 is the water discharge of the spring (m^3/s) at the beginning of the depletion ($t = 0$).

Equation 1 was applied to the discharge of the dry seasons within the investigated period (Fig. 6a). Due to the relatively small number of data, only a rough estimate of the hydrogeological parameters (Q_0, α) for each year can be obtained. Mean values of Q_0 and α for the whole period 1973–1995 are $0.086\ m^3/s$ and $0.002\ day^{-1}$, respectively. According to several authors (e.g. Drogue 1992; Bonacci 1987; Soulios 1985), a low α value indicates a slow drainage velocity of the karstic system, and that the water movement within the aquifer occurs via many microfractures and few wide conduits. At first glance, this inference seems to contradict the widespread karstic phenomena observed at the surface within the carbonate outcrops of the investigated area (Fig. 4).

Moreover, according to the Maillet's formula, it is possible to estimate the dynamic storage capacity, W_0, of the reservoir in the saturated zone at the beginning of the dry season ($t = 0$), corresponding to about $3.7 \times 10^6\ m^3$, and this is the volume that is progressively depleted in the case of no recharge. When this volume is completely exhausted, the spring becomes typically temporary, with water discharge occurring only when precipitation occurs.

A karstic system can be theoretically recharged by (i) infiltration of precipitation, (ii) lateral underground leakage, or even (iii) episodic additional recharge from superficial waters. Following the case, the system reacts accordingly and the discharge commonly follows some fixed law distribution (Mangin 1971, 1975). By analysing the cumulative classified values of the discharge, it is thus possible to understand the behaviour of the system, for example the recharge and discharge (Mangin 1975; Soulios 1985; Bonacci 1993).

From the cumulative distribution of the monthly values of water discharge for the years 1973–1995 (Fig. 9), it is possible to recognize if additional and temporal water outlets exist in the karstic system. Indeed, the curve in Figure 9 is clearly segmented showing two important slope variations, at about

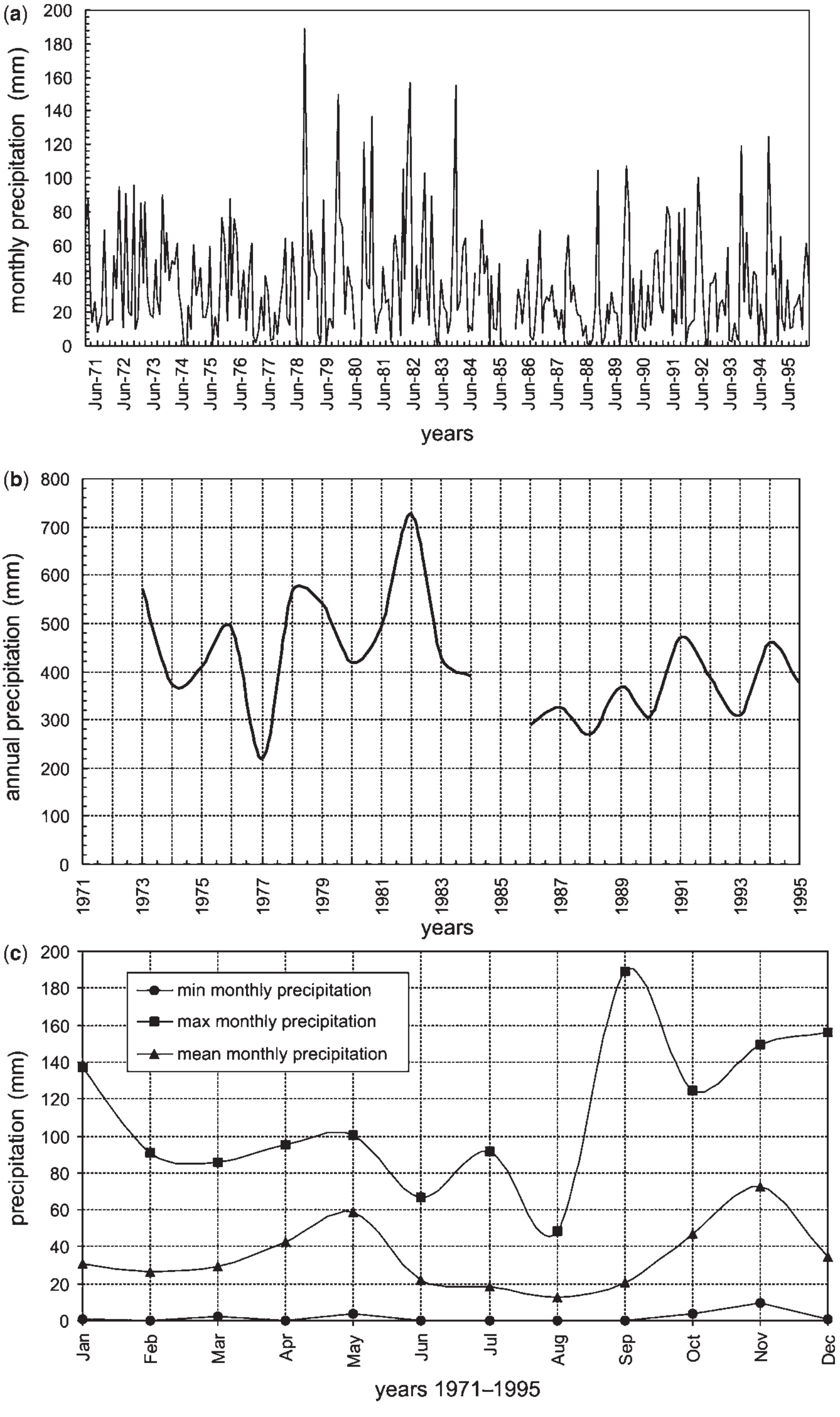

Fig. 6. Graphs representing monthly (**a**) and annual (**b**) precipitation recorded at the pluviometric station of Sotirio, for the period 1971–1995; (**c**) maximum, mean and minimum monthly precipitation for the same period.

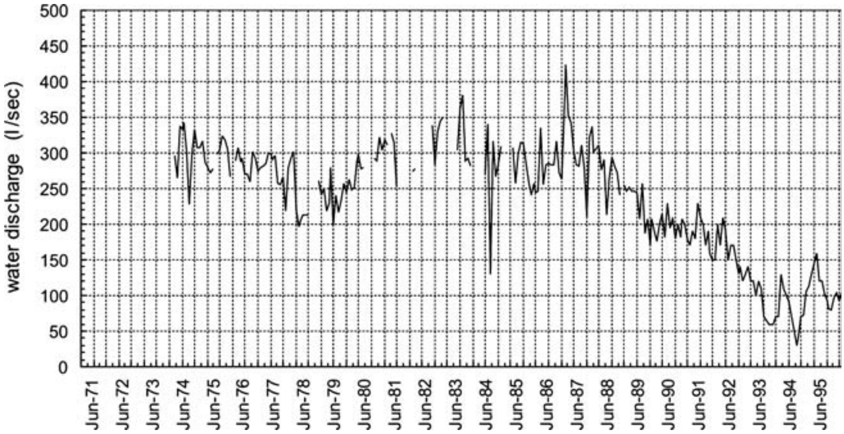

Fig. 7. Graphs representing monthly water discharge recorded at Yperia Krini spring for the period 1973–1995.

90 l/s and 280 l/s. The first variation is probably related to the drilling of new bore-holes that extract part of the water from the karstic aquifer, thus diminishing the water discharge at Yperia Krini. The second variation is possibly due to the water surplus of the karstic system that probably generates temporary outlets (Campi & Dragoni 2000).

Hydrochemical data

More information is available from the chemical analyses, particularly concerning the provenance and quality of the water (e.g. Hem 1985; Freeze & Cherry 1979; Hounslow 1995; Stumm & Morgan 1981; Lloyd & Heathcote 1985; Matthes 1982). In the present work, 12 samples with monthly frequency have been collected from the Yperia Krini spring in the years 1995 and 1996. The results of the statistical analyses are represented in Table 1. The measured pH values indicate a neutral to slightly basic water, while a total hardness of 36.5–37.8°F indicates hard water, typical of the groundwater of limestone formations (Aminot 1974; Bakalowicz 1977).

The absolute and relative values of ion content generally give useful information about the lithological characteristics of the rocks permeated by underground waters (Truesdell & Jones 1974; Plummer *et al.* 1976). The chemical analyses of the samples indicate a Ca/Mg ratio between 3.2 and 4.2, which confirms the importance of calcium-rich materials, such as carbonate rocks, within the aquifer (Hem 1985; Christopher 1992). Similar

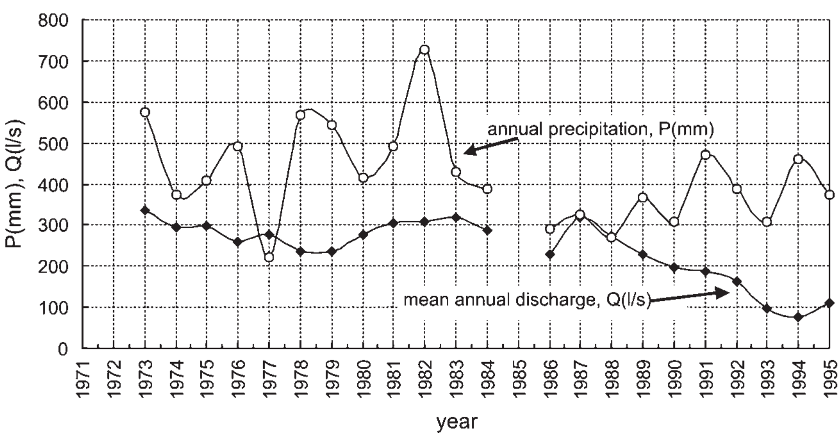

Fig. 8. Graph comparing the mean annual water discharge (Q) recorded at Yperia Krini spring and the mean annual precipitation (P) recorded at the pluviometric station of Sotirio, for the period 1973–1995.

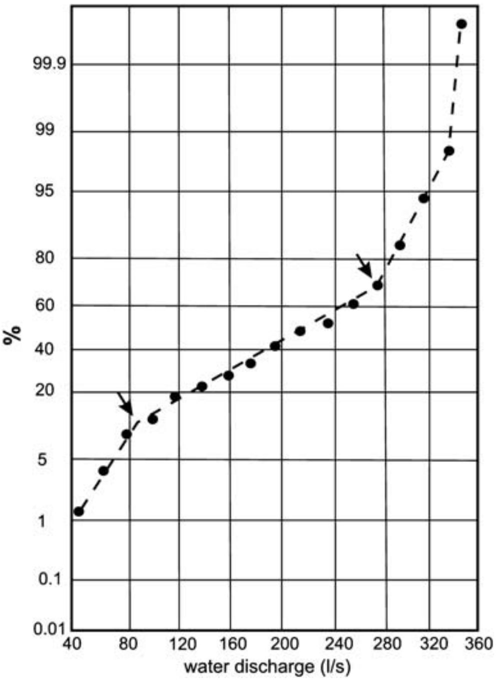

Fig. 9. Cumulative distribution of water discharge of Yperia Krini spring during the period 1973–1995.

conclusions can be inferred from the positive values of the index of saturation of calcite, dolomite and aragonite (Table 1).

According to the triangular diagram of Piper (1953) and to the classification of Davis & DeWiest (1967), the water of the Yperia Krini spring is of calcium bicarbonate type (Ca-HCO_3).

Recharge area

According to the hydrogeological and hydrochemical investigations, the Yperia Krini spring behaves as a typical karstic aquifer. Although the Yperia Krini spring emerges from the Pleistocene terrigenous Red Beds, the bulk of the water volume originates a few metres below, from the Mesozoic limestones of the Malouka Mount. As mentioned above, the carbonate rocks of the Chalkodonio Massif are clearly connected to the hydrogeological system. Nevertheless, even if we assume the maximum infiltration values proposed for karstified rocks in Greece corresponding to about 50% of precipitation (Marinos 1975; Soulios 1985; Manakos & Dimopoulos 1995), the total areal extent of these two major outcrops, which is about 13 km^2, is clearly not sufficient to supply the water discharge observed at the spring. In fact, the 465 mm/a of mean annual precipitation for the years 1971–1984, that is before the dry period and before the widespread pumping of the underground waters, can account for only 95 l/s

Table 1. *Minimum and maximum values of the major ionic concentrations in the water of Ypeia Krini*

	Min.	Max.
Temperature of water (°C)	16.6	17.3
pH	7.5	7.9
Electrical conductivity (μS/cm at 20°C)	580	603
TDS (mg/l)	368	380
Total hardness (°F)	36.5	38
Ca^{++} (mg/l)	114	120
Mg^{++} (mg/l)	17	22
K^{+} (mg/l)	14.5	15.0
Na^{+} (mg/l)	0.58	0.70
$(HCO_3)^{-}$ (mg/l)	410	420
$(SO_4)^{-}$ (mg/l)	7.2	13
Cl^{-} (mg/l)	7.1	10.4
$(NO_3)^{-}$ (mg/l)	3.5	6.0
Ca^{++}/Mg^{++}	3.2	4.5
Saturation Index:		
Calcite	+0.832	+0.850
Dolomite	+1.083	+1.190
Aragonite	+0.670	+0.830

Monthly data for the period 1995–1996.

of water discharge at the spring, against the measured average of 286 l/s. Indeed, consistent with the assumed infiltration coefficient, it is easy to calculate that the required replenishment area is about 38 km^2.

According to the geological and tectonic framework of the region, a possible replenishment area contributing to the aquifer of Yperia Krini is represented by Partes-Pyrgaki Hill, just south of the Chalkodonio Massif (Fig. 2). Indeed, under a thin cover of Quaternary deposits, the two carbonate masses are in direct hydraulic contact, though slightly displaced by the Righeo Fault, which has downthrown the southern block a few tens of metres (Caputo 1995). These additional 17.8 km^2 make a total replenishment area of about 31 km^2, which is comparable to the expected one, especially if we consider further possible minor contributions to the hydrogeological balance given by different non-carbonate superficial lithologies, such as the Red Beds or other terrigenous rocks.

As concerns the system analysis, the low value of the coefficient of recession, α, is possibly related to the presence of cavities within the unsaturated zone of the aquifer system, whose number and importance progressively decrease with depth (saturation zone). Alternatively, the behaviour of the hydrogeological system is probably due to the evident hydraulic bottleneck existing between Mount Malouka and the Chalkodonio Massif (Fig. 2), represented by the carbonate succession which is only a few metres thick and is interbedded with rocks of low permeability. These geometric characteristics of the hydraulic system explain the permanent and generally uniform behaviour of the spring. Indeed, most of the water is contained within the reservoirs of Chalkodonio Massif and Pyrgaki-Partes Hill, but its transfer to Mount Malouka, and subsequently to the spring, is evidently controlled by the abovementioned bottleneck.

Historical observations

The importance the local people attached to the spring and its relationships with the natural and economical conditions of the area can be discerned in the name the Ancients gave to this spring. According to philologists, the toponym Yperia krini (Υπέρια κρήνη) can be interpreted as a composite name by separating υπ from έρεια. Based on έρα (= earth), a term confirmed by εξ-εραω (= to outpour, to vomit, to evacuate), and two notes of Hesychius relative to ερασαι (= vacuous) and έρα (= earth; Chantraine 1966), the term υπ-εραω describes 'the water outpouring from underground resurgence', the Vauclusian fountains, commonly referred to by modern Greeks as *Kephalovrysso* (= major spring). A number of such springs are known in Thessaly. Though of different importance, all these springs marked the natural environment and attracted the establishment of human settlements (Sivignon 1974).

If this is the real meaning of the toponym, other similar springs could have been named in the same way; the term is not exclusive to Pherae, though only at Pherae has it been preserved. Homer described a spring named Yperia in Thessaly belonging to the town of Ormenion in the kingdom of Eurypylos. It is clear that this town cannot be the town of Orminion, placed on the Volos Gulf (Strabon, IX, 5, 18; in Baladié 1996), nor can the Homeric Yperia be associated with Pherae, at that time in the Kingdom of Eumelos. Consequently, in addition to the Yperia of Pherae, at least one other spring named Yperia existed in Thessaly, probably located near Ktouri, within the western plain (Decourt 1990; Helly 1995).

In any case, the inhabitants of Pherae and of Velestino were aware of the very peculiar environment they lived in, because of the Yperia fountain. This environment, which was recognized by all European travellers visiting Thessaly in the previous century, is in clear contrast to the flat and bare lands extending to the edge of the old Karla Lake. This place is characterized by a rich vegetation, small groves, orchards, and by lines of tall trees. Geographers (e.g. Sivignon 1974) define the basin of Velestino as an *huerta*, a typical environment of river deltas and of some alluvial fans. It consists of a mosaic composed of small fields and gardens abundantly drained by an intricate network of channels and numerous paths necessary to reach each field. Eventually, the fields are separated by dry-stone walls and by lines of vegetation, reeds, fig trees and poplar trees marking the pathways and channels.

It is not possible to describe in detail the history and evolution of this environment as we lack any specific information. On the other hand, we certainly know that at the beginning of the second century BC, this characteristic environment was already formed and persisted untill recently. Indeed, we know the citations of Polybius (XVIII, 20, 1, in Paton 1926) and Livy (XXXII, VI, 6–7, in Sage 1961) about the military operations preceding the great battle of Cynoscephalai in 196 BC. The two armies of the Romans and that of Philip V of Macedonia were in the region of Pherae looking for the ideal terrain for fighting. The historiographers say that during the military reconnaissance several 'contacts' occurred between the enemies, when turning around the walls of the gardens. These contacts were unexpected because of the high vegetation and of the noise produced by the water flowing in the channels. Due to the frequency of these accidents and the realization that this environment was clearly not favourable to draw up the armies and especially the cavalry, the two commanders were forced to move towards the hills of Revenia, near Skotoussa, where the final battle took place.

With the exception of this precise witness, our knowledge about possible variations of this environment, its areal extent and even the activity of the Yperia spring is almost nil. Nevertheless, in the imagination of the inhabitants today and in the analyses of the historians, the spring is considered to have always existed, giving abundant and permanent water. The water discharge was probably not constant and past periods of decreasing discharge or even of complete dryness cannot be excluded. Because the written sources do not allow us to solve this problem, we attempt to use three different arguments.

Firstly, from an anthropological point of view, the legends and traditions of the inhabitants of Velestino continuously attest to fear of the disappearance of water. When such a feeling is widespread, we can infer that the phenomenon can really occur. Indeed, it is the anxiety of a collective unconsciousness, whose origins are possibly several centuries back in time. In general, all these legends are born from a situation which actually occurred. Whenever it occurred, long ago or yesterday, the possibility should not be neglected.

Secondly, the recent discovery of the remains of a post-Byzantine or Turkish (fifteenth to sixteenth century) aqueduct, used to bring water from the mountain near Velestino down to the town (Di Salvatore 1994), poses the same question. In fact, the construction of such an edifice clearly indicates the need to improve the delivery of water to the inhabitants and to the environment. This fact can be interpreted in two opposite ways: either the need for water was greatly increased due to a comparable demographic increase, or the spring no longer provided enough water to satisfy the demands of the inhabitants. This situation generates a potential danger for the whole community. The construction of an aqueduct of this importance implies the mobilization of considerable human and economic resources. The latter hypothesis seems to better explain the question we pose about the permanence of the Yperia Krini spring, because it is very likely that the spring experienced periods of high and low water discharge.

Thirdly, a further aspect of the question should be analysed for completeness. As mentioned above the life of the town is closely related to the spring. Therefore, when migrations or phases of desertion occurred during the history of Pherae, such as that recorded at the beginning of the first century BC (Helly 1976, 1980), it is likely that a critical decrease in activity of the Yperia fountain directly caused the decline of the ancient town, and also affected the surrounding territory. Indeed, recent archaeological excavations (Intzesiloglou 1994) clearly show that at least part of the ancient town was abandoned during the first years (or decades) of the first century BC. Moreover, according to Helly (1976), there is evidence that the town of Pherae almost completely disappeared during the reign of Emperor Augustus, at least as a political entity. In fact, an inscription at Larissa states that the entire territory of Pherae and everything living on it, such as men and animals, were totally enclosed within the *patrimonium Augusti* (Helly 1976, 1980). The territory of Pherae was thus integrated within an imperial domain whose administrator, Sebastou Oikonomos, during the second century AD resided in Demetrias.

The eclipse of Pherae as a political entity is also emphasized by the fact that in antiquity the town was never an episcopate, in contrast to all other Thessalian towns which persisted since the end of the fourth century AD (Helly 1976).

Discussion

Based on the analysis of the aquifer system feeding the Yperia Krini spring, it is likely that the strong decrease of water discharge during the last decade was caused by the drilling of numerous new bore-holes for the water supply. In fact, these bore-holes pump significant volumes of water directly from the aquifer, thus exceeding the capacity of the reservoir to recover (Fig. 10).

On the other hand, it seems clear that the present-day Yperia Krini crisis should be included in the more regional hydrogeological problem that affects the broader Larissa Plain. However, as discussed, the spring has already suffered periods of partial (or complete) dryness in historical times, though it is not possible to cite the drilling of bore-holes as a possible cause of previous dry events.

In general, the quality of life and the periods of expansion and reduction of the old town were directly related to the amount of water discharge. Moreover, the disappearance of at least two ancient towns at the end of the Hellenistic epoch or at the beginning of the Imperial epoch, matches that of Pherae. Therefore, although the local conditions are different and there is not perfect chronological coincidence between the three events, the causes responsible for this phenomenon are likely to be regional ones. Indeed, in the case of Pherae, it is likely that the loss of the spring Yperia induced the abandonment of the town.

Two different, but not alternative, explanations for this phenomenon are presented in the following. Firstly, a possible unique natural cause occurring at

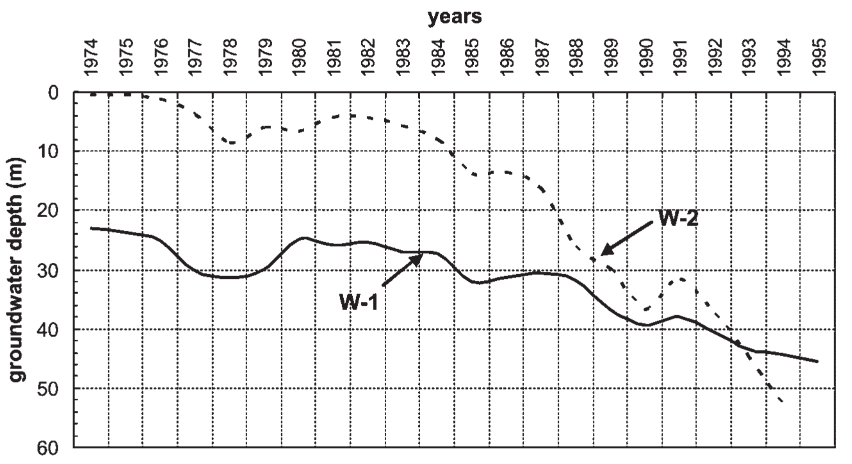

Fig. 10. The groundwater depth of wells W-1 and W-2 between 1974 and 1995 showing the strong drawdown of the piezometric level in the confined alluvial aquifer system east of Yperia Krini. See Figure 1 for location of the wells.

the regional scale and generating instability in this sector of the plain, is the active tectonism which has affected the area since the Middle–Late Pleistocene (Caputo 1990, 1995; Caputo & Pavlides 1993). Although no specific data are available to support this hypothesis for the study area, a close relationship between the variations of the stress field, as induced by earthquakes, and the underground water level is well documented world-wide (Roeloffs 1988; Oki & Hiraga 1988; Kümpel 1992; Asteriadis & Contadakis 1994).

Secondly, it is well known that from the third century BC to the fourth century AD the northern hemisphere underwent the so-called 'Roman climatic optimum' (Panizza 1985; Veggiani 1994). However, if the term 'optimum' can certainly refer to Europe where ice-caps retreated, in contrast in the Mediterranean regions, this warm period evidently generated drier climatic conditions and thus a generally reduced amount of precipitation (Greig & Turner 1986). For this reason, we attempted to analyse the response of our hydrogeological system in the case of dry periods.

Let us consider the relationships between mean annual precipitation and mean water discharge (Fig. 8). For example, just one year of reduced precipitation (1977), with 220 mm/a corresponding to about 50% of the average value for the period 1971–1984 (465 mm/a), caused a decrease of more than 15% in water discharge for as long as two years (1978 and 1979).

Accordingly, whatever the real amount is, it is clear that a decrease of precipitation, of less than 50% but lasting several centuries, such as occurred during the 'Roman' epoch (*c.* third century BC to fourth century AD; Lespez 2003; Bottema 1979; Digerfeldt *et al.* 2000; Dragoni 1998), certainly generated a progressive depletion of the reservoir and a consequent strong reduction in water discharge of the Yperia Krini spring. Based on the flourishing environmental conditions still existing during 196 BC, the degradation phenomenon was necessarily very slow and probably not perceivable during a human lifetime. Nevertheless, in order to induce a migration of the population from Pherae and consequent decay of the old town, it is obviously not necessary for a drastic reduction of the water discharge to have occurred. Indeed, during the first century BC, after 300 years of warm climate, the spring probably did not provide enough water for a large number of inhabitants.

Concluding remarks

The integrated analysis of geological and hydrogeological data indicates that the Yperia Krini spring has little storage capacity and that it is very sensitive to rainfall variations. On the other hand, historical information indicates that important climatic changes have influenced the whole Mediterranean area in the past, affecting the Yperia Krini spring in such a way that the broader area suffered a strong social and political decline.

During the last few decades, the concomitance of a relatively dry period (i.e. low precipitation) affecting the Aegean region and the huge increase of cotton cultivation in large sectors of Thessaly, central Greece, has forced local people to search for new water resources to provide their needs. Accordingly, during the 1970s and 1980s, the synergy of the natural phenomenon with inappropriate agricultural choices was coupled with the drilling of a huge number of bore-holes across the entire region. These water-wells, which mainly exploit the alluvial multi-aquifer system of the Larissa Plain, caused a significant drop of the piezometric level (Fig. 10) that critically exceeded the capacity to regenerate the underground natural hydraulic conditions. As a consequence, a quantitative and sometimes qualitative degradation of the water resources occurred. The process was also associated with differential subsidence phenomena and ground fissuring that caused damage to structures and troubled the local people (Soulios 1997; Rapti-Caputo & Caputo 2004). The recent period of reduced rain suggests the occurrence of low discharge values for the next few years. Therefore, it is highly probable that the tendency to drill new wells to counteract the drought will further accelerate.

According to the scenarios of the Intergovernmental Panel on Climate Change (IPCC 2001*a, b*), an increase of the global surface temperature in the range of 1.7 to 4.0°C will occur by 2100, while the temporal and spatial behaviour of precipitation will be altered.

In particular, the average surface temperature in the Mediterranean area is expected to increase by 0.7–1.6°C per 1°C of global increase (Mitchell & Hulme 2000; Palutikof *et al.* 1996; Cubasch *et al.* 1996; Barrow *et al.* 1995; Palutikof & Wigley 1996; Karas 1998). According to most prediction models, it is expected that precipitation will decrease in several sectors of the Mediterranean area southern of latitude 40–45°N and will increase in areas north of 45°N (Giorgi & Fransisco 2000; IPCC 2001*a*; Palutikof *et al.* 1996).

Based on five climatic models, Mitchell & Hulme (2000) suggest that over the period 1990–2100 the mean temperature in Greece will increase from 3.1°C to 5.1°C with an average of 4.3°C. Changes in precipitation are more difficult to predict, since Greece is located in a transition climatic zone. Notwithstanding some differences, most models agree that precipitation will likely decrease during the summer months, while there is a larger probability that precipitation will increase in the northern Mediterranean sectors (ECSN 1995; Palutikof *et al.* 1996; Perissoratis *et al.* 1996; IPCC 2001*a*; Feidas & Lalas 2001).

As a consequence of (i) the structure of the spring and its small reserve, (ii) the possible future increase of temperature and (iii) the predicted decrease of precipitation, a decrease of the average spring discharge is to be expected, similar to what occurred in past centuries. Accordingly, the drilling of new bore-holes will not efficiently counter the impact of the reduced availability of water, while new strategies, such as rain harvesting, wastewater reuse and aquifer recharge, should be adopted to improve the exploitation of the underground resources.

We are indebted to M. Thanos, former geologist at the Y.E.B. of Larissa, for providing meteorological data of the pluviometric station of Sotirio. G. Soulios and G. Dimopoulos of Aristostotle University of Thessaloniki (Greece) provided facilities for the hydrochemical analyses and many fruitful discussions. Thanks to R. Caputo for discussions on the geological data. Useful comments and criticisms by two anonymous reviewer are also acknowledged.

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Quantifying the impacts of climate change on groundwater in an unconfined aquifer that is strongly influenced by surface water

J. SCIBEK[1], D. M. ALLEN[1] & P. H. WHITFIELD[2]

[1]*Department of Earth Sciences, Simon Fraser University, Burnaby, British Columbia, Canada V5A 1S6 (e-mail: dallen@sfu.ca)*

[2]*Meteorological Service of Canada, Environment Canada, Pacific and Yukon Division, Vancouver, British Columbia, Canada V6C 3S5*

Abstract: A three-dimensional transient groundwater flow model, implemented in MODFLOW, is used to quantify the impacts of climate change on groundwater in an unconfined aquifer with demonstrated strong connection to surface water (Kettle and Granby Rivers). The Grand Forks aquifer is located in a semi-arid region of south-central British Columbia, Canada. Distributed recharge is modelled using HELP, driven by the LARS-WG stochastic weather generator, and stage–discharge curves for rivers are modelled using BRANCH and calibrated to historical data. For recharge modelling, three year-long climate scenarios were run, each representing one typical year in the present, and future (2020s and 2050s), by perturbing the historical weather according to the downscaled CGCM1 global climate model results. By the 2050s the largest increase in recharge relative to present occurs in late spring, by a factor of three or more, a 50% increase in summer months in most areas of the aquifer, a 10–25% increase in autumn, and a reduction in recharge in winter. CGCM1 downscaling was also used to predict basin-scale runoff for the Kettle River. Future climate scenarios suggest a shift in the hydrograph peak to an earlier date, although the peak flow remains the same, and baseflow level is lower and of longer duration. Groundwater levels near the river floodplain are predicted to be higher earlier in the year due to an earlier onset of peak flow, but considerably lower during the summer months. Away from rivers, groundwater levels increase slightly due to the predicted increase in recharge.

Water resources are central to any study on climate change. In areas that rely heavily on groundwater, for example for agricultural, domestic or industrial use, it is important that the potential impacts of climate change be assessed so that adaptation measures can be taken, if needed. One concern of water managers and government officials is the potential decrease of groundwater supplies under future climate change. Another is the potential impact to streams that are fed by groundwater at periods of low flow.

Most research to date has been directed at forecasting the potential impacts to surface water hydrology (e.g. Whitfield & Taylor 1998), while only large, regional and coarse-resolution models have been undertaken to determine the sensitivity of groundwater systems to changes in critical input parameters, such as precipitation and runoff (York *et al.* 2002; Yusoff *et al.* 2002). There are a few exceptions of very small aquifers and detailed investigations of potential impacts of climate change (scenarios) on unconfined aquifer water levels (e.g. Malcolm & Soulsby 2000). Thus, there is still a need for high-resolution, local-scale, realistic models, linked to most up-to-date predictions, which can be useful to groundwater managers.

It is expected that changes in temperature and precipitation will alter groundwater recharge to aquifers, causing shifts in water table levels in unconfined aquifers as a first response to climate trends (Changnon *et al.* 1988; Zektser & Loaiciga 1993). Traditionally, aquifer recharge has been difficult to estimate for large areas; however a variety of methods have been used (Simmers 1998), from statistical empirical models linking precipitation trends to aquifer recharge and groundwater levels (Chen *et al.* 2002), to spatially distributed recharge applied to three-dimensional groundwater flow models (Jyrkama *et al.* 2002). For the purposes of climate change impacts modelling, the recharge rates must be as accurate as possible to accurately represent the small shift from present to future climatic conditions.

Also of interest in climate change impacts studies are coupled hydrologic systems, where changes in surface flow regime and changes in recharge to groundwater interact to affect groundwater and surface water levels. Groundwater may contribute to baseflow in streams; therefore, a change in the groundwater regime could have detrimental environmental effects on fisheries and other wildlife due to the altered baseflow dynamics (Bredehoeft *et al.* 1982; Gleick 1986).

From: DRAGONI, W. & SUKHIJA, B. S. (eds) *Climate Change and Groundwater*. Geological Society, London, Special Publications, **288**, 79–98.
DOI: 10.1144/SP288.07 0305-8719/08/$15.00

Each aquifer has different properties and requires detailed characterization and, eventually, quantification (e.g. numerical modelling) of these processes, as well as linking of the recharge model to climate model predictions (York *et al.* 2002). In practice, any aquifer that has an existing and verified conceptual model, together with a calibrated numerical model, can be assessed for climate change impacts through simulations. The accuracy of predictions depends largely on scale of project and availability of hydrogeologic and climatic datasets.

This paper summarizes the methodology and describes the results of a climate change impacts study of an unconfined aquifer that is strongly influenced by surface water. We present a case study of a small regional unconfined aquifer (34 km^2), contained within the mountainous valley of the Kettle River near the City of Grand Forks in south-central British Columbia (BC), Canada (Fig. 1). The aquifer consists of glaciofluvial sediments overlying glaciolacustrine sediments, which partially infill steep and variable bedrock topography. The climate is semi-arid and most precipitation occurs in summer months during convective activity. Groundwater is used extensively for irrigation and domestic use (Wei *et al.* 1994).

The study was motivated by the Canadian government's initiative to assess impacts of climate change and develop adaptive strategies for climate change under the auspices of Natural Resources Canada's Climate Change Action Fund. The goal of the study was to permit a more comprehensive evaluation of water budgets, incorporate seasonal changes in demand for groundwater, and provide a better understanding of the direct impact of climate change on alluvial aquifers. This study extends the methodology used by Allen *et al.* (2004*a*), through the use of spatial analysis tools in a GIS environment and the development of a transient groundwater flow model.

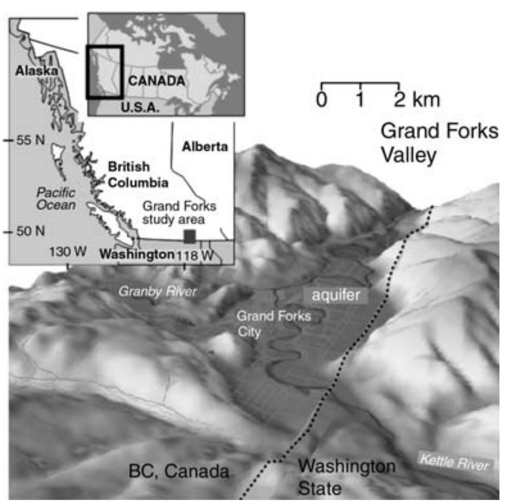

Fig. 1. Shaded relief map of the Grand Forks valley in British Columbia (BC), Canada. Inset maps show the location of the study area within BC and within Canada.

Aquifer model development

The valley shape was modelled using profile extrapolation, constrained by well lithologs and geostatistical interpolation. The valley attains a maximum depth of 250 m below ground surface, but typical sediment thickness is about 100 m. The hydrostratigraphy of the aquifer was interpreted from selected high-quality well lithologs, with layering constrained by the Quaternary depositional history of the valley sediments. Approximately 150 well lithologs are used to constrain the hydrostratigraphic model, mostly for shallow groundwater wells (Fig. 2a). The unconsolidated sediments thicken toward the middle of the valley, and have presumed horizontal stratigraphy. The topmost coarse-grained sediments form the Grand Forks aquifer.

Hydrostratigraphic units were modelled in three dimensions from standardized, reclassified, and interpreted well borehole lithologs. Solid models were constructed using GMS software v. 4.0 (Brigham Young University 2002), converted to a five-layer system underlain by solid bedrock (Fig. 2b), and imported into Visual MODFLOW (Waterloo Hydrogeologic Inc. 2004), as is typically done with complex multi-layer aquifer systems (Herzog *et al.* 2003). Details of model construction are described in Allen *et al.* (2004*b*). Representative homogeneous and isotropic hydraulic properties were initially assigned to each layer, based on values determined from pump test data, but later modified slightly during model calibration.

GCM climate predictions

Downscaling

For many climate change studies, scenarios of climate change derived directly from global climate models (GCMs) are of insufficient spatial and temporal resolution. Spatial downscaling techniques (Hewitson & Crane 1996; Wilby & Wigley 1997) are used to derive finer resolution climate information from coarser resolution GCM output. The fundamental assumption behind all these methods is that the statistical relationships, linking observed time series to GCM variables, will remain valid under future climate conditions.

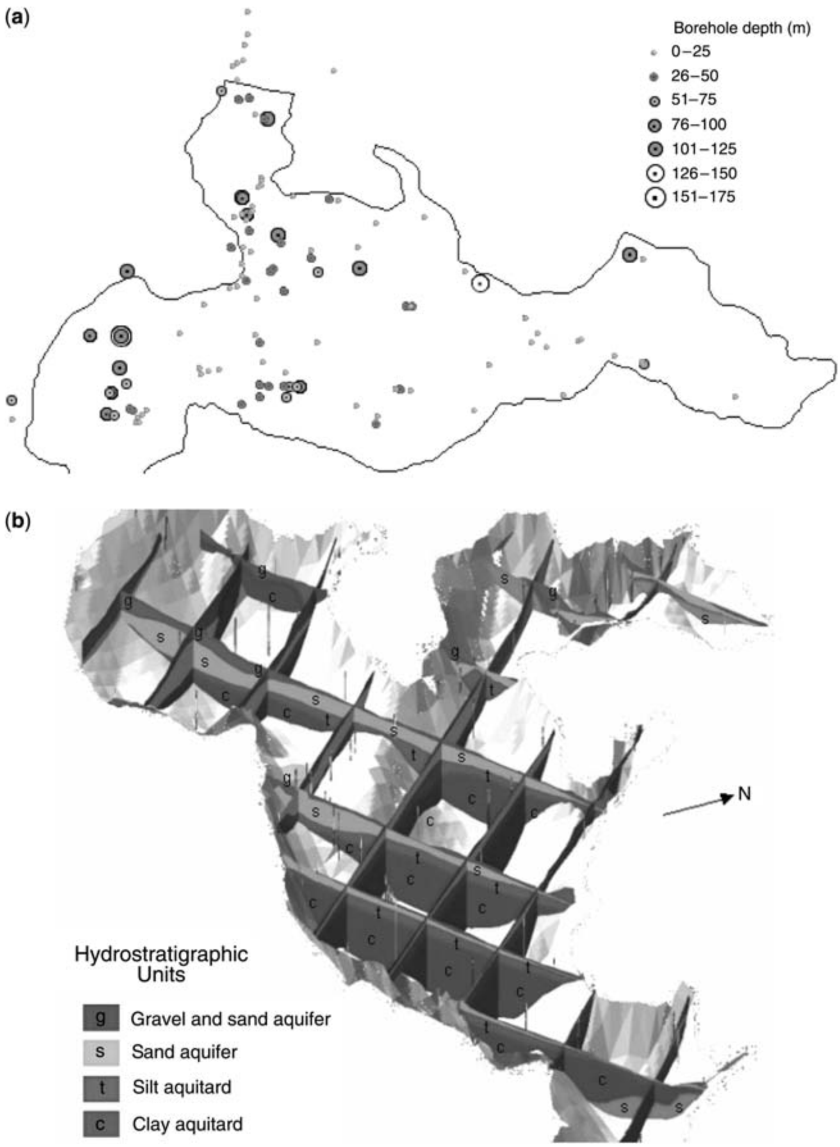

Fig. 2. (**a**) Borehole locations and depths in Grand Forks valley. Only boreholes with lithologs in BC water well database are shown, (**b**) Fence diagram of hydrostratigraphic units in the Grand Forks valley.

GCMs do not accurately predict local climate, but the internal consistency of these physically based climate models provides most likely estimates of ratios and differences (scaling factors) from historical (base case) to predicted scenarios (Loaiciga *et al.* 1996) for climatic variables, such as precipitation and temperature.

Climate scenarios for modelled present and future conditions were taken from the Canadian Global Coupled Model (CGCM1) (Flato *et al.* 2000) for the IPCC IS92a greenhouse gas plus aerosol (GHG + A1) transient simulation. CGCM1 predictions are valid for Canada and fall in the average of other GCMs. These include relative and absolute changes in precipitation and temperature. Precipitation variables were: mean, median, maximum, variance, dry/wet spell length, and percentage wet days in the month. Temperature statistics included: mean, median, minimum, maximum, variance, and interquartile range. Four daily data sets for CGCM1 were obtained from the Canadian Institute for Climate Studies

(CCIS 2004) for a grid location nearest to Grand Forks (Y = 11, latitude 50.09°N and X = 16, longitude 120°W; Grand Forks is at 49.1N and 118.2W). Three were CGCM1 scenarios, each with data for a number of potential predictor variables. The 'current climate' scenario was generated by CGCM1 for the period 1961–2000. The subsequent 'future climate' experiments using CGCM1 with GHG + A1 were for the 2020s, 2050s and 2080s.

The fourth data set was a calibration data set, which contains observed daily data for 1961–2000, derived from the NCEP (National Centre for Environmental Prediction) re-analysis data set (Kalnay *et al.* 1996) for the period 1961–2000. This dataset provides large-scale climate variables that can be used to define analogues with GCMs for climate modelling purposes. Most climate modelling experiments in North America use the NCEP datasets for calibration of downscaling models. Monthly means and other statistics were calculated from mean daily values, and the NCEP dataset had 10% or smaller bias to observed precipitation at Grand Forks (compared monthly means), thus we have high confidence in using NCEP data for calibration of downscaling model. The NCEP dataset includes relative humidity, whereas CGCM1 datasets do not, so specific humidity was used when calibrating the model.

The downscaling of CGCM1 results was accomplished using two independent methods: (1) statistical downscaling model (SDSM) software (Wilby *et al.* 2002) and (2) principle component K-nn method (e.g. Yates *et al.* 2003; Whitfield & Cannon 2000); the results of both were compared. Four climate scenarios (30 years of daily weather) were generated using each calibrated downscaling model: current climate (1960–1999), 2020s climate (2010–2039), 2050s climate (2040–2069), and 2080s climate (2070–2099).

Downscaled daily temperature time series were analysed for (1) mean, and (2) standard deviation. Predictions for mean monthly temperature are almost identical between SDSM and K-nn and calibration bias is small (Fig. 3). Similarly, both methods agree in the magnitudes and directions of temperature change, and represent an increase of approximately 1 °C per 30 years for all months. Downscaled daily precipitation time series were analysed for: (1) mean monthly precipitation, (2) standard deviation in daily precipitation, (3) percentage wet days, (4) dry series length, and (5) wet series length. SDSM results for precipitation differ from K-nn downscaling results, with mean precipitation variation being somewhat better represented by SDSM compared to observed, especially for winter and spring seasons (Fig. 4). SDSM predicts an increase in summer precipitation in future climate, but K-nn predicts a decrease at the same time. K-nn also predicts larger precipitation increases in winter than SDSM, and more precipitation variability into the future.

Clearly, the choice of downscaling method is very important for interpreting predictions of GCM models, as GCMs do not directly model precipitation at a local site. It should be noted that at Grand Forks, the local climate is not modelled very well in the CGCM1 grid cell, probably due to local convective precipitation and valley–mountain–rainshadow effects, which have a strong influence on local precipitation. The poor

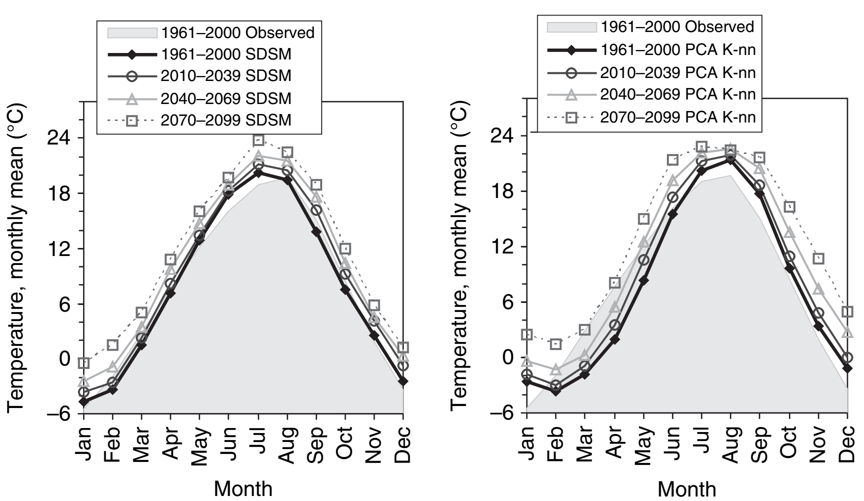

Fig. 3. Mean monthly temperature at Grand Forks, BC: observed and downscaled from CGCM1 model runs for current and future climate scenarios using (**a**) SDSM and (**b**) K-nn.

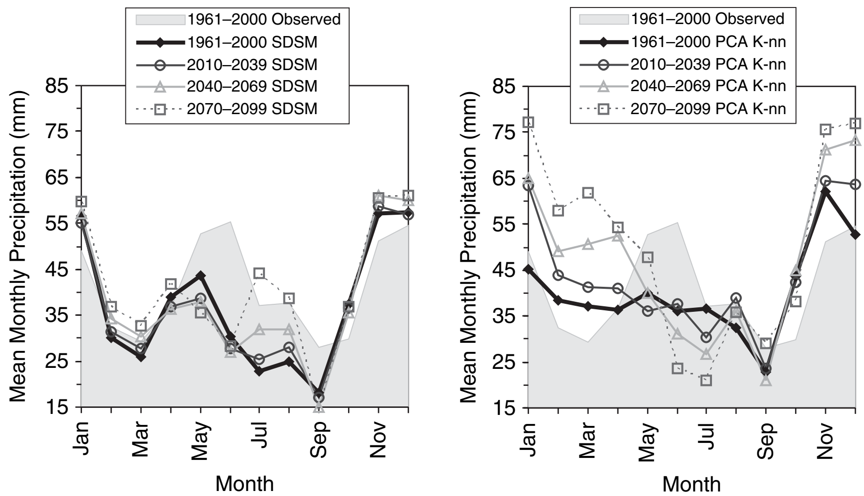

Fig. 4. Mean monthly precipitation at Grand Forks, BC: observed and downscaled from CGCM1 model runs for current and future climate scenarios using (**a**) SDSM and (**b**) K-nn.

downscaling results for precipitation did not allow us to use these data directly in a recharge model. Our approach was to compute change factors (absolute for temperature and relative for precipitation), and redistribute them to daily time series using a stochastic weather generator. An important assumption is made that the GCM can predict these absolute and relative changes, which then can be used to perturb current weather to arrive at future weather conditions. Although this uncertainty limits the predictive aspect of this (and similar) studies, it does not detract from the study's usefulness as a realistic sensitivity analysis to potential climate change, whatever the actual climate changes in each month will be in the future.

For the purpose of this study, only SDSM downscaled results were selected for further modelling of weather and recharge as inputs to groundwater flow models. Details of the SDSM downscaling, calibration and a comparison of the results are provided by Allen *et al.* (2004*b*). CGCM1 downscaling was also used to predict basin-scale runoff for the Kettle River upstream of Grand Forks (Whitfield & Cannon 2000) as described later in this paper.

Weather generation using LARS-WG

At present, output from GCMs is of insufficient spatial and temporal resolution and reliability to be used directly in hydrologic models. A stochastic weather generator, however, can serve as a computationally inexpensive tool to produce multiple-year climate change scenarios at the daily time scale, which incorporates changes in both mean climate and climate variability (Semenov & Barrow 1997). Stochastic weather ensures that daily values of variables are realistic, consistent, site-specific, and preserve both values and variability predicted to change from current to future climate scenarios by GCMs.

LARS-WG is a stochastic weather generator that can be used for the simulation of weather data at a single site (Semenov *et al.* 1998). LARS-WG is based on the series weather generator, which utilizes semi-empirical distributions for the lengths of wet and dry day series, daily precipitation and daily solar radiation. According to Semenov *et al.* (1998), the wet/dry time series are better represented in LARS-WG than in WGEN (Richardson & Wright 1984) and other similar weather generators. WGEN, which is the weather generator included in HELP (the hydrologic model used in this study for estimating recharge), has been known for inadequate modelling of persistent wet or dry periods (Wilks & Wilby 1999). In contrast, the serial weather generators (e.g. LARS-WG) avoid this shortcoming. These models determine sequences of dry and wet series of days, and then generate other climatic variables.

A comparison of LARS-WG and WGEN was undertaken as part of this study and, ultimately, the LARG-WG was found to more accurately reproduce the historic dataset. The base case is here defined as the average of the entire historical period, assuming that it is representative of pre-climate change conditions. Then, climate change scenarios were generated by perturbing the generated weather using the change factors to modify the base case. Each scenario consists of 100 years of generated weather, noting that although generated weather runs of 1000 years converge better to specified climate 'normals', there are diminishing returns of performance after 100 years. The length of generated weather time series is not meant to model actual changing climate year-to-year, but rather to model climate change step-wise for each scenario, and to generate a long enough weather time series to preserve and properly represent statistical properties for the site and the specified climate for the scenario. Averages were computed for monthly and annual data.

Simulated precipitation data compare favourably to observed normals for precipitation mean monthly amounts and precipitation variability (Fig. 5a), although variability in May and July were under-predicted. The LARS-WG reproduced air temperatures very precisely compared to the observed records (Fig. 5b). However, in winter months, LARS-WG produced 0.5 to 1.0°C cooler minimum temperatures than observed, when comparing variability in monthly values. Solar radiation was similarly very well reproduced using LARS-WG (Fig. 5c). Modelled mean solar radiation values were within 1% of observed values. Daily variability in daytime solar radiation was also reasonably well preserved in the stochastic weather model, although daily values were under-predicted by 5 to 10% compared to observed. This under-prediction might cause small error in evapotranspiration estimates in the HELP recharge model, once the LARS-WG weather is input into HELP.

Weather generation for climate change

As stated earlier, the base case is defined as the average of the entire historical period, assuming that it is representative of pre-climate change conditions. Then, climate change scenarios are generated by modifying the base case climatic time series (perturbing the weather) within LARS-WG. The following parameters were modified in LARS-WG according to the downscaled GCM results: precipitation relative change (future/base) or (base/base); wet spell length relative change; dry spell length relative change; absolute temperature change; standard deviation of relative temperature change; and absolute change in solar radiation. Each scenario consists of 100 years or more of simulated weather, from which monthly averages are calculated.

Recharge modelling

Spatially distributed recharge

There are many methods for recharge modelling (York *et al.* 2002), but the methodology presented here generates spatially distributed and temporally varying recharge zonation using a GIS linked to the one-dimensional software HELP (US EPA Hydrologic Evaluation of Landfill Performance model) (Schroeder *et al.* 1994). The program WHI UnSat Suite Plus (Waterloo Hydrogeologic Inc. 1999), which includes the sub-code Visual HELP, is used to estimate recharge to the Grand Forks aquifer. HELP is a versatile quasi-two-dimensional model that can be used for estimating groundwater recharge rates. Inputs consist of a representative sediment column with defined soil and sediment properties, surface slope, meteorological conditions, and evaporation controls. HELP uses numerical solution techniques that account for the effects of surface storage, snowmelt, runoff, infiltration, evapotranspiration, vegetative growth, and soil moisture storage. The natural water balance components that the programme simulates include precipitation, interception of rainwater by leaves, evaporation by leaves, surface runoff, evaporation from soil, plant transpiration, snow accumulation and melting, and percolation of water through the profile.

The approach used is similar to that of Jyrkama *et al.* (2002), in which a methodology was developed for estimating temporally varying and physically based recharge using HELP for any MODFLOW grid cell. Our approach also depends on high resolution GIS maps used to define recharge zones, and linking them to MODFLOW model grids. It differs from previous distributed-recharge methods in that we also estimate the distribution of vertical saturated hydraulic conductivity in the vadose zone, the thickness of the vadose zone, and irrigation return flow, all at high spatial resolution. The temporal inputs are derived from the LARS-WG stochastic weather generator, as opposed to the internal weather generator, WGEN, with parameters derived from downscaled GCM predictions. The GIS software used was ArcGIS version 8.13 (ESRI 2004), and all map processing was done on 20 m raster grid cells. Details concerning the distributed recharge modelling are provided in Scibek & Allen (2006).

Soil thickness was interpolated from 55 well lithologs that contained soil depth information, and dozens of soil pits. With the exception of a few anomalous locations, the soil thickness is rather similar over the valley; the mean for the interpolated soil thickness surface is 0.92 ± 0.21 m and thus the soil thickness was assumed to be 1.0 m in all percolation columns for recharge modelling (Fig. 6a).

The soil map for the aquifer was digitized (as polygons), converted to raster format with 20 m resolution, and then reclassified into soil rating categories based on permeability ratings (1 to 10)

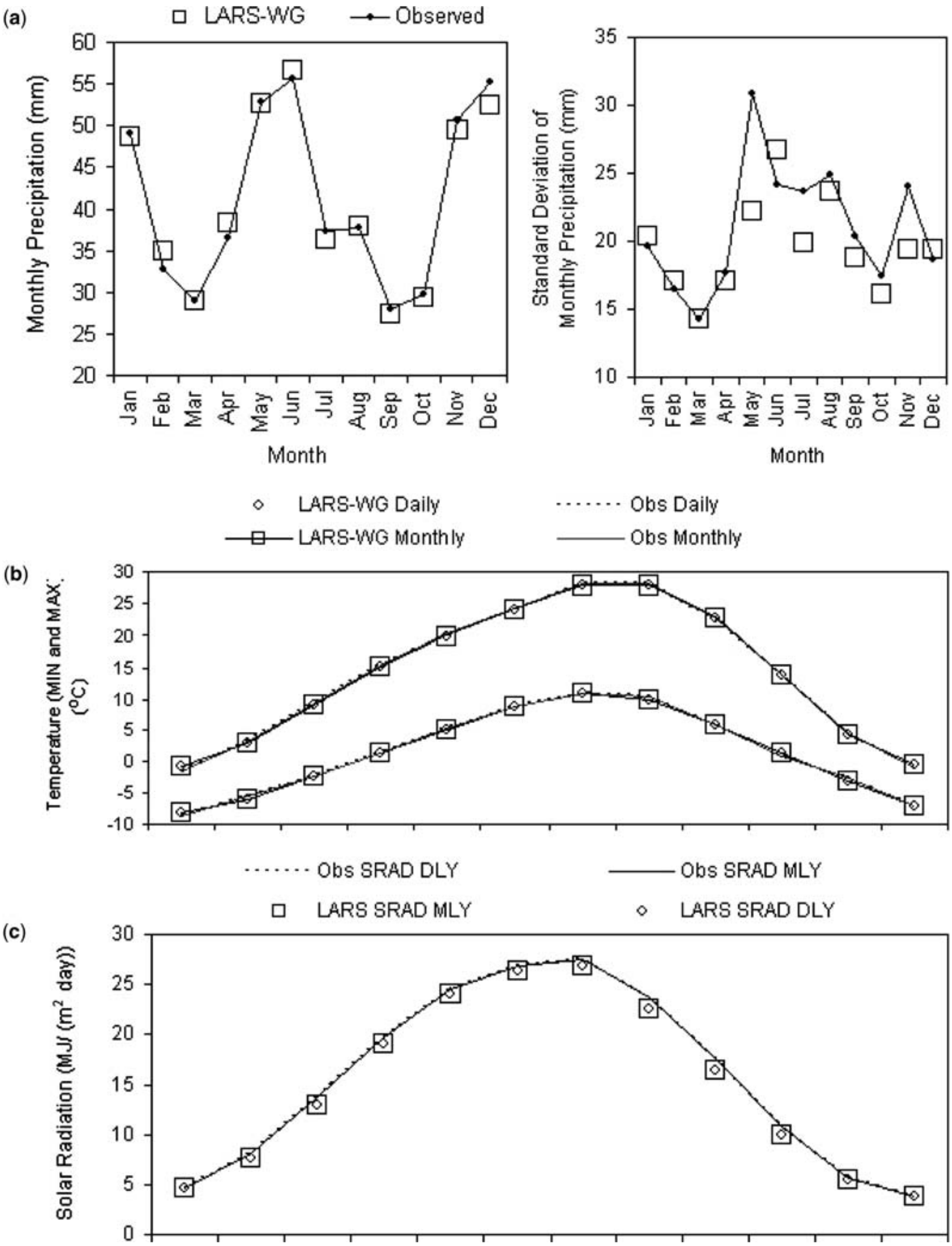

Fig. 5. Weather at Grand Forks, BC, observed for the period of record 1975–1995 (base climate scenario) compared to that modelled with the stochastic LARS-WG weather generator. (**a**) Monthly rainfall; (**b**) monthly maximum and minimum temperature; (**c**) monthly and daily solar radiation.

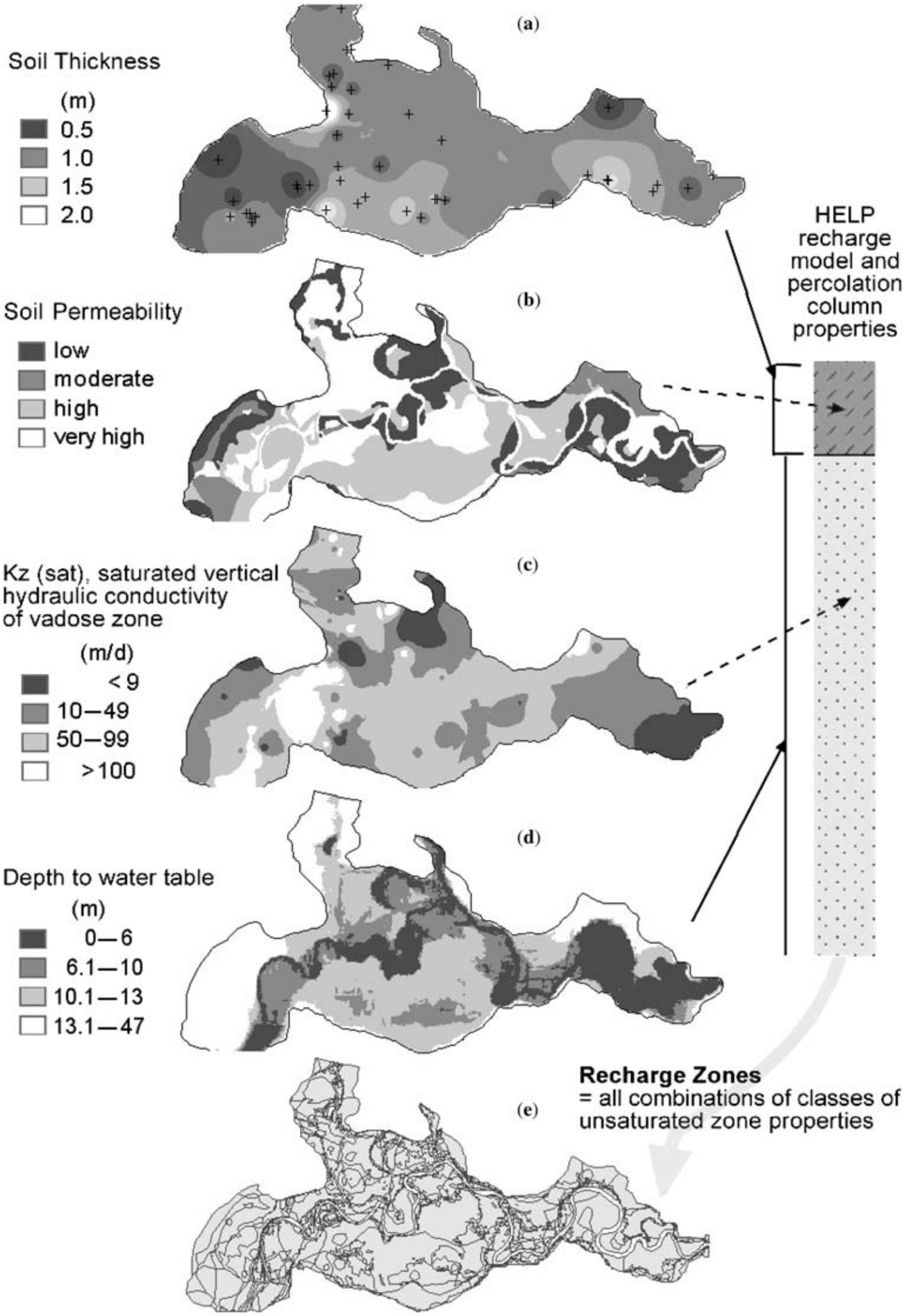

Fig. 6. (**a**) Soil thickness; (**b**) soil permeability classes; (**c**) reclassed *Kz* map of unsaturated zone above water table in Grand Forks aquifer; (**d**) depth to water table classes; (**e**) resulting recharge zones (Scibek & Allen 2006).

(Fig. 6b). Similar soils were combined into one category to reduce the number of categories to four, including very high, high, medium and low permeability, based on the spatial extent of each permeability class, to preserve the most representative soil types over the aquifer extent. Soil permeability maps were modified by land use to account for less permeable areas.

Saturated hydraulic conductivity estimates were estimated for geological units encountered above the water table. Well lithology data were standardized and classified using a custom code in order to simplify the data. Up to three material descriptions were retained for each depth interval. Saturated (assumed vertical) hydraulic conductivity (Kz), specific storage (Ss) and specific yield (Sy) for each material type were assigned based on values in the published literature, and constrained by pumping test results. Geometric means of the Kz values were calculated for each layer in each well where more than one material type was recorded. Equivalent Kz was computed for each well point location, assuming homogeneous and isotropic 'units', given no other data. Kz values in 285 wells ranged from a maximum of 1000 m/day to a minimum of 1×10^{-6} m/day, median of 13 m/day and quartile values of 100 and 0.14 m/day. The Kz values in the vadose zone were interpolated using inverse distance weighed interpolator, and computed on representative vertically averaged log Kz values at all available point locations where lithologs exist. After interpolation, the inverse logarithm (i.e. 10 ∧ [Log Kz]) of the interpolated raster was computed, and converted to units of metres per day. Four classes were chosen as 1×10^{-6} to 9 m/day, 10 to 45 m/day, 50 to 99 m/day, and 100 to 1000 m/day (Fig. 6c). The representative Kz values for each material in the HELP soil columns were 5, 30, 75, and 500 m/day (mid-value in each class).

Depth to water table was estimated via raster computations between ground surface and a numerically derived static groundwater table (Allen *et al.* 2004*a*). This assumption is reasonable as the depth to the water table is usually much larger than the variation in groundwater level, except in the low-lying river floodplain region where river effects dominate the water levels as opposed to recharge, as will be discussed later. Depths to the water table in 285 wells ranged from 1.5 m to 46.8 m, with median of 10.1 m and quartile values of 6.1 m and 12.9 m. The depth classes were based on quartiles of distribution: 0 to 6 m, 6.1 to 10 m, 10.1 to 12.9 m, 13.0 to 47.0 m, with roughly 25% of aquifer area in each of the four categories (Fig. 6d). Representative sediment columns were assigned representative mid-class depths of 3, 8, 11 and 25 m.

Recharge zones were defined for a 50 m raster grid through cross-classification of maps of all important variable distributions (Fig. 6e), resulting in 65 zones (zone 1 was the default zone in MODFLOW with no recharge). The number of combinations of water depth, Kz, and soil permeability was $4 \times 4 \times 4 = 64$ scenarios of soil columns. Soil thickness was assumed the same for all columns. Higher spatial resolution would require more sub-classes in each variable and result in many more recharge zones. In this study, the limiting variable is soil type (originally soil polygons), and the most uncertain is Kz, whereas depth to water table can be represented at 20 m grid or smaller with reasonable accuracy. A recharge zone is any unique combination of soil permeability class, hydraulic conductivity class, and depth to water table class. More combinations were used in sensitivity analyses of HELP model performance. There is a degree of uncertainty in each of these properties because data come from various sources and formats.

The final step involved transferring recharge values into the transient groundwater flow model. Each MODFLOW cell had an independent schedule for recharge. Evapotranspiration was calculated within HELP directly. Precipitation was assumed to be uniform over the aquifer in this valley. Custom codes were written to update MODFLOW files with recharge 'zones' and schedules. Appropriate frequency, magnitude and duration of precipitation and other events are modelled in daily HELP model input/output.

Irrigation return flow estimates

A better representation of recharge takes into account the amount of water that is returned to the aquifer when the land is irrigated. This is commonly referred to as return flow (from the aquifer perspective). In all pumping scenarios, the recharge zones were modified by including estimated irrigation return flow to the aquifer. Generalized estimates of return flow were obtained through consultation with experts in irrigation practices. Irrigation district boundaries were mapped as polygons in GIS, converted to raster coverage, and used in the recharge zone classification. Irrigation and pumping were applied jointly only from June to August (Julian day from 155 to 242). To represent the 'modified' recharge schedules that account for irrigation return flow, a new recharge zone was created out of every combination of original 64 (non-irrigated) recharge zones from HELP model, and areas of irrigated fields with their different return flow estimates. In total 161 recharge zones (unique recharge schedules) were applied to transient Grand Forks recharge model.

Sensitivity of HELP to model inputs

It was important to evaluate the sensitivity of modelled recharge to HELP input parameters. The HELP model results showed a very small effect (<5% change) for type of stand of grass, wilting point, field capacity, and initial moisture content. A moderate effect was found in soil thickness and porosity of percolation layer. As soil thickness increased, the modelled recharge decreased, but only very strongly from April to June as this effect is precipitation- and temperature-dependent. The strongest effect on HELP model recharge results was for depth of vadose zone or percolation layer. Similar effects, but with different magnitudes, were observed for different soil types or permeability of soil, and for saturated vertical hydraulic conductivity of the vadose zone. The effects were seasonal and most pronounced in spring to early summer, again due to a combination of precipitation and temperatures that control evapotranspiration and infiltration rates (together with unsaturated zone properties). The high sensitivity of recharge models such as HELP to unsaturated zone properties suggests that spatial distribution of such properties must be accounted for in recharge modelling for climate change impacts assessment in surficial aquifers.

Historical recharge results

Previous recharge modelling (Allen *et al.* 2004*a*) used a uniform annual recharge value for the Grand Forks aquifer of 135.5 mm/a, or approximately 28% of precipitation. This value was derived using HELP, assuming uniform soil and aquifer media properties and a uniform water table depth, using weather generated stochastically with the WGEN weather generator in HELP. Recharge was remodelled as part of this study, again using HELP, but with LARS-WG, and also to consider a spatial distribution of recharge zones.

Recharge values were modelled for present climate and two future climate scenarios (2010–2039, 2040–2069). All predicted values were expressed as monthly average recharge. These values were summarized and mapped for 64 recharge zones, for each climate scenario. Mean annual recharge varies considerably across the 64 recharge zones, ranging from near 30 to 120 mm/a or between 10% and 30% of mean annual precipitation. The western and the northwestern portions of the aquifer receive the lowest recharge, while the highest recharge is received in the central and eastern portions of the aquifer. The thick depth of gravels in the terraces can absorb a large amount of rainfall before reaching saturation, and because this region is relatively dry, the modelled recharge is low in areas with large depth to water table. Thus, precipitation is insufficient to recharge the aquifer where there are thick sand and gravel terraces around the periphery of the valley – most of the precipitation simply changes moisture content. This situation would be different if this were a wet climatic zone – most recharge would occur in most permeable areas with less influence on depth of sediment to water table.

Monthly recharge varies from <2 to > 12 mm/month, or from 10% to 80% of monthly precipitation. This constitutes a larger variation than in the case of annual precipitation (a more averaged value). Most of the recharge is received in spring and summer seasons, while in winter the ground is frozen and snowmelt does not occur. The autumn season is relatively dry. In spring time, by monthly value, the aquifer receives 40% to 80% recharge from precipitation, depending on soil properties and aquifer media properties; in summer the values are 30% to 50%. During late summer the aquifer receives 60% to 90% of precipitation, but the overall recharge amount is small because rainstorms are infrequent. The LARS-WG preserves the intensities of rain events; we observe that if a high intensity event (such as a thunderstorm) occurs during late summer, it rains heavily and most of the water infiltrates the aquifer. If it were to rain slowly and over a longer time, much more of it would evaporate. This type of relation may be very different in other climate regions and in other aquifers where high intensity rainfall events may lead to increased runoff and less infiltration.

Predicted recharge changes for future climates

Recharge as a percentage of precipitation increases in future climates (Fig. 7). The 2010–2039 climate scenario shows a 2 to 7% increase (spatially variable) from present mean annual recharge (Fig. 7a). Monthly recharge results, shown for selected recharge zones, have the lowest recharge occurring in January to May, the highest recharge occurring in June to September, and October to December receiving moderate recharge (Fig. 8). The 2040–2069 climate scenario shows an 11 to 25% increase (again spatially variable) from present mean annual recharge (Fig. 7b). Monthly recharge results (Fig. 8) have the lowest recharge occurring in January to May, the highest recharge occurring in June to September, and October to December receiving moderate recharge.

Overall, the largest predicted increase due to climate change is in late spring, which suggests an increase of a factor of three or more from present levels. Predictions suggest 50% increase in recharge

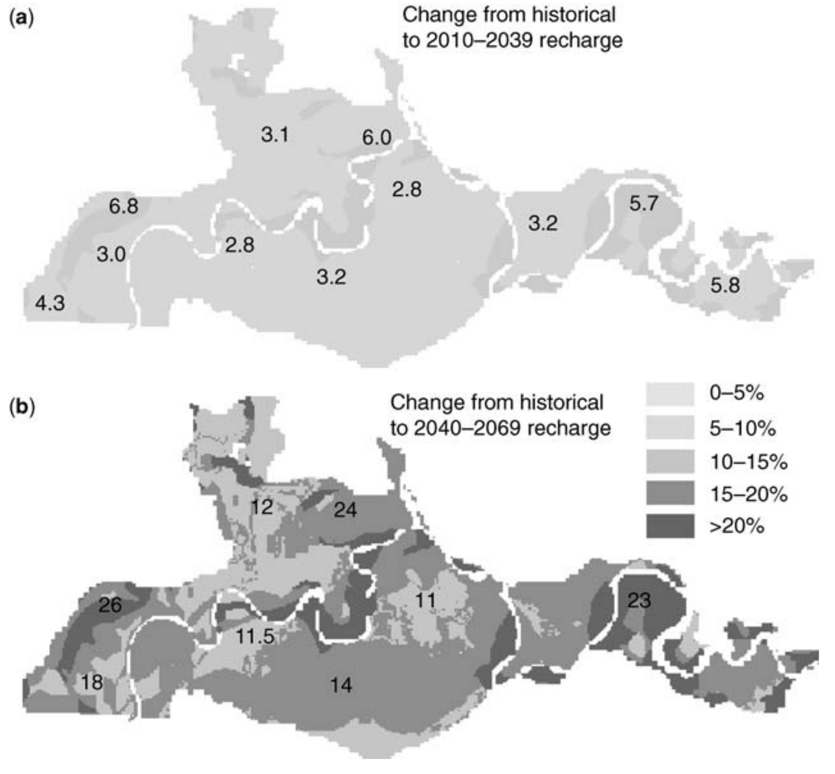

Fig. 7. Percentage change in mean annual recharge to the Grand Forks aquifer modelled in HELP and assigned to recharge zones: between (**a**) 2010–2039 and historical, (**b**) 2040–2069 and historical. Historical climate scenario (1961–1999) (Scibek & Allen 2006).

in summer months, and 10 to 25% increase in autumn. Irrigation return flow contributes 10 to 20% of recharge.

Hydrologic modelling

In order to model the interaction between groundwater and surface water in the valley, stage elevations are required as a function of time for each river node in the groundwater flow model for each climate scenario. The challenges in constructing the model were firstly, balancing the discharge volume in the valley, given that hydrometric stations are located outside the valley and have different periods of record; secondly, modelling basin-scale discharge from downscaled GCM outputs; and thirdly, accurately modelling stage variation in river branches, such that stage could be linked to the groundwater flow model and used to predict impacts on groundwater levels.

Hydrology of Kettle and Granby Rivers

The Kettle River system drains 8300 km^2 within BC and 1500 km^2 in Washington (WA) State, USA. The Grand Forks valley widens near the City of Grand Forks, where the Granby River flows into the Kettle River (Fig. 9). The Granby River has a drainage area of 2050 km^2 at its confluence with the Kettle River. In the Kettle River drainage area, the snow pack increases over the winter until early April, and melts between April and the end of June, with the end date of the snowmelt season varying from mid-May to mid-July. The hydrological response is extremely sensitive to seasonal patterns. During years with unusually warm winters, the system shifts from a snowmelt-dominated regime to a bimodal regime, where there is an increasing number of days of high flow due to rain, but a decreasing number of days of high flow due to snowmelt.

Whitfield & Cannon (2000) analysed data from hydrometric stations in southern BC over two decades (1976–1985 and 1986–1995). The study determined that these streams are currently snowmelt-dominated. Observed changes in Kettle River discharge over the last decade suggest a shift in peak flow to an earlier date, although the peak flow magnitude remains the same. Similar responses were observed in other streams in the

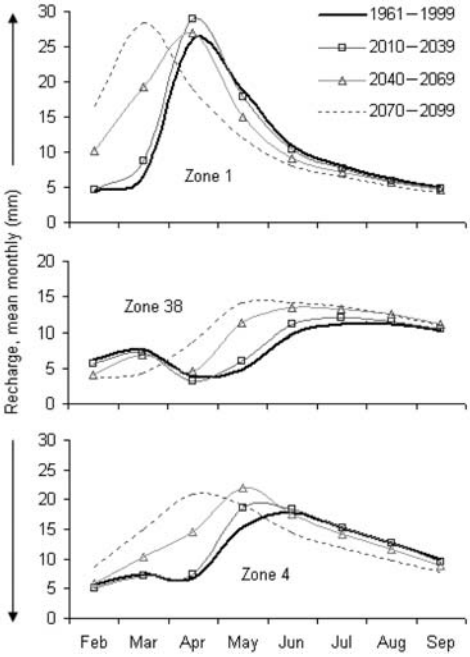

Fig. 8. Changes in recharge due to climate change in three different recharge zones (1, 38 and 4 located in the middle of the Grand Forks valley, and differing by depth of vadose zone and soil permeability). Results are for monthly mean recharge without irrigation return flow.

province. The low-flow period now begins earlier in the summer and baseflow levels are lower. In addition, flow is higher in the early autumn due to higher rainfall. Streams in this region are expected to become increasingly bimodal as a result of predicted winter warming associated with climate change.

Daily discharge records were supplied by Environment Canada. As most river gauges record only water elevation, the discharge records are calculated from stage–discharge rating curves. Representative annual hydrographs, averaged for the period of record, were plotted for each hydrometric station. The available hydrometric stations (shown in Fig. 9) in the valley have non-overlapping periods of record; the longest records are at the Ferry (WA) and Laurier (WA) gauges on the Kettle River. Therefore, it is necessary to scale these discharge records to represent flow at points between these two gauges in the Grand Forks valley. To determine the runoff at a location downstream of a gauge, the observed daily flows at the upstream station were adjusted by the drainage area ratio of downstream/upstream stations, following the methodology of Leith & Whitfield (2000). Thus, the streamflow records at Laurier were scaled to represent the streamflow hydrographs in the Grand Forks valley downstream of the confluence of the Kettle and Granby Rivers. The upper section of Kettle River in the valley

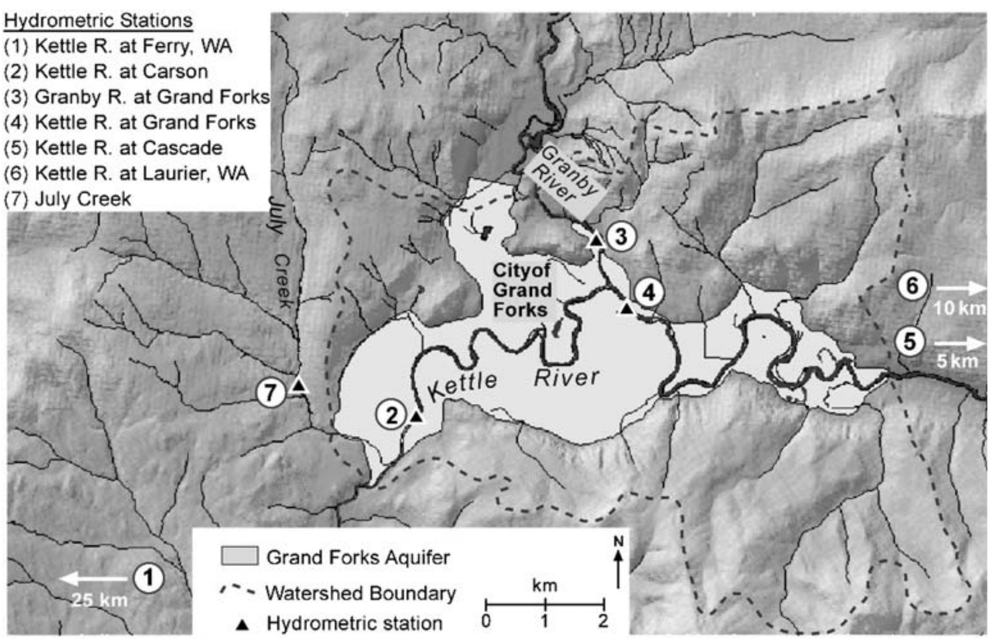

Fig. 9. Grand Forks valley catchment area and hydrometric stations near the Grand Forks aquifer (modified from Scibek *et al.* 2007).

was then modelled using scaled discharge values from the gauge at Ferry.

The mean annual discharge of the Granby River is 30.5 m^3/s, and for the Kettle River, upstream of Grand Forks, it is 44.3 m^3/s. Past the confluence, mean annual discharge is 72.8 m^3/s. Therefore, at the confluence of these rivers, the Granby contributes approximately 40% of the flow, and the Kettle contributes 60% of the flow to the Kettle River. In most years, at low flow in August, the Kettle River maintains a discharge of between 10 and 14 m^3/s, compared to a minimum discharge of 0.0137 m^3/s for the creeks in Grand Forks catchment. Thus, during the low-flow or high-flow conditions, the small tributaries contribute only 0.64 to 0.91 m^3/s mean annual discharge to the larger Kettle River, within the extent of the Grand Forks aquifer, or approximately 1% of the combined Kettle and Granby River discharge.

River boundary conditions

Based on previous steady-state groundwater flow modelling (Allen *et al.* 2004*a*), discharge to the Kettle and Granby Rivers is not measurably affected by baseflow from the aquifer. Thus, the combined aquifer and tributary contributions to the rivers have a very small effect on Kettle and Granby River water levels. In contrast, the river water levels have a strong effect on groundwater levels in the aquifer. Therefore, the rivers can be represented as specified head boundaries, such that the head schedules will represent the modelled river stage in transient Grand Forks aquifer model.

The bottom sediments of the Kettle and Granby Rivers above the Grand Forks aquifer consist of mostly gravels, with very little fine sediments. In effect, the aquifer is in direct contact with the river channel and there is no impediment to flow. The constant head nodes do not have any conductance coefficients, and thus assume perfect hydraulic connection between the river and the aquifer. The river can leak and receive water to and from the aquifer, but the river stage will not change as a result of such interaction. In other words, the river will act as an inexhaustible supply of water and will influence the aquifer water levels, but the aquifer will not have any effect on river discharge and stage. The head is held at a constant value for the duration of a time step, but changes to a different value with successive times.

Simulating river flow and input to the groundwater model

The BRANCH model (Schaffranek *et al.* 1981) is a one-dimensional model that is broadly applicable, and is intended for operational use to compute unsteady flow and water-surface elevation (stage) of either singular or interconnected channels. The time-dependent variables are the flow rate and the water-surface elevation. Water-surface elevations and flow discharges are computed at segment nodes and branch junctions. Limitations of the model include no account for channel storage, variations of channel roughness with stage, and backing-up of water along unsurveyed sections of the channel, which could impact the surveyed locations.

A new user interface was developed for the BRANCH code, where all inputs and outputs are included in a single spreadsheet file (Microsoft Excel). A new module was written to allow for hydrograph generation and to create boundary value data series in any time increments to simulate the hydrograph wave form based on monthly values. Finally, software was developed, which allows mapping of the channel network into a raster grid as defined by the MODFLOW grid, divides the channel into segments, and uses BRANCH output to update the MODFLOW boundary value file for specified-head boundary schedules for any number of cells. The new version of BRANCH was verified successfully with USGS sample data.

The model was applied to sections of the Kettle and Granby Rivers in the Grand Forks valley. Boundary conditions were specified at three external nodes and river stage was computed at 67 channel cross-sections (British Columbia Ministry of Environment 1992). Stage and discharge (rating curves) were calculated for all river cross-sections at 1-minute time intervals over the specified number of 10 000 time steps. River channels were represented in three dimensions using a high grid density (14 to 25 m) in MODFLOW. River segments were mapped onto MODFLOW cells in a GIS system (to mid-points of cells), providing a database link between river water levels and appropriate river boundary cells. For each segment, the program located the nearest upstream and downstream cross-section location, and the stage–discharge rating curve for that cross-section was used to calculate water elevation from discharge. River water elevation was interpolated between cross-sections with the fitted channel profile. River stage schedules along the 26-km-long meandering channel were imported at varying, temporal resolution (one to five days) for every cell location independently. The channel width of Kettle River was two to four cell lengths at most locations. The actual thalweg, or water-filled and flowing channel width, may be less than two cell widths during low-flow months, but this schematization does not adversely affect the groundwater flow model.

Two problems were encountered during groundwater flow modelling, which were related to the poorly resolved topography. First, the channel bottom elevation profile, representing the minimum elevation at each cross-section along the length of the river, appeared to have a jagged appearance because there are local depressions in the river channel, or perhaps surveying errors. Consequently, the river channel bottom profile was smoothed out to ensure that calculated minimum and maximum stage were always decreasing downstream. Secondly, river stage should also be below local floodplain elevation (since extreme floods are not modelled). Floodplain elevations were read from the most accurate source available – floodplain maps for Kettle River. The digital elevation model (DEM) (20 m grid) was rather inaccurate in the valley. River floodplain elevations were too low in many places and the river channels were poorly defined. Thus, MODFLOW layers were edited along all river channels to put all constant-head boundary cells in first layer (gravel) of the model. The channels were also deeper than on the original DEM surface of the valley, but were similar to the surveyed channel profiles.

Downscaling of climate data and calculation of river discharge

Models for streamflow generation from catchment areas can be calibrated to present conditions, and extrapolated to predict future conditions. These include physically based catchment models, empirical or statistical models, relating hydroclimatic variables to streamflow, and empirical downscaling models, where local or regional-scale variables (e.g. streamflow), which are poorly described by coarse-resolution GCMs, are related to synoptic- or global-scale atmospheric fields (Landman *et al.* 2001). A review of applications of downscaling from GCM to hydrologic modelling can be found in Xu (1999).

The dimension of the large-scale climate dataset (CGCM1) was reduced using principal component analysis (PCA) by Environment Canada. A *k*-nearest-neighbour analogue model was used to link principal component scores (explained variance $>90\%$) of the climate fields with the maximum temperature, minimum temperature, and precipitation series (of NCEP dataset). The PCA linked the climate fields over BC and the eastern Pacific Ocean with daily discharge values for Kettle and Granby Rivers. The end product was sets of daily discharge data at the three sites for the simulated 1962–2100 period: Granby River at Grand Forks (BC), Kettle River at Ferry (WA) and Laurier (WA).

The less than ideal fit between the downscaled and observed hydrograph for 1971–2000 (Fig. 10) can mostly be attributed to biases existing between the GCM simulated climate fields and the observed climate fields from the NCEP–NCAR reanalysis. The downscaled CGCM1 data underestimate temperature in the late winter and early spring periods and overestimate temperature in the late autumn and early winter periods. Consequently, the onset of freshet is delayed. The model bias is similar for all three hydrometric stations, but the model bias is greater for median discharges than for mean discharges. Therefore, only mean hydrographs were considered in future analyses.

Where the model bias is unacceptable, the downscaled results could be used as a basis for adjusting the observed historical hydrograph to match the simulated changes. However, such an approach might be hard to justify, especially for future scenarios, and the GCM bias should be explicitly shown, along with the resulting impact on the subsequent hydrologic simulations. The comparisons of impacts of future climates is then always between the unadjusted GCM-driven hydrologic simulations for future time periods, and the unadjusted GCM-driven simulations for the baseline period.

In the future climate scenarios (Fig. 11) the hydrograph peak is shifted to an earlier date, although the peak flow remains the same. Changes to the river hydrograph are predicted to be much larger for the 2040–2069 scenario than the 2010–2039 scenario, compared to the historical period. The two locations on the Kettle River and one on Granby River had very similar responses to climate change.

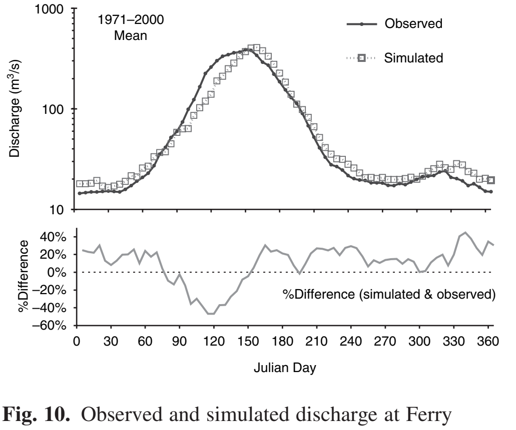

Fig. 10. Observed and simulated discharge at Ferry (WA) on Kettle River, downscaled from CGCM1 (observed data from Environment Canada, 2002) showing model bias (Scibek *et al.* 2007).

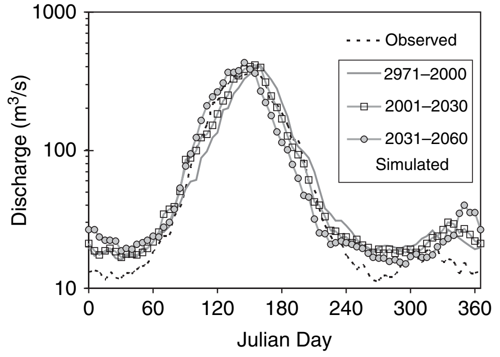

Fig. 11. Predicted discharge for future climates in Kettle River at Laurier, WA (Scibek *et al.* 2007).

Model calibration

Model calibration involved comparing the simulated water levels to observed historic water levels for both steady-state and transient model runs. Figure 12 shows the calibration graph for observation well 217, which is the only provincial observation well in the valley. The graph displays the observed long-term monthly mean water elevation and modelled groundwater elevation after model calibration (1961–1999). Also shown are observed and simulated discharge hydrographs for the nearby Kettle River for the corresponding time period. There is a regular seasonal pattern, similar to the stage hydrograph of the Kettle River. The groundwater level in the well varied between 1 and 1.8 m over the period of record, whereas the river experienced stage fluctuation of 3 m. The mean monthly water table elevation varied only by about 1 m, with a standard deviation of 0.2 m. The shape of the well hydrograph is similar to the Kettle River hydrograph, but the amplitudes of seasonal fluctuations are damped, which would be expected with increasing distance away from the river channel.

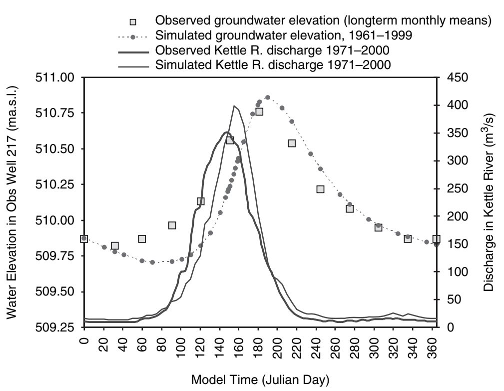

Fig. 12. Mean hydrograph of water table elevation (total head) in observation well 217 in Grand Forks aquifer and water surface elevation of Kettle River approximately 400 m from well 217 (Scibek *et al.* 2007).

The groundwater model is reasonably well calibrated for transient flow at the location of observation well 217. Calibration residuals for static (steady-state) water levels across the entire aquifer had an acceptable error distribution, but residuals tended to be high near the model boundaries, which might be anticipated due to lack of physical data in these areas with which to constrain the conceptual model (Fig. 13a). Observation wells where residuals were very large (>5 m) were examined in detail, and compared to nearby observation wells, the possible range of river water levels if the well was adjacent to the river, ground surface elevation, and the expected water table surface in that area. The root mean square (RMS) error for model was 8% (Fig. 13b).

Responses to climate change

Water budget results

In order to conduct a water balance assessment, the model domain was divided into several water budget zones (Fig. 14), and Zone Budget (ZBUD) was run. ZBUD (McDonald & Harbaugh 1988) calculates sub-regional water budgets using results from MODFLOW simulations. In the top two aquifer layers, ZBUD zones were delineated for all cells. Zone 1 includes the City of Grand Forks (GF) and other areas considered as background to the main irrigation districts. In the five irrigation districts (zones 3 to 7), polygons of irrigated areas (fields) were also used for ZBUD zone delineation, taking into account areas that are irrigated in the large districts. Flow budget zones in these districts only apply to actually irrigated areas, and not whole districts. The river floodplain was given a separate zone (2). It is meant to account for river–aquifer exchanges in the low-lying areas that have head values very similar to river elevations and react very quickly to changes in river water levels. The underlying silt (7) and clay beneath the silt (8, not shown) were similarly assigned as zones.

During spring freshet on the Kettle River, the rise in river stage causes inflow of water to various ZBUD zones (after passing through the floodplain area). This excess water is stored in the aquifer. Mass balance calculations indicate that storage rates are less than 50% of inter-zonal groundwater

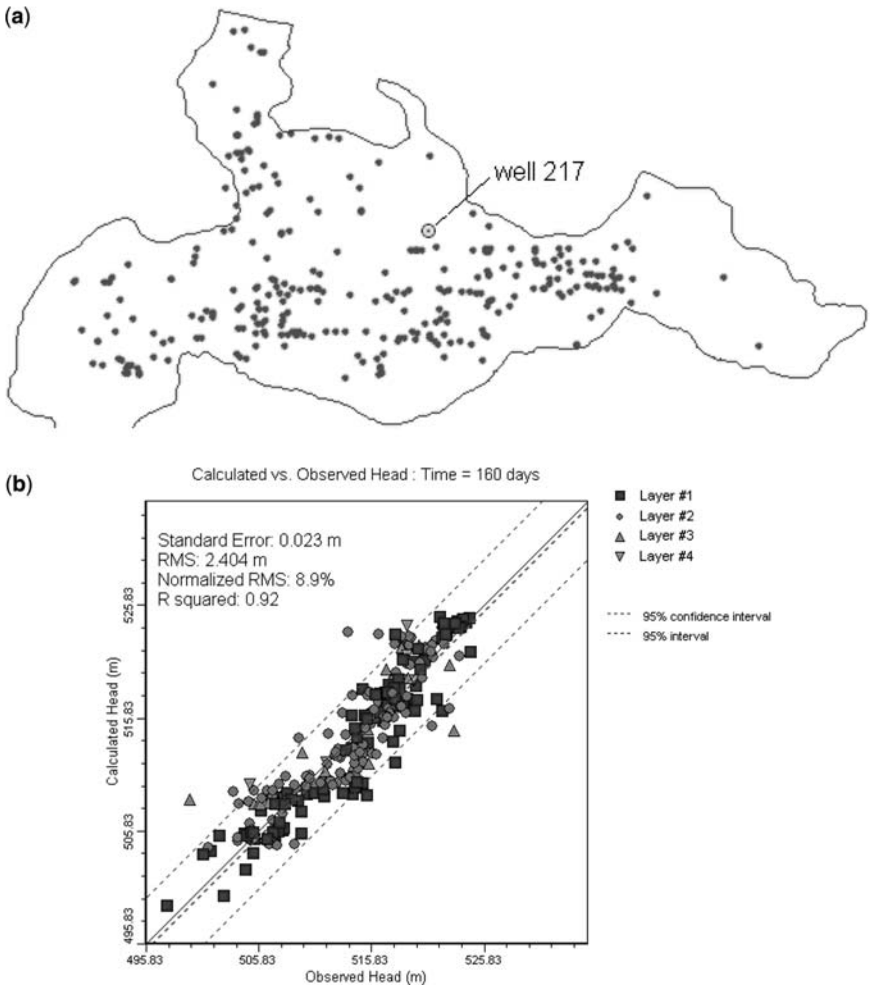

Fig. 13. (**a**) Wells with static groundwater levels in BC well database, and location of observation well 217 with monthly water records. (**b**) Residuals at model time 160 (Julian day) from transient model run for 1961–1999 climate.

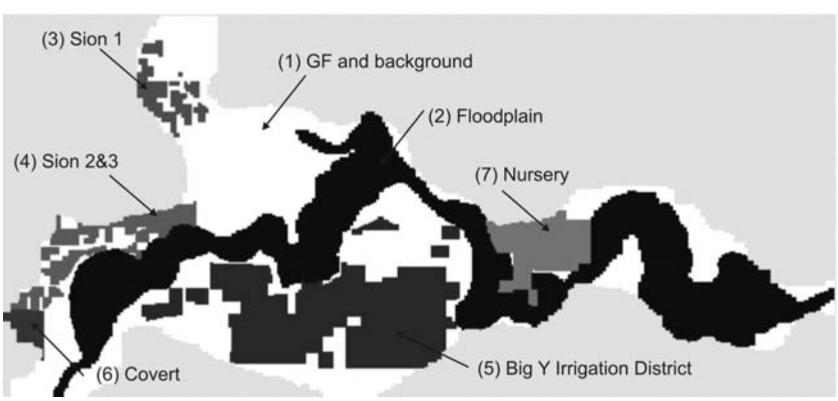

Fig. 14. Water budget zones (numbers in brackets are the zone numbers referred to in the text) in MODFLOW model of Grand Forks aquifer (top layer shown).

flux, and 15 to 20% of river–aquifer flux. As river stage drops, the hydraulic gradient is reversed and water is released from storage and leaves mostly to the floodplain zone as it returns to the river as baseflow. The rate of inflow of groundwater from the river and into the aquifer along the floodplain zone follows the river hydrograph very closely during the rise in river stage. As the river stage levels off and begins to decrease, the flow direction is reversed within 10 days, and the rate of inflow from the aquifer to the river begins to rise, and then dominates for the rest of the year, as water previously stored in the aquifer drains back to the river as baseflow seepage. As most of the pumping water is lost to evapotranspiration on irrigated fields, there is a small reduction in the baseflow component to the Kettle River during the pumping period.

The river–aquifer interaction has a maximum flow rate of 41 m^3/s, which translates to between 11 and 20% of river flow during spring freshet. Thus, the river puts about 15% of its spring freshet flow into storage in Grand Forks valley aquifer, and within 30 to 60 days most of that water is released back to the river as baseflow. During the freshet conditions, 15% of river flow (in m^3/s) occurs through the surficial aquifer, mostly in the floodplain area. As the river peak flow shifts to an earlier date in a year for the future climate scenarios, the 'hydrographs' for aquifer water levels also shift by the same interval. Impacts are smallest in zones least connected to the river (away from the river and at higher elevation).

Volumetric recharge accounts for 1 to 7% of other flow components, such as flow between zones and storage. In most zones, recharge increases from winter to summer, then remains high until late autumn and decreases towards the winter months. Irrigation return flow begins after day 150. In zones 4, 5 (shown) and 7, recharge increases by 10 to 20% due to irrigation return flow. Return flow is constant for any climate scenario because it was calculated from present irrigation use of groundwater in each district (return flow component does not increase with climate change even if recharge changes). Figure 15 shows recharge for zone 5, the Big Y Irrigation District, which is the largest irrigation district in the valley. Irrigation return flow can be observed by differences between pumping and non-pumping recharge for any one climate scenario.

The predicted future climate for the Grand Forks area from the downscaled CGCM1 model will result in more recharge to the unconfined aquifer from spring to summer seasons. The largest predicted increase is from day 100 to day 150, when it is predicted to increase by a factor of three or more from present levels, according to ZBUD results from all zones. In the summer months, recharge is also

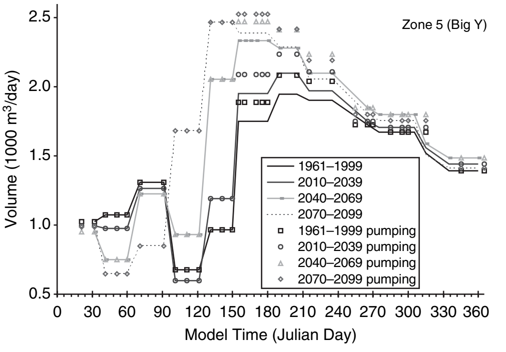

Fig. 15. Transient model recharge flow volumes for Zone 5 (Big Y Irrigation District as shown in Fig. 14).

predicted to be approximately 50% greater than at present (in most zones). In the autumn season the recharge is predicted to increase (10 to 25%) or remain the same as present, depending on location in the valley. In the winter the CGCM1 weather predictions suggest less precipitation in winter and, thus, less recharge to the aquifer.

Within an annual cycle and between climate scenarios the results show different spatial and temporal distributions in groundwater conditions. Head difference maps (Fig. 16) were prepared to show differences due to climate change between future climate scenario model outputs and present climate scenario model outputs. At present, the flow patterns are influenced by river channel profile, and generally follow valley floor topography. In this particular aquifer, the effect of changing recharge on groundwater levels is very small compared to changes in timing of basin-scale snowmelt events in the Kettle River and the subsequent shift in hydrograph.

In the 2010–2039 scenario, water levels rise and fall with the river hydrograph at different times because of a shift in river hydrograph peak flow to an earlier date. Elevated water levels up to 30 to 40 cm persist along the channel and drain within a month. From late summer to the end of the year, water levels are similar to present conditions, with small increases observed due to the increase in recharge in areas away from the river channel. Overall, the climate change effects for the 2010–2039 scenario relative to present are limited to the floodplain, and to the early part of the year when the river hydrograph shifts and is at peak flow levels. A small increase of water levels due to the increase in recharge is forecast for the 2010 climate.

In the 2040–2069 climate scenario, the hydrograph shift is larger than in the 2010–2039 climate scenario, resulting in up to 50 cm change in groundwater levels. Parts of the valley aquifer that are strongly connected to the river have the largest climate-driven changes: the 'hydrographs'

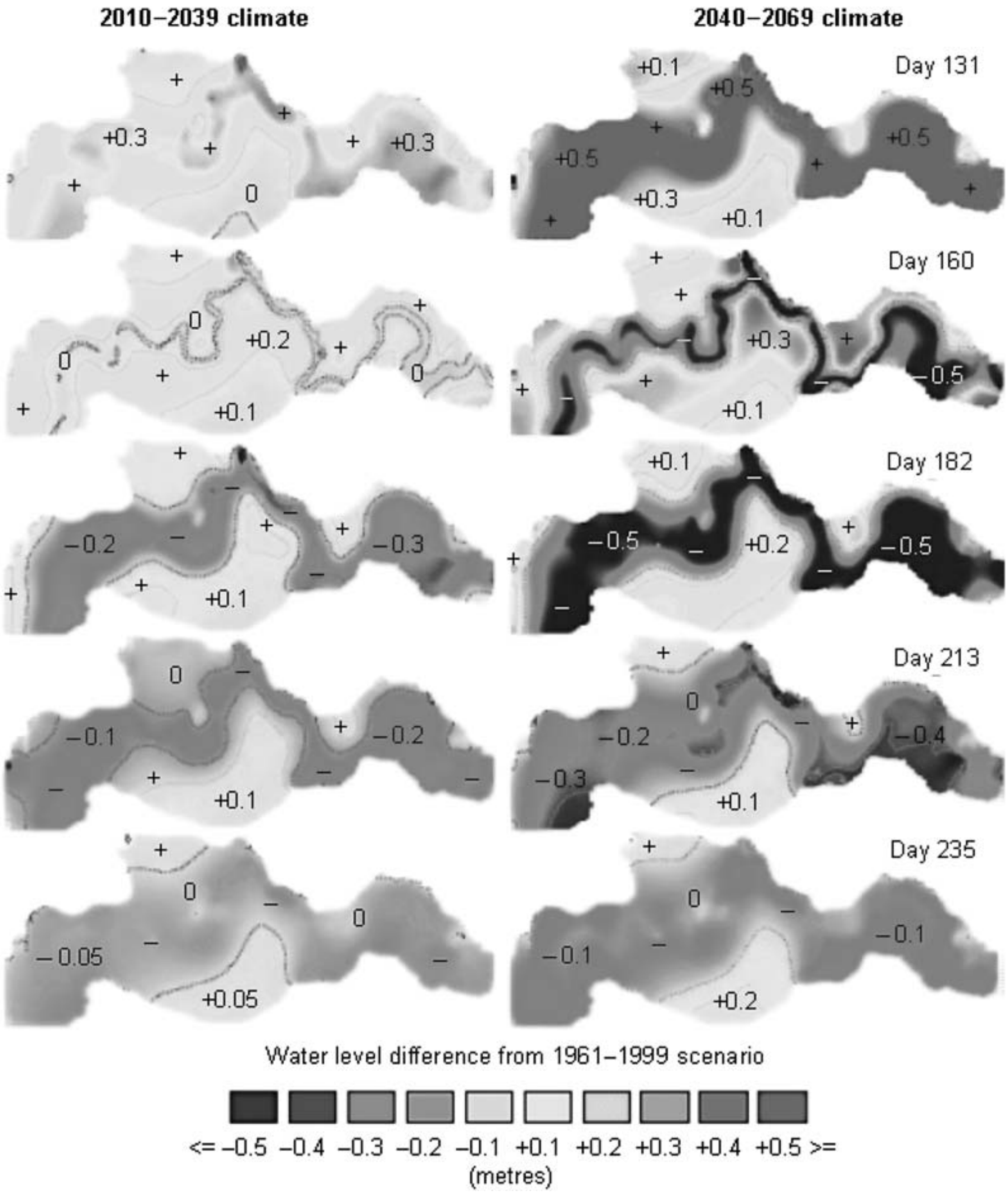

Fig. 16. Water level differences between future (2010–2039) and present climate, and future (2040–2069) and present climate under pumping conditions. Maps by time step in days 131 to 305. Positive contours are shown at 0.1 m interval. The zero contour is a dashed line. Negative contours are not shown. Darkest colours indicate values <-0.5 m (along rivers only). At day 101 (not shown), difference map has values within 0.1 m of zero (Scibek & Allen 2006).

for groundwater levels also shift similarly to the river hydrograph between climate scenarios.

Conclusions

The predicted future climate for the Grand Forks area from the downscaled CGCM1 model will result in more recharge to the unconfined aquifer from spring to the summer season. The largest predicted increase is from day 100 to day 150, when it is predicted to increase by a factor of three or more from present levels. In the summer months, recharge is predicted to be approximately 50% greater than at present (in most zones); in the autumn season the recharge is predicted to increase (10 to 25%) or remain the same as present, and in the winter recharge is expected to decrease.

Irrigation return flow begins after day 150 causing an increase of aquifer recharge by 10 to 20% in most irrigation district zones.

The groundwater flow model is sensitive to recharge only away from the river floodplain, and the maximum change expected (within the range of recharge values between the 65 recharge zones) in water table elevation is between 10 and 50 cm, but typically about 20 cm. Areas of the aquifer where temporal variation in recharge does not significantly affect model output are along river floodplains. There, water levels are almost entirely controlled by river water levels.

The predicted temporal shifts in river hydrographs cause changes in aquifer water levels compared to the present, if compared to the same day of year. Differences are less than 0.5 m away from floodplain, but can be over 0.5 m near the river. However, the overall hydrograph shape remains the same. As the river peak flow shifts to an earlier date in a year, the 'hydrographs' for groundwater levels also shift by the same interval. Storage rates are less than 50% of inter-zonal groundwater flux, and 15 to 20% of river–aquifer flux. The river–aquifer interaction has a maximum flow rate between 11 and 20% of river flow during spring freshet – the river puts about 15% of its spring freshet flow into storage – and within 30 to 60 days, most of that water is released back to the river as baseflow.

The hydrograph shift for the 2040–2069 climate is larger than for the 2010–2039 climate scenario, therefore the computed differences to historical climate are similarly larger. The maximum water levels associated with the peak hydrograph are very similar to present climate because the peak discharge is not predicted to change, only the timing of the peak.

Financial support for this research was provided by the Natural Sciences and Engineering Council of Canada (NSERC) in the form of an undergraduate research award for J. Scibek; Natural Resources Canada under the Climate Change Action Fund (CCAF) Program; and the BC Ministry of Water, Land and Air Protection (BC WLAP). Technical support has been provided by Environment Canada (climate change predictions for the Kettle River), and BC WLAP (water well database).

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Origin of a fresh groundwater body in Cholistan, Thar Desert, Pakistan

M. A. GEYH[1] & D. PLOETHNER[2]

[1]*Rübeland 12, 29 308 Winsen/Aller, Germany (e-mail: mebus.geyh@t-online.de)*

[2]*Bundesanstalt für Geowissenschaften und Rohstoffe (BGR), Hannover, Germany*

Abstract: The Thar Desert of Pakistan stretches along the border to India and is one of the most densely populated deserts in the world. Brackish to saline groundwater prevails. A locally restricted fresh groundwater resource was discovered by a comprehensive hydrogeological, geophysical, and isotope hydrological survey conducted from 1986 to 1991. The origin, recharge mechanism and age of the fresh groundwater resource were assessed. There is only fossil groundwater and this must be mined. Sodium is the predominant cation. Present groundwater recharge is absent or extremely low as the annual precipitation rate and the potential evapotranspiration rate amount to less than 200 mm/a and about 2700 mm/a, respectively. The investigations comprised a hydrogeological well inventory, electrical resistivity transects on the ground and an air-borne electromagnetic survey, followed by a test-hole drilling programme combined with geophysical borehole logging, aquifer testing, and groundwater sampling for both chemical and environmental isotope analyses. The results of this study delivered a hydrogeological concept on the origin and recharge of the fresh groundwater body. We found that the fresh groundwater was indirectly recharged during flash floods in low lands during the last pluvial period rather than directly replenished in the high mountain areas far in the east.

Cholistan, in the northern part of the Thar Desert in Pakistan, is a vast sandy area of about 26 000 km^2. A fresh groundwater resource was discovered during our study. It is located within the area 29°0′ to 29°15′ N and 72°0′ to 73°0′ E, between Fort Abbas in the east and Fort Mojgarh in the SW (some 60 km SE of Bahawalpur). It stretches along the southern bank of the Old Hakra River traced by a chain of forts: Phulra (Abbas), Mirgarh, Jamgarh, Marot, Mojgarh, Dingarh and Derawar, remnants of the Moghul Period (AD 1526 to 1857).

In the north, the irrigated land of the Hakra Right canal system borders the project area. The canal water comes from the Sutlej River, about 90 km north of Fort Abbas. The southwestern part of the study area (some 450 km^2) was under irrigation between 1927 and 1932. The land surface slopes from 140 m a.s.l. in the east to 125 m a.s.l. in the SW. The target area represents a sub-recent stream bed and meander-belt floodplain built up by the Old Hakra River (Geological Survey of Pakistan 1964; Fig. 1). Between 4000 and 3500 years BP the discharge of the Old Hakra River gradually decreased due to the southward shift of its headwaters. The river stopped running perennially around 2500 years BP while flash floods reached the target area until mid-thirteenth century. The river eventually ceased running in the early sixteenth century (Wilhelmy 1969). This former fluvial environment exhibits a flat, hard pan surface on which dunes irregularly rest up to 10 m high.

The top sediments of the floodplain consist of silty to sandy loam or silty clays. They are of low permeability. Thus, rainwater accumulates in natural depressions and may remain for a couple of days, or is stored in artificial ponds for stock watering over a period of up to four months. The pans have no vegetation, while shrub, grass or single small trees grow on the dunes. The vegetation is typically xerophytic.

The study area receives a mean annual rainfall of less than 200 mm. Two-thirds of the precipitation, in the form of high intensity downpours, takes place during the monsoon period lasting from July to September. The rainfall is, however, so erratic that continuous droughts are often experienced for two to five years at a time. The potential evaporation rate amounts to about 2700 mm/a.

The mean annual maximum and minimum air temperatures average 34°C and 18°C, respectively. The daily maximum temperature rises above 40°C during the hottest months of May and June.

Hydrochemical situation

A detailed overview on the spatial distribution of the fresh, brackish and saline groundwaters in the study area was obtained by an electrical resistivity and an air-borne electromagnetic survey. The 18-Ωm line of the apparent resistivity of the

From: DRAGONI, W. & SUKHIJA, B. S. (eds) *Climate Change and Groundwater*. Geological Society, London, Special Publications, **288**, 99–109.
DOI: 10.1144/SP288.8 0305-8719/08/$15.00

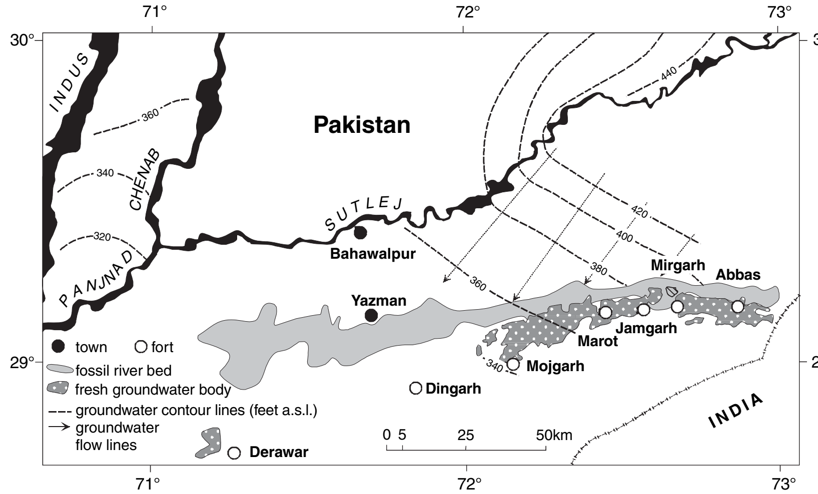

Fig. 1. Map of northern Cholistan. The former bed of the Old Hakra River stretched from Fort Abbas to the west of Yazman (light grey). The fresh groundwater body has been shifted to the south due to the southwesterly flow direction of the groundwater (dark grey with white dots).

water-saturated aquifer, related to 50 m depth below surface, traces a fresh groundwater body. It comprises a 14-km-wide and 98-km-long strip and is directed from east to SW a few kilometres south of the bed of the Old Hakra River (Fig. 1). This apparent resistivity corresponds to an electrical conductivity of groundwater alone of less than 1500 μS/cm corresponding to 900–1200 mg/l total dissolved solids (TDS). The average thickness and volume of the fresh groundwater aquifer is about 100 m and 10 km^3, respectively. This resource is almost completely embedded in brackish and saline groundwater (Fig. 2). The electrical conductivity (EC) of the latter ranges from 3400 μS/cm at the top to 29 000 μS/cm at a depth of 100 m, and 52 000 μS/cm in the east.

Sodium is the predominant cation. At EC $<$ 5000 μS/cm bicarbonate is the dominant anion, while at EC $>$ 5000 μS/cm chloride prevails. The fresh groundwater is a cation-exchange water of the $Na\text{-}HCO_3$ or $Na\text{-}HCO_3\text{-}Cl$ water type. It changes to the $Na\text{-}Cl\text{-}HCO_3$ and $Na\text{-}Cl\text{-}SO_4$ type and finally to the Na-Cl water type with increasing depth and salinity.

The physical and chemical parameters of the fresh groundwater vary in the following ranges: pH = 7.0 to 8.8; T = 24.5 to 31.5°C; EC = 700 to 1200 μS/cm; TDS = 510 to 970 mg/l; sodium = 120 to 270 mg/l; calcium = 2 to 30 mg/l; magnesium = 2 to 40 mg/l; chloride = 30 to 220 mg/l; sulphate = 10 to 150 mg/l; bicarbonate = 200 to 500 mg/l. This fresh groundwater meets the chemical standards laid down for drinking water and stock watering. However, its suitability for irrigation purposes is limited due to the high proportion of sodium of up to 90 meq%.

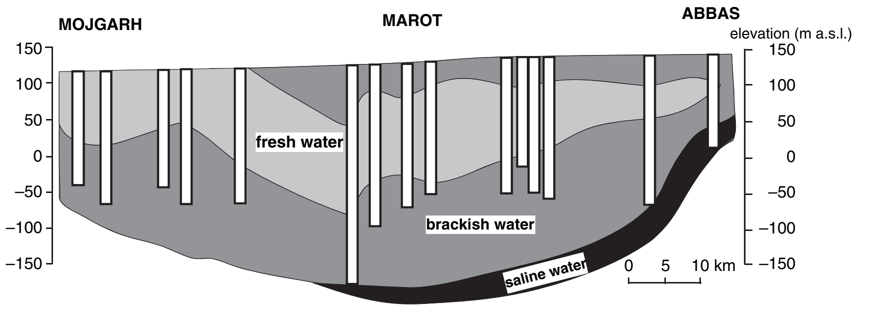

Fig. 2. Hydrogeological longitudinal section of the Old Hakra aquifer and the spatial distribution of the groundwater salinity. The white columns represent wells tapping the groundwater bodies of differing salinity. The salinity boundaries were spatially derived from the air-borne electromagnetic survey and calibrated by the hydrochemical analyses of single samples.

The hydrochemical depth section shown in Fig. 2 may reflect past changes in precipitation: the higher the salinity of the groundwater the drier the climate. Conjunctive use through blending groundwater with surface water from the Sutlej River can reduce both the salt and the mining impacts. But these measures can only minimize the increasing problems with the fresh water supply, mainly used for irrigation, resulting in an acceleration of desertification (Scholz 1997). However, on the other hand the sustainable management of deep groundwater resources may accelerate the socio-economic development of the Thar Desert and may even lead to de-desertification gradually with time (Zaigham 1999). This optimistic view becomes weakened dramatically as isotope analyses show that the studied groundwater resources are fossil and the gradients of the piezometric surface are relicts of former pluvial conditions.

Hydrogeological situation

The study area was explored up to maximum depth of 500 m by 32 test holes, of which 16 tapping fresh groundwater were converted into tube wells. The unconsolidated sediments are mainly of fluvial origin and consist of slightly calcareous and micaceous, fine-grained sand covered and interbedded by sandy–silty clays (Fig. 3). With increasing depth the sands become fine- to medium-grained, locally even coarse-grained. Those sediments form a typical semi-confined aquifer system down to a depth of about 100 m. It has a storativity ranging from 3×10^{-4} to 7×10^{-3}. The horizontal hydraulic conductivity K increases from 5 m/day in the east to 35 m/day in the west. The total porosity of 40% was determined by grain size analysis, while the effective porosity was estimated to be 15%. The groundwater table is about 20 m below surface. The hydraulic gradient between east and SW is about 0.3‰ (Ashraf & Ismail 1994).

Recharge rate of the fresh and saline groundwater

The initial hydrogeological working hypothesis was that the fresh groundwater is still being replenished. However, at three different places in the Thar

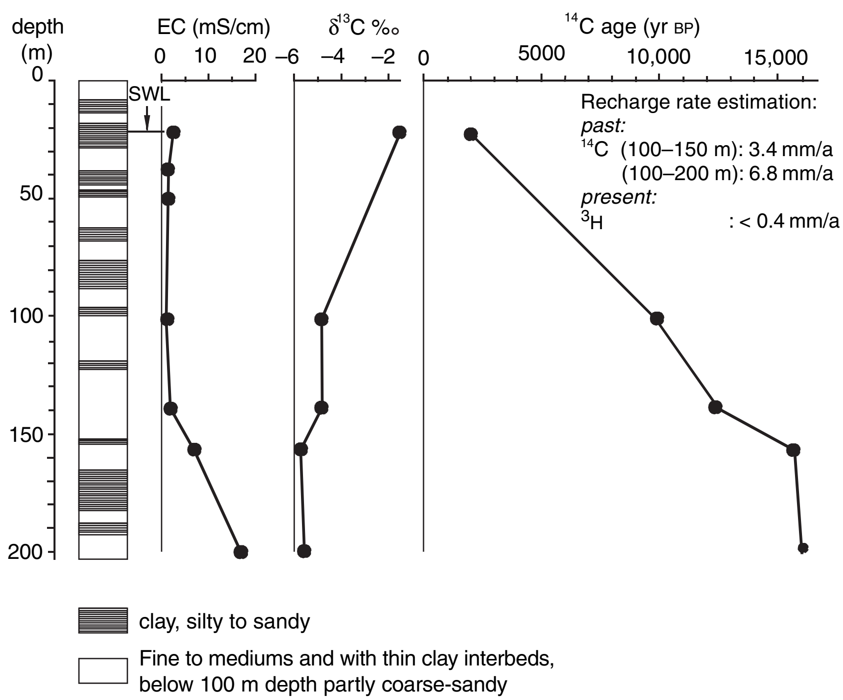

Fig. 3. Layer-wise sampling during test hole drilling yielded the ^{14}C age/depth gradient corresponding to a past recharge rate between 3.4 and 6.8 mm/a which exceeds the present-day maximum recharge rate of 0.4 mm/a by a factor of 8–16. Groundwater recharge commenced with the last long-lasting pluvial period in this region at 13 000 year BP. The shallow groundwater sample was contaminated during sampling by isotope exchange with atmospheric CO_2 shown in increased $\delta^{13}C$ and ^{14}C values (Geyh & Ploethner 1995).

Desert, Western Rajasthan, India (Chandrasekharan *et al.* 1988; Sukhija *et al.* 1996), analysis of tritium and cobalt-60 tracer profiles revealed that the minimum groundwater recharge rate amounts to a minimum 1% of the annual rainfall of about 300 mm, except at one place where it is 13%. The potential evaporation rate is 2000 mm/a. In the Indian part of the Thar Desert Sharma & Gupta (1985, 1987) found higher groundwater recharge rates of between 5 and 12% of the annual rainfall of 200–400 mm, using the tritium-tagging method. However, south of the above-mentioned area, in Gujarat, 3 to 11% of the mean annual rainfall of 500 mm contributes to groundwater recharge, determined by the peak and total tritium methods (Sukhija & Rama 1973; Sukhija & Shah 1976).

We determined the recharge rate of our study area using the tritium value of water samples collected from dug wells. Bomb tritium is present in measurable quantities in the hydrological cycle since 1955 and is used to determine the present recharge rate of groundwater. It was measured without enrichment with a 6-l Oeschger-type counter filled with 3 bar of C_2H_6. The counting gas ethane was prepared from commercial fossil acetylene and hydrogen produced by reduction of a 20 ml water sample with hot magnesium. The background counting rate of the detector is 0.3 cpm and the ^{3}H sensitivity 30 TU/cpm. The detection limit is about 1.2 TU for a counting time of 16 h.

Groundwater recharged after about 1960 contains tritium which can be detected with high reliability. Neither the fresh groundwater down to a depth of 130 m nor the brackish to saline groundwater down to 200 m contained tritium above the radiometric detection limit. Even the samples from dug wells were free of measurable tritium. From 1965 to 1990 rainwater contained 15–30 TU (IAEA 1981, 1983); we measured 16 to 48 TU in surface water and shallow bank storage from the Sutlej River in 1986. The decay-corrected total accumulated tritium (sum of the product of annual precipitation and its tritium value) c_{int} was estimated to be about 330 (TU × m) for Pakistan for the period between 1950 and 1990.

These results allowed us to calculate the maximum recharge rate of groundwater. If the thickness of the water column in a dug well is h, the measured ^{3}H value of groundwater is c, and the total porosity n is known, then the groundwater recharge rate r is given by:

$$r = \frac{c \times h \times n}{c_{\mathrm{int}}} \times p$$

with p as annual precipitation of 200 mm. For a water column of 1 m in the dug wells, a total porosity of 40% and a tritium value of <1.2 TU, a present recharge rate of <0.4 mm/a is obtained which makes up <0.2% of the mean annual rainfall of 200 mm. Consequently, according to these tritium measurements the present recharge of groundwater in Cholistan is negligible.

Chandrasekharan *et al.* (1988) found similar low recharge rates of around 1% of the annual precipitation in western Rajasthan. Edmunds (1998) applied the chloride method at many sites in the Sahara Desert and states that groundwater recharge ceases at an annual precipitation <200 mm/a.

A methodically independent estimate of the recharge rate necessitated the ^{14}C analysis of dissolved inorganic carbon (DIC) extracted from horizon-wise sampled fresh (depth <155 m) and saline groundwater samples (depth 155–200 m) from the test hole T/H-26 (Fig. 3). The ^{14}C age/depth gradient for the depth interval of 100 to 150 m (100–200 m) representing 5730 years (5920 years) of recharge is, however, only 115 a/m (59 a/m). Adopting the total porosity of 40%, the inverse value equals the average recharge rate of 3.4 mm/a (6.8 mm/a). This rate is larger by a factor of 8 (17) than the present recharge rate of the study area derived from ^{3}H measurements. This result provides evidence that the discovered fresh groundwater was not recharged under present climatic conditions. A change in the palaeohydrologic situation might have been responsible.

The recharge rates found in other studies are shown in Fig. 4. There is a trend of an increasing

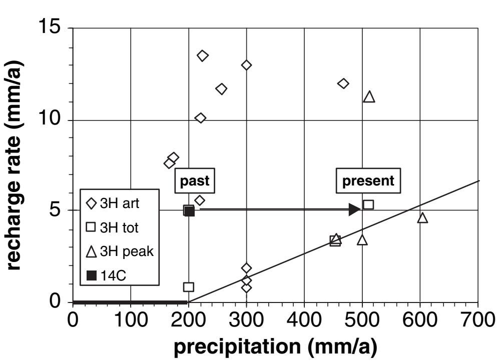

Fig. 4. Isotopically determined present recharge rates from Gujarat and Rajasthan in India. Below 200 mm/a precipitation Edmunds (1998) assumes zero recharge (black line). Between 200 and 600 mm/a there is a clear trend of an increasing present recharge rate. The higher values determined by Sukhija & Rama (1973) and Sukhija & Shah (1976) cannot be explained. The recharge rate of fossil groundwater in the Thar Desert is shown as a black square. From the linear trend a past annual precipitation of about 550 mm may be derived, in agreement with the expectation by Enzel *et al.* (1999).

recharge rate of 0–5 mm/a with annual precipitation of 200–600 mm/a. However, there are also values between 5 and 15 mm/a which were determined between 1973 and 1976 by Sukhija and co-workers. The reason for this discrepancy in results is not known. However, recharge rate estimates generated by the Water GAP Global Hydrological Model (Döll *et al.* 2003) and presented in the map *Groundwater Resources of the World* (BGR & UNESCO 2004) may yield an explanation. The model calculates groundwater recharge as a fraction of total runoff; however, for semi-arid and arid areas the model has been tuned against estimates of groundwater recharge derived from chloride and isotope data. The average annual diffuse groundwater recharge (1961 to 1990) of the Thar Desert of Pakistan is estimated to range from 0 to 5 mm/a, which increases towards India in the east from 5 to 20 mm/a.

Palaeohydrologic situation in the Thar Desert

The different recharge rates of the shallow and deep groundwaters in the study area may be explained by means of the palaeohydrologic situation. Reliable temporal information exists from the eastern part of the Thar Desert in India. Sedimentological and geochemical evidence as well as ^{14}C dates of salt-lake deposits of Sambhar (Thar Desert), Lunkaransar (India) and Didwana Lake (SE Cholistan) revealed that pluvial conditions existed between 13 000 and about 4000 years BP (Singh *et al.* 1972, 1974; Wasson *et al.* 1984; Fig. 5). Two pluvial periods might have existed which lasted from 13 000 to 8000 years BP and from 7000 to 4000 years BP. However, groundwater dating was necessary to gather palaeoclimatic information of our study area. It was done on the basis of

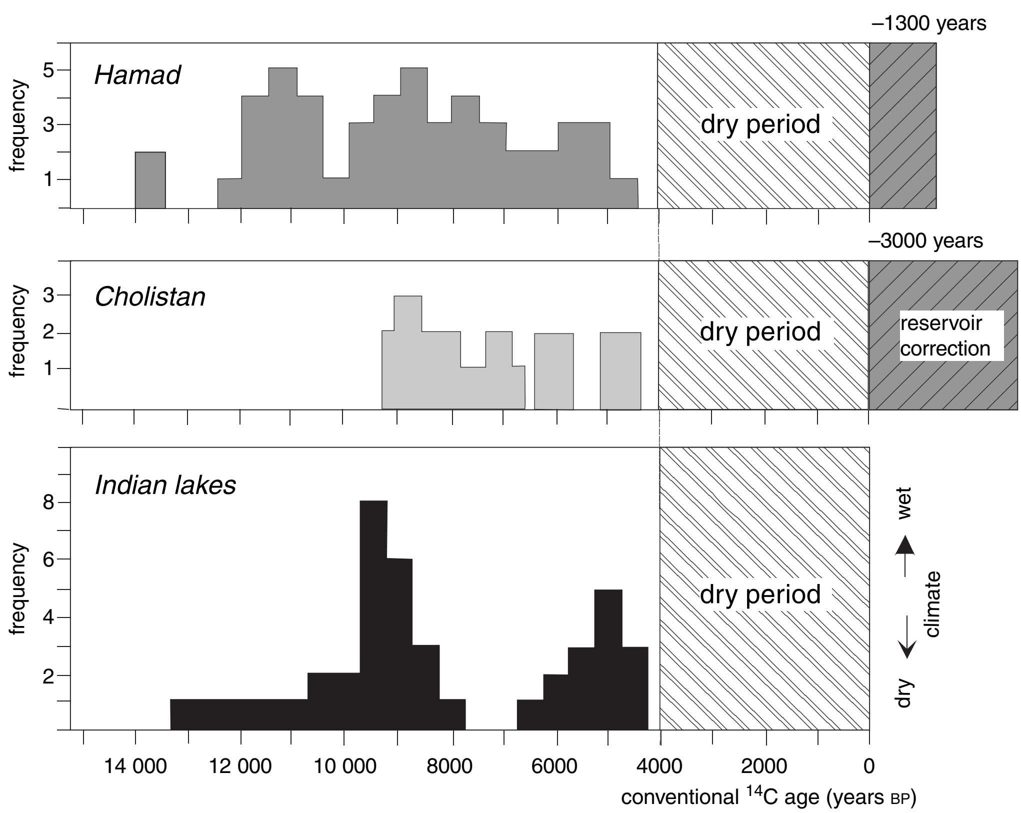

Fig. 5. Histograms of ^{14}C dates reflecting the deterioration of the pluvial conditions within the desert belt between North Africa and the Thar Desert at about 4000 years BP. (**Bottom**) ^{14}C dates from organic sediments from ancient salt lakes in the Indian desert (Singh *et al.* 1972, 1974). (**Middle**) DIC ^{14}C dates for groundwater samples collected in Cholistan (Geyh & Ploethner 1995). (**Top**) DIC ^{14}C dates of groundwater samples from the Hamad Basin (Iraq, Jordan, Syria and Saudi Arabia). The reservoir corrections for the Cholistan and Hamad groundwater samples of −3000 and −1300 years, respectively, were empirically estimated (Geyh *et al.* 1985).

the radioactive decay of radiocarbon in DIC ($= HCO_3^- + CO_2$) of the discovered groundwater body. Carbon-14 in biogenic carbon dioxide is involved during the process of groundwater recharge. A ^{14}C age range of about 40 000 years is covered. The participation of fossil or almost fossil soil lime necessitates a reservoir correction of the conventional ^{14}C ages of DIC. From this, the time scale of the groundwater is transferred to calendar years. The precision of the ^{14}C dates may not be better than about ± 500 years. DIC was precipitated from 60 litres of groundwater with a saturated barium hydroxide solution in the field. The ^{14}C measurements were done on acetylene as counting gas prepared via decomposition of DIC to CO_2, transformation to lithium carbide and C_2H_2 with tritium-free water. The ^{14}C measurements were carried out with proportional counters of different size. The DIC ^{14}C age of 16 fresh to saline groundwater samples covers a time span of 7700 to 15 900 years BP (Table 1).

The reliability of the uncorrected DIC ^{14}C ages was verified by a geohydraulic estimate. The trend of ^{14}C ages from east, at tube well T/W-9, to west, at tube well T/W-8, over a distance of 18 500 m allowed estimation of the tracer velocity v_{tracer}. We obtained 6 m/a or 0.016 m/day based on the ^{14}C age of 3000 years. It follows according to Darcy's law:

$$v_{Darcy} = K \times I = v_{tracer} \times n_e$$
$$K = (v_{tracer} \times n_e)/I$$
$$K = (0.016 \times 0.15)/0.0003$$

Using the regional hydraulic gradient $I = 0.0003$ and the effective porosity $n_e = 0.15$, a value of $v_{tracer} = 6$ m/a is obtained which leads to the hydraulic conductivity $K = 8$ m/day. This is the lowest K established by aquifer testing. Applying the total porosity (40%) as appropriate for mass transport calculations with ^{14}C (Geyh *et al.* 1984; Maloszewski & Zuber 1985) the K value is around 20 m/day which is the mean K found by aquifer testing. This means that the tracer velocity reliably reflects the regional geohydraulic situation.

In order to reconstruct the palaeoclimate of the study region, the reservoir correction had to be determined. The limnologic time marks suggest a reservoir correction of -3000 years resulting in a range of the actual water age from 12 900 to 4700 years BP.

A methodically independent estimate of the reservoir correction was possible, with the ^{14}C age of 2100 years BP (= 76.8 pMC) of the shallow groundwater sample taken from the test well T/H-26 beside an artificial pond. In Cholistan slightly calcareous soils prevail. In 1990 the atmospheric and biogenic CO_2 had a specific ^{14}C activity of about 114 pMC. The lowering of the ^{14}C value to 77.4 pMC in DIC of recent groundwater, or 32%, is due to dissolution of soil lime with a ^{14}C activity of 0 pMC. It corresponds to a reservoir correction of -3100 years. This value is comparable with that mentioned above. The same process might have been effective during recharge of the discovered fossil groundwater body as the pedologic conditions did not change during the last 15 000 years.

Hydrochemical corrections of the ^{14}C values were not necessary as the ranges of the bicarbonate concentration and the $\delta^{13}C$ values of 200 to 500 mg/l and -4.9 to -5.8‰, respectively, are small. The uncertainty of the ^{14}C ages resulting from these ranges is already inherent in the uncertainty of ± 500 years of the reservoir-corrected ^{14}C groundwater dates.

The validity of the methodically differently determined ranges of the pluvial periods applying -3000 years as reservoir correction is supported by the obvious synchronism of the groundwater recharge period in the sub-Indian continent (e.g. Cholistan) and in the Middle East (e.g. Hamad Basin; Geyh *et al.* 1985; Geyh 1994; Fig. 5). The time span lasting from 13 000 to about 4000 years BP includes the Neolithic Pluvial (9000 to 4500 years BP), geomorphologically proven by Kaiser (1973) and Kaiser *et al.* (1973). A similar palaeoclimatic scenario prevails in the Eastern Sahara where in Western Nubia the widespread distribution of stable freshwater lakes is documented by intense lacustrine sedimentation lasting from 11 500 to 5500 years BP. The deterioration of the palaeohydrological conditions between about 5000 and 4000 years BP within the desert belt extending from North Africa (Claussen *et al.* 1999; Pachur & Hoelzmann 2000) to the Thar Desert supports the concept that the Old Hakra River most likely disappeared due to the shift of the monsoonal belt southwards, as concluded by Wilhelmy (1969). Consequently, seepage of water from the Old Hakra River during periods of inundation is considered as the source of the fresh groundwater resource in Cholistan. According to Radhakrishna & Merh (1999) tectonic events also played a role. It has been assumed (Singh *et al.* 1972, 1974; Wasson *et al.* 1984) that the end of this pluvial period coincided with the decline of the Harappan Culture around 4000 years ago. But Enzel *et al.* (1999), while establishing a more reliable chronology, came to the conclusion that the end of the lake phase preceded the early and mature Harappan Culture by more than 800–1000 years.

The scatter of the ^{14}C water ages along the fresh groundwater body may be explained by meandering

Table 1. *Isotope results from Cholistan, Pakistan*

Location	Well	Depth (m)	$\delta^{18}O$ (‰)	$\delta^{2}H$ (‰)	d_{ex} (‰)	^{3}H (TU)	$\delta^{13}C$ (‰)	^{14}C (pMC)	^{14}C age (year BP)	^{14}C age corrected
Mojgarh Forest	T/W-0	42–91	−5.08	−40.8	−0.2	<3.1	−4.9	24.9 ± 0.6	11160 ± 200	8200
Mojgarh Forest	T/W-0	42–91	−4.89	−40.2	−1.1		−4.9	25.1 ±0.6	11080 ±190	8100
S of Mirgarh Fort	T/W-4	61–120	−5.38	−41.8	+ 1.3	<1.9	−3.4	22.6 ±0.4	11965 ±155	9000
N of Wakarwala Toba	T/W-5	52–146	−4.79	−40.1	−1.8	<2.1	−4.3	23.1 ±0.6	11760 ±200	8800
S of Januwali	T/W-6	51–105	−5.25	−40.9	+ 1.7	<2.6	−5.8	38.0 ±0.7	7780 ±145	4800
W of Mojgarh	T/W-8	56–99	−5.11	−39.8	+ 1.1	<4.6	−5.0	22.6 ±0.4	11960 ±130	9000
SSW of Chapuwala Toba	T/W-9	58–86	−4.94	−38.2	+ 1.3		−5.1	32.7 ±0.4	8965 ±105	6000
N of Mojgarh	T/W-10	50–104	−3.37	−39.6		<2.4	−5.4	38.6 ±0.7	7655 ±145	4700
S of Chapuwala Toba	T/W-11	45–99	−2.68	−37.7	−12.6		−5.0	32.6 ±0.5	8990 ±115	6000
Pirwala Toba	T/W-12	55–108	−4.78	−40.4	−16.3		−4.4	27.6 ±0.6	10360 ±175	7400
Karawala Tibba	T/W-14	42–79	−5.40	−42.9	−2.2		−3.8	28.5 ±0.5	10100 ±135	7100
NE of Haiderwala Toba	T/H-26	21–24	−5.25	−42.0	+ 0.3	<2.1	−1.5	76.8 ±0.7	2125 ± 75	0
NE of Haiderwala Toba	T/H-26	101–104	−5.35	−40.9	+ 1.9	<1.3	−4.9	29.0 ±0.3	9930 ± 75	6900
NE of Haiderwala Toba	T/H-26	137–140	−5.18	−39.4	+ 2.0	<2.0	−4.9	21.6 ±0.3	12325 ±105	9300
NE of Haiderwala Toba	T/H-26	155–158	−5.23	−41.7	+ 0.1	<1.8	−5.8	14.2 ±0.2	15665 ±115	12700
Toba	T/H-26	198–201	−5.34	−41.4	+ 1.3	<2.3	−5.7	13.9 ±0.5	15850 ±315	12900
NE of Haiderwala Mojgarh	H/P-200	35–38	−5.28	−40.2	+ 2.0	<2.1	−4.5	21.8 ±0.5	12230 ±175	9200
Mansura/Ranger's post	H/P-162	37–40	−4.32	−34.3	+ 0.3	4.2 ±0.7	−4.8	68.4 ±0.9	3060 ±105	0
Chak-315 at Mosque	H/P no	19–20	−7.66	−47.6	+ 13.7		−11.6	96.8 ±1.0	255 ± 85	
Hakra Left at RD 47	Canal	0	−7.44	−46.8	+ 12.7		−5.0	95.9 ±1.2	335 ±100	
Minor L-3	Canal	0	−7.79	−51.6	+ 10.7		−5.0	89.8 ±1.2	865 ±105	
Sutlej River / South. Bank	River	0	−6.55	−44.3	+ 8.1	16.4 ±1.3		94.9 ±1.4	415 ±120	
215 m N at Sutlej River	Irr.-Well	15–64	−8.02	−54.6	+ 9.6	9.0 ±1.2	−8.7	94.4 ±0.7	465 ± 65	
565 m N at Sutlej River	Handpump	7–8	−8.42	−56.3	+ 11.1	47.5 ±1.4	−13.3	97.2 ±0.7	230 ± 60	
620 m N at Sutlej River	Irr.-Well	15–64	−8.11	−56.4	+ 8.5	40.5 ±1.2		93.4 ±0.7	545 ± 55	

Source: Geyh *et al.* (1995).

of the Old Hakra River. As the seepage could have occurred within decades, the ages reflect the time when river water seeped into the subsurface. Moreover, one has to keep in mind that the sampled wells might pump young groundwater from the aquifer below the old bed of the Old Hakra River rather than the youngest. This may explain some deviation from the historical records of its perennial behaviour.

An alternative interpretation is, however, also feasible. The youngest groundwater occurs around Khirsar village in the west where agricultural irrigation was applied. The corrected water ages here range from 4700 to 6000 years BP coinciding with low $\delta^{13}C$ values.

Origin of the fresh groundwater

Information on the origin of the fresh groundwater recharged by local precipitation or an influx of groundwater recharged far away can be obtained from the stable isotope compositions of oxygen, hydrogen and carbon. The $\delta^{18}O$ and $\delta^{2}H$ values may yield information on the altitude of the catchment area and the processes that occurred at the ground surface before groundwater recharge. We determined the isotope compositions of oxygen and hydrogen of 101 samples by means of two MICROMASS mass spectrometers (602C for the first two and 602E for hydrogen) and CO_2 exchange with 20 ml of sample water. Hydrogen isotopes were determined on hydrogen produced by reduction of 20 μl water with zinc. The results are given in the common delta-notation referring to the Vienna-SMOW standard. The $\delta^{13}C$ values were determined on CO_2 prepared from the DIC sample and referred to the marine PDB standard. The relationship between $\delta^{18}O$ and $\delta^{2}H$ values of surface water, shallow bank storage of the Sutlej River and fossil groundwater is shown in Fig. 6. The $\delta^{18}O$ and $\delta^{2}H$ values of the surface water and related recent groundwater form a narrow cluster on the global meteoric water line (GMWL) ranging from -6.3 to -8.4‰ for $\delta^{18}O$ and from -38 to -56‰ for $\delta^{2}H$. This water originates from the Sutlej River with a catchment area at a mean altitude of 3700 m a.s.l. of the foothills of the Himalaya Mountains. The $\delta^{18}O$ and $\delta^{2}H$ values of rainwater taken in the Hindukush Mountains. (5300 m a.s.l.) and at the Khyber Pass fit a local meteoric water line of $\delta^{2}H = 8 \times \delta^{18}O + 15$. This is in accordance with the observation that groundwater recharge in western Pakistan is predominantly derived from the winter monsoon but also through westerly storms from the eastern Mediterranean with higher deuterium excess of $+22$‰ (Gat & Carmi 1970). Some rainwater samples that we took at Lahore and Marot/Cholistan during the monsoon period delivered $\delta^{18}O$ and $\delta^{2}H$ values which also scatter around the GMWL.

The majority of $\delta^{18}O$ and $\delta^{2}H$ values obtained from 101 fossil fresh and saline groundwater

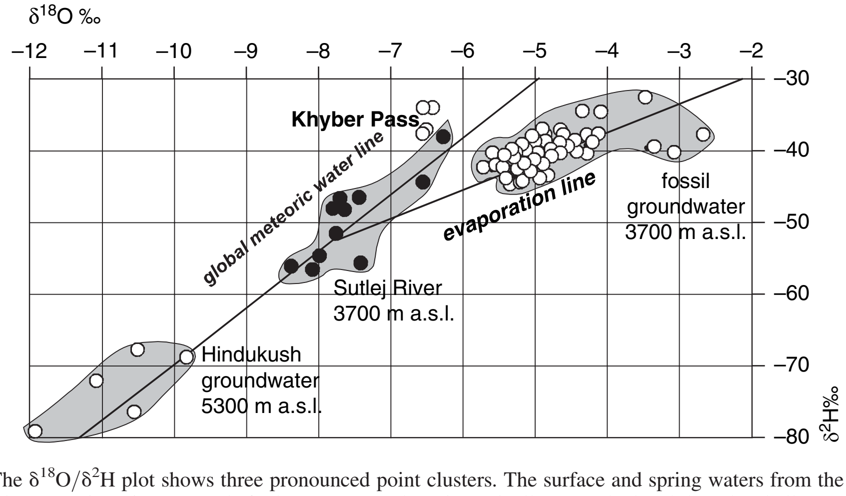

Fig. 6. The $\delta^{18}O/\delta^{2}H$ plot shows three pronounced point clusters. The surface and spring waters from the high Hindukush Mountains (5300 m a.s.l) fit the GMWL and are isotopically more depleted than the shallow groundwater and canal water derived from the Sutlej River with a mean altitude of the recharge area of 3700 m a.s.l. The isotope data of the latter also form a cluster on the MWL which intercepts with an assumed evaporation line through the cluster of isotope data of the explored fossil fresh groundwater body. This finding is a strong argument that this groundwater was recharged by seepage from the Old Hakra River. Water samples from the Khyber Pass provided evidence that Mediterranean storms reach this region.

samples from test holes and tube wells form a cluster ($\delta^{18}O$, -5.7 to -4.2‰; δ^2H, -44 to -37‰) along an evaporation line: $\delta^2H = 4.1 \times \delta^{18}O - 20$. This evaporation line originates from the GMWL with an intercept at $\delta^{18}O = -7.4$‰ and $\delta^2H = -49$‰ within the cluster of the corresponding values of the young groundwater of Sutlej River. The deuterium excess ($= \delta^2H - 8 \times \delta^{18}O$) between $+8$ and $+12$‰ (Table 1) provides evidence that some of the river water might have seeped into the ground without evaporative enrichment of the heavy oxygen and hydrogen isotopes. As the stable oxygen and hydrogen isotope compositions of the fossil groundwater of Cholistan vary only slightly and are not correlated with the salinity of the water, it provides evidence that the fresh and brackish to saline groundwaters are of the same origin and were recharged under similar climatic conditions. The wide range of salinity is explained by the varying salt concentration in the sediments.

The recharge process of the fossil groundwater becomes understandable if we compare the evaporation line with that of the recent groundwater of the New Delhi area, India: $\delta^2H = 4.8 \times \delta^{18}O - 17.5$ (Das *et al.* 1988). This shallow groundwater is recharged within alluvial sediments during which inundation of the Yamuna River, lasts many weeks. The phenomenon that the heavier stable oxygen and hydrogen isotopes of water become enriched under semi-arid and arid climate conditions before and during groundwater recharge has been known for a long time (Gonfiantini *et al.* 1974; Münnich *et al.* 1984; Müller *et al.* 1984).

It can be ruled out that the $\delta^{18}O/\delta^2H$ cluster of the fossil groundwater samples joins a meteoric water line with a deuterium excess less than $+10$‰ as this is typical for groundwater recharged during a cooler and wetter climate than that of the present day (Job *et al.* 1975). Such depleted stable isotope compositions with a MWL $\delta^2H = 8 \times \delta^{18}O + 4$ were found for groundwaters in the Thar Desert of Western Rajasthan, India, where a conventional ^{14}C age of 18 000 years BP is considered as evidence for recharge during a cooler and wetter climate (Chandrasekharan *et al.* 1988).

The alternative explanation of the presence of groundwater with enrichment of heavy isotopes is diffuse recharge during relatively wet periods and partial evaporation through the unsaturated zone. This process would not be in conflict with the missing correlation between the deuterium excess and the salinity of the groundwater. However, this interpretation does not explain why the present fossil groundwater is restricted to the former course of the Old Hakra River.

The stable isotope composition of carbon of the DIC delivers complementary hints on the origin of the groundwater. The $\delta^{13}C$ value is controlled by the interaction of biogenic carbon dioxide with lime in the topsoil or other sources.

The $\delta^{13}C$ value of atmospheric carbon dioxide amounts to around -7‰, those of biogenic carbon range from -9 to -35‰ depending on the type of photosynthesis pathways. Under different environmental conditions the $\delta^{13}C$ values of C_3 plants (these are generally plants of the humid zone, e.g. wheat) range from -20 to -35‰ (-25 to -28‰). The $\delta^{13}C$ values of C_4 plants (e.g. steppe grass, maize, sorghum, sugar cane) range from -9 to -16‰ (-12 to -14‰). Characteristic values are given by Merwe (1982) and Hillaire-Marcel (1986).

The $\delta^{13}C$ values of DIC of groundwater recharged in sedimentary catchments range between -10 and -14‰, while in crystalline areas $\delta^{13}C$ values up to -20‰ are found. If marine carbonates were dissolved in the subsurface as the result of secondary chemical reactions, and / or if the recharge area is covered by C_4 vegetation, or seepage from rivers or canals takes place, then $\delta^{13}C$ values in groundwater less negative than -10‰ are found.

The $\delta^{13}C$ values of all groundwater samples taken in Cholistan vary from -3.4 to -5.8‰. The main reason for this relatively positive $\delta^{13}C$ value might be that seepage from the former Old Hakra River was the predominant source of groundwater recharge. The $\delta^{13}C$ values of groundwater tend to become more negative downstream (Fig. 7). This trend might indicate that the influence of biogenic carbon dioxide becomes more dominant in the down-gradient direction. This interpretation fits with the concept that the Old Hakra River became shallower and wider towards the Derawar inland delta and therefore vegetation on river

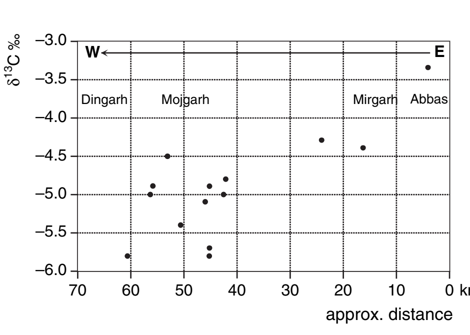

Fig. 7. Decline of the $\delta^{13}C$ values of DIC along the palaeochannel of the Old Hakra River. The *x*-axis shows the east-west distance between the sampling points.

banks was more abundant in the west. The $\delta^{13}C$ values of the surface water lie between -8.7 and $-5.0‰$ while related recent groundwater attains $\delta^{13}C$ values ranging from -13.3 to $-11.6‰$.

The groundwater study was carried out by the Hydrogeology Directorate, Water and Power Development Authority (WAPDA), Lahore, and the Federal Institute for Geosciences and Natural Resources (BGR), Hannover, in the framework of a Pakistan–German Technical Cooperation Project. The project was funded by the Ministry of Water and Power, Government of Pakistan, and the Federal Ministry for Economic Cooperation and Development (BMZ), Government of the Federal Republic of Germany. Special gratitude is extended to the WAPDA drilling crews, field assistants and hydrogeologists who strongly supported groundwater sampling in the field. Thanks to the staff of the NLfB isotope laboratory in Hannover. We also thank Dr S. K. Gupta for reviewing our paper and valuable comments.

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Transient response of groundwater systems to climate changes

K.-P. SEILER[1], W.-Z. GU[2] & W. STICHLER[1]

[1]*GSF National Research Centre, D-85758 Neuherberg, Germany (e-mail: seiler@gsf.de)*

[2]*Institute of Hydrology, Hohai University, Nanjing, P. R. China*

Abstract: Groundwater flow is steered by both the groundwater recharge rate and by discharge altitudes above or below sea level; it is further controlled by the hydraulic properties of the aquifer system and often contains a transient flow component affected by natural hydrologic processes. All present groundwater discharges have both recent (<100 years) and past groundwater recharge components (>100 years). The ratio of the present to the past groundwater recharge depends on the climate zone: it is large in humid and small in arid areas, hence at low recharge rates transient, and at high groundwater recharge rates steady-state conditions prevail. Developing groundwater management strategies while neglecting any transient response of groundwater resources, and conducted in sensitive recharge/discharge areas like dry lands, results in either over-estimates or under-estimates of safe yields of groundwater resources, and thus may lead to non-sustainable resource development. The consequence of this would be groundwater depletion and often also a deterioration of the hydraulic properties of the aquifer system by subsidence, which both take place only after a long period of time.

Few issues have produced as much scientific and political attention and controversial debate as the increase of greenhouse gases (CO_2, N_2O and CH_4) in the atmosphere from combustion processes, their effect on global warming, and its consequences for the life and health of human beings and ecosystems including groundwater regime.

The global warming scenario, from the beginning of the industrial age until the end of the twentieth century, caused an increase in the average climate temperature of about 0.6°C and an average sea level rise of about 0.1 to 0.2 m. In the same period of time rainfall increased in some areas in the north of the northern hemisphere and decreased in some areas of the tropics and subtropics of Africa and southeast Asia (IPCC 2001).

Climate modelling predicts for the end of the twenty-first century a temperature increase of a minimum of 1.4°C to a maximum of 5.8°C as compared to the beginning of the industrial period (IPCC 2001). The consequences of this include: an increase of evaporation, transpiration and precipitation, but also aridity; a higher variability of the rate and intensity of precipitation, and further decrease of precipitation in South Africa, Australia, the Near East, Mediterranean and in Central America; a shift of the magnitude and time of peak stream flow in mountain areas from smooth to strong, from mid- or late spring at present, to the end of the winter season in the future; increased melting and calving of polar ice; the Arctic Sea will be partly or completely free of ice during the summer period, leading to a decrease in the overall albedo; a further mean sea level rise from 1990 to 2100 of 0.1 m to 0.9 m, exceeding the rate observed in the twentieth century, hence damaging coastal lowland areas and moving the salt water–fresh water interface landwards, a reduction of the permafrost in the northern hemisphere by about 0.5 m and the disappearance of lakes in grasslands of the same areas.

Thus, in the future global warming is expected to have a major impact on the magnitude of both surface and subsurface runoff. Experience has shown that such changes show up instantaneously in compartments with short turn-over times such as rivers, but in freshwater reservoirs with long turn-over times, such as groundwater, they appear with delay times of decades or even millennia. Table 1 compiles mean turn-over times (MTT) for important surface and subsurface freshwater reservoirs. It can be seen that surface water has MTTs of days, shallow groundwater of years, deep groundwater and glaciers of centuries to millennia; MTTs of lakes are in between those of surface and subsurface reservoirs.

Globally, the predicted trend of precipitation changes related to the present water availability on continents would not result in serious problems for the Americas and Australia (>15000 m^3/capita/year; Table 2) as compared to a total water demand (direct and indirect water) of 1000 to 1500 m^3/capita/year. In contrast, some areas in southern Europe, north Africa, and the Near and Far East already suffer, or will suffer in the near future, from water scarcity.

We are living in an exceptional geological period, the Cenozoic glacial period, with natural,

From: DRAGONI, W. & SUKHIJA, B. S. (eds) *Climate Change and Groundwater*. Geological Society, London, Special Publications, **288**, 111–119.
DOI: 10.1144/SP288.9 0305-8719/08/$15.00

Table 1. *Mean turn over times (MTT) of waters in different reservoirs*

Reservoir type	MTT
Cold glaciers	>100 000 years
Temperate glaciers	<500 years
Connate water	>1 000 000 years
Deep groundwater	≫100 years
Shallow groundwater	<100 years
Lakes	*c.* 15 years
Rivers	*c.* 15 days
Atmospheric water vapour	*c.* 10 days

dramatic climate changes, to which man-made short-term changes become an additional, accelerating challenge. Against this background, short-term solutions to management and protection of freshwater resources have been discussed in detail, but developments of water resources, lasting for more than a few generations, have not yet been well considered. To compile such information, past long-term changes of groundwater discharge and heads can be analysed in water-sensitive areas such as dry, cold or flood areas, using geohydraulic and environmental isotope techniques. Then, these results can be extrapolated to the future and allow a better estimate of the consequences of expected climate changes for subsurface waters. Generally the focus is on short-term strategies for development of groundwater resources to address the challenges caused by climate changes rather than long-term strategies for long-lasting solutions. This paper deals with one of the many aspects of the response of subsurface water resources to climate changes (e.g. global warming), especially the transient behaviour/response of subsurface water in dry lands.

Table 2. *Available water (1921–1985) in different world regions as related to a yearly bulk discharge from continents of 44 800 km^3*

Continent	Percentage of bulk discharge	Available water in (m^3/capita/year)
Asia	31.5	3920
Europe	6.7	4200
Africa	9.8	5720
N America	18.4	17 400
S America	28.0	38 200
Australia & Oceania	5.6	83 700

Data from UNESCO (1999).

Characteristics of groundwater flow based on field and modelling studies

Groundwater recharge and discharge altitudes are driving forces of groundwater flow. Groundwater recharge occurs in all climate zones, albeit at different rates. In desert regions groundwater recharge is small (<5 mm/a) and occurs irregularly (Verhagen *et al.* 1979). In semi-arid regions it is >5 mm/a and <25 mm/a and undergoes significant annual fluctuations. In tropical regions it ranges mostly between >25 mm/a and <100 mm/a and again varies considerably year-to-years. In temperate humid regions groundwater recharge may reach many hundred millimetres per annum. Groundwater recharge may also occur in permafrost areas (Michel & Fritz 1978), but at very low rates.

Supposing that an equivalent of 44 800 km^3/a of continental runoff (UNESCO 1999) recharged groundwater resources (total volume about 8 200 000 km^3) homogeneously throughout the profile, a minimum mean residence time of 183 years would ensue; this is a minimum number, because continental discharge does not completely contribute to groundwater recharge. The result of this bulk calculation, however, is in obvious disagreement with all field observations on pollution, hydrochemistry and environmental isotopes in groundwater resources. Hence, flow in the subsurface cannot be homogeneous throughout the vertical profile, as has been qualitatively stated by Toth (1995), Freeze & Witherspoon (1967) and many others, and has been quantified by numerical modelling by Seiler & Lindner (1995).

Groundwater flows in aquifer systems with particular hydraulic properties. To investigate this influence, numerical modelling was performed (Fig. 1) with: a groundwater movement between an underground water divide (left border in Fig. 1) and a river, collecting totally subsurface discharge (right border in Fig. 1); model dimensions of $z = 400$ m, $x = 15\,000$ m and $y = \infty$ m; a constant groundwater recharge rate of 150 mm/a over the entire catchment; and typical distributions of aquifer conductivities as show in Figure 2 (left column) and a hydraulic effective porosity of 25%.

From about 100 numerical scenario simulations (Fig. 2) it was found that generally more than 85% of the groundwater recharge rate (right column in Fig. 2) joins surface discharges through near-surface aquifers (active groundwater recharge zone or shallow groundwater) and less than 15% reaches deep aquifers (passive groundwater recharge zone or deep groundwater) (Seiler & Lindner 1995). This holds both for uniform and stratified aquifer systems, though is accentuated for stratified aquifer systems.

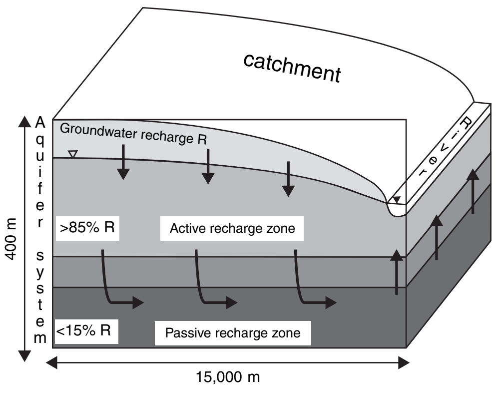

Fig. 1. Block diagram of a catchment underlain by an aquifer system ($z = 400$ m, $x = 15000$ m, $y = \infty$ m). Numbers indicate the average percentage of turn-over of groundwater recharge (R), resulting from some 100 numerical runs with different distributions of hydraulic conductivities.

From modelling of the propagation of both pollutants and environmental isotopes in non-stressed aquifer systems it was found that tritium and surface pollutants propagate in quasi-identical profile sectors. Hence tritium, in concentrations above the measuring accuracy, is considered a good indicator for shallow groundwater, and at non-measurable concentrations for deep groundwater, for groundwater of very short-term and low susceptibility to pollution. Tritium traces subsurface water quasi-exclusively by means of infiltration. The interface between shallow and deep groundwater was therefore defined by the tritium nought line (TNL) (Seiler 1983), separating shallow, young (<100 years) and deep, old (>100 years) groundwater (Fig. 3).

Shallow and deep groundwaters are both of meteoric origin and belong to a hydraulic continuum; both have been identified by field investigations in tropical humid (Alvarado *et al.* 1996), temperate humid (Andres & Egger 1985) and semi-arid climate zones (H. Raanan, Ben Gurion University, pers. comm.). In dry and hyper-dry areas shallow groundwater often has a patchy distribution, because of extended catchment sizes as well as an uneven and sporadic occurrence and distribution of precipitation. Sometimes a thick percolation zone also plays a role in arid areas, through which groundwater recharge flows with very low apparent percolation velocities (<1 m/year); in such dry areas it is therefore often difficult to detect the shallow, young groundwater, but it is always easy to find old, deep groundwater.

In semi-arid regions, shallow groundwater has a thickness of a few metres, in the tropics typically a few decametres and in temperate climates up to 100 m; this thickness depends essentially upon the rate of groundwater recharge as well as the porosities and hydraulic conductivities. Hence all general changes in groundwater recharge rates will lead to an increase or depletion of the thickness of the active groundwater recharge zone. In the case of a reduction of groundwater recharge because of global climate changes, many wells in the active groundwater recharge zone will shift into the passive groundwater recharge zone, leading to a change in the hydraulic behaviour from instantaneous to transient.

Under hydraulically non-stressed and stressed conditions, transient conditions in groundwater flow play only a random role in shallow groundwater as long as groundwater abstraction does not exceed the safe yield. In contrast, exploitation from deep groundwater produces regularly transient conditions (Fig. 4) over decades to millennia as a response to: groundwater extraction at rates exceeding the depth-related, available groundwater recharge rate; and changing meteorological or tectonic and eustatic boundary conditions. This transient response results from the slow adaptation of the groundwater flow field to recharge/discharge changes to reach a new steady-state condition.

The passive groundwater recharge zone is underlain by connate or formation water (von Engelhardt 1960), which is mostly of non-meteoric origin; it has been entrapped in sediments since sedimentation times, and has not returned to the atmosphere.

Groundwater in desert areas

Field observations

Humid areas have typically high groundwater recharge and small catchment sizes; in contrast drylands have extended catchment areas with very low groundwater recharge.

In many dry lands, low salinity lakes occur as well as freshwater springs adjacent to salt lakes or sabkhas. In the Badain Jaran Shamo (Gobi desert) more than 70 lakes with a total surface of area about 27 km^2 are known, having either high (180 g/l) and low salinities (<1 g/l) (Chen *et al.* 2004); the same area also hosts fresh water in dug wells adjacent to saline lakes. In Jordan the Azraq desert freshwater springs yielded about 1.2 m^3/s to the Azraq Sabkha before pumping by gravity wells close to these springs started to supply Amman city with drinking water (Almomani & Seiler 1996). The Chott area of south Tunisia represents a terminal discharge area and sabkha, which is fed by fresh water.

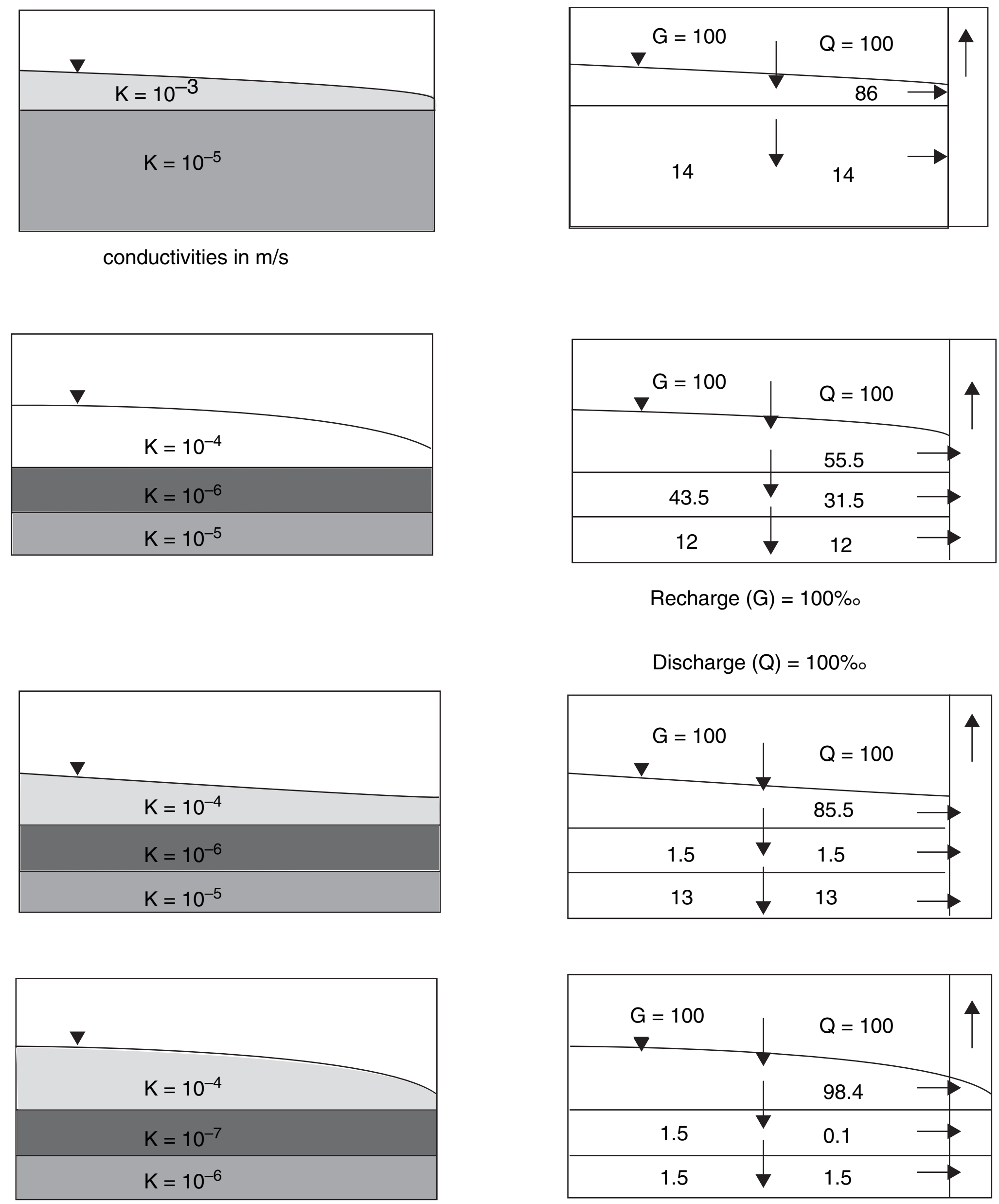

Fig. 2. Selected examples of frequently occurring aquifer systems (left columns) and the distribution of the groundwater recharge (in percentage of infiltration) upon the individual aquifers. G, Infiltration; Q, discharge to rivers.

Radiocarbon and tritium dating of such fresh waters revealed that they are mostly free of tritium, hence older than 50 to 100 years, and carbon-14 ages range from several thousand to 10 000 years. Statistic evaluations of radiocarbon groundwater ages in north Africa and the Arabian peninsula resulted in two frequencies of water ages (25 000 years BP and 5000 years BP), which have been interpreted as periods with significant groundwater recharge (Verhagen *et al.* 1987). Between these two periods only limited groundwater recharge is expected.

Groundwater chemistry from samples collected from dug wells and springs around lakes in the

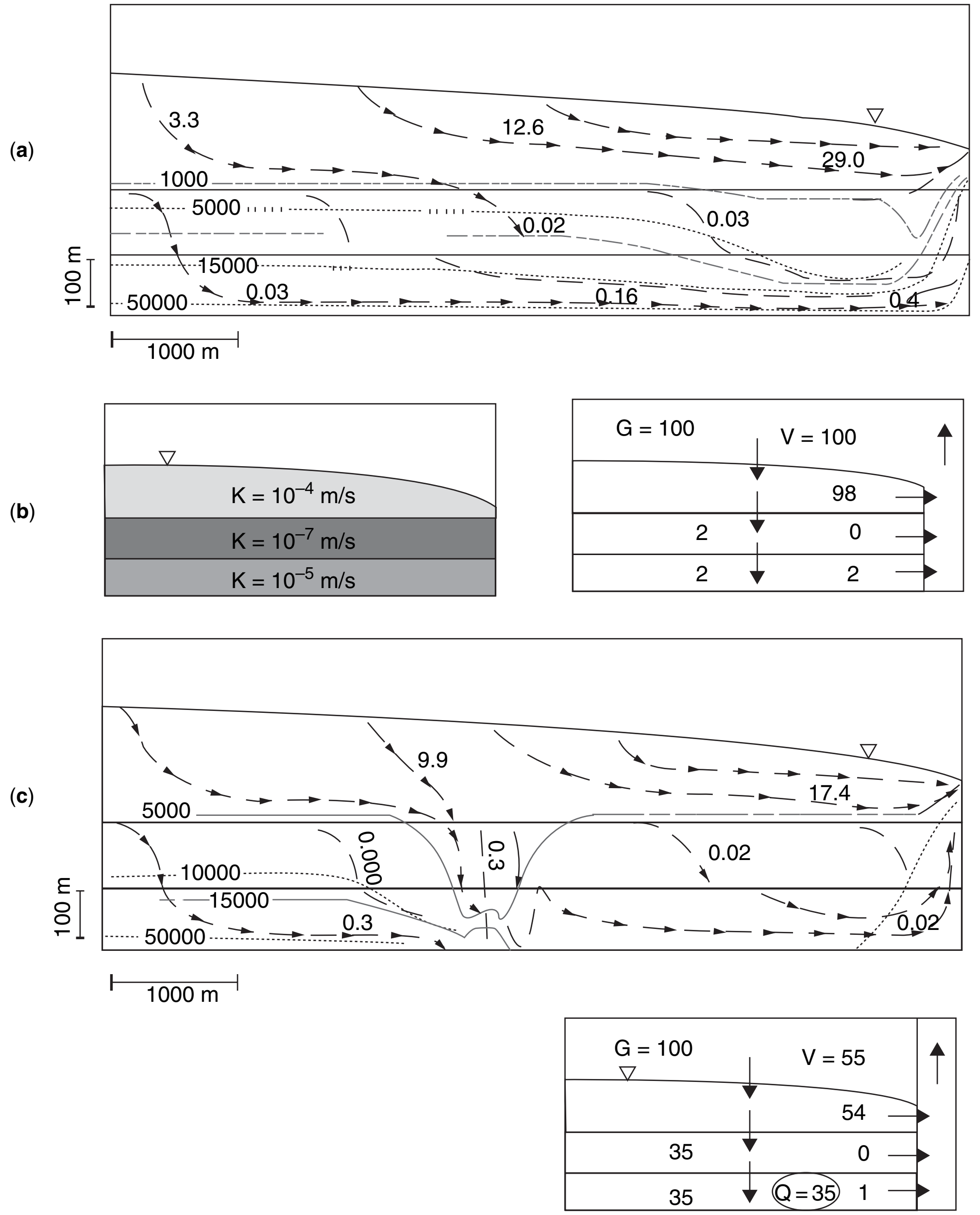

Fig. 3. Flow (arrows) and age distribution (broken lines) in an aquifer system with hydraulic conductivities and relative discharges (**b**, left and right), with (**a**) and without (**c**) groundwater abstraction from deep groundwater.

Badain Jaran Shamo doubtless indicates that the lake waters represent uncovered groundwater, flowing through the lakes with low flow velocities. The present lakes in the Badain Jaran Shamo are residuals of more extended lakes in the historic and geological past. This is deduced from shores 40 m to 50 m above the actual lake levels; the shores are traced by Palaeolithic artefacts, carbonate precipitates from plants growing in shallow water, fossil freshwater snails and organic lake sediments. Carbonates have been dated by ^{14}C to 15 000 years (Hofmann 1999). These relicts and ages show the decay of a groundwater resource beneath the Badain Jaran Shamo, emptying

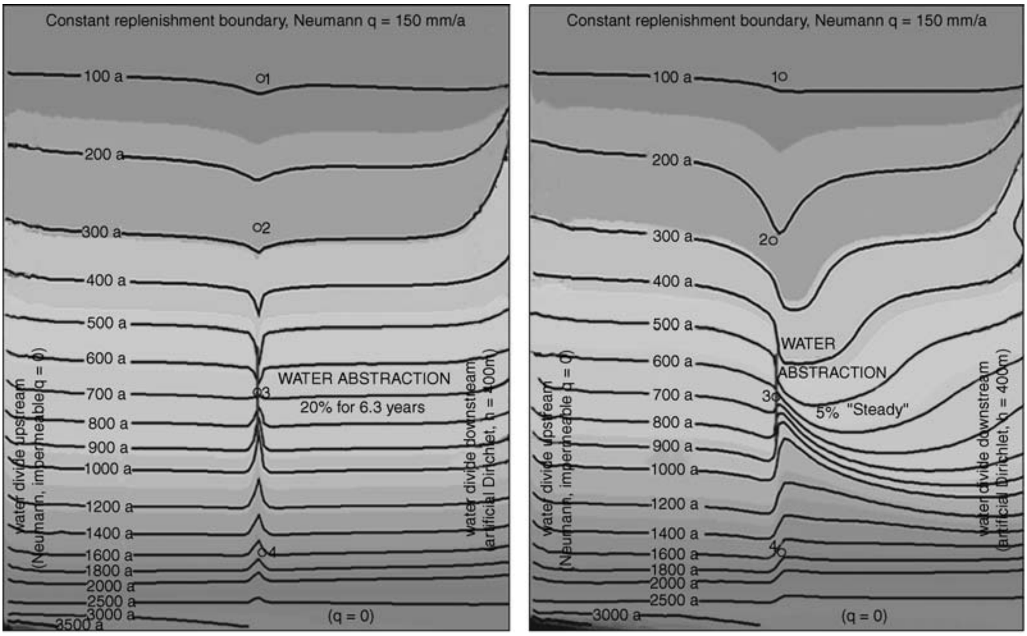

Fig. 4. Response of deep groundwater on exploitation of 20% of groundwater recharge over a run of 6.3 (left) and of 5% of groundwater recharge until reaching steady-state conditions (right). Lines represent water ages in years.

continuously or discontinuously over millennia. Any discontinuous decay can be attributed to climate changes as well as to changes in the size of the subsurface catchment, feeding the discharge; the latter are typical when the subsurface flow system changes, e.g. from exorheic to endorheic.

It is interesting to note that the lakes in the Badain Jaran Shamo have old groundwater without tritium, but most shallow waters from dug wells contain significant amounts of tritium (Tables 3 and 4). This finding may be considered as a qualitative confirmation of results of scenario modelling. Here present groundwater recharge exists, but is of patchy pattern and appears only as a thin cover over groundwater without tritium. As indicated by the high water contents (>2 vol%) at shallow depth (50 cm) below the ground surface found everywhere, and because of the often steep dune slopes, this groundwater recharge occurs through the unsaturated zone, reaches local fresh-water springs and discharges to the lakes.

A first estimate of present groundwater recharge in the Badain Jaran Shamo, applying chloride balances, results in 0.5 to 1 mm/a. This low recharge

Table 3. *Stable isotope measurements and tritium content in groundwater samples of the Badain Jaran Shamo*

Sample	$\delta^{18}O$ (‰)	$\delta^{2}H$ (‰)	d(‰)	^{3}H (TU)
Free of tritium				
ZEGT (S)	−2.13	−40.8	−23.8	0.7 ± 0.7
LT (DW)	−2.62	−46.0	−25.1	1.1 ± 1.1
Containing tritium				
NRT	−4.87	−20.1	18.9	1.5 ± 0.7
LT (S)	9.85	2.7	−76.1	1.6 ± 0.7
SGJL (DW)	−5.39	−61.7	−18.6	2.9 ± 0.7
HBT	−3.02	−49.5	−25.4	4.6 ± 0.7
LT (IR)	8.00	−2.0	−66.0	4.7 ± 0.7
NRT-S	−5.21	−62.2	−20.5	11.4 ± 0.9
BACP	−4.08	−23.6	9.0	16.1 ± 1.2

Sampling carried out in September 2002. DW, dug well; IR, irrigation dug well; S, spring.

Table 4. *Carbon-14, $\delta^{13}C$ and tritium in groundwater samples of the Badain Jaran Shamo*

Sample	C^{14} Activity pMC (corrected)	$\Delta^{14}C$ (‰)	^{14}C Age yrBP	$\delta^{13}C$(‰)	^{3}H (TU)
LSW (DW)	98.76 ± 0.33	−18.3 ± 3.3	100 ± 25	−10.64 ± 0.19	–
HNGD	88.35 ± 0.22	−121.2 ± 2.2	995 ± 20	−11.21 ± 0.07	0.7 ± 0.7
LT	80.54 ± 0.37	−199.6 ± 3.7	1740 ± 35	−12.82 ± 0.24	0.9 ± 0.7
LSW (S)	31.25 ± 0.27	−689.3 ± 2.7	9340 ± 70	−10.36 ± 0.25	–
SGIL (S)	23.17 ± 0.17	−769.7 ± 1.7	11 750 ± 60	−4.08 ± 0.13	0.8 ± 0.7

Sampling carried out in September 2002. DW, dug well; S, spring.

rate together with slow shallow groundwater fluxes from remote areas, however, cannot compensate the evaporation losses from lakes over the last 15 000 years, hence a groundwater decline of about 1 mm/a results.

It is well known that subsurface reservoirs with low groundwater recharge rates have a long transient response through the decay of groundwater mounds that have been built up by past groundwater recharge. As the outflow of a reservoir is also associated with enlargement of the size of the catchment area, by pulling down former subsurface water divides and thus changing groundwater flow directions, transient outflow lasts longer and even becomes discontinuous. Hence groundwater flow or the decline of groundwater heads is oscillating on the one hand, but follows a general trend of decay on the other hand.

Description of transient groundwater flow

In a first approximation, the outflow (Q_t) of a subsurface reservoir with fixed boundary conditions and missing groundwater recharge can be characterized by the Maillet function:

$$Q_t = Q_o e^{-t/T} \qquad (1)$$

where Q_0 = initial discharge, T = mean turn-over time, t = time variable. The MTT (T) in this equation can be determined by means of ^{14}C in dissolved inorganic or dissolved organic carbon. As can be seen from Figure 5, a 100% discharge approaches slowly or quickly to 37% according to a high (20 000 years) or low MTT (5000 years). In terms of shallow and deep groundwater this result shows that although both belong to a hydraulic continuum, young groundwater reaches steady-state conditions much more quickly than deep groundwater, when hydrodynamic boundary conditions change. With regard to sustainable groundwater exploitation, i.e. beneficial for several generations, Figure 6 further shows that under natural boundary conditions hydraulic heads or discharge declined according to low or high MTTs within 200 years of observation and without further groundwater recharge by 85% (MTT = 100 years), 30% (MTT = 500 years) or by only 3% (MTT = 5000 years). This transient behaviour of groundwater systems in dry lands or deep aquifers is mostly neglected when developing management strategies for groundwater extraction under changing input or emergency conditions.

Typical examples of such transient groundwater behaviour are known from the Arabian peninsula (Verhagen *et al.* 1987) and north Africa (Sonntag 1985) and play a role in groundwater resources adjacent to the Jordan graben (Dead Sea) (Salameh & Udluft 1984) as well as in the Molasse basin in south Germany (Lemcke 1976). In the two latter cases the decline of pressure heads were remobilized during geological time periods by entrapped hydrocarbons, which now appear as asphalt in terminal lakes (e.g. Dead Sea), or as methane or high sulphur contents in springs along the regional discharge base in waters from two deep wells in the Molasse basin, south Germany (Lemcke 1981).

Water balance studies mostly refer to the meteorological water balance:

$$P = D + ET + \Delta S \qquad (2)$$

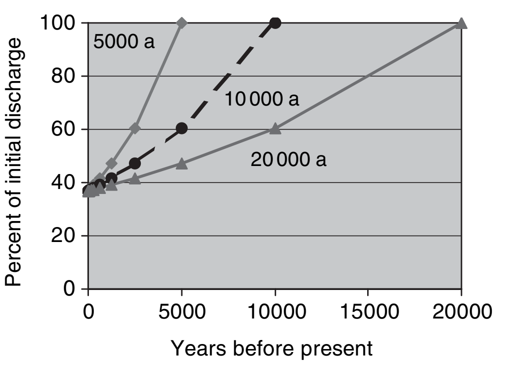

Fig. 5. Outflow of a spring according to an exponential age distribution in groundwater. Mean turn-over-times of 20 000, 10 000 and 5000 years.

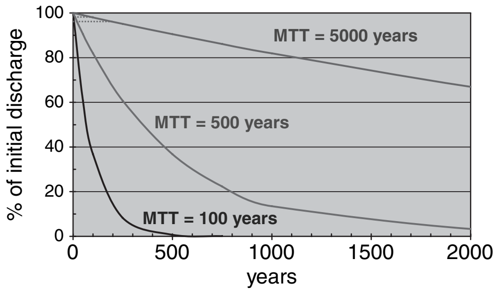

Fig. 6. Discharge or hydraulic head changes in groundwater with different mean turn-over-times.

Precipitation (*P*) and discharge (*D*) are measured over many years and discharge is further analysed by hydrograph or chemical/isotope separation methods. Potential evapotranspiration (*ET*) is estimated using meteorological data sets and reducing it to actual *ET* according to empirical transformations. Stored water (ΔS) is generally considered as belonging to the percolation zone; it includes, however, water stored over long periods of time in emptying parts of aquifers, and in transient systems contributes much more to the water balance than the small water in the percolation zone. Hence water balances over-estimate groundwater recharge in decaying, and under-estimate it in filling up systems.

As a consequence, from water balances based on Equation 2 it becomes evident that not only is a representative number of years necessary to neglect any storage in the subsurface, but it should also be ascertained that the discharge has short MTTs, to reliably calculate actual groundwater recharge. If the turn-over time is high ($>$100 years), the present discharge does not represent actual groundwater recharge and vice versa.

As an example, the Azraq springs in the Jordan Badia region discharged about 1.2 m^3/s before systematic groundwater abstraction by tube-wells started close to the spring area; today the springs do not exist any more. The springs drained an orographic catchment area of 12 700 km^2; for this catchment, with a discharge of 1.2 m^3/s, a recharge rate of 0.1 l/(s km^2) or about 3.2 mm/a is calculated. However, this calculation is actually incorrect because the water age is about 5000 years and the contribution of actual recharge was calculated by other methods to be less than 1 mm/a. This interpretation is supported by tritium analysis of water samples from shallow wells. Only a few samples out of dug wells close to wadis yielded detectable ^3H concentrations or water ages of less than 100 years; all other wells were free of tritium except recently-recharged groundwater.

Considering that discharge from the Azraq springs was three times higher 5000 years ago (Fig. 4), the same size of catchment resulted in a recharge rate of 0.3 l/(s km^2) or 10.6 mm/a for the past. A smaller size of the catchment would shift the groundwater recharge rate to an even higher number, close to the groundwater recharge under present semi-arid climate conditions, that perhaps prevailed 5000 years ago in this area.

All these examples make clear that it is quite difficult to recognize any reduction of groundwater recharge rates in semi-arid or arid areas by immediate changes of discharges, hydraulic heads or environmental isotope methods. Such changes are therefore mostly disregarded in developing exploration and exploitation strategies for these areas.

Since it is often difficult to gain reliable values on safe yield in drylands, spring discharges or well yields are mostly taken as a base for developing management strategies. However, this always leads to an over-estimate of real water availability, because hydraulic aquifer properties have no direct link to groundwater regeneration and transient systems give misleading information.

Transient flow and groundwater management

Transient groundwater flow is characteristic of low-recharge, mostly dry areas, which underwent in the past millennia more serious climate changes than other areas. Because of the small natural recharge rates, small input changes over long periods of time are finally expressed in strong responses. A detailed consideration of this transient flow, however, leads to a special aspect of groundwater management for small consumer units per square kilometre.

Transient groundwater flow in dry lands ends in oceans or terminal lakes or appears in through-flow lakes and springs. There, it is completely or partially lost by evaporation, or it becomes saline and creates groundwater contamination downstream of the lakes; hence such water losses or degradations are unproductive.

Therefore, apart from water balance studies, to better management of these unproductive, natural water losses for human or ecosystem use could also be considered. In this exercise, however, one principle must be followed: avoid groundwater over-exploitation in excess of natural losses. To reach this goal, the groundwater surface should either be removed from the influence of evaporation, thus be kept a few metres below the ground surface, or groundwater should be used before it becomes exposed to evaporation. However, groundwater exploitation should always take account of the exponential decline of groundwater discharges or hydraulic heads as characterized by the analysis of discharge or hydraulic head recession.

Approximating, for example, the lake surfaces of the Badain Jaran Shamo to about 27 km^2 and the potential evaporation from lakes as 3000 mm/a leads to unproductive water losses of 81 000 000 m^3/year; this is a minimum number because it does not include evaporation losses from groundwater resources close to the land surface.

Avoiding and using unproductive evaporation losses, indeed, is insignificant for water supply; however, it is sufficient for small communities in widespread, water-scarce areas.

Conclusions

Even with a drastic reduction of greenhouse gas emissions now, the consequences of the present trend of climate changes will continue in subsurface waters beyond the twenty-first century, because of the delayed response of all large reservoirs such as the atmosphere, oceans, ice shields and groundwater systems. All input changes in such large reservoirs will trigger a transient behaviour, as is well known in hydrogeology from desertification processes and the change of hydrogeologic systems from exorheic to endorheic.

Every assessment of future strategies of groundwater management in a world of climate changes has to consider the transient response of subsurface systems, which leads to an over- or under-estimate of the groundwater yield. This is specially valid for fragile ecosystems such as dry (arid, semi-arid) and cold lands (permafrost) and should also be debated in connection with deep groundwater extraction everywhere (Ghergut *et al.* 2001).

Since emerging groundwater in dry lands is lost by unproductive evaporation, on the one hand, and evaporation increases groundwater salinity, on the other hand, it is also recommended that these resources are explored and exploited in accordance with the transient outflow characteristics. In this way over-exploitation of the groundwater resource exceeding ‘natural groundwater stress’ can be excluded.

Hence, the re-assessment of groundwater resources with respect to climate changes is not only a question of present input/output relations but also of past, transient input responses.

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Application of isotope techniques in groundwater recharge studies in arid western Rajasthan, India: some case studies

U. K. SINHA & S. V. NAVADA

Isotope Hydrology Section, Isotope Applications Division, Bhabha Atomic Research Centre, Trombay, Mumbai - 400 085, India (e-mail: udayks@apsara.barc.ernet.in)

Abstract: Environmental isotope studies of groundwaters carried out in the arid region of Rajasthan, India, suggest that recharge to shallow groundwater occurs as a result of direct infiltration of precipitation and/or through river channels during episodic floods. In many parts of Rajasthan deep fresh groundwater is available, which was recharged in the past when the climatic conditions were more favorable than at present. However, negligible modern recharge and over-exploitation of groundwater resources have led to the deterioration of groundwater quality in the Bhadka–Bheemda area of Barmer district. Buried river courses (palaeochannels) could be potential sources of groundwater since radiocarbon measurements of groundwater along palaeochannels in Jaisalmer district revealed that it was old water recharged a few thousand years ago.

There is evidence that the climate has changed repeatedly during the Earth's history and, in the past, the present deserts were not always so arid, so devoid of life and vegetation (Singh *et al.* 1974). As groundwater is the only available source of water in most of the desert regions, exploitation of these resources could help to solve the water requirement of these regions. Sound management of these groundwater resources requires data on the source, dynamics and quality of groundwater, which could be obtained using isotope techniques, hydrochemical and hydrogeological studies. Isotope techniques have successfully been used by many investigators to study some of the water resource problems in arid regions (Fontes & Edmunds 1989). A basic problem in arid areas, which often cannot be solved easily with conventional hydrological techniques, is to determine whether a given body of groundwater is actively recharged, i.e. whether it is a renewable resource. Environmental tritium and radiocarbon (^{14}C) can be used to identify recharge pertaining to modern-day precipitation or precipitation in the past (up to 45 000 years) (Sukhija & Rama 1973; Sukhija *et al.* 1996; Allison & Hughes 1974, 1978; Dincer *et al.* 1974). Several hydrological studies carried out in arid zones (mostly in the Sahara) using isotope techniques have shown that the deep groundwater (generally $>$200 m) is generally old, with radiocarbon ages $>$20 000 years BP (Sonntag *et al.* 1979; Mabrook & Abdel Shafi 1977; Abdelkader & Zuppi 1999). This was supported by variations of the stable isotopic composition (^{2}H and ^{18}O) in groundwater.

Study area

This paper presents four different studies from arid regions intended to highlight applications of isotopic studies in arriving at specific hydrologic understanding. These studies employed environmentally stable isotopes ^{2}H, ^{18}O, ^{13}C and radioactive isotopes ^{3}H and ^{14}C.

The Thar Desert extends from the western side of the Aravalli mountain ranges in India up to the limit of the Indus valley in Pakistan (Fig. 1). It covers about 60% of the area of Rajasthan state in the northwestern part of India. Having *c.* 38% of the state's population with a density of 84 persons/km^2, this is one of the most populated desert region of the world. The constantly increasing human and livestock population exerts tremendous stress on the available natural groundwater resources.

The land is characterized by sand dunes with interdunal plains in the north, west and south, and alluvium in the central and eastern parts. Streams are very few, ephemeral in nature and confined mostly to the rocky parts of the desert; the most prominent is the Luni river in the SW region. The average annual precipitation is low (100 to 300 mm per year) and erratic with frequents droughts. The average annual maximum and minimum temperatures are 45°C and 8°C, respectively.

The main source of water in the area is groundwater. The region, with mean annual potential evapotranspiration ranging from 180 to 200 cm, faces an acute shortage of potable water during the summer months. At many places the groundwater is brackish or saline. Efforts are being made

From: Dragoni, W. & Sukhija, B. S. (eds) *Climate Change and Groundwater*. Geological Society, London, Special Publications, **288**, 121–135.
DOI: 10.1144/SP288.10 0305-8719/08/$15.00

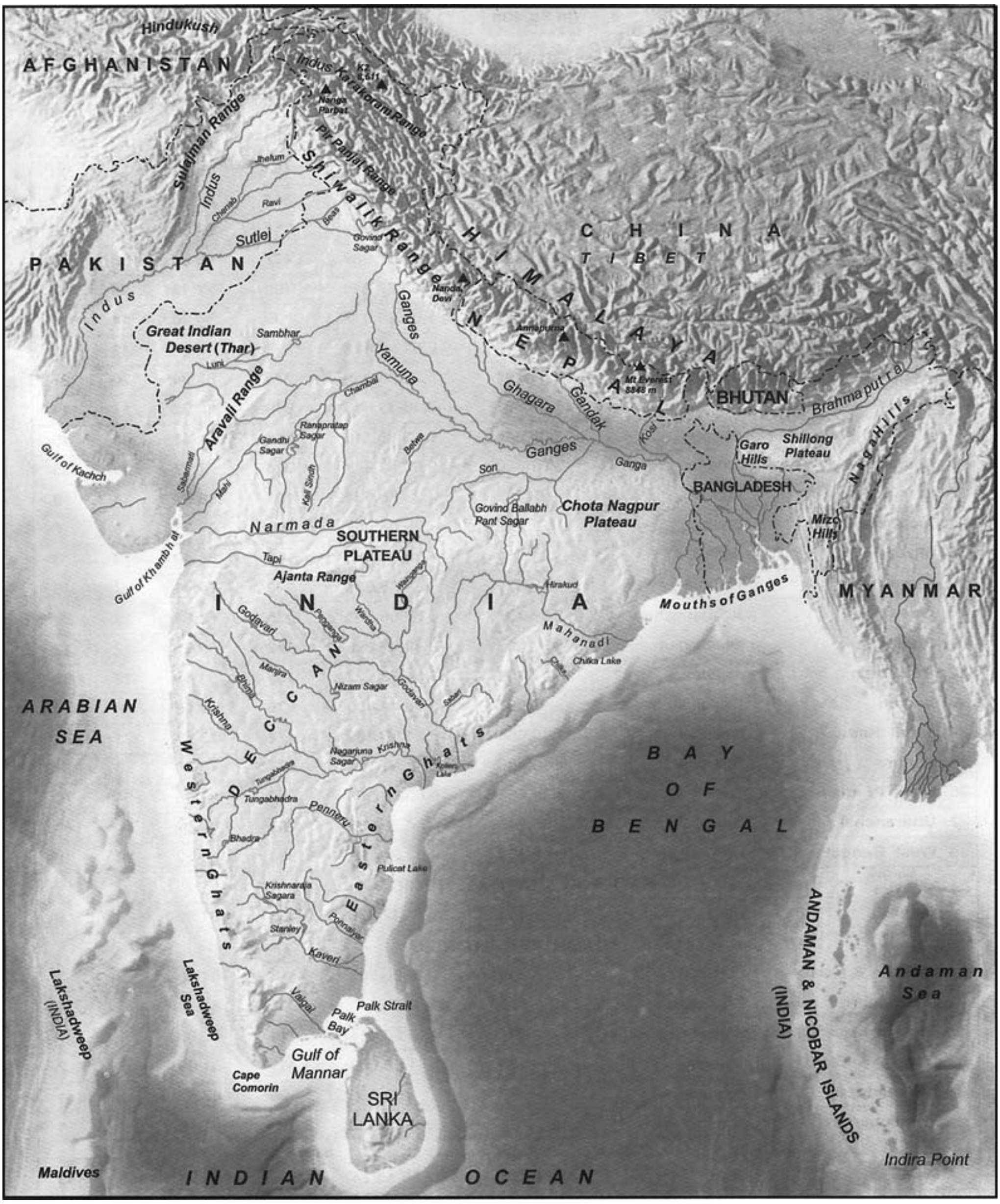

Fig. 1. Physical map of India showing the Thar Desert.

by the state groundwater department to study known groundwater resources and explore potential ones in the region.

The possible groundwater recharge processes in arid areas like Rajasthan are: (a) direct recharge of precipitation through the unsaturated zone; (b) indirect recharge through river channels from flash floods or from irrigation canals etc. and (c) palaeoclimatic recharge during humid episodes in the past.

Groundwater recharge studies

Direct recharge of precipitation to the groundwater was studied during 1982–1984 and 1990–1992 in Jodhpur and Barmer districts of western Rajasthan using artificial ^{3}H tracer as tritiated water (HTO) in the unsaturated zone (Shivanna *et al.* 1994). The results showed negligible groundwater recharge due to low rainfall (*c*. 200–300 mm) during the study periods. Since the artificial ^{3}H tracer method determines recharge for the study period only, other methods such as environmental ^{3}H, ^{36}Cl and chloride methods, need to be employed to determine long-term recharge (Gaye & Edmunds 1996; Cook *et al.* 1994). In the Barmer district of Rajasthan, mean annual recharge of *c*. 14 mm was obtained using environmental chloride and natural tritium methods (Navada *et al.* 2001).

In arid areas, indirect recharge through wadis (river channels) could be an important mechanism for groundwater recharge (Darling *et al.* 1987). Such an observation was made in the Jalore area of western Rajasthan (Navada *et al.* 1993) using isotope studies of groundwater, which showed that shallow groundwater near Sukri river is recharged from river channels during flash floods. Buried river courses (palaeochannels) are important as they have good groundwater potential. In Jaisalmer district of Rajasthan, a Holocene age freshwater channel superimposed on older groundwaters was identified. Direct head water connection to the groundwater in the study area from present-day Himalayan sources appeared to be remote (Nair *et al.* 1999).

An environmental isotope study in the Bhadka–Bheemda area in Barmer district, Rajasthan (Navada *et al.* 1996), showed that the deep groundwater (depth >100 m) has depleted δ^2H and δ^{18}O compared to the shallow groundwater. The deep groundwater has negligible ^{3}H and the ^{14}C model (Pearson) ages range from 4000 to 9500 years BP, hence it is palaeowater recharged in the past. Palaeontological studies have proposed the following dry and wet periods in Rajasthan (Das 1968):

10 000–6000 years BP	moderately humid, rainfall greater than present
5000–3000 years BP	humid period
3000–1100 years BP	low rainfall
1100 BP–Present	dry conditions.

Hence the observed ^{14}C ages of groundwater in the Bhadka–Bheemda area represent recharge during humid periods.

Over-exploitation of groundwater resources in arid regions with limited water resources reveals the adverse effects of rapid lowering of the water table and deterioration of water quality. For example, in Bikaner town in Rajasthan the aquifers are being used extensively, resulting in declining groundwater levels. An environmental isotope (^{18}O, ^{2}H and ^{3}H) study (Navada 1988) showed mixing of shallow and deep zone waters due to heavy exploitation of groundwater in the area. The effect of over-exploitation of groundwater was also studied in the Bhadka–Bheemda area (Nair *et al.* 2007) in Barmer district, and the limestone belt of the Jodhpur–Nagaur district (Nair *et al.* 1993) using isotope techniques.

Environmental isotope measurements

Measurement of ^{2}H, ^{18}O and ^{13}C

For measurement of δ^2H in water, a 5-μl water sample was reduced to H_2 gas using BDH zinc shots in a closed evacuated tube at 450°C (Coleman *et al.* 1982). The H_2 gas produced was analysed using a 602E mass spectrometer (VG Isogas, UK). The δ^2H values reported were with respect to SMOW (standard mean ocean water) standard. The precision of measurement was 1‰ (2σ). For δ^{18}O measurement of the water, a 5-ml sample was taken in a flask and equilibrated with CO_2 gas at 25°C for 8 hours (Epstein & Mayeda 1953). The equilibrated CO_2 was analysed using a 602E Auto mass spectrometer (VG Isogas, UK). The δ^{18}O was reported with respect to SMOW. For δ^{13}C analysis, the bicarbonates/carbonates in water were precipitated as barium carbonate by using saturated barium chloride solution in alkaline medium. The carbonate sample was treated with 100% phosphoric acid (McCrea 1950) and the carbon dioxide gas released was analysed using the 602E Auto mass spectrometer. The δ^{13}C value was reported with respect to PDB (belemnite of Pee Dee formation). The precision of measurement of both δ^{18}O and δ^{13}C was 0.2‰ (2σ).

Measurement ^{3}H

Measurement of ^{3}H of the water sample was carried out by electrolytic enrichment followed by liquid scintillation counting (Taylor 1981).

A 250-g sample of distilled water was transferred into an electrolytic cell comprising mild steel perforated cathodes and stainless steel anode. A 0.5-g sample of Na_2O_2 was added to 250 g of distilled water (2 g/l). A total of 708 ampere hour charge was passed, which reduced the volume from 250 g to 12 g. The temperature of the cells was maintained at *c*. 5 ± 2°C throughout the electrolysis.

Packard (now Perkin Elmer) Hisafe Scintillator was mixed with electrolysed and neutralized water samples (8-ml water sample + 14-ml scintillator) in a 24-ml high-density polythene vial and counted in a LKB 1200 Quantulus liquid scintillation counter.

The minimum detection limit obtained was 0.5 TU.

Measurement of ^{14}C

Measurement of ^{14}C was carried out by direct absorption of carbon dioxide (obtained from the field precipitated carbonate sample) into an absorber–scintillator mixture (11 ml + 11 ml carbasorb and Permaflour V or its equivalent) followed by liquid scintillation counting (Nair *et al.* 1994). Carbon dioxide obtained from oxalic acid (secondary standard calibrated with respect to primary NIST standard) was also absorbed in the same quantity of absorber and scintillator mixture. Samples and standards were counted in a LKB 1200 liquid scintillation counter and each sample was counted for 1000 minutes.

Knowing the standard, background and sample count rate, ^{14}C activity could be calculated in per cent modern carbon (pMC), where 100 pMC = 13.6 dpm/g of carbon. Once corrected and normalized sample activity (A_{SN}) and normalized standard activity (A_{ON}) are known, the uncorrected age of the sample can be calculated using the equation:

$$t = 8268^{*} \ln A_{ON}/A_{SN}$$

Minimum detection limit was 1 pMC, which corresponds to a maximum measurable age of 38 000 years BP (Nair *et al.* 1994). Model ages can be estimated by calculating A_{ON} values using different model equations.

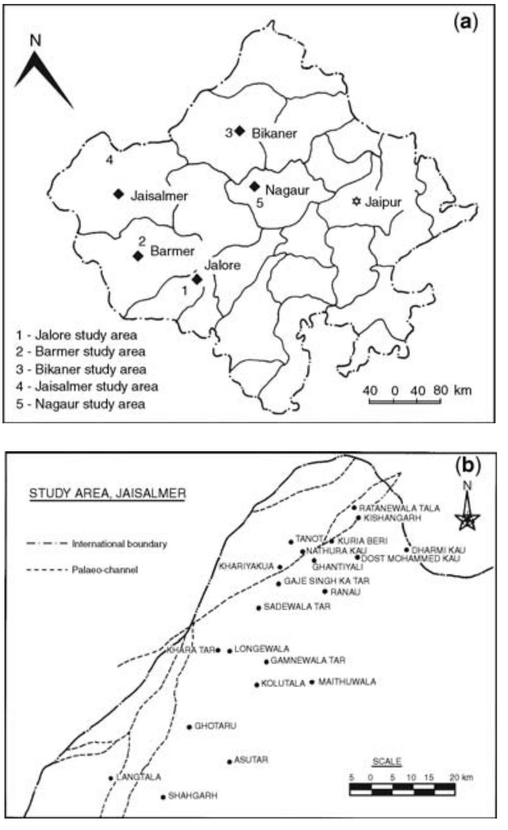

Fig. 2. (**a**) Map of Rajasthan. (**b**) Location map of the study area.

Environmental isotope studies along an 'identified' palaeochannel

Interpretation of satellite imagery of the western parts of the Jaisalmer district (western Rajasthan) 'revealed' the buried course of a river orientated in the NE–SW direction (Kar 1986; Bakliwal & Grover 1988) (Fig. 2a). In spite of the highly arid condition of the region, comparatively good quality groundwater is available along the course below 30 m depth. The aquifer consists of medium to fine sand with very little clay. A few dug wells do not dry up even in summer and the tube wells do not show reduction in water table, even after extensive utilization for human as well as livestock consumption. Groundwater away from this course is saline. The course is thought to belong to the legendary river Saraswati of Himalayan origin, mentioned in many early literary works and known to have existed 3000 years BP (Ghose *et al.* 1979; Valdiya 1996).

Quaternary alluvium composed of gravel, sand, silt and clay in varying proportions is the most widespread lithological unit. The thickness of this unit extends to a depth of 200 m. Much of the area occupied by this formation is either unproductive or contains saline groundwater. The central part of Jaisalmer district is occupied by Cenozoic and Mesozoic sandstone. Another widespread lithological unit encompassing about 52% area of the potential zone is Lathi sandstone (Department of Science and Technology 1999). It is white, yellow or reddish brown, poorly to moderately lithified medium-grained sandstone interspersed with silt, stone and shale. The upper part contains considerable fine-grained material. This unit attains maximum thickness (250 m) near Jaisalmer town.

To determine the origin and age of groundwater along the buried course of the river (palaeochannel), a study was carried out using environmental isotope. Samples were collected in the year 1998 from existing dug wells and tube wells in the area for analysis of δ^2H, $\delta^{18}O$, ^3H, $\delta^{13}C$ and ^{14}C as well as for chemistry. Dug wells were quite deep with water levels below 30 m or more. Tube wells are cased with screens below 60 m depth. Figure 2b shows sample locations. Both dug wells and tube wells show similar characteristics, being evolved towards Na-Cl type (Fig. 3). A plot of δ^2H versus $\delta^{18}O$ for the samples is shown in Figure 4. It is seen that the stable isotopic values of a few groundwater samples from dug wells are enriched with respect to tube well samples. This could be due to the exposure of open water bodies to evaporation in an arid environment. The linear regression line of dug well samples has an equation $\delta^2H = 2.18\ \delta^{18}O - 28.7$, $r^2 = 0.39$, $n = 12$. The δ^2H–$\delta^{18}O$ relationship (leaving aside the three dug well samples with enriched stable isotope values) would support the model of Allison *et al.* (1984), wherein direct infiltration during high rainfall event mixes with evaporated soil moisture and the mixed parcel eventually reaches the water table. The stable isotope (δ^2H and $\delta^{18}O$) values clearly show that the deep and shallow zones of the aquifer are generally not interconnected. Since no lithological separation exists between the deep and shallow zones, the deep aquifer receives a small fraction of vertical recharge from precipitation; this is spatially variable in accordance with the coarse grain content of the upper layer, following Allison *et al.* (1984) model. The regression line is parallel to the meteoric water line, which indicates that this recharge persisted over a long time. Both dug well and tube well samples are enriched compared to present-day Himalayan rivers ($\delta^{18}O$: −11‰ to −9‰), which indicates that palaeochannel waters are not in dynamic contact with present-day Himalayan rivers. Figure 5 indicates that the tube well samples have negligible tritium, indicating the absence of modern recharge. However, a few dug wells do show small components of modern recharge.

Most of the dug well and tube well samples along the palaeochannel have electrical conductivity (EC) less than 4000 μS/cm (Table 1), while groundwater away from this channel is highly saline, with conductivity more than 10 000 μS/cm. This indicates that salinity increases as one moves away from the channel.

The ^{14}C age using an empirical model (Table 1) of the dug well samples is less than 5000 years BP (Fig. 6), which shows that they are old groundwaters, but no flow pattern is discernible from the results. The ^{14}C age of tube well samples is more than 5000 years BP. The maximum age is from the saline pocket, Sadewala. There is a trend of an

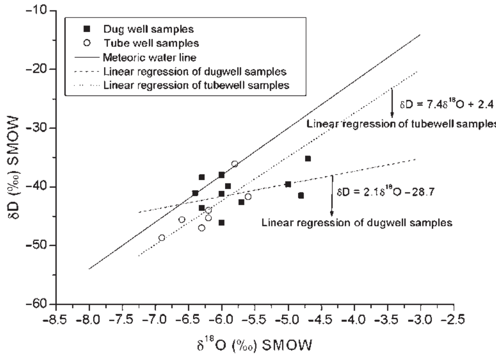

Fig. 4. Plot of $\delta^{18}O$ versus δ^2H for Jaisalmer samples.

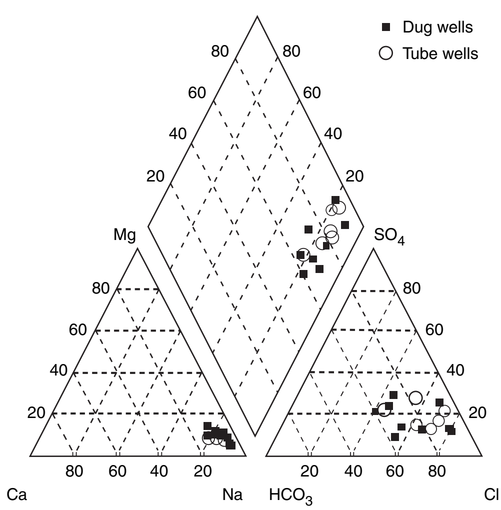

Fig. 3. Piper trilinear plot of Jaisalmer groundwater samples.

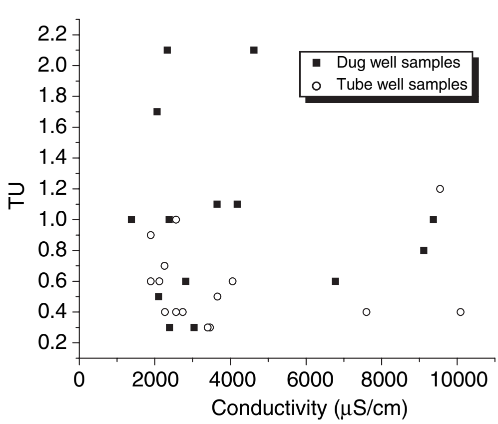

Fig. 5. Plot of conductivity versus tritium content (TU) of Jaisalmer samples.

Table 1. *Isotopic and conductivity data of palaeochannel samples*

ID no.	Location	Well type*	EC (µs/cm)	$^3H \pm 1\sigma$ (TU)	$^{14}C \pm 1\sigma$ (pMC)	^{13}C (‰)	^{18}O (‰)	^2H (‰)	^{14}C (model age) 91.9 pMC as A_0
D1	Dharmikua	DW	2330	2.1 ± 0.3	79.5 ± 2.2	−9.6	−7.5	−47.1	1198
T1	Kishengarh	TW	3460	0.3 ± 0.1	47.3 ± 1.4	−5.7	−5.6	−41.7	5492
D2	Kishengarh	DW	4180	1.1 ± 0.2	91.9 ± 1.7	−10.7	−6	−40.9	0
D3	Kurriaberi	DW	2100	0.5 ± 0.2	58.8 ± 1.6	−8.3	−5.7	−42.6	3692
D4	Nathurakua	DW	3040	0.3 ± 0.2	69.3 ± 1.8	−7.9	−6.3	−38.4	2334
T2	Ghantiyali	TW	3660	0.5 ± 0.2	31.2 ± 1.2	−4.0	−6.6	−45.6	8932
D5	Ghantiyali	DW	2820	0.6 ± 0.2	54.9 ± 1.5	−7.1	−6.0	−41.2	4260
D6	Khairakua	DW	8900				−4.8	−41.5	
D7	Gajesingh Ka Tar	DW	4620	2.1 ± 0.3	64.9 ± 1.9	−7.7	−4.7	−35.2	2876
T3	Ranau	TW	1890	0.6 ± 0.2	48.8 ± 1.5	−7.4	−6.2	−45.3	5233
D8	Ranau	DW	2060	1.7 ± 0.3			−6.0	−46.1	
T4	Sadewala	TW	7600	0.4 ± 0.2	6.6 ± 0.9	−7.7	−3.4	−26.3	21775
D9	Sadewala	DW	9120	0.8 ± 0.2			−6.3	−43.6	
T5	Loungewala	TW	2740	0.4 ± 0.2	10.4 ± 0.9	−5.6	−6.2	−44.0	18015
D10	Loungewala	DW	9370	1.0 ± 0.2			−5.9	−39.9	
T6	Gumnewala	TW	4060	0.6 ± 0.2			−6.1	−30.0	
T7	Ghotaru	TW	2270	0.4 ± 0.2	20.7 ± 1.0	−7.3	−6.9	−48.7	12324
D12	Ghotaru	DW	3650	1.1 ± 0.2	62.7 ± 1.6		−6.4	−41.1	3161
T8	Asutar	TW	2560	0.4 ± 0.2	36.1 ± 1.3	−7.5	−6.3	−47.0	7726
D13	Asutar	DW	2390	0.3 ± 0.1					
D14	Langtala	HP	3400	0.3 ± 0.1	68.6 ± 2.0	−6.2	−5.0	−39.6	2418
D15	Langtala	DW	2380	1 ± 0.2	64.8 ± 1.7	−6.9	−6.0	−46.1	2889
T9	Shahgarh	TW	10090	0.4 ± 0.2			−6.0	−38	
D16	Ratnewala	DW	10330			−4.7			
D17	Dost Mohammad Ka Kuan	DW	1380	1	49.7 ± 1.5	−1.2			5082
D18	Mituwala	DW	6780	0.6	57.9 ± 1.7	−8.5	−5.8	−36.1	3820

*DW, dug well; HP, hand-pump well; TW, tube well.

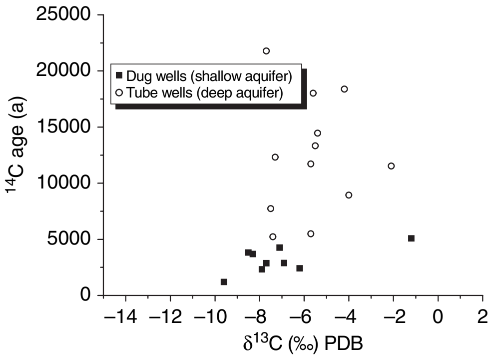

Fig. 6. Plot of $\delta^{13}C$ versus ^{14}C age of Jaisalmer samples.

increase in the apparent ^{14}C age for groundwaters from Kishengarh to Loungewale, along the buried course. From the relative radiocarbon ages, a groundwater velocity of about 5 m/a may be inferred, which is a normal value expected under similar desert conditions. The $\delta^{13}C$ content of shallow groundwater (dug wells) is depleted compared to other tube well samples.

Isotopic values indicate that this channel is not connected with the area across the border covering the dry Ghaggar River bed (Geyh & Ploethner 1995). The results of the isotope study indicate that direct headwater connection to the groundwaters in the study area from present-day Himalayan sources is remote. The stable isotope (^2H, ^{18}O), ^3H and ^{14}C values indicate that they are palaeowaters with negligible modern recharge component. Deep and shallow groundwaters are generally not interconnected, but some recharge to the deeper zones is possible. This indicates accumulation of groundwater over a 10^3–10^4 year time scale. The water quality difference may be related to small variations in the coarse grain fractions, which may be more along the paleochannel permitting slightly higher recharge and less evaporative enrichment prior to recharge.

The authors feel that the palaeochannel contains comparatively fresh water, which can be used for drinking and irrigation processes. Since the channel is presently not connected with any of the Himalayan rivers, over-exploitation may lead to depletion of groundwater resources.

Recharge studies in Jalore, Rajasthan

Jalore district (Fig. 7) is situated adjacent to Barmer in the SW part of Rajasthan. An environmental isotope investigation was undertaken to understand the groundwater recharge mechanism at the study area (Fig. 8). The region receives a mean annual rainfall of *c*. 380 mm and is drained by Sukri river, a tributary of the Luni river system, which is ephemeral in nature. The younger alluvium which is present along the river course is unconsolidated to semi-consolidated coarse to fine sand and gravel. Older alluvium is of sub-Recent to Pleistocene age. Subsurface geology (Navada *et al.* 1993) reveals the presence of a fault in a NE–SW direction along the Sukri river.

Younger and older Quaternary alluvium is the major hydrogeological unit. The younger alluvium comprises unconsolidated sediments – sand, silt and clays – in varying proportions and is found in floodplains. Older alluvium includes piedmont alluvial sediments, unsorted rock fragments, gravel, and sand with higher clay percentage; and higher terrace deposits – upper sandy loam and kankar layers and lower fine to medium sand, gravel and silt. In the central part along Sukri river in the east–west direction, there is a thick clay horizon on the northern bank and medium to coarse sand and gravel interbedded with clay lenses on the southern bank, where coarse sediments increases towards the east.

A number of samples from shallow (<50 m) and deep wells were collected and analysed for environmental isotope as well as chemical analyses.

Deep groundwater near the river course is generally fresh and Na-HCO_3 type. Shallow and deep groundwater, away from the river course, are brackish and are Na-Cl type. Figure 9 shows the δ^2H versus $\delta^{18}O$ plot for the samples. The samples appear to fall into three groups. Shallow groundwaters along the river course are enriched in stable isotopes and fall into group C. They show a typical evaporation trend and have high ^3H contents (5 to 20 TU) indicating a modern recharge component. This group also contains two deep well samples, including an artesian well, which are brackish, have negligible ^3H and low ^{14}C values. Shallow groundwaters, which are located away from the river course in the western and southwestern parts, have comparatively depleted stable isotope values and fall into group B. These samples have ^3H concentrations ranging from 1 to 4 TU. A few fresh, deep well samples, from near the river courses, also have similar stable isotope (^2H, ^{18}O) values and measurable tritium contents, indicating that they have modern recharge components. Group A contains shallow well samples that are the most depleted in stable isotope values. A few samples from the deeper zone, which are brackish and have negligible ^3H content, indicating absence of any recent recharge, are also included in this group.

It is observed that samples with depleted ^{18}O values have low ^3H contents and vice versa. It is also seen that old waters with low ^3H values have

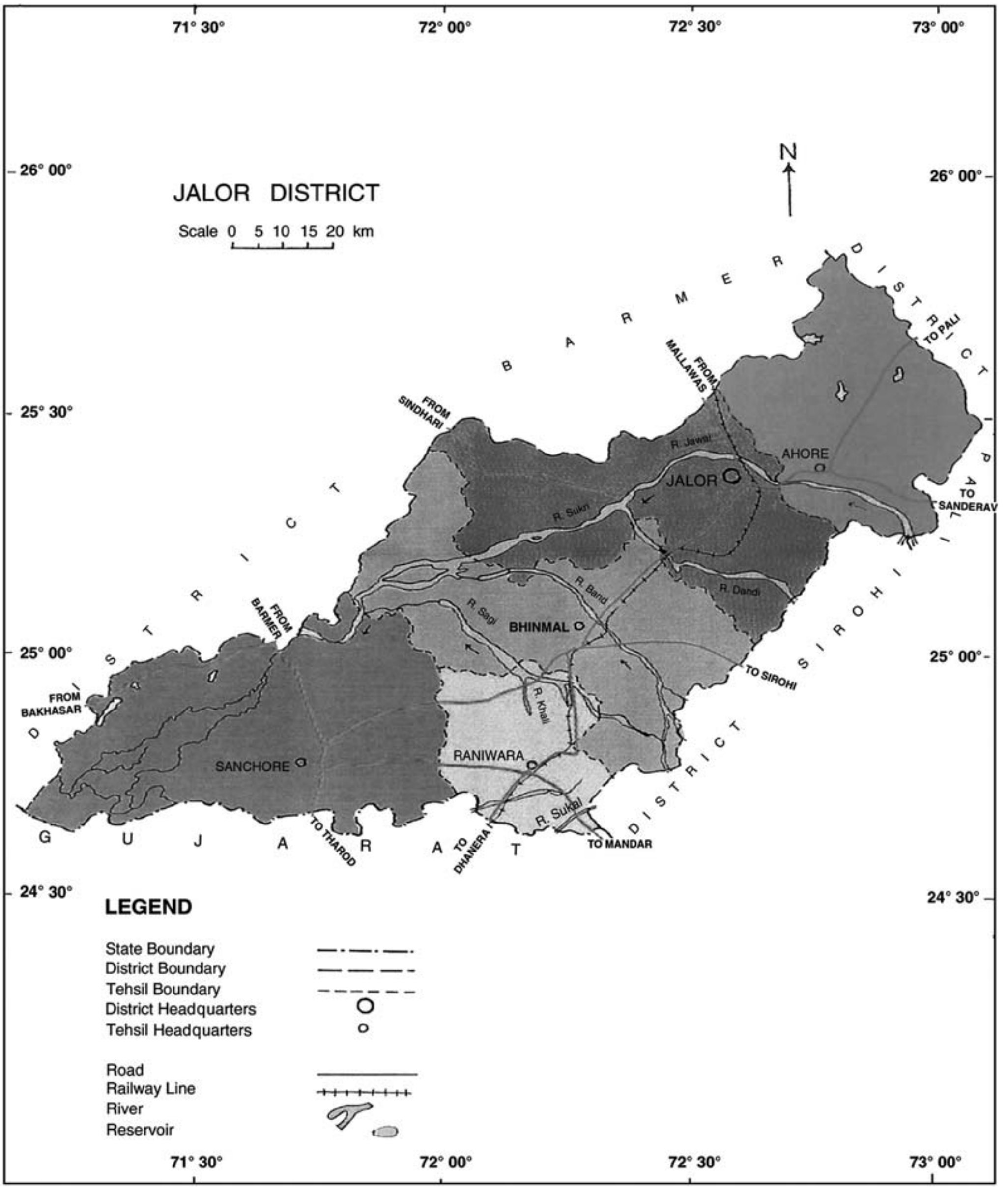

Fig. 7. Map of Jalore district.

high chloride contents (800 to 1000 ppm), whereas recent waters with high ^3H values have low chloride content. This indicates that the groundwater near the river course is fresh water with enriched stable isotopic composition and high tritium value, showing the presence of modern recharge. The groundwater away from the river course is brackish, has depleted stable isotopes (^2H, ^{18}O) and low ^3H contents and thus represents older water. Some interconnection between the shallow and deeper aquifer along the course is indicated by the stable isotopes (^2H, ^{18}O), ^3H and ^{14}C values (Navada *et al.* 1993).

The study indicates that the shallow aquifer receives recharge through vertical infiltration of local rainwater through the soil column, having spatially variable vertical permeability. Less permeable zones have less recharge, higher salinity

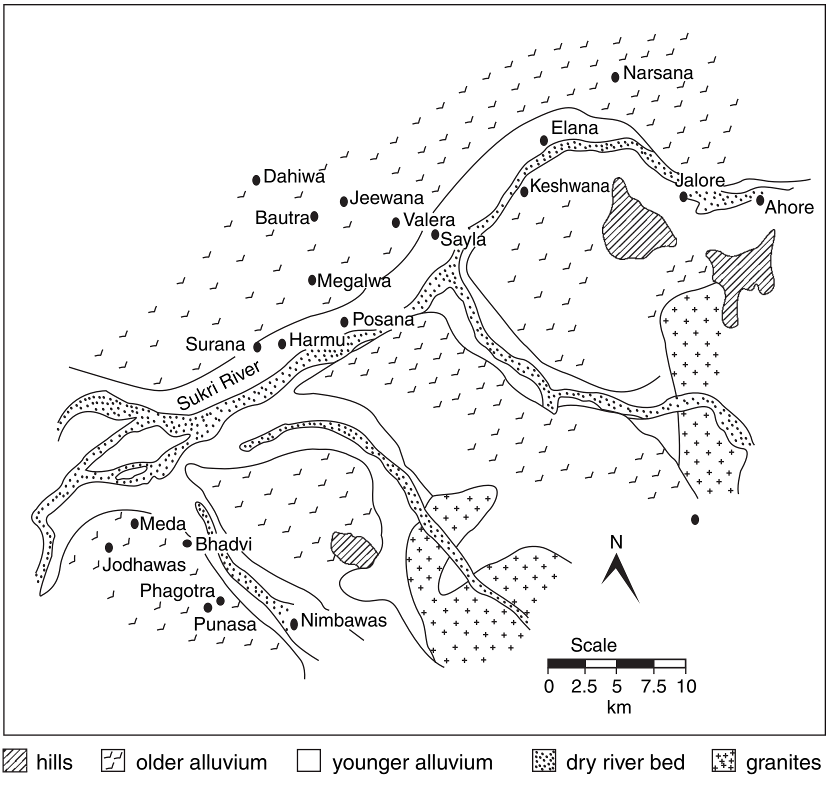

Fig. 8. Sample location map and geological formation of the study area.

of groundwater and low ^{3}H. Stable isotope values reflect the evaporation signal overprinted on the primary isotopic signal of the rainfall at different times in the past.

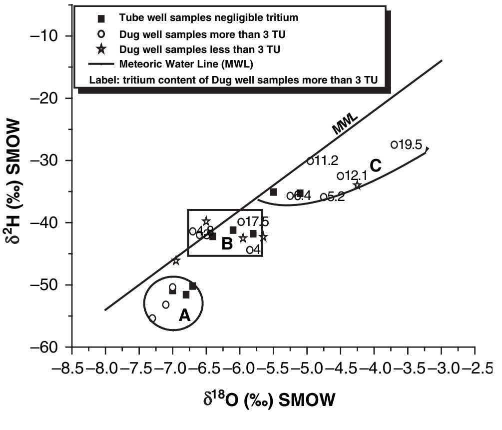

Fig. 9. Plot of δ^2H versus δ^{18}O for dug well and tube well samples with label of high tritium content of dug well samples.

Isotope investigation of dynamic changes due to long-term exploitation of groundwater

The Bhadka–Bheemda area in Barmer district is situated in the SW part of Rajasthan. At the request of the Groundwater Department, Rajasthan, environmental isotope investigations were undertaken in the 1980s to understand the recharge processes and dynamics of groundwater in these areas. A fresh look at the isotopic and chemical characteristics of groundwater was attempted subsequently in order to understand their response to large-scale exploitation of groundwater.

Figure 10 shows the geology of the area with sample locations. It receives a mean annual rainfall of 280 mm and the central portion is mainly underlain with Tertiary sediments. The northern part is covered by Lathi sandstone (Jurassic age). The other parts are covered by the Malani suite of igneous rocks (Precambrian–Lower Palaeozoic) (Groundwater Department Jodhpur 1983). Fresh groundwater forms a basin in the central portion. Here, the shallow aquifer is under phreatic conditions while the deeper aquifer is under

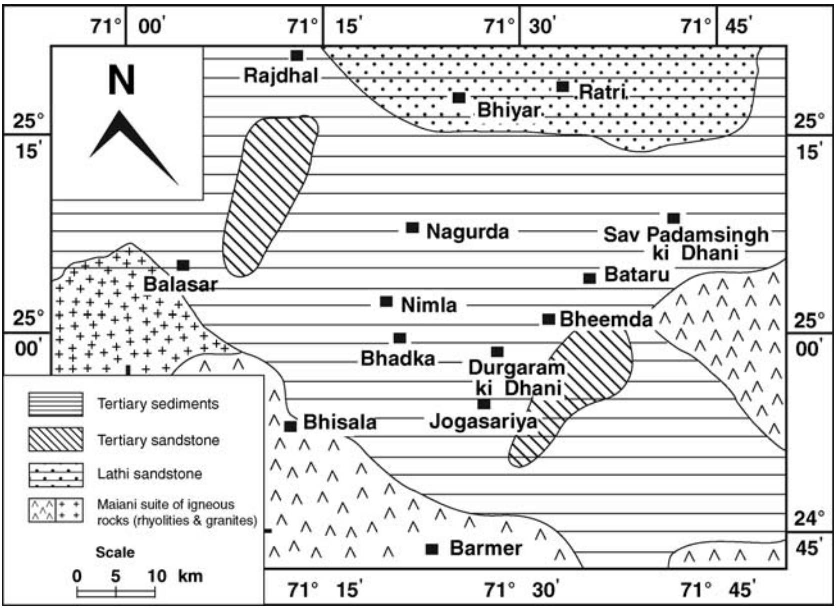

Fig. 10. Map of Bhadka–Bheemda area in Barmer district.

semi-confined or confined conditions. The shallow groundwater is generally brackish. These waters, along with the deep brackish groundwaters, are of Na-Cl type. The deep groundwater, which is relatively fresh, is $Na\text{-}HCO_3$ type, which is being exploited at a rate of over 20 million cubic metres per year.

A set of samples, collected in 1987 as analysed for ^{2}H, ^{18}O, ^{3}H, ^{14}C and chemistry. It was observed that most of the deep (>100 m) fresh groundwater is isotopically depleted and plots along the meteoric water line on the $\delta^2H-\delta^{18}O$ plot (Fig. 11). This group also contains a brackish sample from Bhadka as well as a freshwater sample from Rajdhal from the northern dunal area. Negligible ^{3}H and model ages (Pearson's model) ranging from 4000 to 9500 years BP (Navada *et al.* 1996) indicate that these are palaeowaters, probably recharged during a cooler and/or pluvial phase in the Holocene. The shallow and deep groundwaters (including samples from Lathi sandstone), which are brackish, fall along an evaporation line. Some samples in this group have measurable tritium indicating some components of modern recharge. Similar isotopic and chemical characteristics of shallow as well as deep samples at some locations indicate aquifer interconnection (e.g. Durgaram ki Dhani).

A second set of samples was collected from the area in 1996, and a third set in March 2000. They have been analysed for δ^2H, $\delta^{18}O$, $\delta^{13}C$, ^{3}H and ^{14}C as well as for chemistry. It was observed that most of the dug wells in the study area, sampled earlier, have been abandoned or have dried up. A few existing dug well samples were collected in 1996 for ^{14}C and tritium measurement. In 2000, due to unavailability of dug wells, only tube well samples were collected. Many deep wells, which were sampled earlier, have been abandoned due to either failure of the wells or deterioration of water quality. In such cases, other nearby wells have been sampled.

Figure 12a shows electrical conductivities of tube well samples for all three sampling campaigns. It can be seen that the groundwater becomes more brackish with time except at Bhadka and Ratari where a reverse trend has been noticed. Deviation from the general trend may be due to sampling of a nearby well (in the case of failure of the previous well), which may not represent the groundwater of the failed well. Figure 12b clearly shows evaporative enrichment of deep groundwater with time. Exceptions are Ratari which is situated in the Lathi sandstone, and Rajdhal. Also, a slight increase in ^{14}C values, from 58 pMC in 1996 to 61 pMC during 2000, is an indication of the contribution of younger water in the Ratari area. Figure 11 is the $\delta^2H-\delta^{18}O$ plot for all the samples collected during the three sampling campaigns in 1987, 1996 and 2000. Rainwater samples (monthly average of year 2000) show depleted values. The general trend of deep and shallow groundwater

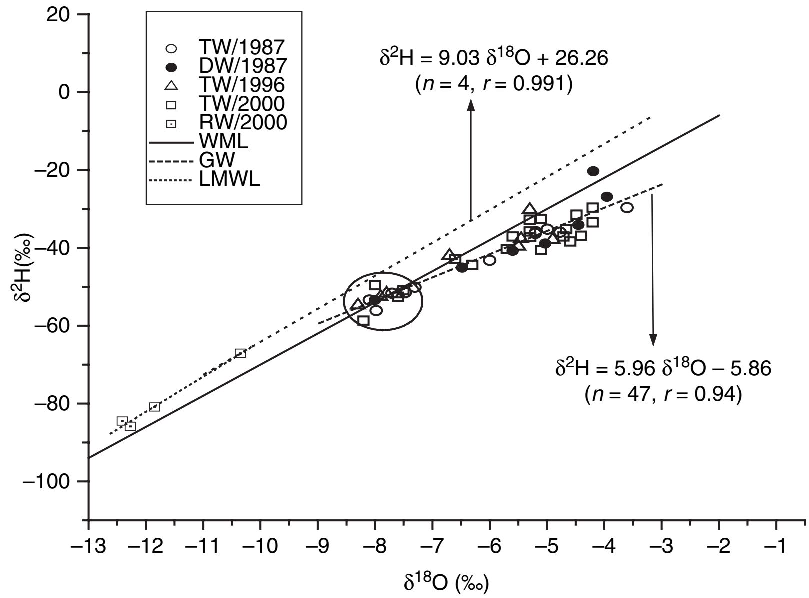

Fig. 11. Plot of δ^2H versus δ^{18}O for all samples collected from Barmer area.

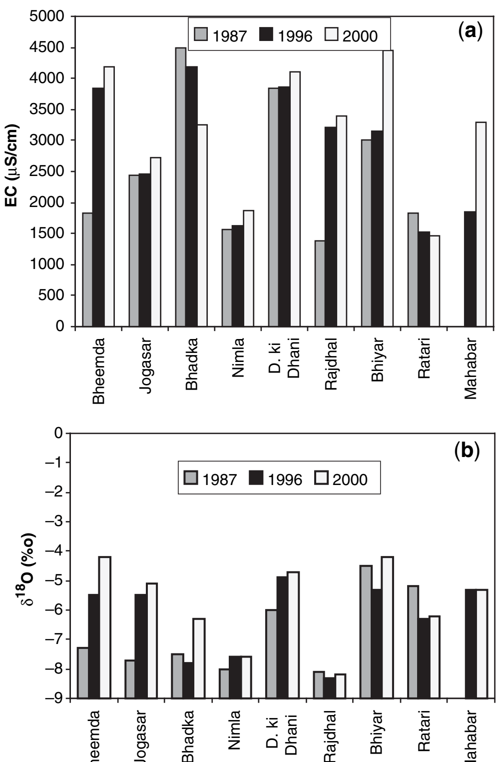

Fig. 12. Bar graphs showing electrical conductivity (**a**) and δ^{18}O values; (**b**) of samples collected during 1987, 1996 and 2000.

apparently remains the same. The distribution of stable isotopic data indicates the possibility of dispersed recharge through vertical infiltration with some evaporation prior to recharge during different climatic phases during the post 12 ka. However, from the δ^{18}O versus ^{14}C plot (Fig. 13), variations in ^{14}C values have been observed at Bheemda, Bhiyar, Nimla, Jogasar and Durgaram ki Dhani, and negligible variations at Bhadka, Rajdhal and Ratri. Although variations in the ^{14}C values are seen at Bheemda, Bhiyar and Jogasar, a general trend is absent which indicates that different wells have been sampled at different times. Although there is a general tendency for samples to show lower ^{14}C values, the variations are not significant at most locations. Small variations in their ^{14}C content are indicative of lateral flow within an aquifer that has spatial variations in water quality and isotopic character inherited from vertical infiltration. Though mining of groundwater is apparent from the declining groundwater table, it is well reflected in the isotopic and chemical characteristics of the groundwater mixing with isotopically enriched saline waters. With the rate of abstraction increasing, the quality of water is bound to suffer further in the future, indicating a need for proper planning of groundwater resource management.

Groundwater recharge in a limestone belt

The limestone belt of Jodhpur–Nagaur district of Rajasthan extends from Bilara (SE of Jodhpur) to

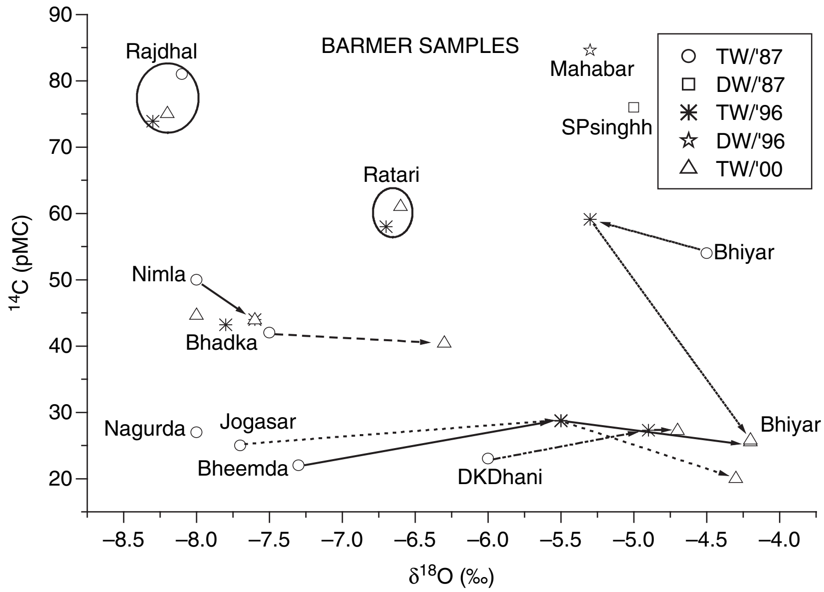

Fig. 13. Plot of ^{14}C versus δ^{18}O for all samples collected from Barmer area.

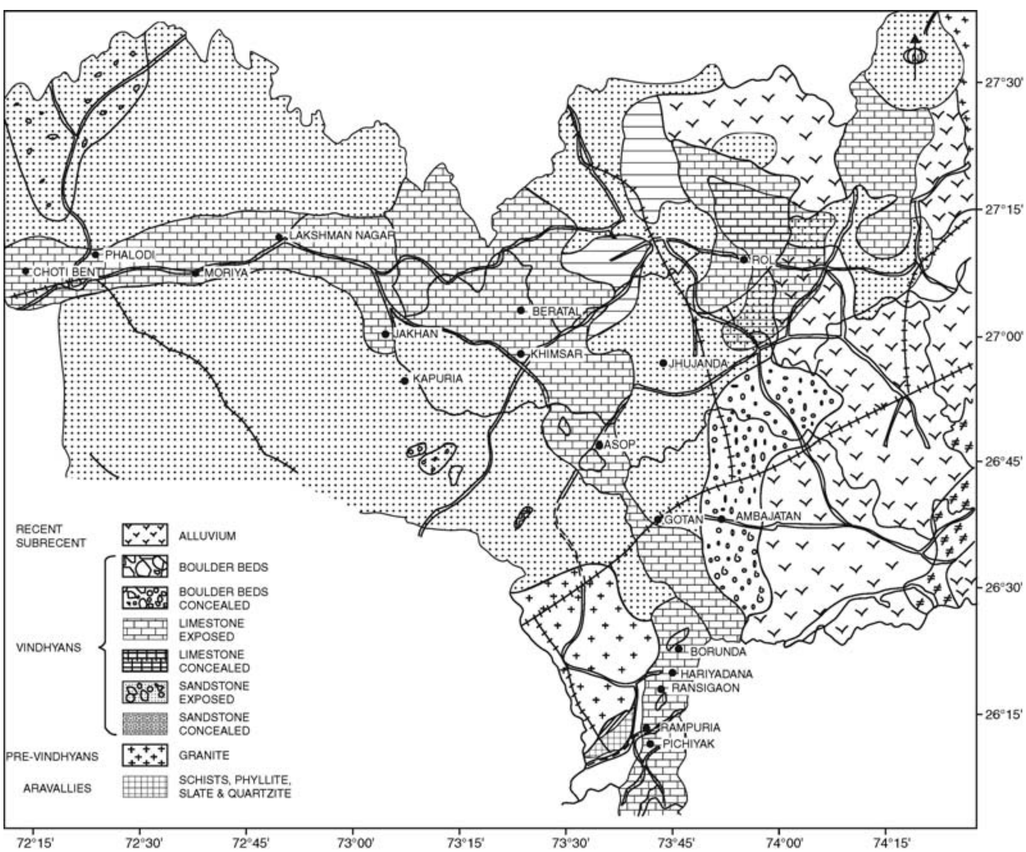

Fig. 14. Sample location map in the limestone belt of Jodhpur–Nagaur district of Rajasthan.

Phalodi (north of Jodhpur) shown in Figure 14. In the southern part of the study area a controlling fault has channelized the flow of groundwater in the cavernous and fractured limestone. Caverns are widespread and have large dimensions in some parts, and are the source of copious groundwater. A number of tube wells are located in Ransigaon, which supply water to Jodhpur town. Groundwater contours in the area in April (UNDP 1971) show that the general flow of groundwater is southwards in the dry season, and the controlling fault acts as a sump. Groundwater contours for October indicate that the recharge during and after the monsoon is temporarily from the Luni river, which is located in the southern parts of the area (i.e. the flow is northwards). The Tertiary alluvium to the east of the limestone area partially contributes groundwater inflow to the limestone aquifer. The main source of recharge to the limestone aquifer is rainfall falling within the drainage basin of the Luni river and its tributaries.

Groundwater samples were collected mainly from tube wells for environmental ^{3}H and ^{14}C analysis. The results obtained are presented in Table 2. The ^{3}H results show that the tube well sample from Pichiyak (located in the southern part of the area) has high ^{3}H of 10 TU showing modern recharge. The tube well is located near the Pichiyak lake which is fed by the Luni river, and hence lake water contribution to groundwater is possible. Other tube wells from Ransigaon, Borunda, Amba Jatan in the southern part and Jhujanda, and Rol in the northerneastern part of the limestone belt show ^{3}H contents of 2–4 TU, indicating some component of present-day recharge. Tube well samples from northern and northwestern parts such as Khimsar, Beraltal, Jakhan, Laxmannagar, Moriya, Phalodi and Choti Benti show negligible ^{3}H, indicating absence of modern recharge.

The ^{3}H data of samples collected in 1992 for Borunda, Gotan, and Khimsaar are lower than those collected in 1990. This may be due to over-exploitation of groundwaters in the Borunda area leading to the influx of older waters from the neighbouring areas.

Model ^{14}C ages (Pearson's model) are also given in Table 2. In Ransigaon and Borunda in the southern part, ^{14}C data show that they represent waters younger than 1000 years. Recent waters are also found in the northern parts at Jhujanda and Rol. In Phalodi area the groundwaters are old with ^{14}C ages in the range of 2500–4000 years. These groundwaters were probably recharged during the last pluvial episode. The ^{14}C ages of the groundwater in Beratal and Jakhan in the central part are 1750 and 1900 years, respectively, and the ^{3}H content is about 1.1 TU, suggesting that they are a mixture of young and old waters (Nair *et al.* 1993).

It may thus be concluded that modern recharge to the groundwater in the limestone belt occurs mostly from the southern parts. In the Phalodi area, the groundwaters are about 2500–4000 years old and were probably recharged during a pluvial period in the past.

Thus isotope techniques are useful in understanding groundwater recharge processes in arid

Table 2. *^{3}H and ^{14}C data of groundwater samples collected from the limestone belt (Jodhpur – Nagaur)*

Location	Date sampled	^{3}H (TU) ($\pm$0.5)	^{14}C (PMC) ($\pm$1)	^{13}C (‰) ($\pm$0.2)	Age (years)
Pichiyak	August 1990	10.3	–	–	–
Ransigaon	August 1990	3.4	–	–	–
Borunda	August 1990	3.9	–	–	–
	January 1992	2.8	64.6	−8.6	900
Hariadana	January 1992	1.9	60.1	−9.7	2400
Amba Jatan	August 1990	2.7	–	–	–
Gotan	August 1990	1.5	–	–	–
	January 1992	0.5	–	–	–
Rampuria	January 1992	1.4	–	–	–
Asop	January 1992	1.0	–	–	–
Khimsar	November 1989	2.0	–	–	–
	January 1992	1.0	50.7	−8.1	2400
Beratal	January 1992	1.1	54.3	−8.0	1750
Jhujanda	January 1992	3.3	82.9	−10.0	Modern
Rol	January 1992	4.3	71.1	−9.5	900
Kapuria	November1989	2.0	–	–	–
Jakhan	April 1992	1.2	55.2	−8.3	1900
Laxmannagar	April 1992	0.5	45.5	−7.0	3100
Moriya	April 1992	0.8	54.6	−8.9	2500
Phalodi	November 1989	1.3	–	–	–
	April 1992	1.2	54.8	−8.9	2500
Choti Bhenti	April 1992	0.5	44	−8.6	4000

areas. The absence of modern recharge and over-exploitation of aquifers observed in some of the above studies stress the need for proper management of this scarce resource.

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A tale of two cities in ancient Canaan: how the groundwater storage capacity of Arad and Jericho decided their history

A. S. ISSAR

Ben Gurion University of the Negev, Hameshoreret Rachel 13, Jerusalem 96348, Israel
(e-mail: issar@bgumail.bgu.ac.il)

Abstract: The history of two ancient cities, Arad and Jericho, sheds light on the role of groundwater storage in deciding the survival of settlements in arid and semi-arid regions when climate changes take place. Arad, which was dependent on a local perched horizon, was deserted during the global warm periods, which spelled dryness in the Middle East. This was in spite of one of the earliest systems of artificial recharge to groundwater developed by its inhabitants. On the other hand Jericho, which depended on the supply from a perennial spring, fed by a regional aquifer, was almost continuously settled from prehistoric times to the present.

The impacts of climate changes on global hydrology during the Holocene were investigated in the framework of a research project sponsored by UNESCO and the Institute of Desert Research of Ben Gurion University of the Negev (Issar 2003; Issar & Zohar 2004). The aim was to try to forecast the future impact of global climate change on the hydrological cycle of various parts of the globe. The rationale behind this research was that although past climate changes were due to natural causes, while the present change is mostly anthropogenic, the impact of past warm climates on the hydrological cycle may be replicated.

In the first stage of this project the sequence of the global climate changes during the Holocene had to be established. For this purpose, time series of proxy data (sea, lake and river levels; chemical characteristics of sediments in lakes, caves and riverbeds; changes in pollen and faunal assemblages; environmental isotope ratios; and changes in the rates of sedimentation) in the Middle East were investigated. The reason for choosing the Middle East as a starting point was because this region has been intensively investigated by historians, archaeologists and geographers during the last two centuries. This region also yielded the most ancient written records. From all these data the impact of climate changes on the natural environment and on the socio-economic systems could be observed.

The stratigraphic columnar section for the Holocene for this region was established, and was compared step by step with other regions around the world, where time series of proxy data were available. The most prominent periods of change on a global scale were chosen as key chrono-zones and a global correlation between these was established. One of the chrono-zones that could be correlated on a global scale was the climate change at *c*. 4000 BP. During this time period observations showed that it was an extremely warm period. In the Mediterranean region and especially the Middle East this warm climate spelled dryness, while in most tropical and subtropical regions it brought more rain and floods.

Generally speaking, the Middle East (and within it, Israel) is influenced by the Mediterranean climate system, which lies in the transition zone between the Westerlies low-pressure system to the north and the Azores high-pressure zone of the subtropics to the south. The Westerlies system, which moves to the south, is responsible during the winter for the rain-bearing low-pressure depressions. This causes the weather to become humid along the south and southeastern regions bordering the Mediterranean. In summer, the high-pressure belt moves northward, which brings higher temperatures and aridity to these regions.

The movement rate of the belts southwestward, and thus the number of rainstorms reaching the region, varies from year to year. In years when the belt of high pressure remains over the area, the rainstorms are less abundant and there is a drought.

These two factors, namely, the southern desert belt and the trajectories of the low-pressure, moisture-bearing storms, affect the mean annual quantity of rain as well as its variability from year to year. As can be seen from the multi-annual precipitation map (Fig. 1) aridity increases from the north towards the south and east.

This is even more pronounced in the Jordan–Dead Sea–Arava rift valleys because they are located in the rain shadow caused by the mountain range stretching from south to north along the central part of Israel and Palestinian territory (Fig. 2). The scarcity of rains, on the one hand, and its variance, on the other hand, become more

From: DRAGONI, W. & SUKHIJA, B. S. (eds) *Climate Change and Groundwater*. Geological Society, London, Special Publications, **288**, 137–143.
DOI: 10.1144/SP288.11 0305-8719/08/$15.00

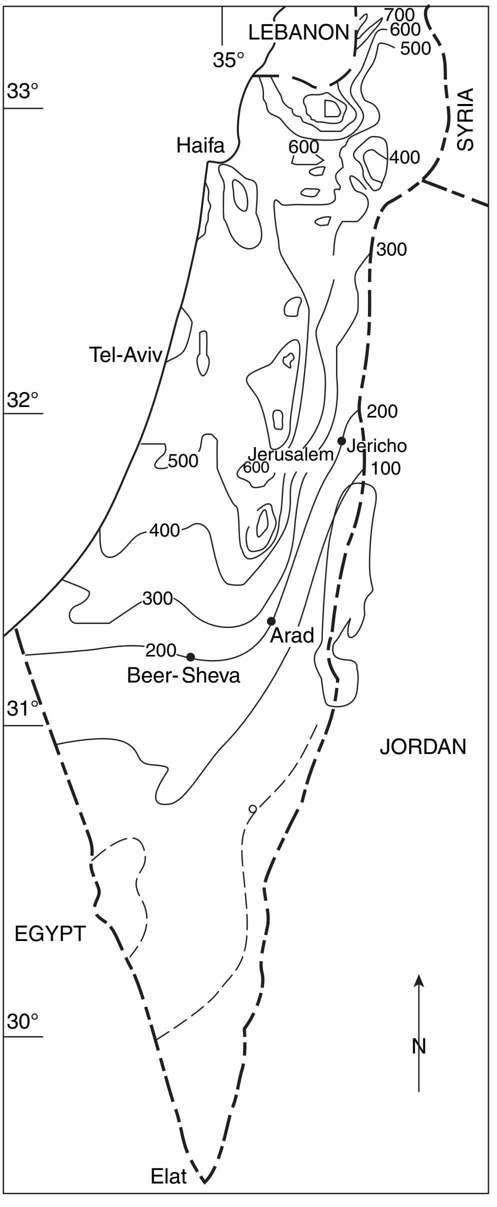

Fig. 1. Precipitation map of Israel.

extreme further south and east, namely in the more arid regions.

The two cities discussed, Jericho and Arad, lie in the proximity of the 200-mm/a isohyet line, which is the borderline between the more humid part and the arid part of Israel. (Fig. 1) This line moved northward and southward during historical periods in accordance with global climate changes during the Pleistocene and Holocene (Issar 2003). Its movement during historical periods had an impact on the welfare of the peoples of the Fertile Crescent and thus on their history. These changes can be traced by correlating time series of proxy data such as oxygen and carbon isotope ratios in cave and in lake deposits, as well as fluctuations of sea and lake levels with historical and archaeological records (Issar & Zohar 2004) (Fig. 3).

The left-hand curve in Fig. 3 shows changes in precipitation gathered from changes in the composition of environmental isotopes (^{18}O and ^{13}C) in the stalagmites in Soreq Cave near Jerusalem (Bar-Matthews *et al.* 1998). These data enabled estimation of the quantity of rain during historical periods. This was done by sampling contemporary rain and investigating the ^{18}O and ^{13}C isotope ratios and correlating it with the annual quantity of rain. Periods of high precipitation corresponded with periods of low temperature.

The middle curve represents ancient levels of the Dead Sea as deciphered from shoreline deposits and erosion lines in the salt caves of Mount Sodom, which is a salt plug in the vicinity of the Dead Sea. The dates were established by carbon-14 dating of wood, which was deposited along the ancient shorelines (Frumkin *et al.* 1991). The altitudes of these shorelines are very important for reconstruction of the ancient hydrological regime of the Dead Sea and thus the palaeoclimates; however, it should be taken into consideration that younger but higher shorelines may obliterate the marks of lower, older shorelines.

The right-hand curve provides information about former Dead Sea levels and is based on the mapping of sediments which were exposed due to the recent regression of the Dead Sea (Bookman *et al.* 2004).

The correlation between the three curves (allowing for minor differences on the time scale due to differences in type of sediments and their dating) leads to the conclusion that periods of high precipitation, which correspond with colder climate, correspond with high levels of the Dead Sea and vice versa. The main conclusion is that during cold periods the Middle East became humid, while during warm periods this region became drier.

Correlating the major climate changes with historical and archaeological data showed that during cold humid periods the Fertile Crescent flourished, while warm dry periods spelled socio-economic crises, desertion of urban centres, especially along the desert margins, as well as processes of desertification. The warmer the climate (as can be seen by the oxygen-18 ratios) the drier it was (as can be seen from the lake data) and greater the severity of the socio-economic crisis (archaeological data).

A study of the archaeological data from various sites situated along desert margins in the Fertile Crescent, which went through a crisis during

Fig. 2. Landsat image of Central Israel (courtesy of NASA). As can be seen, Jericho and Arad are east of the central mountain range and thus in the rainshadow with regard to rains coming from the Mediterranean Sea.

periods of dryness, shows that not all places shared the same fate. Urban and rural centres which got their water supply from non-perennial streams or perched local aquifers, such as Arad, Avdat and most of the Decapolis cities in Trans-Jordan, did not survive. On the other hand sites such as Jericho and Beit Sha'an, which were supplied with water from rivers or springs fed by regional aquifers, survived, and when totally deserted it was only for a short while. The perennial springs, which mitigated the impact of periods of dryness and also enabled rapid recovers, emerged in most cases from limestone and dolomite aquifers of Mesozoic age and in some cases of Eocene age. The aquiferous Mesozoic rocks build the backbone of the anticlinorial structures of the Taurides, Zagroids and Syrian arch, thus forming the high mountainous regions, which receive high amounts of precipitation and recharge the regional aquifers. These emerge as perennial springs feeding the main rivers of the Fertile Crescent.

It can be concluded that these geological pre-conditions, i.e. deposition of limestone and dolomite rocks during the Mesozoic and the post-Mesozoic folding, generated the hydrogeological conditions that helped the civilizations of the Fertile Crescent to survive during periods of warm and dry climate.

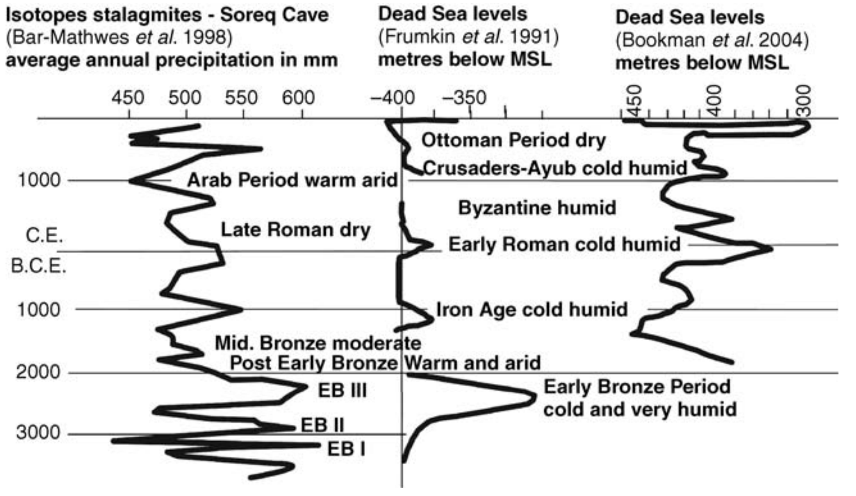

Fig. 3. Curves representing part of the proxy data which were used in order to decipher the nature of climate changes during the last 7000 years, and their impact on the socio-economic systems in ancient times.

As well as the natural pre-conditions, one should not neglect the cultural pre-condition. An important lesson that can be drawn from investigating the archaeological sites, is that in many cases a crisis resulting from the decline of flow from a spring has been postponed or entirely averted by human resourcefulness and innovation. This was achieved by various methods for augmenting and lengthening the period of flow, by tunnelling, or by inventing methods of pumping from wells.

The case of Arad

The ruins of the ancient city of Arad are located on the southern semi-arid flanks of the Hebron Mountains.

The average annual precipitation is about 200 mm, and there is no perennial water source in the region. The bedrock is composed of chalks of Eocene age (Avedat Group) overlain by loess of Late Pleistocene age (Issar & Bruins 1983). The chalks are mostly impervious, enabling the ancient and present inhabitants to dig cisterns for collecting rainwater flowing along the hillslopes. At the same time, dissolution processes have developed dissolution channels along fractured layers, causing the rock to become semi-pervious and thus water-bearing (Lerner *et al.* 1990). The chalks are underlain by a clay–marl impervious layer of Paleocene age (Takiya Formation).

The Early Bronze Age (2950–2650 BCE) walled city (Fig. 4), the area of which reached about 10 hectares, spread over the slope of a hill. At the lowest part of the hill within the walls of the city, a shaft was found and was excavated to a depth of about 20 m, but not reaching the bedrock. Its upper part is walled by masonry, while at the bottom the chalk bedrock is exposed (Amiran *et al.* 1987). Although the shaft is dry, its lower exposed part is moist. Since its excavation in the 1960s various opinions have been expressed about whether this shaft served as a cistern or a well. Rosenan (1978), Yair & Garti (1996) and Tsuk (2000) investigated the purpose of this shaft; they claimed that it functioned as a cistern collecting the runoff from the city draining from the streets, and sloped towards a sedimentation basin, and drained into the shaft. Yet, their calculations regarding the quantity of runoff, and taking into account the present value of precipitation, showed that the amount of runoff would not have been sufficient to supply the needs of the ancient city, which had a the population was about 3000 people. Amiran (1991) suggested that the climate was more humid and thus more water was available. Yair & Garti (1996) did not accept the climate change hypothesis, and suggested that in addition to the shaft, which was a communal project, each household collected its own water from its roof, which was made rain-proof and had an elevated rim and apertures from which the water flowed into big jars, many of which were found in the ruins.

The author's hydrogeological investigations in this region brought him to the conclusion that this shaft was an ancient well reaching a perched water table. This well was part of a complex system, which also served as an artificial recharge system. In other words the people of ancient Arad were innovative enough to develop a central water

Fig. 4. Aerial photo of the partly excavated Early Bronze Age city of Arad.

supply project, which utilized groundwater but at the same time recharged it by diverting the surface flow from the city into the shaft. The archaeologist who is presently carrying out excavations at this site, arrived at the same conclusion (Govrin, Y., pers. comm. 2005). At the same time, most probably each household collected its supply from its own premises.

The shaft, which is located at the lowest part of the city, touched the perched groundwater table on the clay–marl layer. This table was undoubtedly higher during the time in which this city flourished. As can be deducted from the palaeo-climate curve shown on Figure 3, the annual precipitation at Arad during the Early Bronze Period could have reached about 300–400 mm. Taking into account an order of magnitude of 3% recharge on this type of rock (Issar *et al.* 1984) and an area of about 1 km^2 for local recharge, the annual amount of groundwater available to be pumped by this well would be in the order of 9000 m^3. Based on the yield of shafts into the same rock in Israel, it can be assumed that such a well could have supplied about 15–20 m^3 per day, i.e. 5200 to 7200 m^3 per year.

Tsuk (2000), summarizing various estimations, came to the conclusion that in the pre-Byzantine periods the average demand for water per capita per year was about 5 m^3, with an additional quantity of 25 m^3 per year for livestock. Assuming that the latter were watered outside the city in the pasture areas from cisterns, and that the population, according to the archaeologists' estimates, reached about 3000 people, the total water demand could have reached 15 000 m^3 per year. Thus the supply from the well could not cover the demand unless augmented artificially. Yet adding to this about 4500 m^3 from the 15% runoff (Issar 1981) flowing from the built-up area extending over 10 hectares, and taking the higher value of the quantity supplied by the well, i.e. 7200 m^3 per year, then the demand is nearly balanced by the supply, taking into account that all these figures represent order of magnitude estimates.

This delicate balance between demand and supply explains the fact that as this region became drier towards the middle of the third millennium before Christian era (BCE), the city was deserted, without any significant evidence of its destruction by war. It should be noted that an ash layer shows that at *c*. 2800 BCE the city was conquered and burnt, but was immediately rebuilt as the climate and thus living conditions were optimal.

As can be seen on Figure 3, optimal climate conditions reoccurred at *c*. 1000 BCE. The favourable climate enabled the kings of Judea to build a fortress atop the acropolis, with a temple. The water supply of the citadel came from the collected runoff, which was diverted into two cisterns with capacity amounting to 160 m^3 (Govrin, Y., pers. comm. 2005). This is more or less the amount of runoff that could be collected from the area of 0.35 hectares of the citadel. A covered trench leading from the citadel outside may have served as a supply path for hauling water from the re-excavated well in the valley. Ash and ruin layers are evidence for destruction by war, which took place during the history of the Judean kingdom, and correlate quite well with events mentioned in the Bible. The citadel was totally destroyed *c*. 587 BCE when the king of Babylon conquered and destroyed Jerusalem and took many of its people into exile (Aharoni 1992).

The case of Jericho

Ancient Jericho (Tell es-Sultan) is located in the wide plain of the Jordan valley about 16 km NW of the northern shore of the Dead Sea and just to the east of the mountains of Judea. At its maximum height on the NW side, the mound rises 24 m, and its area is approximately 4 hectares. The fertile plain in which the site is situated is artificially irrigated by the spring of 'Ain es-Sultan' or Elisha's Fountain. It is fed by the regional aquifer formed by the 800-m-thick Judea Group, which constitutes the main karstic aquifer of the area. The Judea Group is formed mainly of limestones and dolestones and is divided into two main sub-aquifers. The Turonian Upper Cenomenian aquifer feeds perched springs in the vicinity, such as the springs of the Wadi el Qilt. The Lower Cenomanian Albian aquifer is the main source of the spring of Ein el Sultan (Fig. 5), the annual yield of which is about 6×10^6 m^3. The yield and hydrochemistry remain almost constant, even in years of low precipitation.

The site lies 255 m below sea level. It was inhabited in pre-historic times and was established as an

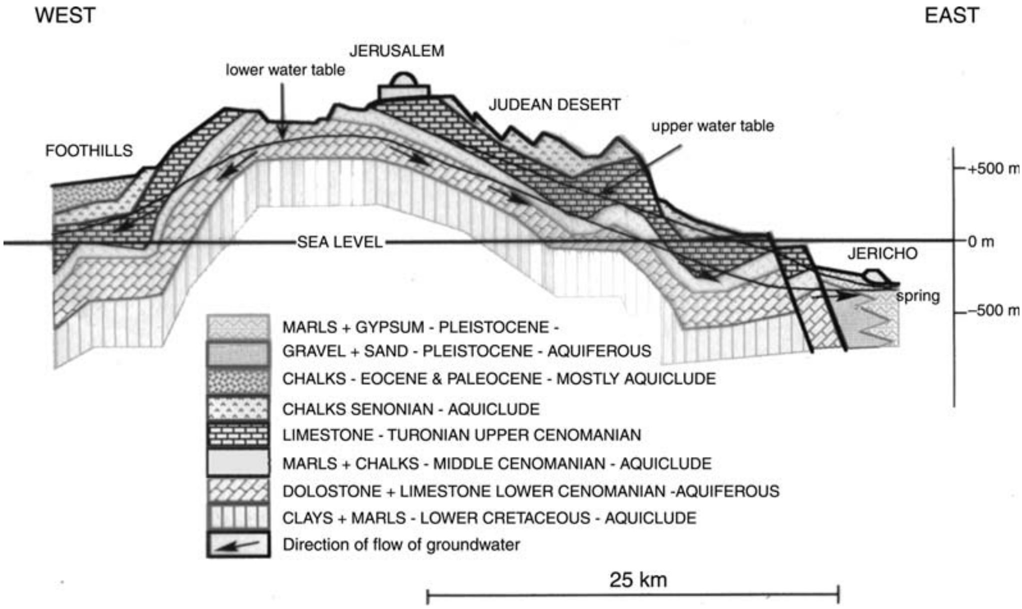

Fig. 5. Hydrogeological cross-section explaining the emergence of the spring feeding the oasis of Jericho.

agricultural community at the beginning of the Pre-Pottery Neolithic Period *c*. 10 000 years ago. Due to the aridity of the area, caused by its location in the rainshadow of the Judean Mountains, its Neolithic inhabitants practised irrigation while taking humanity's first steps in the field of agriculture.

The city flourished during the Early Bronze Period and was destroyed at *c*. 2200 BCE by war. It was rebuilt at *c*. 1800 BCE, when the climatic conditions became favourable, as can be seen on the climate section (Fig. 3). It flourished during the Middle Bronze Period until it was destroyed around 1550 BCE. The settlement during the Late Bronze period, a relatively dry period (around 1400 BCE), was poor and not permanent. The city was rebuilt during the later Iron Age (around the seventh century BCE) and survived until the Babylonian conquest at 587 BCE (Fig. 6).

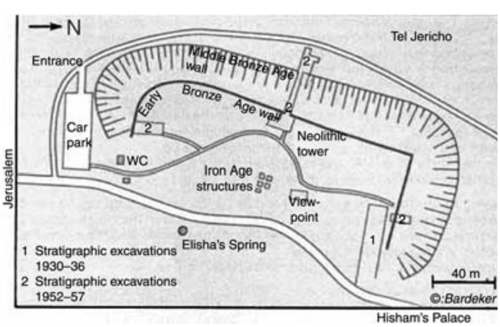

Fig. 6. Archaeological map of Jericho.

Thus although the flourishing and decline of Jericho followed the pattern of climate changes, which also decided the fate of Arad, the abandonment was not total, and during most periods the site was inhabited. The reason was undoubtedly the existence of a perennial spring, which, in the midst of a warm desert area, created ideal oasis conditions.

Summary and conclusions

Two case histories of survival of human societies in an arid region have been discussed. In both, creativity and invention enabled survival when climate conditions were moderate. Yet, when extreme negative climate changes occurred, human innovation was not sufficient to enable survival.

At ancient Arad during the Early Bronze Period (third millennium BCE) the city flourished due to the application of artificial recharge of a shallow local perched water table. At Jericho the transition from a socio-economic system of hunters and gatherers to that of farmers was enabled by the technical innovation of irrigation. In both cases, as in many other cases of settlements in arid regions, the reliance of human society on the natural system was dependent on the physical properties of the natural system for storing and delivering water. Because this was marginal at Arad, it did not survive when severe negative climate changes took place. On the other hand the ancient settlement of Jericho was able to withstand many climate changes because its supply of water for irrigation derives from a spring fed by a limestone aquifer drawing from a large area of the eastern Judean Mountains.

In general, the ability of a society to withstand the impact of climate change and its consequences depends on the total resilience of its sub-systems, both societal and natural. Yet, the magnitude of the impact and its duration play the decisive role.

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Impacts of climatic change on water resources: the future of groundwater recharge with reclaimed water in the south of Europe

M. SALGOT & A. TORRENS

Institut de l'Aigua de la Universitat de Barcelona, Laboratori d'Edafologia, Facultat de Farmàcia, Institut de Recerca de l'Aigua, Universitat de Barcelona, Joan XXIII, s/n. 08028 Barcelona, Spain (e-mail: M. Salgot: salgot@ub.edu)

Abstract: In an era of increasing contest for limited water resources, the wise joint management of conventional and non-conventional water resources must be considered. Water scarcity is aggravated in coastal zones which are often characterized by high population densities and intense economic activities making heavy seasonal water demands. In this context, the use of non-conventional water increases the availability of water supplies. Non-conventional water resources of lower quality could be directed to meet additional needs. As a consequence, significantly more potable water would be available to meet human demand for safe water.

Non-conventional water resources are described: waste water reclamation and reuse, and its potential application for increasing groundwater resources, as well as several practical applications.

Water demand is increasing in the south of Europe due to the economic development. The southern parts of Portugal, Spain, France, Italy and Greece are strongly associated with tourism activities. Coastal resorts, golf courses, and other water-consuming facilities are spreading throughout the tourism-related areas. Agricultural water demands as well as industry and services in localized areas are also growing in several parts of these countries. Population all around the Mediterranean borders is experiencing additional population growth because of foreign and European immigration (the latter comprise mainly retired persons).

Nevertheless, a difference arises when comparing coastal areas with the interior ones. Population has a strong tendency to inhabit coastal areas for several reasons. These include a mild climate compared to inland, important economic activities related to transportation, and other benefits such as landscaping aesthetics, flat areas for living and access to bathing waters. Also, due to the ease of transport by ship, road or railway, a lot of industries are located near harbours on the coastline.

Among the richest agricultural lands, deltaic and similar areas occupy a prominent site. Vegetables, rice, corn and other rewarding crops can be grown in such areas, using at the same time huge amounts of water. Citrus and other orchards also need to be located in the vicinity of the sea or in places with mild climate.

Mass or select tourism demands sunny, politically safe areas with good transportation facilities, located close to tourist and cultural sites. The people involved, and the associated golf courses, parks and the like, are demanding water in large quantities. An important part of the Mediterranean, including its islands, is meeting all or almost all of these conditions.

Considering all these activities in terms of water resources, we notice a concentration of demand in time and space. This happens for the majority of cases mainly during summertime and in a fringe reaching a few kilometres inland from the sea. Demands in the inland areas are usually more evenly distributed during the year, although it depends on the existence of irrigated areas and agricultural patterns.

Climatic changes will affect the pattern of water resource availability, thus making it more difficult to cope with water demands and increasing health hazards (McMichael 1997). Molles (2002) indicates that the global circulation models predict an Earth more subject to drought, particularly in the interior of continents. Water availability is expected to be limited for populations in many regions because of changes in precipitation and increases in evaporation (Findlay 2003).

These circumstances make it even more attractive to find non-conventional solutions to fulfil the water demands. Among the solutions, treated wastewater reuse through groundwater recharge presents a number of advantages, namely the increase of water availability, the use of aquifers for water storage, and the long hydraulic detention times which contribute to the advanced treatment.

It is important to recognize that there is consumption of recycled water from both non-intentional and intentional recharge (Jiménez 2003).

From: DRAGONI, W. & SUKHIJA, B. S. (eds) *Climate Change and Groundwater*. Geological Society, London, Special Publications, **288**, 145–168.
DOI: 10.1144/SP288.12 0305-8719/08/$15.00

Water resources

The usual way to cope with water demand is to use conventional water sources, namely surface and groundwater. Since surface waters are not evenly distributed along the coastline and groundwater resource is limited, water quantity must be increased through water transport or by using non-conventional resources. A classification of the different types of water resources based on their origin is shown in Table 1.

Although it seems that a balance between demand and supply should be obtained by increasing supply, the most sustainable way to do it is through the implementation of water-saving measures. This sometimes means a reduction in demand increasing efficiency, or a reduction in the extraction of natural water through the use of several non-conventional water resources.

In coastal arid and semi-arid climate locations, usually corresponding to areas where agriculture and tourism are located, surface waters (streams and lakes) can offer a limited amount of water resources. The limitation is due to the uneven distribution of rivers, a lack of streams on islands, a coincidence of minimal stream flows with peak demands, the need to maintain a minimal environment-related flow, and several other factors. Groundwater can offer a limited water supply because of low flow velocity, the need for pumping, and the need to maintain water barriers to control seawater intrusion. The storage capacity of aquifers must not be forgotten (Dillon *et al.* 2005).

New conventional resources can be obtained through infrastructures for water transportation from other basins. Nevertheless, this solution becomes difficult because of non-availability of resources, non-acceptance from donor areas, or financial limitations. Additionally, all existing conventional water resources are nowadays fully allocated in the developed countries.

As indicated above, when the availability of conventional water resources reaches a limit, additional supplies can be obtained from other water sources, mainly non-conventional ones, which include seawater and brackish water desalination, rainwater and runoff water, and reclaimed wastewater. Other possibilities are transportation using trucks, railways or ships. Socio-economic features are a key issue for implementing such alternatives. Transportation costs, energy demand and availability (desalting), and socio-economic factors when dealing with wastewater reuse are the limiting factors.

Water balance

From the viewpoint of balancing supply and demand of water resources, excess demand can be dealt with either by increasing supply or by limiting demand. Increasing supply will not create a balance, because there is a false sensation that authorities will facilitate an ever-increasing amount of water, which is by definition impossible.

On the demand side, water authorities can act by pricing (increasing prices, which are in relation to what is called 'demand elasticity') or by supporting water demand reduction through the use of water-saving devices and education campaigns (Griffin 2001) or through the use of taxes. It seems that important water savings can be obtained in agriculture, since this activity is the main water consumer in arid and semi-arid climates. Nevertheless, reduction in water uses in agriculture is not as clear a practice as it seems, because excess water in the agricultural fields contributes to the leaching of salts from the soils. When insufficient water is applied, salts can accumulate, several crops cannot be grown anymore, and soils can finally become completely unproductive.

Industry in developed countries can be persuaded to reduce water use through pricing and other easily implemented measures, while savings in domestic/urban demand are practicable in specific cases (e.g. improving distribution networks,

Table 1. *Classification of water resources depending on the origin*

Inside the basin		External to the basin*
Conventional	Non-conventional	
Surface water Groundwater	Brackish water Seawater Runoff Reclaimed wastewater Dew, frost, fog, etc.	Classical transfers among basins Other movements (ship, railway, etc.)

From Salgot & Folch (2003).
*The classification among conventional and non-conventional can also be performed here, if necessary.

installing in-house devices), although the savings are not as important as those of agriculture.

Coastline specificities

Because water-consuming activities on the coastline are located at the end of basins, there is almost always an association of scarce flows (in relation to demand) with scarce quality of existing flows, because of previous use and disposal in the basin. Apart from this, it is difficult to build water storage facilities near the points of use on the coastline. There is a tendency to cope with demand by first using surface waters (easy to obtain from the physical and economic aspects) and afterwards using groundwater, especially when peak demands are to be covered. Nevertheless, a lack of planning procedures for water extraction is usual in these cases, especially while existing resources can cover the demand easily. Planning efforts start when drought or scarcity appears, which in most cases cannot solve the problems.

Water resources and wastewater need to be conveyed to the point of use or to the point of treatment/disposal. As the territory is usually fully occupied by housing developments along the coastline, the only space usually free to install infrastructures is the shore. It is therefore typical to find water transportation infrastructures (pipelines, channels) along the coastline, especially for sewerage. The water for distribution reaches the inhabited areas, the industries and the agricultural areas in a localized way. Afterwards, there is the need to build sewerage facilities for non-consumptive uses, mainly urban and industrial.

Agriculture can also generate water-related pollution problems. As water needs to be applied in excess of water demand, infiltration or irrigation tails are at all times heavily charged with salts, nutrients and pesticides. This generates a negative impact on existing water resources, either groundwater or surface water (diffuse pollution), and occasionally seawater if the agricultural area is located in the vicinity of the shoreline.

Sewerage facilities on the coastline present additional problems because of the relationships with seawater. Seawater can enter sewers through discontinuities or through old outfalls, as tides or strong winds force seawater into the conduits. Building activities on the shoreline tend to pump water (mainly seawater under the sand) and throw it into the sewerage.

When groundwater is extracted to excess in the vicinity of the interface with seawater, the consequence is seawater intrusion into the coastal aquifers (Custodio & Bruggeman 1987). This situation is typical in locations with water stresses all over the world. In recent years, several solutions have been proposed, from an integrated management of aquifers to groundwater recharge using reclaimed wastewater. Aquifers can play a key role for water storage in coastal areas, as it is impossible to build important surface storage infrastructures near the coastline due to the lack of suitable locations.

Water use in coastal areas normally has uneven patterns throughout the year. Peak demands from different users (agriculture, tourism, year-round and seasonal inhabitants) tend to concentrate during the summer months, thus exerting a strong pressure on water distribution facilities and on water authorities. It could happen that several users exert demand in two or more different sites, if they own two or more permanent or temporary residences along the coast in the same country or in different ones.

Urbanization of the coastline (resorts, cities, industries, surface and ground urban infrastructure) disturb natural water circulation in three different ways. The first is related to ‘natural structures’ for groundwater circulation from mainland to the sea; underground railways, roads, etc., create artificial barriers. The second is the implementation of infrastructures that modify the surface water circulation (high buildings, urbanization of inundation areas and wetlands) or are built over rivers and small stream beds. The third is the occupation of infiltration surfaces with impervious structures (car parks, buildings, streets). Then, groundwater and surface flows are reduced, and the normal circulation of water is not allowed, creating a different water flow model at the land–sea interface, which can extend to thousands of metres and even kilometres towards the interior. It needs to be managed if specific events or natural hazards occur. Then, so-called ‘geohazards’ can appear and generate ‘natural disasters’, which are to some extent facilitated by human infrastructures (Coch 1995).

As a result, wastewater from industries and towns and stormwater are concentrated at the end of the pipes and need to be evacuated, usually to the sea, because there are no other solutions except reuse. Submarine outfalls are the classic solution for wastewater, but stormwater needs different solutions. Urban runoff coming from rainwater does not necessarily imply clean water, because runoff is effectively cleaning the entire town, from streets to sewerage systems, including roofs and other impervious surfaces. Since there is no possibility for water to infiltrate, such water reaches the sewerage if unitary systems are implemented, or the rainwater elimination systems if two separate networks exist (Metcalf & Eddy, Inc. 2003).

In the case of heavy rains, runoff reaches the sea along the coastline through alleviating devices or by surface infrastructures, thus polluting seawater. This creates problems in bathing areas, because of water quality problems during several days or hours (depending on the tides and currents). The problems are due not only to faecal contamination, but to suspended solids that can include dead animals or huge suspended solids. Apart from the aesthetic problems, health-related risks appear temporarily.

The existence of harbours and marinas in the coastal environment should not be forgotton. Those facilities can exert a certain influence through the dissemination of wastewater and loss of paints and other chemicals from ships. Then, neighbouring systems (bathing water, sand) become heavily contaminated, but the contamination is not of land origin. Also, the facilities located on the coastline modify the circulation of solids along the coast, and can affect the extension or form of beaches.

Mainland specificities

Usually, water users in the upper and middle parts of the basins experience fewer problems than users located at the end of the basins. The reasons are numerous, from the better quality (water has not been used before) to higher amounts of water available. There are other features to be considered, such as availability of specific sites for building dams and diverting water to the presence of wildlife. Groundwater is also used when water needs are located away from the surface water sources. Irrigated polygons also appear in these areas, where water is diverted from rivers, dams or lakes to usually open channels for water transportation and distribution. Usually, the demands are covered with the existing resources.

As in the coastal areas, agriculture can generate pollution, usually through non-point sources. There are also point source pollutions, derived from cattle facilities, towns and industry. In comparison with coastal areas, the territory is not fully occupied, except in the case of big cities.

Water uses in inland areas have a pattern of use derived from water demands; but scarcely related to tourism. Main demands are associated with irrigated agriculture and existing towns. Peak demands are not as strong as in the case of coastlines. Infrastructures are distributed along a more extended territory and so their impact is ‘diluted’ in some way.

In any case, there is a need to find a place for the final disposal of effluent generated in domestic, industrial or cattle wastewater treatment facilities. Rivers or lakes are the most usual disposal sites, but this creates point pollution. Alternatively, soil–plant systems can be used, or reuse is a possibility if advanced (reclamation) systems are implemented.

Climate change and groundwater resources

Although effects of climate change are still largely hypothetical, it seems that changes in meteoric water will become evident and affect the amount of water available for human activities. Precipitation patterns will change and it seems obvious that the relations between surface waters and groundwater will experience changes. Modifications in recharge, quantitative and qualitative, are also to be expected, although it is not clear if the amount of groundwater will be affected in a global basis. Local situations will change if precipitation is reduced or augmented in a noticeable way.

The role of groundwater as a storage system for water will be enhanced. Natural recharge will acquire more importance and artificial recharge will be supported and enhanced. Recharge through soil/subsoil systems will be studied further and the possibilities for water contamination through the solid matrix must be considered and managed adequately.

The final result will be an increase in quantity and change in quality in some places. Because up to now we are referring to theoretical models, it is not possible to establish clear patterns. As a consequence, groundwater recharge could play an important role in specific places, especially where non-conventional resources are available and soil and geological formations are adequate for artificial recharge.

Non-conventional water resources

If conventional resources are not usually available to cover excess or peak demands, non-conventional resources are employed. Among natural non-conventional water resources several possibilities are considered, but the usual ones are runoff water, natural brackish water, or desalinated seawater (Georgopoulou *et al.* 2001). Non-conventional water resources coming from the anthropic water cycle can also be used; reclaimed wastewater will be considered (Asano & Levine 1998).

Runoff water

In the Mediterranean basin, a lot of streams carry water only after rain events; this is due to high rain intensity, the small size of coastal basins, and the imperviousness of important parts of the basin. Due to the disappearance of coastal lagoons, marsh and wetlands (the natural ways to control excess flows), and sometimes to the transformation of free-flowing streams in pipeline systems, there is an increase in the quantity of runoff per unit time.

Because this runoff is concentrated in short periods of time and does not recharge groundwater at all, it can be used to increase water supplies if appropriate treatments are performed. This is the case, for example, for Palma de Majorca, where occasional runoff is collected, treated and distributed as tap water for the city (Terrassa & Cadenas 2000). There is an economic consideration here: is it worth building treatment infrastructures for a scarce amount of resources? The answer is not easy and demands good economic assessments.

When there is sufficient space to install constructed wetlands using marginal lands or areas without any specific use such as motorway surroundings, rainwater can be diverted to such areas until it percolates and recharges groundwater using the soil/subsoil system (vadose zone) for additional treatment. In other cases, small or medium-sized storage pools or dams can be used to retain such waters if there is enough space.

Desalinated water

If there is no other water available, water containing high amounts of salts can be used. The salt contents can be eliminated using several technologies, but mainly membrane-related ones (Medina 2000). Several crops can use waters containing certain amounts of salts during the growing period. Sometimes this improves the quality of the crop, because it forces the plant to produce more sugar or other substances.

Apart from evaporation, which consumes excessive amounts of energy for Mediterranean standards, reverse osmosis (RO) and electrodialysis reversal (EDR) are being used nowadays for water desalination. RO is preferred when using seawater as affluent, while for salt contents up to 10 g/l EDR seems to be the best technology. As the cost of water desalination could exert a strong influence on the choice of technology even at the planning stage, decision-support systems should be employed in order to choose the best system for specific circumstances.

Water from RO cannot be used directly, but needs to be mixed with other water containing salts or directly with salts (e.g. lime). The RO water can be assimilated with distilled water, which is not adequate for direct human consumption or for irrigation. There is an additional problem in relation to RO and other desalting technologies, which is the salt generated during the separation processes. If salt is in a solid state the problem is not so complex; nevertheless, the brine generated by EDR must be disposed of. The usual destination has been the sea, through outfalls; however, there is increasing concern about the impact of such brines in the sea biotope, because it seems that *Posidonia* and other components of the marine ecosystem suffer from this excess of salts.

When working with integrated management of water in areas where desalination is being practised, reuse is an imperative because after the first use water has still a low content of salts, and can be easily employed after advanced treatment for irrigation. This is a way to spare energy and resources.

Reclaimed wastewater

Reclaimed wastewater is the most commonly used non-conventional water resource. As wastewater reclamation and reuse have been studied extensively since the 1920s, and great amounts of treated wastewater are usually available at the coastline and in inland towns and industries, reclaimed wastewater should be fully studied and employed.

Reclaimed water is defined as wastewater purposely treated for reuse. However, it seems that nowadays 'water recycling' is the preferred term for generic water reclamation and use in view of the acceptance and success of other urban recycling programmes (AATSE 2004). As with the term 'biosolids' for sludge to be reclaimed, 'water recycling' could be a marketing definition, because of the disappearance of the term 'waste' and the appearance of the term 'recycling', which has positive connotations.

Apart from the conventional wastewater from urban areas, which consists mainly of domestic wastewater with small amounts of industrial wastewater, there is the possibility to separate different types of wastewaters at their source: the houses. Ledin *et al.* (2001) describe the possibility to reuse/infiltrate so-called 'diluted', 'light' or 'grey' wastewater. All three terms refer to wastewater produced in households, office buildings and schools as well as some types of industries, where there is no contribution from toilets or heavily polluted process water. Grey wastewater is wastewater from baths, showers, hand-basins, washing machines and dishwashers, laundries and kitchen sinks. This type of wastewater has been estimated to account for about 73% of the volume of combined residential sewage. In general terms, grey wastewater has lower concentrations of organic matter, some nutrients (e.g. nitrogen) and microorganisms than combined wastewater. However, the concentration of phosphorus, heavy metals and xenobiotic organic pollutants are around the same level.

Other non-conventional resources

Other non-conventional resources are water transported through non-classical methods (ships, railways), and water obtained from condensation; usually, these resources will not offer significant amounts of water and just serve for specific situations (see Table 1), where localized problems can be solved in this way.

Wastewater reclamation and reuse

Wastewater to be reclaimed can be generated by towns, industries and agriculture in the broad sense (i.e. cattle, forestry, aquiculture). Nevertheless, the most studied reclamation procedures are those with domestic or urban wastewater. In this case, the whole flow can be assimilated to domestic wastewater with respect to quality. As the EU Directive (91/271) on wastewater is theoretically implemented and the water generated has at least secondary treatment, wastewater-related impacts are reduced to a certain acceptable level.

As water is used for many purposes in a developed society (Table 2), it should be considered which of these uses can be filled with reclaimed wastewater. Practically all uses can, theoretically, have reclaimed wastewater as a source. Nevertheless, several of them are usually discarded (e.g. potable uses), although there is a description of reclaimed wastewater being used for tap water in Windhoek, Namibia (Odendaal *et al.* 1998) for more than 30 years. Full studies of this possibility are described by WEF & AWWA (1998).

Due to the several possible uses of reclaimed water, there is a need to attain a certain quality (of reclaimed water) depending on the type of use. For functional reasons, five groups of reclaimed water use (excluding tap water) are considered:

- urban uses (other than tap water),
- agricultural,
- industrial,
- environment and leisure, and
- groundwater recharge.

The main use of reclaimed wastewater at the global scale is in agriculture. This is because agricultural use of water for irrigation is the most usual. Due to the fact that the main production of wastewater is in towns and big cities, which become more and more distant from agricultural areas, there is excess reclaimed water near towns. If this water is needed in the country, there are two main solutions, namely transportation from the production area to the places where it can be used, or the search for uses other than agricultural ones near or in towns.

In the first case, there is the need to build a network for transportation and final distribution. A classic case is in Israel, where reclaimed wastewater from Tel Aviv area is conveyed through a water carrier to the Negev desert (Chikurel *et al.* 2001).

The second case is being developed through the use of reclaimed water for other purposes, like urban parks and gardens irrigation, industrial use or groundwater recharge. If we consider additional uses, as indicated in Table 2, we can describe, in relation to the coastline, streamflow augmentation and landscape-related uses, which in turn can contribute indirectly to groundwater recharge.

Reclaimed water quality

Reclaimed water quality is nowadays defined by a few analytical parameters, even in regulations, but the need arises to improve the determination of qualitative aspects, perhaps not increasing the number of analyses but implementing complementary tools, like risk assessment or good reuse practices. However, there is a lack of physical-chemical parameters and toxicity-related ones.

Treated wastewater needs to be reclaimed by advanced treatment before reuse. For the specific purposes of recharge, two different qualities need to be established depending on the type of recharge. Direct recharge means that water reaches the aquifer directly, without natural barriers, while indirect recharge is based on the passage through soil and subsoil.

Several rules and regulations could establish the different qualities needed for any specific case, but it is absolutely necessary that water in the aquifer does not reduce its quality. When reclaimed water is used for creating a barrier against seawater intrusion, the quality does not seem at first sight to be so important. Yet, this is not acceptable because of the possibility of reclaimed water reaching wells inland which could be used for obtaining tap water or water for other purposes.

Theoretically, water quality is improved while remaining in the aquifer and for this purpose long water residence times are suggested or mandated. Nevertheless, due to the specific features of aquifers, one should not expect great activity of the solid matrix on the water quality.

Water quality considerations may be classified into biological, chemical and physical. From the biological point of view, the number of water-related pathogens is important (Table 3) but it is impossible to examine the water to detect the presence of them all. Instead, indicators are employed. In this sense, *E. coli* and the coliforms (total and faecal) are used to establish the microbiological quality of reclaimed water and the efficiency of the treatment. Nevertheless, the mentioned indicators are not at all reliable to guarantee the absence of viruses and parasites. For this reason other indicators are being employed (e.g. nematode eggs) or suggested (e.g. bacteriophage), and several pathogens are determined directly (e.g. cysts and

Table 2. *Possible uses of water, observations of quality, and resources adequate for the defined use*

Type of use	Specific use	Observations	Useful resource
Urban domestic 'potable'	Drinking Hygiene Cooking/food-related	Maximum quality (suitable for drinking purposes)	Conventional Occasionally reclaimed wastewater (if there are no other possibilities) with or without blending
Urban commercial	Drinking, cooking and hygiene in hotels, restaurants, and similar		
Urban fountains	Drinking		
Urban 'general' not for drinking, but related to domestic	Air conditioning Toilet flushing	Disinfected, especially for *Legionella* Disinfected	Conventional and non-conventional Conventional and non-conventional
Urban not domestic	Fire protection Irrigation of public spaces Irrigation of private spaces Urban cleaning (streets. . .)	Disinfected	Conventional and non-conventional
	Sewerage management	Secondary treatment if wastewater	Non-conventional if possible
Industry	Food-related Pharmaceuticals and similar	Tap-water quality	Conventional
	Cooling water* Boiler feed Process water Heavy construction	Constituents related to scaling, corrosion, biological growth, and fouling to be controlled	Conventional and non-conventional
Agriculture	Food crops Non-food crops	For reclaimed wastewater: rules or regulations. River water and freshwater usually do not have quality-related rules	Conventional and non-conventional
	Aquaculture	Specific rules (WHO)	
Non-agricultural irrigation	Golf course irrigation	For reclaimed wastewater as indicated by rules or regulations	Some countries are forcing these facilities to use only non-conventional resources
	Landscape Forest	Freshwater usually don't have quality-related rules	Conventional and non-conventional
Livestock	Watering and dairy operations	Should be 'potable'	Conventional mainly, and non-conventional other than reclaimed water
	Fish farming	Specific rules (WHO) for the use of reclaimed water	Conventional and non-conventional

(*Continued*)

Table 2. *Continued*

Type of use	Specific use	Observations	Useful resource
Groundwater recharge	Direct recharge	Advanced tertiary treatment if wastewater Pre-potable if other source	Conventional and non-conventional
	Indirect recharge	Through soil formations	
Water-related sports, leisure activities	Contact allowed	Specific rules and regulations if reclaimed wastewater is used	Conventional and non-conventional
	Contact not allowed Snowmaking Leisure boats maintenance		
Stream and water – body recharge	Habitat wetlands Lakes and ponds Enhancement of marsh and similar Streamflow augmentation	Toxicity for aquatic and water-related wildlife	Conventional and non-conventional
Thermoelectric power use	Power generation	No need of quality rules, but resource management is essential	Usually conventional, but stream augmentation with reclaimed wastewater is possible

From Brissaud *et al.* (2005).
*Disinfection for *Legionella*.

Table 3. *Pathogens potentially present in wastewater and their associated diseases*

	Pathogen	Associated disease
Bacteria	*Salmonella typhi*	Typhoid fever
	Salmonella paratyphi	Paratyphoid fever
	Salmonella spp.	Salmonellosis
	Shigella spp.	Shigellosis (bacillary dysentery)
	Vibrio cholera	Cholera
	Vibrio parahaemolyticus	Gastroenteritis from seafood
	Campilobacter jejuni	Gastroenteritis
	Pathogenic *Escherichia coli*	Gastroenteritis
	Enterobacter aerogenes	Gastroenteritis
	Klebsiella pneumoniae	Pneumonia
	Proteus mirabilis	Urinary tract infections
	Serratia marcescens	Opportunistic infections
	Haemophilus influenzae	Meningitis, other pediatric diseases
	Coxiella brunetti	Q fever
	Chlamydia psittaci	Psittacosis
	Mycoplasma pneumoniae	Primary, atypical pneumonia
	Staphylococcus aureus	Food poisoning, skin infections
	Streptococcus pyogenes	Pharyngitis, skin infections
	Enterococcus faecalis	Opportunistic infections
	Bacillus anthracis	Anthrax
	Clostridium botulinum	Botulism, food poisoning
	Clostridium perfringens	Gas gangrene, food poisoning
	Clostridium difficile	Gastroenteritis, colitis
	Listeria monocytogenes	Meningitis
	Corynebacterium diphtheria	Diphtheria
	Actinomycetes israelli	Actinomycosis
	Mycobacterium tuberculosis	Tuberculosis
	Mycobacterium avium	Pulmonary disease, disseminated disease in immunocompromised
	Pseudomonas aeruginosa	Wound, burn, urinary tract infections
	Brucella abortus	Brucellosis
	Brucella melitensis	Brucellosis
	Brucella suis	Brucellosis
	Bordetella pertussis	Pertussis (whooping cough)
	Francisella tularensis	Tularemia
	Yersinia enterocolitica	Yersiniosis (diarrhoea and septicaemia)
	Legionella pneumophila	Legionellosis
	Leptospira spp.	Leptospirosis
Viruses	Polioviruses	Poliomyelitis
	Picornaviruses (animal viruses)	Paralysis, common cold, myocarditis
	Togaviruses (animal viruses)	Encephalitis, yellow fever
	Paramyxoviruses, rhabdoviruses (animal viruses)	Measles, mumps, rabies
	Orthomyxoviruses, arenaviruses (animal viruses)	Influenza, haemorrhagic fevers
	Hepatitis A virus	Infectious hepatitis
	Hepatitis E virus	Hepatitis
	Rotaviruses	Gastroenteritis
	Retroviruses (animal viruses)	Leukaemia, tumours, AIDs
	Adenoviruses	Respiratory diseases
	Herpesviruses (animal viruses)	Oral and genital herpes, chickenpox, shingles, mononucleosis
	Poxvviruses (animal viruses)	Smallpox, cowpox
	Papovaviruses (animal viruses)	Warts
	Parvoviruses (animal viruses)	Roseola in children, aggravates sickle cell anaemia
	Norwalk agent	Gastroenteritis
	Reoviruses	Gastroenteritis

(*Continued*)

Table 3. *Continued*

	Pathogen	Associated disease
	Astroviruses	Gastroenteritis
	Caliciviruses	Gastroenteritis
	Coronaviruses	Gastroenteritis
	Coxsackie A	Meningitis, fever, respiratory illness, herpangina
	Coxsackie B	Myocarditis, rash, meningitis, fever, respiratory illness, pleurodynia
	Enteroviruses 68-71	Meningitis, encephalitis, respiratory illness, rash, diarrhoea, fever
	Echoviruses	Meningitis, encephalitis, respiratory illness, rash, diarrhoea, fever
Protozoa	*Entamoeba histolytica*	Amoebiasis (Amoebic dysentery)
	Naegleria fowleri	Primary meningoencephalitis (PAM)
	Acanthamoeba spp.	Meningoencephalitis, eye lesions, respiratory and skin lesions
	Giardia intestinalis	Giardiasis (diarrhoea)
	Cryptosporidium parvum	Cryptosporidiasis (diarrhoea)
	Isospora spp.	Diarrhoea
	Balantidium coli	Balantidiasis (diarrhoea, dysentery)
	Cyclospora spp.	Intestinal diseases
	Toxoplasma spp.	Toxoplasmosis
	Enterocytozoon bieneusi	Diarrhoea
	Encephalitzoon councili	Disseminated disease of lungs and liver
	Encephalitzoon intestinalis	Disseminated disease of lungs and liver
	Phyllum *Microspora*	Microsporidiosis (intestinal and nervous diseases)
Helminths	*Ascaris lumbricoides* (N)	Ascariasis (roundworm infection)
	Ancylostoma duodenale (N)	Anaemia, intestinal diseases
	Necator americanus (N)	Anaemia, intestinal diseases
	Clonorchis spp. (T)	Clonorchiasis
	Taenia spp. (C)	Taeniasis
	Enterobius vermicularis (N)	Enterobiasis
	Hymenolepis nana (C)	Hymenolepiasis
	Trichuris trichiura (N)	Trichurasis
	Schistosoma spp. (T)	Schistosomiasis (bilharziasis)
	Strongyloides stercoralis (N)	Diarrhoea, abdominal pain, nausea
	Toxocara canis (N)	Fever, abdominal pain
	Toxocara cati (N)	Fever, abdominal pain

From: Metcalf & Eddy, Inc. (1991); Rowe & Abdel-Magid (1995); Yates & Gerba (1998); Haas *et al.* (1999) and Salgot (2002).
N, Nematodes; T, trematodes; C, cestodes.

oocysts of *Giardia* and *Cryptosporidium*) and are designed as index organisms.

From the physical and chemical points of view, the number of wastewater constituents is huge, and it seems impossible to deal with such a huge number of them. Detail can be found in Tables 4 and 5. For the physical parameters, the number is relatively reduced and several of them can be determined in a continuous way (e.g. temperature, pH and conductivity). It is not possible to find indicators for chemicals, and there is the need to know the origin of the treated wastewater and decide which types of chemicals are most important on a case-by-case basis. For example, when determining volatile organic contaminants, Romero *et al.* (2003) describe a method to make an initial assessment, without having to determine the whole spectrum of such chemicals.

Rules and regulations

Although almost all the laws governing wastewater reuse refer to its use for agriculture, there are several countries that support groundwater recharge and have established limitations for such purposes. It is necessary to distinguish between regulations, i.e. actual rules that have been passed and are enforceable by government agencies, and guidelines, which are not enforceable but can be used in the development of a reuse programme (Salgot & Angelakis 2001).

Table 4. *Chemical agents potentially present in municipal wastewater*

Group of chemical	Chemical agent	Effect/observations
Easily biodegradable organic compounds	Proteins, carbohydrates	Loss of dissolved oxygen from aquatic ecosystems (anoxic conditions). Generation of hydrogen sulphide and methane gases
Hardly biodegradable organic compounds	Greases, phenols, cellulose, lignin and similar	Residual COD, loss of dissolved oxygen
Xenobiotic compounds	Several formulations of synthetic compounds	Bioaccumulation, toxicity, interferences with the life cycle
Nutrients (macro)	Nitrogen, phosphorus, potassium	Eutrophication, loss of dissolved oxygen, toxic effects
Nutrients (micro)	S, Fe, Mn, Cu, Zn, Co, B	Plant toxicity
Metals	Hg, Pb, Cd, Cr, Cu, Ni, Al, Fe, Mn, Zn	Bioaccumulation, toxic effects
Radioactive compounds		(See Table 5)
Dissolved salts	Chlorides, sulphides, nitrates	Effect on agricultural uses, risk for human health (nitrates)
Other chemical compounds (organic and inorganic)	Pesticides, plaguicides, organic halogens, residual chlorine/disinfection by-products	Carcinogenic, teratogenic and/or mutagenic effects

From Metcalf & Eddy, Inc. (1991); Rowe & Abdel-Magid (1995) and Crook (1998).

Table 5. *Physical pollution potentially present in urban wastewater*

Physical parameter	Agent	Effect/observations
Radioactive compounds	Radon, radioactive isotopes	Bioaccumulation, toxic effects
Residual heat (thermal pollution)	Water temperature above normal level	Effects on aquatic life Reduction of dissolved oxygen concentration
Odour	Gases (hydrogen sulphide, mercaptans, cyanide, ammonia)	Nuisance effects on human health
Colour	Natural metallic ions (iron oxides, manganese oxides), humic acids, lignin derivatives	Aesthetic and nuisance effects
Taste	Phenols, dissolved salts, disinfection by-products	Effects on human health (see Table 4)
Total solids: suspended solids (settleable, non-settleable), filterable solids (colloidal, dissolved)	Suspended solids (organic matter, clay, silt) Filterable solids (organic matter, dissolved salts)	Diverse effects (see Table 4)

From Metcalf & Eddy, Inc. (1991) and Rowe & Abdel-Magid (1995).

Usually, reclaimed wastewater quality is established independently from other considerations, using standards. Standard figures depend on several concepts such as (Salgot & Angelakis 2001): economic and social circumstances, legal capacity from different entities and implicated administrations, human health/hygienic degree (endemic illnesses, parasitism), technological capacity, previously existing rules and/or criteria, crop type, analytical capacity, risk groups possibly affected, technical and scientific opinions and other miscellaneous reasons.

Three types of factors can be distinguished: technological (analytical, treatment methods, and capacity, knowledge); legislative and economical (criteria, socio-economical, legal competence); and health-related (sanitary state, diseases, risk groups). Standards and quality regulations have been a matter of discussion among scientists, health and legislation officers, and engineers, because of the parameters to be controlled. Much controversy has been aroused among research teams and regulating bodies on the quality that reclaimed water must meet before its reuse with an acceptable degree of risk.

Existing wastewater reuse regulations have been based traditionally on biological quality considerations, and only during the last few years have chemical and toxicological concerns appeared.

At the European level, the only reference to reuse is article 12 of the European Wastewater Directive (91/271/EEC): 'Treated wastewater shall be reused whenever appropriate'. In order to make this statement reality, common definitions of what is 'appropriate' are needed. A complete revision of the guidelines and regulations can be found in Salgot & Angelakis (2001), although several modifications appeared after this revision. The most important changes have been the appearance of a new edition of the USEPA *Guidelines for Water Reuse* in 2004, and the WHO drafts on reuse (2002 and 2005, for Europe and the world, respectively).

The USEPA (2004) guidelines refer to planned groundwater recharge, and indicate the rules in several states, mentioning specifically California, Florida, Hawaii and Washington.

California and Hawaii do not specify required treatment processes and determine requirements on a case-by-case basis. The California and Hawaii health services departments base evaluation on all relevant aspects of each project including treatment provided, effluent quality and quantity, effluent or application spreading area operation, soil characteristics, hydrogeology, residence time and distance to withdrawal. Hawaii does require a groundwater monitoring programme.

Washington has extensive guidelines for the use of reclaimed water for direct groundwater recharge of non-potable aquifers. It requires Class A reclaimed water defined as oxidized, coagulated, filtered and disinfected. Total coliform content is not to exceed 2.2/100 ml as a 7-day median and 23/100 ml in any sample. Weekly average biological oxygen demand (BOD) and total soluble solids (TSS) limits are set at 5 mg/l. Turbidity is not to exceed 2 NTU as monthly average and 5 NTU in any sample. Groundwater monitoring is required and is based on reclaimed water quantity and quality, site-specific soil and hydrogeological characteristics, and other considerations. Washington also specifies that reclaimed water withdrawn for non-potable purposes can be withdrawn at any distance from the point of injection and at any time after direct recharge.

Florida requires that TSS do not exceed 5.0 mg/l in any sample, achieved prior to disinfection, and that the total nitrogen in the reclaimed water be less than 12 mg/l. Florida also requires continuous on-line monitoring of turbidity; however, no limit is specified.

Concerning Europe, Brissaud (2003) describes the criteria for health-related guidelines in groundwater recharge with recycled municipal wastewater. He describes four different cases.

Direct recharge for indirect potable use. If the recharge is direct, then the injected water should be potable. Moreover, the water injected should also be treated to prevent clogging around the injection wells, long-term health risks linked to mineral and organic trace elements and the degradation of injected water quality into the aquifer.

Indirect recharge for indirect potable use. The quality of infiltrated water may be dramatically improved when percolating through the vadose zone, thanks to several processes, which include retention and oxidation processes. These processes affect organic matter, nutrients, micro-organisms, heavy metals and trace organic pollutants, among others. When transfer through the vadose zone is part of the treatment intended to bring injected water up to potable water, a case-by-case approach should be highly recommended.

Direct recharge of non-potable aquifers. The recycled water should have been upgraded to the standards and limits required for the intended applications. A high degree of treatment is also necessary to make the injection sustainable. Suspended solids and organic matter should have been drastically reduced to avoid clogging around the injection wells.

Indirect recharge of non-potable aquifers. This requires a less treated injectant and is easier to implement. SAT is an appropriate treatment to meet the required water quality, provided it is properly designed and managed.

The discussion on how strict the standards must be is still on going and no agreement has been achieved so far among the different points of view. Nevertheless, it seems that the WHO is promoting a new perspective, using DALYs (disability-adjusted life years) instead of numeric standards (G. Kamizoulis pers. comm. 2005). This new way seems promising, but needs a better understanding, focused research and a good communication policy.

There are other ways to progress, like the development of good reuse practices (GRP) or the risk analysis approach, which is connected with the DALY approach.

In several countries, such as Spain, so-called 'autocontrol' procedures – using tools derived from classic industries, mainly the HACCP (hazard analysis and critical control points) approach – are gaining momentum, as described below. Again, this is related to risk developments.

Reclamation technologies

Nowadays, there are no technological problems to reaching any desired water quality when wastewater is used as a raw material. Nevertheless, there is a clear relationship between the degree of treatment needed and the economy of the whole procedure, and additionally the question of who must pay for the treatment. So it appears that the main problem is economic. Another consideration is that there is a frontier between what is a public service (wastewater treatment) and the manufacture of a product (reclaimed water). The end-product (reclaimed water) must compete in an open market (theoretical) with the conventional water resources. Both conditions (economy and technology) must be improved with adequate research, development and innovation (R + D + i). Two main lines of R + D + i are required (Salgot & Vergès 2003): (a) new management tools for water resources management which include risk analysis and management; (b) new wastewater reclamation technologies, capable of reducing the treatment costs to a level compatible with the socio-economic characteristics of the reuse site.

All this raises several questions on the level of reclamation that a country can accept from the economy and risk points of view. There is not a clear solution, and several discussion points arise (Salgot & Vergès 2003). First to consider is the socio-economic context in which the reuse takes place: where wastewater treatment is a brand new, or nearly new, element in the anthropic water cycle, the taxation effort supported in order to fulfil the legal requirements for wastewater treatment has usually been important. It seems difficult in this context to implement new taxes to reclaim secondarily treated wastewater.

The second point is that several users obtain a benefit from the reclaimed wastewater. This aspect is included in the main perception that irrigated agriculture is not using water resources efficiently. In many places, water is not paid for, or is paid for at an inadequate price. This low price does not promote the saving or efficient use of water. Other users (e.g. golf courses) have been the subject of heavy criticisms from a number of stakeholders on the grounds of suspected wasteful use of water resources. Nevertheless, such users usually pay for the resources and the comparative economic benefits for society of 'sumptuary' users and the 'classic' users should be examined and discussed.

A third point, true in mid-2005, that should be stated. The water authorities around the Mediterranean promised that wastewater reuse will be promoted and backed up. Yet, this promise has not exactly been fulfilled, with a few honourable exceptions (e.g. Tunisia and Israel).

Finally, the wastewater reclamation technologies usually implemented have mainly been intensive, using large amounts of energy. It seems that from the point of view of the ecological impact, it would be more logical to use, especially for small- and medium-sized facilities, extensive, less energy-consuming technologies.

Brissaud (2003) indicates that it is common agreement that recharge should not lead to additional or supplementary treatment after withdrawal to meet the standards related to the intended water applications. Meeting the standards at the point of use is not enough; qualitative requirements have to be satisfied within the aquifer.

Asano & Cotruvo (2004) indicate that four water quality factors are particularly significant in groundwater recharge with recycled wastewater: microbiological quality, total mineral content (total dissolved solids), presence of heavy metal toxicants, and the concentrations of stable and potentially harmful organic substances.

The same authors establish that some basic questions that need to be addressed include: What treatment processes are available for producing water suitable for groundwater recharge? How do these processes perform in practice at specific sites? How does water quality change during infiltration–percolation and in the groundwater zone? What do infiltration–percolation and groundwater passage contribute to the overall treatment system performance and reliability? What are the important health issues to be resolved? How do these issues influence groundwater recharge regulations at the points of recharge and extraction?

Table 6. *Best available technology conditions*

Wastewater characteristics
Size/type of the facility
Existing rules and regulations
Available technology
Landscape integration
Economy (town, county, country)
Social acceptance
Centralisation vs. decentralisation
Desired final quality of water (following rules and regulations)
Reclaimed water reuse possibilities
Political decisions

What benefits, problems and successes have been experienced in practice?

In all cases the 'best available technology' (BAT) concept needs to be applied. The BAT (Table 6) is used considering all possible options with respect to technology, economy and social aspects. It starts with a systematic analysis of the technical, economic, environmental and financial factors necessary to select a cost-effective wastewater management plan (Metcalf & Eddy, Inc. 2003).

BAT procedures describe, at the planning levels, the technology most suitable for a given location or town. Sometimes, this approach is described as BATEA (best available technology economically achievable), BPWTT (best practicable technology currently available) or BPCTCA (best practicable control technology currently available) for industries (Rowe & Abdel-Magid 1995).

Risk-related aspects

When reusing reclaimed wastewater it is obvious that hazards will appear (Salgot 2002). The hazard is due to the presence in the reclaimed wastewater of several contaminants, pathogens and chemicals. Pathogens include protozoa, bacteria and viruses (Table 3), while chemicals include the simplest (e.g. nitrates) to the more complex molecules, e.g. trihalomethanes (Table 4).

Pathogens usually generate infectious illnesses that may or may not be apparent (clinical and non-clinical), and usually appear in the short term, while chemicals are related to toxicity, either acute or long term. It is obvious that the better the reclaimed water quality, the less the risk related to both types of contaminant. When dealing with risk, two steps should be considered: assessment and management. It seems basic to have a guide on how to determine the risk assessment. Nevertheless, this does not exist in full, and is nowadays an object of intensive research (Salgot & Pascual 1996; Vergés & Salgot 2002).

Several tools are being developed and becoming available for hazard or risk assessment. The best known is the HACCP system. This system, developed for the food industry, focuses on the detection of relevant control points, whose establishment increases safety while reducing the costs through a better use of analytical work. HACCP is conducted considering seven main points as indicated in Table 7.

HACCP procedures must theoretically be performed for all planned reuses. In specific cases, as for bathing water or shellfish cultivation, it is necessary to establish contacts with other authorities implicated, e.g. health departments. In any case, the risk is related to the presence of a contact between the target (man, animals, plants) and reclaimed wastewater.

The solutions to reduce risks to an acceptable amount are: reduction of contacts between the pathogen and the target organism; and good reclaimed water quality. The reduction of contact could be reached through what is called good reuse practices, intended partly to reduce the possible contacts. While considering that 'contact' in a broad sense could imply either direct or indirect contact (e.g. skin or mucous membrane contact, aerosols entering the respiratory tract), any contact reduction will imply risk reduction. Good reclaimed quality is obtained using standards, which when enforced must guarantee a quality of water theoretically good enough to reduce the risk to an acceptable level.

Groundwater recharge

The definition of artificial recharge is: 'the techniques or operations which have the main objective of allowing a better aquifer management by increasing the water resources and creating reserves, by means of a direct or indirect intervention in the natural water cycle' (Custodio & Llamas 1996). Groundwater recharge with wastewater is always artificial recharge.

Dillon (2005) describes the management of aquifer recharge (MAR) as the intentional banking and treatment of water in aquifers. The same author discusses the meaning of 'artificial recharge' indicating that the adverse connotations of 'artificial', in a society where community participation in water resources management is becoming more prevalent, suggests that it is time for a new name. The old name implied that the water was in some way unnatural. Managing recharge is intentional as opposed to the effects of land clearing, irrigation and installing water mains, where recharge increases are incidental. MAR has also been called enhanced recharge, water banking and sustainable underground storage (Dillon 2005).

Table 7. *Main points to be developed in HACCP systems*

Item	Observations
1. Conduct a hazard analysis	Needs to be developed for wastewater treatment and reclamation facilities.
2. Identify the critical control points (CCP)	Not clearly defined if it should be in the entire wastewater management (i.e. from the point of wastewater generation to the point of use and beyond: fate of generated products like vegetables).
3. Establish target levels and tolerances	It is extremely important to set not only the percentage of samples that cannot to comply, but also the tolerances (maximum deviation allowed).
4. Establish a monitoring system	Needs to be issued taking into consideration the analytical capacity (e.g. complex chemicals) and the cost of the whole control procedure.
5. Establish corrective actions	When problems are detected and all the information is available, due to the risk to be generated, corrective actions should be implemented. In the meantime it should be considered if the facility has to be stopped and reclaimed wastewater is not to be served.
6. Establish documentation	Necessary to identify all problems during the lifespan of the project.
7. Establish verification procedures	The whole procedure must be analysed in a continuous and periodic way.

Among the main objectives of recharge, several are related to recharge with reclaimed water: supplement groundwater resources available; reduce or eliminate (even increase) groundwater level drop; compensate natural recharge lost by human activities; improve coastal aquifer situation; use aquifers as a storage medium for water, instead of using surface facilities; improve joint uses of surface and groundwater; avoid movements of bad quality waters inside the aquifer; increase good quality water availability, through blending; increase leaching of salts and other contaminants; use the soil aquifer system for water treatment; reclaim treated wastewater, store it, and complete the treatment using the soil/aquifer system; reduce, mitigate or even eliminate subsidence caused by over-exploitation of groundwater; compensate negative effects of hydraulic and civil works; maintain flows in a stream or levels in lakes during low waters; and use the aquifer as a water transportation media.

There are basically two types of groundwater recharge/application: on-surface (indirect, over the soil); and deep injection (direct injection onto the aquifer). These are several techniques, described in Table 8 and depicted in Figure 1.

Surface or deep works can be employed for recharge, as summarized in Table 9. In both cases, the reduction of the recharge capacity (clogging) can be attained by soil surface alterations, addition of too much suspended matter, or biological activity. The causes of clogging are usually the presence of suspended solids and/or gas bubbles in the recharge water or bacterial growth in the well and surrounding it. Other causes can be chemical precipitation in the water, soil and well, clay swell or dispersion, and soil structure erosion and subsequent aquifer obstruction. Reduction of recharge capacity, social acceptance, pollution of aquifers used for potable water supply, hazard and risk increases and extraction abuses are the main resulting problems.

Once reclaimed water reaches the aquifer, there are several phenomena that can occur, namely: organic matter reduction; water odour and taste correction; and adsorption of some organic matter compounds. Reactions in saturated media are much slow than in non-saturated ones. It should also be noted that the bacterial flora in groundwater is usually scarce.

In general, groundwater recharge presents several disadvantages:

- huge (surface) application areas are needed;
- if reclaimed water is injected, energy is necessary;
- recharge is increasing the groundwater pollution risks;

Table 8. *Major techniques of groundwater recharge*

Technique	Acronym	Description
Aquifer storage and recovery	ASR	Injection of water into a well for storage and recovery from the same well
Aquifer storage transfer and recovery	ASTR	Injection of water into a well for storage and recovery from a different well, generally to provide additional water treatment
Bank filtration		Extraction of groundwater from a well or caisson near or under a river or lake to induce infiltration from the surface water body thereby improving and making more consistent the quality of water recovered
Dune filtration		Infiltration of water from ponds constructed in dunes and extraction from wells or ponds at lower elevation for water quality improvement and to balance supply and demand
Infiltration–percolation	IP	Infiltration and percolation of partially treated wastewater in sand formations for further treatment. Water may or may not be extracted later. Can be considered as a variant of SAT.
Infiltration systems		Ponds constructed usually off-stream where surface water is diverted and allowed to infiltrate (generally through an unsaturated zone) to the underlying unconfined aquifer
Percolation tanks		A term used in India to describe harvesting of water in storages built in ephemeral wadis where water is diverted and infiltrates through the base to enhance storage in unconfined aquifers and is extracted down-valley for town water supply or irrigation
Rainwater harvesting		Roof runoff is diverted into a well or a caisson filled with sand or gravel and allowed to percolate to the water-table where it is collected by pumping from a well
Soil aquifer treatment	SAT	Reclaimed water is intermittently infiltrated through infiltration ponds to facilitate nutrient and pathogen removal in passage through the unsaturated zone for recovery by wells after residence in the aquifer
Soil aquifer plant treatment	SAPT	If plants are implemented in SAT or similar systems, there is an additional improvement of the quality of the water applied because of the action of plants (e.g. further removal of nutrients)
Sand dams		Built in wadis in arid areas on low permeability lithology, these trap sediments when flow occurs, and following successive floods the sand dam is raised to create an 'aquifer' which can be tapped by wells in dry seasons
Underground storage		In ephemeral streams where basement highs constrict flows, a trench is constructed across the streambed keyed to the basement and backfilled with low-permeability material to help retain flood flows in saturated alluvium for stock and domestic use
Stream releases		Dams on ephemeral streams are used to detain floodwater and uses may include slow release of water into the streambed downstream to match the capacity for infiltration into underlying aquifers, thereby significantly enhancing recharge

Modified from Tuinhof & Heederik (2003).

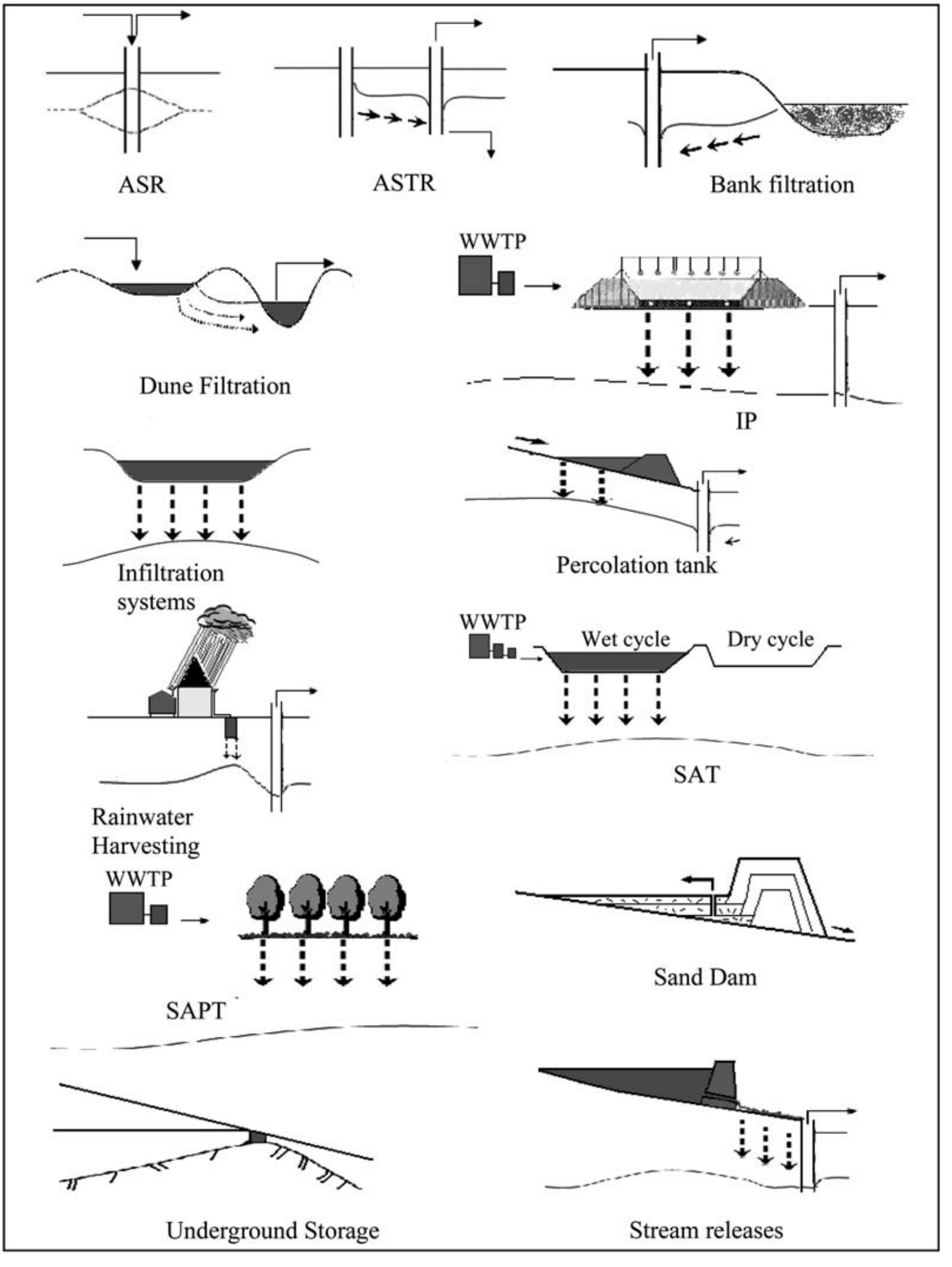

Fig. 1. Major techniques of groundwater recharge (modified from Dillon 2005). ASR, Aquifer Storage and Recovery; ASTR, Aquifer Storage Transfer and Recovery; WWTP, Wastewater Treatment Plant; IP, Infiltration Percolation; SAT, Soil Aquifer Treatment; SAPT, Soil Aquifer Plant Treatment.

Table 9. *Works for groundwater recharge*

Surface works for recharge	Deep works for direct injection
Lagoons or ponds	Wells
Channels, trenches and furrows	Absorption or diffusion wells
Pits	Drains or galleries in the bottom of a well
Areas for surface infiltration	Trenches filled with gravel reaching the phreatic level
River bed actuations	Natural sinks, ravines or fissures in karst areas

- not all water added can be recovered;
- there is great surface demand for the system being operative;
- instant demands cannot be satisfied (low rate answer);
- there are problems with the water legal status.

Direct aquifer recharge with reclaimed water. Direct recharge means that the reclaimed water is introduced directly into the aquifer through injection wells. Rowe & Abel-Magid (1995) make several statements and describe direct recharge indicating that treated wastewater is pumped under pressure directly into the groundwater zone, usually into a well-confined aquifer. WEF & AWWA (1998) also describe recharge into wells by using gravity flow.

Groundwater recharge is practised, in most cases, where groundwater is deep or where the topography or existing land use makes surface spreading impractical or too expensive. This method of groundwater recharge is particularly effective in creating freshwater barriers in coastal aquifers against intrusion of seawater.

Rowe & Abel-Magid (1995) also state that locating the extraction wells at as great a distance as possible from the injection wells increases the flowpath length and residence time of the recharged groundwater, as well as mixing the recharged water and the other aquifer contents.

In any case (State of California 1978), it seems clear that wells are the least desirable method of groundwater recharge, largely because of problems of pore clogging, well silting, air entrainment, bacterial and algal growths, and deflocculation caused by reaction.

Indirect aquifer recharge. Indirect recharge is feasible only for unconfined aquifers; the recharged water is spread on the land – over irrigation – or on infiltration basins or by other means, and infiltrates through the vadose zone down to the water table. The unsaturated layer (including the soil) behaves as a filter and a natural reactor, providing an additional treatment and allowing the percolating water quality to greatly improve.

On-surface irrigation has the advantage of employing the treatment capacities of the soil, which constitutes an additional barrier, while direct injection has the main disadvantage of introducing water directly into the aquifer. Direct injection is considered more hazardous than indirect (Table 10).

Another form of indirect groundwater recharge with reclaimed water is the ‘collateral’ recharge which happens when this water is used for irrigation. The excess water which infiltrates and later percolates through the soil/subsoil can reach the aquifer. For irrigation to be sustainable, the chemicals (especially salts) carried with the reclaimed water must not be allowed to accumulate in the root zone of the soil. Then, an extra amount of water (leaching water) needs to be applied in order to control the salinity.

Practical application of groundwater recharge

Many groundwater recharge applications (Pyne 1995) can be easily identified; however, for simplicity, only the basic applications are briefly described here (Anderbouhr & Pérez-Paricio

Table 10. *Groundwater recharge uses of reclaimed water*

Type of use	Specific use	Observations
Groundwater recharge	Direct recharge	Potable quality if treated for domestic uses Pre-potable if other source
	Indirect recharge	Through soil formations

Modified from Brissaud *et al.* (2005).

1999). More than one of the possibilities mentioned can be true for the same application site.

Aquifer water levels recovery. This is by far the most typical cause of groundwater recharge: if the extraction patterns or the natural losses caused by human activities (e.g. river beds channelling, water abstractions, civil works) overwhelm natural recharge, the solution is supplying additional water flows to the aquifer.

Minimization of seawater intrusion. Injection–extraction barriers or just injection are used. As a consequence, the freshwater level can increase and the seawater–freshwater front can become stabilized.

Reclaimed water storage. The aquifer can be used for storage of reclaimed water or as a management facility to deal with uneven demands.

Storage facility for tap water. This function is usually performed for additional storage of tap water near the site of production and use. Although it is not fully accepted, the capacity of the aquifer to be used for advanced treatment to reclaimed water should not be forgotten.

Fighting against subsidence in areas where excess groundwater extraction is common.

Improvement of water quality by passage through the aquifer. Two possibilities are indicated: (1) when the quality of groundwater is defective (e.g. brackish) reclaimed water can improve the quality; and (2) the water which enters the aquifer receives additional treatment.

Tap water generation. This is usually performed when trying to improve 'natural' water that has been polluted. Several experiences have been described around the north of Europe (Paris, Dortmund, The Netherlands) and near Toulouse (south of France).

Wastewater reclamation. As described, natural systems can improve the quality of reclaimed wastewater.

Environmental improvements. The negative impacts of dams can be reduced, while not affecting the natural water bodies.

Technical and real-time problems

If water of a different quality from that existing in the aquifer is introduced, technical problems can appear arise, which are mainly related to the following.

Clogging by reduction of the surface infiltration capacity. Chemical, physical and biological problems appear. There are five main processes affecting the permeability on the surface: suspended solid retention; bacterial and algal growth; chemical precipitation; gas formation; and settled material compaction.

Pathogen presence. When viruses and bacteria enter the aquifer, a negative impact is produced, although there is discussion on the real origin of such organisms. There is a relationship between the hydraulic residence time and the elimination of micro-organisms in the aquifer; it means that long residence times favour the decontamination of water.

Organic matter and chemicals. The elimination of organic matter and chemicals in the aquifers is complicated and difficult, due to the lack of oxygen present. Reduction is mainly performed during passage through the vadose zone and not in the aquifer.

Social problems

Groundwater recharge needs to be accepted by society, which means that an adequate effort must be performed by the authorities to make the practice understandable to people. It is recommended that pilot studies are developed and demonstration facilities used before undertaking a full-scale procedure.

Tools for public acceptance are available to contribute to the success of the practice. Full and adequate planning should be prepared and implemented.

The economy of any project is paramount to guarantee the success, as well as adequate legislation to be followed by the managers of the facility.

Environmental problems

The definition of environmental problems associated with groundwater recharge using reclaimed water can be described using the environmental impact assessment tools. First of all, it is necessary to state that the distinction between 'environmental impact' and 'change in an environmental attribute' is that changes in the attributes provide an indication of changes in the environment. In a sense, the set of attributes must provide a model for the prediction of all impacts. The steps in determining environmental impact are: (1) identification of impacts on attributes; (2) measurement of impacts on attributes; and (3) aggregation of impacts on attributes to reflect impact on the environment.

The conditions for estimating environmental impact are measurement of attributes with (positive scenarios) and without (zero scenario) the project or activity under consideration at a given point in time. Consideration of the potential for impact if no action is taken, that is, maintaining the status quo, is called the 'no-action alternative' (the zero scenario, in other words).

While 'affected environment' describes the condition of the environment when the action is proposed to take place, the environment will not remain static over time. If a hypothetical 'proposed action' were implemented, the impact would be the degree of change over time if the action were taken, compared to the condition of the environment over the same span of time if the action were *not* taken. It should be noted that the impact would not be the proposed action over time compared to the ambient environment prior to the point of action.

For other alternatives, the comparison can be to the impacts of the proposed action or all alternatives can be compared to the no-action alternative. Both approaches are used; the only caution is to be consistent throughout the analysis and explain clearly which approach is used.

A difficulty is that data for a 'with activity' and 'without activity' projection of impacts are difficult to obtain, and results are difficult to verify.

In respect to the practical application, environmental impacts of recharge must be related to soil, water, air, flora and fauna, and partly to social, cultural and economic impacts. It should be noted that impacts can be positive or negative with respect to the environment. In Table 11 there is a description of both types of impact.

Control needs

Due to the fact that groundwater recharge is identified as being more hazardous than other reclaimed water uses (e.g. agricultural irrigation), there is the need to strictly control the whole procedure. There are two types of control: analytical, related to the quality of water that will enter into the aquifer formation; and legal, the control of the later uses of the recharged water (recovered water).

It is also necessary to establish differences, with respect to the degree of risk, between direct and indirect recharge. Direct recharge (i.e. making the water enter directly into the aquifer) generates more hazards than indirect recharge (passage through the soil and subsoil). There are usually rules and regulations on this (see above).

Apart from the controls stated in the regulations, the HACCP approach indicates the points where it is essential to perform controls (e.g. the effluent from the reclamation facility, the point of application or the point of recovery of the recharged water).

Apart from the pure controls, marked by rules and regulations, there is the need to establish the periodicity of sampling procedures. This can be stated by the standards or by good reuse practices. This control has a cost that must be fully considered when making economic calculations on groundwater recharge.

Table 12 shows a sampling pattern suggested by a draft version for a Spanish decree on reuse.

Recharge in the south of Europe: the potential

It is necessary to distinguish between the existing recharge, (mainly) non-planned recharge and the possibilities for future development of the technique.

The current situation

As explained, there are two types of recharge using reclaimed wastewater: intentional and non-intentional. When dealing with wastewater reuse the terms are planned and unplanned reuse.

Unplanned reuse is usual in Mediterranean rivers when treated wastewater is disposed of into rivers or lakes. Sometimes, the unique water flowing in most streams is wastewater with or without treatment. Due to the relationship between aquifer and surface flows, the recharge is produced automatically.

Planned reuse is specifically intended to increase the amounts of water present in the aquifer. There are few cases in the south of Europe, because of the lack of rules and regulations on reuse and the fears of stakeholders to support this practice.

Several practical examples of groundwater recharge are described in Table 13.

The future

Bouwer (2006) states that artificial recharge is expected to become increasingly necessary in the future as growing populations will require more water, and as more storage of water will be needed to save water in times of water surplus for use in times of water shortage. The traditional way of storing water has been with dams. However, good dam sites are becoming scarce. In addition, dams have various disadvantages, such as evaporation losses (about 2 m/year in warm, dry climates); sediment accumulation; potential or structural failure; increased malaria, schistosomiasis and other human diseases; and adverse ecological, environmental and socio-cultural effects. New dams are becoming more and more difficult to build because of high cost and public opposition.

In some areas of the Mediterranean, farmers have a policy to build small pools to accumulate water for several days, in case they experience problems with the usual supplies. Farmers' associations could also have the capacity to implement greater reservoirs, from 0.5 to several million cubic metres. In both cases, the management of such systems is related to the energy cost (e.g. pumping by night, when the energy is cheaper)

Table 11. *Groundwater recharge environmental impacts*

Matrix affected	Direct recharge		Indirect recharge	
	Positive impacts	Negative impacts	Positive impacts	Negative impacts
Soil	NA	If excess water is applied, soils can become inundated with uprising water	Increases the amount of water available for plants	Possible anaerobic conditions if excess water is applied. Water losses due to evapotranspiration
Water	Increases the available groundwater. Can positively affect the quality of groundwater. Reduction of seawater intrusion. No subsidence	Can negatively affect the quality of the groundwater	Increases the available groundwater. Can positively affect the quality of groundwater. Reduction of seawater intrusion. No subsidence	Possible pollution (organic matter, micro-organisms)
Air	NA	NA	Increase of evapotranspiration (air temperature effects)	Aerosol formation if water is applied through sprinkling or similar
Flora	NA	NA	Increases water-related species on surface	Aerobic species can suffer from excess water
Fauna	Amount of water increase in water bodies related to aquifers	Reduced amounts of water on surface	More water available on surface	Diseases associated with reclaimed water
Social and cultural	Increases the water resources available	Recovered water could not be accepted	Increases the water resources available	Recovered water could not be accepted
Economy	Increases economic development (more water available)	Additional treatment and control costs	Increases economic development (more water available)	The quality of water can be reduced

NA, not applicable.

Table 12. *Sampling frequencies as established by the draft Spanish Decree on reuse, June 2005*

Parameter	Sampling frequency/number of samples per year			
	Sampling Pattern I	Sampling Pattern II	Sampling Pattern III	Sampling Pattern IV
Nematode eggs	Weekly/52	Fortnightly/26	Fortnightly/26	Monthly/12
E. coli	Three times per week/104	Twice per week/104	Weekly/52	Monthly/12
Suspended solids	Daily/365	Weekly/52	Weekly/52	Monthly/12
Turbidity	Daily/365	Daily/365	Twice per week/104	Weekly/52
Legionella spp.	–	Monthly/12	Monthly/12	–
Taenia saginata and *T. Solium*	–	–	Fortnightly/26	–
Nitrogen – total	Weekly/52	–	–	–
Phosphorus – total	–	–	–	Monthly/ 12
Potentially toxic elements	Monthly	Quarterly	Quarterly for aquaculture and for agriculture when the maximum allowable concentration of sludge is exceeded	Quarterly when the maximum allowable concentration of sludge is exceeded

Table 13. *Examples of groundwater recharge of reclaimed water*

Type	Description	Location	Comments	Reference
Surface spreading	Health studies	Whittier Narrows, Los Angeles County, California, USA	Blended water (storm and river) is used	WEF & AWWA (1998)
ASR	Testing the technical, economic and environmental viability of seasonal storage in the aquifer	Bolivar, Adelaide, Australia	Final horticultural irrigation	Dillon *et al.* (2005)
SAT	Irrigation	Porquerolles Island, France		Rodier & Brissaud (1989)
	Excess irrigation and infiltration.	Pla de Sant Jordi, Majorca Island, Spain	In operation for more than 30 years	Lopez –García (2003)
	Infiltration (residence time >6 months) and extraction	Dan region, Israel	In operation for more than 20 years	Chikurel (2004)
	Subsurface storage for reuse on high-valued horticulture and viticulture	Alice Springs, Australia	Prevent the weather overflows	Dillon *et al.* (2005)

and not to water management, because water is not usually paid for by the farmers.

Apart from the above-mentioned features, dams are physically located in a definite position, and a distribution network needs to be built. It is possible to use an aquifer as a distribution tool, if the area where water is to be used corresponds to the direction of groundwater flow. If comparing the energy uses of surface and underground (aquifer) distribution systems, there is the need to calculate pumping expenses for moving water from one site to another.

With the increasing concern over water management in the Mediterranean region, caused by the structural lack of water aggravated by drought episodes, private storage for small volumes of surface water will be increasingly contested.

The integrated management of water resources requires control of all the water cycle components, including irrigation, groundwater, reclamation and reuse. Therefore, alternative solutions to the ineffective small surface storage are needed. In addition, fighting against seawater intrusion or movement of brackish groundwater will be needed in many Mediterranean areas.

All the mentioned circumstances lead to the need for further studies on the use of aquifers to manage non-conventional water resources. At the same time, planned water reuse is expected to become increasingly applied, for several reasons in relation to groundwater recharge: additional wastewater treatment; increasing the storage capacity without evaporation losses; maintaining groundwater levels; possibly reducing energy use; and transporting water to the specific sites of use.

On the coastline, there are additional reasons: reducing seawater intrusion and storage of water (no dams are possible).

Conclusions

In the present conditions of water demand in the south of Europe, integrated management of water resources becomes increasingly important. In this context, groundwater recharge with reclaimed water can play an important role, especially in southern European countries.

Recharge is paramount for water reuse schemes, because it: improves reclaimed water quality; increases storage capacity and the amount of water available; reduces evaporation losses; and allows the reduction of seawater intrusion, among other positive impacts. Negative impacts include the possible mixture with water to be used for tap purposes, flooding if excessive amounts of water are applied or undue appropriation of resources.

Adequate studies are needed before implementing groundwater recharge schemes. Such studies must include geological, geochemical, hydrological, biological and engineering features. Risk-related, environmental impact and other basic studies are also a necessity.

Social aspects, including communication procedures and economic feasibility, are paramount for the success of the practice.

Inexpensive tools for water management, wastewater reclamation and reuse will be needed in the light of changes that the water cycle will experience due to the climatic change.

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Adaptation to climate change: strategies for sustaining groundwater resources during droughts

B. S. SUKHIJA

National Geophysical Research Institute, Hyderabad-500 007, India (e-mail:bssngri@rediffmail.com)

Abstract: Extreme climate conditions are expected in the twenty-first century in the form of higher maximum temperature (with more hot days) resulting in frequent droughts. The continents of Africa and Asia are anticipated to be extremely vulnerable to droughts. In the impending extreme climate conditions, humanity's sustenance hinges on groundwater as it forms the world's largest freshwater resource. Adaptive and mitigation measures entail well planned strategies for sustained groundwater through extreme climate conditions including droughts. In this paper two such strategies are discussed to overcome the problem of droughts: (i) artificial groundwater recharge using percolation ponds; and (ii) identifying and characterizing deep aquifers resilient to droughts through detailed geophysical, hydrogeological and isotopic studies.

Percolation ponds act as artificial recharge structures which are constructed across monsoon streams with the purpose of harvesting surface runoff caused by monsoon streams. Conventional and tracer methods were developed in India to determine how effective these artificial recharge structures could be. From studies carried out on percolation ponds located in diverse geological formations such as granites, basalts and sandstones, it was concluded that these structures are quite useful for overcoming droughts in semi-arid and arid regions, and it was demonstrated that the role of geology outweighs the effect of climate on such structures. It has been shown that in a similar climatic environment, the percolation ponds in sandstones were far more efficient (efficiency 60%) than those in basalts (efficiency 20–30%).

Recently it has been realized that certain deep aquifers can yield a good quantity and quality of water even during extreme climate events. The Neyveli aquifer in southern India has been demonstrated to be such a representative aquifer for mitigation of droughts. Very extensive and intensive hydrogeological and isotopic studies on the aquifer revealed that the aquifer has distinct characteristics, namely: (i) distinct recharge area; (ii) extensive groundwater regime with high degree of recharge rate; (iii) wide span of radiocarbon ages from Modern to >30 000 years BP indicating modern as well as palaeorecharge; and (iv) minimal changes in groundwater quality despite very heavy and continuous withdrawal during the last four decades. All these criteria provide the necessary ingredients for drought resilient aquifers which can be used to identify similar aquifers elsewhere in the world.

Climate change has always had a profound impact on human society. History has recorded flourishing human society and economy in case of moderate climate and vanishing civilization in the case of extreme climatic conditions. In the context of climate extremes, the responsive sensitivity of human society and its adaptive capacity are the two factors that determine the vulnerability and the degree to which human society succumbs to the adverse effects of climate change. The adaptiveness or adaptive capacity is the ability of a system to adjust to climate change including climate variability and extremes. Such extremes can be expected in every subsystem of the Earth system, including water cycle, sea levels, ocean circulation, ice cover, air quality and even biodiversity. This could lead to altered oceanic circulation, vertical mixing, and reduced sea-ice cover. These changes have a direct effect on society in coastal regions by way of submergence of coastal areas, erosion of shores and associated habitat, increased salinity in estuaries and freshwater aquifers, changes in sediment and nutrient transport. The coastal ecosystem is particularly at risk, for example saltwater marshes, mangrove ecosystem, coastal wetlands, coral reefs, sea-atolls and river deltas. In extreme climatic situations, human welfare hinges mainly on demand and supply of water, food and energy. In the global scenario, climate change will have a significant impact on sustainable development in different parts of the world and in different measures. Human society will rise to meet the challenges induced by climate variability by appropriate adaptation and mitigation to utilize the natural resources to the fullest benefit of society. In cases where human endeavor fails to harness the natural resources then the gap between the 'haves' and the 'have-nots' widens, which further acts as a vehicle for escalating human suffering.

From: DRAGONI, W. & SUKHIJA, B. S. (eds) *Climate Change and Groundwater*. Geological Society, London, Special Publications, **288**, 169–181.
DOI: 10.1144/SP288.13 0305-8719/08/$15.00

It is anticipated (McCarthy *et al.* 2001) that there are very likely to be extreme climate phenomena during the twenty-first century in the form of higher maximum temperature with more hot days and heat-waves over nearly all land areas, and more intense precipitation events leading to increased flood runoff, resulting in frequent droughts and floods. It is anticipated that as a result of increased temperature there will be widespread and accelerated glacier retreat, thus shifting the timing of stream flow and increasing inflows in certain regions. Consequently the water cycle will become intensified with more frequent extreme conditions including increased evaporation. Moreover, perceptible changes will take place such as high variability in quantity of water in time and space, and water quality degradation due to increased water temperature. The adverse effect of increased water temperature shows up on biogeochemical processes lowering the dissolved oxygen concentration of water. This effect will be offset in the case of enhanced stream flow, resulting in dilution of chemical concentration and vice versa in the case of lower stream flow.

The vulnerability of human populations to climate change differs considerably on a regional scale primarily because of differences in the base line climate and different exposures to climate stimuli in different regions. Also the severity of vulnerability depends on the adaptive capacity of the population and their management skills and resource availability. For example, the two continents of Africa and Asia are extremely vulnerable to climate adversity, and are additionally affected due to population growth notwithstanding the diminishing water resources. Africa is the continent with the lowest conversion factor of precipitation to runoff, which averages 15%, mainly due to predominant aridity in the region. Groundwater recharge measurements in India indicate the recharge rate to be as low as 3–5% of rainfall in the arid/semi-arid region (Sukhija *et al.* 1996*a*). Thus, climate change in Asia, which has 60% of the world's population, will have a significant impact as the region is naturally dependent on the monsoon system and is economically weaker resulting in lesser adaptability. Groundwater resources, which form world's largest freshwater resource, become increasingly important in a changing climate scenario. Aquifers have a large capacity and are capable of storing large amounts of runoff water due to increased flood events, are less vulnerable to droughts and in fact are important resources of drinking water supply during calamities as they are less vulnerable to contamination, do not require large-scale engineering structures for their exploitation and even for their augmentation, and are less prone to evaporation. However, as a result of changing climate and human interventions, in certain areas of the world there is a general fall in groundwater levels. However, groundwater provides an opportunity for several adaptation methods by way of artificial recharge or managed aquifer systems to overcome the ill effects of climate change. Artificial recharge (Asano 1985) is carried out with the following purposes for circumventing the problems created by the vagaries of changing climate: (1) to reduce, stop or reverse the declining trend of groundwater levels caused due to drought; (2) to protect the groundwater in coastal aquifers against seawater intrusion; and (3) to store surplus surface runoff floodwater to augment the ground reservoir and to utilize the wastewater for future use.

The purpose of this paper is to review the adaptive methods for sustaining groundwater through droughts, and discuss in detail two methods of adaptive and mitigation measures, namely: (i) artificial recharge through percolation ponds; and (ii) identification, assessment and characterization of deep confined aquifers which are resilient to droughts.

Adaptive methods for sustaining groundwater regime: techniques adopted for artificial recharge

There are several important techniques (Athavale 2003; Sharma 1998) used for artificial recharge as described below.

In the water spreading technique, surface water from perennial rivers/streams is allowed to spread over a large area for a sufficiently long period so that it percolates to recharge the aquifers.

With recharge basin and percolation tanks/ponds, the concept is to maximize the contact area and time of surface water with soil so that surface water percolates down to recharge groundwater. The technique involves construction of small dams or weirs across the river course. Monsoon streams run for short periods because of the limited period of rainfall. Alternatively the flow of the streams is restricted due to check dams, weirs and gully plugs, enabling more contact time for flowing water to infiltrate into the ground rather than go as surface runoff.

Surplus surface water, after treatment, can be artificially recharged through injection wells which have a similar construction to a tube well, but with a purpose of recharging the deep semi-confined aquifer. The same wells can be used to retrieve water during period of droughts and thus are useful for the changing climate.

Artificial recharge can be induced into groundwater, with deliberate creation of a cone of depression along rivers and streams. During flow time the stream can be effectively recharging groundwater which can be utilized during drought periods.

Another important groundwater management practice that is emerging is the identification of certain deep confined aquifers in regions where droughts occur periodically, and such deep aquifers can be utilized during such calamities. Deep confined renewable or non-renewable aquifers, if available in a region, are the most suitable source of safe and usually good quality water available for drinking or even irrigation. Such aquifers generally have a high residence time which can be determined by ^{14}C dating method (Vrba & Verhagen 2006).

Study of sustaining groundwater resources through percolation ponds during droughts in India

The vagaries of climate change are already being felt in many countries including India. Droughts and floods have perhaps become more common. The time series of rainfall data for *c*. 100 years in a semi-arid region of India shows (Fig. 1) that every fourth year is a drought year and every seventh year is a year of surplus water causing floods. Thus it becomes imperative for a country like India, situated in the monsoon region (other monsoon countries include Korea, Philippines, China, Japan, Bangladesh, Pakistan, Sri Lanka, Malaysia) to develop adaptive measures to counter the climate changes.

Percolation ponds as artificial recharge structures

A percolation pond or tank is an artificial recharge structure constructed across a monsoon stream with the purpose of harvesting surface runoff caused during the monsoon period. These structures are simple earthen dams to impound surface runoff (Fig. 2a). Generally these earthen dams are a few hundred metres long and a few metres high. Typically the water area for such ponds is less than one square kilometre and storage capacity is less than half a million cubic metres. Due to evaporation losses, only a certain fraction of impounded water (defined as the percolation efficiency) is expected to percolate through the pond bed to the groundwater. The efficiency of such structures is expected to be governed by factors such as geology, climate, rate of silting, age of tank etc. The purpose of this investigation was to determine how effective these structures are and how various governing factors control the efficiency of these structures. Although there are a very large number of such structures in hard rock areas of peninsular India and a huge investment is made in such structures, there are only a few scientific studies.

A qualitative change in the study of such structures was made as a new method was developed to estimate their efficiency based on the chloride balance (Sukhija *et al.* 1997). This method, in contrast to the conventional water balance method, overcomes the problem of estimating evaporation from the surface reservoir, which is either estimated

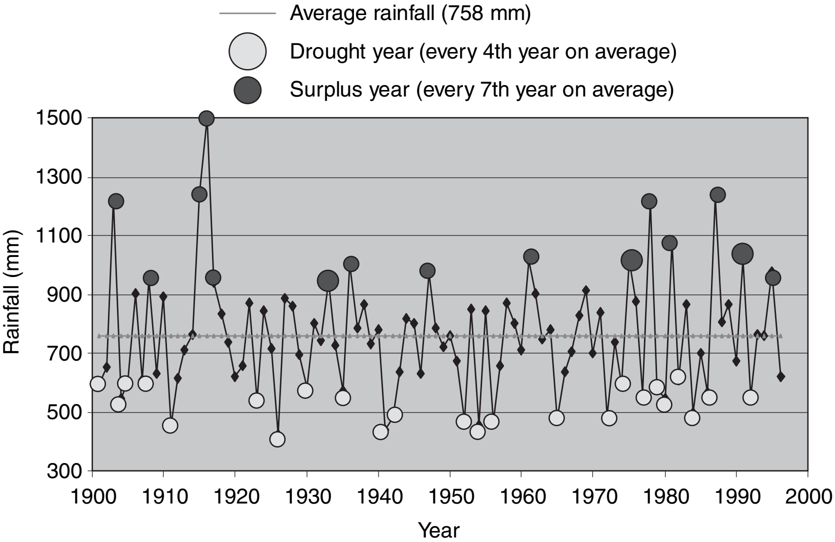

Fig. 1. Rainfall pattern in Nalgonda district (Andhra Pradesh, India) for the last 100 years (1901–1996) showing drought every fourth year and excess rainfall in every seventh year.

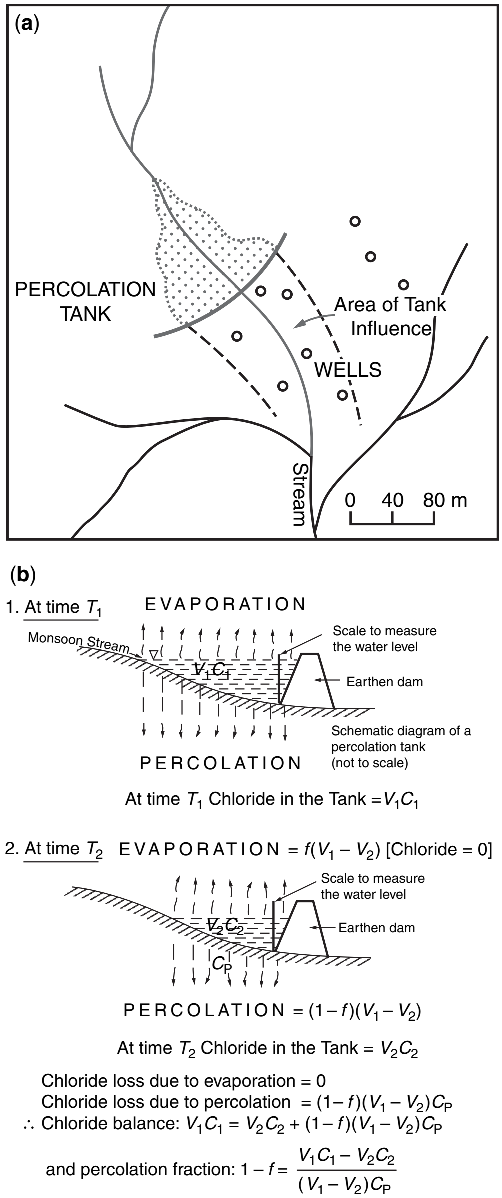

Fig. 2. (**a**) Schematic representation of a percolation pond. (**b**) The principle of measurement of efficiency of percolation ponds based on chloride mass balance.

by empirical relationship of meteorological parameters (humidity, temperature, wind speed, sunshine hours etc.) or is based on small-scale evaporimeters established close to the percolation ponds, which do not provide realistic estimates of evaporation affected by waves, wind speed, turbulence etc. In the following, the methodology of estimation of efficiency of percolation ponds is demonstrated in different geological formations, and the results of efficiency in light of various controlling factors is discussed.

Efficiency of percolation ponds: methodology

The measurement of efficiency of percolation ponds requires the chloride mass balance to be estimated, as shown in Figure 2b. If there are no sources or sinks of chloride, the total amount of chloride in the tank (e.g. at the end of the monsoon period) between time T_1 and T_2 (any time before the tank is empty) either remains in the tank or is charged to groundwater since no chloride can be lost by evaporation. Thus at time T_1 (immediately after the monsoon season), the total chloride in the tank is given by V_1C_1, where V_1 is the volume of water and C_1 is chloride concentration of water in the tank. V_1 can be estimated by computing the average depth of the water in the tank and chloride concentration is measured at two or three sites to get an average concentration, though not much difference in concentration is seen. During the time between T_1 and T_2, it is expected that part of the water (fraction f) of the tank is evaporated and another part (fraction $1 - f$) is recharged, assuming no other losses from the structure. Thus at time T_2, chloride in the tank will be (C_2V_2) and chloride recharged through the tank will be $(1 - f)^*$ $C_p(V_1 - V_2)$ since no chloride is lost by evaporation. Thus for the mass balance of chloride:

$$C_1V_1 = C_2V_2 + (1 - f)C_p(V_1 - V_2)$$
$$(1 - f) = (V_1C_1 - V_2C_2)/C_p(V_1 - V_2)$$

where $C_p = \sum V_iC_i / \sum V_i$ = weighted average chloride concentration in the tank water.

Thus the efficiency of the tank as a recharge structure can be obtained through measurement of the volume of water in the tank and its chloride concentration as a function of time. Figure 3 shows typical curves of volume of water and chloride as a function of time used for such calculations.

Percolation ponds studied

In all four ponds (Table 1), located in diverse geological formations such as granites, basalts and sandstones, were studied with the objective of determining their efficiency and the factors which contribute to varying efficiency from 20–60%.

Climate adaptation: groundwater management during drought. The following discussion focuses on how the percolation ponds can serve to

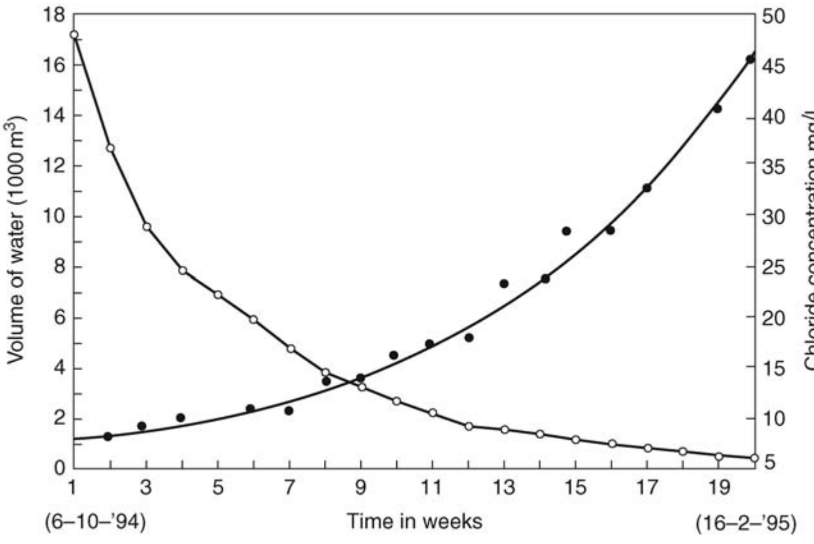

Fig. 3. Change in the volume of water and chloride concentration in the percolation pond as a function of time.

augment groundwater resources using monsoon stream water. Attention was given to one pond (Kalwakurty) situated in granite which has been studied in detail. The pond is situated in a typical semi-arid climate with normal rainfall during 1985–1986, which can also be considered as a base year before the construction of the pond. The positive impact of the percolation pond (Kalwakurty) can be gauged from the fact that during normal and above-normal years of rainfall (1989–1991), the water level in the downstream wells increased by 5–6 m primarily due to contribution from the pond (Fig. 4). It was also observed that groundwater levels in two open wells close to the dam rose to ground surface during the 1989–1991 hydrological years. Such a situation never occurred before construction of the percolation pond. Another significant observation was that, in spite of very low rainfall (drought) during 1992–1993, the wells near the pond showed higher water levels than during the period before pond construction. Thus the groundwater level data qualitatively verify that the percolation pond is effective and it provided sustained exploitation during drought.

Efficiency of percolation ponds: climate versus geological control. One important point that emerges from Table 1 after comparing the percolation efficiencies is that if two ponds are located in the same climatic environment (e.g. Saurashtra region) within a distance of 50–100 km, a drastic change in efficiency is observed. The two ponds are located in areas with sandstone and basalt as

Table 1. *Details of studied percolation ponds and their efficiency using chloride mass balance method*

No.	Geological formation	Name of pond	State & district	Water area (km^2)	Tank dimensions (m), length/height	Tank capacity (Mm^3)	Efficiency (% (year))
1	Granite	Kalwakurty	Andhra Pradesh/ Mehboob Nagar	0.123	600/5	0.133	36 (1994–95) 42 (1995–96) 43 (1996–97)
2	Granite	Singaram	Andhra Pradesh/ Ranga Reddy	0.015	170/2.5	0.01	30 (1992–94)
3	Basalts (Deccan Trap)	Lakanka	Gujarat/Surinder Nagar	0.161	600/5	0.21	20–30 (1995–97)
4	Sandstone	Saper	Gujarat/Surinder Nagar	0.268	600/5.5	0.312	60 (1996–97)

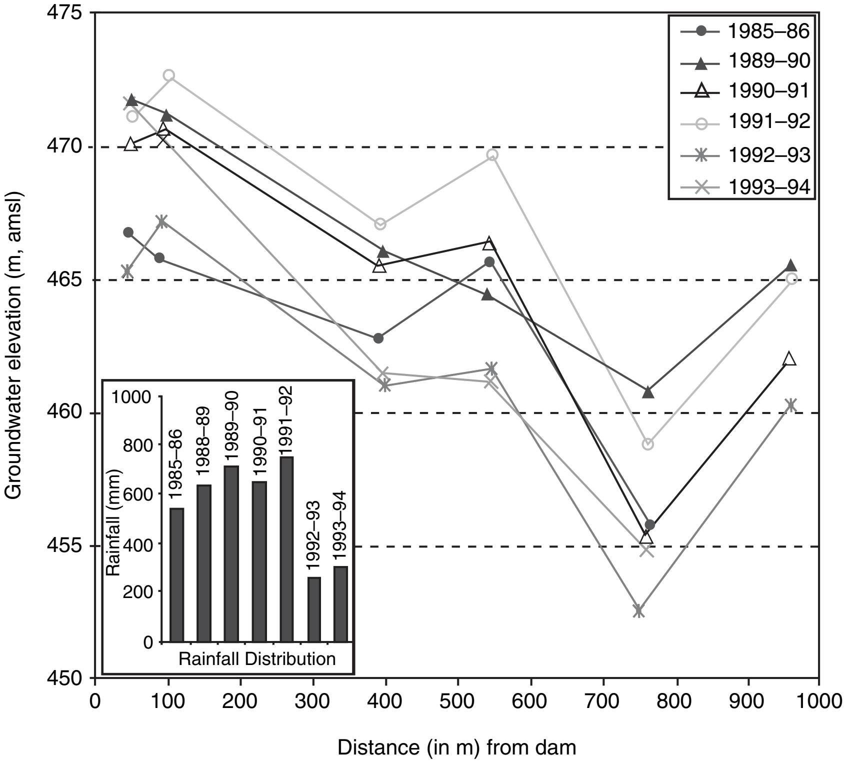

Fig. 4. Spatial and temporal response of groundwater levels before (1985–1986) and after (1988–1989) construction of percolation pond at Kalwakurthy in granite. The water levels in the downstream wells increased by 5–6 m during normal and above-normal rainfall years primarily due to pond construction. Pond influence is seen up to 500 m. Even during drought years of 1992–1993 and 1993–1994, the wells within 200–300 m of the tank have maintained levels higher than those in 1985–1986 (before construction of percolation pond).

bed rock. The pond situated in the sandstone has an efficiency of 60% in contrast to the one situated in basalts which has an efficiency of 20%. This difference in efficiency can be related to the geology, as the climatic effect (evaporation) should be more or less the same. This observation demonstrated that the role of geology outweighs the climate effect. This observation is supported on the basis of higher hydraulic conductivity (11 m/day; UNDP 1976; Karanth 2004) for the sandstones in comparison to the basalts (0.006–0.07 m/day). Further studies on natural recharge indicate higher recharge by a factor of three (20–25%) in sandstones in comparison to granites (7–8%) within semi-arid areas of India (Sukhija *et al.* 1996*a*).

Adaptive measures: sustainable groundwater management by identification of deep confined aquifers

There are different approaches to mitigate drought. The approaches can be preventive, as discussed above, by integrated management of surface and groundwater or using deep confined aquifers which should be identified before the calamity. The important consideration is that the potential aquifer identified is assessed to be resistant against droughts using geological, hydrogeological, geophysical and isotopic methods.

UNESCO (2003) has recently formed an Expert Group called Groundwater for Emergency Situations (GWES) to formulate guidelines for mitigating groundwater problems during emergency situations. The objective of this group has been to look at measures to be undertaken regarding groundwater supply and management during emergency situations such as floods, droughts, earthquakes, landslides etc. One of the recommendations of UNESCO (2003) is to identify aquifers which can yield good quantity and quality of groundwater during the extreme events. Worldwide attempts are in progress for such a task.

In the following, the Neyveli aquifer from southern India is discussed as a representative groundwater resource body that can be taken as an example to be used for mitigation of drought. The purpose of this example is to illustrate how and why this confined aquifer can be considered as useful for climate adaptation. The important criteria for such a selection or identification are: (i) hydrogeology; (ii) groundwater regime in time and space; (iii) characterization of recharge and discharge areas; and (iv) wide span of groundwater ages from Modern to >30 000 years BP. These points are discussed to bring home the usefulness of the Neyveli aquifer.

Regional hydrogeology of the study area

The Neyveli aquifer lies in Tamilnadu state, India, about 200 km south of Chennai (formerly Madras) on the eastern coast (11°15′–11°50′N, 79°10′–79°50′E; Fig. 5). The average rainfall

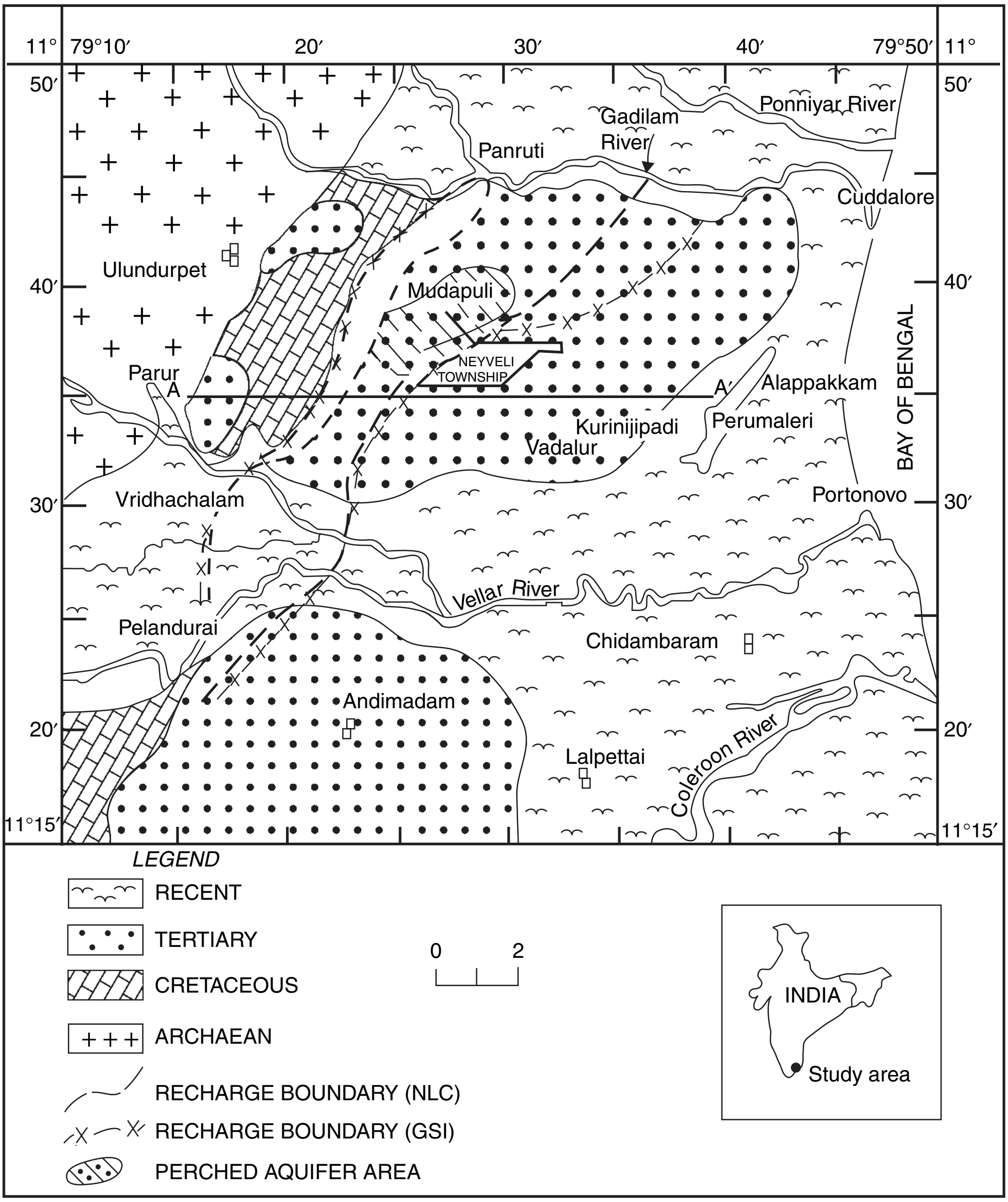

Fig. 5. Geological map of Neyveli groundwater basin.

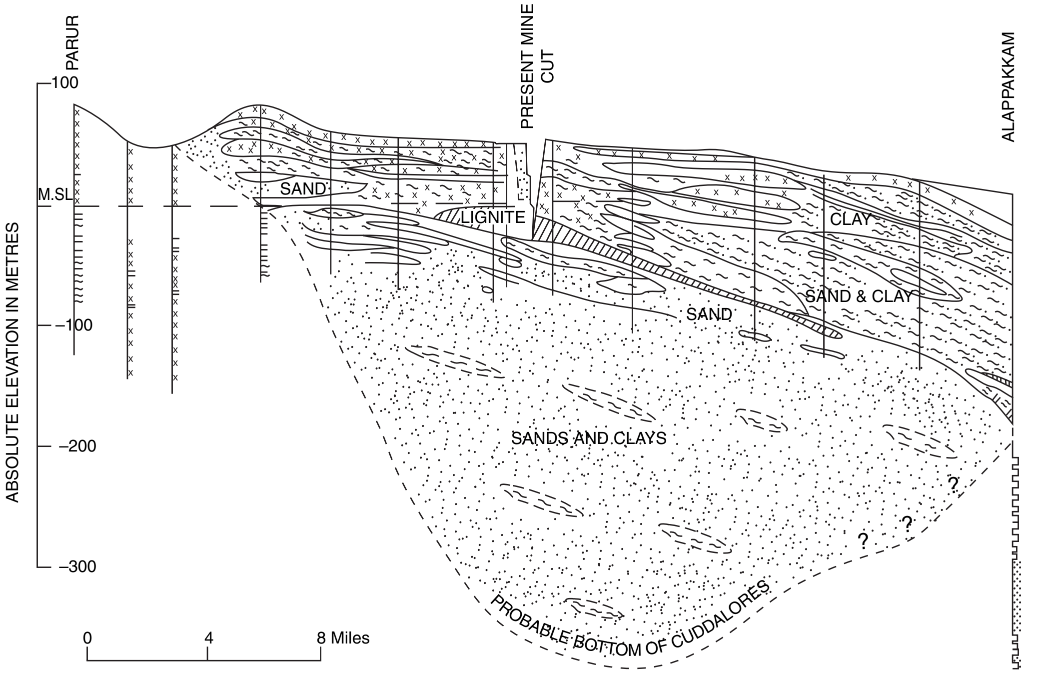

Fig. 6. Neyveli groundwater reservoir cross-section (profile A–A′ of Fig. 5).

in the area is about 1200 mm. The geological cross-section A–A′ in Figure 5 is shown in Figure 6. Thick Upper Miocene sediments, mainly Cuddalore sandstones, serve as an excellent aquifer under unconfined and confined conditions. Extensive clay layers with a thick lignite bed (8–16 m) are responsible for confining conditions (Fig. 6). The aquifer has been stressed with an average groundwater extraction rate of 7000 m^3/h for facilitating open-cast lignite mining. This has resulted in development of a cone of depression around the mine areas. Groundwater occurs under phreatic, semi-confined and confined conditions. Measurements of ^{14}C and ^3H, as discussed later, led to identification of an area trending NE–SW as the recharge area. Under confined conditions two aquifers, upper and lower, are identified below the lignite bed. These two aquifers are confined and act as a single hydraulic unit although they are separated by clay beds. The level of potentiometric surface of the lower and upper confined aquifers was initially approximately the same to justify them being considered as a single unit. Open-cast mining of lignite required lowering of the hydraulic pressure by pumping, which began in July 1961. The aquifer has been pumped continuously for the last 45 years.

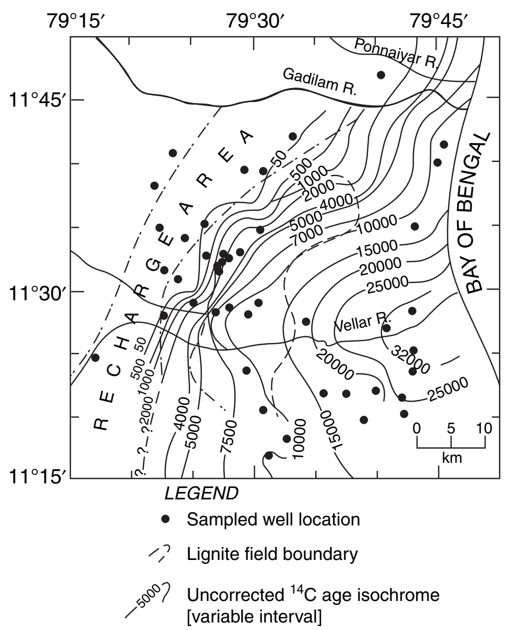

Fig. 7. Systematic lateral distribution of radiocarbon isochrons (1985 sampling).

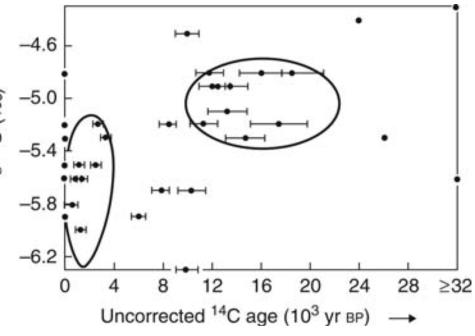

Fig. 8. Plot of $\delta^{18}O$ (‰ SMOW) versus uncorrected ^{14}C ages of deep confined aquifer.

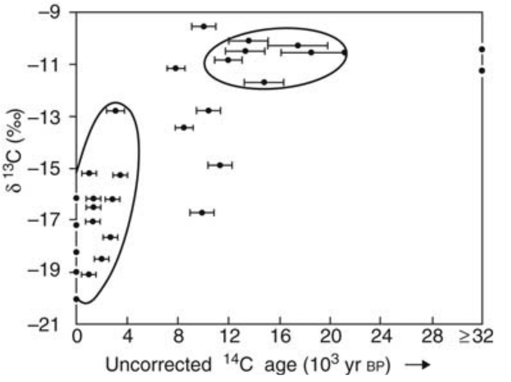

Fig. 9. Plot of $\delta^{13}C$ versus uncorrected ^{14}C age of Neyveli groundwater basin.

Demarcation of recharge area

One of the outstanding questions in understanding the groundwater system of the Neyveli aquifer has been identifying the recharge area for such a powerful artesian aquifer. This question, apart being addressed by the conventional hydrogeological approach, was also addressed through measurements of environmental isotopes of tritium, ^{14}C, $\delta^{13}C$ and $\delta^{18}O$. For this purpose about 60 groundwater samples collected from wells deeper than 50 m were analysed during two campaigns in 1985 and 1991. The basic idea was to study not only spatial changes in isotopic composition but also temporal variation.

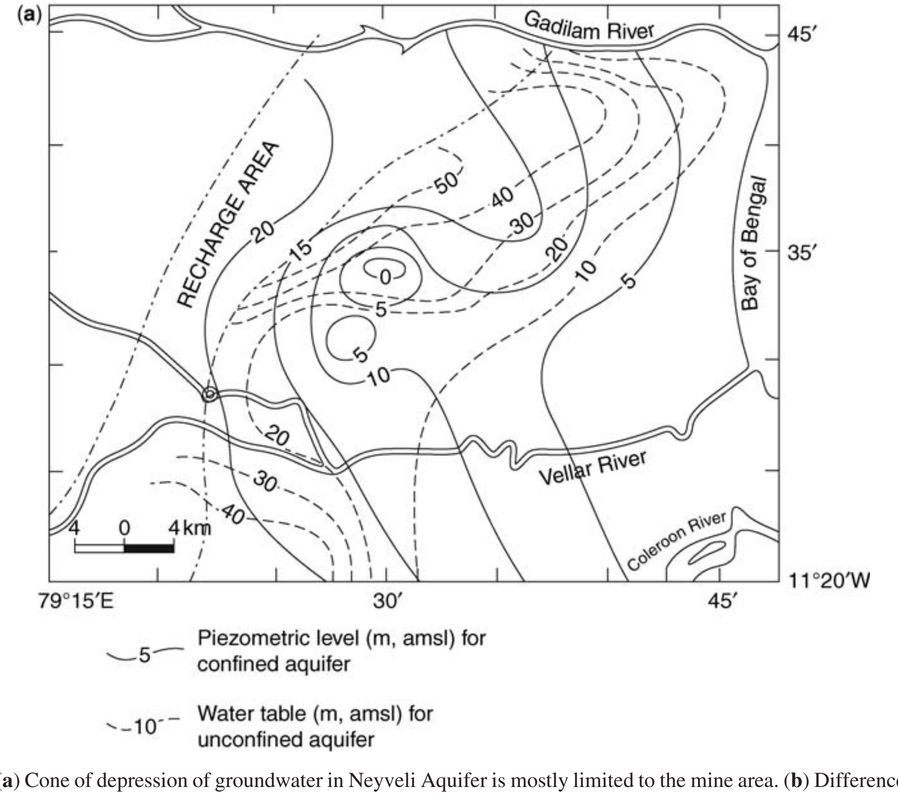

Fig. 10. (**a**) Cone of depression of groundwater in Neyveli Aquifer is mostly limited to the mine area. (**b**) Difference in the piezometric levels (in m) between 1986 and 1990 for Neyveli Aquifer.

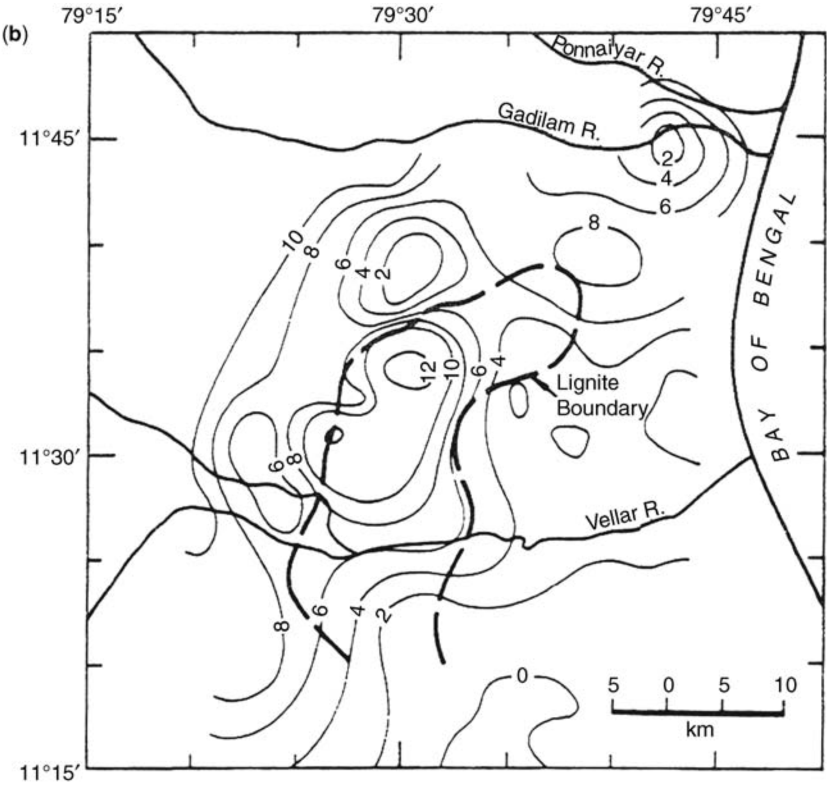

Fig. 10. (*Continued*)

Radiocarbon ages of groundwater

For ^{14}C age calculation, we have assumed an initial activity of 100% modern carbon (pmc) in groundwater, and taken the half-life of ^{14}C as 5730 ± 40 years. The initial activity of 100 pmc was justified as the youngest groundwater with thermonuclear tritium had a ^{14}C concentration ≥ 100 pmc. The ^{14}C ages were calculated on the piston flow model which has a distinct recharge area (Bath *et al.* 1979). Figure 7 shows the systematic lateral distribution of radiocarbon isochrons based on the 1985 sampling. The recharge area of the Neyveli confined aquifer was delineated on the basis of isotopic, geochemical and hydrogeological considerations. Figure 7 shows the delineated recharge area extending from NE to SW. The wells in the delineated recharge area have shown the presence of thermonuclear tritium, as well as radiocarbon indicating a modern age. Away from this area and towards the east, the wells tapping the confined aquifer lack tritium and the ^{14}C ages show a gradual increase in flow direction. A systematic lateral increase in ^{14}C ages from recharge area to discharge area varied from Modern to $\geq 30\,000$ years BP. In addition to these isotopes, the $\delta^{13}C$ and chloride show low values (-23 to -27‰ and 10 to 30 mg/l, respectively) in the identified recharge area. This investigation indicates clearly that the confined aquifer, though very large, has modern recharge feeding in its recharge area for sustenance during drought periods.

Palaeoclimate signatures

From the isotopic and geochemical observations (Sukhija *et al.* 1998) the palaeoclimatic signatures in the deep confined Neyveli aquifers during 20 000 to 12 000 years BP were deduced. Figure 8 shows the plot of $\delta^{18}O$ versus uncorrected ^{14}C ages of the deep confined aquifer. Relatively enriched $\delta^{18}O$ values (-4.8 to -5.3‰) corresponded to an age span of 20 000 to 12 000 years BP encompasing the last glacial period ($18\,000 \pm 2000$ years BP) with aridity. The climate transition from an arid to a relatively humid period about 8000 to 12 000 years BP is marked by rather depleted values (−4.5 to −6.3‰) implying that the groundwater recharge was related to precipitation events having varied isotopic ratios during this climate transition. The late Holocene (4000 years BP) groundwater has $\delta^{18}O$ values of −5.2 to −6.3‰, indicating a rather humid but unstable climate.

Figure 9 shows the plot of $\delta\ ^{13}C$ versus uncorrected ^{14}C ages. Here again, relatively enriched $\delta^{13}C$ values (−10 to −12‰) corresponding to ^{14}C ages 20 000 to 12 000 years BP encompassing the last glacial period are attributed to growth and dominance of C4 plant species indicating an arid climate during this time span. In a similar fashion to $\delta^{18}O$ versus ^{14}C during 12 000 to 8000 years BP, a rather depleting trend of $\delta^{13}C$ (−9.5 to −17‰) indicates growth of vegetation following the C3 pathway of carbon fixation. Further depletion of $\delta^{13}C$ from −13 to −20‰ is seen during 4000 years BP onwards indicating the abundance of C3 plant species. Thus in the Neyveli aquifer very clear signatures of changing climates are preserved.

Resilient nature of the Neyveli aquifer

The resilient nature of the Neyveli aquifer was tested by continuous withdrawal of the aquifer. Because of the characteristic recharge area of the aquifer, confined aquifers get a huge amount of natural recharge of *c.* 110 m^3/year (Sukhija *et al.* 1996*b*; Rangarajan *et al.* 2005). The aquifer has also been studied for change in the pressure heads. Figure 10a and b show that in spite of very heavy continuous pumping, the cone of depression is mostly limited to the mine area. Maximum pressure head difference is about 10 to 12 m and at the top of the aquifer the pressure difference does not exceed a few metres. Similarly the quality variation of the aquifer is studied through chloride variation as chloride is a conservative tracer.

Figure 11a shows the isochlors based on chloride concentration of deep groundwater (1985). During 1985 chloride concentration was 10–30 mg/l in the recharge area, and there was a systematic increase towards the SE, indicating groundwater

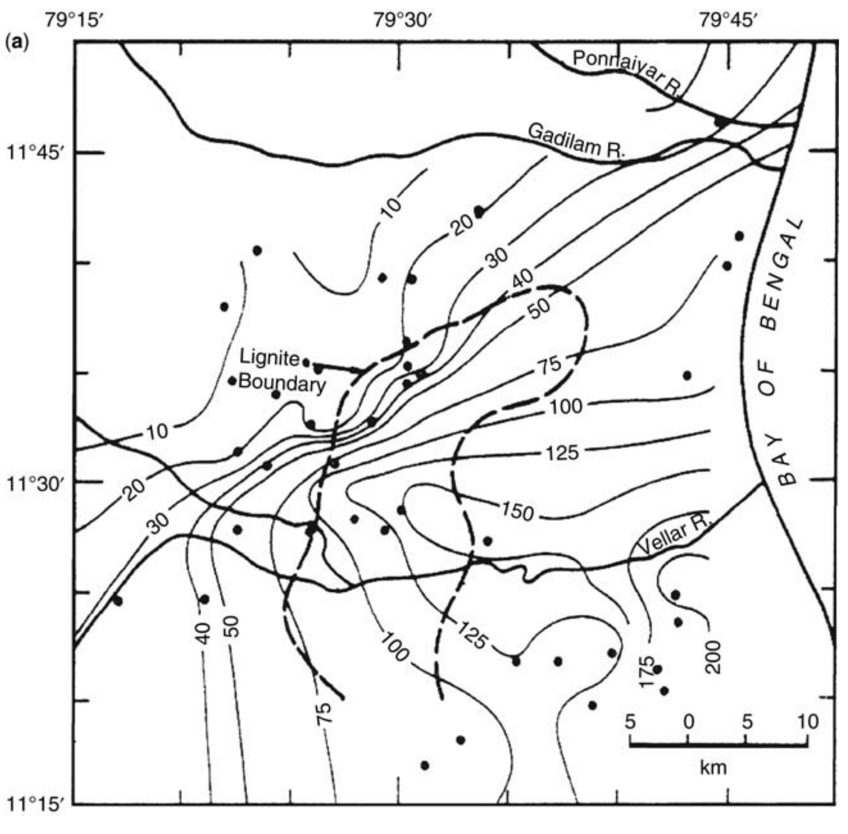

Fig. 11. Isochlors based on chloride concentration (mg/l) of deep groundwater (1985) of Neyveli aquifer in: (**a**) 1985; (**b**) 1991.

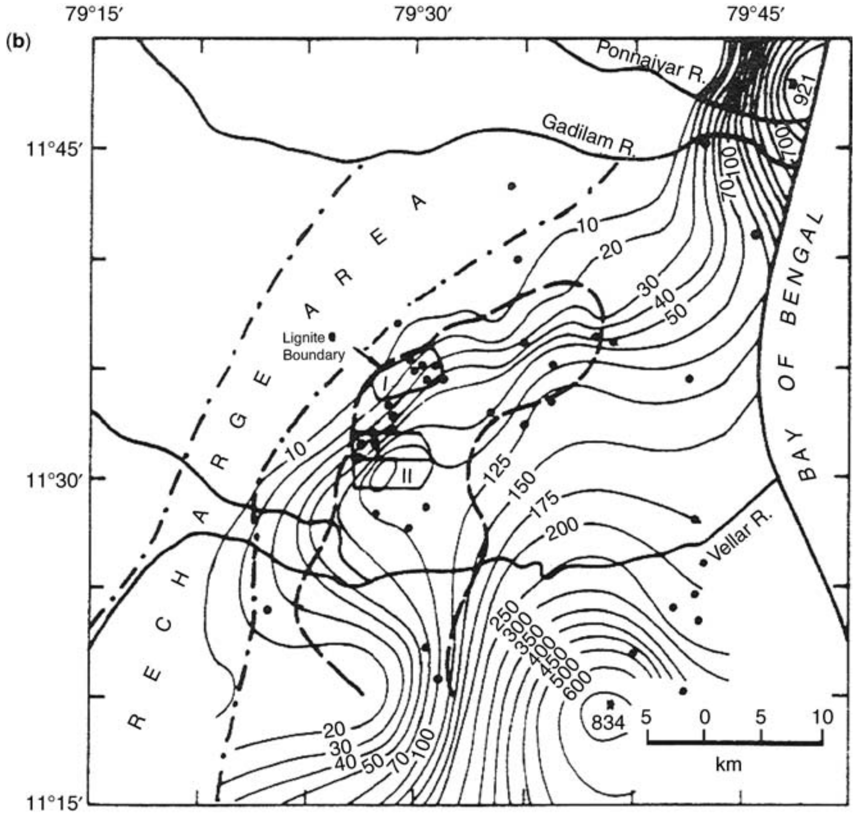

Fig. 11. (*Continued*)

movement towards the SE. During 1991 (Fig. 11b) the general trend of isochlors remains similar except for local perturbation in the NE and SE. Thus there is some change in quality in certain parts of the aquifer, yet the overall quality of the aquifer is not altered significantly, indicating the resilient nature of the aquifer.

Discussion

The following discussion focuses on how the two different approaches of artificial recharge for augmenting groundwater resources, and identifying and characterizing renewable and non-renewable deep aquifers, can prove highly useful for sustaining groundwater resources through droughts.

In the first approach it is shown that percolation ponds constructed across monsoon streams can prove highly successful by augmenting groundwater resources (Sukhija *et al.* 2005). It has been shown that after construction of such percolation ponds the water table during low rainfall years is much higher than the period before construction of such artificial recharge structures. During normal rainfall years the water table had even touched the ground surface. Further it has been shown that there is lot of variation in the efficiency of such structures. However, it is also demonstrated that the control of geology outweighs the climate effect. Thus if such percolation ponds are constructed in suitable geological strata, they can provide significant artificial recharge for groundwater sustenance.

From the results of the Neyveli aquifer studies, we demonstrate how and why the Neyveli aquifer is an outstanding example of an aquifer which can be utilized in extreme conditions of drought. It has been shown that Neyveli aquifer is a vast aquifer having semi-confined and confined groundwater system; the aquifer has excellent hydraulic conductivity and has a large areal extent. The aquifer has a very distinct recharge area with very young (Modern) water indicating excellent recharge conditions. The areal extent of recharge area is about 600–1000 km^2 with a recharge rate of *c*. 110 m^3/year. The vastness of the aquifer is further demonstrated by the ^{14}C age structure as it is very young age (Modern) to $\geq$30 000 years BP old (Sukhija *et al.* 1996*b*). The palaeoclimate signatures have been clearly preserved in the aquifer. It is also shown that despite continued pumping, the groundwater potential of the aquifer is still quite high. Although there is change in the hydraulic head and groundwater chloride, the changes are not alarming, demonstrating the resilient nature of the aquifer against changing climates and increasing withdrawal rate, thus making the aquifer most useful for tackling climate change.

Conclusions

Adaptive measures and mitigation of climate changes are vital activities to counter the effects of droughts and floods. Integrated management of surface and groundwater resources through artificial recharge and water conservation are very important as a precautionary measure. The identification of deep confined aquifers with large groundwater potential which can serve during severe droughts, without much change in the groundwater yield and quality, is another adaptive and mitigation measure which is advocated.

This work was carried out under the CSIR Emeritus Scheme for which the author is very grateful to the Council of Scientific & Industrial Research, New Delhi. The author is grateful to Dr V. P. Dimri, Director, NGRI, for providing facilities at NGRI, Hyderabad. Mr P. Nagabhusanan, Dr D. V. Reddy and Dr N. C. Mondal provided various inputs, and Dr S. K. Gupta has taken pains to review this paper for which author is very grateful.

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Note: Page number in *italic* denote figures. Page numbers in **bold** denote tables.

There is a general consensus that for the next few decades at least, the Earth will continue its warming. This will inevitably bring about serious environmental problems. For human society, the most severe will be those related to alterations of the hydrological cycle, which is already heavily influenced by human activities. Climate change will directly affect groundwater recharge, groundwater quality and the freshwater–seawater interface. The variations of groundwater storage inevitably entail a variety of geomorphological and engineering effects. In the areas where water resources are likely to diminish, groundwater will be one of the main solutions to prevent drought. In spite of its paramount importance, the issue of ‘Climate Change and Groundwater’ has been neglected. This volume presents some of the current understanding of the topic.