

Groundwater prospecting

A HYDRO-GEOPHYSICAL STUDIES

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The success of a borehole or well depends to a great extent on how and where it is constructed.

Prospecting for groundwater is the process that leads to an understanding of the nature and the dynamics of the water resources. This process is a series of steps, carried out using various tools, enabling the prospector to optimise the borehole or well and to estimate the useful life of the ground-water resources.

1 Questions facing the prospector

The principal questions to be considered by the “ACF prospector” are categorised below:

Aquifer geometry

- Estimation of the depths and surface area of the water-bearing zones.
- Determination of the capping-layer quality (clayey or permeable).
- Definition of the unconfined water static level.

Water storage

- Estimation of the quantity of water stored (drainage porosity or storage coefficient).
- Estimation of the water quality.

Table 5.I: General prospecting procedure.

Phases	Objectives	Principal tools
Preliminary study	To accept or reject groundwater as a supply for the population To determine the technical solutions needed to satisfy the local water requirements	Analysis of available data Maps Satellite images Field study
Construction study	To confirm the utility of the selected aquifer To decide on the precise siting of the borehole or well	Photo-interpretation Geology and geomorphology of the study area Geophysics Piezometry Hydrochemistry
<i>Construction of wells or boreholes</i>		
Evaluation of the resource	To simulate the behaviour of the borehole over time To estimate the quantity of water in the aquifer	Test pumping Piezometry Water balance Hydrodynamic modelling
Perfecting procedures	To evaluate the initial hypotheses in the light of results obtained	Use of all the tools shown to be useful

Water flow characteristics

- Estimation of the permeability coefficient.
- Estimation of the transmissivity and of a theoretical exploitable yield.

Resource life

- Estimation of the exploitation potential over various time scales.
- Estimation of the vulnerability of the resource in terms of its water quality.

Equipping boreholes

- Localisation of water-bearing zones before installing equipment.

The procedure for answering these questions is a series of steps or phases, which can be interrupted in the case of failure, and which can be carried out gradually (Table 5.I).

The prospecting procedure to use in a particular case depends on the hydrogeological context. It is therefore first of all necessary to identify the nature of the system in question, before being able to determine which research methods to apply and choose suitable investigation tools.

2 Preliminary actions

Whatever the geological context and the prospecting procedure envisaged, a few simple actions can be rapidly performed, enabling the selection or rejection of groundwater as a supply option. They can also help to identify which technical solutions should be adopted. The aim at this stage is to make a decision and draft a budget.

2.1 Available information

In certain urgent and complex situations, the study of the site should start with the collection, compilation and organisation of existing information. Such information may be available locally, or from national institutions, other NGOs and universities, and also at the international level, from the UN, universities, geological research centres etc.

The principal questions which can be answered in this way are summarised in Table 5.II.

Table 5.II: Main preliminary questions.

Type of geological formation	Alluvial, sedimentary basin, bedrock zone etc.
Climatic data	Average monthly rainfall and temperatures
Groundwater resource	Average depth, yield of existing installations
Possible technical solutions	Wells or boreholes Construction techniques Average depth and diameter Routine methods of siting and construction
Cost/resources	Technical Human Logistic

2.2 Cartography

Topographical and geological maps indicate the main characteristics of the zone under study at the regional level, even if their scales are often too small to give a highly detailed picture.

Figure 5.1 shows an extract from a geological map of the Choluteca region of Honduras (from *Mapa Geologico de Honduras*, M.J. Kozuch, 1991) where ACF was planning a borehole-drilling programme. Despite its small scale (1:500 000) this map reveals:

- contact between non-consolidated sediments (quaternary) and consolidated volcanic rocks (tertiary);
- the principal directions of lineaments (N45° and N330°);
- the relief;
- the hydrographic network (on the topographic map of the region, the linear course of the river Choluteca also indicates a consolidated and fractured formation).

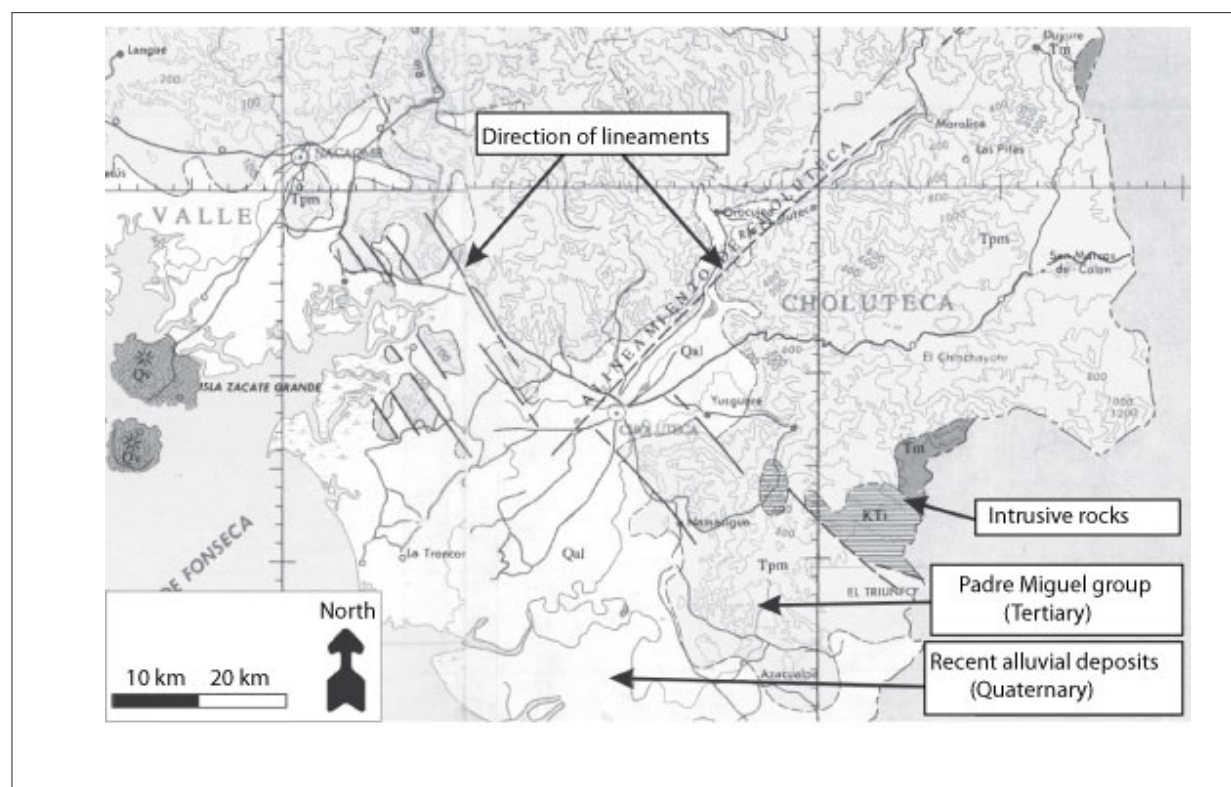


Figure 5.1: Geological map of southern Honduras.

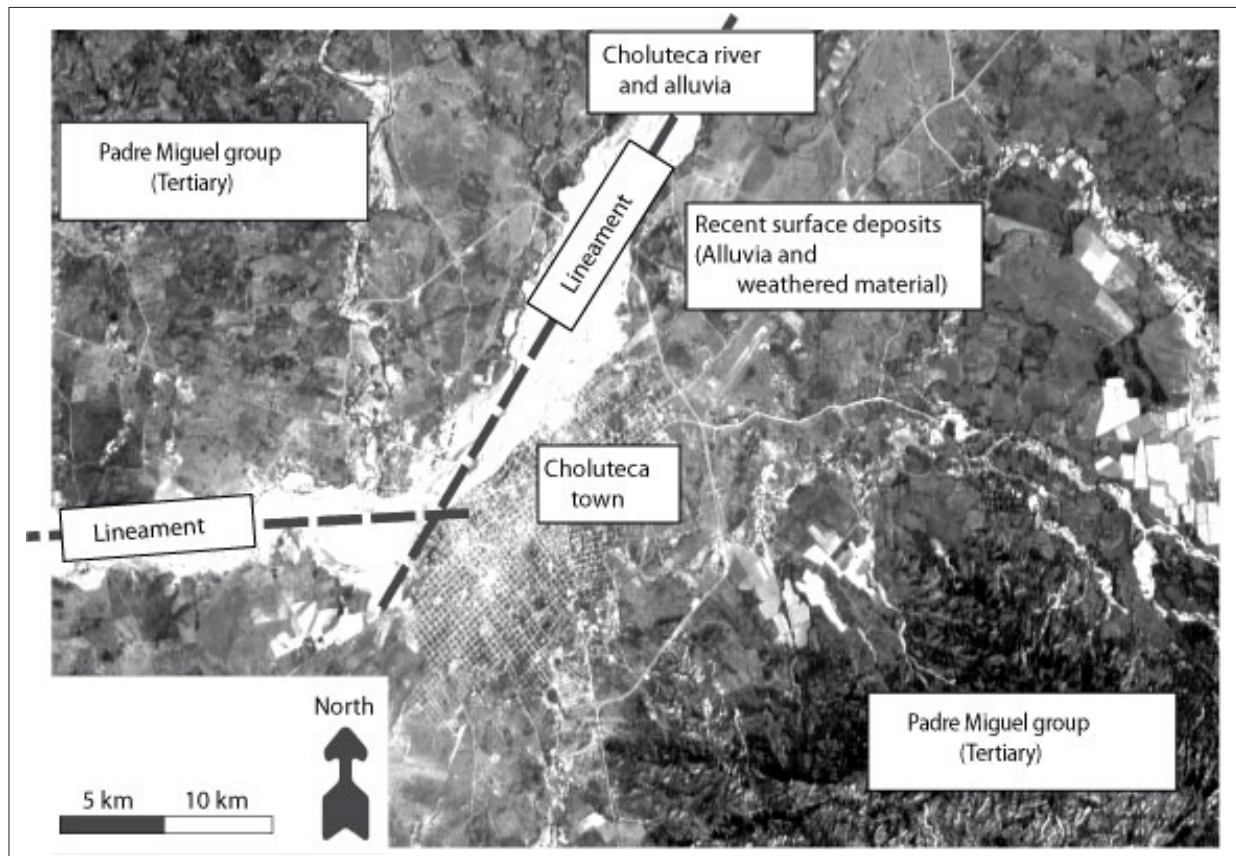


Figure 5.2: Extract from a SPOT (KJ 615-323) image (30/11/1998) centred on the town of Choluteca, Honduras.

2.3 Satellite images

The analysis of satellite images is generally concerned with questions about vegetation, structural parameters and geomorphology. Figure 5.2 shows an extract of a SPOT image of the town of Choluteca in Honduras. The information it contains complements that obtained from the geological map (Figure 5.1):

- contrasts in the plant cover indicate different geological/pedological areas, humid zones and dryer zones (alluvia and recent capping, volcanic rocks in the Padre Miguel group);
- lineaments several kilometres long are clearly visible running across the hydrographic network (they are also shown on the topographic maps);
- the drainage networks (concentration of runoff water) and the zones of runoff (slopes) are clearly visible in the topography.

2.4 Photo-interpretation

The interpretation of aerial photos is based on the analysis of forms of relief which are identified directly, by stereoscopic imagery, or indirectly, by studying the hydrographic network and variations of the contrast levels in the photograph.

– *In a bedrock zone*, the most useful information is that concerning fracturing and weathering, which are signs of the probable presence of water.

- The hydrographic network is often superimposed on the fracture network: stereoscopic images of this network can thus yield structural information. Generally speaking, a dense, many-branched hydrographic network is the consequence of impermeable terrain. In a basement zone, this gene-

Figure 5.3: Aerial photo, Marcelinot Champagnat, region of Choluteca, Honduras (ACF 2000).

rally means that the erosion products are plentiful and clayey. On the other hand, a linear hydrographic pattern indicates a fractured zone, which may have suffered very little weathering.

- The lineaments are not always aligned with by the hydrographic network: they can also be found in terrain running across the relief or vegetation.

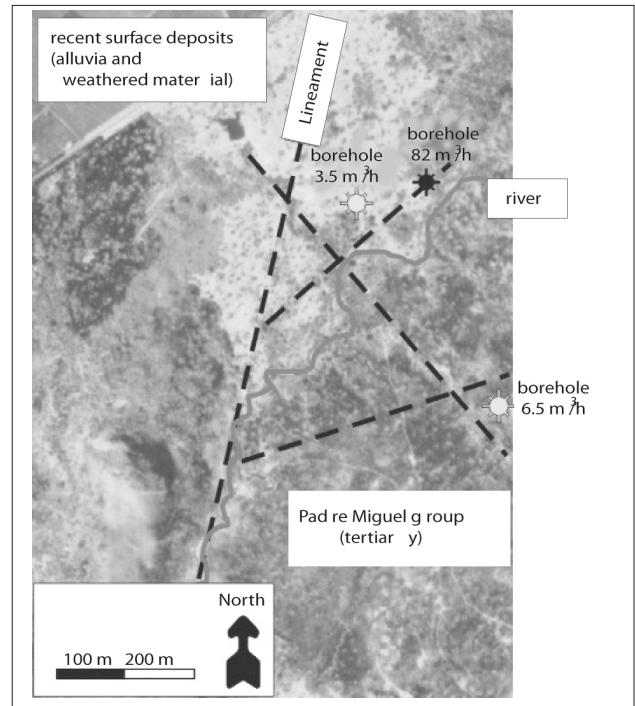
- The lineaments identified must be studied in the light of their direction: the best choice is those directions where the fractures are known to be open, or sometimes the intersection points of two lineaments (nodes).

- In bedrock areas, aerial photos can also be used to distinguish preferential infiltration zones (lowlands, *oued*) from draining zones (hydrographic network).

– *In non-consolidated, sedimentary zones*, the information required is different. The key points to identify are:

- the geological units, indicated by the vegetation and the relief;
- floodable and well-drained areas;
- the hydrographic network, whose previous pathways (paleochannels) are sometimes clearly visible: these are favourable sites for siting wells and boreholes.

Figure 5.3 shows a structural analysis performed using aerial photos of Marcelinot Champagnat Camp (in the north-east of the satellite image in Figure 5.2). The geological context is that of tertiary rhyolitic ignimbrites (Padre Miguel group) capped in the northern part of the photo by quaternary alluvia and erosion products (colluvia). The N45° lineament corresponds to a regional principal fracture direction marked on the geological map (Figure 5.1) and on the satellite image (Figure 5.2). The borehole which was drilled on this structure had an air-lift output of 82 m³/h for a total depth of 32 metres, whereas the previous boreholes drilled outside the fracture zones had much lower outputs for greater depths (6.5 and 3.5 m³/h for 55 and 96 metres depth respectively).



3 Field survey

All the analyses carried out in the office must be continued in the field. In practice, the process is iterative between data collection in the field and data analysis in the office.

3.1 The preliminary visit

It is essential to visit the area and meet the authorities (traditional and/or administrative) and the local population. After introductions, the information needed to understand the situation is sought: this is done by observation, listening, questioning without suggesting the answers, reformulating questions etc. A range of factors need to be understood, including human and technical elements (see Chapter 2):

- the climatic context: rainfall and seasonal patterns of rainfall, temperature, wind etc.;

- the geological context: outcrops, tilts, directions of identified fractures, profiles observed in wells etc.;
- the geomorphological context: relief, hydrographic network, potential infiltration zones, floodable zones;
- the hydrogeological context: state of the existing water resources: cleanliness, durability, measurement of static and dynamic levels, and estimation of outputs, localisation of positive *and* negative wells/boreholes, water quality (conductivity measurements), ‘special’ observations such as the alignment of termite mounds in some direction, the association of certain trees with existing water points, the positions of the most productive water points in relation to the geomorphology, the hydrographic network etc.

The results of this preliminary visit can be summarised graphically on maps or digital satellite images (e.g. using a GIS programme).

3.2 Technical meetings

At this stage in the investigations, meetings should be held with businesses, regional and national technical departments and any other relevant organisations, in order to complement the information gathered during the preliminary visit: borehole reports, pumping tests, hydrochemistry etc.

Dialogue with the local population remains essential: the inhabitants of the zone have a certain amount of knowledge and are aware of what has already been tried and whether or not it worked.

3.3 Additional investigations

Once the potential aquifers have been identified, the next step is to determine in the field the exact sites at which to construct the wells or boreholes.

Existing structures must be visited systematically, in order to measure the total depths, the static and dynamic levels and the production (output:drawdown ratio). The measurement of the water’s conductivity at each site, and that of surface water, can indicate the degree of linkage between the groundwater zones (is there one large homogeneous aquifer, or several isolated systems?). The conductivity can also reveal any problem of excessive mineralisation of the water, or of salinity.

The topography and the plant cover can be used to identify zones of infiltration and run-off.

In bedrock zones, the sites selected by inspection of aerial photos should be visited, to verify the presence of lineaments. An alignment identified on a satellite image or photo may turn out to be a track or a trench, and thus without any particular water-bearing potential. In the field, all the signs which might confirm the presence of an anomaly should be observed: alignments of termite mounds or large trees, changes in vegetation, outcrops etc.

In alluvial zones, the groundwater-river interactions (supply/drainage/silting, should be estimated (see Chapter 3).

4 Hydro-geophysical methods

Geophysical methods can be applied to hydrogeology (hydro-geophysical methods). They may be useful at some stage in the hydrogeological procedure, to answer questions concerning the local geology. Such questions are specific to the context of the study, but they can be grouped as a function of aquifer typology and classed in two categories: the first concerns the aquifer geometry; the second concerns the parameters describing the storage and flow characteristics. Where the salinity of the groundwater may be an issue, the prospector may also wish to measure electrical conductivity.

From 2000 to 2003, ACF ran a research and development programme to appraise the aptitude of geophysical methods to characterise aquifers in the principal geological types (non-consolidated

formations, bedrock zones and carbonated rocks). This work included, notably, a study of the new “MRS” method, whose principal results are presented in this chapter (Vouillamoz, 2003).

4.1 Introduction to the methods

Notation used

Direct Current method	DC method
Resistivity Profiling	RP
Vertical Electrical Sounding	VES
Electrical Resistivity Imagery	ERI
Spontaneous Potential	SP
Electromagnetic method	EM method
Very Low Frequency	VLF
Time Domain Electromagnetic	TDEM
Magnetic Resonance Sounding	MRS
Borehole Electrical Logging	BEL
Transmitter	Tx
Receiver	Rx

4.1.1 THE PRINCIPLE

Hydro-geophysical methods measure the spatial and temporal variations in the physical properties of underground rocks. The physical properties studied are influenced by the nature of the reservoir, the volume of empty space it encloses (its porosity) and the volume, the degree of saturation and quality of the water it contains.

– With all the traditional geophysical methods, the groundwater affects certain measured parameters, but it is never the only influencing factor. The physical quantities recorded by the geophysicist do not enable a direct determination of the presence of groundwater or its quality, but in favourable cases they help to consider the nature and the structure of any aquifers present.

– In comparison with traditional methods, Magnetic Resonance Sounding (MRS) can be classed as a *direct* geophysical method, because it measures a signal emitted by atomic nuclei present in each water molecule. The contribution of MRS to hydrogeology is thus the ability to measure directly a signal indicating the existence of groundwater.

Among the numerous geophysical methods used in hydrogeology, three have been selected by ACF on the basis of the particular requirements of humanitarian programmes:

- electrical resistivity measurement is the reference method, because it can be used in a wide range of contexts and can also give 2-dimensional data;
- electromagnetic methods, notably the VLF method, the Slingram method and TDEM soundings, are easier to use than electrical resistivity measurements but are less versatile;
- MRS is rather difficult to use, but it is the only method that gives direct information on the existence of groundwater.

Finally, it should be noted that whatever the geophysical method used, the validity of the study depends on the quality of the measurements performed in the field, and on the number and variety of other observations and complementary analyses performed.

4.1.2 THE CHOICE OF METHODS

The choice of geophysical methods depends on the geological context in question and the information required. The method, or the combination of methods used must form part of an overall prospecting procedure, the determination of which is presented in Section 5. The principal criteria of choice are the following.

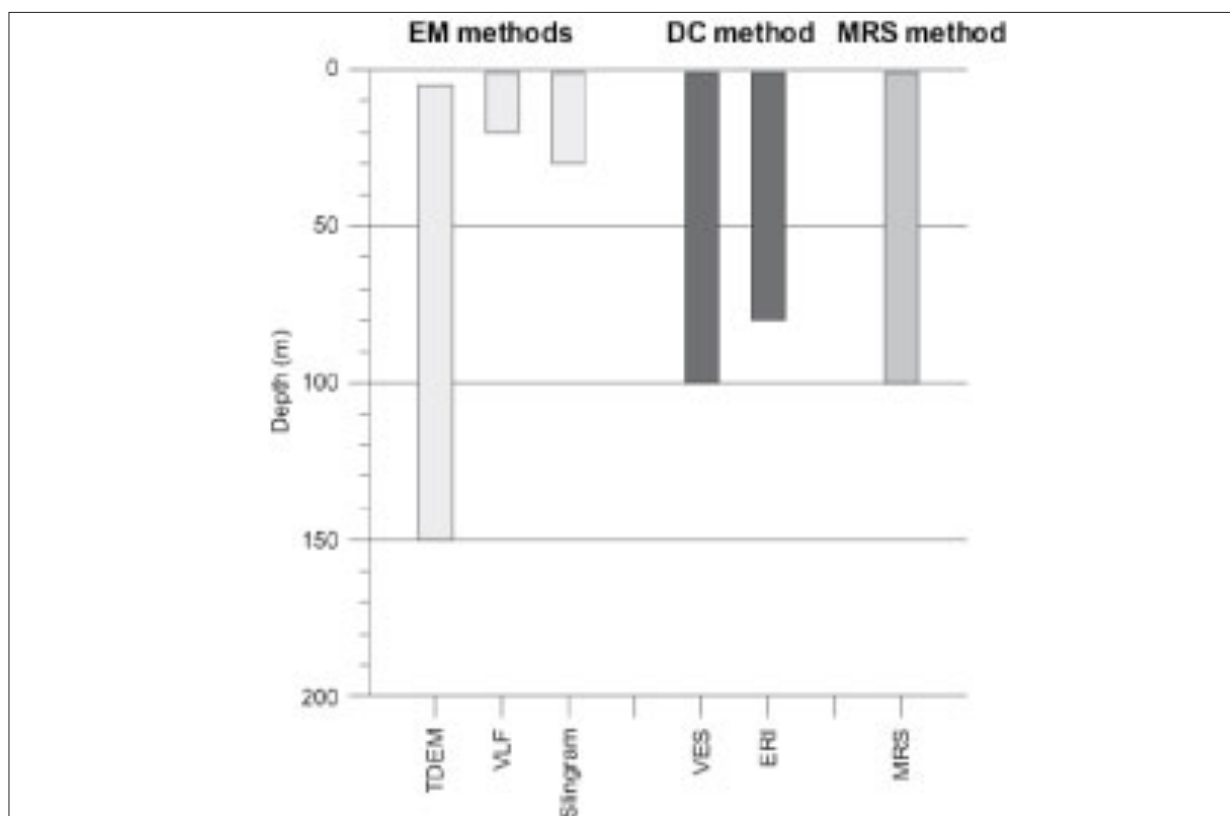


Figure 5.4: Typical sounding depths of the methods and equipment used by ACF.

The nature of the water source being sought

It must have a sufficient effect on the quantity measured in the experiment, which will itself depend on the physical properties of the probed volume (Table 5.III). For example, alluvial ground-water tends to generate proton magnetic relaxation signals that can be detected by MRS. Also, salinity variations in groundwater can cause variations in electrical resistivity that can be detected in TDEM soundings. Table 5.III shows clearly that only MRS measures geophysical parameters directly linked to groundwater.

The required precision

This cannot exceed the resolution of the method, or more specifically, that of the equipment to be used. The resolution determines the capacity of a method to detect and characterise a water reserve. It is a function of the sensitivity of the apparatus, but it is also limited by the sensitivity of the equipment used and by the conditions of measurement (signal-to-noise ratio). It is thus not possible to quote a standard resolution for each method. Instead, a typical range of depth of resolution for each hydrological method is usually given (Figure 5.4).

The scale of the study

This determines a framework for the practical application of the method, which requires the deployment of equipment in the field. Each type of equipment has its particular modes of operation and can be deployed in various configurations, depending on the type of information sought. Table 5.IV summarises all the various possibilities and presents the principal applications of the main apparatuses used by ACF.

Cost

This is best considered as the impact of the geophysical method used on the total cost of the project. Two principal factors must thus be evaluated: on one hand, the cost of carrying out the measurements, and on the other, the savings their use generates, through the reduction in the number

Table 5.III: Principal geophysical methods used in hydrogeology (the properties whose effects are secondary are in brackets) (from Kearey & Brooks 1984, modified).

Method	Measured geophysical parameter	Operational physical property	Influence of groundwater
Electrical	Potential difference due to electric currents	Electrical resistivity	Indirect
Electromagnetic	Electromagnetic signals due to induction	Electrical conductivity (magnetic susceptibility and dielectric permittivity)	Indirect
MRS	Proton magnetic relaxation signal in water	Spin and magnetic moment of the hydrogen nucleus	Direct

Table 5.IV: Typical geophysical methods and domains of hydrogeological application.

Method	Array	Application	Domain of use
DC METHODS	RP (single and double line)	Preliminary study: resistivity profile qualitative 1D interpretation	All formation types, but possible difficulties implanting electrodes and assuring good electrode-earth contact Sounding depth limited if surface is highly conducting (clay, saltwater)
	VES (Schlumberger, Wenner, pole-dipole; pole-pole)	Complementary study: log of resistivity quantitative 1D interpretation	
	Multidirectional soundings	Complementary study: directional log of resistivity qualitative 2-3D interpretation	
	ERI (Wenner, pole-pole, pole-dipole, dipole-dipole)	Complementary study: section of resistivity quantitative 2D interpretation	
EM METHOD	VLF (pole-dipole, dipole-dipole)	Preliminary study: profile and map of iso-resistivity qualitative 1-2D interpretation	All except highly resistive formation types Sounding depth limited if surface is highly conducting for VLF and Slingram
	Slingram (multi-frequency)	Preliminary study: profile and map of iso-conductivity qualitative 1-2D interpretation	
	TDEM sounding	Preliminary & complementary study: log of resistivity quantitative 1D interpretation	
MRS	1D sounding	Complementary study: qualitative 1D interpretation or quantitative if calibrated (log of storativity and hydraulic conductivity)	Continuous formations Impossible in the presence of magnetic rocks

of negative wells and boreholes. Knowing the cost of the geophysical studies and of that of a negative borehole calculated on the same basis, one can define the economic domains of use of the various methods, or combinations of methods. The relation can be written:

$$r \cdot bh = r \cdot bh^+ + (1 - r) \cdot bh^- + \rho$$

$$bh = bh^+ \frac{1 - r}{r} \cdot bh^- + \frac{\rho}{r}$$

where r is the borehole success rate (%), bh the average cost of an exploitation borehole, bh^+ the average cost of a positive borehole, bh^- the average cost of a negative borehole and ρ the average cost of geophysical studies per borehole. Systematic geophysical studies will save money in the programme if:

$$\rho \leq bh^- \cdot \left[\frac{r_p}{r} - 1 \right]$$

where r is the borehole success rate without geophysical studies, and r_p the borehole success rate with geophysical studies. These rates can be estimated from the experience of local people or from previous ACF programmes.

However, the cost evaluation of programmes alone is not sufficient to decide what actions should be undertaken in an ACF programme. The calculation does not take into account the fact that geophysical studies often enable the construction of successful boreholes in difficult zones in which the population's water requirements are substantial. ACF's objective is to respond to the needs of vulnerable populations, even where this may not seem 'economically' justified.

4.1.3 GEOPHYSICAL PROCEDURES

Using the criteria of choice of methods, and with ACF's experience, standard geophysical procedures can be constructed. However, these procedures must form part of a hydrogeological investigation strategy, whose conception is explained in Section 5. These standard procedures are thus suggestions, which can be modified (Table 5.V). Examples of real cases are given in Chapter 5B.

4.2 The Electrical Resistivity method

To improve readability, the scientific references used in the following Section are not noted, but they can be found in the reference list.

4.2.1 THE PRINCIPLE

The principle is to feed a direct current flow into the ground, and use it to measure the apparent resistivity of the formation. The nature and structure of the aquifers are then deduced on the basis of the variations (contrasts) in the calculated resistivity.

The electrical resistivity of a medium is the physical property that determines its capacity to oppose the passage of electric current. In rocks, the flow of current by electron movement is rare ('electronic' or 'metallic' conductivity in certain mineral seams) and the charge transport is essentially due to ions moving in solution (electrolytic conductivity). Thus, the resistivity of the rocks depends essentially on:

- the nature and the weathering of the rock (electrolyte distribution in the ground);
- the water concentration (saturation of the rock with electrolyte);
- the water quality (mineralisation of the electrolyte);
- the temperature (electrolyte viscosity and ion mobility).

The ranges of real resistivity values found generally in the field are presented in Table 5.VI.

Table 5.V: Standard geophysical procedures.

FORMATION	Continuous (sedimentary non-consolidated)	EM Resistivity mapping VLF or Slingram	Delimitation of hydrogeological domains Characterisation of the capping layer (clay content)
		VES if formation is resistive TDEM if formation is conducting	1D description of the structures Estimation of the water mineralisation
		ERI Estimation of the water mineralisation MRS	2D description of the structures
			Confirmation of the presence of groundwater Characterisation of the storativity and transmissivity of the aquifers
		BEL	Localisation of productive zones Choice of borehole drilling equipment
	Discontinuous (basement)	Resistivity Slingram or RP profile	Estimate of the thickness and the nature of weathering
		VES (occasionally TDEM)	1D description of the structures
		ERI MRS	2D description of the structures Confirmation of the presence of groundwater Characterisation of the storativity and transmissivity of the aquifers
		BEL	Localisation of productive zones Choice of the borehole drilling equipment
	Karstic	EM Resistivity mapping VLF or Slingram	Revealing shallow fractures
		ERI	2D estimation of the structures
		MRS	Localisation of saturated zones (epikarst and shallow dissolution patterns)

Table 5.VI: Real resistivities generally encountered in various formations.

Formations	Resistivity of saturated formations* (Ωm)	Resistivity** (Ωm)
Clays	5 to 10	0.5 - 5
Sands	50 to 400	40 to 300
Gravels	150 to 500	200 to 500
Crystalline shales	100 to 10 000	
Solid gneiss		1 000 to 10 000
Weathered dry gneiss		300 to 600
Weathered wet gneiss		120 - 200
Solid granites	100 to 50 000	1 000 to 10 000
Weathered dry granite		300 to 1 000
Weathered wet granite		100 - 300
Chalk	100 to 10 000	50 - 300
Seawater	< 0.2	
Saline groundwater	< 1	
Fresh surface-water	0 to 300	
Limit of potability	2 to 6	

* According to Meyer 1991. ** ACF experiments.

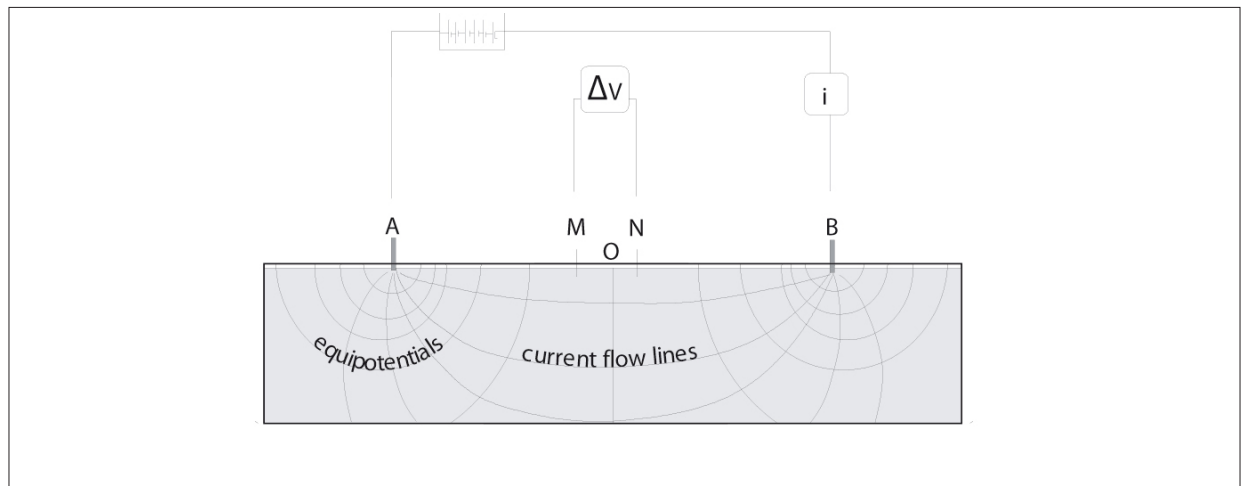


Figure 5.5: Principles of geophysical electrical measurements.

4.2.2 MEASUREMENTS

To measure resistivity, a direct current is fed into the ground via two electrodes A and B, and the potential difference (ΔV) generated is measured between two electrodes M and N (Figure 5.5).

The electrical resistivity of the formation through which the current passes is calculated using the formula $\rho_a = K \cdot \Delta V_{MN} / I_{AB}$ with ρ_a the apparent resistivity in ohm-metres (Ωm), ΔV_{MN} the potential difference in volts (V), I_{AB} the direct current in amperes (A) and K a geometric factor such that:

$$K = \frac{2\pi}{1/AM - 1/BM - 1/AN + 1/BN}$$

The resistivity ρ_a is called the “apparent resistivity”, because it corresponds to the resistivity of the whole set of formation types through which the current flows, which may be different from the real resistivities of each formation type. These are calculated later, from the apparent resistivities, during the data analysis phase of the work.

The penetration depth of the current-flux lines (and consequently the depth probed) is a function of the separation of electrodes A and B.

There are four types of measurement, each associated with a particular area of investigation:

- Vertical Electric Sounding (VES) explores the layers of ground lying between electrodes M and N. It can therefore measure the resistivity as a function of depth, by progressively increasing the distance AB (one-dimensional sounding);
- Resistivity Profiling (RP) identifies the horizontal development (at constant depth) of a set of formations along a profile. It is therefore possible to measure variations in thickness of a formation along a profile. The thickness of formations being tested is given by length AB, which is maintained constant throughout the profile;
- Electrical Resistivity Imagery (ERI) gives a cross-section of resistivity measured and interpreted in two dimensions. It therefore collates information given by the sounding (vertical variation) and by the traverse (horizontal variation);
- Square Sounding reveals anisotropy. A coefficient and direction of anisotropy can thus be calculated, as well as average resistivity.

A number of electrode arrays are commonly used:

- the *Schlumberger array*, with four electrodes placed in line such that $AB > 5 MN$ (Figure 5.6), is the most frequently used for VES. It offers the advantage over the Wenner method of having to move only two electrodes (A and B) and offering a good depth of penetration;

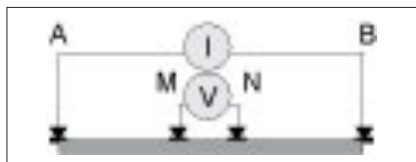


Figure 5.6: Schlumberger array.

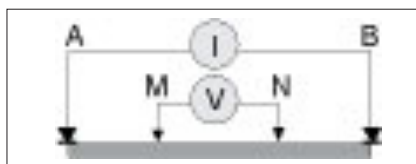


Figure 5.7: Wenner array.

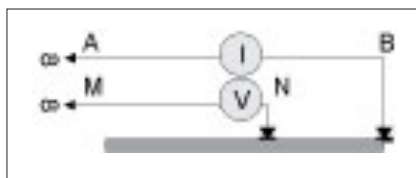


Figure 5.8: Pole-pole array.

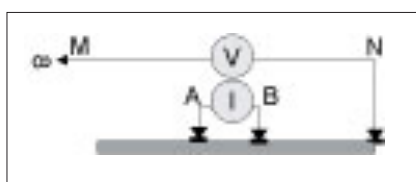


Figure 5.9: Half-Schlumberger array.

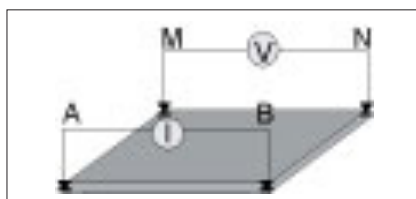


Figure 5.10: Square array.

- the Wenner array, with four in-line electrodes such that $AM = MN = NB$ (Figure 5.7), is widely used for electrical resistivity imagery (symmetrical array);
- the pole-pole array consists of placing two electrodes (M and A) at ‘infinity’ and only moving B and N (Figure 5.8). It offers a high signal-to-noise ratio and thus allows considerable investigation depths to be reached, and is more sensitive to vertical anomalies than the quadrupole array;
- the pole-dipole or half-Schlumberger array can be used in VES and ERI. By doing two ‘inverted’ measurements, it is possible rapidly to obtain information about the heterogeneity of the formation (Figure 5.9);
- the square array, with $AB = MN = AM = BN$ (Figure 5.10), is mainly used to perform soundings. It allows anisotropy to be estimated, notably directions of fracture, but it is difficult to carry out in the field.

For all these configurations, exchanging the positions of the current electrodes A & B with those of the potential electrodes M & N does not in theory affect the results. A large spacing between the potential electrodes gives a strong measured signal (ΔV): this is an advantage in environments without much noise, in which the signal-to-noise ratio will thus be enhanced, but it is a disadvantage in noisy zones (stray currents, telluric currents) in which noise will also be increased.

4.2.3 THE METHOD IN PRACTICE

4.2.3.1 Equipment and personnel

ACF has developed a device suited to the particular conditions in which it operates: the Ω mega resistivity-meter (Figure 5.13). Instructions for use are given in Annex 8A. The operating method

Box 5.1 Electrical geophysics.

Resistivity

The current across a conductor is equal to the voltage, divided by a constant – the resistance. Ohm's Law is expressed as:

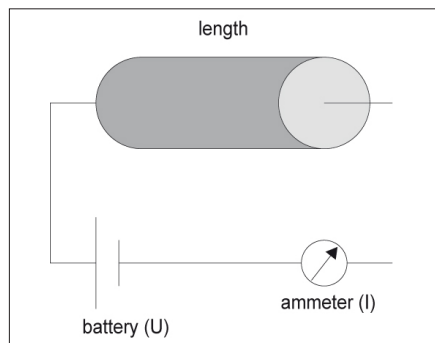
$$U = RI$$

where U is voltage (volts), R resistance (ohms), and I current (amperes). This law is only strictly valid for metallic conduction, but it is an acceptable approximation for electrolytic conduction.

The relation between resistance R and resistivity is:

$$R = \rho \cdot \frac{L}{S}$$

where R is resistance (Ω), ρ resistivity ($\Omega \cdot m$), L conductor length (m) and S conductor cross-sectional area (m^2) (Figure 5.11).



Most rocks are isotropic – they have the same resistivity in all directions. However some, e.g. metamorphic rocks, have oriented structures, and are sufficiently anisotropic for this simplification not to be valid.

< Figure 5.11: Resistivity, function of the conductor geometry.

Geometric coefficient K

A current I passing through an electrode located in an infinite and isotropic space (Figure 5.12) creates a potential V at point M, such that:

$$V_M = \rho \cdot \frac{I}{4\pi r}$$

< Figure 5.12: Infinite and semi-infinite spaces.

In the semi-infinite space of a hemisphere defined by the ground surface:

$$V_M = \rho \cdot \frac{I}{2\pi r}$$

If a current I passes through two electrodes A and B, it is possible to measure the potential difference between two other electrodes M and N due to the joint action of A and B as:

$$V_M = V_A + V_B = \frac{\rho I}{2\pi AM} - \frac{\rho I}{2\pi BM} = \frac{\rho I}{2\pi} \left(\frac{1}{AM} - \frac{1}{BM} \right)$$

$$V_N = V_A + V_B = \frac{\rho I}{2\pi} \left(\frac{1}{AN} - \frac{1}{BN} \right)$$

$$\Delta V_{MN} = V_M - V_N = \frac{\rho I}{2\pi} \left(\frac{1}{AM} - \frac{1}{BM} - \frac{1}{AN} + \frac{1}{BN} \right)$$

The resistivity is therefore obtained by:

$$\rho = \frac{\Delta V_{MN}}{I} K$$

where K is a geometric coefficient:

$$K = \frac{2\pi}{1/AM - 1/BM - 1/AN + 1/BN}$$



Figure 5.13: Ω mega-resistivity-meter (ACF, Cambodia, 1998).

consists of taking successive measurements for different electrode spacings. The measurement files given in Annex 8B show values of current, ΔV , and apparent resistivity (R_{oa}) calculated for every measurement point in the field. Values of the geometric coefficient are pre-calculated for standard proposed measurement steps. It is nevertheless possible to modify them in the light of experience, and to prepare a sheet of values adapted to local conditions. It is important to calculate the apparent resistivity values and to draw curves directly in the field to check the consistency of the measurement. This equipment is used to perform one-dimensional measurements (soundings and profiles).

It is of course possible to perform measurements in the field with other equipment. In particular, ACF has successfully used Syscal resistivity-meters and Iris Instruments multi-electrode systems (Figure 5.14). With this equipment, the measurements can be automated, the values of I , ΔV and R_{oa} are recorded in an internal memory and the quality of the measurements is improved considerably (higher voltmeter sensitivity, advanced spontaneous potential (SP) correction, measurement dispersion analysis etc.). In view of these possibilities, this equipment is particularly well suited to two-dimensional measurements (ERI).



Figure 5.14: Syscal R1 resistivity-meter (ACF, Mozambique, 2000).

A compass is needed to determine the measurement direction, as well as two 100-m measuring tapes strong enough to be used to measure the spacing of electrodes.

To carry out a sounding or a profiling, about twenty electrodes allow part of the array to be put in place before taking measurements. This helps to avoid induced currents created when the electrodes are set up. When setting up ERI, the number of electrodes required depends on the length of the section to be measured and the capacity of the equipment: 64, 72 or 96 electrodes are generally necessary.

A team of five people can perform between 2 and 10 measurement sessions per day, depending on local constraints: when safety requires short days on site, it is difficult to carry out more than two sets of measurements per day. An engineer must define the parameters of the measurement and interpret the data. Four operators are put in charge of setting up the electrodes and spacing the coils.

4.2.3.2 Constraints

Studies in the field are limited by the extent of the exploration zone: it is therefore essential to choose a potentially favourable zone rather than to attempt to cover a very large sector. Similarly, it is necessary to try to carry out studies in a relatively flat area, where vegetation and constructions do not prevent the placing of sufficiently long A-B lines (300 to 800 m). Flooded zones and periods of heavy rain should also be avoided, because the electric measurements may be disrupted in such conditions.

The nature and thickness of capping layers are sometimes limiting factors in the use of this technique. Conducting surface formations require significant injection powers to permit the current to penetrate very deeply into the ground. Thus, resistivities of 5 to 10 Ωm allow survey depths of 30 to 50 m with the Ω mega-resistivity-meter in Schlumberger surveys (ACF Cambodia, 1998).

Conversely, in some unconsolidated sands it is necessary to wet the contact pegs in order to improve contact and facilitate passage of the current.

4.2.3.3 Vertical Electrical Sounding (VES)

Implementation

Schlumberger soundings are carried out by progressively separating electrodes A and B, without changing M and N: this increases the volume of ground through which the majority of the current passes, thereby increasing the depth of investigation (Figure 5.15).

Increasing the investigation depth while maintaining the same injection power causes the measured ΔV value to decrease progressively. To maintain it at an acceptable level (typically between 1 and 5 mV), the injection power is increased. When the power of the resistivity-meter is no longer sufficient to obtain adequate ΔV values, overlapping measurements are carried out. This consists of increasing the separation of electrodes M and N, and measuring ΔV over a greater volume of earth. The values obtained can be compared with those recorded with the initial spacing. It is essential to carry out the overlap on a minimum of two measurement points in order to check for accuracy.

The example shown in Figure 5.16 corresponds to the survey performed in a gneiss rock area in Sudan. Two overlaps were carried out:

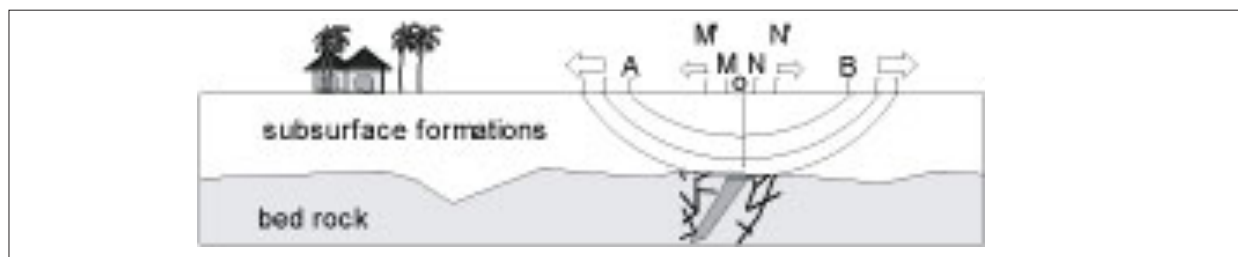


Figure 5.15: Implementation of Schlumberger VES.

- the first, at $\Delta V = 26$ mV for $MN = 2$ m, and $\Delta V = 266$ mV for $MN = 10$ m. This overlap was carried out at three measurement points ($AB/2 = 15, 20$ and 25 m);
- the second at $\Delta V = 5.6$ mV for $MN = 10$ m and $\Delta V = 24.6$ mV for $MN = 40$ m. This overlap was carried out at two points.

The two overlaps are satisfactory, in that the apparent resistivity values and slopes of curves before and after the overlap are close (Figure 5.16).

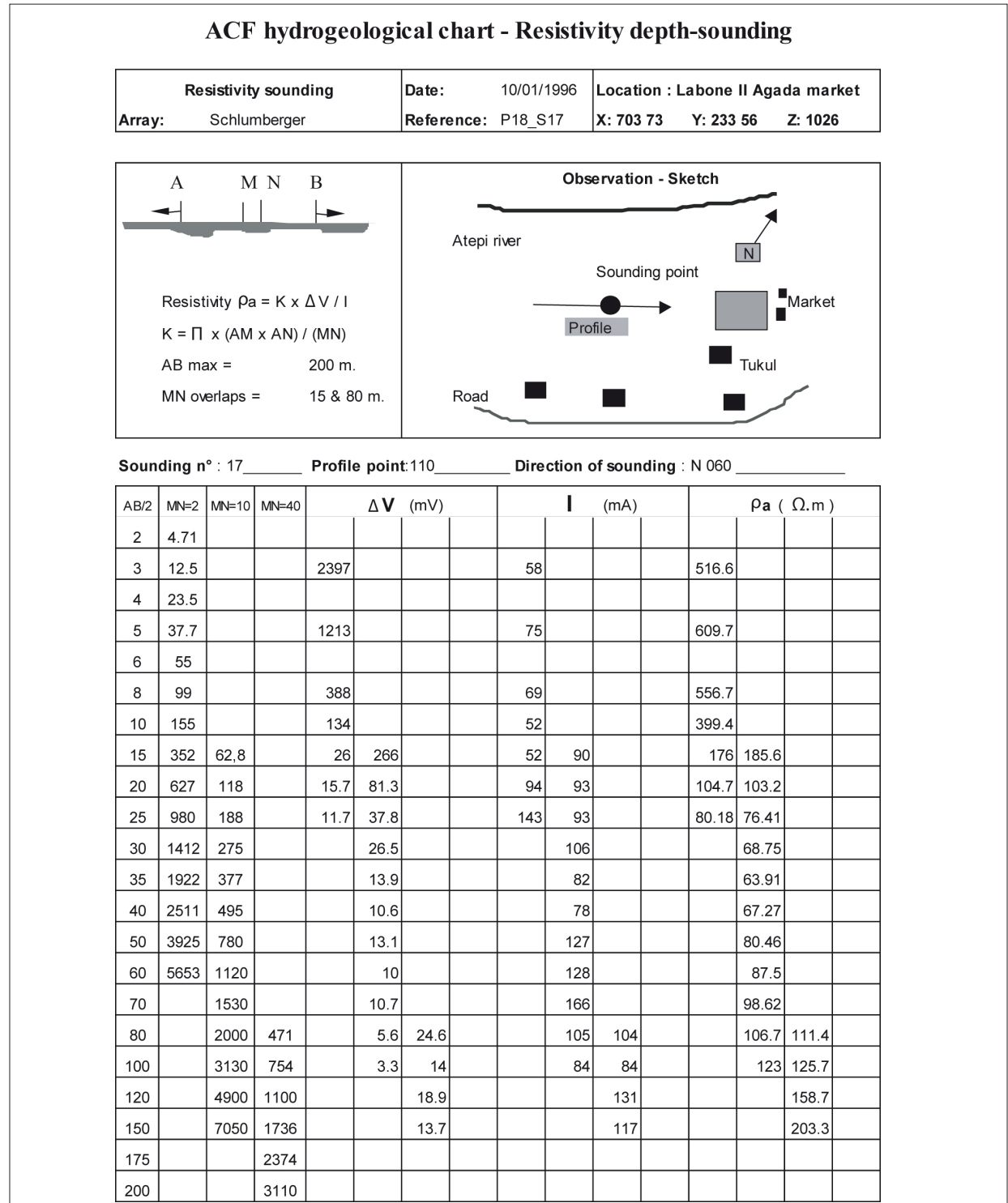


Figure 5.16: Example of Schlumberger VES in bedrock (ACF, Sudan, 1996).

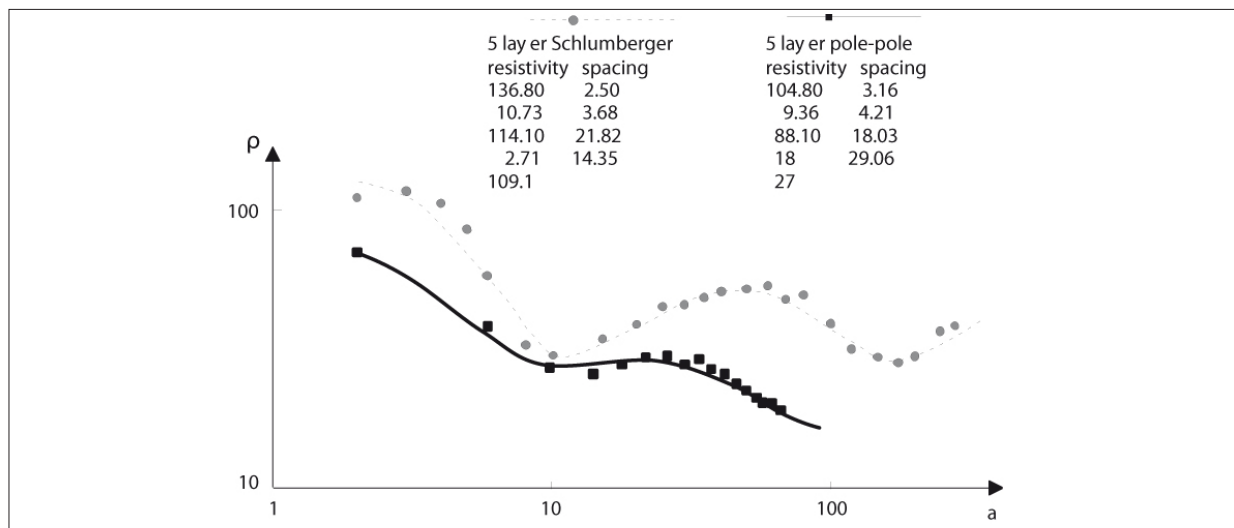


Figure 5.17: Example of a pole-pole survey (ACF, Siem Reap, Cambodia, 1998).

The *Wenner* array avoids the need for overlaps because between each measurement the distance MN is progressively increased along with AB.

The *pole-pole* array is rarely used in Vertical Electric Sounding, because its resolution is not as good as that of quadrupole arrays (Schlumberger or Wenner). Additionally, it requires the electrodes to be located at ‘infinity’ (A and M), which means a distance of at least 20 times the largest B-N spacing which is not always easy in the field. The investigation depth is given by the spacing (a) of mobile electrodes B and N.

The main advantage of the pole-pole survey method lies in its rapidity (only 2 electrodes to move, no overlap) and its depth of penetration, which is greater than that of a Schlumberger survey of the same line length.

Examples of pole-pole and Schlumberger soundings taken by ACF in Cambodia are shown in Figure 5.17. The results prove to be comparable for the two configurations. It is also noticeable that, for $a = 80$ m, the investigation depth of the pole-pole method is much greater than that of the Schlumberger method for $AB/2 = 40$ m. On the other hand, contrasts are much more clearly defined by the Schlumberger method.

The *pole-dipole* array, or half-Schlumberger sounding has the advantage over the other arrays of enabling two soundings to be taken at the same location by simply reversing the position of the infinite distance electrode (Figure 5.18). In a one-dimensional medium the two soundings are identical; any difference means that the medium is heterogeneous and one-dimensional soundings are inadequate (see Box 5.2).

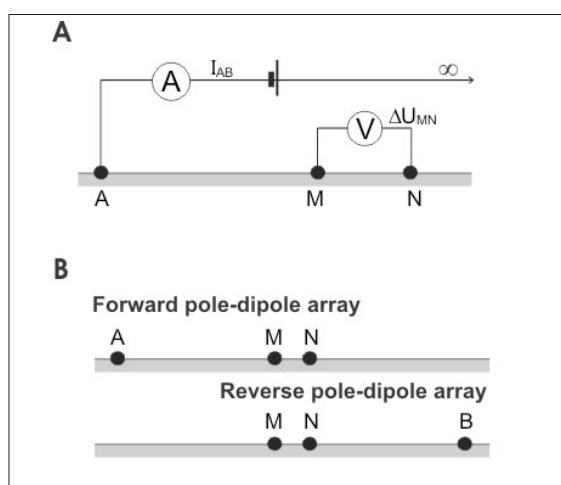


Figure 5.18: Half-Schlumberger sounding. A: principle. B: arrays.

Interpretation

The number of layers, their true resistivities, and their respective thicknesses may be estimated from the apparent resistivity measurements. There are several methods of interpretation. The simplest is the auxiliary curve or Hummel method, but interpretation software is also available (see Box 5.2).

Whatever the interpretation method, a calibration process has to be applied for every new geological zone. This means taking soundings from existing boreholes, and also on visible outcrops, to calibrate the results of the interpretations (number of layers, calculated resistivities and thicknesses) with the lithological data. Only after this calibration process can the interpretation of the geophysical measurements have any meaning.

The example shown in Figure 5.19 is from a crystalline bedrock context in Sudan. The aquifer potential is provided by an 18-m thick layer with a resistivity of 95 Ωm . Its calculated resistivity

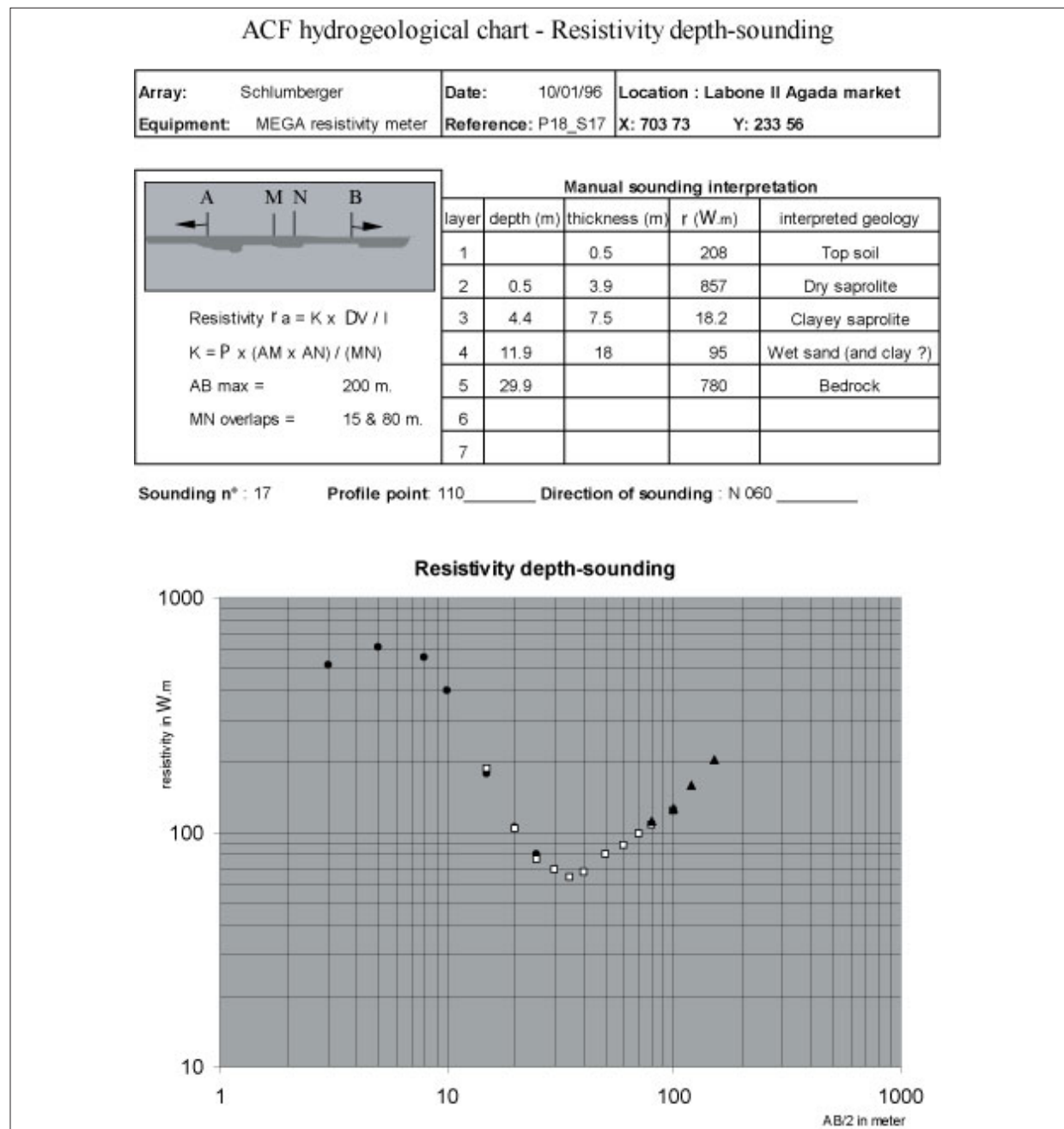


Figure 5.19: Results of VES interpretation (ACF, Sudan, 1996).

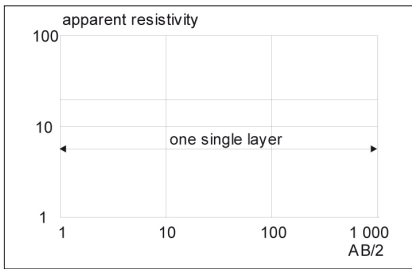
Box 5.2

Interpretation of VES.

Results are plotted on log-log paper with values of $AB/2$ (abscissa in meters) plotted against apparent resistivity values measured in the field (ordinate in $\Omega.m$). The experimental curve of the survey is obtained in this way. This curve is then compared with the theoretical curves in order to estimate the true resistivities and formation thickness.

Nomenclature

- Apparent resistivity: ρ_a
- Calculated resistivity of formation n : ρ_n
- Calculated resistivity of theoretical formation: ρ_f
- Thickness of formation n : e_n



Curves

An electrical survey carried out on a single isotropic layer gives an experimental curve (Figure 5.20) that is simply a horizontal straight line (ρ is constant as a function of depth, which is given by $AB/2$).

<

Figure 5.20: Single-layer curve.

A survey carried out on a two-layer profile gives a relatively simple curve (Figure 5.21). The first section corresponds to the first layer (curve from one layer), and the slope of the zone of influence of the two formations is given by the ratio ρ_2/ρ_1 . When $\rho_1 < \rho_2$, the curve is rising; if $\rho_1 > \rho_2$, it is falling. The end of the curve, as $AB/2$ tends towards infinity, tends to a value corresponding to the resistivity of the second formation.

A survey of three layers is also characterised by the ratio of the resistivities. The example corresponds to a curve for which $\rho_1 < \rho_2 < \rho_3$ (Figure 5.22).

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Figure 5.21: Two-layer curve.

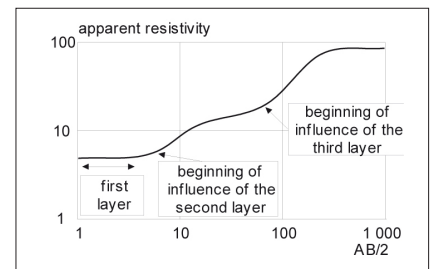
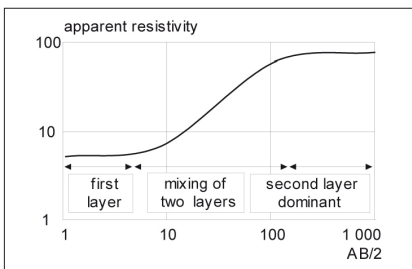


Figure 5.22: Three-layer curve.

Principles of manual interpretation, nomograms

The idea is to compare the experimental and theoretical curves replotted as nomograms. There are various types of nomogram used in the interpretation of electrical surveys. The operating methods however are very similar for all. The Cagniard nomograms, used for the Schlumberger array, are included in Annex 8 C. There are in fact two sets of nomograms: the 'two-formation' nomogram and the three auxiliary nomograms.

The two-formation nomogram is a set of theoretical curves drawn on log-log paper that indicate ρ_a as a function of $AB/2$ for different values of the ratio ρ_2/ρ_1 .

When the ground includes more than two layers, the auxiliary nomogram method is used. This consists of reducing any survey of n layers to a succession of soundings of two layers: all soundings start with the interpretation of a survey of the first two layers. These two layers are then replaced by an electrically-equivalent theoretical layer. With the third layer, this theoretical layer forms a pair of theoretical layers (principle of reduction). This then continues as an iterative process until the last layer is reached.

Operational method

The experimental curve is drawn in the same way on transparent log-log paper at the same scale as the nomograms ($AB/2$ as abscissa, ρ_a as ordinate).

The values of ρ_1 , ρ_2 and e_1 are then estimated by simple superposition of the "two-formation" nomogram and the beginning of the experimental curve, selecting the theoretical curve that best corresponds to the experimental curve, while maintaining the axes parallel. The origin of the curves in the nomogram, termed the left cross, gives as abscissa the thickness of the first formation e_1 and as ordinate the true resistivity of the first formation ρ_1 . On the theoretical curve selected, ratio ρ_2/ρ_1 enables ρ_2 to be calculated. The position of the left cross is indicated on the experimental curve.

To evaluate ρ_3 , the auxiliary nomogram is taken as a function of ratio ρ_3/r_1 . The experimental curve is superimposed on the auxiliary nomogram with its origin coinciding with that of the previously-drawn left cross. On the experimental curve is traced the curve of the auxiliary nomogram corresponding to the previous ratio ρ_2/ρ_1 ; this curve represents the geometric location of the origin of the two-formation nomogram. The two-formation nomogram is then taken while maintaining its origin on this curve and adjusting it to the form of the experimental curve. By marking a new cross as the origin of the two-layer nomogram, the value r_f of the theoretical formation (electrically equivalent to the first two formations) is obtained as ordinate. From the ratio ρ_3/ρ_f read from the selected theoretical curve, ρ_3 is obtained.

e_2 is estimated by replacing the auxiliary nomogram on the experimental curve, ensuring that its origin coincides with the origin of the first left cross. The site of the second cross of the two-formation nomogram has as abscissa the ratio e_3/e_f . The value of e_f is given as the abscissa of the first left cross position.

This process is iterated up to the end of the experimental curve in order to obtain all the true resistivity values (ρ_n = ordinate of the left cross $\times$ ratio ρ_{f_n}/ρ_f) of layers and all their thicknesses (e_n = abscissa left cross $\times$ ratio e_n/e_f), except for the last layer.

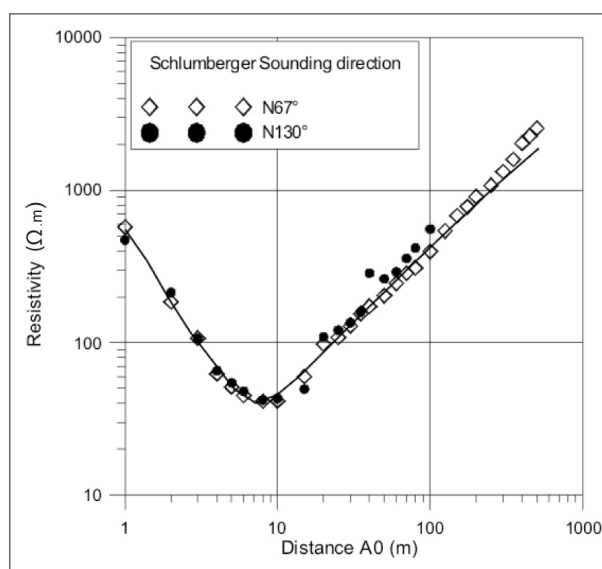
Errors of interpretation

A little practice in the use of nomograms quickly makes them easy and practical to use. One of the common mistakes made in the early stages is to try to superimpose the experimental curve perfectly on the nomogram, which leads to a multiplication of the number of formation types. In practice, it is preferable to retain a theoretical curve that includes a maximum number of points on the experimental curve, and that also follows its trend well. If this proves difficult, it is always possible to select a curve that corresponds to an intermediate ratio ρ_2/ρ_1 not actually present in the nomogram.

The solution obtained after interpretation is not unique. There are in fact several solutions, known as equivalences, corresponding to different thicknesses and resistivities of layers that give the same experimental curve (ρ/e or $\rho \cdot e$). It is not possible to choose the correct solution without extra information (thickness of a formation ascertained by borehole, resistivity determined by measurement on an outcrop etc.).

It is also possible that a formation type which is present may not be revealed by the interpretation: this phenomenon of 'suppression' sometimes occurs for formation layers that are not very thick, contained between two other formations of similar resistivity. Again, this can only be discovered through external information.

Also, many soundings lead to erroneous interpretations, because they are carried out in an anisotropic environment (non-parallel and non-horizontal layers, or large lateral variations). It is often possible to account for these during interpretation, when rising curve slopes are greater than 45° or curvatures of maxima and minima are too pronounced to be handled by nomograms. Carrying out a second sounding rotated by 90° and centred precisely on the same point confirms this problem of anisotropy (different curves obtained) and offers a more exact solution (Figure 5.23).



Finally, interpretation using common sense often eliminates certain solutions (aberrant value of ρ , number of different formations not corresponding to known geological context etc.) and enables the proper solution to be obtained.

Figure 5.23: Verification of the homogeneous nature of the zone. Two Schlumberger soundings centred on the same point (o) but in almost perpendicular directions bedrock zone in Burkina Faso, ACF, 2003).

Computer interpretation

There are a number of computer programmes which provide rapid interpretation of electrical soundings, although they are all confronted with the same problems as manual interpretation (suppression, equivalence, number of formation types). Software does not provide the same 'feeling' as manual interpretation, but it does have several advantages, notably speed and flexibility, that make it easy to change models and to visualise equivalences and suppressions.

For those with little experience therefore, their use is recommended only for checking the validity of manual solutions. Staff who are experienced

in manual interpretation can, on the other hand, take full advantage of the speed of direct computer interpretation.

ACF has chosen to use the IPI2WIN programme, developed by Moscow State University's Geophysical Laboratory. It is well suited to use in the field, and can be downloaded as freeware (http://geophys.geol.msu.ru/rec_lab3.htm). The programme has been modified to suit ACF's requirements, and it is now a very user-friendly, high performance package (comparison between soundings, exploration of the domains of equivalence etc.). It can also interpret data acquired using any of the principal equipment set-ups (Schlumberger, Wenner, pole-pole, pole-dipole).

is however a little low, and it may be that this formation, which corresponds to a fissured/weathered bedrock zone, may be only slightly argillaceous, or that the water is not very mineralised. Only experience and observations in a zone can allow these hypotheses to be confirmed or rejected. It is also possible that this potential aquifer is covered by an argillaceous layer (resistivity 18 Ωm on 7.5 m) that provides good protection of the groundwater from surface pollution: it is therefore probable that the aquifer is confined.

4.2.3.4 Square sounding

Implementation

A square survey can be used to complement the standard survey in an anisotropic environment. The principle is the same as for the Wenner survey. The main difference is the fact that, for every depth investigated, the apparent resistivity is measured in different directions, the array being turned before separating electrodes A and B to probe to a greater depth.

The array is defined by $AB = MN$ and $OA = OB = OM = ON$. It is located as follows (Figure 5.24), directions being measured from the centre O:

- electrode A positioned at N315;
- electrode N positioned at N135;
- electrode B positioned at N045;
- electrode M positioned at N225.

While carrying out measurements in this configuration, the direction of measured resistivity is N090/N270 (parallel to the AB direction and MN). The angle between North and the direction of measured resistivity is then $\alpha = 90^\circ$. Reversing the connections of electrodes M and B on the resistivity-meter (Figure 5.25), a new configuration is obtained so that the apparent resistivity direction is N000°. The angle $\beta = 0^\circ$ is then defined. To obtain supplementary directions, the square array is rotated through an angle, e.g. 30° (Figure 5.26).

The array A1,B1,M1,N1 gives apparent resistivities for directions $\alpha_1 = 90^\circ$ and $\beta_1 = 0^\circ$, the A2,B2,M2,N2 array for directions $\alpha_2 = 120^\circ$ and $\beta_2 = 30^\circ$, and the A3,B3,M3,N3 array for directions $\alpha_3 = 150^\circ$ and $\beta_3 = 60^\circ$.

In the field, the angle of rotation of the array depends on the required degree of precision: in general, 30° or 45° is used. If the direction of fracture considered is N 000°, and the angle chosen between each measurement is, for example, 30° , the parameters specified in Table 5.VII are obtained.

When measurements have been taken for a given length AB in all directions, the side of the square may be increased in order to investigate greater depths.

Interpretation

The resistivities are calculated by the standard formula:

$$\rho_a = \frac{\Delta V \times K}{I}$$

Table 5.VII: Parameters of the square array in rotation.

Direction of lines implemented (direction OB)	Direction of lines obtained by reversing M and B (direction OB)	Direction of resistivities measured with respect to N000°
N 045	N 225	$\alpha_1 = 90^\circ, \beta_1 = 0^\circ$
N 075	N 255	$\alpha_2 = 120^\circ, \beta_2 = 30^\circ$
N 105	N 285	$\alpha_3 = 150^\circ, \beta_3 = 60^\circ$

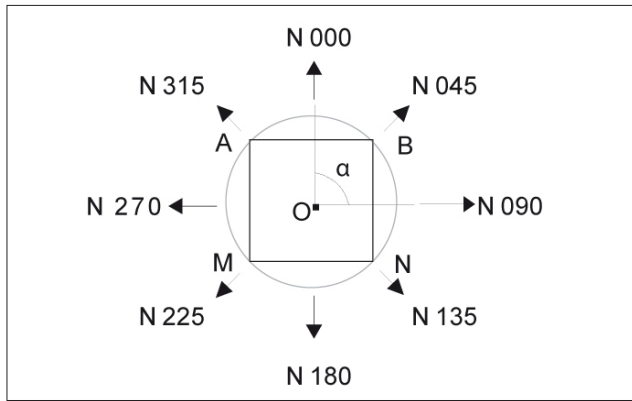


Figure 5.24: Layout for square survey.

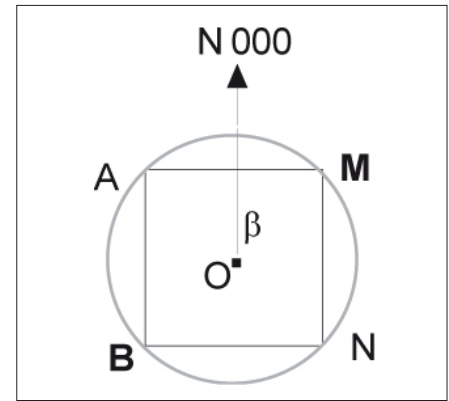


Figure 5.25: Inverted square layout.

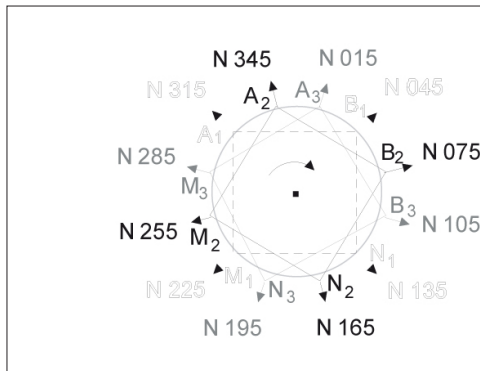
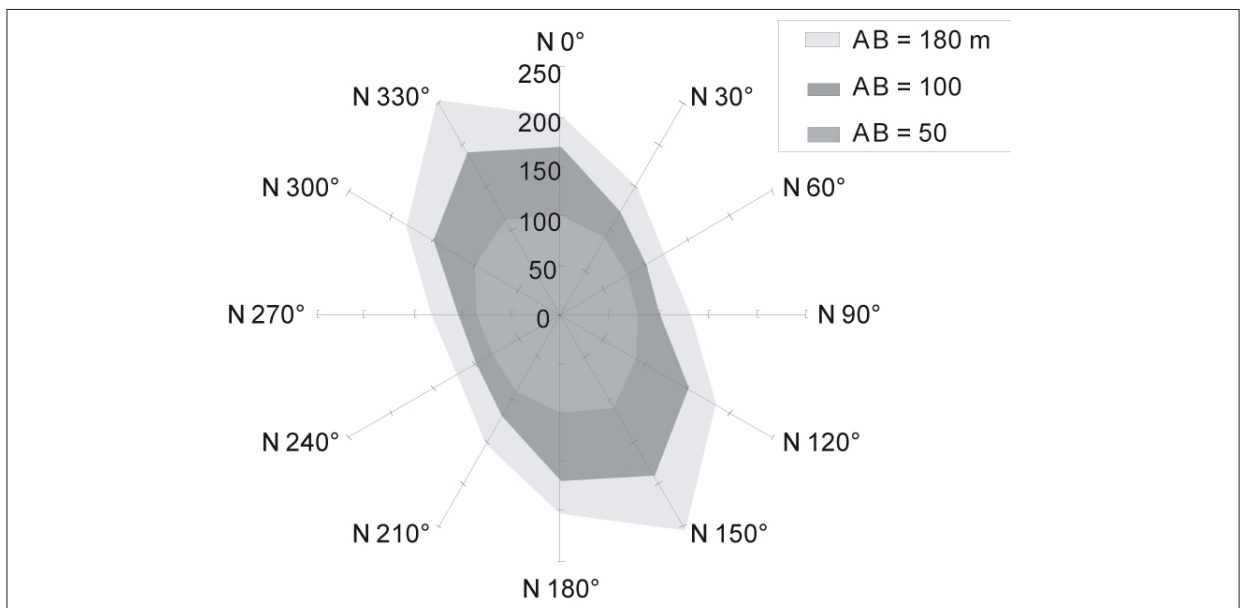


Figure 5.26: Rotation of square layout.

Figure 5.27: Result of square survey.

v



And:

$$K = \frac{2 \cdot \pi \cdot AB}{2 - \sqrt{2}}$$

Table 5.VIII gives several values of K for lengths AB normally used. Values of resistivity are plotted as a function of direction (Figure 5.27). The direction of anisotropy at various depths is clearly shown graphically. In Figure 5.27 the anisotropy (corresponding to a fractured bedrock zone) in direction N 60° increases with depth.

Table 5.VIII: Values of K.

AB (metres) (= MN = AM = BN)	K
180	1 931
140	1 502
100	1 073
70	751
50	536

The descriptive parameters are defined as follows:

- direction of the anisotropy θ ;
- coefficient of anisotropy $\lambda = \rho_a/\rho_b$ (the higher the value of this coefficient, the greater the anisotropy);
- mean resistivity $\rho_m = (\rho_\alpha \rho_\beta)^{1/2}$.

For every direction and depth of investigation, it is therefore possible to know the magnitude of the anisotropy (λ), its direction (θ) and the mean resistivity (ρ_m).

4.2.3.5 Resistivity profiling (RP)

Implementation

By moving a device of fixed length AB–MN in a given direction (Figure 5.28A), a profile of apparent resistivities of a slice of ground of approximately constant thickness is obtained.

The direction of the profile is chosen on the basis of the supposed directions of the anomalies detected in the field or by photo-interpretation. The ideal procedure is to take a perpendicular slice through the anomaly, in order to determine its width and to estimate its tilt angle (Figure 5.28B).

The profile ρ_a shown in Figure 5.28B is obtained for a small length AB, whereas profile ρ_a corresponds to a larger AB ($\rho_1 < \rho_2$). The electrode spacing AB is therefore determined by the depth under investigation. The measurement step-sizes are a function of the desired precision: a step of 10 m may be taken as standard.

Interpretation

The apparent resistivity measurements are plotted on millimetre graph paper, with the length AB/2 as abscissa and the apparent resistivities as ordinate (Figure 5.29).

In bedrock zones, a study by *Burgeap* (1984) shows that anomalies indicated by ER are increasingly favourable when:

- the width of the anomaly (measured between the two points of inflection) is less than about 50 m;
- the minimum apparent resistivity is between 50 and 120 Ωm^* ;
- the resistivity contrast (ratio of apparent resistivity of the surrounding rock to the minimal resistivity of the anomaly) is greater than 1.5. (This value must be larger for wider anomalies.)

4.2.3.6 Electrical resistivity imagery (ERI)

Implementation

Electrical resistivity imagery is performed by carrying out a series of measurements in two dimensions. For this purpose, a number of pegs are placed along the profile. Whereas a normal VES needs 4 electrodes, the example shown in Figure 5.30 uses 64 pegs. Each peg is connected to the resistivity-meter via a particular address, and can therefore be used as an electrode for injection (A or B)

* This spread of resistivity is approximate because it is a function of local conditions. With a little experience of a given zone, it is easy to redefine it.

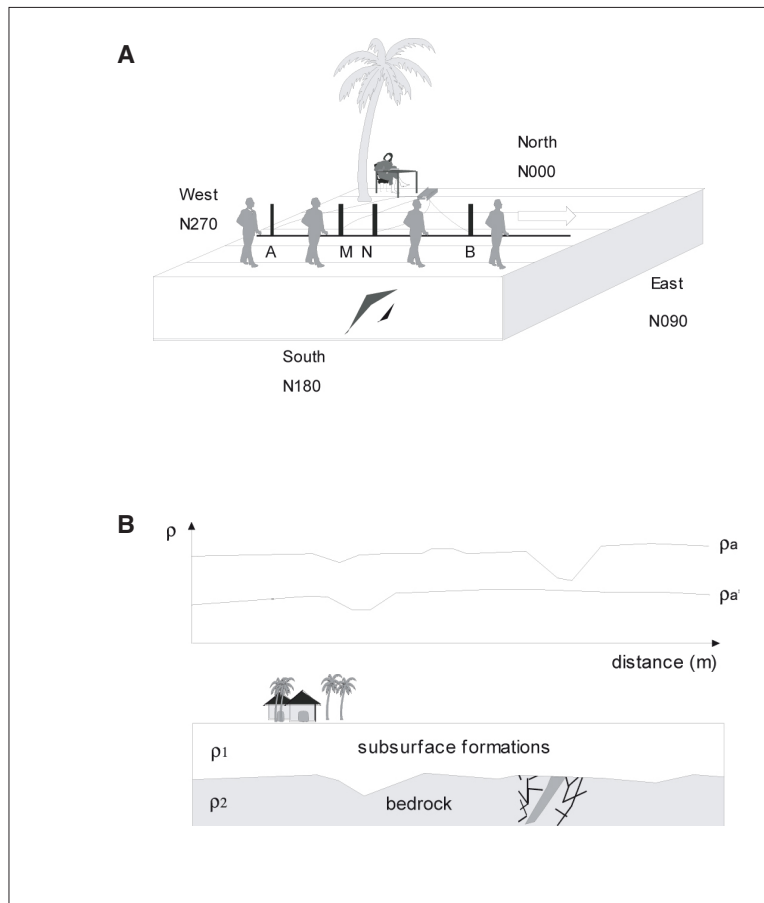


Figure 5.28: Resistivity profiling.
A: setting up. B: example.

or for potential measurement (M or N). A sequence is first recorded in the resistivity-meter memory, defining which pegs will be used for each measurement: all possible combinations of electrodes are thus used for exploring at different depths and at various points along the section. Symmetrical arrays are generally used, such as the Wenner (sensitivity to both lateral and vertical variations) and the dipole-dipole (higher sensitivity to vertical anomalies).

Interpretation

The series of measurements is interpreted using software to calculate the resistivity cross-section of the formation. Figure 5.31 shows a cross-section obtained using resistivity imagery performed by ACF in Mozambique, with a Wenner array (128 electrodes, 4-metre spacing) in a clayey sandstone formation. The usefulness of this method is evident in this type of heterogeneous formation: the calculated resistivity contrasts reveal the geological structures that are interpreted on this example as a zone of clayey sandstone without aquifers in the north-west and a sandy, potentially water-bearing zone to the south-east.

ERI is the only operational method which can be used routinely in humanitarian programmes to perform two-dimensional measurements. It is therefore the method of choice in all heterogeneous environments.

Two programmes are used by ACF to perform ERI. The freeware programme X2IPI developed by the Geophysics Laboratory of Moscow State University (http://geophys.geol.msu.ru/rec_lab3.htm) is used to prepare the sequences of measurements which will be recorded by the resistivity-meter and to control and analyse the recordings (analysis of measurements, suppression of noisy data etc.). The RES2DINV programme is used to interpret the apparent resistivities (measurements previously analysed by X2IPI) and to obtain a cross-section of calculated resistivity (Figure 5.31). A basic version of this programme, sufficient for simple data interpretation, is available as freeware (<http://www.goelectrical.com>).

Setting up an ERI survey differs from a standard VES in that all the electrodes must be positioned before beginning the measurements, which take about 45 minutes for a 64-electrode Wenner sequence.

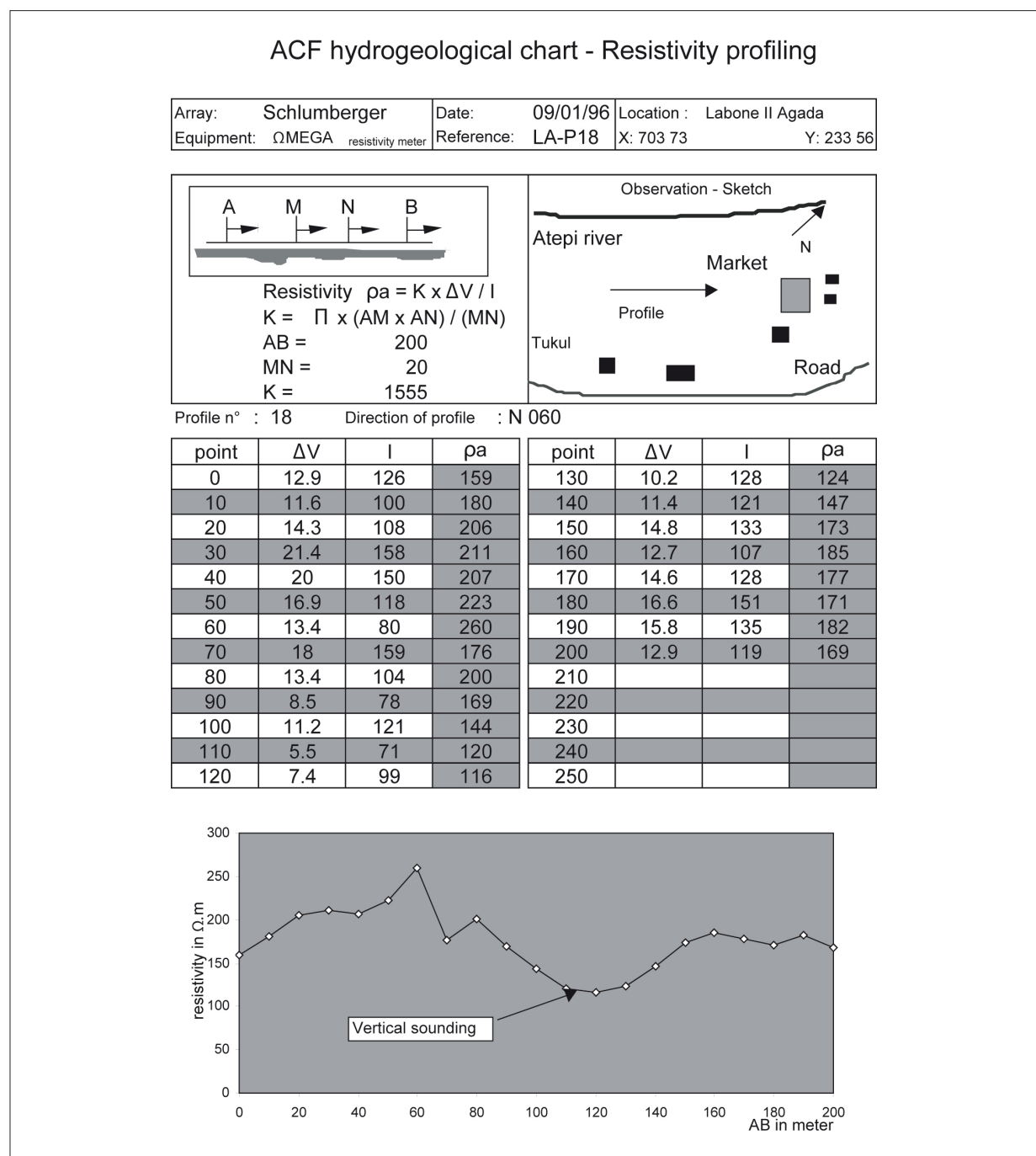


Figure 5.29: Example of a traverse in a bedrock zone (ACF, Sudan, 1996).

This example concerns granito-gneissic formations. The Schlumberger array chosen was AB = 200 m, MN = 20 m, with measurement points every 10 m. In the bedrock zone studied, the objective was to identify the largest zones of weathering, as these are the most likely to be an aquifer. Solid (and sterile) bedrock has a high resistivity, while the weathered water-bearing sections are conductive, so that it is possible to identify, at a distance of about 110m in the profile, a zone where the apparent resistivity is lower and which is therefore likely to be weathered to a much greater depth. The VES presented in Figure 5.16 was performed at this 110m point.

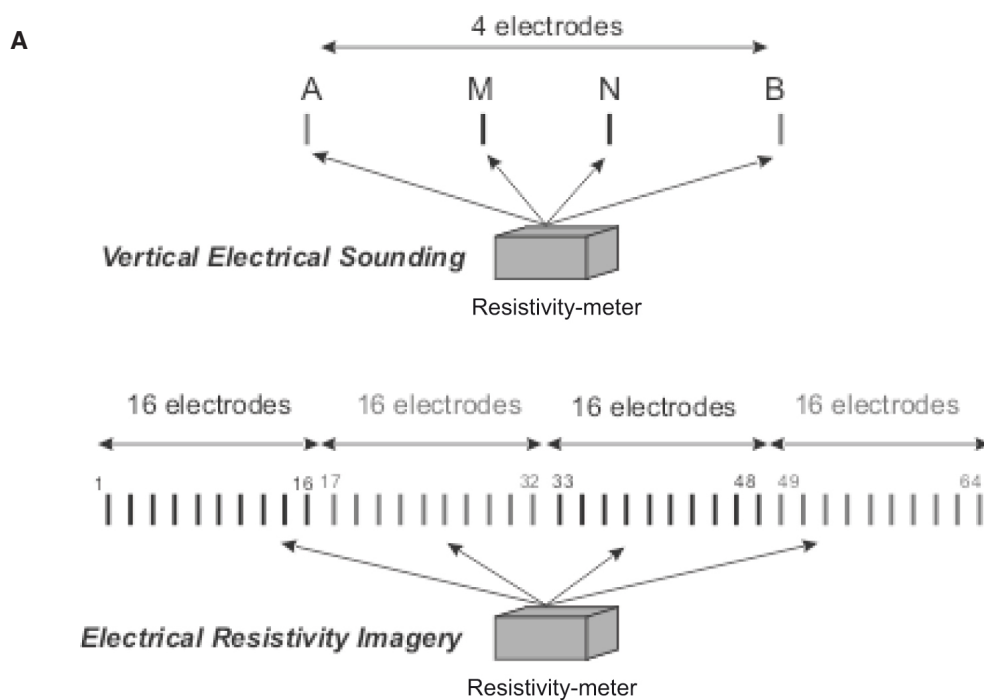


Figure 5.30: ERI measurement array.

A: comparison between a normal VES array and an ERI array. **B:** Wenner array, 64 electrodes and 4-metre spacing used in Honduras (ACF, 2000).

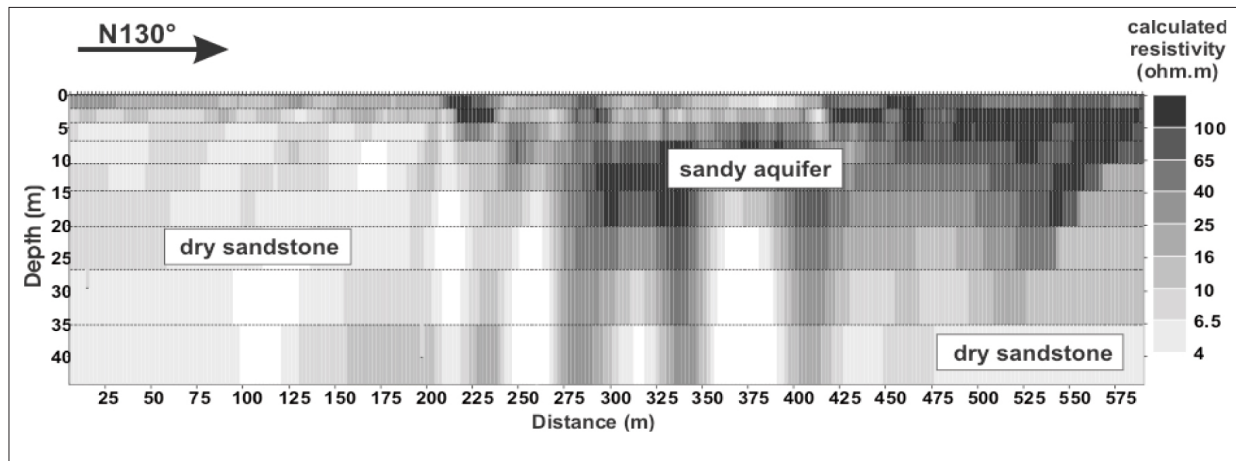


Figure 5.31: Electrical Resistivity Imagery (ACF, Mozambique, 2000).

4.2.3.7 Borehole Electrical Logging (BEL)

Borehole logging is not strictly speaking a method of groundwater exploration, but it forms a part of electric geophysical methodology. Based on the same principle as surface methods, it consists of using different arrays of electrodes to measure resistivities directly in the borehole. When carried out right at the end of the drilling operation, it provides high-precision localisation of water-generating zones, and therefore defines the optimal position of screens.

The main value of this method lies in obtaining the maximum amount of information from reasonably productive boreholes in sedimentary zones, where the passages from sandy to argillaceous zones are progressive. It also provides a clear indication of water mineralisation problems, and therefore of salinity.

Implementation

Various types of measurements are possible in electrical logging. The experience of ACF in Cambodia showed that normal probe and fluid resistivity surveys enable the collection of the main information being sought.

Normal probes (Figure 5.32) use a pole-pole configuration: the electrodes B and N are therefore set at infinity (according to Chapellier (2000), B and N can be regarded as infinite when $MN > 14 AM$). The AM spacing determines the 'lateral depth' of investigation. A spacing of 20 cm (N20 probe) gives acceptable precision at layer boundaries, but will be greatly affected by drilling mud. The N80 probe corresponds to a spacing of 80 cm, less accurate in vertical definition, but with a measured resistivity that approximates to the true resistivity.

The fluid resistivity measurement probe uses the Wenner configuration ($a = 2$ cm) with the four electrodes immersed in the fluid. The fluid resistivity measurement is essential for the interpretation of the other logging measurements. Also, it can indicate, where necessary, the zone of water-intake (which generally exhibits resistivity variations).

The probe developed by ACF can perform temperature and fluid resistivity measurements, spontaneous potential (SP), mono and 8, 16 and 32-cm normal electrodes. This probe is manually lowered into the borehole. The measurement intervals are 50 cm, or less if a more precise definition is required. Measurements are taken at the surface using a specific resistivity-meter.

Direct on-site completion by an operator of an Excel spreadsheet furnishes curves of the measured parameters as a function of depth. It is important to measure the resistivity and the temperature of the fluid and the SP during the descent, when the water in the borehole has not been agitated too much, and that of the normal and mono probes during the ascent.

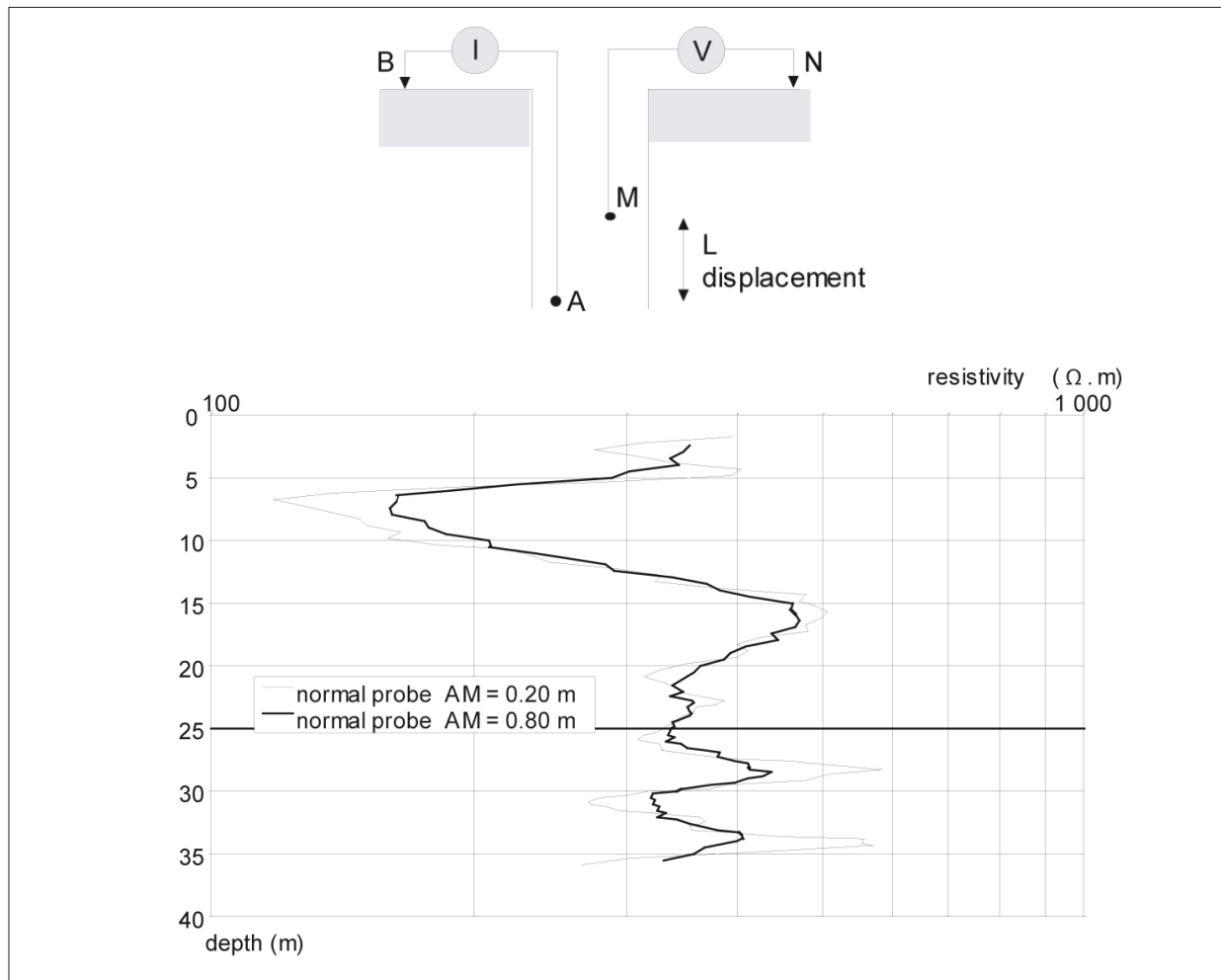


Figure 5.32: Electric logging (ACF, Siem Reap, Cambodia, 1998).

Interpretation

Resistivity values are calculated *in situ*. Note: the geometric coefficient K corresponds to a complete sphere and not to a hemisphere; it is therefore given by:

$$K = \frac{4\pi}{1/AM - 1/BM - 1/AN + 1/BN}$$

An example of curves for N20 and N80 probes is shown in Figure 5.32. It is from a borehole sunk by rotary drilling (see Chapter 8) using a polymeric mud (ACF Cambodia, 1998) in a clay-sand zone. During the descent, the fluid-resistivity probe did not show any strong contrast at 35 m depth in the borehole (logging carried out some minutes after withdrawal of the drilling pipes) and took one measurement of the resistivity of the drilling mud of 280 Ωm (about 35 $\mu\text{S/cm}$). The normal probes used during the ascent indicated two zones: a conductive zone centred around 7 m depth, interpreted as argillaceous (= 150 Ωm), and a more highly resistive zone (= 350 Ωm) at 12 to 35 m. Within this more resistive zone, three passages proved to be distinctly more resistive (> 400 Ωm) and are interpreted as aquiferous sands. This interpretation, confirmed by the analysis of cuttings, enabled optimal exploitation of the borehole ($Q = 5 \text{ m}^3/\text{h}$).

To exploit BEL information to the full, the results should be interpreted only after applying a certain number of corrections (layer thicknesses, resistivities etc). Then the various measurements should be correlated with each other. See bibliography for more details.

4.3 Electromagnetic methods

4.3.1 PRINCIPLE

Electromagnetic (EM) methods use those physical properties of the ground that affect the magnitudes of time-varying currents, to study the distribution of conductivity values in the ground (conductivity is the inverse of resistivity).

EM phenomena are classically described by Maxwell's equations, which state that any time-varying electric current produces an electromagnetic wave, which consists of an electric field and a magnetic field propagating in space and oscillating at the same frequency, perpendicular to each other.

In contrast to the method of electrical resistivity measurements, which uses DC current and requires the use of electrodes, EM methods use the induction properties of electromagnetic waves, and do not require direct contact with the ground. They can thus be implemented very quickly. These methods generally have a higher resolution than DC methods in conducting formation types (the phenomenon of induction is much stronger in conducting formations), but resistive formation types are less well defined.

4.3.2 MEASUREMENTS

Many techniques can be used to measure the response of the underground formations to an EM stimulus. However, most are based on the same sequence of events:

- *Stimulus*: the 'primary', time-varying (generally oscillating) electric current is created in an electric cable. This current generates a 'primary' magnetic field.
 - *Reaction*: when this primary magnetic field traverses the underground formations, an electromotive force and an associated current are created by induction in these formations. This 'secondary' current, in turn, generates a secondary magnetic field.
 - *Reception*: the secondary magnetic field also traverses the underground rock formations and induces a signal in a receptor placed at the surface. Various quantities can thus be recorded: the resultant field of the primary and secondary fields, the secondary field on its own, the electric or magnetic components etc. These data are then analysed to yield information about underground conductivity.
- ACF programmes have employed several EM methods:
- the VLF method;
 - the Slingram method;
 - TDEM soundings.

4.3.3 THE VLF METHOD

VLF (very low frequency) antennas carry alternating electric current oscillating at frequencies between 15 and 30 kHz, which are used by the military for long-distance communications.

Maxwell's equations show that all oscillating electric currents generate electromagnetic waves, which appear as an electric field and a perpendicular magnetic field of the same frequency (Figure 5.33).

Difficulties with the method

The principal limitation of the VLF method is its dependence on the primary field. The use of VLF transmission appears to be diminishing (certain transmitters have been shut down) and this method may no longer be operational in a few years time. In addition, the profile directions of VLF measurements are determined by the position of the primary-field-emitting antenna, which can be a severe limitation in the field.

Moreover, the sounding depth of the VLF method is rarely more than 20 metres, and may only be a few metres if the surface layer is highly conducting (clay, saltwater).

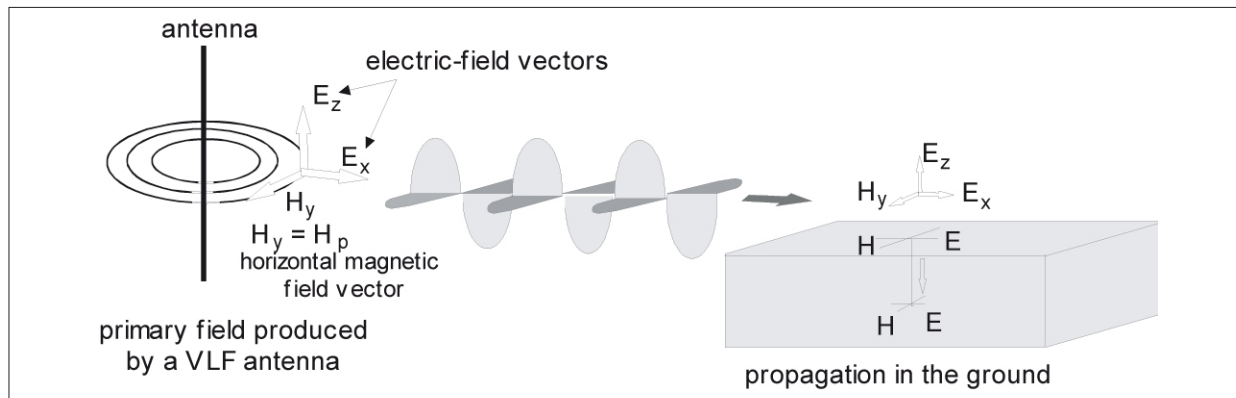


Figure 5.33: Primary electromagnetic field.

Vector E_x corresponds to the horizontal component of the primary electric field (same direction as the antenna). Vector E_z corresponds to the vertical component of the same field. Vector H_p (or H_y in a system of x, y, z co-ordinates) corresponds to the horizontal component of the primary magnetic field (at right-angles to the direction of the antenna).

However, the rapidity of VLF measurements means they are still quite widely used, particularly to measure profiles in fractured media (using ‘Wadi’ equipment), or to map apparent iso-resistivity in sedimentary media (using ‘TVLF’ equipment).

Magnetic field measurements using Wadi equipment

Originally designed for ore exploration, the Wadi equipment (from the ABEM Company) has been adapted for use in exploration for water, notably in fracture zones, where discontinuities are significant. It does not give useable results in sedimentary zones (sedimentary structures are too laminar and discontinuities insufficiently marked). As with other VLF equipment, the anomaly signal becomes too weak to be measured when argillaceous cover is significant (recorded values are around 0).

When the primary electromagnetic field intersects a target conductor, this conductor creates in turn a secondary electromagnetic field (H_s) of the same frequency, but out of phase with the primary field. When the joint effect of the primary and secondary fields is measured (Figure 5.34), the resulting field, instead of oscillating in a linear manner, describes an ellipse (the ‘ellipse of polarisation’).

Measurement devices give:

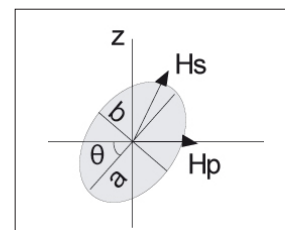
- the ratio $b:a$ of the minor and major axes of the ellipse (in%), which provides information on the conductivity of the anomaly (defined as the real/imaginary ratio of the Wadi signal).
- the angle θ between the horizontal and the major axis of the ellipse of polarisation. The cosine of this angle multiplied by the major axis will give an approximation to the real component, which is used to locate anomalies.

Wadi is essentially used for the location of fractures in lightly-covered bedrock zones, implementing profiles as RP. Compared to electrical methods, implementation is much more rapid, allowing larger exploration zones to be defined.

The interpretation software gives information on the depths and shapes of anomalies (Box 5.3). The equipment gives the same information in a more schematic form directly in the field. Whatever the information obtained, it is strongly recommended that VES measurements should be carried out on anomalies to confirm their potential and to specify depths.

An example of well location in Mali using Wadi is shown in Figure 5.35.

Figure 5.34: Ellipse of polarisation of the field resulting from the interaction between primary and secondary fields.



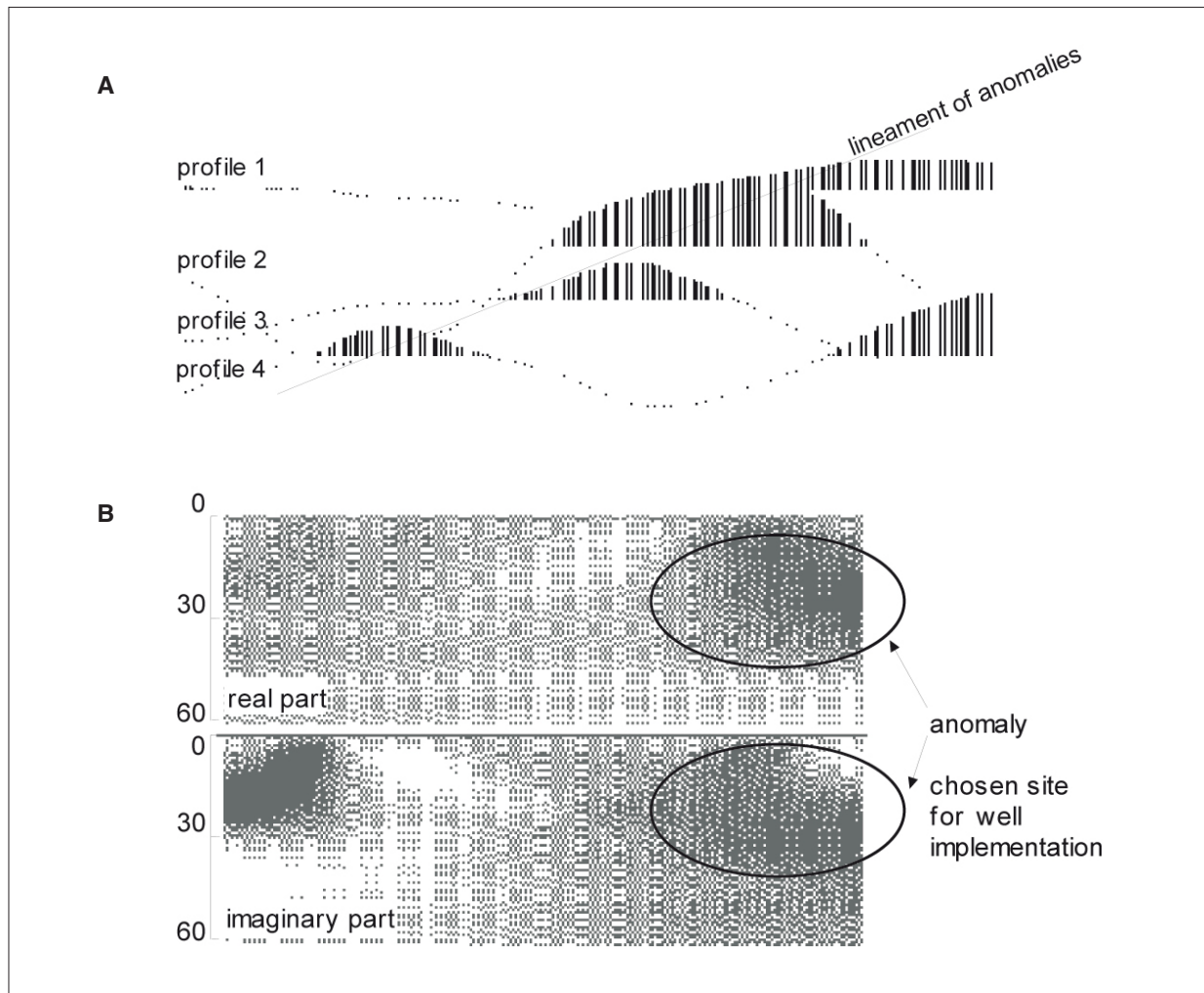


Figure 5.35: Well sunk after Wadi survey (ACF, Mali, 1998).
LS = 10.2 m; depth = 15.2 m. A: multiple profiles Re. B: vertical section.

Electric-field measurements using TVLF equipment

The measurement of the horizontal components of the resulting electric (E_x) and magnetic (H_y) fields allows one to calculate the apparent resistivity of layers traversed, using the relation:

$$\rho_a = \frac{1}{2\pi \cdot \mu_0 \cdot f} \times \frac{E_x^2}{H_y^2}$$

where μ_0 is the magnetic permeability in vacuum ($= 4\pi \cdot 10^{-7}$) and f the antenna frequency.

The TVLF (from Iris Instruments) can simultaneously measure E_x and H_y , and calculates the apparent resistivity for the antenna frequency used. To measure E_x , two electrodes planted in the ground at every station are used: if the measurement point is electrode 1, electrode 2 is planted at a distance of 10 m, in line with the direction towards the antenna (and therefore of the profile). H_y is measured perpendicular to the direction of the profile.

The phase shift between E and H is also measured and is represented by the value j . In homogeneous environments, its value is 45° or -135° depending on the sense of the profile or the angle of the electrodes. It indicates how the resistivity varies. When $j < 45^\circ$ (or -135°), resistivity increases with depth; conversely, when $j > 45^\circ$ (or -135°), resistivity decreases with depth.

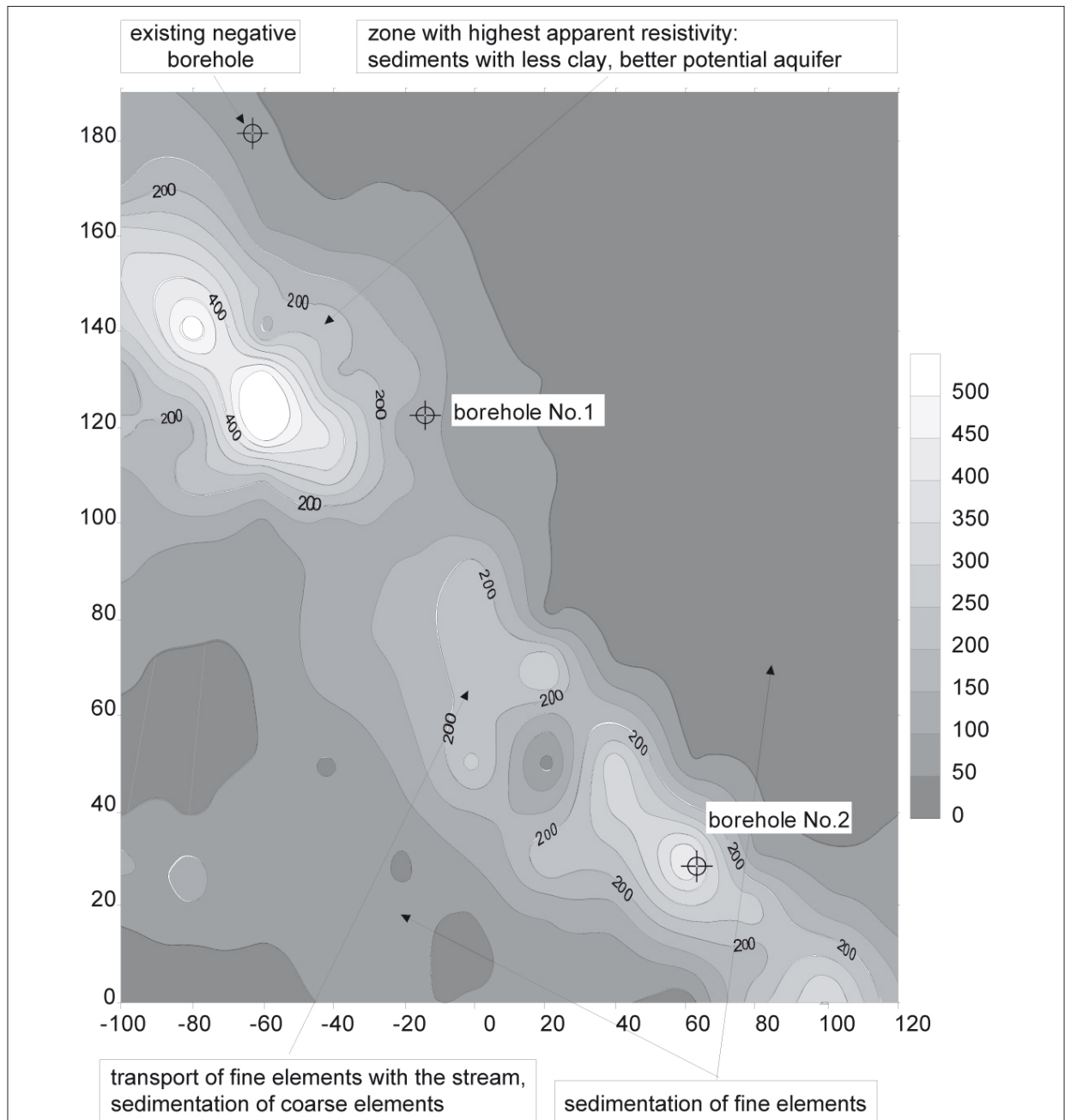


Figure 5.36: Iso-resistivity map (ACF, St André Farm, France, 1995).

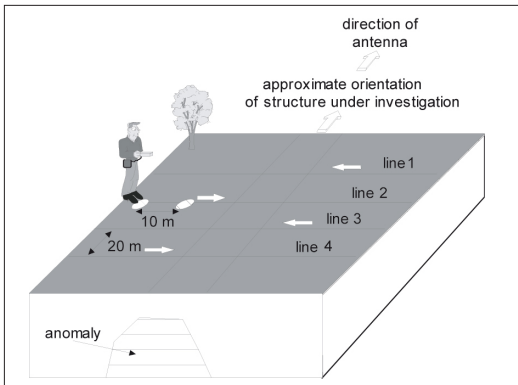
TVLF in the resistive mode is mainly used in sedimentary environments to map variations in apparent resistivity. An example of borehole implementation in alluvial deposits from the farm at St André (France, ACF, 1995) is given in Figure 5.36. Borehole No. 1 was abandoned (formations included silt-sand sediments with gravel passages of widely varying thicknesses, and also including pebbles). Borehole No. 2 revealed the same type of profile (more pebbles), with incoming water at a depth of 44 to 52 m. The old borehole had encountered neither pebbles nor gravel, but much more clayey deposits down to a depth of 60 m. The iso-resistivity map can be interpreted as indicating the presence of an old riverbed, with deposits of coarse elements (pebbles, gravel, sand) where the current was strongest. The deposit of finer elements (silt) occurs at the riverbanks.

Box 5.3

Operating the Wadi

The Wadi measures the magnitude of the resultant of the primary and secondary fields, which is transformed into a complex number function with imaginary and real components. The imaginary component is 90° out of phase with the primary field, whereas the real component is in phase with this field.

To be able to detect a conductive anomaly, the primary magnetic field must pass perpendicularly through the anomaly, i.e., the direction of the transmitter is the direction of the anomaly. In other words, the antenna must be perpendicular to the direction of the displacement and parallel to that of the anomaly. It is necessary to take multiple profiles in order to confirm the information (Figure 5.37). It is important to maintain the same measurement direction for a series of profiles, but the direction itself is unimportant.



The distance between profiles depends on the width of the anomaly, the area to be covered, and the precision required. A distance of 10 m between each measurement is a good compromise between precision and time taken to make the measurement.

It should be noted that the Abem-Wadi software has its own system of coordinates, and it is therefore recommended that the programmed distances should be maintained, and that the 'north' of the Wadi should be made to correspond to the direction of the initial measurement.

Figure 5.37: Anomaly/antenna/profile directions, VLF.

Results can appear in two forms, real or imaginary, each being either filtered or unfiltered. The filtered real curve gives an initial interpretation and indicates the observed anomalies in the field. For a good interpretation (Figure 5.38), the anomaly must be completely covered. The sign (+ or -) of the imaginary curve depends largely on the thickness of surface formations and their conductivity. It is therefore necessary to try to observe the amplitude of the curve rather than the sign.

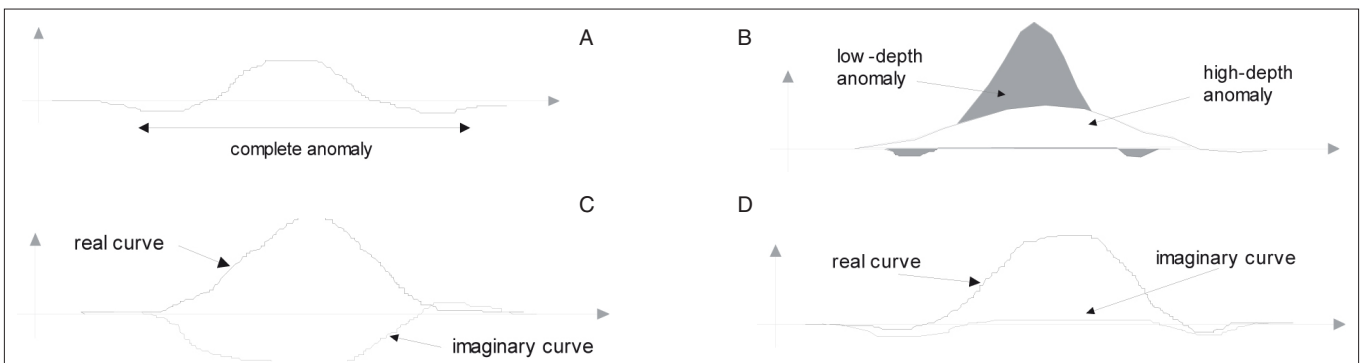


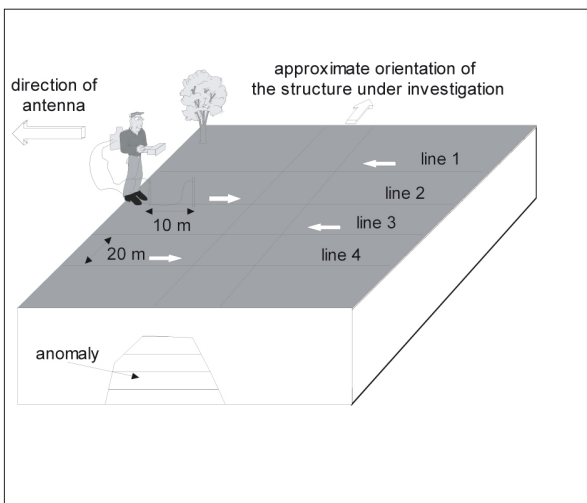
Figure 5.38: Real and imaginary curves of anomalies detected by Wadi.

A: real curve of an observed anomaly.

B: modification of the shape of the real curve according to anomaly depth.

C: comparison of real and imaginary curves in the case of an imaginary part of high amplitude indicating an anomaly of high conductivity (saline water, clay etc.).

D: comparison of real and imaginary curves in the case of an imaginary part approaching zero which indicates the presence of poorly conductive ground (fracture full of lightly mineralised water).



Operating the TVLF

A parallel series of profiles oriented in the direction of the antenna is taken perpendicular to the structure of interest. In the absence of any indications as to the direction of structures (no geomorphological clues, outcrops etc.), a pair of profiles is taken using two antennas set at right angles, and the frequency which gives the most information is used.

In all cases, it is preferable to select only one frequency (antenna) per series of measurements. The profile directions must be verified using a compass, and the data must be rigorously entered in the TVLF. For example, to record a 300-m profile, station intervals of 10 m, with 20 m between each profile is suggested (Figure 5.39). If profile no. 1 is defined as LINE = 00100 m, profile no. 2 will be LINE = 00120 m. At the time of data capture, STATION 10 of LINE 00120 will correspond to STATION 290 (300 m profile) of LINE 00100.

Figure 5.39: TVLF profile in resistivity mode.

4.3.4 THE SLINGRAM METHOD

This method consists of measuring the ratio of the signal emitted by an transmitter to that detected in a receiver a few metres or tens of metres away (Figure 5.40).

The transmitter and the receiver used by ACF (Geonics EM34) are coils of cable about one metre in diameter. Measurements can be made at 3 frequencies (6.4, 1.6 and 0.4 kHz) corresponding to 3 transmitter-receiver distances ($a = 10, 20$ and 40 m) and using 2 orientations: the horizontal dipole (HD), in which transmitter and receiver are placed vertically in the same plane, and the vertical dipole (VD) with transmitter and receiver horizontal in the same plane.

These various configurations give a variation in the investigation depth from a few metres (HD and $a = 10$ m) up to a few tens of metres (VD and $a = 40$ m).

This method can be used to construct profiles of apparent resistivity or maps of apparent iso-resistivity.

Figure 5.41 shows two maps made by ACF in northern Uganda (Gulu), in a weathered, granito-gneiss bedrock zone. The measurements were all performed with the same receiver/transmitter spacing ($a = 20$ m), but using the two configurations, horizontal dipole (HD) and vertical dipole (VD). These maps clearly reveal the presence (at a depth of about 20 m as measured by VD) of a conducting zone to the East, which probably corresponds to an argillaceous weathered area, whereas the more resistive zone to the West could be an aquifer.

Figure 5.42 corresponds to two profiles made using the EM 34 equipment on the abscissa $x=0$ of Figure 5.41. On the same graph is also shown an RP recorded using the Wenner configuration with the distance $AB/2=40$ m. The investigation depths of the RP and the EM34 VD measurements are comparable, and the information furnished by the profiles is similar. But the EM profile acquisition is much quicker (500 m per hour in these conditions compared with 200 m per hour for the RP) and requires only 2 people (1 for the receiver, 1 for the transmitter) compared with 4 for the RP (1 for each electrode)

The Slingram method may be preferable to the RP method at shallow investigation depths (10 to 40 m maximum) when the surface formation is not too conducting. But if this is not the case, the sounding depth is generally too low to reach typical hydrogeological structures of interest. On the other hand, this method is more versatile than VLF (measurements easily made at different depths, no problem of profile orientation etc.) and does not rely, like VLF, on an external transmitter whose signal may be difficult to pick up.

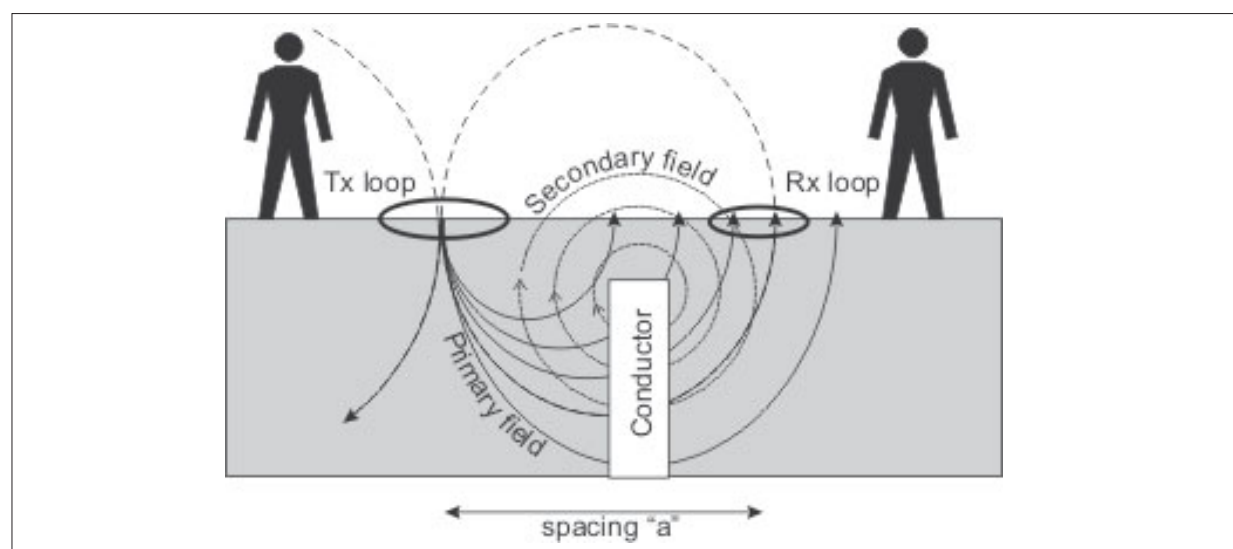


Figure 5.40: Slingram configuration (VD), EM 34.

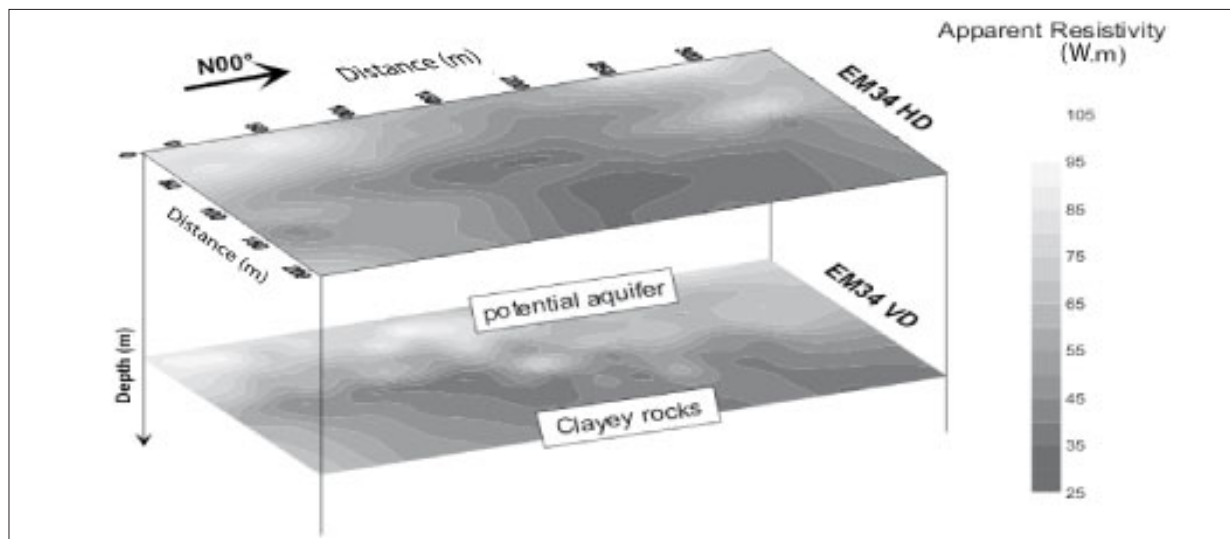


Figure 5.41: Apparent isoresistivity map (ACF, Uganda, 1999).
EM34 $a = 20$ m, 20-m steps.

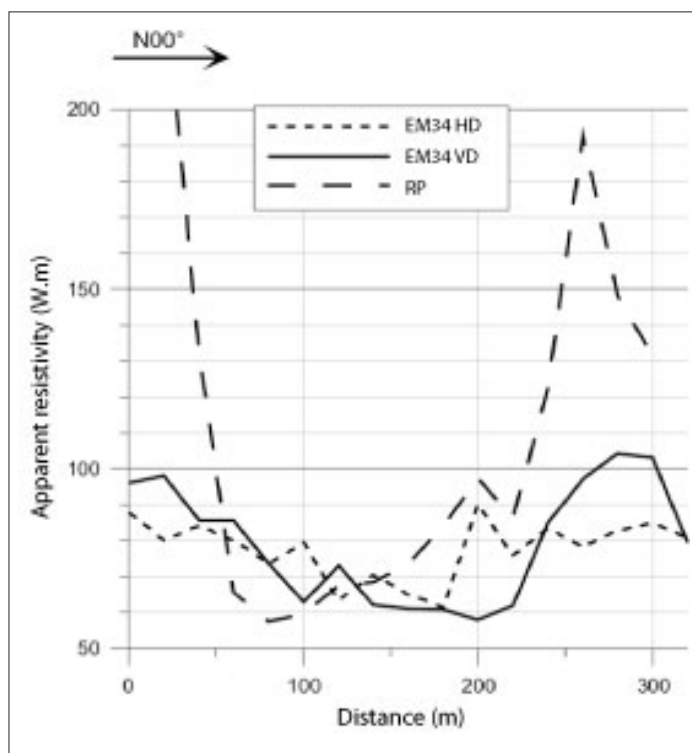


Figure 5.42: Comparison between RP and EM34 profile.
Profile recorded on the abscissa $x=0$ of Figure 5.41. EM34 $a = 20$ m, RP $a = 40$ m, 20-m steps.

4.3.5 TDEM SOUNDINGS

The TDEM (Time Domain Electro-Magnetism) method is different from the EM methods described above, in that the investigation depth changes with time (rather than with frequency, as in the other methods presented). It is the easiest way in which to perform very deep soundings.

A square loop of electrical cable is usually used as a transmitter (Tx) (Figure 5.43). It is placed on the ground, and its dimensions are chosen to suit the required investigation depth (usually from 40 m on a side for shallow investigations and up to 200 m on a side for deep surveys). A square-wave current (from 1 to 30 A) is passed through the loop; it is cut off suddenly to produce a variation in the

Figure 5.43: TDEM sounding principle.
The dashed lines represent the secondary current at different delay times (t_1 , t_2 , t_3 ...) after the cut-off of the primary current (t_0).

flux of the primary magnetic field, which in turn produces induced currents in the ground (Foucault currents). The amplitude of these induced currents decreases quickly with time. This decrease itself generates another induced current, but at a greater depth. The primary excitation is therefore capable of propagating itself to great depths.

These induced currents generate secondary magnetic fields that can be measured at the surface. As the currents penetrate more deeply with time, measurement of the secondary field at the surface at different times gives information from different depths. The receptor (Rx) is usually a pre-wound multiple loop, a simple loop of cable placed on the ground that could be the same loop as that used previously as Tx.

The injection rate of the primary current is chosen according to the required investigation depth. In practice, measurements start as soon as possible after primary current cut-off in order not to lose the initial information representing the shallow formations.

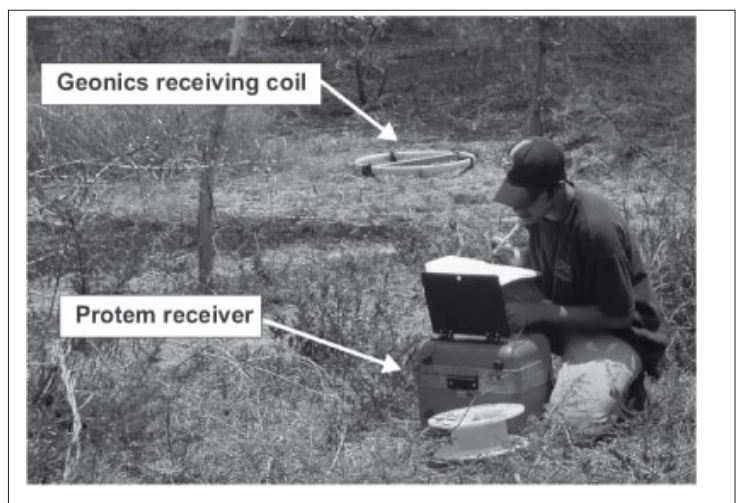
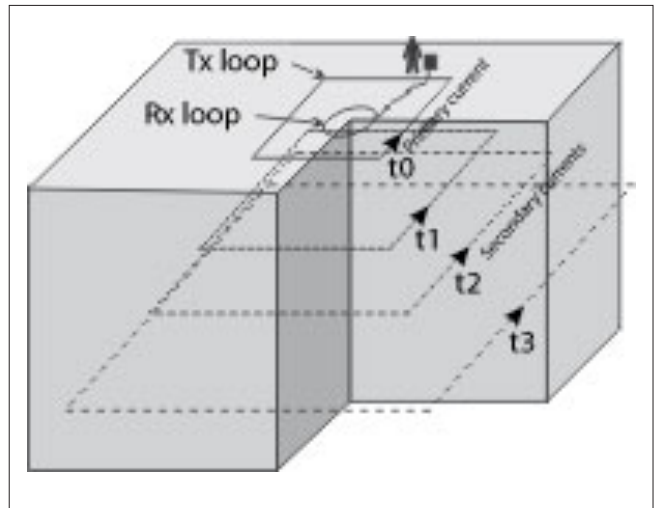
The process is controlled from a transmitter/receiver box that records the set of measurements. The approximate set-up time for a TDEM survey is about 30 minutes for 2 people, and the measurement time is about 15 minutes. The maximum investigation depth ranges from 100 to 300 m, depending on the Tx loop dimensions, the primary current amplitude and the local EM noise level.

Two types of equipment have been used by ACF: the Protem system (from Geonics, Figure 5.44) and the Temfast system (<http://www.aemr.net>). The Protem can be more powerful (1 to 30A depending on the generator) but has a total weight ranging from 50 kg (12V battery as generator) to 100 kg (220V AC generator), while the Temfast produces a primary current of 1 or 4 A, depending on the model, for a total weight of under 10 kg.

The interpretation of results is performed at the base, using software. The procedure is similar to that used for interpreting electric soundings, and gives families of curves: field records (voltage/time and apparent resistivity/time), and interpreted curves (calculated resistivity/depth and equivalences).

Figure 5.45 shows an example of a TDEM survey carried out by ACF in Cambodia in a clay-sand formation type, using a Tx loop 75 m on each side, and a receiver coil of area 31.4 m². This survey covers four levels, the first of which, 0 to 40 m depth, shows the highest resistance (= 80 Ω m) and could be water-bearing, whereas the others are very conductive and are probably argillaceous (20 to 2 Ω m).

Figure 5.44: Protem equipment
(ACF, Honduras, 2000).



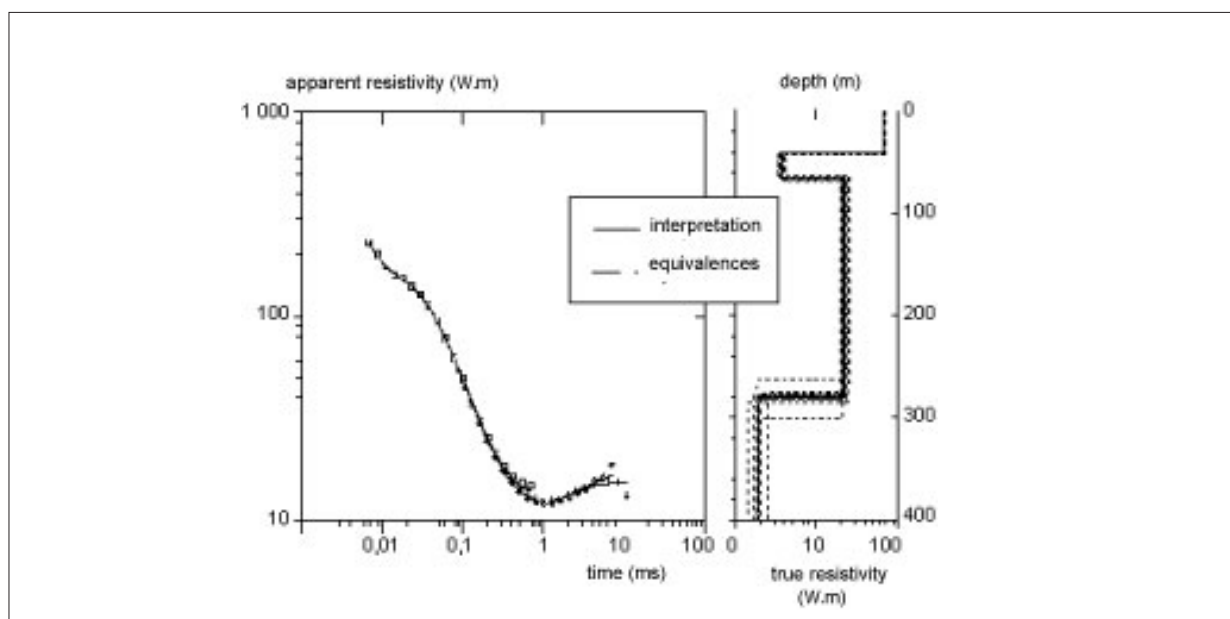


Figure 5.45: Example of TDEM sounding (ACF, Cambodia, 1998).

This example clearly illustrates the advantages and disadvantages of the method in the configuration used by ACF. It is fast and easy to carry out and can provide fairly deep soundings (with no pegs to plant or cable to deploy), and it gives better resolution in conductive formations than DC methods. However, TDEM does not distinguish the uppermost level because primary signal cut-off times, even when very short, do not allow measurements of the induced field to begin instantaneously. In Figure 5.45, for example, the first layer analysed is 40-m thick with a resistivity of 80 Ωm , but is in fact composed of different levels. The use of a smaller transmitter loop allows the primary signal cut-off time to be reduced, and therefore to begin measurements earlier, but even then it is not possible to obtain a detailed image of the first 5 to 10 m. TDEM does not work well in resistant environments, in which little or no induction occurs.

4.4 Magnetic Resonance Sounding

The main advantage of the Magnetic Resonance Sounding (MRS) method, compared to other geophysical methods for groundwater prospecting, is that the measured signal is from groundwater molecules: this reduces the degree of ambiguity in interpreting the data, since any measured signal proves the presence of groundwater.

Developed in the early eighties in Russia, MRS initially provided only the geometry and the water content of water-saturated layers. From 1999 to 2003, ACF conducted a research programme on MRS, in collaboration with various partners, showing that it is also possible to estimate the storativity and transmissivity of aquifers from MRS records calibrated using hydrogeological data.

ACF considers MRS now to be a very useful (and specific) tool for characterising the saturated zones of aquifers.

4.4.1 PRINCIPLE OF THE METHOD

It is known from nuclear physics that the hydrogen nucleus (proton) possesses an angular momentum and a magnetic moment μ . In a homogeneous magnetic field B_0 (such as the earth's geomagnetic field) the proton acts as a magnetic dipole and experiences a torsional moment that attempts to align it with the direction of the field. Therefore the angular momentum of the proton causes a precessional motion of μ around B_0 with an angular velocity ω_0 known as the Larmor frequency (Figure 5.46A).

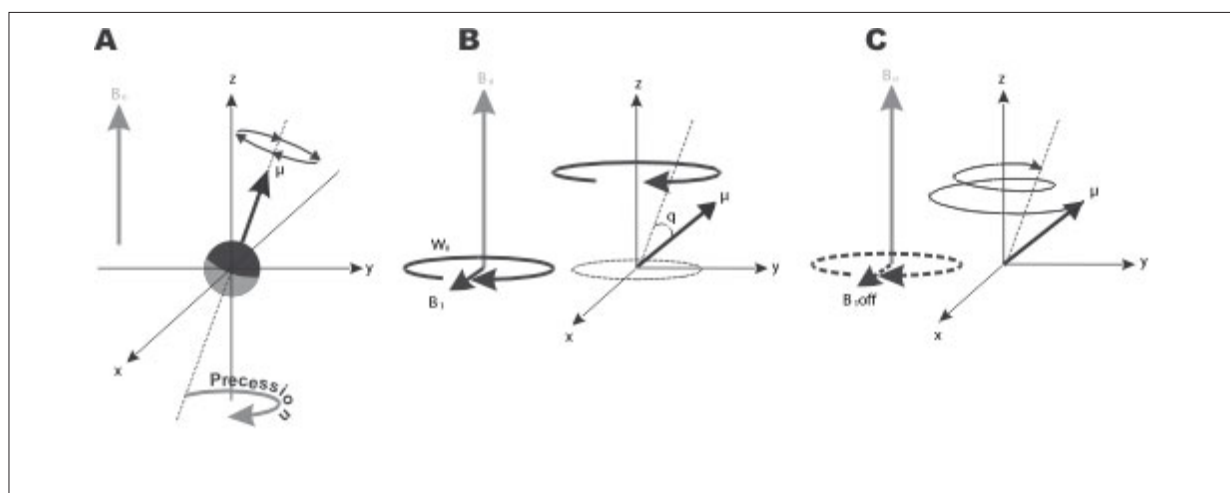


Figure 5.46: MRS principle.

A: hydrogen nucleus's precessional motion around the static field B_0 . **B:** hydrogen nucleus excitation by an oscillating field B_1 . **C:** hydrogen nucleus relaxation.

To carry out a sounding, an energizing field B_1 is created in a transmitter loop by a pulse of current oscillating at the Larmor frequency. At resonance, i.e. when the energizing frequency is the local Larmor frequency, an interaction between the nuclear angular momentum and the excitation field B_1 occurs, which deflects the magnetic moment μ from its equilibrium position (Figure 5.46B).

When the excitation field B_1 is cut off, the magnetic moment goes back to its equilibrium position, sending out a magnetic resonance signal, which is detected by a receiver loop (Figure 5.46C). This signal has an exponential envelope decaying with time (Box 5.4). Three types of information can be obtained from this relaxation signal:

- its initial amplitude is directly related to the number of hydrogen nuclei and therefore to the quantity of groundwater;

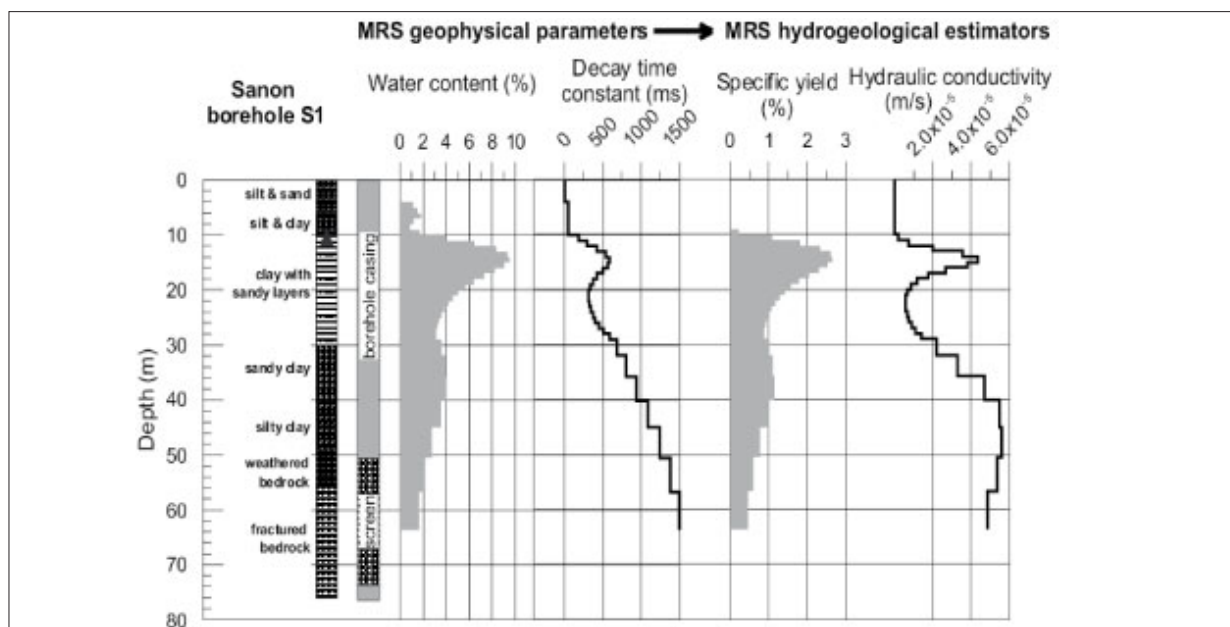


Figure 5.47: MRS geophysical parameters (the presented decay time is T_1 , see Box 5.4) and hydrogeological indicators, example of Sanon S1 Borehole in Burkina Faso, weathered granite (ACF, 2002).

- the time decay constant of the relaxation signal is related to the mean size of water-containing pores, and therefore to porosity;
- the phase shift between the relaxation signal and the excitation current is linked to the ground resistivity.

After interpretation of MRS field data, two geophysical parameters are obtained: the variation of the water content and of the decay time with depth (Figure 5.47). If a calibration process is performed, i.e., if some MRS measurements are performed around boreholes with known lithology and pumping-test results, two hydrogeological estimators can be obtained from the geophysical parameters: the aquifer's local storativity and transmissivity.

4.4.2 PERFORMING MEASUREMENTS

To generate the excitation pulses, an electric wire loop is placed on the ground and is energized by a pulse of alternating current at the Larmor frequency (Figure 5.48). The Larmor frequency depends on the local geomagnetic field amplitude (Box 5.4) which needs to be measured at the sounding location using a common proton magnetometer (Figure 5.49A). Figure 5.50 shows the theoretical Larmor frequency calculated for the year 2002 at sea level.

To carry out a sounding, i.e. to perform measurements at various depths, the amplitude of the excitation current is increased: the higher its amplitude (the current intensity), the greater the sounding depth. Sixteen depth steps are used in a typical sounding. When the excitation pulse is switched off, the magnetic resonance signal is recorded through the same loop for each depth step. The only commercial MRS equipment available is the Numis range (from Iris Instruments, Figure 5.50B). The model used by ACF for hydrogeological purposes is the powerful system called Numis Plus. It is driven by a laptop computer, which commands both the excitation and the recording sequences.

The Tx/Rx loop is typically square in shape (easy to set up in the field) with a side length of between 40 and 150 m. At a given location, this length determines the maximum depth of investigation, which is in fact roughly equal to the side length.

The implementation time for an MRS measurement depends mainly on the local signal to noise ratio: it was about 1 hour in Cambodia (non-consolidated sediments) and but around 20 hours in Burkina Faso (weathered granite). The total weight of the equipment used is about 350 kg, so that a vehicle is required for transportation. A team of 4 people is the most convenient, 2 unskilled workers, 1 field geologist to direct the setting up of the loop and 1 hydro-geophysicist to manage the whole MRS process. ACF's experience is that a local team can be trained to implement standard MRS independently in about 1 month.

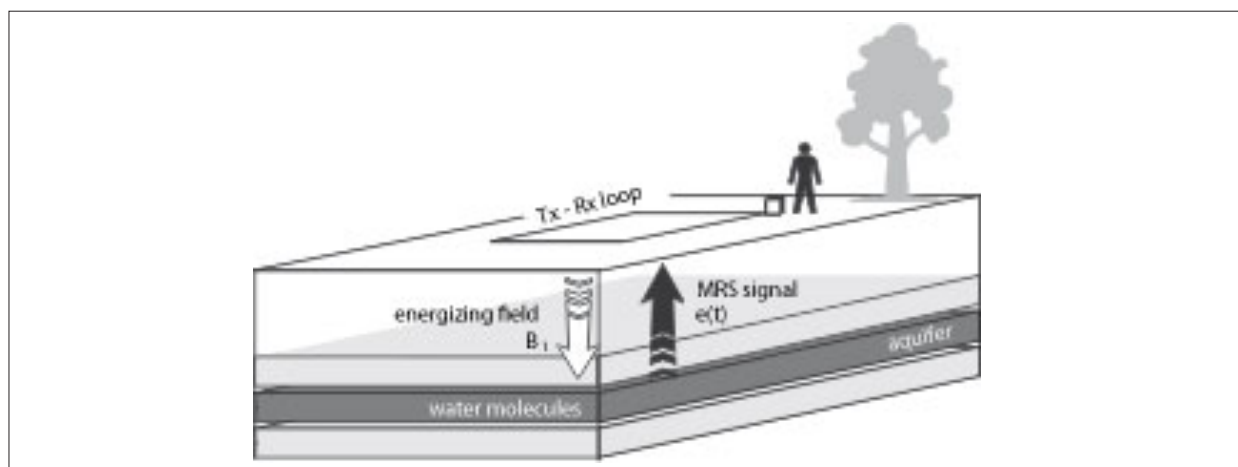


Figure 5.48: Schematic diagram of MRS measurement array.



Figure 5.49: MRS equipment.

A: geomagnetic field measurement using a proton magnetometer. B: Numis Plus equipment, ACF 2002.

4.4.3 INTERPRETATION

At the end of the data-gathering period, it takes only a few seconds to interpret the records directly on site using inversion software included in the equipment package (Samovar). The main curves obtained are:

- the sounding curve, which is the initial amplitude of the relaxation signal as a function of the pulse moment (i.e., depth step, see Box 5.4). This gives, in qualitative terms, the amount of groundwater as a function of pseudo-depth (Figure 5.51A);
- field record curves, which help to check the records' quality and the quality of the model fit;
- geophysical interpretation curves, showing the water-content and the relaxation decay time as a function of depth (Figure 5.51B).

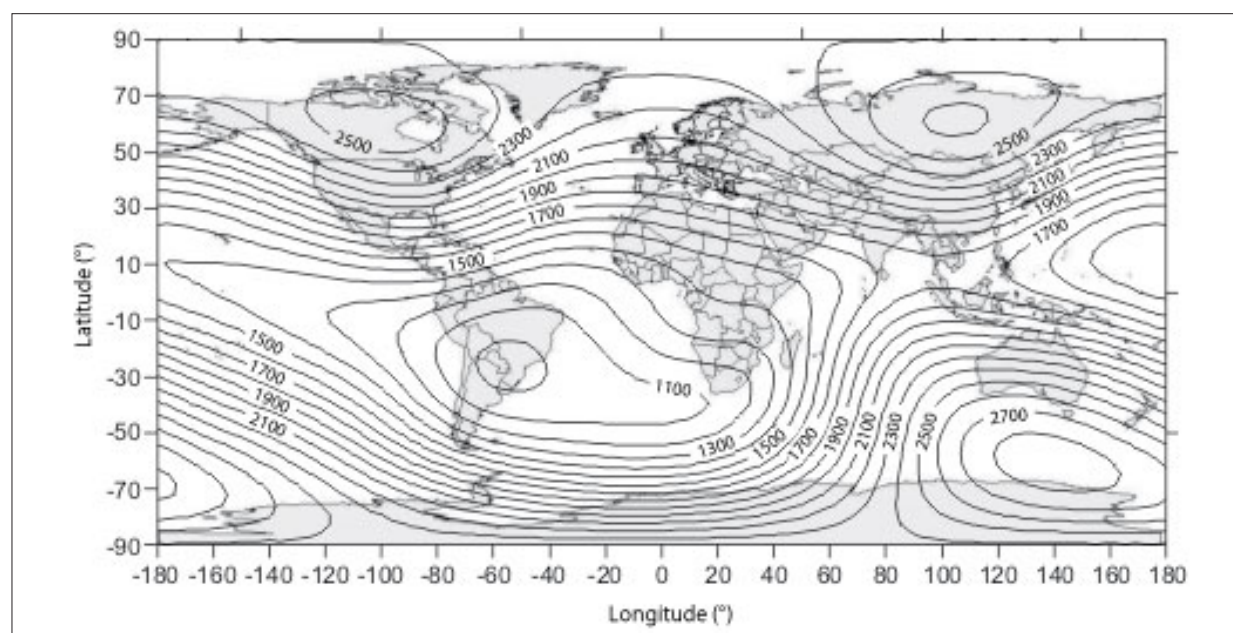


Figure 5.50: Theoretical Larmor frequency, year 2002 at sea level.

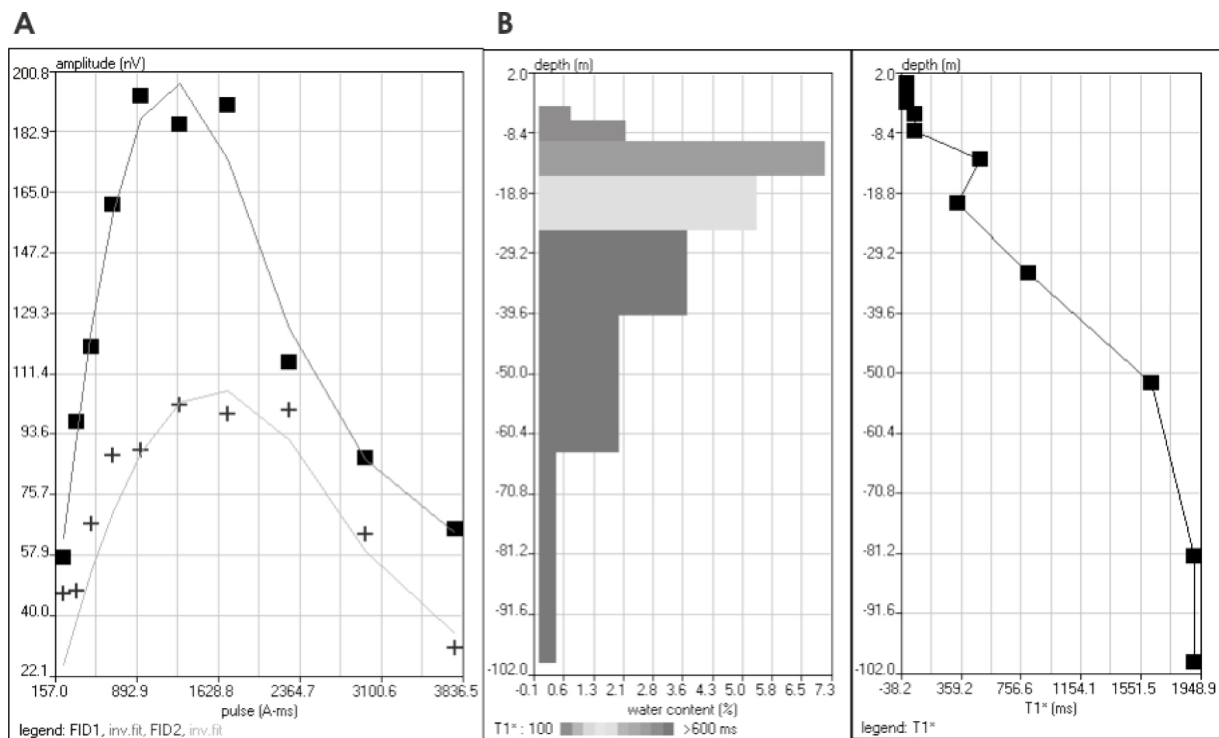


Figure 5.51: MRS curves.

A: field records (raw data). B: geophysical interpretation curves, example of Sanon S1 Borehole in Burkina Faso, weathered granite (ACF 2002).

If a calibration process is performed, hydrogeological estimators can be obtained from geophysical parameters, i.e., the local storativity and transmissivity of the aquifer (Figure 5.47).

ACF has evaluated the quality of aquifer characterisation by MRS in several geological contexts. The MRS-measured saturated-reservoir geometry, storativity and transmissivity have been compared to those obtained from boreholes and pumping tests. The main results are presented below.

Saturated-reservoir geometry

Figure 5.52 compares the one-dimensional geometry of saturated reservoirs estimated from MRS with the geometry estimated from the borehole lithology. The location of the top of the saturated reservoir is estimated by MRS with an average difference of 24%, i.e. 3.9 m, from that given by the borehole, and the bottom of aquifer is estimated with an average difference of 13%, i.e. 11.4 m. These results show that MRS gives a sufficiently accurate picture of the geometry of the saturated zone, but cannot replace genuine piezometers. In fact, MRS does not give any information about static levels, only about the depths of the saturated zones: the two are equivalent in unconfined aquifers, but are of course different in confined aquifers.

Geophysical parameters: water content and relaxation time

By comparison with the total porosity, the MRS water content is defined as $w = V_{\text{long}}/V_{\text{total}} \cdot 100$, where V_{long} is the volume of water with sufficiently long relaxation constant (T_2^* , see box 5.4), and V_{total} is the total volume of the sample. The MRS water content differs from the total porosity of saturated media because the relaxation effect can make the MRS signal too short for the equipment to detect (currently the instrumental ‘dead time’ is 40 ms). Because the relaxation time is longer for flowing water (some tens to thousands of ms) than for non-flowing water (some units to tens of ms), the MRS water content is a rough estimation of the aquifer’s kinematic porosity.

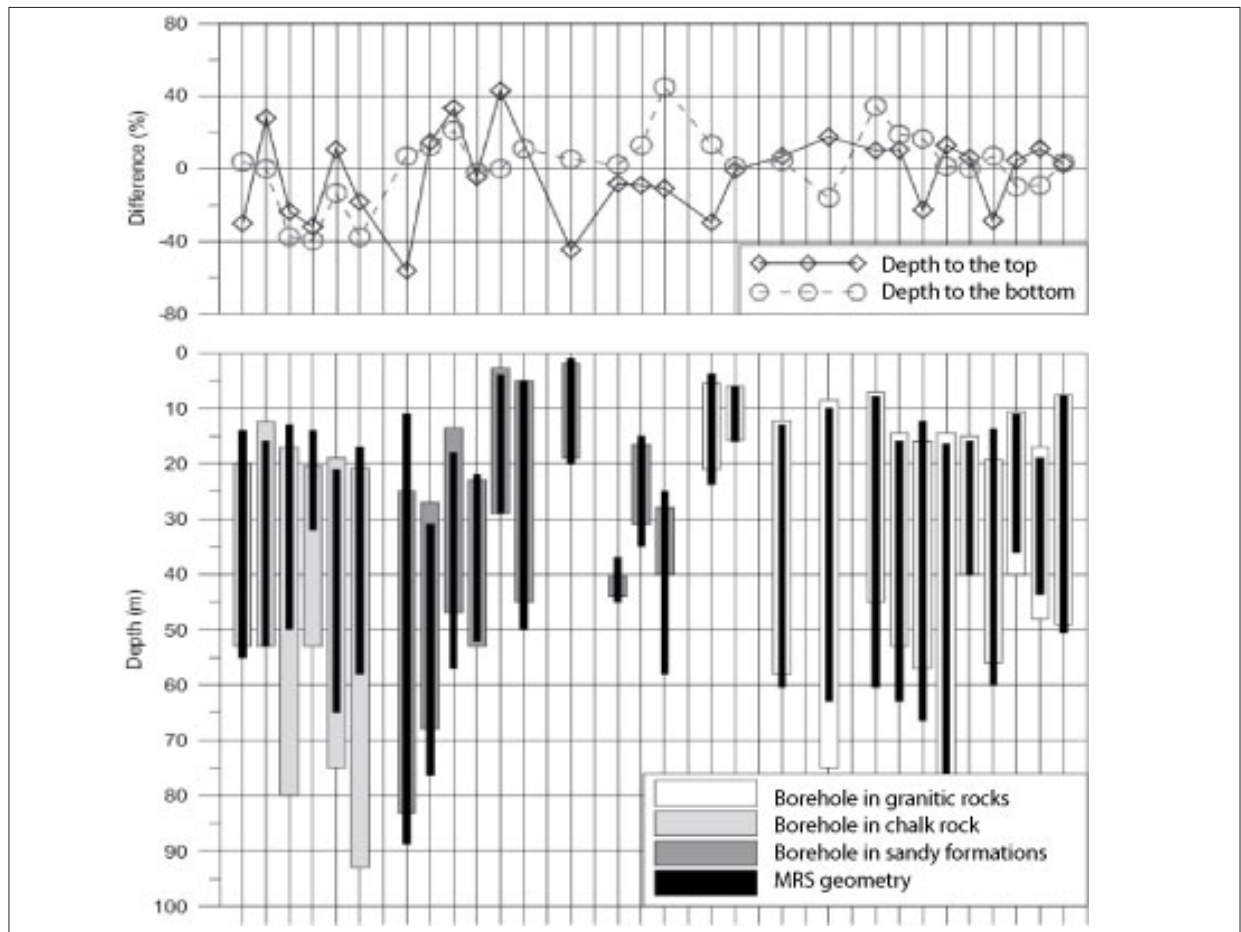


Figure 5.52: Comparison between saturated reservoir geometry as estimated from MRS and from boreholes.

Two decay-time constants can be estimated from MRS records (Box 5.4). ACF's experience is that T_1 is the more useful for hydrogeological purposes and should be estimated. T_1 is linked to the mean size of the pores containing water: the larger the mean pore size, the longer T_1 .

Hydrogeological parameters: storativity, permeability (hydraulic conductivity) and transmissivity estimators

The water content $w(z)$ and the relaxation time $T_1(z)$ derived from MRS records can be used to estimate the storativity S , the hydraulic conductivity K and the transmissivity T of aquifers:

- In confined aquifers, the storativity can be represented by the storage coefficient:

$$S_{\text{MRS}} = (w \cdot \Delta z) \cdot C_1.$$

- In unconfined aquifers, the storativity is almost exactly equal to the specific yield, represented by $n_{\text{eMRS}} = w \cdot C_2$.

- The hydraulic conductivity can be estimated as $K_{\text{MRS}} = C' \cdot w(z) \cdot T_1^2(z)$.

- The transmissivity is therefore calculated as:

$$T_{\text{MRS}} = \int_{\Delta z} K_{\text{MRS}}(z) \cdot dz$$

where w is the MRS water content, T_1 the MRS longitudinal decay constant, Δz the aquifer thickness and C_1 , C_2 and C' empirical constants that need to be determined from pumping-test data.

Some empirical constant values are proposed by ACF, but new calibrations need to be performed for new geological contexts and site locations (Table 5.IX and Figure 5.53).

Table 5.IX: Empirical constants used for MRS hydrogeological estimator calibration.

	C1	C2	C'
Weathered granite in Burkina Faso	$4.3 \cdot 10^{-2}$	0.28	$1.3 \cdot 10^{-9}$
Unconsolidated sediments (silt, sand, gravel) in France			$4.9 \cdot 10^{-9}$
Chalk in France			$3.5 \cdot 10^{-8}$

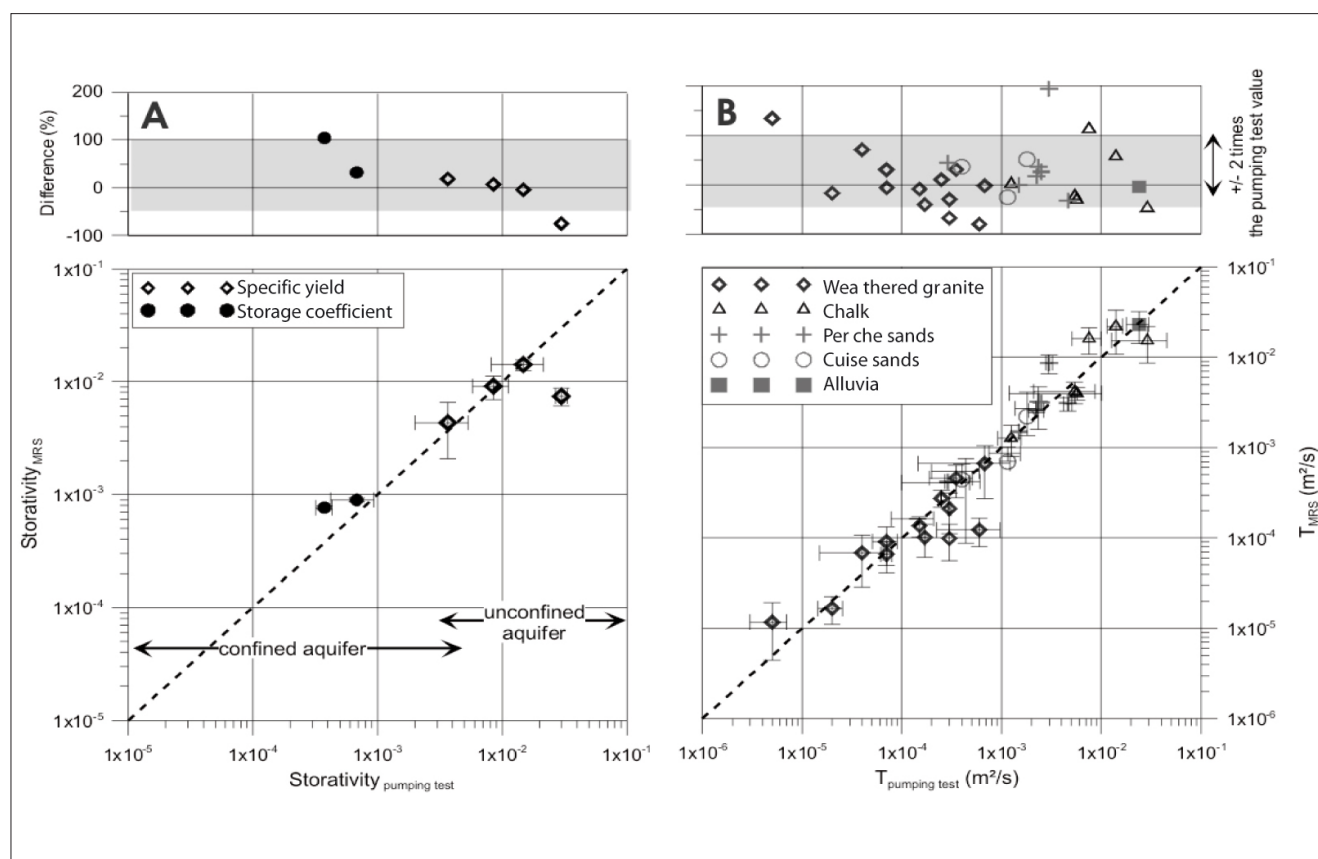


Figure 5.53: MRS hydrogeological estimators.

A: storativity estimators for weathered granite in Burkina Faso. B: transmissivity estimator.

The transmissivity estimator is fairly robust – it gives an estimate of the productivity that is equal, within a factor of 2, to the value given by the test pumping in 75% of cases. In addition, its range of transmissivity values is wide, corresponding to outputs from a few hundred litres per hour ($T = 10^{-6} \text{ m}^2/\text{s}$) to several hundred cubic metres per hour ($T = 0.1 \text{ m}^2/\text{s}$).

The storativity estimators require further confirmation (insufficient calibration data) and are more difficult to use, since one needs to know *a priori* whether the aquifer is confined or unconfined in order to use the correct estimator (specific yield or storage coefficient).

Qualitative analysis: permeability and transmissivity estimators

If the calibration process cannot be carried out, either because the geological context is obviously not one-dimensional at the observed scale, or because no pumping-test data are available, a qualitative approach using two estimators can still be used to characterize the groundwater.

A permeability estimator can be calculated as:

$$k_x(z) = \frac{w_x(z) \cdot T_{lx}^2(z)}{w_r(z) \cdot T_{lr}^2(z)}$$

A transmissivity estimator can be calculated as:

$$T_x = \frac{\int_{\Delta z} w_x(z) \cdot T_{lx}^2(z) \cdot dz}{\int_{\Delta z} w_r(z) \cdot T_{lr}^2(z) \cdot dz}$$

where x and r are two of n MRS stations ($1, x, r, n$). The reference station r can be selected so that $k_{x\max}(z) \leq 1$ or $T_{x\max} \leq 1$ for all soundings ($x = 1$ to $x = n$).

The estimators k_x (k-estimator) and T_x (T-estimator), which are normalized estimators for the permeability and the transmissivity, can be used as contrast indicators to compare several (n) soundings. An example of such a qualitative approach is presented in Chapter 5B.

Principal limiting factors in MRS

Like all geophysical methods, MRS has various limitations:

– *Equivalences*: two models are equivalent when they give the same signal. In MRS, layers 1 and 2 are equivalent if the product of saturated thickness and water concentration are similar in both. In other words, one cannot determine at the same time both the water concentration and the saturated layer thickness.

– *Resolution and probing depth*: the resolution is the capacity of the method to detect and characterise a saturated level. It is a function of various factors, including principally the intensity, the tilt angle of the geomagnetic field, the electrical resistivity of the formation and the size of the Tx-Rx loop used. Taking the example of a small reservoir, equivalent to a 1-m thick layer of pure water, i.e. a 6.5-m thick layer of fine sand with 15% kinematic porosity, or – a geologically equivalent model – a water-filled cylindrical pipe 10 m in diameter, crossing a Tx-Rx loop with 100-m sides. In the most unfavourable conditions (close to the magnetic equator and on conducting ground) this aquifer could be detected by MRS down to a depth equivalent to 40% of the diameter of the Tx-Rx loop. In the most favourable conditions (close to the magnetic poles and on insulating ground) this same aquifer could be detected down to a depth equivalent to 150% of the diameter of the Tx-Rx loop. However, detection and characterisation are not the same thing: the errors in the estimate of the water concentration and the thickness of the saturated zone can be large, even at small depths, but the error in the estimate of their product (equivalence) is less than 5% for depths equivalent to half the size of Tx-Rx loop and 15% beyond that depth. In the context of ACF programmes, this means that it is generally possible to detect saturated levels down to depths of about 100 m, and to characterise them properly down to about 70 m.

– *Sensitivity and one-dimensional nature*: the maximum volume integrated in an MRS measurement is given approximately by a circular area of diameter 1.5 times that of the Tx-Rx loop multiplied by a depth equal to the loop diameter. For a 100-m diameter loop, this volume is roughly 1.8 million m³. The fact that the signal is integrated over this volume means of course that the method cannot reveal any heterogeneity inside such a volume, nor can it detect signals generated by small numbers of hydrogen nuclei.

– *Suppression*: when the volume of water is very small compared to the volume explored by the sounding, the signal is too weak to be measured: this is known as suppression. It can occur in fractured media or in dissolution patterns, which contain only small volumes of water but which are nevertheless highly productive.

Box 5.4

Principle of MRS.

The hydrogen nucleus, i.e. a proton, possesses an angular momentum S and a magnetic moment μ . The two quantities are related through the expression $\mu = \gamma \cdot S$, where the gyromagnetic ratio γ is a constant characteristic of the hydrogen nucleus. In a homogeneous magnetic field B_0 , the angular momentum of the proton causes a precessional motion of μ around B_0 with an angular velocity known as the Larmor frequency $\omega_0 = \gamma B_0 / 2\pi$.

To meet the resonance condition, an energizing field B_1 of frequency equal to the local Larmor frequency is created in a transmitter loop by a pulse of alternating current $i(t) = I_0 \cos(\omega_0 t)$, $0 < t \leq \tau$, where I_0 and τ are respectively the pulse amplitude and duration. Therefore, an interaction between the nuclear angular momentum and the excitation field B_1 occurs, deflecting the magnetic moment μ from its equilibrium position.

When the excitation field B_1 is cut off, the magnetic moment goes back to its equilibrium position, sending out a relaxation signal, which is recorded in a receiver loop. This signal oscillates at the Larmor frequency and has an exponential envelope decaying with time, which can be approximated by (Figure 5.54A):

$$e(t, q) = E_0(q) \exp(-t/T_2^*(q)) \cos(\omega_0 t + \phi_0(q)) \quad (1)$$

where $q = I_0 \tau$ is the energising pulse parameter, ϕ_0 is the phase, T_2^* is the signal decay time (called the transverse relaxation time in the usual terminology) and $E_0(q)$ is the initial signal amplitude:

$$E_0(q) = \omega_0 M_0 \int_V B_{1\perp} \sin(1/2 \gamma_p B_{1\perp} q) w(r) dV(r) \quad (2)$$

where M_0 is the nuclear magnetisation for the protons, $B_{1\perp}$ is the primary magnetic field component (normalised to 1A) perpendicular to the static field B_0 , r is the co-ordinate vector and $w(r)$ the water content.

From equations (1) and (2) it can be understood that:

- The initial amplitude of the signal E_0 is related to the water content $w(r)$. Solving equation (2) leads to a direct estimate of the water content of the investigated volume.
- The spatial contribution of the signal is determined by q , i.e. the depth of investigation is controlled by the pulse intensity I_0 (maintaining the pulse duration constant).
- The signal decays with time at a rate governed by T_2^* . This constant is linked to the mean water-containing pore size, but it is also influenced by local inhomogeneities in the static field, often induced by the magnetic properties of rocks.

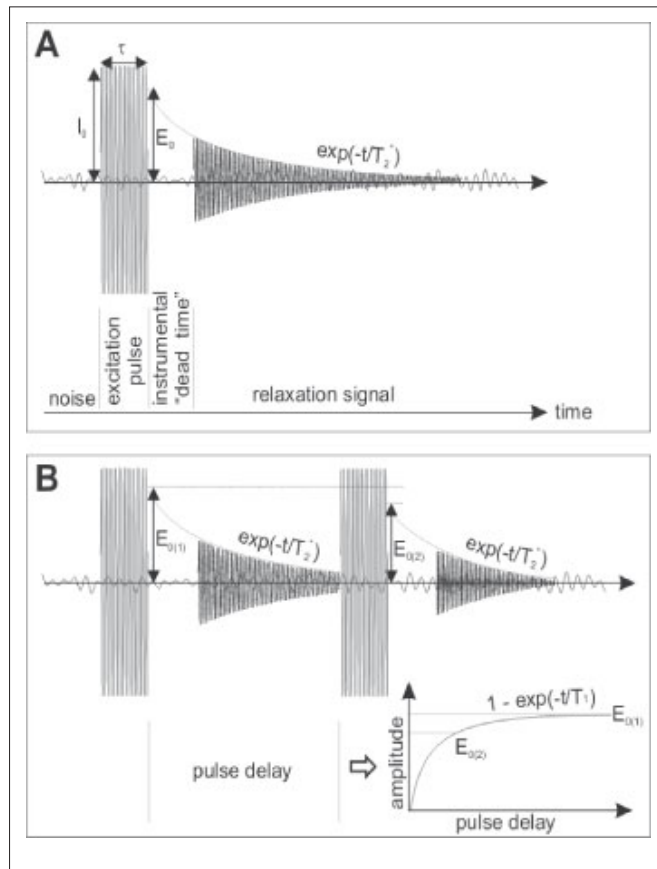
To access a more reliable parameter linked to the pore size, a modified excitation sequence using two pulses, known as saturation recovery, can be used (Figure 5.54B). It gives an estimate of T_1 , (usually called the longitudinal relaxation time), which is linked to the mean pore size of the aquifer: $T_1 = V_p / \rho \cdot S_p$, where V_p and S_p are the volume and surface area of the pore, and ρ the surface relaxivity of the rocks.

When the excitation pulse is switched off, the magnetic resonance signal is recorded through the antenna for each pulse q . Assuming a horizontal stratification, a modified form of equations (1) and (2) is used to convert the recorded quantities $E_0(q)$ and $T_2^*(q)$ into water content $w(z)$ and decay times $T_2^*(z)$ and $T_1(z)$ as a function of depth.

Figure 5.54: Time sequence of the MRS signal.

A: single-pulse sequence (estimation of T_2^*).

B: two-pulse sequence (estimation of T_1).



– *Magnetic rocks*: current MRS equipment measures a macroscopic signal, which is the sum of the signals emitted by each hydrogen nucleus. When the geomagnetic field is heterogeneous on the scale of the sounding, the hydrogen nuclei have different Larmor frequencies, and this generates phase differences between the individual signals, which can reduce the measured time constant sufficiently to prevent the measurement of the resultant signal. ACF was thus unable to measure MRS signals in very productive reservoirs in Honduras (2000), since the magnetic properties of the volcanic rocks present greatly perturbed the geomagnetic field. One should therefore estimate the feasibility of the measurements before carrying out the sounding (variations in the measured field at the surface, magnetic susceptibility of outcrops and stability of the Larmor frequency).

5 Prospecting procedures

The sequence of investigation proceeds as a dynamic process that allows the hypotheses advanced on the basis of the results obtained to be tested. It is also at this stage that deeper investigation into the water resource begins.

5.1 Exploration boreholes

An exploration borehole verifies hypotheses and furnishes information essential for tasks such as checking VES interpretations and calibrating MRS hydrogeological estimators. It also facilitates logging and pumping tests.

Every result is used as a source of information: a dry borehole must be “used” in the same way as one that produces water, since the reasoning that led to sinking the borehole must be reviewed or developed. The feedback between the previously-envisaged presence of water and the results obtained increases the accuracy of the overall picture.

5.2 Assessment of the resource

Pumping tests help to estimate the potential of the boreholes (well tests), and the hydraulic characteristics of the aquifers (aquifer tests, see Chapter 6). At this stage it becomes possible to draw a map of zones of potential groundwater based on the parameters (transmissivity and storativity) obtained from both the pumping tests and the MRS.

Piezometric monitoring must also be implemented at this stage. It acts as a check on the development of resources that are the object of new exploitation, and gives an estimate of seasonal variations. A sketch of the piezometric map can be drawn in order to visualise groundwater flows. This monitoring can be carried out at the same time as measuring water quality (see Chapter 4).

The groundwater chemistry survey can also reveal a great deal about the water-flow system, as well as the relationships between different aquifers, recharge mechanisms and links between groundwater and surface-water bodies (see Chapter 4).

Finally, bases for the construction of a hydrological balance can also be created: rainfall and piezometric monitoring measurements, surface flow-rate studies, and determination of the survey catchment basin should be carried out.

5.3 Storage of results and analytical tools

To avoid significant information loss when key staff change, and also to capitalise on the experience gained by staff and share information in the most effective way, it is essential to archive results in a relatively standardised manner. Furthermore, proper storage ensures that information is managed with more care.

ACF has developed tools for managing and analysing hydrogeological information. This process is in constant evolution and the main tools shown in Table 5.X are only a guide.

Table 5.X: Tools for managing and analysing hydrogeological information.

Existing documentation	Technical libraries of reference material on various subjects in Paris and Madrid Reports and programme descriptions archived at various sites ACF technical experience collected in the form of manuals and specific reports
Geological maps	Geological map base available at the Paris office of ACF, showing most countries where the agency works, generally at small scales
Well-test and aquifer-test Interpretation software	WHI AquiferTest_Pro software tested and recommended (from Waterloo Hydrogeologic Inc.)
VES and RP Package	Developed by ACF, contains land measurements, maps of presentation of results and sets of Cagniard nomograms IPI2Win freeware included for VES interpretation and results presentation
Interpretation software for geophysical surveys	IPI2Win freeware for interpretation and visualisation of VES recommended QWSEL® software also available for VES interpretation X2IPI freeware recommended for ERI protocol construction and ERI data processing RES2DINV software recommended for ERI data interpretation
Database and geographical information system	Models carried out using Excel® software (limited amounts of data, good system user-friendliness) Mapinfo® software for geographical analyses (thematic map)
Equipment kit	Set of materials and equipment tested and developed by ACF: Ωmega resistivity-meter, electrical-probe logging, drilling rigs, pumping-test kit (see Chapter 6)

5.4 Prospecting procedures

To decide the best prospecting procedure to use, all the elements presented in this chapter should be considered, including: what are the hydrogeological questions? which methods and tools will best answer them? how to compare the usefulness of the various possible techniques?

Each context is of course unique, but, using the example of an ACF project in Mozambique, this section illustrates how to set about finding the best procedure to apply when prospecting for underground water.

5.4.1 HYDROGEOLOGICAL QUESTIONS

ACF ran a programme to set up a drinking water supply in the Sofala province of Mozambique from 1994 to 2001. By December 2000, 178 water points had been constructed (143 boreholes and 35 wells) with an average success rate of 44.5%. The high number of negative boreholes includes constructions which were dry or whose output was insufficient ($Q < 500$ l/h), but also boreholes whose water was too mineralised for human consumption (12% of the boreholes).

A study was performed from October to December 2000, with the aim of improving knowledge of the aquifers, in order develop a prospecting method that might increase the borehole success rate.

The general context of the study: collating available information, and additional investigations

The districts of Caia and Chemba are in the northern part of Sofala province, on the west bank of the Zambezi River (Figure 5.55).

The region is in the sedimentary basin ‘Nord of Save’. The ‘Sena’ geological formations date from the upper Cretaceous and are composed of continental sandstone up to 2 500m thick (information from Mozambique’s National Directorate for Water Affairs, Figure 5.56). Sandstone samples prepared by ACF in the form of thin slices reveal a siliceous formation containing calcic cement, with little or no porosity. This observation is confirmed by the high failure rate of boreholes sunk in the sandstone.

Figure 5.55: Location of the study zone in Mozambique.

The region is traversed by temporary rivers flowing from west/south-west to east/north-east; its eastern limit is the Zambezi river. Analysis of the information from ACF's borehole drilling programs shows that the sediments deposited by these surface waters constitute reservoirs of highly variable quality. The Zambezi's alluvia are clayey in some places, but generally provide lightly mineralised water; the sediments deposited by the temporary rivers are coarser but can contain highly mineral-rich water in the north, while in the south they are less permeable and the water is often less highly mineralised.

Samples of the water in the region (rainwater, surface water and groundwater) were collected by ACF, and analyses of the major ions were performed. The interpretation of these analyses shows that the high mineral levels in the groundwater (electrical conductivity in the range 1 500 to 9 000 $\mu\text{S}/\text{cm}$) are due to the dissolution by the infiltration water of recently formed minerals present in the superficial horizons. This interpretation is confirmed by the observation *in situ* of carbonated precipitates in superficial cracks in the sandstone, and by the presence of halite in the dead zones of the temporary rivers where water loss by evaporation is intense.

The hydrological balances measured by ACF, using data collected from local partners, are also in agreement with the observations above. They indicate that the northern zone is subject to high rates of evapotranspiration, which explains why the underground water there is highly mineralised (Table 5.XI).

It was not possible to measure the surface run-off, and therefore the fraction of effective infiltration in the total run-off could not be estimated. However, it seems, from the (admittedly incomplete) information available, that the use of low-output (handpump) systems does not significantly affect the replenishment of the water resources, since no borehole ran dry and no excessive drop in the static levels was reported in the period from 1994 to 2000.

According to the results of an exhaustive study carried out by ACF in 1999 and 2000, the districts of Caia and Chemba have populations of 157 000 and 52 000 respectively, of whom 27% and 15% respectively have access to drinking water. In this context, access to drinking water is defined as the possibility for a family to take its water from a properly constructed water point, delivering drinkable-quality water, situated less than 1 hour's walk away and serving a maximum of 300 users (i.e. an average of 20 litres of water available per person per day for a well or borehole equipped with a handpump).

Access to water is of crucial importance to these rural populations, and the only resource used all year round is groundwater.

Figure 5.56: Geological sketch of the districts of Caia and Chemba (Mozambique).

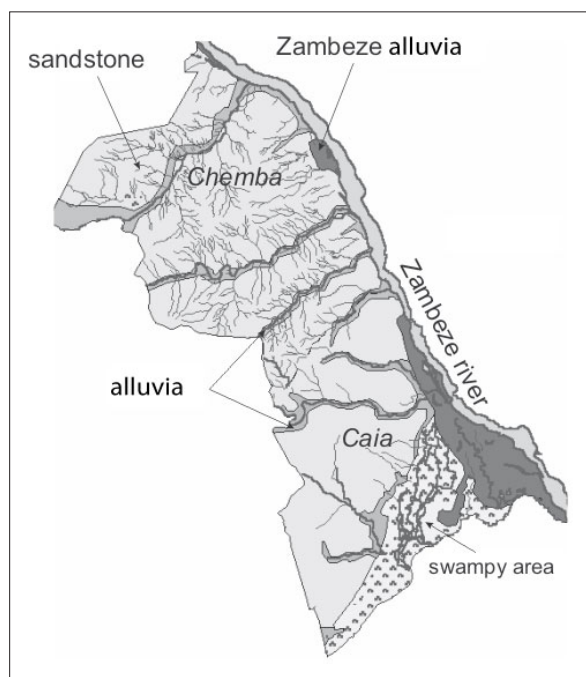
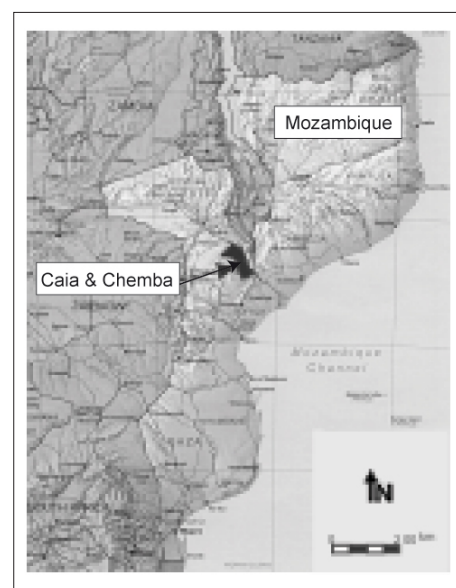


Table 5.XI: Electrical conductivity of water and hydrological data (Thornwaite's formula, steps of one month, period: 1946 to 1955).

Zone	Average electrical conductivity of the groundwater ($\mu\text{S/cm}$)	Average yearly rainfall (mm)	Average yearly temperature ($^{\circ}\text{C}$)	Total yearly run-off (mm)
North (Chemba)	3700	728	34	56
South (Caia)	1500	954	32.3	200

Hydrogeological questions facing the prospector

The aquifers required for the establishment of new water facilities are those permeable layers of alluvia and fissured and weathered zone of sandstone where water mineralisation levels are low enough to provide water fit for human consumption.

In searching for water, the questions facing the prospector were the following:

- which zones in a radius of about 1 km around the villages contain aquifers?
- in these zones, what is the reservoir geometry (where to drill the boreholes)?
- what are its hydraulic characteristics (what will the borehole's output be)?
- what is the electrical conductivity of the water (is it drinkable)?

The standard methods can not satisfactorily answer these questions – the average borehole success rate is low (<50%), even reaching zero for certain zones (sandstone plateaux). Geophysical methods may sometimes prove better able to clarify the situation. The first step is to select which geophysical methods to use. The next is to take the results of experiments in the field to assess the technical and economic utility of the methods, used on their own or in combination with others. Finally a tested hydro-geophysical procedure can be chosen.

5.4.2 THE CHOICE OF GEOPHYSICAL METHOD(S)

This choice is guided *a priori* by the nature of the reservoirs being sought, the required precision and the size of the zones to be prospected (see Section 4).

In the region considered, highly variable water mineralisation levels exist, and so resistivity is the physical parameter most likely to furnish the relevant information, because it is highly sensitive to the electrical conductivity of the water.

The high probability of encountering clayey layers inside the alluvia, and the *a priori* knowledge of the existence of highly mineralised groundwater indicate that the formations studied are good electrical conductors. In these conditions, electromagnetic resistivity measurements are preferable to direct current methods.

The precision must be sufficient to enable a borehole to be sunk with certainty on the extension of the anomaly. TDEM soundings in the light-equipment configuration can measure to depths of around 100 m in this conducting environment. On the other hand, they integrate signal over a large volume (depending on the size of loop used), and in the presence of complex structures, they cannot give a sufficiently precise picture of the lateral variations.

The only method that can routinely be used to perform the two-dimensional resistivity measurements necessary to describe complex geological structures is electrical resistivity imaging (ERI).

MRS can provide information on the presence of groundwater and on the characteristics of the reservoirs.

A TDEM sounding takes only 30 to 45 minutes to perform (including setting up the equipment and taking the readings), compared with 2 or 3 hours for a conventional ERI and 2 to 20 hours for an MRS.

For the specific context of Mozambique, the methods ultimately chosen were:

- TDEM soundings to map the resistivity as a function of depth and indicate the likely aquifer zones (Figure 5.57);



Figure 5.57: Setting up a TDEM measurement (Mozambique, 2000).



Figure 5.58: Traditional well in Nhamago (Mozambique, 2000).

- ERI to map the resistivity in two dimensions and thus determine the geometry of the potential reservoirs and the electrical conductivity of the water contained;
- MRS to confirm the existence of the reservoirs and determine their hydraulic characteristics.

5.4.3 APPLICATIONS OF THE GEOPHYSICAL METHODS

Searching for potential sites: TDEM soundings

After some field reconnaissance, the geophysical prospecting zones were chosen as near as possible to the villages that were considered as top priority in terms of the population's water requirements. Aerial photos were of limited use, since it was often impossible to identify any location on the images! (The area had been a war zone for a long time and none of the infrastructure visible on the photos, including the roads, was identifiable on the ground any longer.) So it was mainly the geomorphology and discussions with the villagers that informed the choice of potential aquifer zones.

When the zones to be prospected are extensive, it is advisable to perform some quick measurements in order to class the various sites according to their aquifer-bearing potential.

In the region concerned, TDEM soundings were well adapted to this purpose. They are easy to set up and their results are sufficient to indicate the zones on which more detailed (but more difficult) research should be concentrated.

Figure 5.59 shows the example of the village of Nhamago situated in a uniform sandstone environment. A calibration sounding was first performed around the traditional well used by the local people (shown in Figure 5.58 and marked as '*Nm well*' in Figure 5.59), and the calculated resistivity values were compared to direct measurements and observations in the well:

- The resistivity of the sandy aquifers was ill-defined because the TDEM method is insensitive to the first few metres in depth. These sands are nevertheless noted as being much more resistive than the material lying underneath them.
- The electrical conductivity of the water (at 25°C) was 770 $\mu\text{S}/\text{cm}$, or 13 $\Omega\cdot\text{m}$.
- A clayey layer identified at the bottom of the well showed a calculated resistivity of less than 10 $\Omega\cdot\text{m}$.
- The sandstone substratum crops out near the well. It is also highly conducting and presents a calculated resistivity of less than 15 $\Omega\cdot\text{m}$.

The geophysical target zone was thus identified as a resistive mass resting upon a conducting substratum.

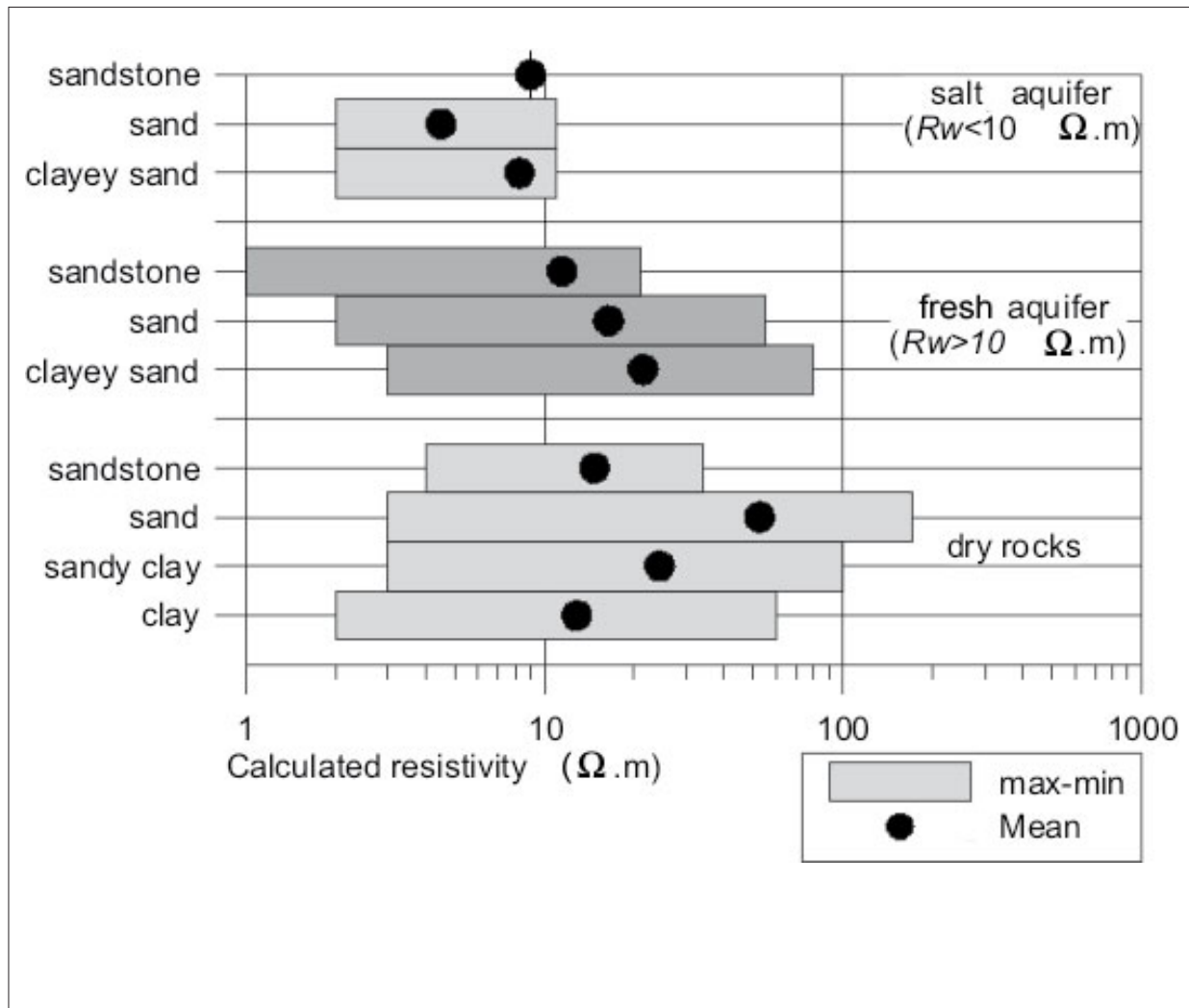


Figure 5.59: TDEM soundings, village of Nhamago (Mozambique, 2000).
 R_w is the resistivity of the water.

Eleven TDEM soundings were performed over two days on suspected anomalies (determined by the geomorphology). The results of the interpretations are presented in Figure 5.59. Two groups of curves can be distinguished: the first group, for which the calculated resistivities are under $10 \Omega.m$, is interpreted as clayey sandstone; the second represents the more resistive surface formation lying above the conducting substratum. The most favourable sites are those whose resistive layers are the most clearly evidenced, in terms both of their thickness and their calculated resistivity value (e.g. Nm3 and Nm4).

Revealing complex structures: ERI

Even if the calculated resistivity values are difficult to correlate with rock types, it may still be possible to use the resistivity contrasts to show up geological structures.

Figure 5.60 presents the calculated resistivity values obtained from 43 soundings performed in Mozambique. The dispersion of the values around the mean and the overlap of the different ranges (dry rocks, freshwater aquifers and saltwater aquifers) mean that a resistivity value cannot solely be associated with a particular type of rock. The presence of clay and contrasted mineralised water explains this phenomenon.

In these circumstances, two-dimensional measurements can furnish additional information through the qualitative interpretation of the variations in resistivity.

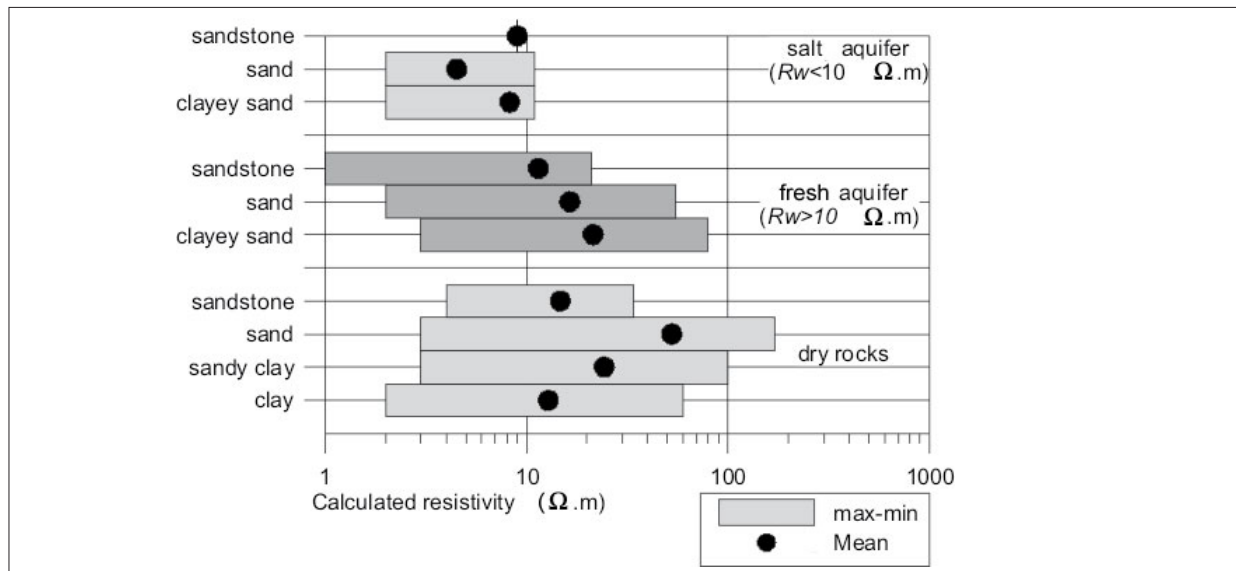


Figure 5.60: Calculated resistivity as a function of rock type, population of 43 measurements. R_w is the resistivity of the water (Mozambique, 2000).

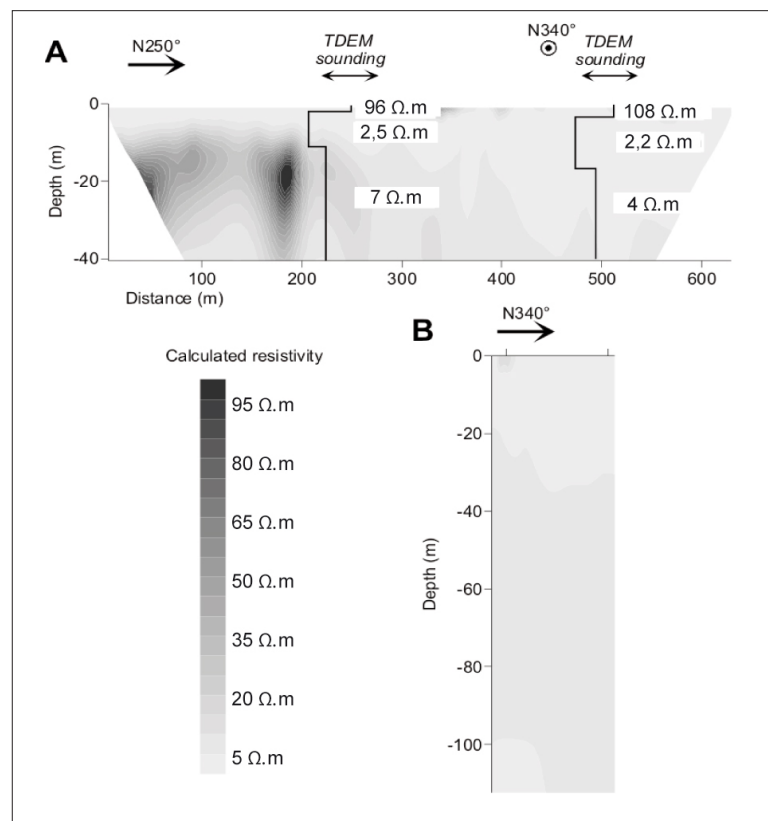
Figure 5.61 shows the resistivity measurements performed in the village of Chivulivuli, situated in the north of the zone, at the limit between a sandstone zone and the alluvial capping layer from the Zambezi river 3.5 km away. The set of TDEM soundings carried out for the preliminary study did not exhibit any highly-marked contrast — the results of two typical soundings are presented in graph A. However, an ERI was performed to see whether this weak contrast might nevertheless prove significant.

Resistivity variations (contrasts) are clearly revealed by the two-dimensional measurement; they suggest the following structures (Figure 5.61, graph A):

- The shallow, highly conducting (1 to 5 $\Omega.m$) layer present across the whole section clearly indicates a clayey stratum.

- In the western part of the ERI, the low resistivities continue at depth and may be indicative of a clay-silt medium.

- Finally, the deeper resistive zone to the east of the ERI (30 to 80 $\Omega.m$) is certainly more sandy.



**Figure 5.61: TDEM soundings and ERI, Chivulivuli site (Mozambique, 2000).
A: section of the calculated Wenner alpha resistivities (RMS = 2.9%).
B: perpendicular cross-section of the calculated pole-pole resistivities (RMS = 2.3%, inversion performed with the RES2DINV programme).**

Table 5.XII: Boreholes and MRS, Chivulivuli site (Mozambique, 2000).

Borehole	Total depth (m)	Screen position (m)	Static level (m)	Air-lift discharge (l/h)	MRS water concentration (%)	Time constant T_2^* (ms)	Water electrical conductivity ($\mu\text{S}/\text{cm}$)	Aquifer resistivity ($\Omega\cdot\text{m}$)
8/FCH/98	28.5	22-27	2.2	1000	17	230	650	40
2/GCH/00	36	4-9	2.2	1200	8	200	33000	2
9/FCH/98	10.5	4-9	4.5	1000	3	180	11000	2
1/GCH/00	37	3-8	4.5	1200	3	180	10000	2

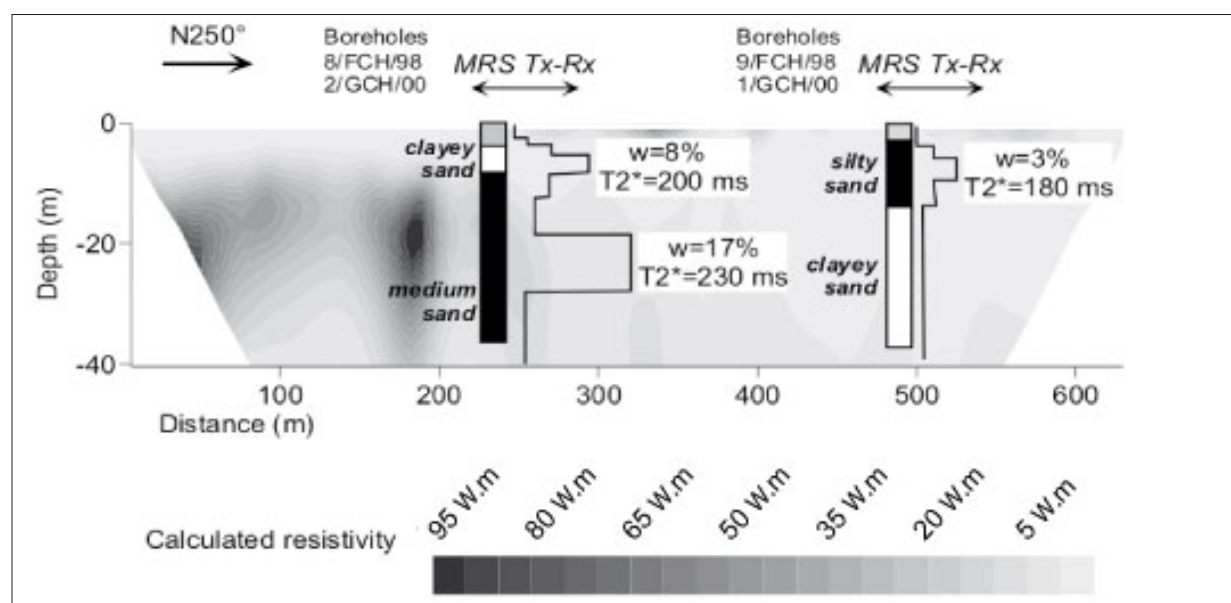
A second measurement, perpendicular to the first and centred on a point 470 m from the first ERI measurement was performed using a pole-pole apparatus, giving a greater penetration depth (all other parameters remaining constant – graph B). The resulting image is much more uniform, confirming that the structures are two-dimensional (east-west) and that the domain of validity of the method is respected.

In conditions such as these, ERI thus yields additional information about the existing structures, and enables the hydrogeologist to envisage the geometry of the potential reservoirs. But the existence of underground water remains uncertain because resistivity values alone do not unambiguously indicate the presence of water.

Confirmation of the presence of underground water: MRS

On this same site at Chivulivuli, MRS tests were carried out on the two formations revealed by the calculated cross-sectional resistivity. The interpretation of these tests was checked by drilling two pairs of boreholes. The results are presented in Table 5.XII and Figure 5.62.

MRS reveals the presence of a reservoir over the whole profile, between 2 and 10 m in depth. The very low resistivity of this horizon (2 $\Omega\cdot\text{m}$) clearly indicates highly mineralised water (and not a clayey layer as one might initially imagine from the ERI results). The three boreholes which exploit this aquifer confirm these findings: the reservoir exists (air-lift yield of about 1 m^3/h) and the water is highly mineralised (10 000 to 33 000 $\mu\text{S}/\text{cm}$).

**Figure 5.62: ERI and MRS, Chivulivuli Site (Mozambique, 2000).**

W is the water fraction and T_2^* the measured decay constant.

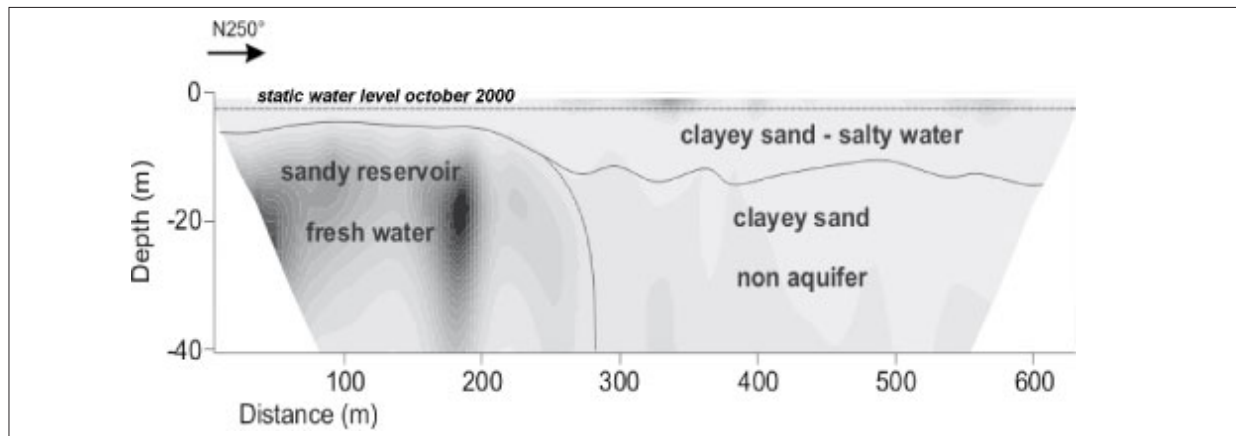


Figure 5.63: Representation of the Chivulivuli aquifer (Mozambique, 2000).

A second reservoir is revealed by MRS to the east of the profile. The much higher water concentration, for a slightly longer signal decay time, indicates a more productive formation than the surface reservoir. The resistivity of this volume is much higher and one may expect it to contain moderately mineralised water.

Two boreholes were drilled to confirm these findings, but only one could be used because the other one, situated to the west, collapsed and so was not finished (1/GCH/00). The remaining borehole, which exploited the formation to the east of the profile, confirmed the geophysical information: the aquifer exists and delivers water that is only lightly mineralised for this region (650 $\mu\text{S}/\text{cm}$).

Aquifer characterisation by ERI+MRS

Using the complementary information given by MRS and the ERI, it is possible to propose a hydrogeological model for the site (Figure 5.63). This model is qualitative rather than quantitative, because the number of measurements is not sufficient to enable a calibration of the MRS hydrogeological estimators. Examples of calibrations are given in Chapter 5B.

5.4.4 PROPOSING A HYDRO-GEOPHYSICAL PROCEDURE

Geophysical tools can be useful for two reasons:

- They provide information which enable positive boreholes to be drilled in failure zones, thereby meeting the water requirements of the population.
- They can save money for the programme, by reducing the number of salty, dry or low-output boreholes drilled.

The overall success rate

ACF built 178 water access points between 1994 and 2000. The negative boreholes have only been recorded systematically since 1997. Since this date, the annual averages are:

- 11.3 negative boreholes and 9 positives (55.6% failure rate);
- 7.3 dry boreholes (35.8% of the negative boreholes);
- 2.5 salty boreholes (12.3% of the negatives);
- 1.5 boreholes abandoned because of drilling problems (7.4% of the negatives).

In 1999, the introduction of one-dimensional geophysical techniques (Schlumberger VES) increased the borehole success-rate from 29 to 54% in similar geographical zones. This improvement is essentially due to a reduction of 16% in the number of salty boreholes, whereas the number of dry boreholes scarcely diminished. There are thus limits to the usefulness of resistivity measurements in this context: the resistivity values enable identification of reservoirs containing highly mineralised water (low resistivity), but do not signal the simple presence of water.

A geophysical programme using new methods was therefore organised from October to December 2000. The study-sites selected (4 sites with a total of 15 boreholes) corresponded to zones in which all the boreholes previously drilled had been negative. These were thus the most difficult sites encountered since the beginning of the project in 1994. The combined use of the TDEM, ERI and MRS methods yielded a success-rate of 66%.

Economic analysis of geophysical studies

Table 5.XIII presents the cost of geophysical studies per site in Euros. The calculation is performed using real ACF costs in 2000, on the following basis:

- 4x4 vehicle, used partly for geophysical studies, bought new, depreciation to 0 over 5 years;
- geophysics equipment bought new, depreciation to 0 over 5 years (except the VES equipment: over 3 years);
- team of technicians and full-time geophysics workers;
- 9 months work in the year, with 10 days per site, giving 27 sites per year.

Table 5.XIII: Cost of geophysical studies in Euros (ACF, Mozambique, 2000).

Method	National staff		International staff		Geophysical equipment				Vehicle		Number of studies per year	Admin. cost %	Cost per site (€)
	number	cost/study	number	cost-study	purchase cost	amortisation (year)	misc. cost/study	cost/study	number	cost/study			
VES	4	327	1	311	4 200	3	10	56	1	218	27	48	649
ERI	5	373	1	311	30 000	5	10	227	1	218	27	65	883
ERI + MRS	8	691	1	311	141 000	5	20	1 053	1	218	27	157	2 119
ERI + MRS + TDEM	10	840	1	700	183 510	5	30	1 373	1	327	27	203	2 743

The cost of a negative borehole calculated on the same basis is 3 655 Euros, and the cost of a positive borehole is 4 483 Euros. Using these costs, one can determine the domain of economic utility of the various methods or combinations of methods as a function of the improvement in the borehole success-rate they produce (Figure 5.64).

On the scale of ACF programmes, and for the geophysical campaign in 2000, it can be noted that:

- the use of the VES method, which increased the borehole success-rate by 24% on its introduction 1999, is economically viable whatever the borehole failure-rate;
- the combined use of the 3 methods (TDEM, ERI and MRS) is justified financially on difficult sites (success-rate below 40%) if the success-rate is doubled, which seems realistic.

Recommendation

The decision concerning the type of geophysical procedure to use as part of a hydrogeological study (Table 5.XIV) is based on technical evaluations and economic analysis.

Technically, the merits of the combined use of several geophysical methods in this region are illustrated by the examples cited above:

- TDEM soundings enable rapid prospecting over vast zones judged promising in the land survey (photo-interpretation, geomorphology and geology). They indicate the sites on which to perform ERI.
- Two-dimensional resistivity measurements are used to trace underground structures and reveal the potentially aquifer-bearing zones on which MRS can be performed.
- MRS confirms the presence of groundwater and furnishes information on the hydraulic parameters of the reservoir.
- Finally, the interpretation of the MRS parameters combined with the electrical resistivity of the aquifers provides an estimate of the electrical conductivity of the water.

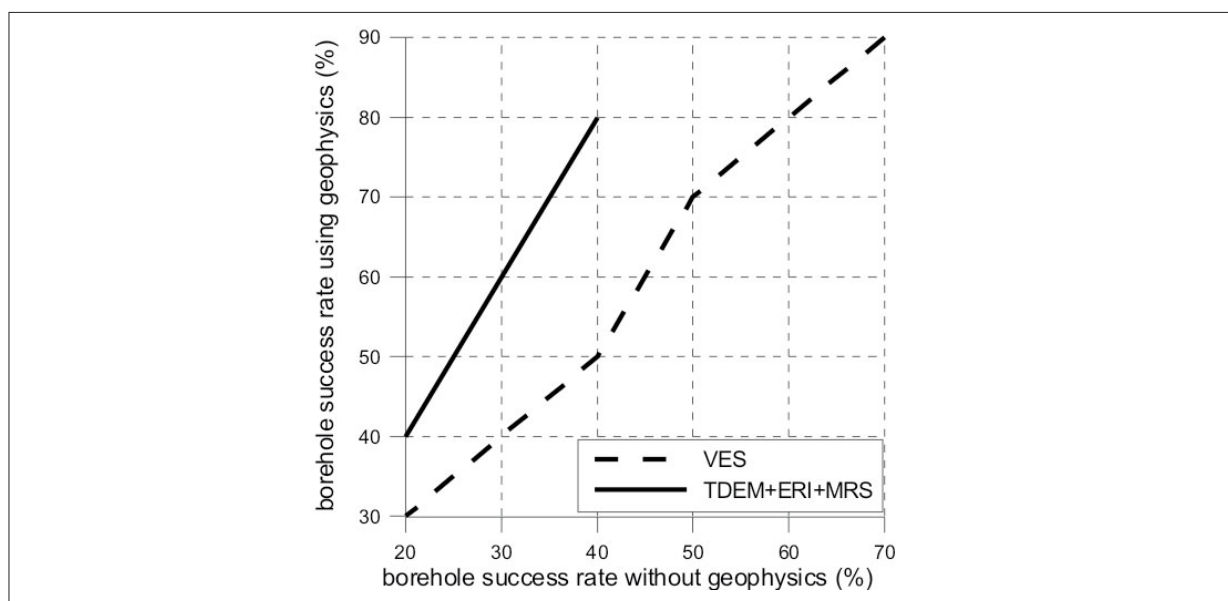


Figure 5.64: Economically favourable domain of geophysical methods (ACF, Mozambique, 2000).

Table 5.XIV: Hydro-geophysical procedure (Mozambique, 2000).

Phase of study	Methods and tools	Objective
Preliminary study	Photo-interpretation (when zone location is possible)	Identification of the geological domains (sandstone/alluvia) Identification of lineaments in the sandstone zones
Field reconnaissance	Geological & geomorphological observations Visit existing water access structures	Confirmation of the local geology and of the hydrogeological targets
Geophysical study	TDEM Soundings ERI MRS	Identification of potential aquifers Definition of the potential reservoir's geometry and the water mineralisation level Confirmation of aquifer's existence and hydraulic characterisation
Complementary study	Reconnaissance borehole Water analysis MRS Pumping test Piezometric measures	Elucidation of the groundwater dynamics Reservoir and groundwater characterisation Evaluation of the borehole's impact on the water resources

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To illustrate the way hydro-geophysical procedures are carried out by ACF when investigating groundwater, three examples are presented that correspond to some of the main geological contexts: recent unconsolidated sediments in Cambodia, crystalline bedrock in Burkina Faso and karstic limestone in France as an example of a highly heterogeneous context.

To improve readability, the scientific references used in Chapter 5B are not noted, but they can be found in the main reference list.

1 Implementing boreholes in unconsolidated sediments

Unconsolidated sediments are a common geological context for ACF programmes. They are of great potential for supplying water for many people because they are widely represented all over the world, and their development is usually easy and accessible to users. However, these groundwater resources can also be difficult to develop because of their clay content, and because they are commonly vulnerable to surface pollution and sometimes subject to excessive mineralisation. To improve its procedures for studying these resources, ACF conducted a test project in Cambodia with the support of a French research institute, IRD (*Institut de Recherche pour le Développement*), and a geophysical equipment company, Iris Instruments.

1.1 Background

ACF started its programme in Siem Reap province of Cambodia in 1992 (Figure 5.65). Its objective was to support the people who had fled into Thailand to return and resettle in their previous home area. By the end of 1998, 900 boreholes had been drilled in rural areas to supply the population

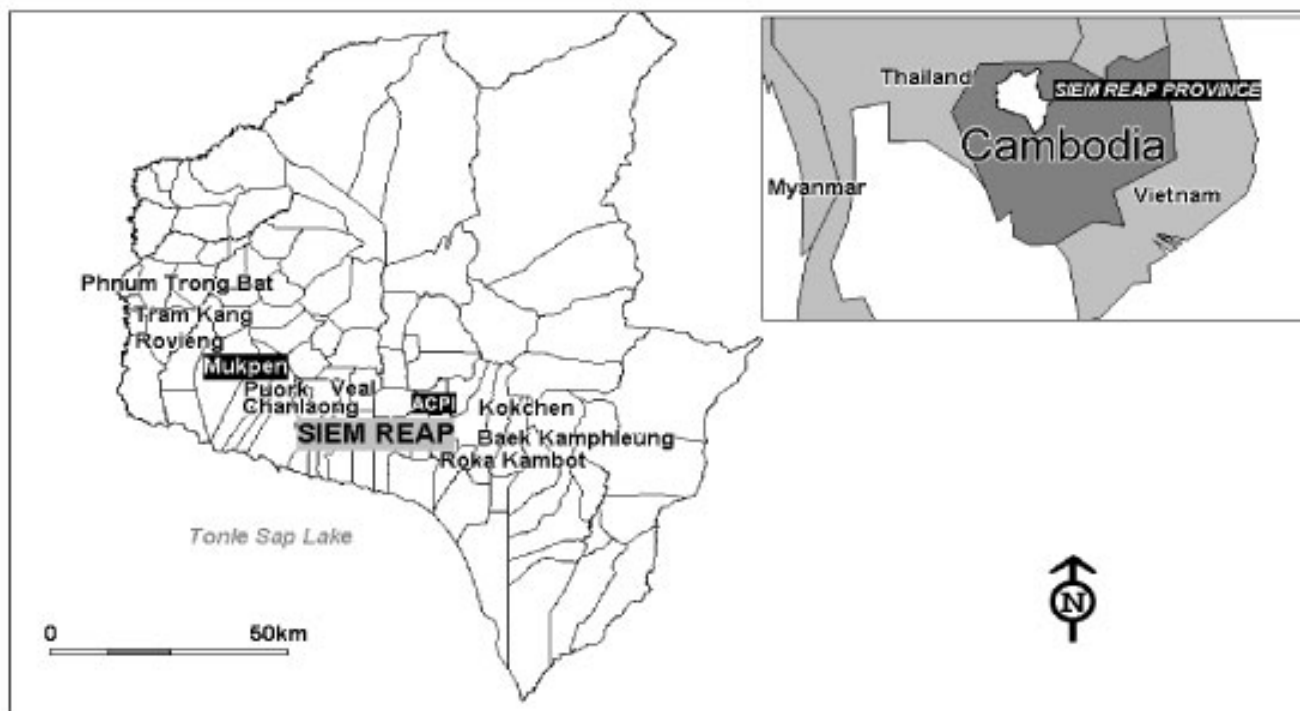


Figure 5.65: Siem Reap Province and main investigated site.

with drinking water. Of these 900 boreholes, 180 were unsuccessful, most of these in particular areas, where the failure rate reached more than 70%.

The province is highly affected by human activities (paddy fields, irrigation infrastructure etc.) and the traditional hydrogeological tools of photo interpretation or field cartography are not useful to site boreholes. Furthermore, the cuttings analysis of the drilled boreholes shows high lithological heterogeneity at the provincial scale (kilometres range) or at the village scale (hundreds-of-metres range). Geophysics is a good technique in this kind of context and several different tools were selected to survey the province for the following reasons:

- 1) The rock resistivity provides geological and groundwater-quality information in many instances. An electromagnetic method, such as TDEM sounding (time-domain electromagnetic) is more efficient in conductive environments than common VES (vertical electrical sounding) because of its high sensitivity to conductive layers.

- 2) The use of one-dimensional measurement is unsuitable in complex geological areas where two-dimensional measurement such as ERI (electrical resistivity imagery) is more effective.

- 3) The resistivity of an aquifer is related to the EC (electrical conductivity) of its water. When the groundwater EC is high, the resistivity of the aquifer can reach the same range as a clayey medium and the resistivity parameter is of no use in determining the aquifer. Here, the MRS (magnetic resonance sounding) method is most appropriate, because of its effectiveness in determining groundwater occurrence, whatever its conductivity.

- 4) To answer the fundamental questions of hydrogeologists implementing boreholes (Where is the groundwater? What could be the expected yield of a drilled borehole? What is the water quality?), the joint use of resistivity measurements and MRS methods could be helpful.

As a consequence, a 6-week test survey was organised at the end of 1998 at 12 sites, using VES, TDEM, ERI, and MRS jointly.

1.1.1 PHYSICAL ENVIRONMENT

Siem Reap province is located North of Tonle Sap Lake, in the North-Western part of Cambodia (Figure 5.65). The total surface area of the province is 15 270 km². Except for the Kulen

mountains in the north (maximum 487 m ASL) and several isolated peaks (known as Phnom), the province is flat. The main rivers flow from the North to the Tonle Sap Lake in the South. The province has a tropical monsoon climate, with two rainy seasons (May-June and August-October), an average annual rainfall of 1 500 mm and an average air temperature of 26.7°C.

1.1.2 HUMAN ENVIRONMENT AND DRINKING-WATER COVERAGE

The population of the province is about 650 000 (1993). The different economic activities are farming and fishing (90%), trading (7%) and services (3%). The main agricultural activity is paddy rice farming.

The potable drinking-water coverage of the province is estimated to be about 25%, provided by shallow wells (3 to 6 m deep) and boreholes (20 to 80 m deep), mostly equipped with handpumps. The 75% of people who don't have access to potable drinking-water draw water from ponds, rivers and open wells. The health statistics shows that 40% of the hospital admissions and 64% of the known deaths in the province are due to water and environmental diseases.

1.1.3 HYDROGEOLOGICAL ENVIRONMENT

The province is mainly located in a sedimentary basin that consists of 20 to 100 m thick layers of late Tertiary to Quaternary sediments that range from coarse sand to clay. These rocks lie on older consolidated sediments (Secondary to Tertiary sandstone to mudstone) and igneous rocks (diorite, granodiorite, basalt and andesite) which outcrop in the Kulem and Phnom mountains.

The aquifers can be subdivided into:

1) Shallow unconfined aquifers (alluvium deposits) which range from 2 to 5 m deep on average. These aquifers are used by the population through shallow wells, which are sometimes temporary and almost always polluted by organic matter.

2) The aquifers targeted by the drilling programme, ranging from 20 to 80 m deep. They are often confined and heterogeneous:

- in the East of the province, the aquifers are often clayey (clayey sand). The average transmissivity is 10^{-6} m²/s and the average drilling success rate is 35%;
- in the central part of Siem Reap, the sediments are more sandy (less clay) and the average transmissivity is 10^{-2} m²/s. The drilling success rate reaches almost 90%;
- in the Western region, the aquifers are mainly sandy silt and the average transmissivity reaches 10^{-5} m²/s. The average drilling success rate is 60%.

1.1.4 TEST-SURVEY METHODOLOGY

Twelve sites were chosen according to human and technical parameters: the water needs of the people, sites where the aquifer was well identified and sites where several boreholes were unsuccessful. The field surveys were conducted using the four geophysical methods: VES, ERI, TDEM sounding and MRS. The interpretation of the field data was conducted firstly independently for each method, and then jointly, using the available tools (Table 5.XV). Based on these results, boreholes were drilled and tested by electrical logging and pumping tests.

The pumping tests implemented were step tests. The objective was to estimate the specific capacity, the linear and the quadratic head loss, and the maximum exploitation yield of the boreholes. Recovery analyses were also done to estimate the aquifer local transmissivity. The pumping tests were interpreted using the Jacob formula. The technical contribution of geophysics to the borehole success rate was estimated from these twelve sites, and the financial contribution of geophysics to the borehole programme was calculated with ACF figures.

Finally, combining the technical and the cost analyses, a geophysical methodology for borehole siting, drilling and testing was proposed.

Table 5.XV: Methods, equipment and interpretation tools.

Method	Array	Equipment	Interpretation tools
VES	Schlumberger and pole-pole	Ω mega (Aquifer)	Cagniard curve and 'QWSEL' software
ERI	Pole-pole	SYSCAL R2 (IRIS Instruments)	'RES2DINV' software
TDEM	Tx: 100 x 100 m & 75 x 75 m Central Rx: 31.4 m ²	PROTEM 47 (Geonics Ltd.)	'TEMIX' software
MRS	TX/Rx: 75x 75 m & 37.5 x 37.5 m	NUMIS (IRIS Instrument)	'Samovar' software
DC logging	Normal probe 8" & 32" Fluid resistivity	ACF	
Drilling / borehole	Mud rotary 6" PVC casing 4" Screen 0.8 mm slot & 7.5% opening area Gravel pack 1-5 mm	PAT-ACF 201	
Pumping test	1 to 4 pumping steps	4" submersible pump	'Jacob' equation

1.2 Results example

1.2.1 A HIGH-YIELD BOREHOLE: ACPI SCHOOL

The objective was to construct a borehole to supply water to a school (Figure 5.66).

One-dimensional resistivity measurement: VES and TDEM

The inversion of the Schlumberger VES is presented in Figure 5.67. The geological interpretation of the VES is that the resistive layer of 477 Ω .m could be a sandy aquifer from 10 to 42 m deep, which overlies a clayey substratum dropping down to 10 Ω .m. The equivalence problems do not allow separate determination of the resistivity and the thickness of this potential aquifer (only its transverse resistance can be determined), and the thickness can range from about 26 to 36 m, which has a great influence on its transmissivity.

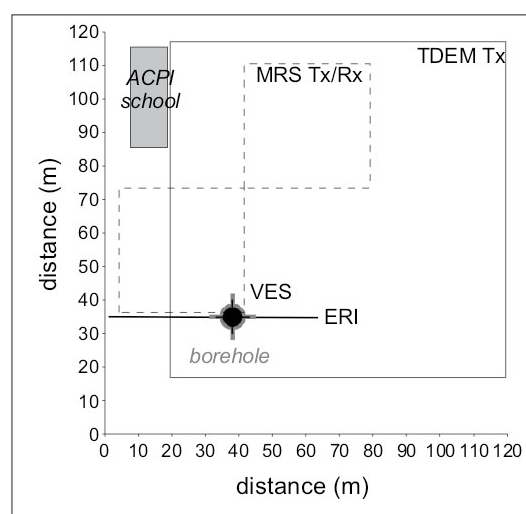


Figure 5.66: Location map of geophysical survey at ACPI school.

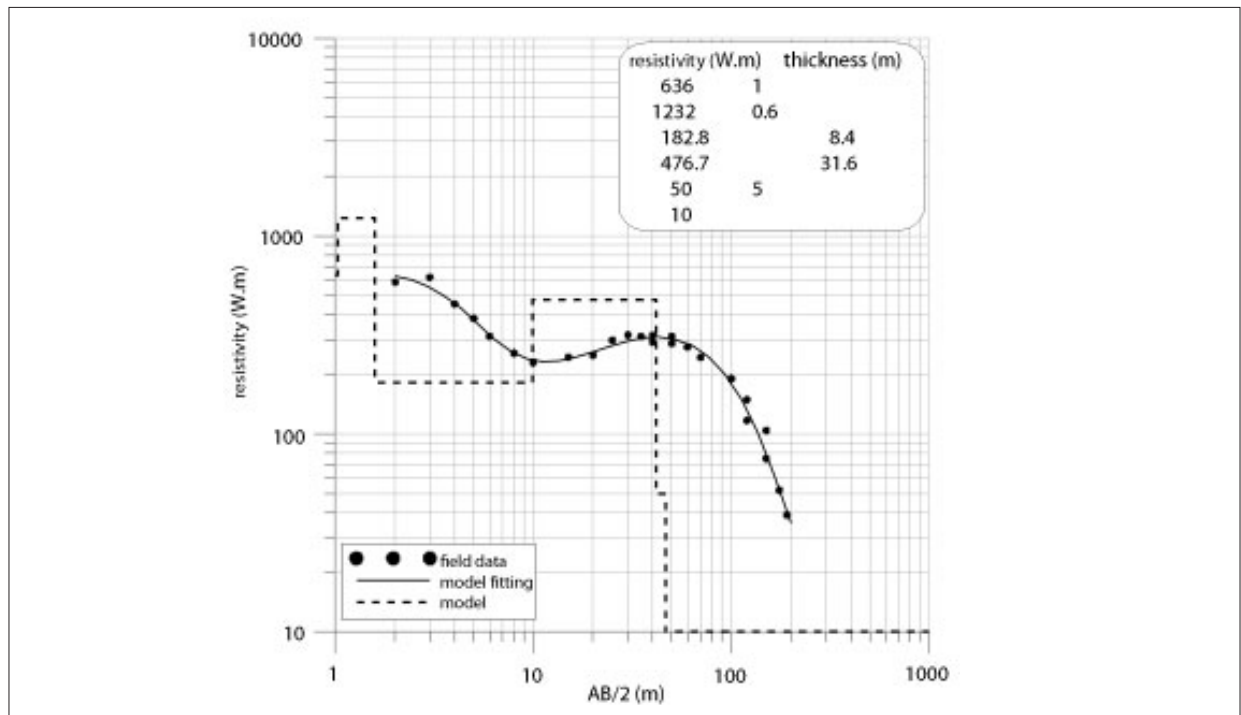


Figure 5.67: VES Schlumberger (rms = 7.5%), ACPI School.

The inversion of the TDEM sounding (Figure 5.68) is similar to that of the VES but: (1) the TDEM enables a better definition of the interface between the top resistive layers and the conductive one to be obtained – the estimation of the bottom of the potential aquifer is then improved; (2) the first 10 m below ground level are defined with less accuracy due to the integration of the shallow information with the TDEM method; (3) the penetration over 200 m is deeper (with the arrays used).

Two-dimensional electrical imaging: ERI

The inversion result is in good agreement with the Schlumberger sounding, but it is smoother (Figure 5.69). It also gives additional information that is important for drilling the borehole: the potential aquifer (resistive layer) seems to be quite homogenous.

MRS

The raw data (Figure 5.70) shows reliable measurements: the signal-to-noise ratio is high (average of 10.6), the Rx frequency of the magnetic resonance signal is very close to Tx frequency (less than 0.5 Hz of difference) and the signal phase changes smoothly as the pulse moment increases.

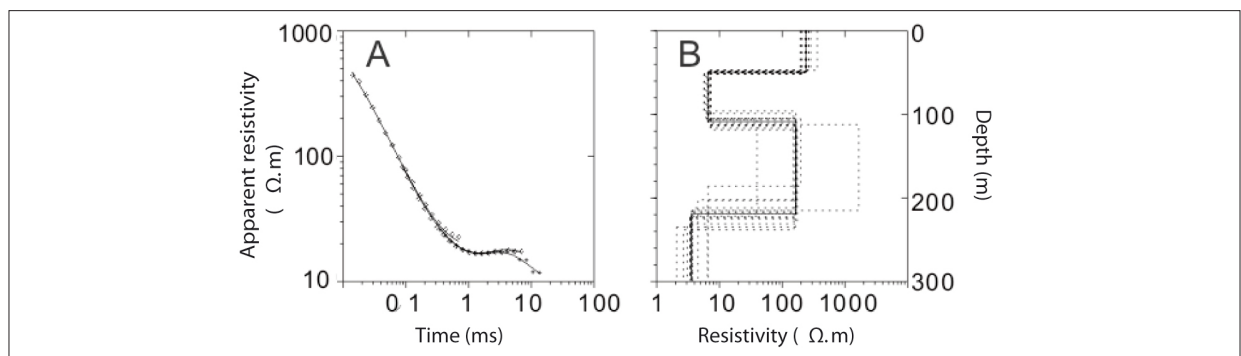


Figure 5.68: TDEM sounding, ACPI school.

A: field data. **B:** inversion (equivalent models, best fit rms error = %).

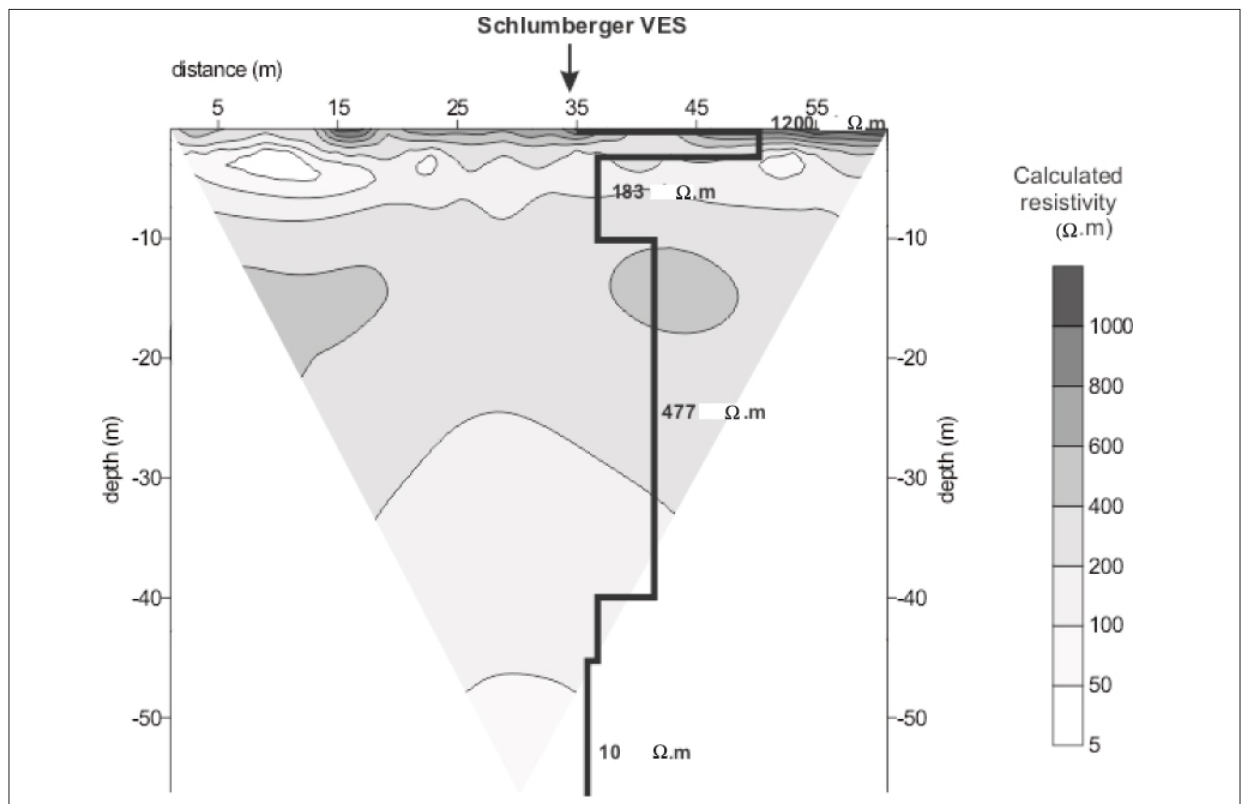


Figure 5.69: ERI (rms error = 3.1%) and VES inversion results, ACPI school.

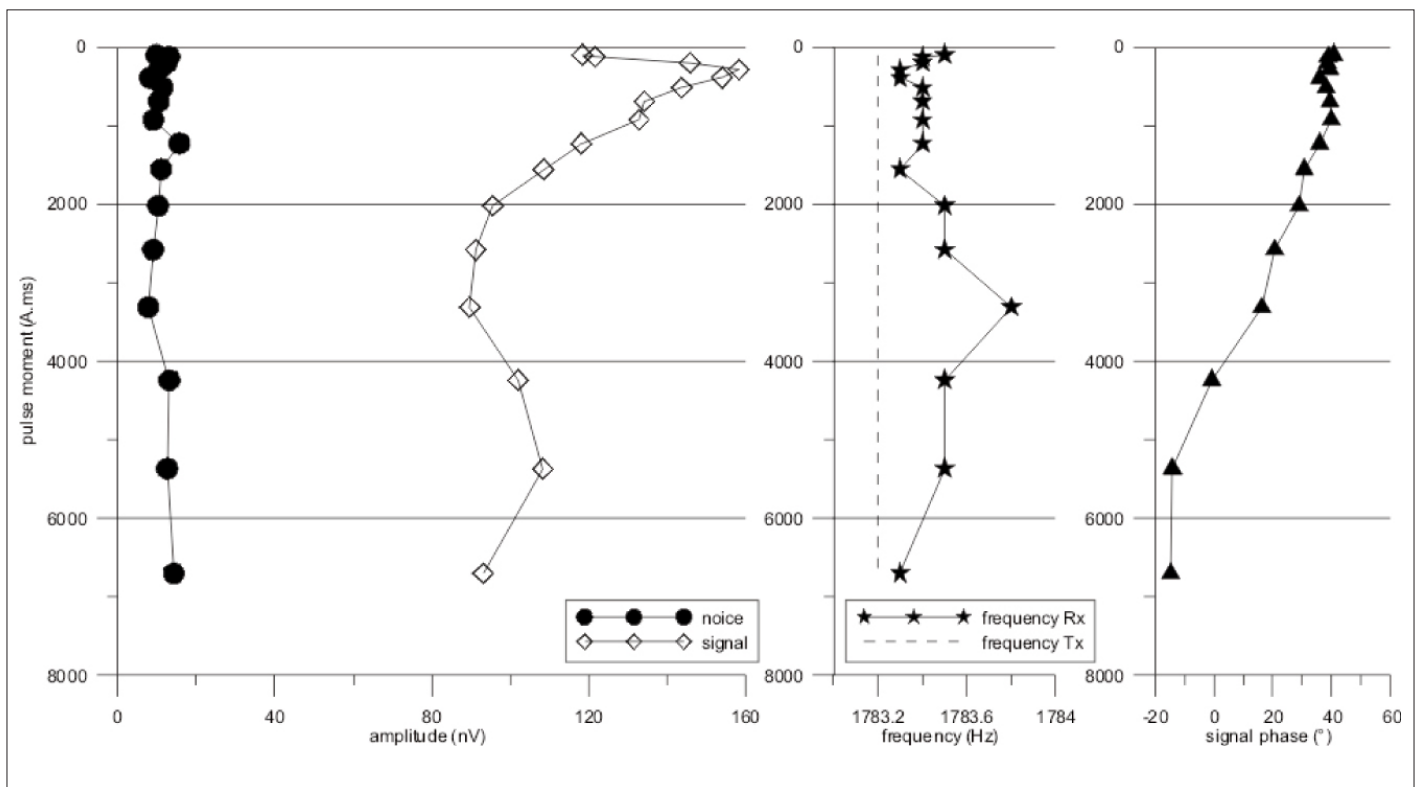


Figure 5.70: MRS raw data, ACPI school.

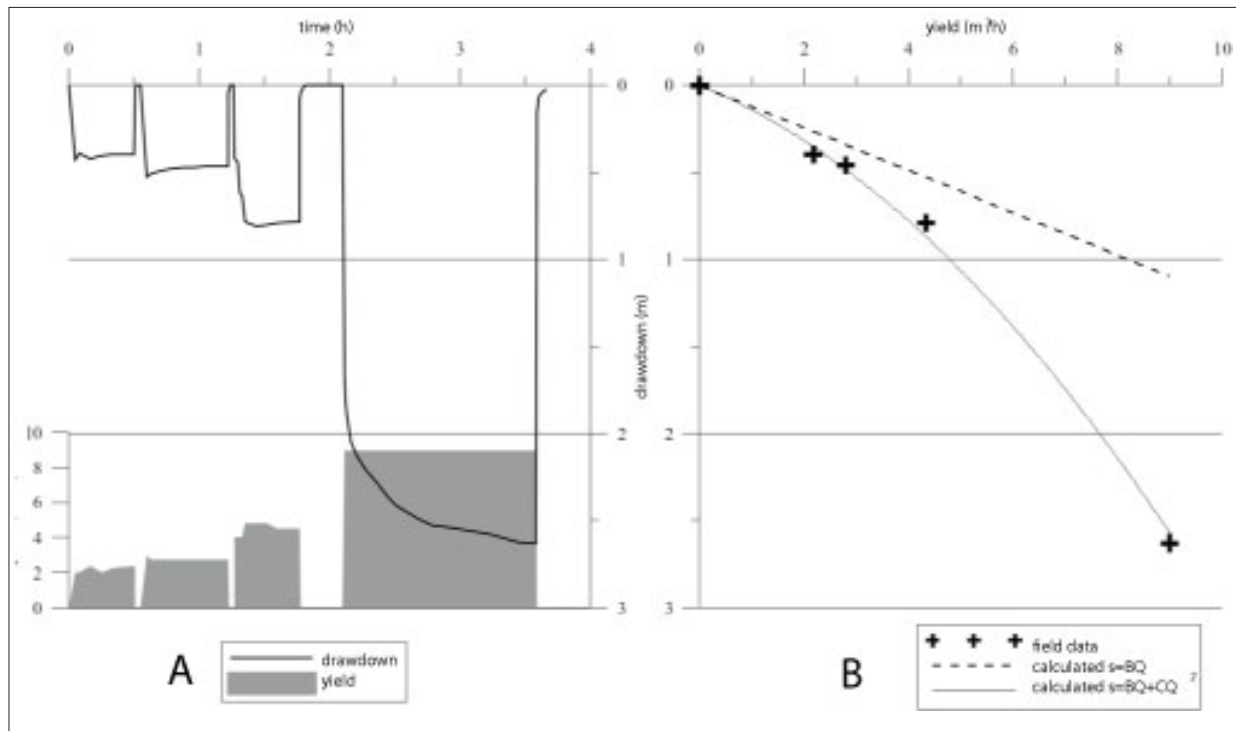


Figure 5.71: Pumping test, ACPI school.
A: field measurements. B: head-loss calculation.

The MRS inversion (Figure 5.72) shows a water-bearing formation from 1 m to the bottom on the investigated depth. The water content ranges from 5 to 20% and the decay time from 160 to 280 ms.

Borehole drilling and pumping test

The drilled borehole was surveyed with by electrical logging that was mainly used to choose the best screen location. After the development of the borehole by air-lift, a 4-step pumping test was conducted (Figure 5.71). The critical yield was not reached during the test, but the quadratic headloss increased strongly when the pumping yield was increased to 9 m³/h. The maximum exploitation yield was fixed at 6 m³/h (beyond this point the quadratic headloss become higher than the linear one).

The local transmissivity was estimated from the recovery analysis to be $1.8 \cdot 10^{-2} \text{ m}^2/\text{s}$ and the relative specific capacity was $1.4 \cdot 10^{-3} \text{ m}^2/\text{s}$. The groundwater resistivity at the end of the pumping test was $350 \text{ } \Omega \cdot \text{m}$ ($29 \text{ } \mu\text{S}/\text{cm}$), which was close to that of the local rainwater.

Comment

For this particular site, it can be seen (Figure 5.72) that the Schlumberger VES gives enough information to site the borehole successfully, even if the existing doubt regarding the aquifer thickness could only be removed with TDEM information. The ERI does not allow more accurate borehole siting, because the reservoir seems to be quite homogeneous. The MRS inversion gives realistic water contents regarding the lithology, but the T_2^* decay times is more difficult to correlate with the rock type.

1.2.2 A HETEROGENEOUS SITE EXAMPLE: MUKPEN VILLAGE

Three unsuccessful boreholes (yield $\approx 300 \text{ l/h}$) were drilled in Mukpen village. A geophysical survey was conducted to locate a new site. Figure 5.73 shows a summary of the surveys carried out both on a previous unsuccessful borehole (Northern site) and on the new site selected for drilling (Western site).

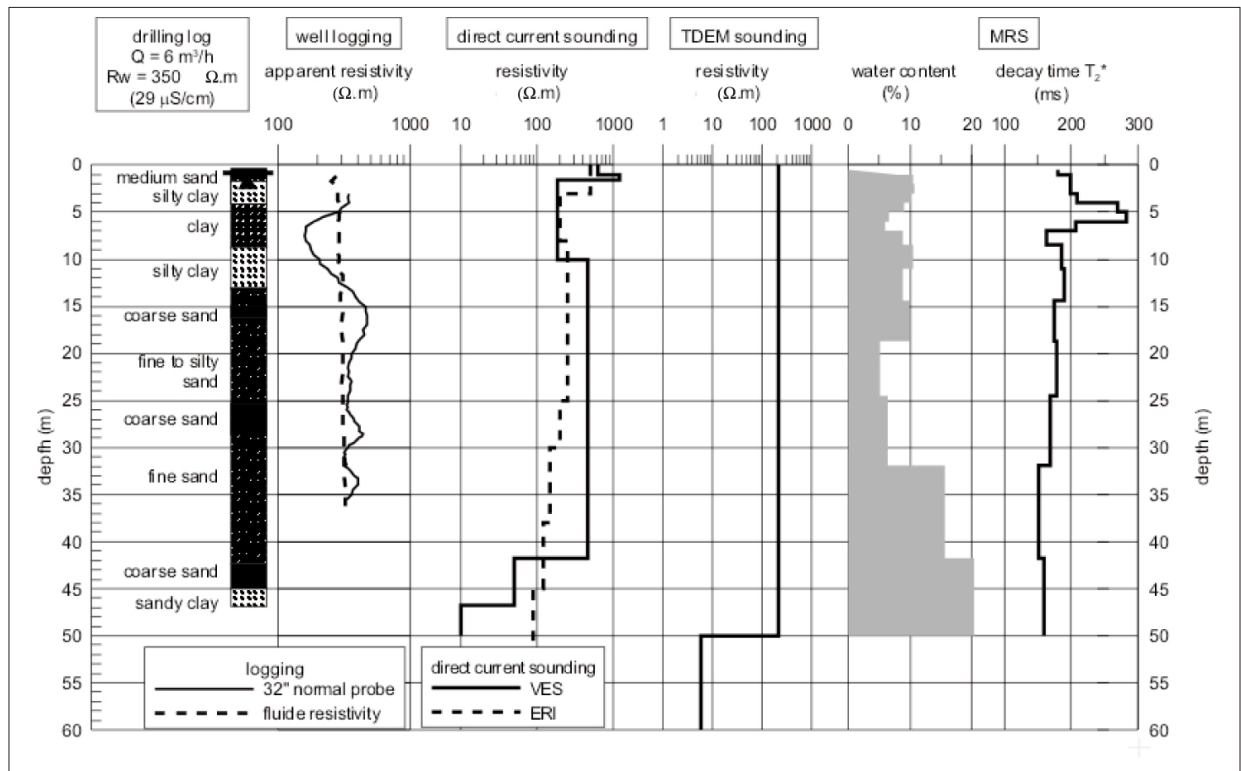


Figure 5.72: Geophysical results summary, ACPI site.

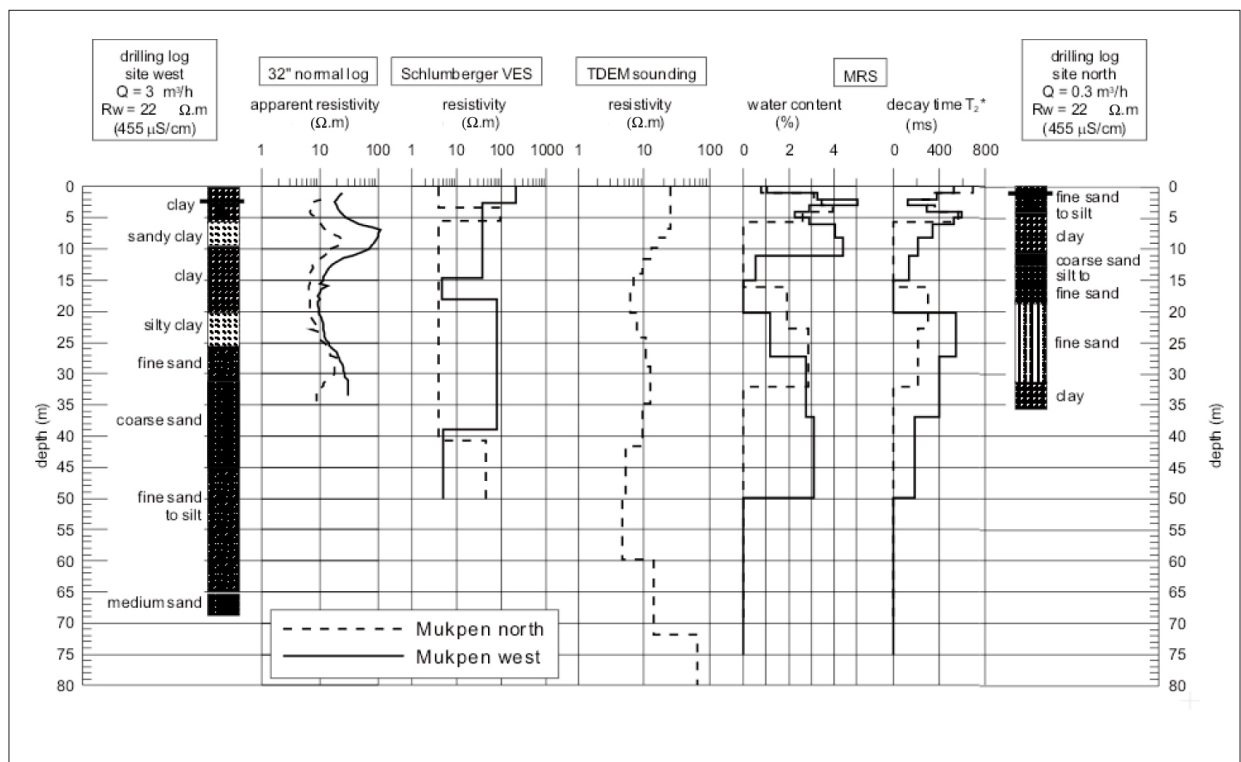


Figure 5.73: Geophysical results summary, Mukpen village.

Northern site: low-yield borehole

The resistivity values indicate a conductive area (less than 10 $\Omega\cdot\text{m}$) which could not be interpreted as an environment with high hydrogeological potential. The MRS inversion gives a realistic estimate of water content corresponding to the lithology and the borehole yield, but the T_2^* decay time does not fit with the expected values (longer time for silty sand than for fine sand).

Western site: medium-yield borehole

The Schlumberger sounding indicates a resistive layer of 80 $\Omega\cdot\text{m}$ between 18 and 39 m deep. This value corresponds to a sandy formation but is much lower than the one at the ACPI school site which was about 480 $\Omega\cdot\text{m}$. This could be explained by a small clay content which had not been noticed or the low water resistivity of 22 $\Omega\cdot\text{m}$ (455 $\mu\text{S}/\text{cm}$) comparing to the 350 $\Omega\cdot\text{m}$ (29 $\mu\text{S}/\text{cm}$) at the ACPI school site. The thickness of the resistive layer is poorly defined by the Schlumberger sounding according to the borehole lithology.

The MRS indicates a water-bearing formation from 20 to 50 m deep, in good agreement with the lithology for the water content, but the T_2^* decay time does not fit with the expected values. As for the ACPI school site, some water with long decay time in a shallow clayey layer is measured.

Comment

In this context of a sandy and clayey formation with conductive water, it can be noted that without *a priori* information on the groundwater electrical conductivity (EC), the resistivity methods do not enable reliable borehole siting because the aquifer resistivity is about the same as a sandy clay material. The MRS indicates higher potentiality, that is higher water content, longer relaxation time and a thicker saturated layer, for the western site. It enabled the successful siting of a borehole in Mukpen village.

1.3 Discussion

The application of geophysics to site a borehole could be useful for two reasons: (1) it helps ACF reach its objective of supplying people with drinking water, (2) it allows money to be saved within the programme by reducing the number of dry boreholes.

1.3.1 TECHNICAL ANALYSIS

MRS

The MRS gave accurate information of water content according to the lithology. There was no clear link found between T_2^* decay time and the aquifer grain-size. According to the literature, the T_2^* decay time varies from 30 ms in clay to 400-600 ms in gravel formations. However, decay times of 100 to 200 ms were often found for clayey material, and less than 150 ms for coarse sand. This could be explained by variations in the magnetic properties of rocks which create local inhomogeneities in the geomagnetic field. The T_1 decay time, which depends mainly on pore size, was not yet available at the time of the survey (1998).

However, MRS is the only method that enabled successful drilling of a borehole in a conductive environment. Figure 5.74 shows the links between the MRS data and local transmissivity of aquifers estimated from pumping tests. The MRS transmissivity was estimated as $T = 1.6 \cdot 10^{-6} \cdot w^4 \cdot (T_2^*)^2 \cdot \Delta z$, where w is the MRS water content in % and Δz the MRS saturated-aquifer thickness in m.

Joint use of the methods

On one hand, the main limitation of resistivity methods in Siem Reap is their high sensitivity to the water EC; on the other hand, this is their main advantage compared to MRS, which is not sensitive to the water quality (the water electrical conductivity is related to the ions which are present in solution).

As a consequence, resistivity methods and MRS can complement each other quite effectively: MRS can be used to locate groundwater, and resistivity methods can be used to check the mineralisation of the water (Figure 5.74).

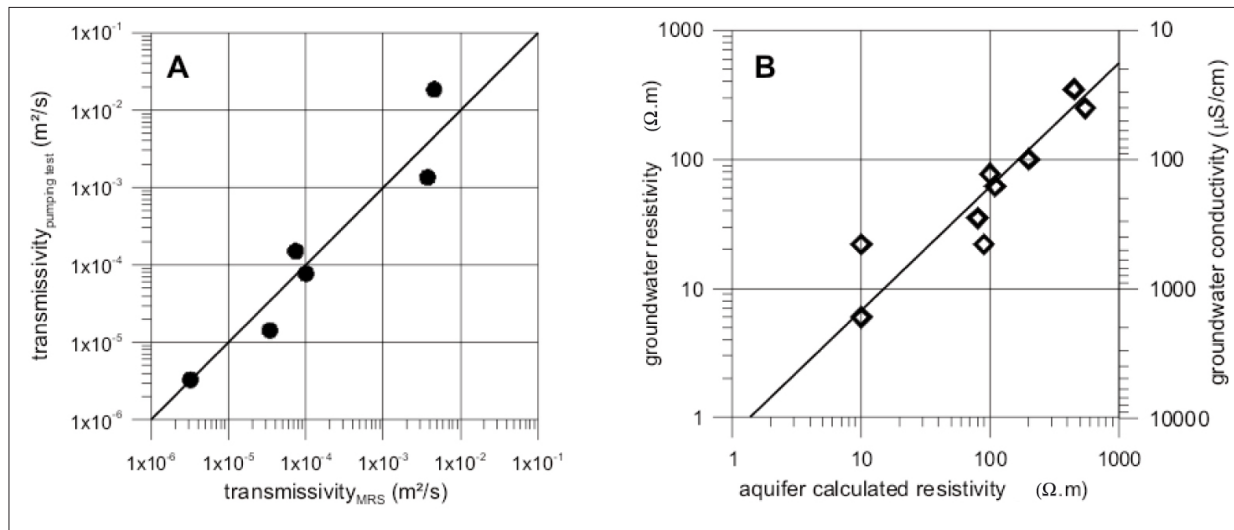


Figure 5.74: MRS hydraulic characterisation.

A: local aquifer transmissivity B: groundwater electrical conductivity.

TDEM is the most convenient method to integrate with MRS in this conductive environment as it can use the same transmitter coil, reducing the set-up time. In a heterogeneous area, an ERI could be used to look for the most convenient site where integrated TDEM/MRS could be carried out.

Technical impact of geophysics

Within the 12 investigated sites, the use of geophysical methods increased the borehole success rate from 56 to 90% for a total of 36 boreholes that were used for this particular test survey. This improvement of success rate was due to: (1) better choice of the drilling sites, (2) better management of the drillers who received guidelines regarding the depth they had to drill, (3) better borehole installation, i.e. screen setting.

1.3.2 COST ANALYSIS

The cost analysis aims to measure the financial impact of geophysics on a drilling programme. Table XVI shows the average costs of geophysical surveys in Siem Reap province. The calculation includes the cost of staff (engineer, technician and labourers), logistics (vehicle purchase, running and depreciation costs) and the geophysical equipment (purchase, maintenance and depreciation). It was

Table XVI: Cost of geophysical surveys, in Euro (€).

cps = cost per survey. dd = duration of depreciation.

	Staff No	cps	Geophysical equipment purchase cost	dd years	cps	Vehicle No	cps	Survey number per year	Admin. cost (8%)	Total cost per survey
VES	4	105	4 615	3	21	1	94	80	18	238
ERI	4	105	30 769	4	98	1	94	80	24	321
TDEM	3	74	53 846	4	151	1	83	90	25	333
MRS	3	74	92 307	4	258	1	83	90	33	448
MRS & ERI	5	136	123 076	4	413	1	100	75	52	701
MRS & TDEM	5	127	146 153	4	460	1	94	80	54	735
MRS & TDEM & ERI	7	188	176 922	4	557	2	187	80	75	1 007

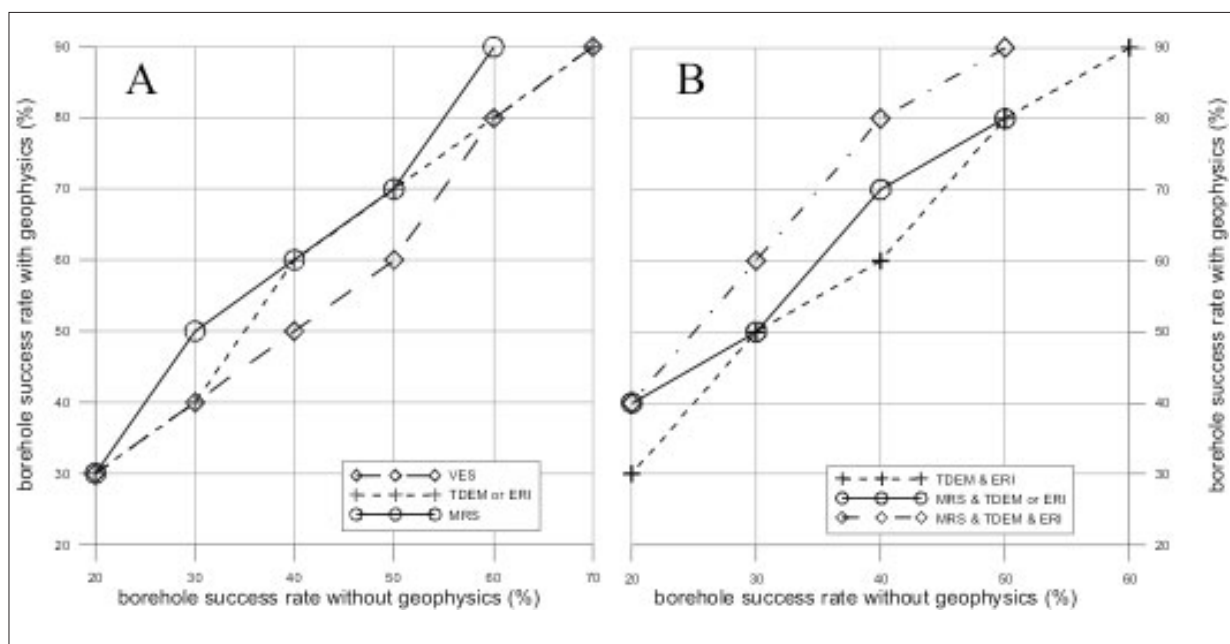


Figure 5.75: Cost-effectiveness boundaries of geophysical methods.

A: use of a single method. B: joint use of methods.

done on the basis of 8 working months per year (dry period), 2 to 3 sites investigated per week, and cost calculated on real ACF expenses in 1998. Using the same calculation method, the average cost of a successful borehole is 2 380 €, compared with 1 260 € for an unsuccessful one.

Figure 5.75 shows the cost-effectiveness boundaries of using geophysical methods for siting and designing boreholes (it was prepared using the formula presented earlier in Chapter 5A, Section 4.1.2). Starting from the borehole success rate without geophysics, the figure shows the minimum success rate which needs to be reached, using geophysics, to save money. For example, for a situation where the borehole success rate without geophysics is 40%: (1) the integrated use of MRS, TDEM and ERI saves money when it increases the success rate to 80%; (2) the joint use of TDEM and MRS saves money when it increases the success rate to 70%.

1.3.3 CONCLUSION: GEOPHYSICAL IMPLEMENTATION METHODOLOGY

Integrating the technical and the cost analyses, it is possible to propose a geophysical methodology for borehole siting and design Siem Reap province. According to the different areas of the province and their known borehole success rate, the following methodology could increase the borehole success rate and save money at the programme scale.

Very difficult area: borehole success rate without geophysics less than 30%

The joint use of ERI, TDEM and MRS could reduce significantly the number of dry boreholes and could save money if the success rate is at least doubled. An average of one site could be surveyed per day.

First, ERI is used to choose the best site according to the vertical and lateral contrasts of resistivity and thickness.

Then MRS is carried out on the potential aquifer site to estimate the borehole productivity using Figure 5.74A. The TDEM sounding is used to accurately define the interface between the resistive aquifer and the conductive clayey substratum (to give recommendations to the driller). The electrical conductivity of the water can be estimated using the resistivity value of the layer identified as the aquifer by the MRS (Figure 5.74B).

Difficult area: borehole success rate without geophysics between 30 and 50 %

The joint use of MRS and resistivity methods (either ERI, to focus on local heterogeneity if necessary, or TDEM, which is easier to integrate with the MRS) is recommended. This joint use of methods could save money if the success rate is increased by at least 20 to 30%. An average of two sites could be surveyed per day.

Common area: borehole success rate without geophysics more than 50 %

VES could be used as a standard method saving money if the success rate is improved by 10 to 20%, which seems to be realistic up to a success rate without geophysics of 60 or 70%. An average of two sites could be investigated daily.

2 Characterisation of crystalline bedrock aquifers

Bedrock aquifers are of particular importance in tropical regions because of their wide spread and accessibility and because there is often no readily available alternative source of water supply. Even in humid tropical regions where surface water is readily available, water quality considerations can favour their use. To improve the current procedures to investigate groundwater resources in crystalline rocks contexts, a methodological survey was conducted in Burkina Faso by ACF with the support of IRD and Iris Instruments.

2.1 Background

2.1.1 CRYSTALLINE BEDROCK AQUIFERS

Bedrock aquifers are developed within the weathered overburden and fractured bedrock of crystalline rocks which are mainly of Precambrian age. The usual conceptual model of the bedrock aquifer describes several zones in the lithological sequence. The alterites (regolith) consist of weathered and decayed rock; their permeabilities vary in accordance with lithology but are usually low; they play the major part of the storativity in the aquifer. The underlying weathered-fissured zone (saprock) and the fractured bedrock present typically low storativity, but permeability commonly increases at lower levels due to a lesser development of secondary clay minerals and a high permeability of open fractures.

There are a number of important constraints to the development of bedrock aquifers. The failure rate of low-yield boreholes for rural water supply is high in the drier regions (typically in the range of 40 to 50%), and the implementation of high-yield boreholes for urban or irrigation purposes is always a challenge for hydrogeologists. Furthermore, the low storativity of bedrock aquifers often leads to unsustainable yield of boreholes.

Therefore, there is an important need both to improve the current methodology for high-yield borehole implementation and to assess more accurately the overall resources and aquifer occurrence to assist development efficiency and sustainable long-term use of the aquifer.

2.1.2 EQUIPMENT AND METHODS

Bedrock aquifers are of significant extent in Burkina Faso (around 80% of the total country surface area). To measure the contribution of the MRS method to characterise these aquifers, a survey was conducted from November 2002 to January 2003 in granite and associated rocks of Precambrian age.

MRS were carried out around recent boreholes drilled in the alterites, in the weathered-fissured zone and in fractured bedrock. All of the 13 boreholes were tested with well-test pumping tests (total pumping duration of 4 hours), and 6 of them were used to conduct aquifer tests (pumping duration of 72 hours).

The aquifers' local transmissivities were calculated from the recovery period of well-tests, and the storativities were calculated from piezometer records with Theis and Jacob methods. The geometry of the aquifers was deduced from borehole reports and the static water level was measured while implementing the MRS. The NumisPlus® equipment was used to set up the MRS with a square loop of a typical 150 m side. The adapted saturation recovery method was used to measure the longitudinal relaxation time, and the storativity and transmissivity of the aquifers were estimated from MRS as:

$$ne_{MRS} = 0.28 \cdot w$$

$$S_{MRS} = 4.3 \cdot 10^{-3} \cdot (w \cdot \Delta z)$$

$$T_{MRS} = 3 \cdot 10^{-7} \cdot (S_{MRS} \cdot (T_1^*)^2)$$

where ne_{MRS} is the specific yield of unconfined aquifer (%), S_{MRS} is the storage coefficient of confined aquifer, T_{MRS} is the MRS transmissivity (m^2/s), w and Δz are respectively the MRS water content (%) and saturated thickness (m), and T_1^* is the observed longitudinal relaxation time (ms).

2.2 Main results

2.2.1 RESERVOIR TYPOLOGY

Field results obtained in Burkina Faso show that the reservoir type could be estimated from MRS data. Table 5.XVII and Figure 5.76 indicate that the average value of water content is higher and the average value of longitudinal relaxation time shorter for water in alterite reservoirs than for water in fissured-fractured reservoirs. It means that the storativity is higher and the transmissivity is lower for the alterites than for the fissured-fractured zones, which is in accordance with the typical hydrogeological conceptual model. However, the values margins are large and ambiguity still remains when interpreting the MRS data alone.

Table 5.XVII: MRS parameter and reservoir type.

Reservoir	Water content (%)			Max.	T_1^* (ms)	
	Max.	Average	Min.		Average	Min.
Alterites	6	3	1	600	400	180
Fissured-fractured bedrock	2.5	1	0.2	1 500	650	350

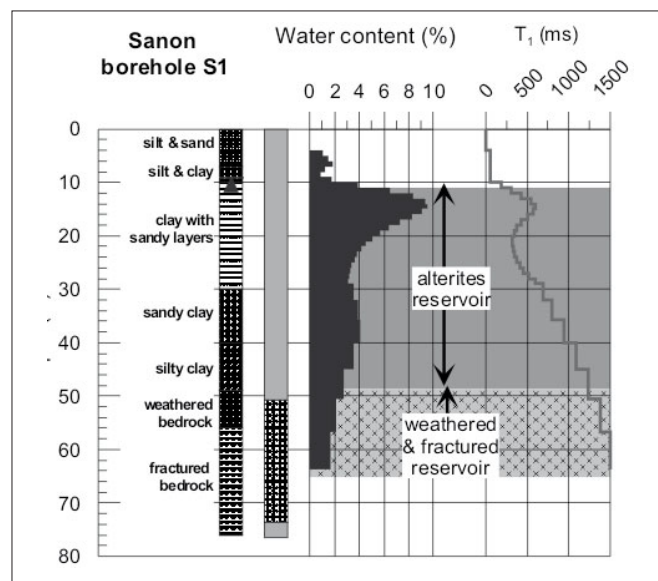


Figure 5.76: MRS and reservoir units, example of Sanon S1 borehole.

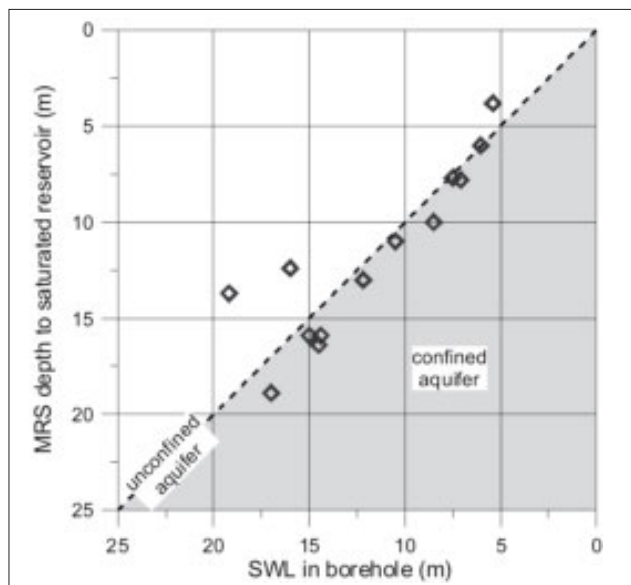


Figure 5.77: MRS and static water level (SWL).

2.2.2 RESERVOIR GEOMETRY

The depth to water-saturated layers obtained by MRS could be compared with the borehole static water level for an unconfined aquifer (Figure 5.77). The average difference with borehole data is $\pm 12\%$, and $\pm 17\%$ for the depth to the unweathered bedrock (Figure 5.78). This characterisation can be used to contribute to both a better implementation of a borehole (thickest alterites reservoir) and an evaluation of the resource (spatial extent of reservoir).

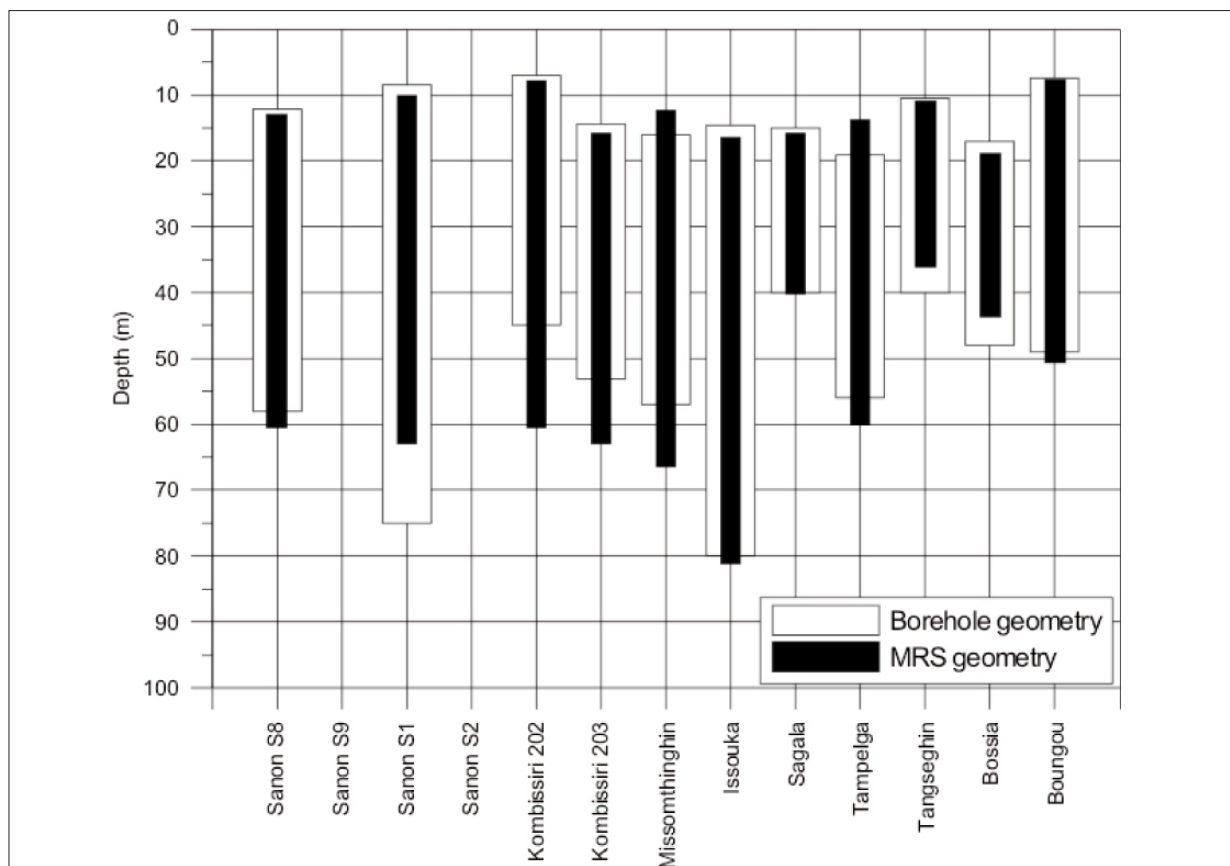


Figure 5.78: MRS and reservoir geometry.

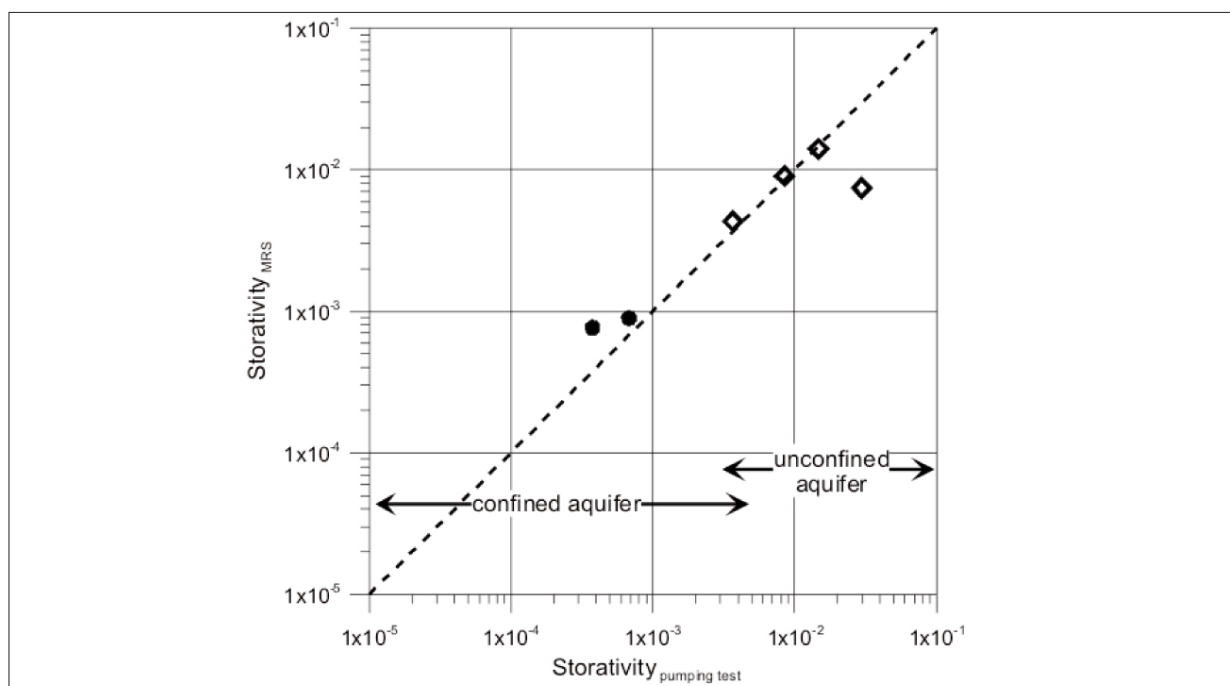


Figure 5.79: Aquifer storativity.

2.2.3 RESERVOIR HYDRAULIC PARAMETERS

The hydraulic properties of the aquifer were estimated after a calibration process with the borehole pumping-test results.

The average difference with properties calculated from pumping-test data are $\pm 80\%$ for the storativity (Figure 5.79) and $\pm 41\%$ for the local transmissivity (Figure 5.80). These first results still need to be confirmed with more data, but they are very encouraging and allowed ACF to propose using MRS as an interpolation tool between boreholes to estimate the hydraulic characteristics of the reservoirs.

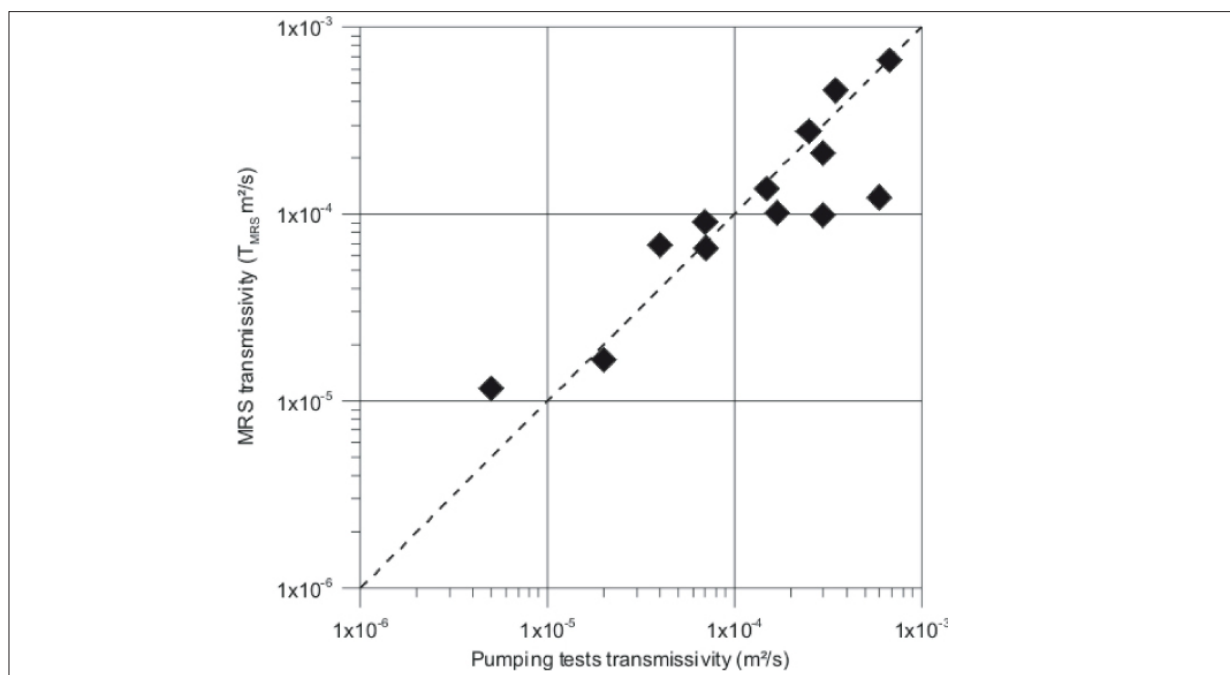


Figure 5.80: Aquifer local transmissivity.

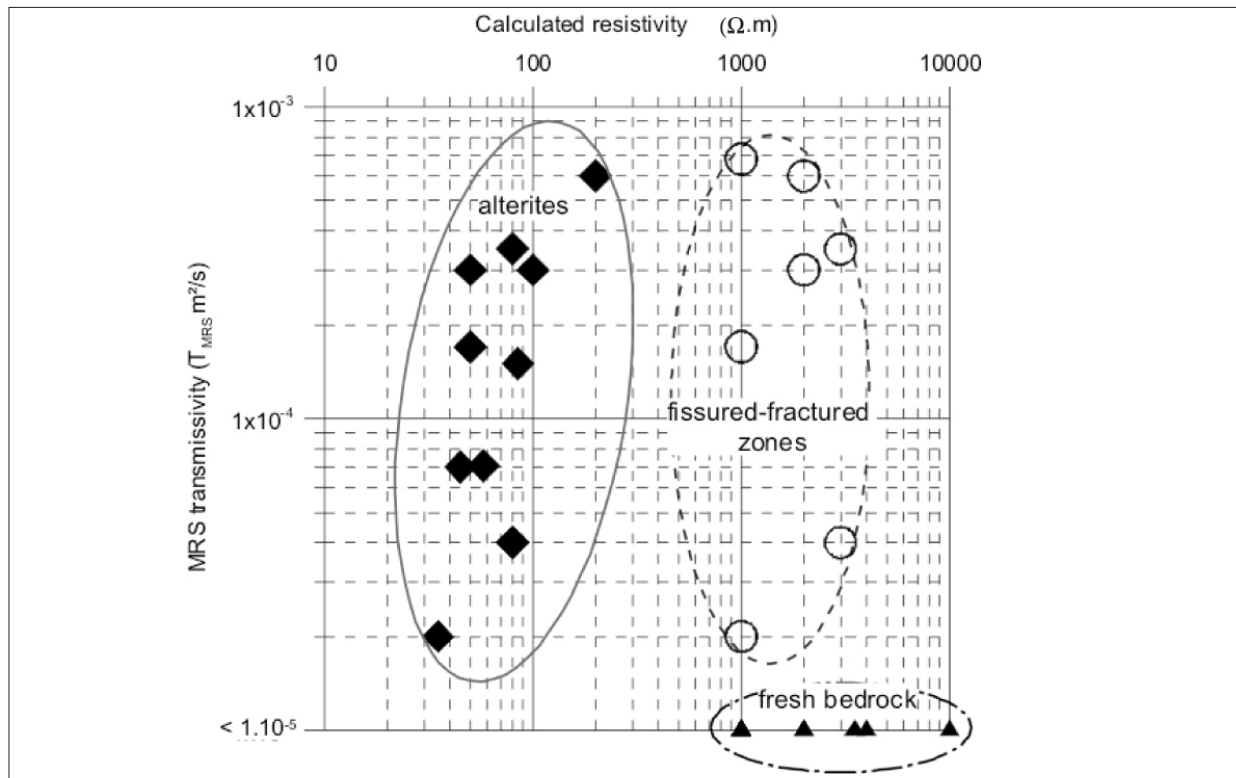


Figure 5.81: Joint use of resistivity and MRS transmissivity to characterise reservoirs.

2.2.4 IMPROVEMENT OF THE AQUIFER CHARACTERISATION

The characterisation of an aquifer is improved if the MRS results are jointly interpreted with electrical resistivity methods. If MRS resolution is not sufficient in depth, one-dimensional electrical soundings are often able to measure the depth of the substratum. In heterogeneous contexts, ERI can highlight the structures of the aquifers.

With the available data, the type of reservoir is well estimated when it is jointly characterised with its electrical resistivity and its MRS transmissivity (Figure 5.81).

2.2.5 MAIN LIMITING FACTORS

The main limiting factors of MRS in such geological contexts are (1) the time required for data collection, which ranges from 6 to 20 hours due to the low signal-to-noise ratio, (2) the one-dimensional measurement, which does not allow description of the reservoir structure over a scale corresponding to the loop size, and (3) the loss of resolution with depth, which does not allow measurement of small signals coming from deep productive fractures (Figure 5.82).

This suppression effect particularly affects the fractures deeper than half of the Tx/Rx side length, with low weathered development. Figure 5.82 shows an example where the fractures identified from an electrical borehole log below 40 m deep are not measured by the MRS carried out with a 75 m side loop (square shape).

2.3 Conclusion

The main conclusions of the comparison between the MRS results and the borehole data are:

- the geometry of the weathered part of the aquifer is well described by MRS;
- storativity and transmissivity can be reasonably estimated from MRS data after calibration with borehole pumping-test results;

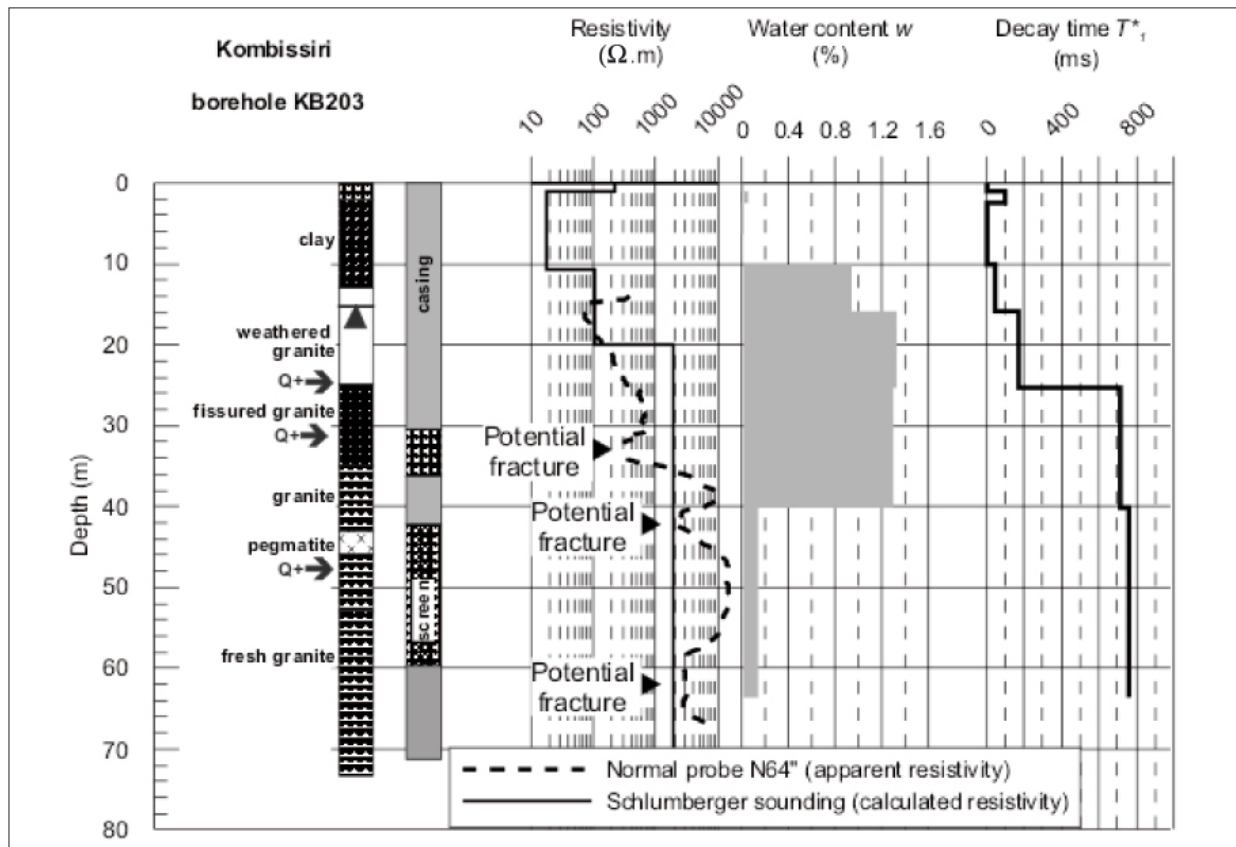


Figure 5.82: Suppression, example of Kombissiri borehole #203.

- the main limitations of MRS are the one-dimensional approximation in highly heterogeneous contexts and the loss of resolution when looking for deep narrow fractures;
- MRS is a useful tool to characterise aquifers in a crystalline context. Its use within a hydrogeological process is a promising support for hydrogeologists for both borehole implementation and estimation of water reserves (see Chapter 5A).

3 Localisation of saturated karst aquifers

The exploitation and management of a karstic aquifer is often a challenge to hydrogeologists. Whatever the stage is when karstified, the heterogeneous structure of karstic systems leads to a complex hydrodynamic behaviour, and this groundwater resource is difficult both to explore for borehole implementation and to protect from pollution. Therefore, specific surveys need to be done to develop and protect these resources for the benefit of everyone. With this in mind, a methodological project was set up between ACF, IRD, the *Bureau de Recherche Géologique et Minière* (BRGM) and the University of Montpellier to evaluate the impact of geophysics on karstic-aquifer investigation.

3.1 Background

3.1.1 PHYSICAL ENVIRONMENT

The Lamalou site is located on the Hortus karstic area, 40 km north of the French city of Montpellier. This limestone plateau covers an area of about 50 km², and its elevation ranges from 195 to 512 m above sea level. It mainly features limestone outcrops, and soil only exists in rock fissures. It is covered with Mediterranean shrubby vegetation. There is no regular surface-water running on the plateau.

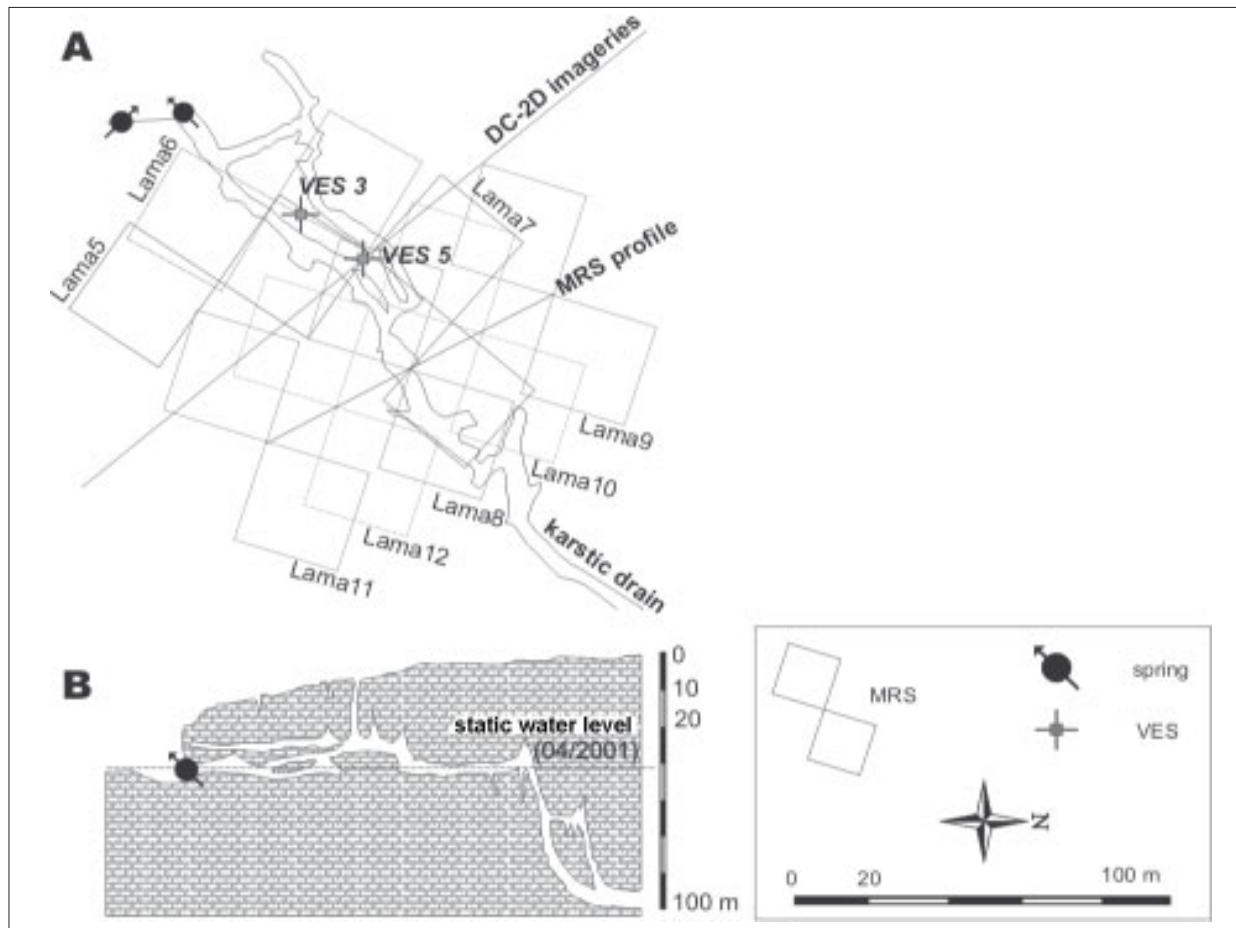


Figure 5.83: The known conduit system of Lamalou karst, and location of geophysical surveys.
A: map. B: cross-section along the conduit.

The Hortus area consists of Upper Valanginian limestone, 80 to 110 m thick, which lies on Upper Berriassien to Lower Valanginian marl. The main recharge area of the Hortus plateau discharges at the Lamalou spring.

The main aquifer is the Upper Valanginian limestone. Its upper part (several metres thick) is highly fissured and weathered: this is the epikarst, which can be saturated in places (see Chapter 3). Under the epikarst, an infiltration zone about 20 m thick drains the water through micro-fissures and joints down to the saturated karst. The porosity of the limestone is very low (1.8%) and the groundwater mainly flows through fissures, fractures and karst conduits. Part of the main conduit was explored and mapped by speleologists (Figure 5.83).

3.1.2 SURVEY OBJECTIVE AND METHODOLOGY

A methodological test was conducted, jointly using MRS and ERI. The main objective was to check how MRS and ERI could help the hydrogeologist to look for groundwater in karst. The specific objectives were to check if MRS could measure and differentiate between the water in the epikarst and the water in the saturated karst, and ERI sensitivity to karstic caves.

Eight MRS and two ERI were carried out on a known system where the depth and the geometry of the main conduit and a cave are mapped (Figure 5.83). Two boreholes were used to measure the static water level and the lithology was known through the drilling logs.

The common hydrogeological parameters, such as permeability or porosity, that are defined for porous media are not usable for karstic systems, which are highly heterogeneous. However, a

qualitative approach could be used to identify the karstic structures because all karstic zones have their own particular geophysical response: the MRS water content and relaxation time are higher for conduits and caves containing water (secondary porosity) than for compact rocks (primary porosity), and the electrical resistivity of empty caves should be infinite compared to the high resistivity of carbonate rocks or to the medium resistivity of saturated structures.

Thus, two MRS estimators k_x (permeability estimator) and T_x (transmissivity estimator) that are normalised MRS permeability and transmissivity could be used to differentiate water in saturated dissolution features (i.e. conduits and caves) from unsaturated and epikarstic zones (Figure 5.84) :

$$k_x(z) = \frac{w_x(z)T_{lx}^2(z)}{w_r(z)T_{lr}^2(z)}$$

$$T_x = \frac{\int_z w_x(z)T_{lx}^2(z)dz}{\int_z w_r(z)T_{lr}^2(z)dz}$$

where x and r are two of N MRS soundings ($1, \dots, x, \dots, r, \dots, N$). The reference r is selected as $\{k_{x_{\max}}(z) \leq 1\}$ or $\{T_{x_{\max}} \leq 1\}$ for all soundings ($x = 1, \dots, N$).

The capability of MRS to detect water according to depth and volume of saturated zone was investigated using $T_1 > 400$ ms as the indicator of dissolution features. Model results show that water in saturated dissolution features can be easily identified at shallow depths (Figure 5.85). For deeper investigation, a larger volume of water is necessary: 100 m³ of water is detectable at 5 m deep, whereas at 15m, at least 250 m³ is required for MRS detection.

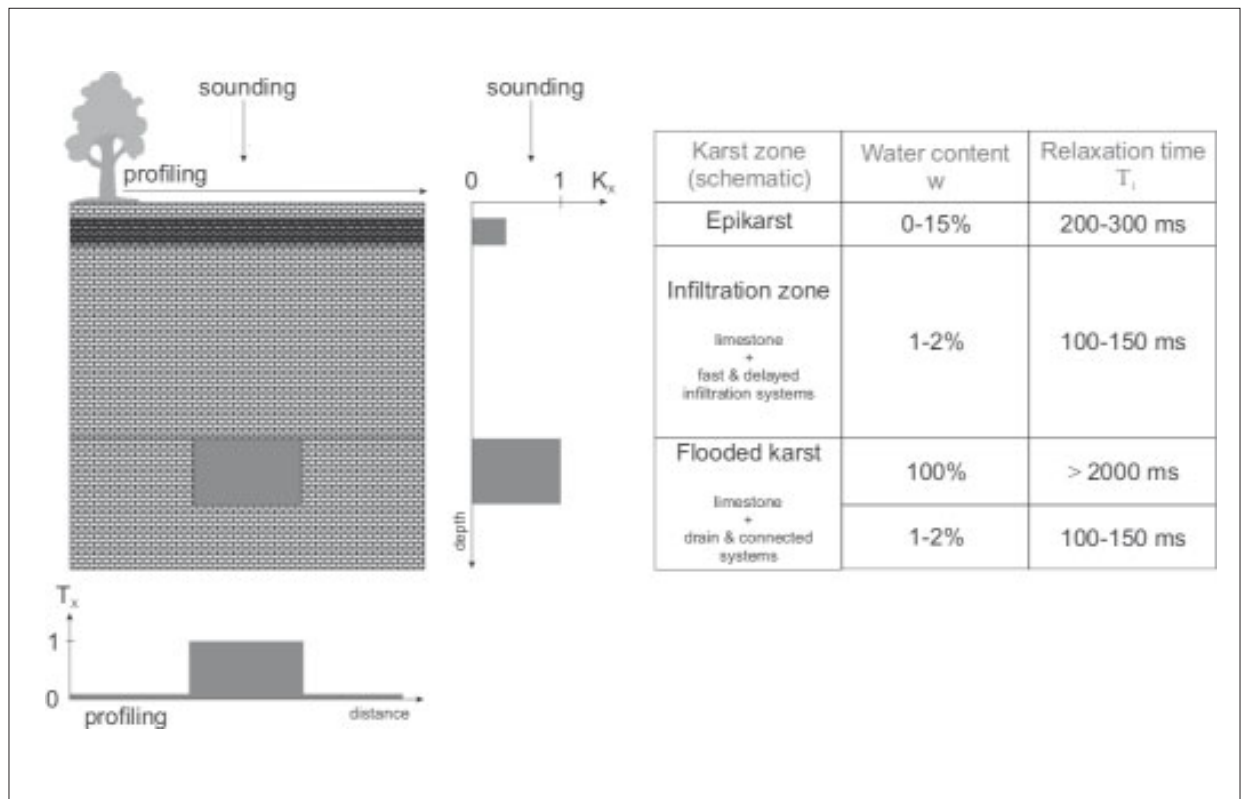


Figure 5.84: MRS estimators as tools to detect saturated karst structures.

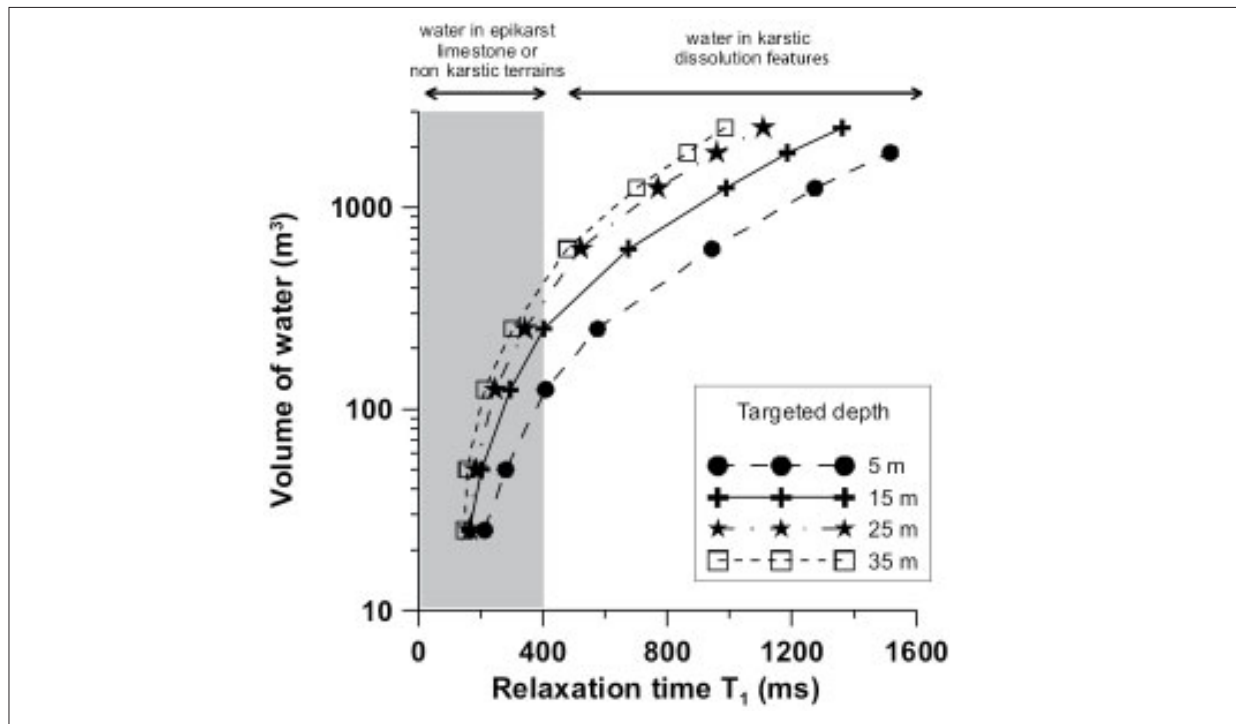


Figure 5.85: Required volume of free water ($w=100\%$ & $T_1=1500$ ms) for MRS measurements in a karstic environment ($w=1\%$ & $T_1=150$ ms).

3.2 Field example

3.2.1 MRS SATURATED KARST LOCALISATION

The NumisPlus® equipment developed by IRIS Instruments and BRGM, and the inversion software ‘Samovar’ were used. An eight-shape loop was used to improve the signal-to-noise ratio, leading the estimation of the investigated volume to be a parallel pipe of 120 x 50 x 50 m. A clear limitation of using MRS, which is a one-dimensional sounding method, to investigate a two- or three-dimensional target, is that it averages the information over the antenna size. To obtain two-dimensional pictures, triangulations with linear interpolation of MRS data were prepared (Figure 5.86).

Karst mapping

The map was drawn using a T-estimator calculated with the MRS Lama7 as reference (Figure 5.86A). It clearly shows a high-transmissivity channel which fits well with the known karstic conduit. The contribution of the different MRS to the map, i.e. their ability to measure groundwater, is linked to the location of the Tx-Rx loop. If the loop surface-area covers enough groundwater, the magnetic-resonance signal can be measured (Lama7, 8, 10), but if the loop only partly covers groundwater, the signal can be too low to be measured (Lama5 and 6).

Karst pseudo section

The water content ranges between 0 and 1.7% (Figure 5.86B). The maximum is reached around the known conduit and represents a thickness of about 1.5 m of water filling the width of the conduit. A second domain extends a few metres below the ground surface for a maximum water content of 0.6% reached around the Lama12 MRS. It could represent the well-known perched saturated area in the epikarst.

The relaxation time section (Figure 5.86B) indicates relaxation times longer than 400 ms for

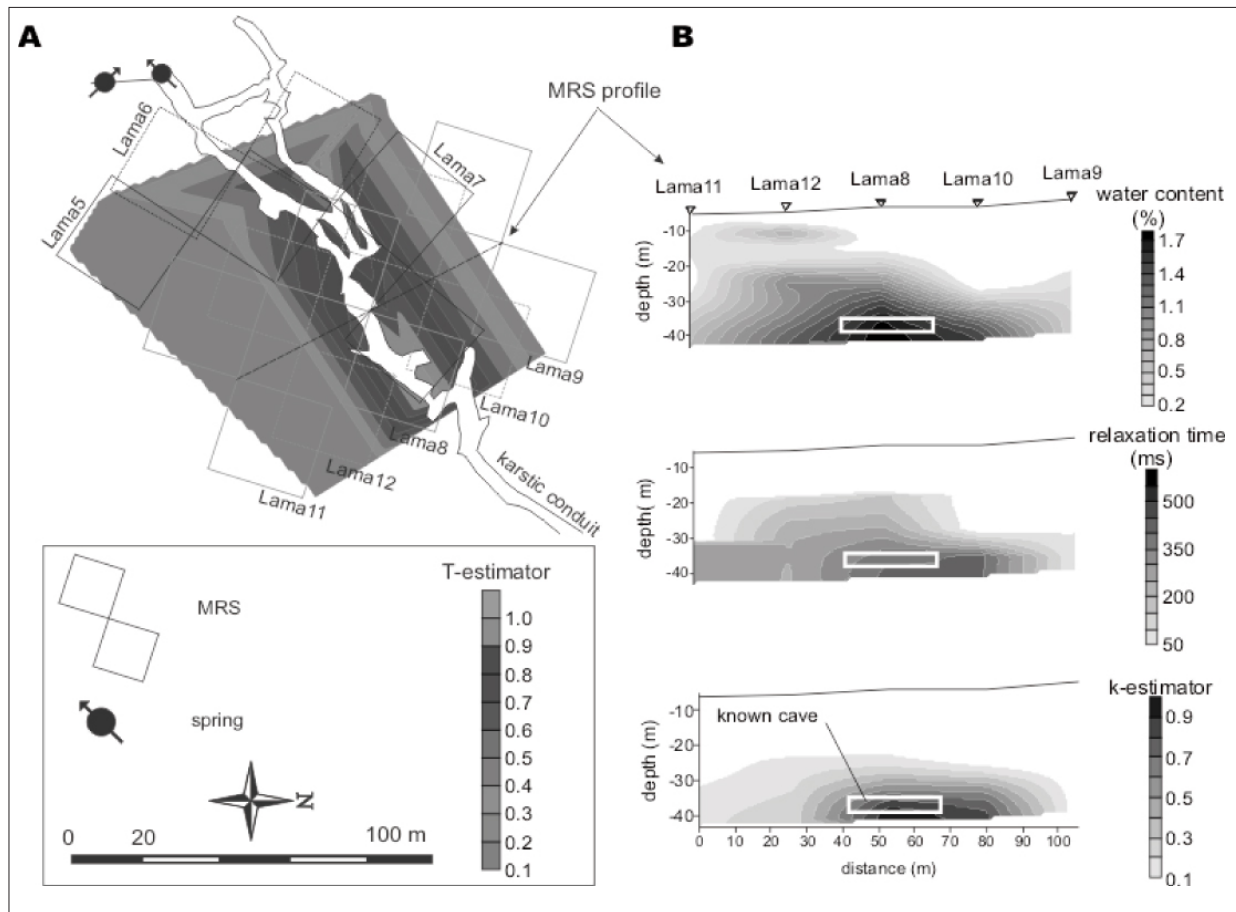


Figure 5.86: MRS Karst localisation.

A: conduit mapping using T-estimator. B: cross-section using k-estimator.

the known conduit area, and shorter than 400 ms for the epikarst and the limestone.

3.2.2 ERI KARST STRUCTURE ESTIMATION

Two ERIs were carried out with the Syscal® R2 system from IRIS instruments with 64 electrodes and an inter-electrode spacing of 4 m. The data was interpreted with RES2DINV software.

Figure 5.87 presents the inversion of the Wenner imageries. Two resistive anomalies of 20 000 $\Omega \cdot m$ located at between 10 and 30 m depth fit quite well with the known cave and with the tilt angle of limestone layers measured both on the outcrops and in the cave itself. However, forward modelling was conducted, which showed that the cave itself could not be identified by ERI. What could be identified was the fracture that stretched from the top of the cave to the surface, and which created an identifiable resistivity anomaly.

3.3 Conclusion

MRS is a useful tool which could have a place in the hydrogeologist's toolbox for karst applications because it can estimate the spatial variations of permeability and transmissivity which underline karstic structures bearing water (such as epikarst, conduits and caves). However, the groundwater quantity has to be enough to send out a signal which can be measured. For practical purposes, the hydrogeologist can use Figure 5.85 to estimate if the geological target can be detected with MRS, i.e. if the groundwater quantity and depth lead to a measurable signal.

One should also keep in mind that the geometry of the three-dimensional structures can only

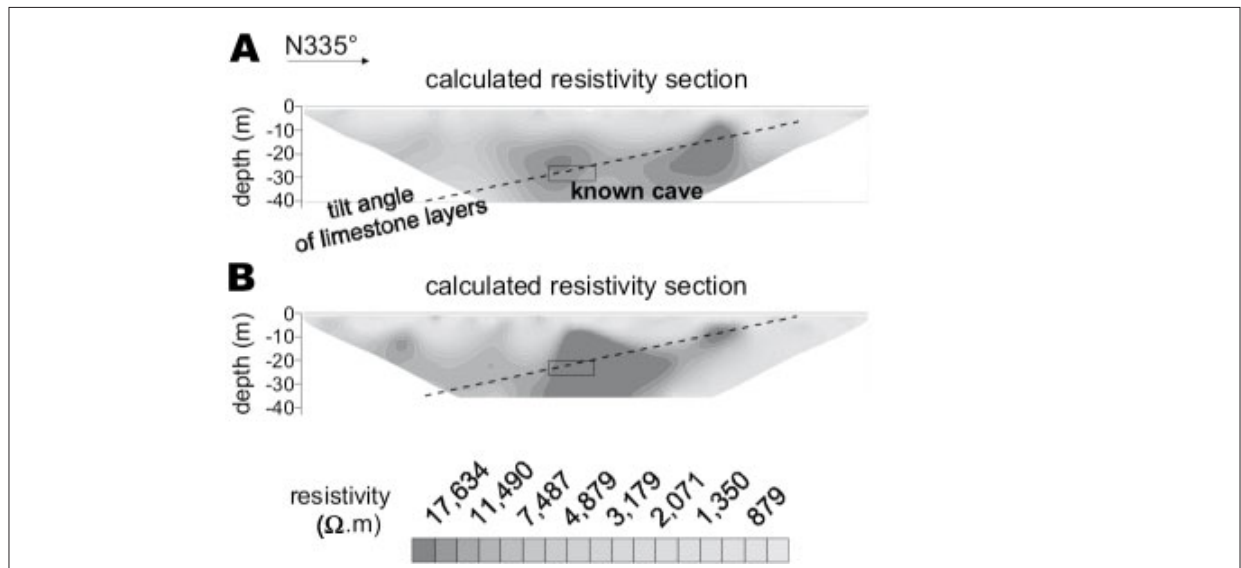


Figure 5.87: ERI interpretation.

A: Wenner array. B: dipole-dipole (Wenner beta).

be approximated with one-dimensional soundings.

For low magnetic signals, i.e. small amounts of water, or water at great depths, the measurement duration of a MRS can reach up to 20 hours (several hundreds of stacks): in a karstic environment, an average of one sounding per day should be planned.

In such a highly resistive context, ERI can measure the signal induced directly by the geological structures down to about 10 m (with the array used). Deeper dissolution structures such as a cave or conduit can be localised if they are associated with shallower anomalies such as fractures.

At the Lamalou site the ground surface is basically limestone. Therefore, to set up the electrodes properly, it was necessary to drill holes into the rocks and fill them with a clayey mixture to secure a proper contact. This was time-consuming, and an average of only one image could be completed per day.