GEOTHERMAL STUDIES OF THE TABLE MOUNTAIN GROUP AQUIFER SYSTEMS

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WRC Report No. 1403/1/07

Water Research Commission

Geothermal Studies of the Table Mountain Group Aquifer Systems

Report to the

Water Research Commission

by

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WRC Report No 1403/1/07

ISBN 978-1-77005-560-5

July 2007

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EXECUTIVE SUMMARY

Recognition of the groundwater potential in the quartzitic sandstones of the Table Mountain Group (TMG), following the development of relatively deep artesian wells in the Citrusdal region, 220 km north-east of Cape Town, has directed considerable attention towards the TMG resource as a potential long-term solution to the increasing metropolitan and urban water scarcity in the Western Cape Province. An incentive for developing new groundwater exploration and reservoir-characterization methodologies has been created, which also provides an opportunity for broadening and deepening the knowledge of rock temperature gradient and heat flux in the area.

Groundwater flowing naturally through an aquifer system transports heat, as well as solutes, and thereby alters the subsurface temperature field. Consequently temperature, which is easy, quick, and inexpensive to measure, compared to chemical and isotopic methods, is useful as a tracer of hydrogeological processes and a means for testing conceptual groundwater-flow models. Large-scale groundwater abstraction such as envisioned for Cape Town, if it leads to a substantial alteration of the deep flow regime, should have near-field impacts on the geothermal regime around a major well-field, and possibly far-field impacts on geothermal regimes in the recharge and discharge areas along a flowpath through the abstraction site. On the grounds that

- impacts cannot be assessed or measured without an actual experiment in large-scale abstraction, and
- the advective transport of heat in the shallow parts of the crust is not only a geophysical, but also an ecological and environmental issue,

the planning for any future groundwater abstraction experiment should include a programme of observation and measurement of geothermal gradient, thermal conductivity and heat flow around the proposed abstraction and environmental monitoring sites.

For a hydrogeological-geophysical experiment with a clear environmental purpose, it is a necessary prerequisite that baseline information on regional heat flow and thermal conductivity properties be obtained from suitable reference sites in the TMG and Cape Fold Belt. Aims of the present study are therefore:

- To establish background geothermal gradients and heat flux in areas unaffected by underground water flow;
- To prepare a theoretical and experimental basis for the monitoring of changes in geothermal gradients and heat flux in regions where groundwater extraction is in progress;
- To prepare specifications for future investigations in which quantitative use is made of heat as a groundwater tracer, through numerical modelling in conjunction with traditional chemical and isotopic methodologies.

Prior to this investigation, no heat-flow studies were available in those parts of the Cape Fold Belt where large-scale groundwater abstraction could be contemplated. Previous borehole geothermal surveys in the Western Cape were mostly unsuccessful because of evident

disturbances due to underground water flow. During the current project an attempt was made to locate as many boreholes as possible that are undisturbed by water flow.

The methods and procedures adopted during the study, such as drilling, field and laboratory measurements and analysis, are described and the results are displayed in a set of tables for boreholes at the Skuifraam site near Franschhoek, the Birkenhead site near Stanford, and the Blikhuis site near Citrusdal. The laboratory conductivity analysis is discussed together with the heat flow results.

Skuifraam results

The best estimate of the background crustal heat flow from the pre-Cape basement terrain in the Western Cape Province is 76 mW m^{-2}, obtained from a thermal gradient of 21.5 K km^{-1} in the deeper (190-290 m) interval of a 300 m borehole into the Cape granite at Skuifraam in the Berg Water Project area. The Cape granite underlying the site has a mean thermal conductivity of 3.55 ± 0.25 W m^{-1} K^{-1}. The thermal gradient in the upper part of the Skuifraam borehole (18.2 K km^{-1}) is lowered by possible shallow heat advection in a fracture intersected by the borehole at ~160 m depth.

A radar image of the borehole obtained during an experimental test of the GeoMole logger, currently under development at the University of Stellenbosch, shows that the granite between the depths of 60 m and 120 m is apparently uniform and homogeneous, but the section centred around 160 m is marked a pattern of radar diffractions, which probably reflects structural heterogeneity. Below this level there are radar reflections inclined to the borehole axis, one of which diverges from it with depth. Testing of a fault hypothesis for this fracture set, through the drilling of a nearby borehole to intersect the zone in an up-dip direction, is proposed.

Birkenhead results

The calculated heat flow in the Bokkeveld Group at the Birkenhead site near Stanford is 50-55 mW m^{-2}, from an observed geothermal gradients ranging between 14.5 and 15.9 K km^{-1}, and the mean thermal conductivity of the Bokkeveld shale (measured in core samples from elsewhere) is 3.43 ± 0.70 W m^{-1} K^{-1}. The lowered apparent heat flow near Stanford is ascribed to the effect of groundwater moving through the TMG aquifer system, which transports substantial quantities of heat and thereby alters the subsurface temperature field.

Blikhuis results

Unambiguous interpretation of earlier thermal-gradient measurements at the Blikhuis Experimental Deep Drilling (BEDD) site, between Citrusdal and Clanwilliam, requires further thermal conductivity data on higher units (Cedarberg shale and Goudini Formation) in the TMG, but observed thermal gradients in the BEDD borehole (14.3 K km^{-1} in the Goudini sandstone and 35.8 K km^{-1} in the underlying Cedarberg shale) are consistent with the same background heat flow as the Skuifraam determination. The thermal conductivity of quartzite samples from drill-core in the Peninsula Formation of the TMG yields a high value of 7.35 ± 0.34 W m^{-1} K^{-1}, which is probably due to its very pure quartzose composition.

The high thermal gradient in the Cedarberg shale is probably related to its low thermal conductivity (~2 W m^{-1} K^{-1}), indicating that this stratigraphic unit is not only an effective aquitard but is also a good insulator or “thermal blanket” above the hydraulically and thermally conductive Peninsula Aquifer.

Equilibrium temperature gradients

The calculations relating to the Blikhuis locality and the inferences drawn about possible lateral heat advection beneath the Birkenhead sites, justify the use of a heat flow of 76 mW m^{-2} as typical for the whole of the pre-Cape basement terranes in the Western Cape Province. Until further background heat flow studies are undertaken in other parts of the Malmesbury Group and the Cape Granite Suite, "undisturbed" temperature gradients in the TMG quartzite (specifically the Peninsula Formation) and the Bokkeveld shale are estimated from the measured conductivity data, as follows:

Bokkeveld shale: $dT/dz = 76/3.43 = 22.2$ K km^{-1}

Peninsula quartzite: $dT/dz = 76/7.35 = 10.3$ K km^{-1}

These thermal gradients are those to be expected in the complete absence of any groundwater flow effects.

Implications for hot-spring sources

The well-known hot springs found in the Western Cape, and a growing set of temperature measurements in boreholes and springs, provide a potentially important source of information about the deep groundwater flow paths within the TMG aquifer system. With recent technological advances, inexpensive waterproof temperature loggers are widely available for subsurface temperature measurements. Groundwater temperature is now measured easily and rapidly, provided that care is taken to ensure the recorded temperature is representative of water in the aquifer and not influenced by movement of water in the borehole. The newer Distributed Temperature Sensing (DTS) technology, based on fibre optics, enables the simultaneous online registration of temperature profiles along one or more boreholes. DTS can be used in conventional wireline logging mode, but a special feature is the potential to install the sensor cables outside the borehole casing. In recent applications, thermocouples and thermistors are used to obtain a time series of measurements remotely, and airborne thermal sensors are used to detect areas of ground water discharge.

Emergence temperatures of spring waters reflect the interaction between advective and conductive transport of heat in the host aquifer systems, aquifer permeability being the key factor. Where rock permeability is low, groundwater flow velocities are low, heat transport is dominantly conductive, and therefore spring temperatures are low. In rocks with high permeability, flow velocities are high and advective transport of heat is the dominant process. However, because high permeability also results in large volumes of circulating water, spring temperatures again remain low. The warmest springs generally occur for an intermediate range of permeability.

3-D coupled heat and flow modelling

Investigations of coupled heat and groundwater flow undertaken elsewhere in the world for the regional estimation of aquifer permeability and hydraulic conductivity, serve as examples for the kind of study that can and should be undertaken within the TMG terrains of the Cape Fold Belt. The value of using water temperature for flux estimation in groundwater-stream systems is now well established elsewhere, and the systematic recording of borehole temperatures and depth from a wide variety of locations within the Cape Fold Belt is proposed. Regular and standardized water temperature measurements taken at key reference sites may provide a simple first order indication of the contribution of deep groundwater systems to base flow.

The example of the hot spring at Brandvlei, near Worcester, where the high outflow temperature (64°C) and voluminous flow (127 l/s or 0.127 m^3/s) implies a geothermal advective power of at least 27 megawatts, illustrates that combined thermal and fluid flow measurements made at springs integrate the signal of geological and hydrological processes over large spatial areas and possibly long periods of time. However, the interpretation of such measurements requires the development of an improved mathematical and computer-based modelling framework, within which a suite of data is related to petrophysical properties and hydrogeological processes. The present report provides information on readily available computer software programmes to solve coupled ground water flow and heat transport problems, provides an overview of their characteristics, and discusses the pro's and con's of using either finite-difference or finite-element methods.

In conclusion, the quantitative spatial mapping and in-situ temporal monitoring of local geothermal gradients and spring-discharge temperatures, in association with the combined modeling of fluid and heat advection in the TMG aquifers, could provide a powerful and relatively inexpensive new tool, both for groundwater exploration (storage and flow determination) and for the monitoring and interpretation of impacts due to large-volume abstraction.

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List of Abbreviations

A	area of the aquifer base
A	Aspect ratio
AAPG	American Association of Petroleum Geologists
atm	atmospheric pressure
B	borehole
B	Basin thickness
BEDD	Blikhuis Experimental Deep Drilling
BHTs	bottom-hole temperatures
BPI	Bernard Price Institute of Geophysical Research
BS	Bokkeveld Shale
BWP	Berg Water Project
CAGE	Citrusdal Artesian Groundwater Exploration
CCT	City of Cape Town
CCTV	Closed Circuit Television
CG	Cape Granite
CHSL	Cape hot springs line
CMWL	Cape meteoric water line
Cps	Count per second
c_w	heat capacity
D	Deuterium
DOC	Dissolved organic carbon
dT/dz	geothermal gradient
dT/dx	geothermal gradient
DTS	Distributed Temperature Sensing
DWAF	Department of Water Affairs and Forestry
E	East
EC	Electrical Conductivity
ENE	East north east
E/W	East - West
FD	finite-difference
FE	finite element
FEFLOW	Finite Element subsurface Flow system
g	gravitational acceleration
GIS	Geographic Information System
GMWL	global meteoric water line
GSNA	Geothermal Survey of North America
GUI	graphical user interface

H	Heat power
J	Joule
k	permeability
k_{NB}	permeability normal to bedding plane
k_{PB}	permeability parallel to bedding plane
K	Kelvin
K_M	Thermal conductivity of rock-fluid matrix
K_m	Thermal conductivity of porous medium
K_t	Thermal conductivity
kg	kilogram
kg/m^3	kilogram per cubic metre
km	kilometer
kyr	Thousand years
L	characteristic length of flow
LBNL	Lawrence Berkeley National Laboratory
L_B	basin length
l/s	litre per second
MODFLOW	modular three-dimensional finite-difference groundwater model
Ma	million years
MW	Mega Watt
mg/l	milligram per liter
min	minute
mS/m	milli Siemens per meter
mV	milli Volt
mW	milli Watt
m^2	metre square
m^{-1}	per metre
m^{-2}	per square metre
m^{-3}	per cubic metre
Mm^3	million cubic meters
m s^{-1}	metre per second
m^3/s	cubic metre per second
NW/SE	Northwest - Southeast
Opa	Pakhuis Formation
Ope	Peninsula Formation
O-Sc	Cedarberg Formation
Opk	Piekenierskloof Formation
Pe	Peclet Number
Pe*	modified Peclet Number

pH	Potential hydrogen
pH_2	Pressure of hydrogen gas in water
pO_2	Pressure of oxygen in water
ppm	parts per million
PMC	percent modern carbon
PMWIN	Pre- and Post processor for MODFLOW
PVC	Polyvinyl chloride
q	heat transferred (mW m^{-2})
$\overline{Q}$	mean heat flux
$\overline{q}$	mean spring discharge
q_w	volumetric fluid flux or flow rate
q_x	horizontal groundwater flux (m s^{-1})
R&S	River and stream
s	Second
S	South
SACS	South African Committee for Stratigraphy
SABS	South African Bureau of Standards
SBM	Steenbras-Brandvlei Mega-fault
s.d.	standard deviation
SE	South East
Sg	Goudini Formation
SGD	submarine groundwater discharge
SHEMAT	Simulator for Heat and Mass Transport
Sn	Nardouw Subgroup
SR	Skuifraam
SRGT09	Skuifraam Borehole
SW	South West
T	Temperature
TDS	Total dissolved solids
TMG	Table Mountain Group
TMS	Table Mountain Sandstone
TOUGH2	numerical simulation program for multi-phase fluid and heat flow in porous and fractured media
T_m	Temperature field measurement
T_p	Temperature after calibration
T_t	Temperature after topographic correction
T_L	Temperature at lower boundary
T_U	Temperature at upper boundary
T_Z	Temperature at intermediate depth

T-z	Temperature – depth profile
USGS	United States Geological Society
v	velocity
W	Watt
WRC	Water Research Commission
W m^{-1}	Watt per metre
z	intermediate depth
z_o	characteristic length in vertical direction
3D	Three dimensional
^{18}O	Oxygen 18 isotope
^{2}H	Deuterium (Hydrogen 2 isotope)
^{14}C	Carbon 14 isotope
δD	relative concentration of Deuterium isotopes
$\delta^{18}O$	relative concentration of Oxygen 18 isotopes
ΔT	temperature change
Δz	water table relief
ρ_w	water density
β	$= \rho_w c_w q_w L / K_m$
κ_e	term that includes the effective thermal conductivity of the rock-fluid matrix
μ_w	Viscosity
°C	degrees Celsius
%	per cent
‰	per mill

1. INTRODUCTION

The Ordovician-Silurian quartzitic sandstones of the Table Mountain Group (TMG - Broquet, 1992; De Beer, 2002) are among the thickest and most extensive of their kind in the world. An early recognition (Joubert, 1970) of the groundwater potential of these major fractured-rock aquifers failed to trigger a concerted response, until the development of relatively deep (>250 m) artesian wells in the Citrusdal region (Hartnady, 1998; Hartnady and Hay, 2000), ~220 km north-east of Cape Town. Since then there has been a strong focus on further research and development on the TMG groundwater resource (Weaver et al., 1999; Hartnady and Hay, 2001; -, 2002a-e; Hay and Hartnady, 2002; Weaver et al., 2002), much of it based upon a better appreciation of the physical properties of compressibility and elasticity (Meinzer, 1928) in its deep-confined artesian basins.

The TMG resource is now receiving considerable attention as a potential long-term solution to increasing metropolitan and urban water scarcity in the Western Cape province (Umvoto, 2005; City of Cape Town, 2004), and there is new incentive for developing exploration and reservoir-characterization methodologies, and the deep drilling expertise required for this unique kind of fractured-rock aquifer. The circumstances also provide opportunity for broadening and deepening the knowledge of rock temperature gradient and heat flux in the area. However, the generally "wet" and often flowing (artesian) boreholes within the TMG terrain are not conducive to the classical modes of geothermal research, which assume that conduction is the principal mode of heat transfer.

Groundwater moving naturally through an aquifer system also transports heat, as well as solutes, and thereby alters the subsurface temperature field. Consequently temperature - the accurate field measurement of which is easy, quick, and inexpensive when compared with chemical and isotopic methods - is useful as a tracer of hydrogeological processes and for testing conceptual groundwater-flow models, as has long been recognised (e.g., Andrews et al., 1982). The temperature of groundwater provides insight into the subsurface geological processes that generate and transport heat. A common observation in mountain terrains, confirmed by computer modelling (Forster and Smith, 1988a, -b, 1989), is that the warmest spring temperatures occur for an intermediate range of fluid-flow velocities and bulk permeabilities, such that groundwater flow removes most of the geothermal heat flux advectively, but the added heat is not diluted by large volumes of water.

Accordingly, in semi-arid terrains such as the TMG in the Cape Fold Belt, which is characterized by an abundance of perennial springs that maintain base flow during dry summer months, the emergence temperatures of spring waters, in combination with their flow rates and discharge elevation (head), can be used within numerical models to solve an inverse problem involving groundwater flow velocity and basin-scale hydraulic conductivity: " ... It appears feasible that temperature measurements can be used for calculating flow velocity and that a combination of head and temperature measurements can be used for calculating aquifer permeability." (Stallman, 1963).

Large-scale groundwater abstraction, *if* it leads to a substantial alteration of the deep flow regime, should have near-field impacts on the geothermal regime around a major well-field, and possibly far-field impacts on geothermal regimes in the recharge and discharge areas along a flowpath through the abstraction site. On the grounds that

1. impacts cannot be assessed or measured without an actual experiment in large-scale abstraction, and
2. the advective transport of heat in the shallow parts of the crust is not only a geophysical, but also an ecological and environmental issue,

the planning and preparation for any future groundwater abstraction experiment should also include a careful programme in the observation and measurement of geothermal gradient, thermal conductivity, and heat flow around the proposed abstraction and environmental monitoring sites.

The geothermal research programme envisaged for the present study therefore follows two subprogramme objectives of a recent WRC research thrust, viz.:

- The development of ***predictive tools*** to assess the impact (or risk) of groundwater abstraction on the environment;
- The use of ***innovative techniques*** to determine the impact of groundwater abstraction on the environment.

It therefore augments a current WRC programme on the eco-environmental impacts of large volume groundwater abstraction in the TMG (Brown et al., 2003). For an efficiently designed hydrogeological-geophysical experiment with clear regional environmental purpose, it is a necessary prerequisite that baseline information on regional heat flow and thermal conductivity properties is obtained from suitable reference sites in the TMG and Cape Fold Belt, which is hitherto devoid of such information. The present study aims to fill that essential scientific requirement.

2. STUDY AIMS AND OBJECTIVES

2.1 STUDY AIM

The ultimate aims of this study are:

- To determine the extent to which monitoring of borehole temperatures can be used to establish underground flow rates and affects on recharge/discharge areas;
- To determine whether, where, and how utilisation of aquifers is likely to affect the subsurface geothermal regime.

2.2 OBJECTIVES

Within the broader geo-environmental context reflected in the above aims, the specific objectives of this investigation are

- To establish background geothermal gradients and heat flux (product of vertical geothermal gradient and rock thermal conductivity) in areas unaffected by underground water flow;
- To prepare a theoretical and experimental basis for the monitoring of changes in geothermal gradients and heat flux in regions where groundwater production is in progress;
- To prepare specifications for future investigations in which quantitative use is made of heat as a groundwater tracer, through numerical computer modelling of underground convective flow systems in conjunction with traditional chemical and isotopic methodologies.

3. REGIONAL GEOLOGY AND GEOTHERMICS

3.1 STRATIGRAPHY AND MAIN HYDROSTRATIGRAPHIC UNITS

The formations of the Cape Supergroup dominate the study area (Figure 3.1), but pre-Cape and post Cape formations are also described in the following sections.

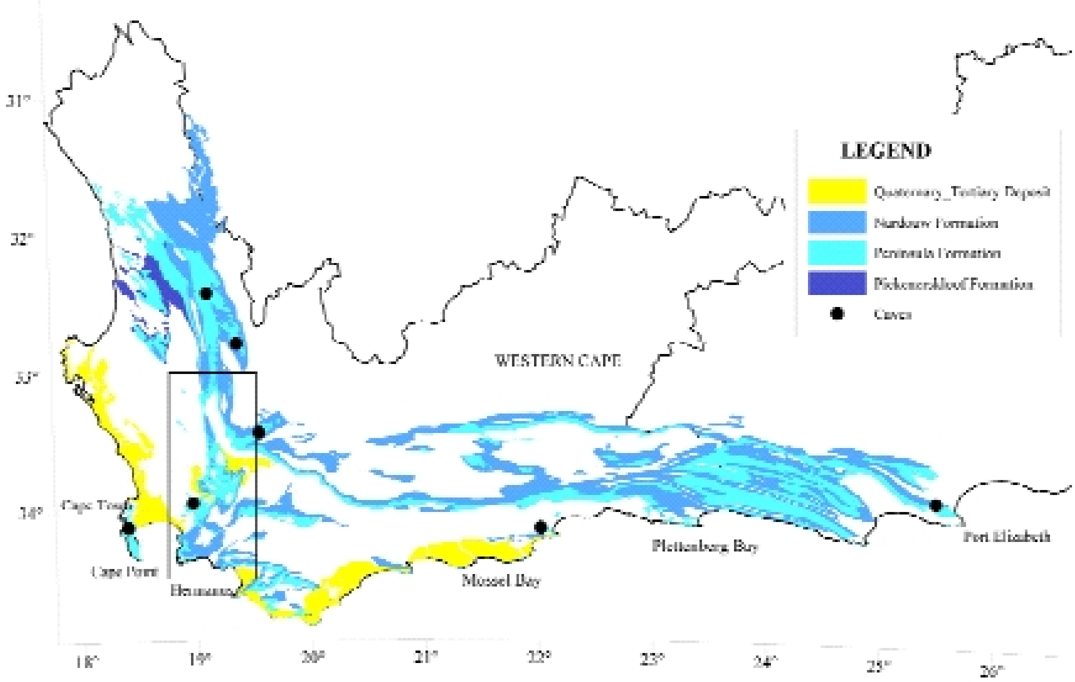

Figure 3.1 Distribution of the Cape Supergroup in the southern part of South Africa, showing the project areas (black rectangle indicates area of City of Cape Town study and red markers show Skuifraam and Stanford sites).

During an episode of compressional deformation lasting from early Permian to Triassic, the “Cape Orogenic Cycle” deformed the pre-Cape, Cape and Karoo rocks. A predominantly east-west structural grain developed in the folded Cape-Karoo strata. Northward-directed asymmetric anticlines in the competent Peninsula Formation form the scenic mountain ranges (cf. cross-section in Figure 3.2). Asymmetric parasitic folds (anticlines and synclines) characterize the overlying less massive and relatively incompetent formations. Plate movements during the late Jurassic and early Cretaceous caused extension along various large south-ward-dipping normal faults in the area (e.g., Worcester Fault) and movement continued intermittently over a very long period (Theron et al., 1991).

3.1.1 Pre-Cape Stratigraphy and structure

An extensive area to the west of the main mountain chain in the study area (Figure 3.1) is underlain by pre-Cape metamorphics and granites (i.e., the Malmesbury Group and Cape Granite Suite; Hartnady et al., 1974). Smaller inliers of pre-Cape basement underlie the wide intermontane valley between Tulbagh and Worcester, valleys in the mountains between Wellington and Villiersdorp, and in the coastal range near Hermanus. Parts of these low-grade metamorphic terrains have been intruded by pre-Cape diabasic dykes and sills. The

Malmesbury units were deformed during the Saldanian episode (Hartnady et al., 1985) of continental crust accretion and mountain-building (orogeny). Narrow outcrops of the sedimentary Klipheuwel Group, a late- to post-Saldanian unit, are mapped along major fault zones to the north and west of the study area.

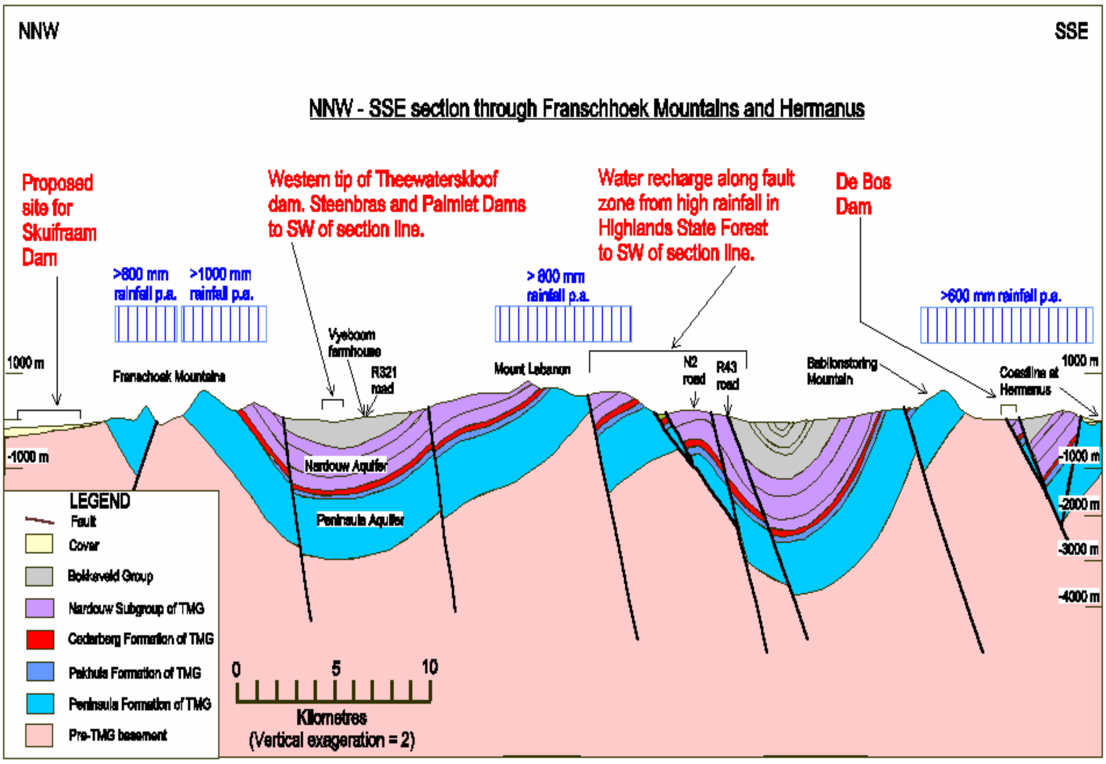

Figure 3.2 Schematic geological cross-section through southern study area

The Cape Granite Suite (cf. Scheepers, 1995) is of particular interest in the context of this geothermal study because it is an important element of the pre-Cape basement terrain beneath the two borehole sites from which thermal gradient data is here reported. The Phase I "S-type granitoids" (555-540 Ma; Scheepers and Armstrong, 2002) are typical of the southwestern (Tygerberg) terrane in the Saldania Belt, whereas the Phase II "I-type" granitoids (540-520 Ma) are characteristic of the central (Swartland) terrane (cf. Scheepers, 1995, Fig. 1). Concerning the main heat-producing elements – potassium (K), uranium (U) and thorium (Th) - in the Phase I granites that underlie the Skuifraam and Birkenhead (Stanford area) sites discussed in the present investigation (see Section 4 below), the K_2O contents of representative S-type samples – mostly obtained from the north-western part of the Tygerberg terrane, rather than the south-eastern part where Skuifraam and Birkenhead are located - range between ~4.1% to 5.2%, the U content between 2 ppm and 20 ppm, and the Th content between 2 ppm and 26 ppm (cf. Scheepers and Armstrong, 2002, Table 3).

3.1.2 Cape Supergroup

Palaeozoic sedimentary rocks of the Cape Supergroup (Broquet, 1992; Figure 3.1) dominate the geology and topography of the area. This supergroup comprises three principal stratigraphic groups, of which the lower unit, the Table Mountain Group (TMG; De Beer, 2002) is the particular focus of this study. The two overlying units are the Bokkeveld Group and the Witteberg Group.

Table Mountain Group

The lowest TMG units in the southwestern Cape, i.e., the Piekenierskloof Formation sandstones and the Graafwater Formation siltstones and shales are not represented in the study area.

The Peninsula Formation (Figure 3.1) constitutes the lower aquifer in the TMG. It is the topographically dominant unit, building most of the high mountain ranges, and it is *hydrogeologically most important* in having *wide areal extent* in the areas of maximum precipitation and recharge potential; and *greatest subsurface volume* of permeable fractured rock.

The Peninsula Formation, approximately 550 m thick in the Cape Peninsula area but reaching ~1300 m in the Citrusdal region, wedges out beyond Clanwilliam in a northerly or north-easterly direction, and is completely overlapped by the Cedarberg Formation and overlying sandstone units. Eastwards within the Cape Fold Belt, however, it thickens to >2000 m in the Oudtshoorn area.

Because of its importance the thickness of the Peninsula Formation has been accurately determined at several sites within the study area (City of Cape Town, 2004). These sites are characterised by complete exposure over a short distance between basal contact and top contact with the Pakhuis or Cedarberg Formations, and by a relatively constant dip along a profile between base and top. The thickest measured section was over 1700 m in an area north of Villiersdorp. The least thickness of ~700 m was determined for a site near Sir Lowry's Pass.

In the Cape Peninsula, the formation is subdivided into a lower Leeukop Member, distinguished by repeated fining- and thinning-upward sedimentary cyclicity, and an upper Platteklip Member, which is generally thick bedded and lacks the aforementioned cyclicity (Broquet, 1992). According to Broquet (op. cit. p.165), the basal Platteklip conglomerate marks an "apparently erosive contact" that can be traced laterally on the Cape Peninsula, and for "purposes of convenience" it may be taken to represent the Lower-Upper Ordovician boundary (~460 Ma). The Leeukop-Platteklip boundary is not yet mapped distinctly in areas other than the Cape Peninsula. In the Hottentots-Holland and Hawequas mountain ranges, however, there is some indication that the Peninsula Formation may be divisible between two members, the upper more massively bedded one of which may be an excellent fractured-rock aquifer.

In the SW Cape, the diamictite or tillite of the Pakhuis Formation overlies the Peninsula Formation, with localized unconformable relationships in zones of "synsedimentary folding" caused by ice movement. This thin (generally <50 m), poorly sorted, compact and impermeable arenaceous unit is widespread in the mountain ranges of the study area. It locally appears in fresh outcrop as a blue-black, fairly massive, pebbly mudstone unit succeeded – in most areas conformably, but in a few places with apparent angular unconformity above the intraformational "folded zone" - by the argillaceous Cedarberg Formation.

In its Western Cape stratotype area, to the north of the present study, the Cedarberg Formation is composed of two members. The lower Soom Member is dominantly black carbonaceous shale, and the upper Disa Member is a cyclically repeated mudstone-siltstone sequence, coarsening and thickening upwards towards the overlying Nardouw contact zone. Exposures of the Cedarberg in the study area are generally poor, and the rock unit is highly cleaved and sheared by folding.

The Nardouw Subgroup (Figure 3.1) consists of three sandstone-dominated formations (Malan and Theron, 1989). The lower Goudini Formation has a transitional contact with the underlying Disa Member of the Cedarberg Formation. It is characterised by repeated sandstone-siltstone cyclicity, and reddish-brown weathering due to iron-oxide content. The middle Skurweberg Formation consists of thick, cross-bedded quartzitic sandstones and is a potentially important fractured-rock aquifer. In parts of the range it reaches a thickness of up to 200 m.

The upper Nardouw, Rietvlei Formation (Theron et al., 1991, p. 34) consists of grey to light-grey feldspathic sandstone, siltstone and mica-rich shale, and is hydrogeologically less prospective than the deeper Skurweberg. In the Hex River Mountains area, the Skurweberg-Rietvlei transition has been drilled during a groundwater exploration and production effort (Hartnady and Hay, 2002d) and shown to be a thin but impermeable, black shale zone (provisionally designated as the Verlorenvalley Member), which is hardly anywhere exposed but is nevertheless photogeologically evident as a “soft band” in the terrain.

Bokkeveld Group

The TMG as a whole (Figure 3.1) is confined in the subsurface by the lowermost shale unit (Gydo Formation) in the overlying Bokkeveld Group. Subdivided into a lower Ceres Subgroup of six formations and upper Traka Subgroup in these parts, the Bokkeveld is dominated by shaly strata. It has little significance as an aquifer, except where the intercalated sandstone formations (Gamka, Hex River, and Boplaas) in the Ceres Subgroup have become fractured and relatively porous in the near-surface weathered zone or “regolith” (generally <50 m depth).

Witteberg Group

Exposures of the Witteberg Group occur in a restricted area in the eastern part of the current study domain, namely, immediately south of the Caledon Fault, south-east of Villiersdorp, and more widely in areas south of Worcester and north of Ceres. These outcrop areas fall outside the present study limits.

3.1.3 Post Cape rocks

The post-Cape stratigraphy of the TMG Project area belongs to two main groupings, viz., older duricrusts (silcrete and ferricrete on pre- or Early-Tertiary land surfaces) and the Bredasdorp Group . Alluvial deposits correlated with the Bredasdorp Group of Tertiary-Quaternary age cover areas of the coastal plain south of the Kleinriviersberge in the Hermanus-Stanford area (Figure 3.1). Here the main Tertiary unit is the Waenhuiskrans Formation.

Young sedimentary strata of Tertiary-Quaternary age occur in extensive flood-plain tracts along certain stretches of major fluvial systems such as the Berg and Breede rivers (Figure 3.1).

3.1.4 Hydrostratigraphic classification

The TMG and its components are officially defined lithostratigraphic units in terms of South African Committee for Stratigraphy (SACS) protocols. Accordingly they are not susceptible to arbitrary redefinition dictated by a specifically hydrogeological perspective. Based on information from the geological map at a scale of 1 : 250 000 (Theron et al., 1991; Gresse, 1997), the DWAF hydrogeological map at a scale of 1 : 500 000 (Meyer, 2000), and work conducted by Umvoto elsewhere in the TMG terrains, a general nomenclature for relating geological (stratigraphic) units to hydrogeological units (aquifers and aquitards) was provisionally proposed (Hartnady and Hay, 2000, Table 1, p. 184).

The nature and composition (lithology) of each of the different stratigraphic units in the TMG and Bokkeveld (Section 3.1.2 above) is here considered from a purely hydrogeological viewpoint, with regard to the relative permeability of the unit, and its corresponding classification either as aquifer, aquitard, or aquiclude (Table 3.1).

TABLE 3.1 COINCIDENT HYDROSTRATIGRAPHIC UNITS OF THE WESTERN TMG

Superunits	Units	Subunits
	Gydo Mega-aquitard	
Table Mountain Superaquifer	Nardouw Aquifer	Rietvlei Subaquifer
		Verlorenvalley Mini-aquitard
		Skurweberg Subaquifer
	Winterhoek Mega-aquitard	Goudini Meso-aquitard
		Cedarberg Meso-aquitard
		Pakhuis Mini-aquitard
	Peninsula Aquifer	Platteklip Subaquifer ? (not yet separately mapped)
		Leeukop Subaquifer ?
	Saldanian Aquicludes	[Klipheuwel Group] [Cape Granite Suite] [Malmesbury Group]

The older pre-Cape aquicludes are major barriers to groundwater movement, except in the shallow weathered regolith zone overlying fresh bedrock. Regional groundwater movements are controlled by the presence of these and other aquitard units.

The TMG quartzites are stratabound aquifers (i.e., having significant fracture porosity and a permeability greater than 10–16 m2), and therefore constitute “coincident” hydrostratigraphic units (Table 3.1), as defined by Al-Aswad and Al-Bassam (1997). The Table Mountain Superaquifer is composed of the larger Peninsula Aquifer (apparent thickness approximately 1,5 km in this area) and the lesser Nardouw Aquifer (with its component subaquifers), and is the principal focus of the present study.

The overlying unit that confines the Peninsula Aquifer in the subsurface, and separates it from the overlying Nardouw Aquifer, consists of a conformable package of three aquitard units (Table 3.1). These are grouped as the Winterhoek Mega-aquitard, following terminology originally used by Rust (1967) to describe the glaciation event recorded by the Pakhuis tillite.

Hydrogeologically, the entire Pakhuis – Goudini sequence is an effective aquitard, although the Goudini Formation is considered part of the Nardouw Subgroup.

This provisional hydrostratigraphic nomenclature recognizes that the Rietvlei Formation in the Nardouw Subgroup may have an upper water-bearing zone (Rietvlei Subaquifer), separated from the larger Skurweberg Aquifer by the Verlorenvalley Mini-aquitard. However, the Peninsula Aquifer and the Skurweberg Subaquifer are the main deep aquifer targets.

The lower three sandstone units of the Bokkeveld Group are not included within the TMG hydrostratigraphic framework (Table 3.1). At some depth (>50-100 m below surface) the Bokkeveld Group as a whole is a mega-aquitard; not only is permeability low, but groundwater quality – even in the cleaner quartzitic zones - is (sometimes exceedingly) poor in terms of Total Dissolved Solids. The only part of the Bokkeveld worth hydrogeological consideration is the "weathered and fractured" zone (categorically distinguished from true fractured-rock aquifers on the 1 : 500 000 DWAF hydrogeological map series), which may also be termed a "regolith" aquifer. The overflow from the deeper Nardouw aquifers generally leaks into and percolates through this superficial Bokkeveld regolith aquifer on mountain slopes downhill from the TMG-Bokkeveld contact, and the same applies to the Saldanian regolith aquifers on slopes downhill from the basal Peninsula Aquifer contact. In these instances of "bounding aquifers", and also in relation to other Tertiary-Quaternary aquifers, which may overlie or be in lateral contact with TMG bed-rock, it may be sensible to adopt a more inclusive informal definition of "TMG-related aquizones".

In relation to the total volume of secondary fracture porosity within the deep, confined Peninsula Aquifer alone, these bounding aquifers constitute a relatively trivial resource, and are probably also the most susceptible to negative environmental impacts on groundwater-surface water interactions. On that account, there is a reasonable case for declaring them, and also the upper Nardouw aquifers, off-limits to any future large-scale agricultural or urban abstraction. It is accordingly important to map them, mainly for ecological Reserve purposes.

3.2 GEOTHERMICS CONTEXT

In the TMG terrains the fluid-mechanical advection and concentration of geothermal energy supports distinctive environmental regimes or niches that are largely buffered against extreme seasonal change, and which in turn allow characteristic microfloral and microfaunal populations to flourish. There is a dearth of hard information on what are these hot-spring microbial ecologies, and how higher floral and faunal elements in the wetland and riparian zones might depend for their nourishment on this geothermally-supported base. By analogy with conditions in other parts of the world, it might also be suspected that vertebrates, such as endemic fish or amphibians, are sensitive to the range of water temperature at certain times of the year.

Furthermore, when measured regularly in a standardised manner at key reference sites, water temperature may be a simple first-order indicator of the contribution of deep groundwater systems to base flow, and provide an inexpensive and independent check on other, more sophisticated methods of base-flow separation. For the above and other reasons, groundwater temperatures and the deeper geothermal regime are not only considered to be important ecological and environmental factors, but ones that have hitherto been largely neglected. One of the principal reasons for this neglect is the complicating role of fluid flow, especially in regions of great topographic contrast.

Finite-element modelling shows that thermal energy introduced into mountain flow systems by regional heat flux defines subsurface temperature distributions that, in turn, influence the nature of fluid flow in mountains (Forster and Smith, 1988a). Under certain circumstances increasing basal heat flux exerts a strong control by lowering water table elevation in the recharge zone (Forster and Smith, 1988b). From modelling results, it is obvious that there is a complex interdependence between regional heat flux, regional permeability and groundwater flow systems. In some parts of the world, attempts are made to calibrate regional permeability using heat flow (Deming, 1993; McPherson et al., 2001).

An extensive monograph on heat flow in South Africa (Jones, 1992) shows that there are no geothermal measurements in those parts of the Cape Fold Belt where large-scale groundwater abstraction might be contemplated. The nearest data in a comparable geological and tectonic setting are from four deep boreholes in the Southern Karoo, between longitudes 21°E and 25°E, around latitude 30.5°S. Heat-flow (q) values here range between 51 and 61 mW/m^2, around an average of 56 mW/m^2 (Jones, 1992).

Within the last few years attempted borehole geothermal surveys in the Western Cape and western part of the Northern Cape were mostly unsuccessful because of evident disturbances due to underground water flow. Apparently undisturbed temperature logs were obtained in three boreholes of the Council for Geoscience:

1. near Calvinia, where temperatures were not obviously disturbed, but there was evidence of underground water flow in the area;
2. at Hans Gat, north of Calvinia;
3. at Lamberts Bay (where the borehole was only about 200 m deep, but there was talk of deepening it).

However, no further work has been done on these results because more data is really needed to make a viable project. In the course of this current project, an initial aim is to locate as many boreholes as possible that are not disturbed by water flow, which is not easy to do from a distance and without field knowledge of the hydrogeological setting.

The modelling of recent climate change in South Africa from perturbed borehole temperatures (Tyson et al., 1998) is also relevant to hydrogeology, in terms of its possible future effects on rainfall patterns - and therefore on recharge to aquifers. Secular change in surface temperature is just one of the causative factors of borehole temperature perturbations, others being "... subsurface water flow, topographic gradients, variations in thermal properties of rocks, (and) transients associated with drilling ... " (Jones et al., 1999, p. 185).

The "signal" of subsurface water flow has to be discriminated from the other factors (as the "noise" in the data), but there is also an opportunity to focus on the mutual interactions between it and the surface-temperature signal of climate change. The conjecture is that, during times of lower surface temperature, both higher rainfall and lower actual evapotranspiration losses are expected in the high mountain recharge areas. Higher average recharge could be expected to increase the hydraulic gradients towards spring discharge zones in the valleys, and therefore increase the rate of deep subsurface water flow and associated advective heat transport. The challenge is to design a field experiment to make quantitative measurements that would test this conjecture in practice. The route to this objective includes the development of realistic 3D groundwater flow models that explicitly incorporate advective heat-transport modules.

3.2.1 Western Cape hot springs

The TMG is well known for the occurrence of hot springs, including the hottest and strongest in South Africa, namely, Brandvlei near Worcester. These geothermal phenomena provide direct evidence for deep circulation of groundwater, but locally also exert a significant control on the ambient temperature of wetland soils and surface-water streams in the discharge areas.

3.2.1.1. The Baths

The first buildings at The Baths were erected in 1739 as a military post for "piekeniers" (pikemen) to contain the Bushmen raids on Dutch farmers. The Dutch East India Company erected the first stone building and thatched bathing huts for visitors. Having gone through several cycles of dilapidation and repair, the British government sold The Baths in 1855. Several owners followed, and in 1895 a commission reported that The Baths had a good reputation for healing. The commissioners themselves decided to buy The Baths, but the Anglo-Boer War intervened, and it was sold in 1903 to James MacGregor, whose descendants still run it today.

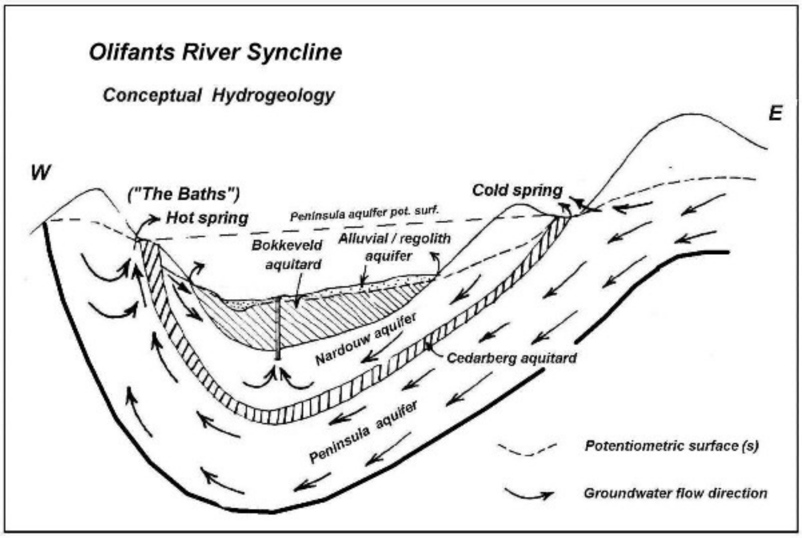

Figure 3.3. Schematic cross-section through The Baths hot spring (left). On the right is shown the fracture from which the water flows at 43°C.

The water emerges at 430C from a fracture on the side of a small tributary valley of the Olifants River. The groundwater flow at a rate of ~30 l/s is believed to derive from rainfall on the recharge area in the high Cederberg to the east, travel downwards and westwards through the Peninsula Aquifer contained by the Cedarberg shale aquitard, and emerge on the western flank of the Olifants River valley (Figure 3.3). Its electrical conductivity (EC) is still <10 mS/m after long-distance transport through the fractured orthoquartzite.

The hot spring at The Baths is clear evidence for the deep circulation of meteoric water, in this case right across the axial zone of the Olifants River Syncline (Figure 3.3). This phenomenon was the scientific inspiration for the Citrusdal Artesian Groundwater Exploration (CAGE) Project (Hartnady and Hay, 2000).

The tracing of water through the hydrometeorological cycle is facilitated by measured concentrations of the "environmental isotopes" ^{18}O and ^{2}H (or deuterium, D). These isotopes label the water naturally, and serve as indicators of groundwater source areas. The carbon isotope ^{14}C is used as a guide to the age or residence time of groundwater. In the CAGE Area, a total of 80 borehole, spring and stream/river samples were selected for ^{18}O and D isotope analyses during a routine hydrocensus survey (Hartnady and Hay, 2000). Several sites were selected for carbon isotope analyses, in order to assist in dating the aquifer water types, in particular to see if there are significant age differences in the active-recharge type TMG aquifers.

The TMG isotope data from the CAGE samples are divided between borehole (B) and surface-water (river and stream; R&S) sources (solid and open blue symbols, respectively; Figure 3.4). Further distinction is drawn between samples collected from Piekenierskloof Formation (Opk), Peninsula Formation (Ope) and Nardouw Subgroup (Sn) geological contexts. Rainfall from cumulative collector devices in the high Agter Witzenberg cluster around the values δD = -33‰, $\delta^{18}O$ = -6,25‰ (Weaver *et al.*, 1999), i.e., more depleted in D and ^{18}O than all but one of the TMG samples (Figure 3.4).

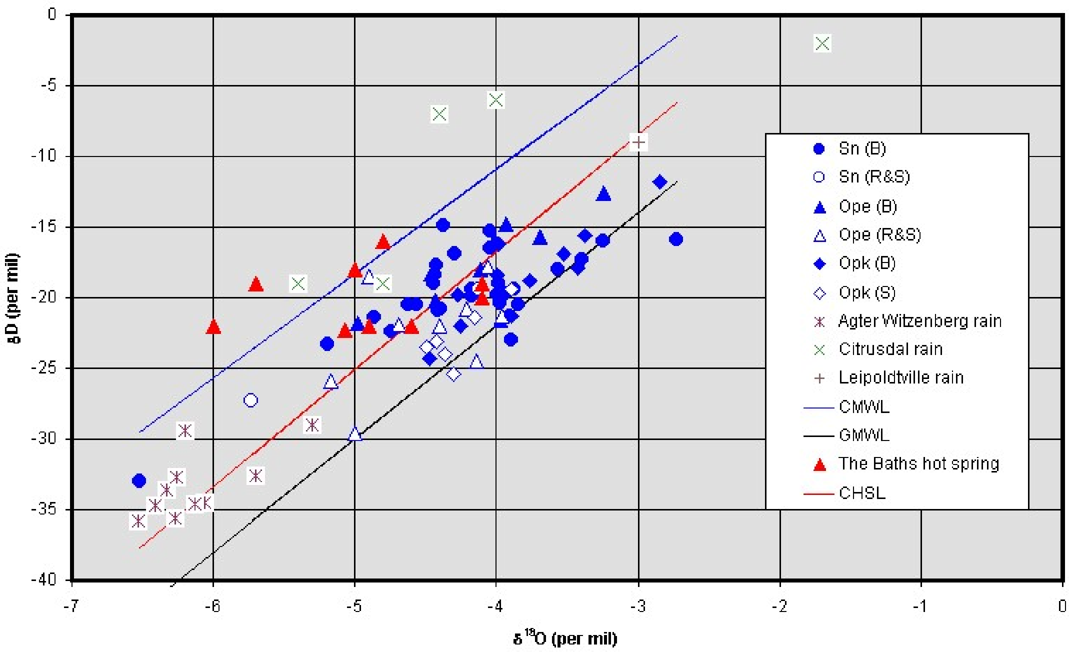

Figure 3.4. TMG isotope plot with rainfall (Agter Witzenberg, Citrusdal and Leipoldtville) and The Baths hot spring comparisons. See accompanying text for elaboration of symbols. The TMG data cluster between the high altitude (~1000 m) Agter Witzenberg rainfall samples (red asterisks) and the low altitude (<200 m) Citrusdal samples (green crosses). All TMG samples fall below the "Cape meteoric water line" (CMWL; Diamond and Harris, 2000), mainly above the global meteoric water line (GMWL), and obliquely straddle the Cape hot springs line (CHSL).

Nine monthly samples of The Baths hot spring were collected between February and October 1995 (red triangles in Figure 3.4); after Diamond, 1997). These data overlap with the TMG data cluster, but extend towards lower $\delta^{18}O$ (i.e., isotopically lighter) values. A ^{14}C result of 70.7 percent modern carbon (PMC; Mazor and Verhagen, 1983) from The Baths indicates an age or residence time along the flowpath of ~2 kyr.

3.2.1.2. Warmwaterkloof

The Warmwaterkloof, approximately 10 km south of The Baths hot-spring resort, derives its name from mixing between a cryptic hot-spring source and cold mountain-stream water on the western side of the Olifants River Syncline. The mixed water has a temperature of 26,4°C, whereas the upstream surface water temperature is in the range 17-19°C. Neither the hot-spring source temperature nor the proportion of surface-ground water mixing at this site is known.

The Warmwaterkloof ^{14}C result of 75.5±1.9 percent modern carbon (PMC; Hartnady and Hay, 2000, Table 3.1) is close to that from The Baths, again indicating an age of ~2 kyr. Depending on the degree of mixing with the stream and shallow groundwaters of more localised flow path, the mean residence time of the hot groundwaters could be much greater than this estimate (B. Verhagen, personal communication).

3.2.1.3. Artesian borehole data

Knowledge of and access to several deep (>250 m) boreholes, and also a data-base of groundwater temperature information generated during the Citrusdal Artesian Groundwater Exploration (CAGE) Project (Hartnady and Hay, 2000), is an asset for the purposes of the present project. In a CAGE Project follow-on, important geothermal data was acquired at unprecedented depth (~700 m) within the main Peninsula Aquifer of the TMG (Hartnady and Hay, 2001).

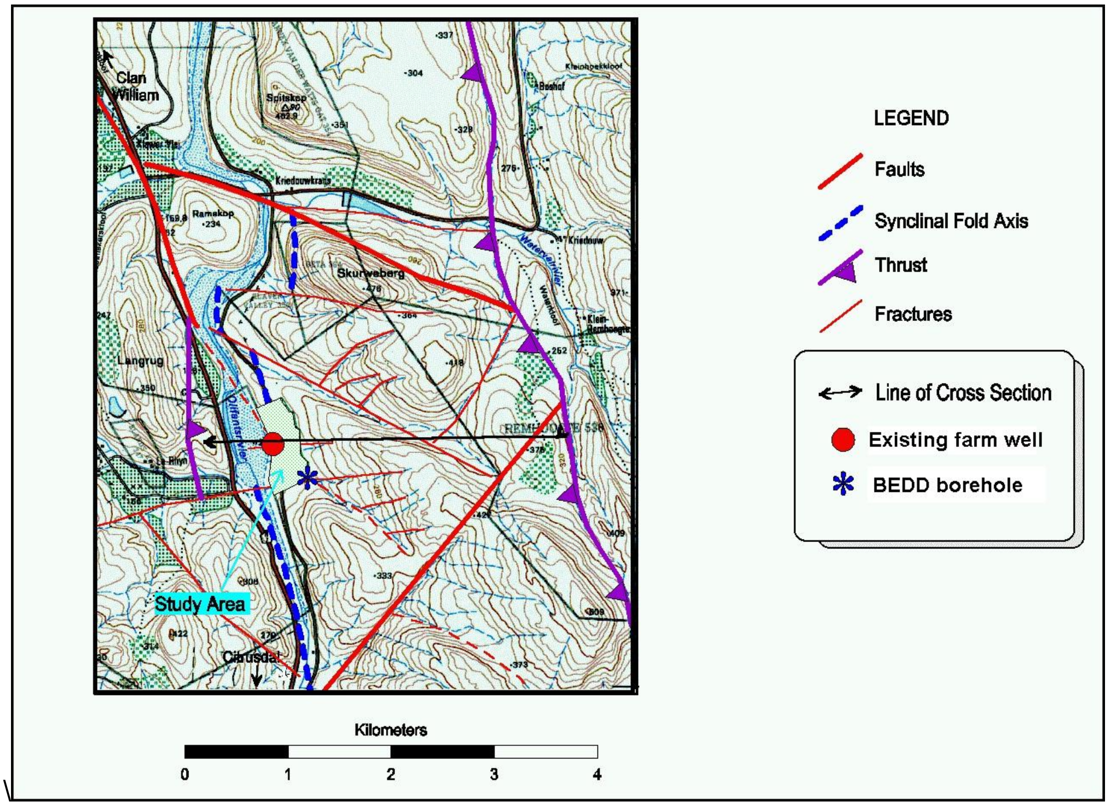

Figure 3.5. Structural map of the Blikhuis Experimental Deep Drilling (BEDD) Project site, showing pattern of faults and fractures (red lines) crossing the main synclinal hinge zone (blue dashed line), and a line of cross-section (following diagram) through the existing (BH1) borehole site (red circle). The BEDD (BH2) site was used to develop an 800 m monitoring borehole into the Peninsula Aquifer. Further drilling also occurred around the original BH1 site.

3.2.1.4. Blikhuis Borehole BH2

At the Blikhuis Experimental Deep Drilling (BEDD) Project site of the Department of Water Affairs and Forestry (DWAF), between Clanwilliam and Citrusdal (Figure 4.3), a temperature profile was measured by Mr B. Venter (DWAF) in the BH2 experimental well when it had been percussion-drilled to just over 300 m, through the Goudini (Sg) and Cedarberg (O-Sc) Formations (Figures 3.5 and 3.6). This BH2 well terminated in black carbonaceous shale; less than a half metre short of the Pakhuis (Opa)-Cedarberg contact, as was later discovered when it was deepened by diamond-drilling. At temporary development to 300 m depth the static water level in the well was ~7 m below collar, and obviously there was no groundwater flow.

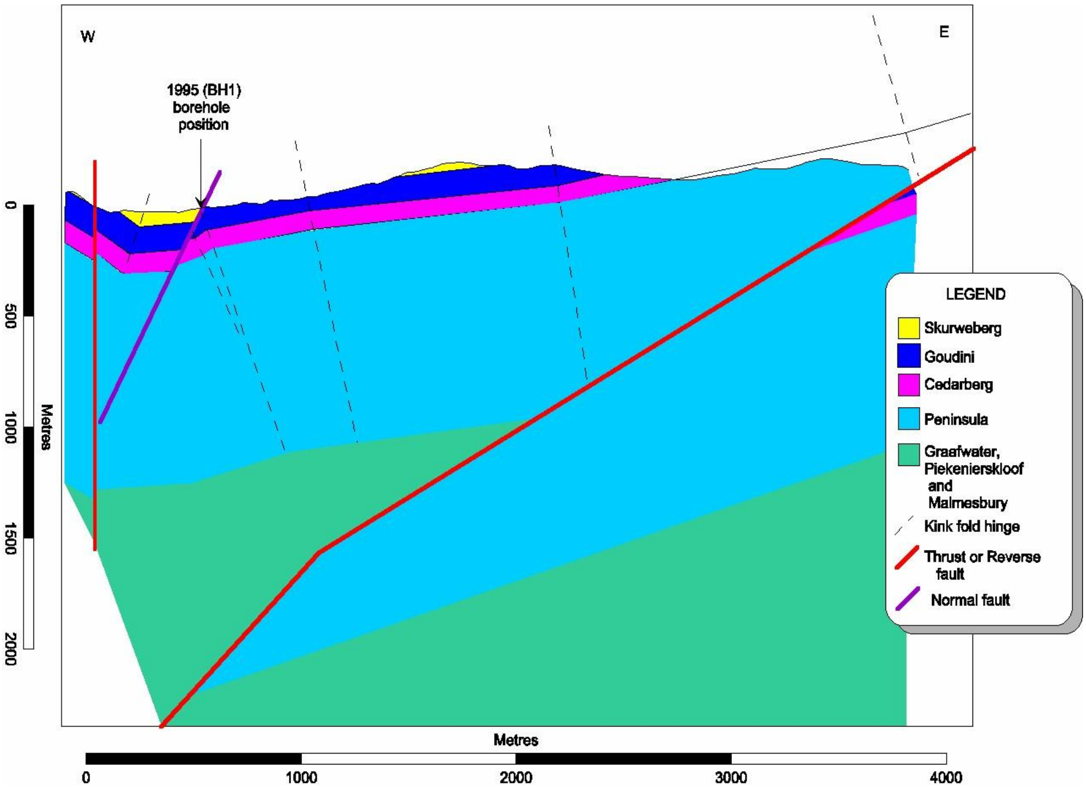

Figure 3.6. Geological cross-section through the BEDD Project area, based only on limited surface structural data and showing the true thickness and structural geometry of the Peninsula Aquifer and the overlying confining aquitard units, the possible projection of a low-angle thrust fault at depth, and fault-bend fold complexities in the hinge zone which enhance permeability in the aquifer. The 0 m level on the vertical scale refers to BH1 collar elevation at ~125 m above mean sea level. Subsequent Drilling at BH2 (BEDD site) has revised the interpreted subsurface structural geometry in finer detail.

In the gamma-ray log of the upper, percussion-drilled part of the Blikhuis BH2 borehole, the Cedarberg-Goudini contact is marked by a distinctive shale bed with high gamma characteristics (mean 71 cps) between 144.3 and 149.7 m depth (lower, blue curve in Figure 3.7). This bed is underlain by a thinner unit (149.8 – 151.8 m) with sandstone-like gamma characteristics (mean 26 cps). Between 151.9 m and 154.6 m the borehole intersects another shale bed with high gamma characteristics (mean 74 cps), overlying a lithological transition zone - possibly alternating shale-siltstone of the Disa Member - in which gamma ranges from ~30 cps at 155 m to ~60 cps at 180 m depth. The neutron log (upper, brown curve in Figure 3.7) generally mirrors the gamma log (high gamma corresponds to low neutron counts).

The pre-artesian (static) water temperature showed a distinct change in gradient around the Cedarberg-Goudini transition at ~140-180 m depth (Figure 3.8). There is a steady rise from 23.2°C at 20 m depth (~15 m below rest water level) to 24.9°C at 140 m depth, corresponding to a geothermal gradient (dT/dz) of 14.3 K km-1. From this point there is a subtle inflection to steeper dT/dz, which is roughly constant at 35.8 K km-1 between temperatures of 25.3°C at 180 m and 28.9°C at 280 m. The change in geothermal gradient coincides with the Cedarberg-Goudini transition, and is consistent with a possible change from lower thermal conductivity (K_t) in the Cedarberg shales to higher K_t in the Goudini sandstone, assuming uniform heat flow ($q = -K_t.dT/dz$).

Sedimentary rocks that are rich in organic matter, such as carbonaceous shales like the (lower part of the) Cedarberg formation, are often characterized by remarkably low K_t in the range of 0.2–1.0 W m^{-1} K^{-1}, lower by a factor of 2 or more than other common rock types (Nunn and Lin, 2002). This natural insulating effect produces elevated temperature gradients in organic-rich, fine-grained sediments, even with a typical continental basal heat flow of ~60 mW m-2. Consequently, underlying rocks will attain higher temperatures than would otherwise occur.

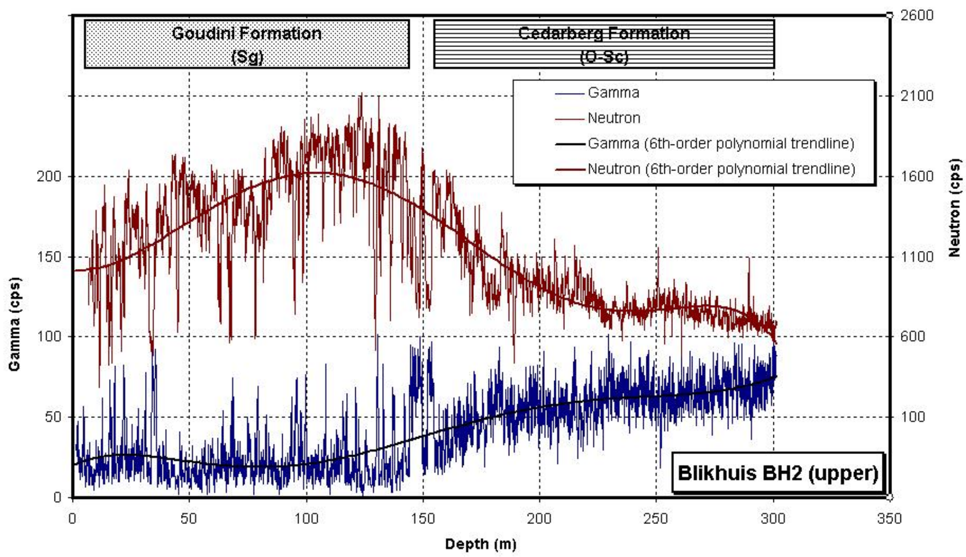

Figure 3.7. Gamma (blue) and neutron (brown) logs of the upper part of Blikhuis BH2

After deepening by diamond drilling through the Pakhuis Formation (Opa) at a depth of 324 m, and encountering the first water strikes in the Peninsula Formation (Ope) at ~340-350 m, the BH2 well began to flow at surface. At the most recent logging (June 2001), the surface flow temperature was ~28.5°C, and the bottom-hole temperature was 31.6°C (Figure 3.8). For a region of moderate heat flux, the apparent geothermal gradient in the Peninsula Aquifer is remarkably flat (~5K/km), probably signifying a convective, adiabatic gradient within a highly permeable unit.

3.2.1.5. Boschkloof (Citrusdal area)

South of Blikhuis, on the eastern side of the Olifants River valley, samples from the Boschkloof well-field near Citrusdal, from boreholes BK1 and BK2 drilled 666 m apart into the top part of the Peninsula Aquifer along a major fault zone following the contact with the overlying, steeply-dipping (>80°) Cedarberg shale aquitard, yield ^{14}C results of 79.3 ± 2.0 and 78.8 ± 2.0 pmc respectively (Hartnady and Hay, 2000). The results from these wells, which reach different depths (174 m and 314 m, respectively) and have different pumping-yield characteristics, are not significantly different.

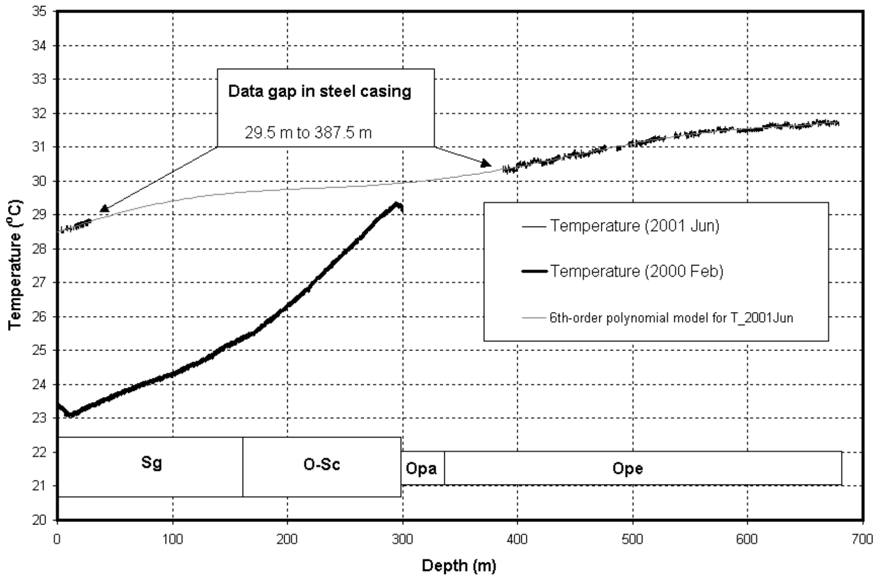

Figure 3.8. Geothermal results from the BEDD borehole, acquired in two stages: (1) in February 2000 after air-percussion drilling through the Goudini (Sg) and Cedarberg (O-Sc) aquitards; (2) in Jun 2001 after diamond-bit core-drilling through the Pakhuis (Opa) aquitard and deep into the Peninsula (Ope) aquifer. The apparent geothermal gradient within the Peninsula Aquifer is scarcely different from the gradient between surface and the base of the steel casing (389 m) in the freely flowing artesian well. The combined results show an impermeable conductive "lid" on a permeable convecting "pot".

The common ^{14}C results for BK1 and BK2, and the elevated groundwater temperatures for all these boreholes, establish a definite influx of groundwater from a relatively deep flow path into these wells. What degree of mixing there may be between this deep water and cooler water of more localised origin, encountered at shallower strikes in the fractured-rock aquifer, is a possible complicating uncertainty for interpretation of the ^{14}C-derived mean residence time. The possible recharge area for groundwater following the deep flow path is ~40 km distant in a south-easterly direction along the Elandskloof-Boschkloof fault zone.

3.2.2 Temperature Data From Surface Waters And Non-Flowing Boreholes

Water temperature measurements in boreholes and streams provide a potentially important source of information about deep groundwater flow paths within the TMG aquifer system, and have therefore been recorded during routine surveys as part of the TMG groundwater exploration programme for the City of Cape Town (CCT). A GIS-based analysis of water temperatures recorded in four hydrocensus surveys at six-monthly intervals between April 2003 and November 2004 revealed five possible areas of geothermal anomaly, defined as the occurrence of borehole and/or surface water temperature in excess of 25°C. These possible geothermal anomalies were located at the following sites:

1. Along the Klein Drakenstein Fault near the south-eastern arm of the Wemmershoek reservoir;
2. Along the Steenbras-Brandvlei Mega-fault (SBM) trend, on the north-western side of the Theewaterskloof reservoir;
3. In the Beer River Valley, near Villiersdorp, on the north-eastern side of the Theewaterskloof reservoir;
4. Along the SBM trend in the area north-west of Grabouw;
5. In an area south-west of the Houw Hoek Pass.

The temperature results from the May/June 2005 hydrocensus (diamond symbols in Figure 4.7) generally discount the previous anomalous temperature indications at the above localities, except (5) which being outside of the exploration focus area was not revisited. In fact, most of the new results from the area between Grabouw and Franschhoek now show anomalously low surface-water temperatures (<15°C), which are ascribed to an early spell of cold and rainy weather that preceded the start of the May/June 2005 hydrocensus.

Three significant thermal springs occur within the CCT hydrocensus study area, namely Brandvlei, Goudini and Caledon. In other areas of the Western Cape, e.g., in the region around the hot spring at The Baths near Citrusdal, there are local occurrences of emergent thermal groundwater mixing unobtrusively with mountain stream flow, in such a way as to cause a relatively sudden and anomalous elevation in the temperature of the surface water flow (see Section 4.2 above). Warmwaterkloof, south of The Baths, and a stream crossing the farm Koringlandshoek, to the north of The Baths, are examples where thermal water inputs to surface flow are diffuse and are generally only detectable through along-channel mapping of water temperature, especially during mid-winter snowfall periods, when the contrast between the surface-water temperature and the groundwater temperature is greatest (Umvoto Africa, unpublished stream-survey data). These water-temperature results from the CCT programme further emphasize the importance of instrument (thermometer) calibration during the hydrocensus surveys, and the need for systematic standardization of temperature-measurement procedures, so as to ensure that measurements are strictly comparable between sites and between seasons.

The logistic requirements of a rapid hydrocensus survey during a limited time period generally preclude the measurement of full temperature depth (T-z) profiles in boreholes, which can be time-consuming. In their place, however, bottom-hole temperatures (BHTs) have often been used in the petroleum industry as a proxy source of information about the geothermal state of the subsurface reservoir. For example, in the early 1970s the American Association of Petroleum Geologists (AAPG) carried out a large effort in the Geothermal Survey of North America (GSNA) that culminated in the production of a massive data base (over 20,000 BHT points from over 10,000 wells in the US, Canada, and Mexico) and resulted in the publication of continent scale maps by the USGS (DeFord and Kehle, 1976).

In North American experience, however, BHTs recorded during the logging of a borehole are found to be not at equilibrium with formation temperatures and require a correction. In general, BHTs from shallow boreholes are too high and BHTs from deep boreholes are too low. Therefore, a BHT correction factor (Deming, 1989) has been applied in many cases to adjust the temperature recorded during logging to the true formation temperature. For example, a plot of BHT's from many wells in a particular reservoir or aquifer might show the expected increase of temperature with depth, but an unconstrained trend line through the data intercepts the surface at a temperature that is greater than the known average annual surface temperature. By forcing the trend through the lower ambient surface temperature a correction factor can be generated from the difference between constrained and unconstrained trend lines (e.g., Carr et al., 2003).

Elsewhere in the world, the utility of temperature measurements in estimating fluxes in groundwater-stream systems is now well established (Stonestrom and Constantz 2003). Temperature measurements have also proved useful in estimating groundwater flux in wetland settings (e.g., Burow et al. 2005), in reservoirs and lakes (e.g., Krabbenhoft and Babiarz 1992), and in coastal aquifers, including estimation of submarine groundwater discharge (SGD; Taniguchi et al. 2003). In the TMG terrains of the Cape Fold Belt, therefore, a systematic recording of BHTs and depth at a wide variety of locations and rock types should be instituted.

At a similar regional scale to the TMG artesian basins, McPherson and Chapman (1996) reported heat flow values from oil-wells BHT data in the Powder River Basin. While this method is not as good as using full temperature surveys, the geothermal regime is well determined given the large number of data (>3000). While the TMG lacks the deep well infrastructure to achieve the same kind of BHT coverage at present, a start can still be made.

4. PRESENT INVESTIGATION

4.1 METHODS AND PROCEDURES

4.1.1 Borehole drilling

Ideally, at least three holes unaffected by water flow are needed to establish the background thermal regime for a region. Other boreholes drilled to study the aquifers can be used for monitoring the thermal effect of subsurface fluid flow. The requirements for a successful heat flow determination are listed below.

- ***Depth.*** At least 250 m of vertical depth is needed in order to get away from near-surface effects such as topography and climatically-induced surface temperature change.
- ***Inclination.*** Preferably vertical, but generally less than 40° from vertical.
- ***Diameter.*** Minimum 40 mm.
- ***Coring.*** Ideally, core samples of minimum diameter ~35 mm are required at ~10 m interval for thermal conductivity measurements. Core should not be split.
- ***Conservation***. Boreholes should be properly conserved, preferably with continuous metal or plastic pipe to prevent collapse. Failing this, just the standpipe is left in, but this does not guarantee preservation. Holes should be capped at surface to prevent undesired access, but permit repeated accessibility for surveying.
- ***Hydrology***. Holes for background heat flow studies should be free of underground water flow.

4.1.2 Field measurements.

These involve lowering a probe down the boreholes and measuring temperature at ~5 m intervals. Normally 2-3 boreholes can be surveyed in a day, depending on accessibility. 'Background' holes should ideally be surveyed at least once approximately 3 months after cessation of drilling to ensure drilling transients have subsided. 'Aquifer' holes should ideally be surveyed at regular (~6 month) intervals to monitor changes. Measurements in the background boreholes are conveniently repeated during these monitoring surveys.

4.1.3 Laboratory measurements and analysis

The procedures used after the gathering of the field data consist of the following:

- ***Thermal conductivity measurements.*** Measurements on core samples are made using a divided bar apparatus in the Bernard Price Institute of Geophysical Research (BPI) laboratory.
- ***Data processing.*** Using In-house BPI software.
- ***Computer modelling.*** Computer software needs to be developed or purchased to analyse effects of aquifer flow on the geothermal regime.

4.2 RESULTS

The temperature data for the Skuifraam and Birkenhead sites are given in Tables 4.1 and 4.3 and laboratory conductivity measurements in Tables 4.2, 4.4 and 4.5. The borehole temperature profile and thermal conductivity measurements are plotted in Figures 4.1 and 4.2, and the thermal conductivity histograms in Figure 4.3. Tables 4.6 and 4.7 respectively summarise the conductivity data and the process of heat flow calculation.

The raw field data and data corrected for probe calibration are listed in the tables. In the case of the Skuifraam borehole it was necessary to apply a topographic correction (also listed in Table 4.1), but this turned out to be quite small.

4.2.1 <u>Borehole temperature profiles and conductivity measurements</u>

4.2.1.1. Skuifraam (Cape Granite Suite)

The Skuifraam or "SRGT09" borehole is drilled into Cape granite on the east side of the reservoir to be filled on completion of the Berg Water Project in the Franschhoek area (Figure 4.1). The temperature log shows a discontinuity with a large increase in thermal gradient dT/dz between 160 m and 190 m (Table 4.1 and Figure 4.2) that must reflect a water flow. The temperature gradients, found by fitting least squares lines to the data above and below this discontinuity (Figure 4.2) yield significantly different thermal gradients (Table 4.7).

There is no obvious lithological change in the granite at the level associated with the rapid increase in temperature gradient (Figure 4.2 left) separating the two sections of different dT/dz. However, there is some indication from the drilling fragments that a fracture system was intersected at a depth around 160-180 m.

Conductivity measurements of the Cape granite at Skuifraam (Table 4.2) were obtained from composite samples formed from the fresh chip fragments from the recently-drilled borehole. These were taken at 1 m intervals and laid out in ordered array on the drilling site. There is no significant difference in thermal conductivity above and below the discontinuity (Figure 4.2) and the overall mean conductivity was used to calculate heat flow (see Section 4.2.3 below).

TABLE 4.1. TEMPERATURE DATA FROM THE SRGT09 BOREHOLE.

Depth, m	T_m, °C	T_p, °C	T_t, °C
30	16.072	16.082	16.243
40	16.054	16.064	16.219
50	16.226	16.235	16.382
60	16.473	16.481	16.619
70	16.572	16.579	16.710
80	16.706	16.713	16.835
90	16.846	16.852	16.966
100	17.009	17.014	17.120
110	17.171	17.176	17.273
120	17.353	17.357	17.447
130	17.561	17.564	17.645
140	17.757	17.759	17.834
150	17.945	17.947	18.014
160	18.146	18.147	18.208
170	18.732	18.731	18.785
180	19.203	19.201	19.248
190	19.525	19.522	19.564
200	19.771	19.767	19.804
210	19.992	19.987	20.019
220	20.217	20.212	20.239
230	20.443	20.437	20.460
240	20.659	20.653	20.672
250	20.872	20.865	20.880
260	21.096	21.089	21.100
270	21.296	21.288	21.296
280	21.504	21.496	21.501
290	21.746	21.737	21.740

T, temperature; m, field measurements; p, after probe calibration;

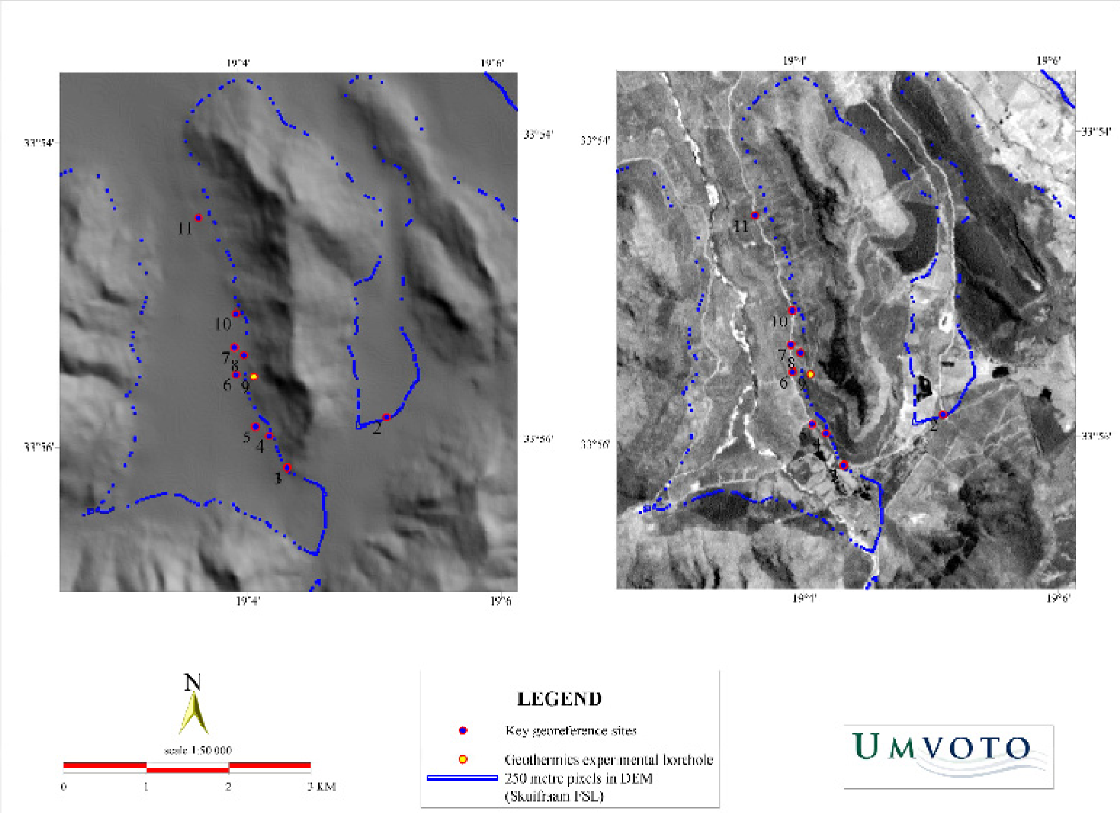

Figure 4.1 Skuifraam Location

TABLE 4.2. THERMAL CONDUCTIVITY RESULTS FOR SAMPLES OF CAPE GRANITE OBTAINED FROM THE SRGT09 BOREHOLE.

Sample	Depth, m	K_t, W m^{-1} K^{-1}
SR1	115	3.58
SR2	120	3.37
SR3	130	3.36
SR4	145	3.81
SR5	150	3.28
SR6	160	3.56
SR7	175	3.35
SR8	180	3.25
SR9	189	3.27
SR10	201	3.35
SR11	210	3.55
SR12	215	3.47
SR14	220	3.70
SR15	235	3.55
SR16	241	3.92
SR17	245	4.01
SR19	250	3.67
SR20	260	2.97
SR21	265	3.76
SR22	269	3.70
SR23	281	3.72
SR24	290	3.90
SR25	295	3.76
SR26	300	3.36

K, thermal conductivity.

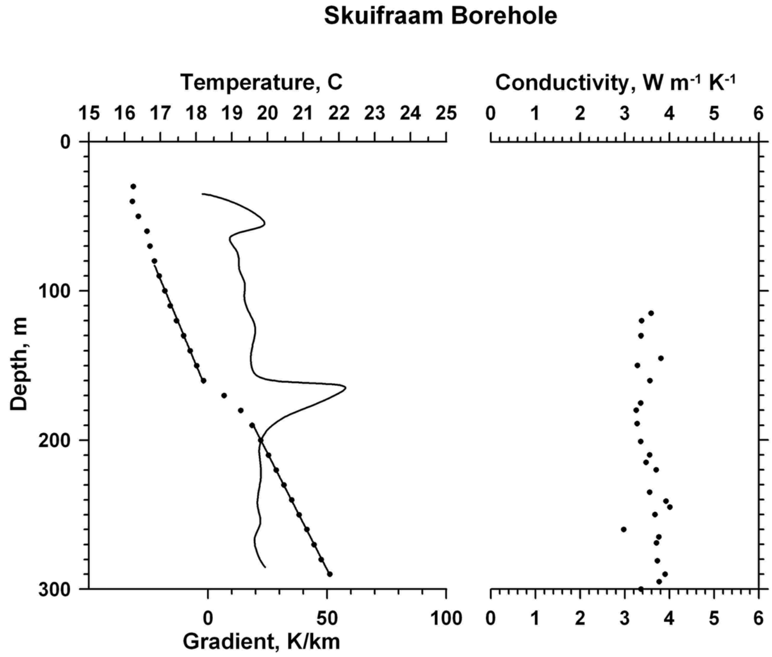

Figure 4.2. Left: Temperature (dots, top axis) and temperature gradient (solid curve, bottom axis) versus depth in the Skuifraam borehole. Lines represent least squares fits to the temperature data. The borehole temperature profile shows a rapid increase in temperature gradient between 160 and 190 m, which is indicative of water flow. Right: Thermal conductivity versus depth.

4.2.1.2. Birkenhead boreholes, Stanford (Bokkeveld Group)

Two abandoned boreholes on the Birkenhead property near the village of Stanford were drilled during May 1999 into Bokkeveld shale in the course of a failed attempt to reach a deeper fractured-sandstone aquifer. The blackish-grey shale is highly weathered to ~5 m depth and moderately weathered down to ~60 m, without obvious indications of veining and fracturing. The deeper unweathered rock is relatively impermeable and the water columns in these wells are static, with the water table about ten metres below ground level.

The temperature data (Table 4.3) from the Birkenhead wells yield relatively uniform thermal gradients (Figure 4.3), except that there is a decrease in the bottom part of the deeper hole (see steepening below the arrow at 190 m depth in graph on left hand side of Figure 4.3). Because the wells were drilled using percussion methods, and no detailed petrographic description of the shale chips was undertaken, it is not known if this gradient change could be related to a slight change in lithology. It is speculatively possible that the decrease in gradient reflects an increase in thermal conductivity, due to an increase in the content of quartzose silt in the shale, perhaps indicating a transition towards the sought-after sandstone aquifer.

TABLE 4.3. TEMPERATURE DATA FROM THE BIRKENHEAD BOREHOLES.

Borehole 1			Borehole 2		
Depth, m	T_m, °C	T_p, °C	Depth, m	T_m, °C	T_p, °C
10	18.288	18.288	10	18.336	18.336
20	18.848	18.847	20	18.663	18.662
30	18.950	18.948	30	18.812	18.811
40	19.037	19.035	40	18.973	18.971
50	19.151	19.149	50	19.134	19.132
60	19.243	19.241	60	19.299	19.296
70	19.377	19.374	70	19.458	19.455
80	19.529	19.526	80	19.614	19.610
90	19.670	19.666	90	19.771	19.767
100	19.813	19.809	100	19.936	19.932
110	19.964	19.960	110	20.089	20.084
120	20.114	20.109	120	20.246	20.241
130	20.253	20.248	130	20.408	20.402
140	20.390	20.384	140	20.554	20.548
150	20.544	20.538	147.8	20.652	20.646
160	20.691	20.685			
170	20.838	20.831			
180	20.983	20.976			
190	21.122	21.115			
200	21.247	21.239			
210	21.370	21.362			
220	21.505	21.497			
230	21.628	21.620			
240	21.742	21.733			

T, temperature; m, field measurements; p, after probe calibration.

TABLE 4.4. THERMAL CONDUCTIVITY RESULTS FOR SAMPLES OF BOKKEVELD SHALE OBTAINED FROM BOREHOLE OL1/69.

Sample	Depth, m	K_t, W m^{-1} K^{-1}
BS1	801.9	4.46
BS2	805.9	3.18
BS3	809.5	4.42
BS4	812.9	3.93
BS5	817.2	3.88
BS6	822.4	3.58
BS7	824.5	3.48
BS8	830.0	2.09
BS9	834.5	3.39
BS10	837.4	4.03
BS11	841.2	3.78
BS12	845.8	3.61
BS14	854.0	2.25
BS15	856.5	4.31
BS16	860.5	4.14
BS17	864.4	1.90
BS19	871.4	3.71
BS20	874.8	3.68
BS21	878.4	2.96
BS22	883.9	3.09
BS23	887.6	2.72
BS24	893.1	3.47
BS25	896.4	2.94
BS26	900.7	3.31

K_t, thermal conductivity.

The temperature data for the deeper borehole shows an upper layer of higher thermal gradient (low water table temperature) down to ~25 m depth (Figure 4.3).

Fresh chip samples of Bokkeveld shale were no longer available at the Birkenhead site; hence proxy conductivity measurements (Table 4.4) were undertaken on samples from a deep stratigraphic borehole (OL 1/69).

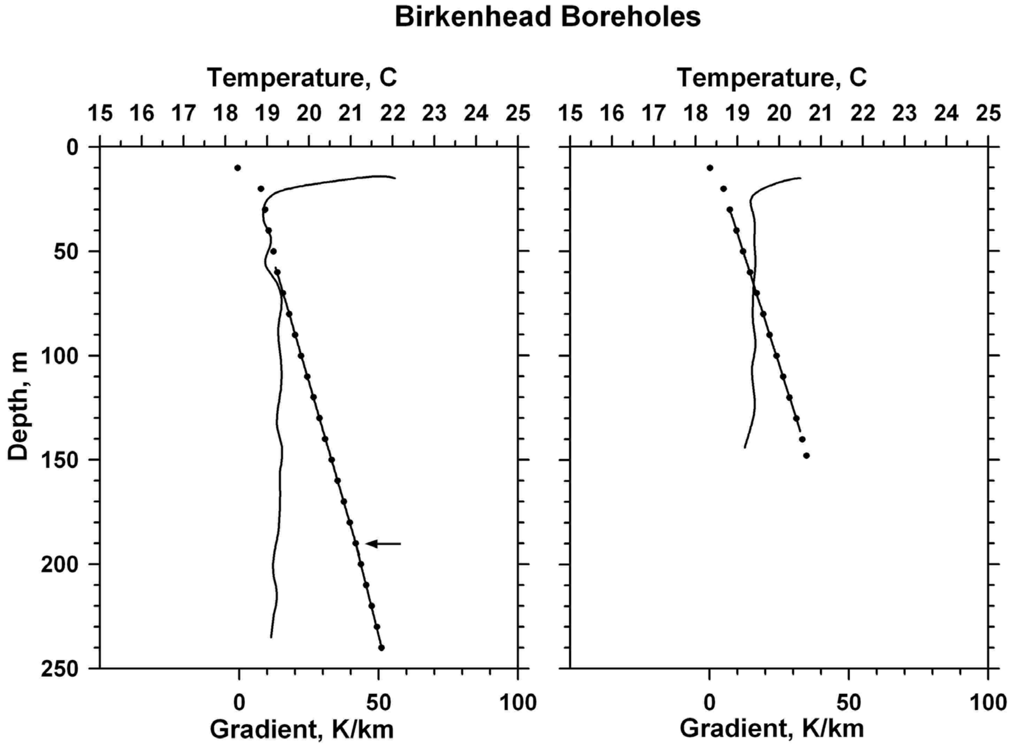

Figure 4.3. Temperature (dots, top axes) and temperature gradient (solid curves, bottom axes) versus depth in the Birkenhead boreholes. There is a decrease in the temperature gradient at 190 m in the deeper borehole (left) indicated by the arrow.

4.2.1.3. Blikhuis core measurements

In addition to the Cape granite (Table 4.2) and Bokkeveld shale (Table 4.4) conductivity data, measurements of thermal conductivity were made on samples of the Peninsula Formation (Table Mountain sandstone, Table 4.5), in order to provide better constraints on the previous interpretation of the thermal gradient data (see Section 4.3.1 below). These quartzitic sandstone samples were selected from the drill core of the BEDD borehole at Blikhuis in the Olifants River Valley, which is stored at a special core curation facility at the University of Stellenbosch, under the academic supervision of Dr H. de V. Wickens.

TABLE 4.5. THERMAL CONDUCTIVITY RESULTS FOR SAMPLES OF TABLE MOUNTAIN SANDSTONE FROM THE BLIKHUIS BOREHOLE

Sample	Depth, m	K_t, W m^{-1} K^{-1}
TMS1	370.0	6.81
TMS2	394.0	7.00
TMS3	418.0	7.51
TMS4	445.0	7.41
TMS5	477.0	7.42
TMS6	508.0	7.59
TMS7	561.0	7.09
TMS8	563.0	7.53
TMS9	570.0	7.37
TMS10	592.0	7.74
TMS11	646.0	7.03
TMS12	670.0	7.91
TMS13	692.0	7.67
TMS14	725.0	6.65
TMS15	759.0	7.19
TMS16	771.0	7.38
TMS17	800.0	7.61
K_t, thermal conductivity.		

4.2.2 Laboratory conductivity analysis

Statistical analysis (Table 4.6) and graphical presentation (Figure 4.4) of the thermal conductivity (K_t) data shows that there is little variation within each rock type, which is an encouraging sign. Results from the Bokkeveld shale (3.43 ± 0.70 W m^{-1} K^{-1}) and the Cape granite (3.55 ± 0.25 W m^{-1} K^{-1}) fall within the normal range for these lithologies, with the Bokkeveld showing a greater variability (Figure 4.4). However, the Table Mountain samples from the Peninsula Formation (7.35 ± 0.34 W m^{-1} K^{-1}), intersected in the deeper (>300 m) section of the BEDD borehole, are characterized by values in a very high range. This thermal conductivity is unusually high, substantially greater even than that for the typical Witwatersrand quartzite (6.3 $Wm^{-1}K^{-1}$; Jones, 1992, Table 2). This result indicates that the Peninsula Formation is composed of very pure quartzite, but it should be noted that these measurements made on dry core samples may not be representative of the in-situ bulk thermal conductivity of water-saturated quartzite with significant effective fracture porosity.

TABLE 4.6. STATISTICS OF THERMAL CONDUCTIVITY DATA.

Rock Type	K_t,	s.d.,	Range,	N
Bokkeveld shale	3.43	0.70	1.90-4.46	24
Table Mountain quartzite	7.35	0.34	6.66-7.91	17
Cape granite	3.55	0.25	2.97-4.01	24

K, mean thermal conductivity; s.d., standard deviation; N, number of measurements.

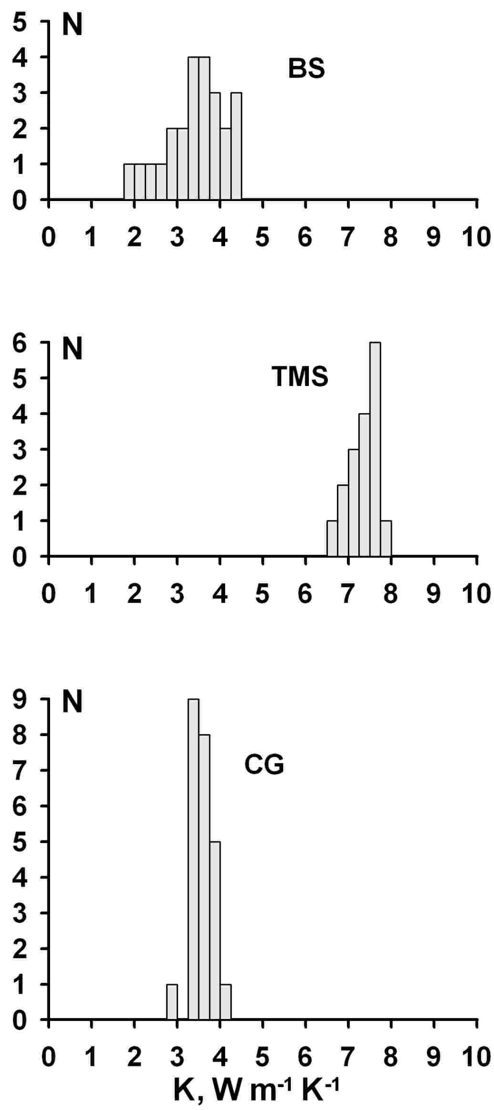

Figure 4.4. Histograms of thermal conductivity data from Cape granite (CG), Table Mountain sandstone (TMS) and Bokkeveld shale (BS).

4.2.3 Heat flow results

The heat flow results obtained in this study (Table 4.7) are calculated from the formula

$$Q = -K_t\,(\mathrm{d}T/\mathrm{d}z),$$

where Q is heat flow, dT/dz is thermal gradient, and K_t is thermal conductivity, and are tabulated according to borehole site and depth range.

For the lower borehole section (from 190m to 290 m depth) at the *Skuifraam site*, the heat flow is determined at 76 mW m^{-2} (Table 4.7). This new result is the current best estimate for the undisturbed or “background” heat flow in the Cape Fold Belt. The upper borehole interval at Skuifraam (from 60 to160 m), yields a significantly lower heat flow (65 mW m^{-2}; Table 4.7). The inferred reason for the changes in geothermal gradient is an advective transfer of heat by fluid flow in the fracture system that was intersected (see Section 4.3.1 below). Its apparent effect is to divert heat from or cool the base of overlying layer and produce anomalously high geothermal gradients (up to 60 K km^{-1}) in the narrow zone around the fracture system.

TABLE 4.7. HEAT FLOW CALCULATIONS.

Borehole	Depth Range,	dT/dz	K_t	Q
SRGT09	80-160	18.2	3.55	65
	190-290	21.5	3.55	76
Birkenhead 1	60-190	14.5	3.43	50
	190-240	12.5	3.43	43
Birkenhead 2	30-130	15.9	3.43	55
Blikhuis	20-140	14.3	7.35 (TMS)	105
	20-140	14.3	6.3 (Wits)	90
	20-140	14.3	5.31	(76) (SR)
	180-280	35.8	2.12	(76) (SR)

The heat flow result (43 mW m^{-2}) from the deeper part (190-240 m) of the Birkenhead 1 borehole may reflect an unforeseen complication due to an increase of thermal conductivity of the rock type. If this result is ignored, the calculated heat flow *Q* for the Stanford region is 50-55 mW m^{-2}, derived from the thermal gradients for upper part of Birkenhead 1 and from Birkenhead 2 (Table 4.7).

This *Q* result for Stanford is substantially lower than the background heat flow at the Skuifraam site, but there is no essential difference in the character of the pre-Cape granite basement below both sites. Both areas fall within the Tygerberg Terrane of the Saldania belt (Hartnady et al., 1974), and both contain the same kind ("S-type") of granite as a major terrane constituent (Scheepers, 1995). In the absence of a lithological or crustal-terrane explanation of the Q difference, it is alternatively suggested that the lower thermal gradients and heat flow of the Stanford region reflect an underground water flow.

This hypothesis postulates that a significant proportion (i.e., the Q difference of 21-26 mW m^{-2}) of the conductive vertical heat flow from the pre-Cape basement aquitards into the TMG aquifer systems is diverted and horizontally advected by a regional groundwater flow within the TMG. The direction of deep advection is unknown, as there are no deep boreholes into the deep confined aquifer in this area. However, but is assumed to follow a roughly E/W trend along the main topographic gradient between a recharge source in the TMG hills along the eastern part of the Stanford basin and a possible submarine groundwater discharge (SGD) zone from TMG sea-floor outcrops in Walker Bay, beyond the coastline around the Klein River estuary.

Prior to the present study, a Southern Karoo Q value of 56 mW m^{-2} was assumed for the BEDD site, and K_t values for Goudini and Cedarberg lithologies of 3.9 W m^{-1} K^{-1} and 1.6 W $m^{-1}K^{-1}$, respectively, were calculated (Hartnady and Jones, 2001). As the derived Cedarberg K_t is lower than typical Karoo sandstone/shale (2.2 W m^{-1} K^{-1}), and the Goudini value is also lower than typical Witwatersrand quartzite (6.3 W m^{-1} K^{-1}; Jones, 1992, Table 2), the Blikhuis Q could be as high as 80 mW m^{-2} and still the Cedarberg-Goudini K_t values would lie within the Karoo-Witwatersrand range.

The use of actual K_t measurements from the TMG Peninsula Formation and the Witwatersrand quartzite, as proxy K_t for the Goudini Formation, yields calculated heat flows of 105 and 90 mW m^{-2}, respectively (Table 4.7). Fixing the Clanwilliam region's heat flow at the Skuifraam (SR) value of 76 mW m^{-2}, a K_t of 5.31 W m^{-1} K^{-1} is obtained, which may be realistic for a ferruginous and feldspathic sandstone-siltstone sequence. Likewise, under the same heat-flow assumption, a K_t of 2.12 W m^{-1} K^{-1} is obtained for the main part (180-280 m) of the Cedarberg shale, which is not far below a typical Karoo shale value, and is consistent with the dark carbonaceous character of parts of the Cedarberg shale. These comparisons (Table 4.7) illustrate the need for further quantitative measurements of fundamental physical properties in the upper (post-Peninsula) TMG rock types.

4.3 INTERPRETATION AND DISCUSSION

4.3.1 Equilibrium temperature gradients

The calculations relating to the Blikhuis locality and the inferences drawn about possible lateral heat advection beneath the Birkenhead sites, justify the use of a heat flow of 76 mW m^{-2} as a typical for the whole of the pre-Cape basement terranes. Until further background heat flow studies are undertaken in other parts of the Malmesbury Group and the Cape Granite Suite, there is no reasonable alternative to this assumption.

Accordingly, it is possible to estimate "undisturbed" temperature gradients in the TMG quartzite (specifically the Peninsula Formation) and the Bokkeveld shale from the measured conductivity data, as follows:

Bokkeveld shale: $dT/dx = 76/3.43 = 22.2$ K km^{-1}

Peninsula quartzite: $dT/dx = 76/7.35 = 10.3$ K km^{-1}

These thermal gradients are those to be expected in the complete absence of any groundwater flow effects.

For the Peninsula Formation at Blikhuis, a pre-artesian thermal gradient is inferred from the difference between the February 2000 bottom-hole temperature (BHT = 29.24°C at 298 m) and the June 2001 measurement (BHT = 31.63°C at 679 m; Figure 3.8). The inferred gradient is thus (2.39 K / 0.381 km = 6.27 K km^{-1}), which is lower by about 4 K km^{-1} than the equilibrium or "undisturbed" thermal gradient calculated in the paragraph above. It is possible to interpret this difference as an advective effect due to the bypass flow of cooling water at deeper levels within the Peninsula Aquifer. Alternatively, there is a system of free convection driven by temperature-induced density differences beneath the "thermal blanket" of the confining Cedarberg shale, which lowers the thermal gradient within the underlying Peninsula Formation.

4.3.2 Fracture identification at Skuifraam

Temperature profiles measured in wells that intercept fractures often exhibit anomalies caused by movement of relatively cool or hot water in or out of the fracture. Hence temperature logs have been used to identify a fracture zone in granite, for example (Drury, 1989). In the case of the SRGT09 profile at the Skuifraam site (Figure 4.2), the sharp temperature inflection near 160 m indicates that this profile may be influenced by an isolated fracture that intersects the borehole.

Confirmation of this idea comes from a radar image of the borehole, obtained during an experimental test of the GeoMole logger, a new instrument that is currently under development at the University of Stellenbosch. The GeoMole log (Figure 4.5) shows that the granite between the depths of 60 m and 120 m is apparently a uniform and homogeneous mass, but the section centred around 160 m is marked a pattern of radar diffractions, which probably reflects structural heterogeneity. Below this level there are radar reflections inclined to the borehole axis, one of which diverges from it with depth.

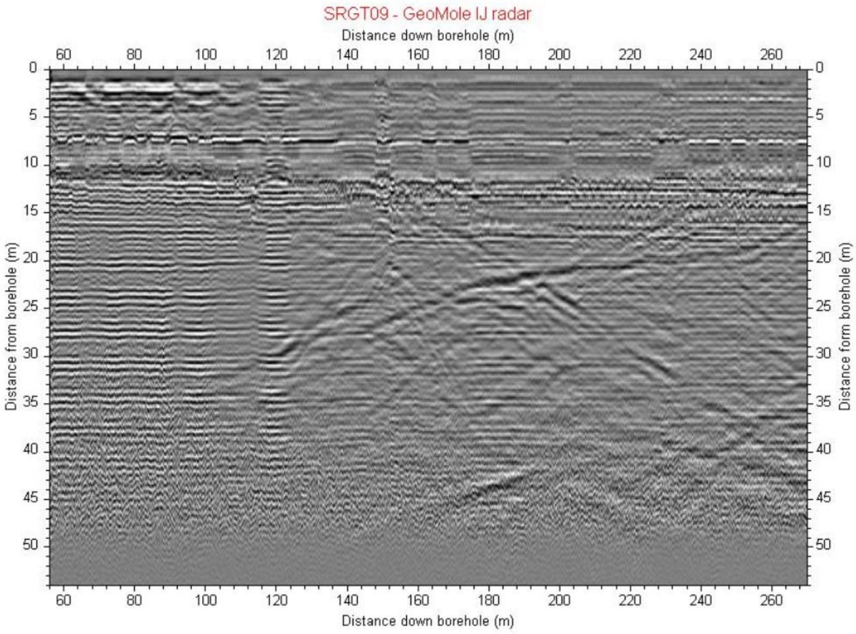

Figure 4.5. GeoMole radar image of the SRGT09 borehole (courtesy of Mr Paul van der Merwe, University of Stellenbosch).

Although the SRGT09 borehole was purposely sited on a subtle ridge feature, away from any structurally-controlled drainage feature, to ensure that it is as far as possible a "dry" borehole in unfractured terrain, it has fortuitously intersected a concealed fracture. In the regional context, this fracture could be a SE extension of a NW/SE-striking fault that downthrows TMG basal unconformity in the Drakenstein Mountains (western) block relative to the Middenberg (eastern) structural block. This fault is not shown on the newer 1:250 000 geological map (Gresse, 1997), but appears on the older 1:125 000 version (De Villiers et al., 1964). Extrapolated from its mapped locus, it runs underneath or close to the dam wall of the Berg Water Project (BWP), as is clearly evident in the displacement of the TMG-granite contact on either side of the Berg River.

The single GeoMole profile (Figure 4.5) does not alone provide an adequate basis for further testing of this fault hypothesis. The core drilling of a new borehole adjacent to the existing SRTG09 borehole is required to intersect and sample the fracture/fault zone in an up-dip direction (i.e., ENE of the present site, if the fault-related hypothesis is correct). Apart from establishing the orientation, width, internal structure, and hydromechanical properties of the postulated fault, the core samples may permit the use of indirect stress measurement techniques, such as the acoustic emission-based "Kaiser effect" (Kaiser, 1953; Kanagawa et al., 1976; Hughson and Crawford, 1986; Seto et al., 1998) to explore its degree of "failure criticality" and the potential for triggered rupture under small additional stress loads, such as the imminent filling of the BWP reservoir.

4.3.3 Implications for hot-spring sources

The emergence temperatures of spring waters reflect the interaction between the advective and conductive transport of heat in the host aquifer systems, mainly a balance between the amount of heat transported advectively and the volume of water that must be warmed (Manga, 2001). As shown in model calculations (e.g., Forster and Smith, 1989), aquifer permeability is the key factor. Where rock permeability is low, groundwater flow velocities are low, heat transport is dominantly conductive, and therefore spring temperatures are low. In rocks with high permeability, flow velocities are high and advective transport of heat is the dominant process. However, because high permeability also results in large volumes of circulating water, spring temperatures again remain low. The warmest springs generally occur for an intermediate range of permeability (Manga, 2001), which is a case inferred for the TMG, where hot springs are a characteristic feature of the terrain (Diamond, 1997; Diamond and Harris, 2000; Harris and Diamond, 2002; Meyer, 2002).

Consider that a mean heat flux $\overline{Q}$ enters the base of an aquifer system, and assume that all this heat power $H = \overline{Q}A$, where A is the area of the aquifer base, is advected horizontally by groundwater flow in the system and is then discharged at a spring. The total heat discharged at the spring is related to the temperature change ΔT by (Manga, 2001):

$$H = \rho_w c_w \overline{q} \, \Delta T$$

where ρ_w and c_w are water density and heat capacity, respectively, and $\overline{q}$ is the mean spring discharge. In the case of the Brandvlei hot spring, there is a discharge of 127 l/s or 0.127 m^3/s at a temperature of 64°C (Meyer, 2002). Assuming that the mountain recharge temperature is about 14°C or less, the temperature change between recharge and discharge zones is not less 50 kelvins (50 K). Given a water density of 1000 kg/m^3 and heat capacity of 4184 joule per kilogram kelvin (4184 J kg^{-1} K^{-1}), the Brandvlei spring discharges at least 26 568 400 watts (~27 MW) of geothermal heat. For comparison with the Brandvlei geothermal output, the output of one unit of the Koeberg nuclear reactor near Cape Town is 900 MW, the total amount of heat discharged by the large and very active hydrothermal system at the large Yellowstone volcanic province in the USA is about 5×10^3 MW (Fournier 1989), and the total worldwide geothermal power being exploited in 1995 was 8.7×10^3 MW (Freeston 1996).

If the mean heat flux or background heat flow to the aquifer base is 76 mW/m^2 (or 0.076 W/m^2) as determined from the deeper Skuifraam result, and it is all advected to the Brandvlei spring by a downward recharge flux, the source area A for this advective flux is ~350 km^2. The Brandvlei hot spring is located within the H40E quaternary subcatchment, which has a gross area of 285 km^2 (Midgley et al., 1994a, p. 8.4). The H40E area is bounded on the west by the H10K subcatchment (194 km^2; op. cit., p. 8.3), which in turn is bounded on the south by the H60B (210 km^2; op. cit., p. 8.4) and H60C (217 km^2; op. cit., p. 8.4) subcatchments. Of this total area (906 km^2), if only one third contributed to the groundwater recharge to Brandvlei, it would account for most of the advected heat output from the spring.

The Brandvlei example illustrates that combined thermal and fluid flow measurements made at springs integrate the signal of geological and hydrological processes over large spatial areas and possibly long periods of time (Manga, 2001). However, the springs do not always directly measure the geological or hydraulic properties of interest. Thus, the interpretation of these measurements requires the development of a model or mathematical framework, within which a suite of measurements are related to physical and hydrogeological processes.

4.3.4 Regional permeability and hydraulic conductivity estimation

Investigations of coupled heat and groundwater flow elsewhere in the world have provided estimates of regional permeability, and these can serve as examples for the kind of study that can and should be undertaken within the TMG terrains of the Cape Fold Belt.

For example, borehole-scale permeability estimates in the North Slope of Alaska (Deming et al., 1992; Deming, 1993) were performed on the basis of data from core measurements made parallel to bedding and well tests compiled for a number of different geologic units. Permeabilities k derived from the arithmetic mean of core-sample measurements range from 2.2×10^{-13} m^2 for sandstones to 1.1×10^{-16} m^2 for limestones. Even if a continuous and impermeable layer of permafrost extends over the part of the model domain representing the entire foothills and coastal plains provinces of the North Slope, the general trend of the heat flow data is explained by the existence of a regional groundwater flow system. A comparison of the arithmetic average of the k measurements on cores with the basin-scale k inferred from heat- and fluid-flow modeling suggests that the average permeability of the basin as a whole does not increase significantly from the core to the basin scale (Deming, 1993).

In a two-dimension study of the Powder River basin, McPherson et al. (2001) hypothesize that meteoric water recharged in the Black Hills causes the depressed surface heat flow in the northeastern part of profile, from which groundwater is topographically driven deep into the basin and warmed by ambient heat. The heated groundwater is subsequently driven up-dip on the southwest side of the basin axis and discharged in the area of the Salt Creek Anticline, elevating the surface heat flow in that area to >200 mW m^{-2}.

A mathematical model of coupled fluid flow and heat transfer is used to test this hypothesis. McPherson et al. (2001) evaluated the system using a two-dimensional finite difference model coincident with the locus of surface drainage in the basin. The model domain is 230 km horizontal by 5 km vertical, consisting of 5000 grid-blocks (100 horizontal by 50 vertical), each 2300 m by 100 m. Eight model units are delineated in the cross-section, each of which represents different regions of homogeneous permeability and thermal conductivity included in the model. The permeability ranges in the aquifers were $k_{PB} = 5.0 \times 10^{-14}$ m^2 to 2.0×10^{-13} m^2 and $k_{NB} = 1.0 \times 10^{-17}$ m^2 to 1.0×10^{-16} m^2 (where k_{PB} = permeability parallel to the bedding plane and k_{NB} = permeability normal to the bedding plane). The shaley units were assigned regional permeability values of $<10^{-17}$ m^2.

In order to match model surface heat flow to observed heat flow at the Salt Creek Anticline, the permeability calibration required the simulation of fractures in the apex of the anticline with higher, isotropic ($k_{PB} = k_{NB}$) permeability. Thus, for the regional groundwater flow system to be the primary cause of the anomalously high heat flow observed over the anticline, the anticline must be fractured enough to provide a high permeability conduit (McPherson et al., 2001).

The application of coupled heat and groundwater flow modeling, of the kind summarized above, to the TMG aquifer system has yet to be undertaken. In the following section, the necessary elements for such an investigation are reviewed.

5. TOWARDS A HYDROTHERMAL MODEL OF THE TMG AQUIFER SYSTEMS

5.1 BASIC THEORY

Until recently the hydrogeological literature lacked a critical synthesis of the larger body of work on heat as a groundwater tracer. A review paper by Anderson (2005) now fills that void and shows that groundwater temperature data and associated analytical tools are currently underused and have not yet realized their full potential.

The thermal advective effects of groundwater flow have been recognized for over four decades (Bredehoeft and Papadopulos, 1965; Domenico and Palciauskas, 1973), and have been quantitatively modelled since the early 1980s. Stallman (1963) conceived the idea that head and temperature measurements could be used jointly in numerical models to solve the inverse problem for ground water flow and hydraulic conductivity. Cartwright (1970) pioneered the use of temperature measurements at the basin scale, using the Bredehoeft and Papadopulos (1965) model to calculate ground water discharge in the Illinois Basin. Since then, numerous modeling studies of hypothetical ground water basins have demonstrated theoretically that ground water flow causes perturbations in the thermal regime in two dimensions (e.g., Smith and Chapman, 1983) and three dimensions (Woodbury and Smith, 1985).

For example, a numerical simulation of a 5-km-deep by 40-km-long intermontane-type basin with a linearly sloping water table and 500 m of total water table relief, a relatively large vertical/-horizontal aspect ratio and large water table slope (Smith and Chapman, 1983), considered a range of permeability k over which heat advection was negligible ($k < 10^{-18}$ m^2) to highly significant ($k > 10^{-16}$ m^2). At $k = 5 \times 10^{-16}$ m^2, heat flow varies from 20 mW m^{-2} at the recharge margin of the basin to 160 mW m^{-2} at the discharge margin. Temperatures at similar depths in the recharge and discharge areas differ by as much as 50°C (cf. Manning and Ingebritsen, 1999, Fig. 4).

Analyses of coupled groundwater flow and heat transport in the upper crust typically infer permeability k in the range of 10^{-17} m^2 to 10^{-14} m^2, with a distribution that is positively skewed about a mean value somewhat greater than 10^{-16} m^2 (Manning and Ingebritsen, 1999) . This mean value is about 2 orders of magnitude lower than the average crustal permeability suggested by Brace (1984) but is reasonably consistent with the compilation of permeability data by Clauser (1992). A permeability of 10^{-16} m^2 is therefore recognized as the approximate threshold value for significant advective heat transport in a fairly wide range of upper crustal contexts. Permeability (k, units of m^2) and hydraulic conductivity ($K = \rho_w g k / \mu_w$, units of m s^{-1}) are related for water density ρ_w and viscosity μ_w at 15°C.

For the particular case of uniform upflow between constant-temperature boundaries:

$$(T_z - T_U)/(T_L - T_U) = f(\beta, z/L)$$

where

$$f(\beta, z/L) = [\exp(\beta z/L) - 1]/[\exp(\beta) - 1],$$

$$\beta = \rho_w c_w q_w L / K_m,$$

T_U, T_L, and T_z are temperatures at the upper and lower boundaries and at an intermediate depth z, L is the distance between the constant-temperature boundaries, c_w is the heat capacity of liquid water, q_w is the volumetric fluid flux or flow rate defined by Darcy's law, and K_m is the thermal conductivity of the porous medium (after Bredehoeft and Papadopulos,1965; Manning and Ingebritsen, 1999).

In a two-dimensional analytical solution of heat transport in a cross section of a ground water basin, Domenico and Palciauskas (1973) showed that significant advective perturbation of the thermal field can be expected when the dimensionless ratio

$$(\rho_w)^2 c_w g k B \Delta z / 2\mu_w K_m L$$

reaches a value of the order of 1, where B is basin thickness, Δz is the water table relief, and L is basin length. There is an analogy between this ratio and the β ratio of Bredehoeft and Papadopoulos, 1963), in which $\rho_w g k \Delta z/\mu_w L$ approximates a basin-scale flow rate and is thus analogous to qw in the expression for the β parameter (Manning and Ingebritsen, 1999).

The Peclet number - a dimensionless ratio of convection to conduction - quantifies the potential for forced convection - the transfer of heat owing to the circulation of ground water driven by recharge and discharge - to perturb the geothermal gradient. The thermal Peclet number, Pe, is given by Anderson (2005, equation 3, p. 955) as:

$$\mathrm{Pe} = \rho_w c_w q L / \kappa_e$$

where q is the "seepage velocity or specific discharge vector", L is a characteristic length, and κ_e is "a term that includes the effective thermal conductivity of the rock-fluid matrix" (op. cit., p. 953). This expression is identical in its form to that for the parameter β in the derivation of Bredehoeft and Papadopulos (1963), where the characteristic length L represents the vertical direction and dimension (z_o in the notation of Anderson, 2005) in an upflowing system, which is also the main direction of heat flux.

For basin-scale flow and in order to include the geometry of the basin, Anderson (2005) multiplies the Peclet number by the aspect ratio, A = z_o/L_B, thus introducing the "modified Peclet number, Pe*" (op. cit., equation 7, p. 961):

$$\mathrm{Pe}^* = (\rho_w c_w / \kappa_e)\,(q_x z_o \mathrm{A})$$

where q_x is the horizontal ground water flux through the basin. This equation is also obtained (Van der Kamp, 1984) by taking the ratio of the amount of heat convected horizontally (approximately equal to $\rho_w c_w q_x z_o \Delta T$, where ΔT is the change in temperature between the top and bottom of the system) to the amount of heat transferred vertically by conduction (approximately equal to $\kappa_e L_B \Delta T / z_o$).

Ground water flow will affect the basin thermal regime when Pe*, the modified or "basin Peclet number" is $>10^{-1}$, and thermal springs occur when the Pe* ~1 (van der Kamp and Bachu 1989). When it is >5, the system will be dominated by ground water flow and essentially isothermal (Woodbury and Smith 1985). In the case of the confined artesian basins of the TMG Peninsula

Formation, the occurrence of thermal springs is a clear indication of a basin Peclet number close to 1.

The form of the basin Peclet equation – respectively involving thermal and hydraulic coefficients in the left- and right- bracketed terms - shows that the importance of convection increases with flow through the basin ($q_x z_o$) and aspect ratio (A = z_o/L_B). In other words, for basins of length L_B, perturbation of the thermal regime is more likely where there is deep circulation of ground water (i.e., large z_o) and/or high velocities (Anderson, 2005). In the case of the Peninsula Aquifer, z_o is indeed large (at least 3000 m), due to the great thickness of the Peninsula Formation (~1500 m) and its synclinal folds that bottom around 3 km below sea-level. For folding with ~10 km half-wavelength, the ratio A has a value around 0.3, and the Darcy velocity remains to be determined.

Assuming standard values for density (1000 kg m^{-3}), heat capacity (~4200 J kg^{-1} K^{-1}) and effective (rock-fluid) thermal conductivity (~7 W m^{-1} K^{-1}), the thermal term in the Peclet question has an approximate value of 6×10^5 s m^{-2} . For a TMG hydrothermal system with modified Peclet number (Pe*) approximately equal to 1, the inferred regional flow velocity q_x is about 0.015 m $s^{-1.}$

5.1.1 Instrumental developments

Inexpensive and widely available, waterproof temperature loggers facilitate the recent revival of interest in subsurface temperatures. Groundwater temperature is now measured easily and rapidly by simply lowering a thermometer down a borehole, provided that care is taken to ensure that the recorded temperature is representative of water in the aquifer and not influenced by movement of water in the borehole. In recent applications, thermocouples and thermistors are used to obtain a time series of measurements remotely, and airborne thermal sensors are used to detect areas of ground water discharge (Becker 2006).

The newer Distributed Temperature Sensing (DTS) technology, based on fibre optics, enables the simultaneous online registration of temperature profiles along one or more boreholes with a maximum spatial resolution of 0.25 m and a minimum temporal sampling interval of 7 s. After an individual calibration of the fibre-optic sensor cables, a resolution of 0.3°C of the measured temperature data is achieved (Henninges et al., 2003). DTS can be used in conventional wireline logging mode, but a special feature is the potential to install the sensor cables outside the borehole casing. In this “smart casing” mode, the fibre-optic cables are attached with cable clamps to the outer side of the casing at every connector, at intervals of approximately 12 m. The clamps enable a defined positioning of the cable around the perimeter of the casing and protect the cable from mechanical damage during installation. After completion of a cased well, the sensor cables are thus located in the cement annulus between casing and borehole wall; so that if access to the well is required for pumping equipment, or if the well is sealed and abandoned, the ongoing monitoring of the full temperature profile is still possible (Henninges et al., 2005).

5.1.2 Computer modeling capacity

Generic codes with user’s manuals and graphical user interfaces (GUIs) are available to solve coupled ground water flow and heat transport problems (Anderson, 2005, Table 2, p. 955).

Some of these codes were designed to solve complex problems involving geothermal systems (e.g., TOUGH2) or geologic processes (e.g., SHEMAT), while others (e.g., FEFLOW) were introduced for simulation of shallow aquifers. TOUGH2 and SHEMAT are both three-dimensional, finite-difference codes, and FEFLOW is a three-dimensional finite-element routine.

In general, the groundwater flow equation is coupled to the heat transport equation through the velocity term. Velocity is dependent on hydraulic conductivity, which is partly governed by fluid density and viscosity, both of which vary with temperature. Because changes in viscosity with temperature are greater than changes in density, some codes consider the temperature dependence of viscosity but assume a constant density fluid. To incorporate the nonlinearity caused by the dependence of hydraulic conductivity on temperature, the ground water flow and heat transport equations are solved iteratively to allow feedback between the solutions within a time step (Anderson, 2005).

The TOUGH family of codes is a suite of computer programs for the simulation of multiphase fluid and heat flows in porous and fractured media with applications to geothermal reservoir engineering, nuclear waste disposal in geologic formations, vadose zone hydrology, environmental remediation, oil and gas reservoir engineering, and other mass transport and energy transfer problems in complex geologic settings. Many modifications and enhancements have been made to TOUGH, at Lawrence Berkeley National Laboratory (LBNL) and elsewhere, from the time it was first released in 1987 to the present TOUGH2 version. TOUGH2 and its various descendants are currently in use in a wide variety and number of research laboratories, private companies, and universities in many countries. McPherson et al. (2001) used the numerical simulator TOUGH2, an integrated finite difference model that simulates coupled heat and fluid transport in variably saturated porous or fractured media (Pruess, 2004; Pruess et al., 1999). Fluid flow is described with a multiphase extension of Darcy's law, and heat flow occurs by conduction and advection.

The 3-D finite-difference (FD) code SHEMAT (Simulator for HEat and MAss Transport) handles the numerical simulation of reactive transport in porous media using the simulation package SHEMAT/Processing SHEMAT. It is an easy-to-use, general-purpose reactive transport simulation code for a wide variety of thermal and hydrogeological problems in both two and three dimensions (Clauser, 2003). User-friendly pre- and post-processors facilitate the setting up, running, and viewing the results a model, all from one platform. The graphic user interface (GUI), Processing SHEMAT, is based on the program PMWIN (Chiang and Kinzelbach, 2001), a widely used interface for the hydrogeological modelling program MODFLOW. SHEMAT is essentially an academic freeware product, and legal access to updated versions of the software is available at nominal cost for the purchase price of the manual (Clauser, 2003).

FEFLOW (Finite Element subsurface FLOW system; Diersch, 2004) is a finite-element (FE) model using a triangulated mesh. FE models can treat complex geology more easily than finite-difference (FD) models, particularly linear and point features such as faults and boreholes. In addition such models are capable of treating density driven and heat flow. FE models are less sensitive to scale problems resulting from a coarse mesh, since they extrapolate head across the mesh between nodal points. The results are more physically real and more accurate than FD models. The issue of scale makes these models more practical for application over large areas, model domains containing many boreholes, 3-D geology and hydrogeology, and relatively rapid lateral and vertical changes in aquifer parameters. A FE model will have to be used for any modelling specific to the TMG terrain.

FEFLOW features simulation capabilities that include:

- 3-dimensional, or 2-dimensional systems (plan view, cross-section, or axisymmetric)
- Transient or steady-state groundwater flow
- Multiple free surfaces (perched water tables)
- Chemical mass and heat transport
- Density dependent flow (e.g., saltwater intrusion)
- Unsaturated flow and transport
- Fracture modelling

The use of FE models allows the development of physically realistic models of the surface and subsurface. The advantage of FE models is that the topographic, geological and hydro-geological settings within the model domain can be translated into a physically real mesh. All relevant processes of water movement in the subsurface are incorporated and can be modelled.

The interaction with surface water bodies can be simulated in a similar fashion as described for the FD models. The advantage of the FE model in this regard is the possible interaction via the unsaturated zone and the possibility to account for changing river width due to flooding.

The current issue FEFLOW is Version 5.2, which was released early 2005.

6. CONCLUSIONS

The following main conclusions arise out of the present study:

1. The best estimate of the background crustal heat flow from the pre-Cape basement terrain in the Western Cape Province is 76 mW m^{-2}, obtained from a thermal gradient of 21.5 K km^{-1} in the deeper (190-290 m) interval of a 300 m borehole into the Cape granite at Skuifraam in the Berg Water Project area;
2. The Cape granite underlying the Skuifraam suite has a mean thermal conductivity of 3.55 ± 0.25 W m^{-1} K^{-1};
3. The thermal gradient in the upper part of the Skuifraam borehole (18.2 K km^{-1}) is lowered by possible shallow heat advection in a fracture intersected by the borehole at ~160 m depth;
4. Groundwater moving through the TMG aquifer system transports substantial quantities of heat (amounting to at least 27 megawatts at the Brandvlei hot spring alone) and thereby alters the subsurface temperature field, so that the measured heat flow in overlying stratigraphic units, such as in the Bokkeveld Group at the Birkenhead site near Stanford, is locally lowered to 50-55 mW m^{-2} (and may locally be raised elsewhere around discharge points);
5. The mean thermal conductivity of the Bokkeveld shale (as measured in core samples from another locality) is 3.43 ± 0.70 W m^{-1} K^{-1};
6. The thermal conductivity of quartzite samples from drill-core in the Peninsula Formation of the TMG yields a high value of 7.35 ± 0.34 W m^{-1} K^{-1}, which is probably due to its very pure quartzose composition;
7. Unambiguous interpretation of earlier thermal-gradient measurements at the Blikhuis Experimental Deep Drilling (BEDD) site, between Citrusdal and Clanwilliam, requires further thermal conductivity data on higher units (Cedarberg shale and Goudini Formation) in the TMG, but observed thermal gradients in the BEDD borehole (14.3 K km^{-1} in the Goudini sandstone and 35.8 K km^{-1} in the underlying Cedarberg shale) are consistent with the same background heat flow as the Skuifraam determination;
8. The high thermal gradient in the Cedarberg shale is probably related to a low thermal conductivity (~2 W m^{-1} K^{-1}), indicating that this stratigraphic unit is not only an effective aquitard but is also a good insulator or "thermal blanket" above the hydraulically and thermally conductive Peninsula Aquifer.
9. The quantitative spatial mapping and in-situ temporal monitoring of local geothermal gradients and spring-discharge temperatures, in association with the combined modelling of fluid and heat advection in the TMG aquifers, could provide a powerful and relatively inexpensive new tool, both for groundwater exploration (storage and flow determination) and for the monitoring and interpretation of impacts due to large-volume abstraction.

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